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Concepts of seismic and sequence stratigraphy as outlined in publications since 1977 made a substantial impact on sedimentary geology. The notion that changes in relative sea level shape sediment in predictable packages across the planet was intuitively attractive to many sedimentologists and stratigraphers. The initial stratigraphic record of Mesozoic and Cenozoic depositional sequences, laid down in response to changes in relative sea level, published in Science in 1987 was greeted with great, albeit mixed, interest. The concept of sequence stratigraphy received much acclaim whereas the chronostratigraphic record of Mesozoic and Cenozoic sequences suffered from a perceived absence of biostratigraphic and outcrop documentation. The Mesozoic and Cenozoic Sequence Stratigraphy of European Basins project, which began officially with an international meeting in Dijon France in 1992, was designed to address the lack of documentation by inviting sedimentologists and stratigraphers to collectively build a documented chronostratigraphic and outcrop record of depositional sequences calibrated across a large number of basins in a geographically restricted area. The choice of Europe as a backdrop to this calibration and documentation effort is rooted in the philosophy that the cumulative stratigraphic data base for European Basins, which have been studied for over hundred years and are home to most Mesozoic and Cenozoic stage stratotypes, is uniquely suited for such a calibration project. European basins offer a variety of climatic provinces and their depositional systems range from siliciclastic systems in the northern part of the study area to carbonate dominated systems in the tethyan area. Sequence interpretations for a large number of European basins were presented at poster sessions in Dijon. Papers in this volume, many of them based on the Dijon posters, form an integral part of the sequence documentation presented here.

Sequence stratigraphy applies the inherent premise that eustasy represents a global signal among the variables that play a role in shaping depositional sequences. This global signal plays an essential role in shaping depositional sequences laid down in response to changes in relative sea level. Because of this global signal, bounding surfaces of depositional sequences (sequence boundaries at their correlative conformity) can be expected to be synchronous between basins. To demonstrate such synchronicity requires a very high stratigraphic resolution and a calibration of all stratigraphic disciplines. Therefore it was deemed essential to express the chronostratigraphic record of depositional sequences relative to standard, up to date, geochronologic scales. The Mesozoic chronostratigraphic framework of Gradstein et al. (1994, 1995) was sponsored by the project and for the Cenozoic the recent framework of Berggren et al. (1995) was selected for this project. These chronostratigraphic frameworks integrate state of the art data on standard stages, magnetostratigraphy and geochronology with high resolution biostratigraphy and are essential to calibrate the stratigraphic position of depositional sequence boundaries in the basins studied as part of this project.

To further stress the importance of well-calibrated chronostratigraphic frameworks for the stratigraphic positioning of geologic events such as depositional sequence boundaries in a variety of depositional settings in a large number of basins, the project sponsored a biostratigraphic calibration effort directed at all biostratigraphic disciplines willing to participate. The results of this biostratigraphic calibration effort are summarized on eight charts included in this volume.

This volume addresses the question of cyclicity as a function of the interaction between tectonics, eustasy, sediment supply and depositional setting. An attempt was made to establish a hierarchy of higher order eustatic cycles superimposed on lower-order tectono-eustatic cycles. Crustal events on a plate-tectonic scale are key factors for controlling timing and architecture of Major Transgressive-Regressive Cycles which are surprisingly synchronous across European basins. This synchronicity suggests these Major Transgressive-Regressive Cycles are caused by tectonic processes that effect the whole of the European craton and most probably affect the volume of oceanic basins as well. Transgressive-Regressive Facies Cycles are primarily caused by basin forming events and changes in sediment supply. The relative synchronicity of these cycles across Europe, although differences occur in some basins, suggests that regional tectonic development may have also have an eustatic component.

The composite stratigraphic record of higher order eustatic sequences shows a significant increase in the number of sequences identified in the various European basins. Entries on the new charts include a composite stratigraphic record of 221 sequence boundaries in the Mesozoic and Cenozoic compared to 119 sequences for the same interval identified by Haq et al. (1987, 1988). This increase reflects the number of investigators as well as the number of basins studied, especially in the Triassic and the Jurassic where the number of sequences identified more than doubled. The number of sequences in the Cretaceous nearly doubled, even though few studies addressed the lower Cretaceous interval. Increase in the number of sequences in the Cenozoic was smaller because parts of the Cenozoic were not restudied as part of this project. The stratigraphic position of sequence boundaries is in general greatly improved relative to the Haq et al., (1987, 1988) record because of the effort placed on biostratigraphic calibration as part of this project. The stratigraphic position of sequences in outcrop sections in the Mesozoic can often be determined to a specific ammonite zone. In subsurface sections stratigraphic positioning of sequence boundaries is much less constrained because of calibration uncertainties between different stratigraphic disciplines. Greatest difficulties in stratigraphic calibration were encountered in the upper Cretaceous Coniacian through lower Maastrichtian interval where sequence boundaries from North America are included on the Cretaceous chart because these could be calibrated to the North American ammonite zones included by Gradstein et al. (1994) in their Mesozoic time scale while none...
of the available stratigraphic disciplines in Europe could be satisfactorily calibrated to that timescale.

Since, no effort was devoted to quantification of falls and rises in relative sea level, no attempt was made to revise the coastal onlap curve and the derived eustatic curves of Haq et al. (1987, 1988). Most of the new Mesozoic-Cenozoic stratigraphic record of sequences is placed in the long term eustatic envelope of Haq et al. (1987, 1988). The middle Eocene to recent sequence record is placed in a short term oxygen isotope record of Abreu et al., (this volume). Below the middle Eocene short term eustasy is not indicated since no new quantitative information is available. Qualitative indications of magnitude (minor, medium and major) of sea level falls and rises are used instead. For comparison with the long term Mesozoic-Cenozoic eustatic envelope of Haq et al. (1987, 1988) a curve of inundated continental area (Ronov, 1994) and a long term eustatic curve based on oxygen isotopes for the Albian to recent interval (Abreu et al., this volume), are included.

We trust this volume will contribute to a further discussion of sequence stratigraphy and lead to a better understanding of this new paradigm. We feel the Sequence Stratigraphy of European Basins Project has been successful in its attempt to describe a good portion of the European Mesozoic and Cenozoic succession in a sequence stratigraphic context and improve the stratigraphic record of its bounding surfaces. Additional efforts focusing on other geographical regions is undoubtedly required to further demonstrate the global nature of these depositional sequences.

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REFERENCES

PART I
INTRODUCTION
INTRODUCTION

The chronostratigraphic charts presented in this paper are the result of an initiative by Peter Vail and Thierry Jacquin in 1990 to analyze and document depositional sequences in European basins and to record their stratigraphic position relative to a state-of-the-art temporal framework accurately calibrated to a biostratigraphic framework. The “Mesozoic-Cenozoic Sequence Stratigraphy of European Basins” project started officially with a meeting in Dijon, France, organized by Jacquin, de Graciansky, and Vail, in May 1992. Sequence interpretations for a large number of European basins were presented at poster sessions in Dijon. Papers in this volume, many of them based on the Dijon posters, form an integral part of the sequence documentation for the chronostratigraphic charts.

Work on the detailed chronostratigraphic charts for the Mesozoic and Cenozoic began eighteen months before the Dijon Meeting, in December 1990 in Paris with a planning meeting attended by a large number of specialists in a wide range of biostratigraphic disciplines from several European countries. At the Paris meeting, all specialists present were invited to participate in the calibration of fossil groups representing non-marine, shallow- and deep-water depositional environments to a revised temporal framework. Invitations were extended to specialists not present at the Paris meeting to complement the expertise in fossil groups essential to the construction of a stratigraphic framework and to the calibration of sequences. Progress was reviewed at workshops in Paris in May and December 1991 and a preliminary biochronostratigraphic framework calibrated to the Gradstein et al. (1987) time scale was presented at the Dijon Conference in 1992. After completion of the Gradstein et al. (1994) Mesozoic time scale and the Berggren et al. (1995) Cenozoic time scale, all biostratigraphic, isotopic stratigraphic and sequence stratigraphic entries were recalibrated to the new time scales.

SEISMIC STRATIGRAPHY/SEQUENCE STRATIGRAPHY

Mitchum et al. (1977) described the depositional sequence as a basic unit for stratigraphic analysis with chronostratigraphic significance. They defined the depositional sequence as follows: “A depositional sequence is a stratigraphic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities or their correlative conformities.” This definition adds the concept of the “correlative conformity” to the unconformity-bounded sequence in the sense of Sloss (1963). Adding the “correlative conformity” to the sequence definition is essential to allow application of sequence stratigraphy in areas of continuous deposition. Even though Mitchum et al. (1977) discussed the chronostratigraphic significance of their sequence, they defined the sequence as a lithologic unit (“A depositional sequence is determined by a single objective criterion, the physical relations of the strata themselves.”) They stopped,
however, short of defining a sequence chronostratigraphic unit even though they defined a geochronologic unit sechron as: “the maximum interval of geologic time occupied by a given depositional sequence, defined at the points where the boundaries of the sequence change laterally from unconformities to conformities along which there is no significant hiatus.”

Here we simplify the lithologic definition of the sequence by Mitchum et al. (1977) as follows: A depositional sequence is a lithologic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities and their conformable equivalents.

We also define the chronostratigraphic unit or sequence chronozone as follows: A sequence chronozone comprises all strata deposited globally during the timespan of a sequence measured at the composite conformity where the bounding unconformities become conformable. A sequence chronozone can thus be viewed as a chronostratigraphic unit which includes all rocks deposited globally during the elapsed time between successive falls in relative sea level.

The geochronologic unit sechron defined by Mitchum et al. (1977) could be simplified as: A sechron spans the total interval of geologic time during which a sequence is deposited.

Sequences and subsequences of Sloss (1963), equivalent to megasequences and supersequence sets of Haq et al. (1987, 1988) are major tectono-eustatic units shaped by plate tectonic events that affect longer term eustatic sea level. Even supersequence boundaries of Haq et al. (1987, 1988), correlate well with times of major changes in plate spreading rate and direction (Ross, 1995). Higher frequency (3rd-order) sequences of Mitchum and Vail (1977), Haq et al. (1987, 1988) are shaped primarily by the interaction of sea-level changes with sediment supply, against the backdrop of basin subsidence. Subsidence/uplift is controlled by complex local and regional tectonic factors and is expected to differ from place to place; eustasy, however, represents a global signal. The higher frequency of a glacio-eustatic signal holds promise for high-resolution global stratigraphic correlation, provided its signal can be reliably deduced from the sediment record. The likely mechanism behind these higher frequency sea-level changes is, in the Eocene to recent interval, almost certainly glacio-eustasy (Miller et al., 1987, 1991a), Abreu et al. (this volume), Abreu and Haddad (this volume), Abreu and Anderson (in press). For higher frequency sea level changes prior to the Eocene, Abreu et al. (this volume) postulate the possibility of glacial episodes during the Aptian and Maastrichtian although the Cretaceous and Paleocene glacial history remains largely unknown.

Sequence stratigraphy evolved from seismic stratigraphy (Vail et al., 1977), when the realization was made that packages of sediments observed on reflection seismic data could also be identified in wells and outcrop sections. Stratigraphers and sedimentologists seized on this opportunity that opened new dimensions to their respective disciplines. Stratigraphers sensed the enormous potential of a high frequency eustatic signal for global stratigraphic correlation and focussed on the chronostratigraphic position of the bounding surfaces. Haq et al. (1987) proposed a chronostratigraphic record of Mesozoic and Cenozoic sequences, mostly based on the temporally well-constrained classic stage type and reference sections in Europe.

This record expanded on the Vail et al. (1977), uppermost Triassic to Pleistocene, chronostratigraphic record of depositional sequences identified on seismic sections and dated with available well control.


The principal focus of this paper is to revisit the stratigraphic aspects of sequence stratigraphy and provide a chronostratigraphic record of sequence boundaries to complement the Haq et al. (1987) sequence record with new data provided by the contributors to the “Mesozoic-Cenozoic Sequence Stratigraphy of European Basins” project.

**Sequence Boundaries and Correlative Conformities**

Terrigenous sediments transported offshore accumulate relatively close to the basin margin and are shaped in packages (sequences or systems tracts) bounded by surfaces (sequence boundaries) as a response to the principal variables of sediment supply, subsidence and eustasy. Farther offshore the influence of terrigenous sedimentation decreases and a more pelagic, but not necessarily continuous, sedimentation dominates in which the sequence and systems tract packages and their bounding surfaces are often not well expressed and sequence boundaries become correlative conformities. In any given section a sequence boundary may be deduced from changes in lithofacies across physical surfaces (subaerial-erosional truncation surfaces and flooding or transgressive surfaces of onlap) and from vertical facies relationships (downward shift) Van Wagoner et al. (1990). In basinal settings, where changes in lithofacies are subtle, sequence boundaries or their correlative conformities may be identified from biotic analysis, well logs and/or geochemical analyses. In theory, the chronostratigraphic position of a sequence boundary is determined at the point where the bounding unconformity becomes conformable. The chronostratigraphic position can only be determined by comparing its stratigraphic position with other well-calibrated stratigraphic disciplines either biostratigraphy, magnetostratigraphy, chevron stratigraphy or preferably, a combination of those disciplines. In practice, the correlative conformity may not be recognizable in outcrop and the stratigraphic position of a sequence boundary is determined by choosing a section where lowstand deposits are developed, and the sequence boundary is identified within a biostratigraphic zone of a fossil group with high-stratigraphic resolution. Comparing the stratigraphic position of a sequence boundary in different settings in different basins will eventually reveal the stratigraphic position of the correlative conformity.

In outcrops along slowly subsiding margins with moderate sedimentation rates, prevalent in many basins of western Europe, the most often recognized surface is the combined sequence boundary and subsequent flooding (transgressive) surface. Most standard stage type sections located in passive-margin settings, have a transgressive surface as their lower boundary. However, the lowstand portion of the sequence and unknown portions of the previous highstand and transgres-
sive deposits are missing in that position, but should be present farther down-dip in the basin. In the North Sea Basin for example, transgressive deposits and highstand deposits are found along the basin margins and lowstand deposits are concentrated in the deeper parts of the basin. When ocean drilling established a composite stratigraphic record for the Cretaceous and Cenozoic, hiatuses in the onshore standard record were the source of considerable debate on the placement of stage boundaries. Currently, the Global Boundary Stratotype Section and Point (GSSP) effort by the Commission on Stratigraphy is underway to define stage boundaries in settings where sedimentation is continuous across stage boundaries.

1987 CHRONOSTRATIGRAPHIC SEQUENCE RECORD

The stratigraphic record of Mesozoic and Cenozoic depositional sequences, calibrated to a temporal framework presented by Haq et al. (1987), is based on the sequence stratigraphic premise that deposition is controlled by the principal variables of subsidence/uplift of the basin floor, sediment supply and eustasy (Hardenbol et al., 1981; Jervey, 1988). Subsidence/uplift, controlled by tectonics on a plate tectonic to basin scale, and sediment supply, controlled by tectonics and climate, is expected to differ between basins or even parts of the same basin. Eustasy, on the other hand, whether caused by volume changes of oceanic basins or by sequestering of water in the form of continental ice and in inland seas and lakes, is controlled by tectonics and climate as well, but its effects are global. This global effect, recognizable in the rock record, represents a synchronous stratigraphic signal. Haq et al. (1987) presented a stratigraphic record of hundred nineteen Mesozoic and Cenozoic sequences and their relative onlap calibrated to a temporal scale which expanded on the Vail et al. (1977), uppermost Triassic to Pleistocene, chronostratigraphic record of depositional sequences. Haq et al. (1987) identified considerable more sequences in outcrop than were identified by Vail et al. (1977) from seismic records. Sequence resolution is a function of local sedimentation rates but is often lower on seismic records than in outcrop sections deposited at similar rates. In general, sequence resolution increases in the direction of depocenters. Establishment of the temporal position of sequence boundaries identified from seismic records requires well control in sediments conducive to reliable chronostratigraphic analysis.

Haq et al. (1987) placed shifts in coastal onlap and changes in sea level in three categories of relative magnitude: major, medium and minor (determined from seismic and outcrop records). Short-term sea-level changes derived from relative onlap and magnitude were expressed within an envelope of long-term sea-level change. The long-term sea-level envelope was then calibrated to its highest position of about 260 m in the early Turonian (Kominz, 1984; Pitman, 1978) and modern sea level at 60 m (which assumes no icecaps). In addition, sequences were tentatively ordered in a hierarchical system of 1st-order megasequences nearly identical to the sequences proposed for the North American craton by Sloss (1963, 1988), and 2nd-order supersequences which are subsequences of Sloss (1963, 1988). Second-order supersequences sets and 3rd-order sequences do not have Sloss (1963, 1988) equivalents. Sequences with lowstand submarine fans were indicated as type 1 and all others as type 2 sequences.

Most sequences in the Mesozoic record of Haq et al. (1987) were calibrated to the temporal scale through first-order calibrations to ammonite biostratigraphy. The limited number of radiometric dates in the Mesozoic, prior to the upper Cretaceous, are often not precisely calibrated to ammonite zones but to sub-stages. To subdivide stages, ammonite zones within a stage were allotted equal duration. Few of the other fossil groups on the Mesozoic portion of the Haq et al. (1987) record represent first-order correlation with ammonite zones. Sequences in the Cenozoic portion of Haq et al. (1987) were calibrated to the temporal scale through first- and second-order calibrations with an integrated framework of planktonic foraminifera and calcareous nannofossils.

To facilitate the calibration of sequences to the temporal framework, Haq et al. (1987) focused on extensively studied and biostratigraphically documented stage type and reference sections in Europe. Many type sections are selected in deposits laid down in shallow-marine environments and facies changes across sequence boundaries are rather well expressed, although lowstand deposits are often absent.

MESOZOIC-CENOZOIC CHRONOSTRATIGRAPHIC CHARTS

Temporal Framework

Developments in geochronology since the publication of the Haq et al. (1987) Mesozoic-Cenozoic time scale, such as new 40Ar/39Ar dates for the upper Cretaceous of the North American Western Interior Seaway (Obradovich 1993) and the selection and dating of a boundary stratotype (GSSP) for the Eocene-Oligocene boundary, rendered all published time scales out of date, at least to some extent. For the Cenozoic, a new integrated time scale (Berggren et al., 1995) was made available for the calibration of the Cenozoic bio- and sequencechronostratigraphic record. In order to incorporate new 40Ar/39Ar dates (Obradovich 1993) for the upper Cretaceous and integrate new magnetostratigraphic and bio-stratigraphic calibrations, a separate time scale effort was initiated which resulted in an improved Mesozoic time scale (Gradstein et al., 1994).

Cenozoic Time Scale

The Cenozoic time scale (Berggren et al., 1995) integrates an extensive DSDP/ODP record on magnetostratigraphy, planktonic foraminifera and calcareous nannofossil biostratigraphy and standard stratigraphy with selected radiometric dates to produce a well-calibrated temporal framework (see appendix). Sequences are positioned relative to the Berggren et al. (1995) temporal framework primarily with calcareous nannofossils and planktonic foraminifera (Chart 2). The calibration of fossil groups to this integrated framework (Chart 3) is not documented in this volume and is the responsibility of the coordinator(s) for that particular fossil group. Manuscripts with biostratigraphic documentation submitted by coordinators are or will be published in the Bulletin de la Société Géologique de France: Larger Foraminifera (Cahuzac and Poignant, 1997; Serra-Kiel et al., 1988 in press). Brief summaries submitted by several coordinators are, because of space constraints, included in an appendix.

Mesozoic Time Scale

The Mesozoic time scale (Gradstein et al., 1994) integrates standard stratigraphy, magnetostratigraphy and ammonite biostratigraphy with high-temperature radiometric dates to produce an updated temporal framework (see appendix). The composite ammonite zonation of Gradstein et al. (1994) is, except for the
upper Cretaceous, based on highest resolution zonal or subzonal subdivisions from tethyan and boreal areas provided by coordinators for ammonite biostratigraphy of the Sequence Stratigraphy of European Basins Project. For the upper Cretaceous, Gradstein et al. (1994) used the high-resolution ammonite zonation of the Western Interior Seaway Basins in North America (Cobban et al., 1994; Obradovich, 1993) because the $^{40}$Ar/$^{39}$Ar radiometric dates of Obradovich (1993) were directly calibrated to the North American ammonite record. Calibration of the North American ammonite record to the standard stages and the European ammonite record remains tentative. Sequences are calibrated to the Gradstein et al. (1994) temporal framework primarily with ammonites (Charts 4, 6, 8).

Calibration of fossil groups, provided by coordinators, to the temporal framework of Gradstein et al. (1994) is not documented in this paper. The provided information is plotted on (Charts 5, 7, 8). Some coordinators submitted manuscripts with biostratigraphic documentation; these will be, or are already published in the Bulletin de la Société Géologique de France: Jurassic calcareous nanofossils (De Kaenel et al., 1996), Cretaceous benthic foraminifera (Magniez-Jannin, 1995), Jurassic dinoflagellates (Riding and Ioannides, 1995), Mesozoic-Cenozoic charophytes (Riveline et al., 1996), Cretaceous planktonic foraminifera (Robaszynski and Caron, 1995) and Triassic ammonoids (Mietto and Manfrin, 1995). Jurassic Brachiopods (Almeras et al., 1994) appeared in Geobios. Cariou and Hantzpergue (1997) coordinated an effort of the “Groupe Français d’Étude du Jurassique” to improve stratigraphic calibration of many of the same fossil groups addressed in this study. Summaries submitted by several coordinators are included in an appendix.

Sequence Record

The primary objective of this volume is to provide a state-of-the-art stratigraphic record of sequences identified as part of the Mesozoic-Cenozoic Sequence Stratigraphy of European Basins project. Independent records of sequences in tethyan and boreal basins calibrated to their respective ammonite zonations are summarized on the Mesozoic sequence chronostratigraphic charts from the base of the Triassic through the Turonian (Charts 4, 6, 8). Even though a comparable number of sequences were identified in the tethyan and boreal basins, synchronicity can only be demonstrated with the help of independent stratigraphic tools. In the Jurassic, ammonite records between boreal and tethyan basins considered for this project are much better calibrated than in the Triassic or lower Cretaceous. As a result, sequence records for boreal basins resemble those for tethyan basins closely in most of the Jurassic but the agreement is not as close in the Triassic. The lower Cretaceous interval shows major gaps in the sequence record because fewer papers were submitted while the calibration between boreal and tethyan ammonite zonations is much more tentative.

Jacquin and de Graciansky (this volume) identify four “Major Transgressive-Regressive Cycles” (MTR cycles) in the Mesozoic e.g., Eastern Tethys Cycle (Triassic), Ligurian and North Sea Cycles (Jurassic) and North Atlantic/Biscay Cycle (Cretaceous). Boundaries between MTR cycles do not coincide with system or series boundaries. In the Cenozoic, two additional unnamed MTR cycles are identified but not named. MTR cycles reflect the response of the western portion of the Eurasian plate to major plate tectonic phases in the opening of the Atlantic Ocean (Ziegler, 1990). These major tectonic phases affect the volume of ocean basins and hence global sea level and thus produce synchronous tectono-eustatic MTR cycles which are essentially identical for tethyan and boreal basins. Differences in sediment supply or correlation problems between tethyan and boreal ammonite zonations could explain the offset of the start of the regressive phase, as is the case in the middle Triassic, lower and upper Jurassic.

Jacquin and de Graciansky (this volume) also introduce the concept of “Transgressive-Regressive Facies Cycles” (TRF cycles), which describe sediment response to basin forming events resulting from regional and more local tectonic activity. The resulting tectono-eustatic effects are still producing a number of synchronous TRF cycles across Europe, but exceptions caused by local tectonics are rather ubiquitous as suggested by the numerous differences between boreal and tethyan basins on the sequence chronostratigraphic charts. Gianolla and Jacquin (this volume) describe the evolution of the principal TRF cycles (1 to 4) in Triassic basins from the Alps to the Barents Sea. Jurassic TRF cycles (4 to 6) are documented by de Graciansky et al. a, b (this volume), TRF cycles (7 to 10) by Jacquin et al. (this volume). Lower Cretaceous TRF cycles (11 to 15) are summarized from European basins by Jacquin et al. (this volume). Triassic (Chart 8), Jurassic (Chart 6) and Cretaceous (Chart 4) sequence chronostratigraphic charts carry the Major Transgressive-Regressive Cycles proposed by Jacquin and de Graciansky (this volume) although their numbering system is not used on the charts. Their Transgressive-Regressive Facies Cycles are included on the Triassic, Jurassic and lower Cretaceous charts as well. TRF cycles in the Cenomanian, Turonian and Maastrichtian are based on outcrop records in northwestern Europe and the tethyan area, whereas Coniacian through Campanian TRF cycles are based on the Gulf Coast outcrop record (modified from Young, 1986). TRF cycles on the Cenozoic sequence chronostratigraphic chart are based on outcrop records of stage type areas in Europe.

As in Haq et al. (1987), sequences and subsequent flooding events are placed in three categories of relative magnitude: major, medium and minor. No attempt was made to organize sequences in a hierarchy of different orders of cyclicity even though some of the authors in this volume mentioned a hierarchy in their individual papers. A better understanding of the underlying mechanism and an independent measure of magnitudes are required before any hierarchical classification is justified. No distinction is made between Type 1 and Type 2 sequences, because local subsidence cannot be easily distinguished from the eustatic signal. Submarine fans are identified for essentially all sequences in the Paleocene and lower Eocene of the central North Sea Basin (Neal et al., this volume). Since no effort was devoted to quantification of falls and rises in relative sea level, no attempt was made to revise the coastal onlap curve and the derived eustatic curves of Haq et al. (1987, 1988). The new Mesozoic-Cenozoic stratigraphic record (except middle Eocene to recent) of sequences (Chart 1) is placed in the long term eustatic envelope of Haq et al. (1987). The middle Eocene to recent sequence record is placed in a short term oxygen isotope record of Abreu et al. (this volume).
the middle Eocene short term eustasy is not indicated since no new quantitative information is available. Qualitative indications of magnitude (minor, medium and major) of sea level falls and rises are used instead. For comparison with the long term Mesozoic-Cenozoic eustatic envelope of Haq et al. (1987) a curve of inundated continental area (Ronov 1994) is shown on Chart 1. In addition, a long term eustatic curve based on oxygen isotopes for the Albain to recent interval (Abreu et al. this volume), is added in Chart 1.

The sequence stratigraphic entries on the new charts include a composite stratigraphic record of 221 sequence boundaries in the Mesozoic and Cenozoic. Haq et al. (1987) listed 119 sequences for the same interval. The increase in the number of sequences reflects the increase in the number of investigators as well as the number of basins studied, especially in the Triassic and the Jurassic where the number of sequences identified more than doubled. The number of sequences in the Cretaceous nearly doubled, even though few studies addressed the lower Cretaceous interval. In the upper Cretaceous (Coniacian through Campanian) sequence boundaries identified in boreal and tethyan basins could not be calibrated reliably to the temporal framework. Instead, for the Coniacian through lower Maastrichtian interval, a record of sequence boundaries from North America is included on the chart which could be calibrated to the North American ammonite zones included by Gradstein et al. (1994) in their Mesozoic time scale. Sequences of Haq et al. (1987) in the Coniacian through lower Maastrichtian interval were also based on the North American record and were tentatively calibrated to the standard stages. The increase in the number of sequences in the Cenozoic was smaller because parts of the Cenozoic were not re-studied as part of this project, and the Cenozoic was already studied in more detail by Haq et al. (1987). For comparison, sequences of Haq et al. (1987) are included on the charts calibrated to the new chronostratigraphic record.

Individual sequence boundaries are identified on the new charts by the first two to four letters of the name of the stage in which the sequence boundary is identified and numbered from old to young. For example, C1e1 represents the first sequence boundary in the lower Cenomanian and is situated within the mantlel ammonite zone. The next sequence boundary is C2e in the uppermost dixoni ammonite zone. The Cenomanian deposits below C1e are in sequence A11 which has its lower bounding sequence boundary in the uppermost dispar ammonite zone in the Albanian. If additional sequence boundaries were to be identified between sequence boundaries C1 and C2 those could be identified as C1e1, C1e2, an additional sequence boundary below C1e1 could be identified as C0e, etc.

Calibration of Sequence Boundaries, Bio-zonations and Isotope Data

Calibration of sequence boundaries to a temporal framework requires a stratigraphic discipline with a high resolution. Ammonite biostratigraphy represents the best calibrated, highest resolution stratigraphic discipline in the Mesozoic interval of the European basins studied. Ammonites are ubiquitous in the sedimentary record of many European basins and are extensively studied. Ammonite subdivisions are also well calibrated to the standard stages because they were traditionally included in their definition. Therefore, sequence boundaries on the Mesozoic charts are calibrated to ammonite zones or subzones and can be calibrated from basin to basin as long as the same ammonites are present. Unfortunately, ammonite assemblages differ from basin to basin as a function of the biogeographic provinces in which the basins are located. Calibration of ammonite zonations for different biogeographic provinces is a cooperative process and is still in progress. To preserve apparent differences in stratigraphic position of sequence boundaries between boreal and tethyan basins, all sequence boundaries, from the base of the Triassic to the top of the Turonian, are calibrated separately to ammonite records for boreal and tethyan provinces. In intervals with good agreement in ammonite calibration between boreal and tethyan provinces (Sinemurian through middle Oxfordian and Cenomanian through Turonian), sequence boundaries agree better than in intervals where differences in ammonite calibration are more pronounced (Triassic, upper Oxfordian through Tithonian and much of the Cretaceous). Other factors affecting agreement in sequence calibration are geographic distance between basins, the number of available studies (lower Cretaceous), the way ammonite zones are defined, hiatuses in shallow-water sections and the decision whether an ammonite appearance or disappearance is biozonal or chronozonal.

Synchronicity of sequence boundaries can only be demonstrated in the presence of high-resolution stratigraphic methods. In field observations, sequence boundaries can be positioned either within or at the boundary between ammonite zones. Those positioned at zonal boundaries are especially subject to further scrutiny of the completeness of the stratigraphic record at that location. Cenomanian sequence boundary C3 appears to fall between the Mantelliceras dixoni and Acanthoceras rhotomagese ammonite zones on the platform in the type area of the Cenomanian in France. The sequence boundary coincides, however, with the transgressive surface, and the lowstand deposits are not present on the platform. In basins where a lowstand is developed the sequence boundary occurs in the uppermost dixoni ammonite zone in the boreal realm (Robaszynski et al., this volume). However, in a tethyan realm (Robaszynski et al., 1993), the genus Mantelliceras persists to the sequence boundary but the first representatives of the genus Acanthoceras appear later in the lowstand deposits. The interval without Mantelliceras nor Acanthoceras was placed in a new Cunningtoniceras inerm zone and sequence boundary C3 was placed at the base of that zone. The evolutionary appearance of the planktonic foraminifer Rotalipora reicheli just below or just above the C3 sequence boundary and its disappearance close to the subsequent maximum flooding surface in sections in Tunisia, northwestern and southeastern France provides additional biostratigraphic evidence that sequence boundaries in this example are synchronous.

Calibration of the upper Cretaceous sequence boundaries identified in European basins to the temporal framework and to the North American ammonite record is relatively well understood for the Cenomanian and Turonian Stages but proved to be a challenge for the Coniacian through lower Maastrichtian interval. Western Interior seaway ammonite assemblages are mostly endemic and have very few counterparts among the European upper Cretaceous ammonites. The incomplete ammonite record in the type areas and the lack of calibration between the North American ammonite record and “cosmopolitan” fossil
groups such as planktonic foraminifera and calcareous nannofossils, precludes the calibration of sequence boundaries identified in European basins in the Cenozoic through Campanian interval. A record of North American sequence boundaries identified in the Gulf Coast area and calibrated to Western Interior seaway ammonite zones is included instead.

Stratigraphic calibration of sequence boundaries in the Cenozoic presents a very different challenge. Planktonic foraminifera and calcareous nannofossils are the fossil groups best suited for long-distance calibration. Both groups prefer low latitude and relatively deep water paleoenvironmental settings. Many of the basins studied for the MCSSEB project represent rather shallow, middle latitude basins (North Sea Basin, Pannonian Basin) in which the record of planktonic foraminifera and calcareous nannofossils is incomplete. Tethyan basins such as the Piedmont Basin in northern Italy and the western Pyreneen and Tremp basins in northern Spain provided a more complete record.

Oxygen and Strontium Isotopes

Chemostratigraphy is evolving rapidly into an independent discipline in stratigraphy. Strontium isotope data from published sources are included on the Cenozoic, Cretaceous and Jurassic Sequence Chemostratigraphic charts (Charts 2, 4, 6) to provide an additional discipline for stratigraphic calibration. Oxygen isotope data are included on the Cenozoic and Cretaceous charts (Abreu et al., this volume) and represent an additional approach to determine the stratigraphic position and magnitude of sea-level changes, especially if the case can be made that the observed fluctuations in the Paleogene and Cretaceous isotope records reflect changes in ice volume (Abreu et al., this volume).

Strontium isotope values calibrated to other chronostratigraphic records are available from the literature for the Jurassic through Cenozoic interval. Unfortunately, there are too few stratigraphically well-constrained strontium isotope data for the Triassic to justify including them on the Triassic chart (Chart 8). Strontium isotope ratios on the Jurassic chart (Chart 6) are derived from Jones et al. (1994a, b). These data are precisely located in standard ammonite zones in measured sections in Great Britain, so that they could be readily calibrated to the chronostratigraphic framework on the chart. Data for the lower Cretaceous interval, derived from Jones et al. (1994b) and calibrated to boreal ammonite zones in Great Britain, cannot be calibrated as precisely owing to the tentative nature of the correlation between boreal ammonite zones and tethyan standard zones on the chart. Strontium isotope data in the upper Cretaceous are primarily from the work of McArthur et al. (1994) in the Western Interior of North America. These data are calibrated to Western Interior ammonite zones which are included on the chart. Precise positions within zones are not available and data are averaged by ammonite zone and plotted at the midpoint of the zone. These upper Cretaceous data are supplemented in the upper Campanian and Maastrichtian with data derived from magnetostratigraphically-constrained ODP sites from the work of Barrera (1994), Barrera et al. (in press) and Sugarman et al. (1995). Cenozoic data for the Paleocene and lower Eocene are from Hess et al. (1986). Data for the Lutetian to the present are from Miller et al. (1988), Oslick et al. (1994), Mead and Hodell (1995) and Farrell et al. (1995). These data are derived from ODP sites and the original calibration to calcareous nannofossil biostratigraphy, oxygen isotope stratigraphy or magnetostratigraphy was recalibrated to the temporal framework on the Cenozoic Sequence Chemostratigraphic chart (Chart 2) as appropriate.

Strontium ratios on all three charts are adjusted to a single standard where NIST-987 is 0.710250. The Cenozoic Sequence Chemostratigraphic chart (Chart 2), includes a composite smoothed oxygen isotope curve for the entire Cenozoic compiled from Abreu and Haddad (this volume) and Abreu and Anderson (in press). The composite smoothed isotope curves of Abreu and Haddad (this volume) and Abreu and Anderson (in press), simulate a sea-level curve from the Cretaceous-Tertiary boundary to the recent.

The Cretaceous Sequence Chemostratigraphic chart (Chart 4) includes a smoothed (7 points least square method) isotope record based on bulk rock samples from Cenomanian through lower Campanian outcrops in England (English Chalk) and Italy (Gubbio) (Jenkins et al., 1994) and an upper Campanian to Maastrichtian record from central Tunisia (Abreu et al., this volume). The Cretaceous chart also includes an Aptian through Maastrichtian isotope record of deep water benthic foraminifera compiled from published data (Abreu et al., this volume).

Composite oxygen isotope curves from the Aptian to the present (Abreu et al., this volume) show the lightest values in the lowermost Turonian and a gradual change towards the heavier values of the Quaternary. The long-term trend in the upper Cretaceous and Cenozoic oxygen isotope record towards more positive values is explained by progressive cooling and glaciation at the poles (Savin et al., 1975). Rather than a continuous process, the long-term cooling seems to be made up of several shorter-term steps in the isotope values that can be related to changes, either in ice volume or in bottom water temperatures (Abreu et al., this volume; Abreu and Anderson in press). The long-term evolution in the oxygen isotope values mimics the change in long term sea level proposed by Haq et al. (1987). Higher frequency shifts in the oxygen isotope record are proposed as proxy indicators for glaciation and sea-level fluctuations. Abreu and Haddad (this volume) demonstrate a strong stratigraphic relationship between higher frequency shifts in the oxygen isotope record and sequences proposed from the rock record. Oxygen isotope curves may well provide an independent method for stratigraphic calibration of major eustatic changes (Miller et al., 1987) and demonstrate synchrony of depositional sequences on different continents.

CENOZOIC SEQUENCE CHRONOSTRATIGRAPHIC RECORD

The Cenozoic in Europe consists of two “Major Transgressive-Regressive Cycles” (Chart 2) controlled by steps in the opening of the Atlantic Ocean (Ziegler, 1990). The opening of the Atlantic (Reykjanus) and the failed rifting of the North Sea resulted in a major transgressive phase in the upper Paleocene and lower Eocene. The middle Eocene through lower Oligocene represents an overall regressive phase. A second transgressive episode from the upper Oligocene to the middle Miocene is related to the opening of the North Atlantic. The Neogene from the middle Miocene to the present is mainly regressive. Basin-forming events in the Cenozoic of Europe are controlled by...
episodes in the opening of the Atlantic Ocean and the resulting compression between Europe and Africa. Eight “Transgressive-Regressive Facies Cycles” are identified from outcrop records of stage type areas in Europe (the regressive early Paleocene is still part of the late Cretaceous “Major Regressive Cycle” and the late Maastrichtian “Regressive Facies Cycle”).

Eleven papers on the Cenozoic sequence stratigraphic record submitted for publication in this volume permit a substantial revision of the Haq et al. (1987) record in the Paleocene through lower Eocene, the Oligocene through middle Miocene and the Plio-Pleistocene intervals. The middle through upper Eocene and the upper Miocene are unchanged from Haq et al. (1987). The Paleocene to lower Eocene stratigraphic record of sequences is as the Haq et al. (1987) record based on the southern onshore North Sea Basin sections in southern England and Belgium (Neal et al., this volume; Vandenberghe et al., this volume) with seismic stratigraphic support from the offshore central North Sea basin. The Oligocene through lower Miocene sequence record is now calibrated to the Pannonian and Piedmont Basins (Vakarc et al., this volume; Gnaocclini et al., this volume; Haq et al., 1987) based their record for this interval primarily on the southern North Sea Basin (Belgium) and the Aquitaine Basin (France). The middle Miocene record is also calibrated to the Piedmont and Pannonian Basins (Vakarc et al., this volume; Gnaocclini et al., this volume) whereas Haq et al. (1987) is primarily based on the Piedmont Basin record. The Plio-Pleistocene record is as Haq et al. (1987), calibrated to the Calabrian and Sicilian deposits in Italy supplemented with offshore Gulf of Mexico data. Sequences at or near stage boundaries are identified with both stage prefixes to allow for future changes in the definition of stage boundaries as a result of the ongoing Global Boundary Stratotype and Point (GSSP) effort of the International Committee on Stratigraphy (ICS). Introductions to the Neogene (Vandenberghe and Hardenbol, this volume) and Paleogene (Neal and Hardenbol, this volume) summarize the papers submitted for the Cenozoic chapter of this volume and represent the principal documentation for the Cenozoic sequence chronostratigraphic record.

**CRETACEOUS SEQUENCE CHRONOSTRATIGRAPHIC RECORD**

The Cretaceous in western Europe is characterized by one Major Transgressive-Regressive Cycle named the North Atlantic/Biscay Cycle. Jacquin and de Graciansky, this volume. The earliest Cretaceous (Berriasian) represents the continuation of the regression that started near the Kimeridgian/Tithonion boundary. The onset of the opening of the North Atlantic (Ziegler, 1990) marks the beginning of an overall transgressive phase that continues until the early Turonian and is followed by an overall regression that lasted into the early Cenozoic. Transgressive-Regressive Facies Cycles (TRF cycles) which describe sediment response to basin-forming events of a more local significance punctuate these Major Transgressive-Regressive Cycles (MTR cycles). Jacquin et al., (this volume) describes five TRF cycles (11 to 15) in the lower Cretaceous portion of the (North Atlantic/Biscay, MTR cycle, Jacquin and de Graciansky, this volume). TRF cycles in the upper Cretaceous are discussed below.

**Upper Cretaceous Sequences**

The upper Cretaceous introduced by Hardenbol and Robaszynski (this volume) summarizes the sequence stratigraphic information contained in five papers submitted for this part of the volume. The chronostratigraphic record of sequence boundaries in the upper Cretaceous is well calibrated in the Cenomanian and Turonian. Cosmopolitan ammonite assemblages in the Cenomanian and Turonian facilitate calibration between basins in different paleogeographic settings. As a result the record of sequence boundaries is better calibrated in the Cenomanian and Turonian than in any other Cretaceous interval. Cenomanian and Turonian deposits in western Europe suggest two Transgressive-Regressive Facies Cycles. The first TRF cycle begins at the Albian/Cenomanian boundary and includes the early Cenomanian. The second TRF cycle starts close to the base of the middle Cenomanian and includes the remainder of the Cenomanian and the entire Turonian. Cenomanian and Turonian sequences and TRF cycles on the Cretaceous chart (Chart 4) are based on records for tethyan and boreal areas described in Robaszynski et al. (this volume), Robaszynski et al., 1990, 1993.

The chronostratigraphic record of sequence boundaries in the Coniacian through Maastrichtian interval of European sections is poorly established. In contrast to the Cenomanian and Turonian the stratigraphic record of sequence boundaries for the Coniacian through Maastrichtian interval is the least calibrated of the entire Mesozoic-Cenozoic chronostratigraphic framework. Reliable first-order calibration between Campanian and Maastrichtian standard stages and biostratigraphic zonations, based on more cosmopolitan groups such as ammonites, planktonic foraminifera and calcareous nannofossils, are essentially non-existent. Even second- and third-order calibrations are scarce. Campanian/Maastrichtian strata in the boreal type areas of western Europe are mostly shallow-water deposits and do not contain diagnostic planktonic foraminifera and calcareous nannofossils. Outcrops are scattered over wide areas and assembling a composite section is problematic. Ammonites are scarce in outcrop and most of our current understanding is from a compilation of historical ammonite information from museum collections (Kennedy 1986). Ammonites suggest the lower and upper Campanian to be present. However, there seems to be no record for deposits between *Bostrychoceras polyplectum* and *Nostoceras hyatti* which in North America spans a period of 6–7 my.

The Coniacian through Maastrichtian record of sequence boundaries and TRF cycles on Chart 4 are, because of these unresolved uncertainties in the calibration of European biostratigraphic zonations with the Gradstein et al. (1994, 1995) temporal scale, based on a North American record. North American sequences identified along the Gulf Coast in Texas and Arkansas are calibrated to the North American ammonite zones of Cobban et al. (1994). Sequences in the Coniacian through lowermost Maastrichtian interval are based on a tentative sequence-stratigraphic interpretation of outcrop sections described in published records from the Gulf Coast areas in Texas and Arkansas (Young, 1986; Kennedy and Cobban, 1993a, b, c; Cobban and Kennedy 1992a, b, 1993, 1994). Most Maastrichtian sequences (Ma2 to Ma5) are interpreted from outcrops in the area of the Maastrichtian stratotype.

Coniacian through lower Campanian sequences are identified in the Austin area of central Texas. Young (1986) describes three significant transgressions onto the San Marcos platform in central Texas e.g., near the Santonian/Campanian boundary,
upper Dessau Formation; middle Campanian, Pecan Gap Formation, and lower upper Maastrichtian, upper Corsicana Formation. These transgressions and two additional transgressions, one in the early Coniacian (onlap of Austin Chalk) and one in the upper Campanian (Bergstrom Formation) which yield no evidence of covering the San Marcos platform, are carried on the chart as Transgressive-Regressive Facies cycles.

The base of the Austin chalk onlaps Turonian strata, and the earliest Coniacian is not present in the Austin area. The transgressive base of the Austin Chalks is the base of the Atco Formation. The basal sequence boundary may actually be in the uppermost Turonian = Tu4. Other sequences in the Austin area are: base of the Vinson Formation = Co1, base Jonah Formation = Sa1, base Dessau Formation = Sa2, base upper Dessau Formation = Sa3, base Burditt Formation = Cam1 and base Sprinkle Formation = Cam2, Young (1986). The stratigraphic position of the sequences identified in the Austin area remains tentative because of uncertainties in the calibration of Young’s ammonite zonation with Cobban’s Western Interior ammonite zonation.

Lower Campanian to lowermost Maastrichtian sequences are based on well dated surfaces described in a series of papers on ammonite-bearing deposits in north eastern Texas and Arkansas by Kennedy and Cobban (1993a, b, c) and Cobban and Kennedy (1992a, b, 1993, 1994). The deposits described are obvious transgressive deposits associated with major flooding surfaces from which much of the ammonite record in the Gulf coast area is reported. The ammonite localities are: Ruxton Formation = Cam2 (Cobban and Kennedy, 1992), North Sulpher River = Cam3 (Cobban and Kennedy, 1992), Wolfe City Sand Formation = Cam4 (Cobban and Kennedy, 1993a), Pecan Gap Formation = Cam5 (Cobban and Kennedy, 1994), Annona Chalk Formation at Okay = Cam6 (Kennedy and Cobban 1993), Annona Chalk Formation at Yancy 1 = Cam7 and Yancy 2 = Cam8 (Kennedy and Cobban, 1993a), Saratoga Formation = Cam9 (Kennedy and Cobban, 1993) and Nacatomoch Formation = Ma1 (Cobban and Kennedy, 1995).

Maastrichtian sequences Ma 2 to Ma 5, based on outcrop data from the type area of the Maastrichtian Stage in The Netherlands and Belgium are calibrated to belemnite zones which are also poorly calibrated to the Gradstein et al. (1994) time scale.

**Lower Cretaceous Sequences**

Two papers concerning the lower Cretaceous were submitted for publication in this volume. To complement the documentation for the sequence stratigraphic record of the lower Cretaceous, Jacquin et al. (this volume) provide an overview of sequences and Transgressive-Regressive Facies cycles comprised in the transgressive phase of the Cretaceous Major Transgressive-Regressive cycle. Jacquin et al. (this volume) describe sequences in TRF cycles (numbered 11 to 15), from the northern North Sea to southern Italy. These TRF cycles represent sediment response to eustatic events caused by regional tectonic events superimposed on major intra-pllate reorganizations. Jacquin et al. (this volume) summarize the lower Cretaceous sequence record in the context of his Transgressive-Regressive Facies cycles. Hoedemaeker (this volume) describes sequences in the Berriasian-Barremian interval in southeastern Spain. Rufell and Wach identify several sequences in the Albo-Aptian of southern and eastern England.

**JURASSIC SEQUENCE CHRONOSTRATIGRAPHIC RECORD**

The Jurassic in Europe is characterized by two Major Transgressive-Regressive Cycles (Jacquin and de Graciansky, this volume). The transgressive portion of the first cycle (Ligurian Cycle) begins in the uppermost Norian Stage (upper Triassic) becomes regressive near the base of the middle Toarcian in the lower Jurassic and ends at the base of the upper Aalenian in the middle Jurassic. The transgressive portion of the second cycle (North Sea Cycle) begins in the middle Jurassic and becomes regressive near the top of the Kimmeridgian in the upper Jurassic and ends in the earliest Cretaceous (uppermost Berriasian). De Graciansky et al. a, b (this volume) and Jacquin et al. (this volume) describe Transgressive-Regressive Facies cycles (4 to 10) in the Jurassic. The number of TRF cycles and differences in the stratigraphic position of their bounding surfaces in tethyan and boreal basins reflect differences in sediment response to regional and more local tectonic activity. The record of individual sequences in the Ligurian MTR cycle (TRF cycles 4 to 6) is discussed in de Graciansky et al. (this volume) and for the North Sea MTR cycle (TRF cycles 7 to 10) boreal and tethyan basins is discussed in Jacquin et al. (this volume). Sequence stratigraphic interpretations for Jurassic Basins in Great Britain include Stephen and Davies (this volume) for the Moray Firth Basin, van Buchem and Knox (this volume) for the Cleveland Basin in Yorkshire and Hesselbo and Jenkyns (this volume) for the Wessex, Bristol Channel, Cleveland and Hebrides Basins. Gygi et al. (this volume) summarize sequence stratigraphic interpretations in the Oxfordian and lower Kimmeridgian of northern Switzerland. Sequence stratigraphy of rift related basins in tethyan settings are by Leinfelder and Wilson (this volume) for the Lusitanian Basin in Portugal and Dumont (this volume) in the western Alps in southeastern France.

**TRIASSIC SEQUENCE CHRONOSTRATIGRAPHIC RECORD**

Sequence stratigraphic interpretations of Triassic deposits in European basins include contributions from the Dolomites and Lombardy in the Southern Alps, the Western Southern Alps, Northern Calcareous Alps, Paris Basin, SE France, Germany and SW Barentz Sea. Sequence interpretations in the different basins are calibrated either to a boreal (coordinator Van Veen), or a tethyan ammonoid biozonation (coordinators Mietto and Manfrin). Gianolla and Jacquin (this volume) summarized the contributions and calibrated the various sequence interpretations to these boreal and tethyan ammonoid biochrozones. Jacquin and de Graciansky (this volume) identify a Major Transgressive-Regressive Cycle starting low in the Triassic and ending in the uppermost Norian (Eastern Tethys Cycle) where a second MTR cycle (Ligurian Cycle) begins that continues to the upper Aalenian in the middle Jurassic. Gianolla et al. (this volume) identify four TRF cycles (cycles 1 to 4) and 22 depositional sequences in Triassic basins from the Alps to the Barentz Sea. The lowermost TRF cycle (cycle 1) of Gianolla and Jacquin (this volume) may still be part of the Permian MTR cycle. Sequences carried on Chart 8 are discussed in Gianolla et al. (this volume) and Skjold et al. (this volume). Gianolla
and Jacquin (this volume) summarize and calibrate all papers submitted for the Triassic chapter of this volume.

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REFERENCES


MAJOR TRANSGRESSIVE/REGRESSIVE CYCLES:
THE STRATIGRAPHIC SIGNATURE OF EUROPEAN BASIN DEVELOPMENT

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ABSTRACT: Four Mesozoic major transgressive/regressive cycles have been recognized within Western European basins. They are named as: (1) Eastern Tethys Cycle, (2) Ligurian Cycle, (3) North Sea Cycle, and (4) North Atlantic Cycle, following the four main phases of rifting that affected the whole Western European Craton and its borders during Mesozoic times. Such cycles are bounded by major unconformities, whose names from oldest to youngest are: Hardegsen or Solling (Scythian), Early Cimmerian (Late Norian), Mid-Cimmerian (Aalenian), Late Cimmerian (Berriasian) and Laramide (Paleocene). Major transgressive/regressive cycles record outcrops the area of every individual basin, which suggests that local tectonic features were not the principal causes.

INTRODUCTION

The distribution of stratigraphic features, involving both carbonates and siliciclastics within sedimentary basins, primarily depends on changes in shelfal accommodation. Accommodation changes are caused by relative sea-level changes (the algebraic sum of subsidence or uplift and eustatic rise or fall; Vail et al., 1991). Shelfal accommodation changes can operate at any geological time scale and are a cyclic, aperiodic phenomenon that can be characterized by different stratigraphic signatures of various time duration. Five types of cycles longer than 100,000 years in duration have been observed in the stratigraphic record of European basins (Table 1). They are: (1) continental encroachment cycles, (2) major transgressive/regressive cycles, (3) transgressive/regressive facies cycles, (4) sequence cycles, and (5) parasequence cycles.

Continental encroachment (Pangean) cycles are described by Duval et al. (this volume) and should not be confused with our major transgressive/regressive facies cycles (Table 2). The objective of this paper is to document major transgressive/regressive cycles within Western European basins.

DEFINITION AND CAUSES

Major transgressive/regressive cycles are defined on the basis of long-term displacement of the shoreline, landward and seaward respectively. They are bounded by major subaerial unconformities which may extend over the entire basins. These major erosional unconformities, resulting from the major downward shift of the coastal onlap, are often associated with significant time gaps. Faulted, folded or uplifted strata are frequently linked with these major unconformities. The peak transgression is a major flooding event covering widespread areas. The gradual landward encroachment during the transgressive phase leads to the development of a condensed interval in the distal part of the basin by restriction of sediments. That interval identified on seismic lines as major basin-scale downlap surfaces can provide good source rocks such as the Toarcian Paper Shales in the Paris Basin.

It appears that these cycles have a time duration greater than 30 Ma (Table 1). The amplitude of shelfal accommodation change is in the order of 10^3 m. Geohistory analysis shows that these cycles are caused by 1st order relative sea-level changes mostly created by changes in the long term thermal subsidence (Fig. 1). Apatite fission track analysis also shows a good relationship with thermally controlled regional warping producing kilometer-scale uplift and erosional processes. The wavelength of subsidence versus uplift anomalies linked with this type of deformation could be in the order of N10^3 km. These cycles generally coincide with major steps in the evolution of individual basins, but they also record at distance the complex interaction of processes affecting other basins located far away, together with long-term eustatic changes.

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**TABLE 1.**—HIERARCHY AND PRINCIPAL CHARACTERISTICS OF STRATIGRAPHIC CYCLES: BOTH 3RD AND 4TH ORDER SEQUENCE CYCLES ARE CONTROLLED BY ACCOMMODATION-SPACE CHANGES. EFFECTS OF SUBSIDENCE INCREASE FROM FOURTH TO FIRST ORDER.

<table>
<thead>
<tr>
<th>Duration</th>
<th>Amplitude</th>
<th>Wavelength</th>
</tr>
</thead>
<tbody>
<tr>
<td>PANGEAN (1st-Order cycle)</td>
<td>&gt;250 MA</td>
<td>N 10^3 m</td>
</tr>
<tr>
<td>MAJOR T/R Cycles (1st-Order Sub-cycle)</td>
<td>&gt;30 MA</td>
<td>n 10^3 m</td>
</tr>
<tr>
<td>T/R FACIES Cycles (2nd-Order)</td>
<td>3 to 30 MA</td>
<td>n 10^3 m</td>
</tr>
<tr>
<td>SEQUENCE Cycles (3rd and 4th Order seq.)</td>
<td>0.5 to 3 MA</td>
<td>n 10^3 m</td>
</tr>
</tbody>
</table>

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**TABLE 2.**—COMPARISON BETWEEN PANGEAN AND MAJOR TRANSGRESSIVE/REGRESSIVE CYCLES CHARACTERISTICS.

**PANGEAN CYCLES (1st order cycle)**

Characteristics of boundaries:
- Maximum of progradation on continental margins
- Maximum of landward encroachment, with a downlap surface basinward
- Major source-rocks (Ordovician and Turonian)

Causes and effects:
- 1st order eustatic cycles created by changes in oceanic basin volume induced by, Break-up and subsequent gathering of the Proto-Pangea supercontinents

**MAJOR TRANSGRESSIVE/REGRESSIVE CYCLES (1st order subcycle)**

Characteristics of boundaries:
- Maximum regression: major erosional enhanced unconformity and correlative basin-scale onlap surface separating transgressive from regressive deposits.
- Peak transgression: major downlap surface in the distal parts of the basin with important starvation and major flooding and correlative strata updp separating transgressive from regressive deposits.

Causes and effects:
- 1st order relative sea-level changes mostly created by changes in the thermal long-term basin subsidence
- Regressive and transgressive phases are defined on the basis of long-term displacements seaward and landward respectively of the shoreline
Four major transgressive/regressive cycles have been recognized within Western European basins (Fig. 1, 2). They are named: (1) Eastern Tethys Cycle, (2) Ligurian Cycle, (3) North Sea Cycle and (4) North Atlantic Cycle, following the four main phases of rifting that affected the whole western European crust and its borders during Mesozoic times. Such cycles are bounded by major unconformities, whose names from youngest to oldest are: Hardegsen or Solling (Scythian), early Cimmerian (Aalenian), late Cimmerian (Ladinian) and Triassic long-term transgression can be dated as uppermost Scythian. In the Barents Sea, the Triassic pre-Spathian succession was deposited in continuity with the Late Permian layers, which records a phase of Late Permian extension (Skjold et al., this volume; Ziegler, 1988). The maximum of this long-term progradation is reached during uppermost Smithian times, where a minimum marine influence and strong terrestrial input have been identified (Hammerfest basin; Skjold et al., this volume). This progradation is followed by a major unconformity probably controlled by an Early Spathian compressional event (Van Veen et al., 1993).

An analogous trend can be documented from “Germanic” Triassic strata in Germany, Holland and France. In the most

**Fig. 1.**—Paris Basin geohistory analysis. Diagram illustrating major transgressive-regressive cycles and long-term evolution of the accommodation as reconstructed from the backstripping analysis method (Steckler et al., 1978) from the Chapton well in the central part of the Paris basin. Each point on the curve corresponds to a sequence boundary that has been identified by physical criteria and tied to the European sequence chart. Eustasy is taken as constant through time which implies the lower curve of the diagram represents the total accommodation (i.e. the tectonic subsidence + loading subsidence + eustasy). The middle curve is representative of the accommodation controlled by the external factor (i.e. tectonic subsidence + eustasy). The curve comprises four concave upwards segments, the four major transgressive/regressive cycles. Each of them show a long-term increase in accommodation during the transgressive phase and a long-term decrease during the regressive phase. The general slope changes clearly from one segment to the following one which shows that the strain development differed from one major transgressive/regressive cycle to the other. Major transgressive/regressive cycle boundaries are marked on the curve by uplift which induced hiatuses.

The lowermost unconformity’s definition is uncertain. It has been considered classically to coincide with the Permian/Triassic boundary (Haq et al., 1987, 1988; Duval et al., this volume). But available data from various European basins, including Boreal and Arctic domains, show that the onset of the Triassic long-term transgression can be dated as uppermost Scythian. In the Barents Sea, the Triassic pre-Spathian succession was deposited in continuity with the Late Permian layers, which records a phase of Late Permian extension (Skjold et al., this volume; Ziegler, 1988). The maximum of this long-term progradation is reached during uppermost Smithian times, where a minimum marine influence and strong terrestrial input have been identified (Hammerfest basin; Skjold et al., this volume). This progradation is followed by a major unconformity probably controlled by an Early Spathian compressional event (Van Veen et al., 1993).

An analogous trend can be documented from “Germanic” Triassic strata in Germany, Holland and France. In the most
subsiding parts of these basins, there are no significant breaks between the Permian red beds and “Buntsandstein” facies. The most important unconformity in the Swabische and Saxony Basins is within the upper part of the Buntsandstein. It is called “Hardeggen” (Trusheim, 1961) or “Solling unconformity” and occurs at the base of the transgressive Solling Folge (Röhl- 1992), which is dated as Spathian age. In the West Netherlands Basin, particularly on the Nederland Swell, the Harde- 1992). In the Paris Basin, the unconformity corre- sponds to the lower surface of the Conglomérat principal that rests unconformably over both Buntsandstein facies (Grès Vass- 1992). In Lombardy (Northern Italy), extensive carbonate platforms were established during uppermost Scythian times following a period of renewed tectonic activity with deep erosion close to the Smithian/Spathian boundary (Gaetani et al., this volume). We think that the globally uppermost lowest of Permian sea- level (Haq et al., 1987, 1988) could be extended into Scythian time. The ambiguity probably results from the existence of a higher-frequency (2nd-order) transgressive/regressive cycle within Scythian times, which is in continuity with the long- term Late Permian-Early Triassic regression. The huge amount of clastics (Buntsandstein) deposited within the grabens together with the erosion of the last post-Hercynian highs record this major regression.

Transgressive Phase

The transgressive phase begins everywhere at the Smithian/ Spathian transition. In the Barents Sea, extremely thick Spe- thian deposits accumulated, indicative of the amount of accom- modation space created. Source rocks also start to develop in this Arctic basin setting. During Late Spathian times, the Te- thys sea transgressed northward through the Polish-Dobroudja graben into northwest European basins (Ziegler, 1988). In the German Basin and its embayments towards Holland and the Paris Basin, the Tethys sea also transgressed during late Scyth- ian and Anisian times, forming widespread time-transgressive evaporites and then the Muschelkalk carbonates. A first link between Arctic and Tethyan faunas was reached along the so-called Cycloides bank, just above the fin- ing upward Untere Nodosus Shichten and below the coarsening upward Obere Nodosus Shichten (Aigner and Bachmann, 1992).

Regressive Phase

The regressive phase is dated as Late Triassic. It is charac- terized by:

1. A renewal of extensional tectonic activity in relationship with the southward propagation of the Tethyan rifts and the northward propagation of the Arctic and North Atlantic rift systems (Ziegler, 1988). This resulted in a complex network of horsts and grabens in which clastics and evaporites accumulated mainly during Carnian times.

2. An overall restriction of the depositional environments, with a strong influx of clastics in the Germanic, Boreal and Arctic domains and pervasive dolomitization on the Tethyan car- bonate platforms that produced the Norian Haupt Dolomit.

3. A generally poor Norian biostratigraphic resolution from the Arctic to the Tethyan domains, as a consequence of the restricted environmental settings.

LIGURIAN CYCLE: MAJOR TRANSGRESSIVE/REGRESSIVE CYCLE 2

Main Characteristics

The lower boundary is dated as Late Norian (211 Ma). The upper boundary is dated as Late Aalenian (177 Ma; Ludwigia murchisonae/Graphoceras concavum ammonite zonal bound- ary). The peak transgression is dated as around the Lower to Middle Toarcian boundary, at the top of the Harpoceras falciferum ammonite Subzone (187 Ma). The duration is 46 Ma.

Lower Unconformity

The lower unconformity is recognized as a major tectonically enhanced unconformity and is known as the early Cimmerian unconformity (Stille, 1924). It affected all European basins, with frequent erosional hiatuses, faulted and uplifted strata above and below. The unconformity separates the Norian re- gressive successions (such as the Haupt Dolomit or Dolomia Principale Formation in the Tethyan realm, the Hauptsandstein Formation of the German Triassic, the Hegre clastic Formation in the northern North Sea) from the Rhaetian transgressive de- posits that are the precursor of the Liassic transgression. The margins of the Paris and Lower Saxony Basins perfectly illustrate these conditions (Fig. 4). Unfortunately, the exact timing of the unconformity cannot be accurately documented. This is due to the poor biostratigraphic resolution of the Norian terrig- enous clastics and frequent pedogenetic dolomitization in Western Europe, such problems being enhanced by overall subaerial depositional conditions. However, in Lombardy (Southern Alps), the timing of the unconformity can be approached. It
FIG. 3.—First order peak transgression of the Eastern Tethys cycle. Correlations are well-log data from the upper Muschelkalk carbonate sequence M2 in Alsace (eastern Paris Basin; Goggin and Jacquin, this volume) and an outcrop section in southwestern German basin (Garnberg quarry; modified from Aigner and Bachmann, 1992). The subdivision of Ceratites zones from Urlichs and Mundios (1987) is also indicated. The third order maximum flooding surface is identified as occurring in the enodis Zone by correlation and is the major drowning event of the Muschelkalk carbonate platform.

separates the regressive uppermost Dolomie Principale from the transgressive Riva di Solto Shales (Gaetani et al., this volume) which yield Sevatian faunas (Gaetani; pers. commun., 1994)

Transgressive Phase

The transgressive phase is marked in the basinal areas of the North Sea by huge terrigenous influx dated as Rhaetian through Lower Sinemurian, which sustained balance with subsidence and long-term eustatic sea-level rise. This led to the development of widespread alluvial plain deposits of the Statfjord Formation and its time equivalents that blanketed the whole area between Norway, Greenland and Germany (Steel, 1993). This sand-prone fluvio-deltaic series formed the reservoir for major hydrocarbon accumulations in the northern North Sea and mid-Norway Basins (Ziegler, 1988).

In more southern areas, depositional environments are less uniform even though fluvial to shallow marine siliciclastic settings still predominate during Rhaetian times. Following the transgression, open marine pelecypods (Avicula = Pteria) appeared during Rhaetian deposition, rapidly followed by ammonites in the so-called Planorbis beds during lowermost Hettangian time (Mouterde et al., 1980). The area of shallow-water carbonate development extended northward, towards the London-Paris Basin, as the overall transgression progressed northward during Hettangian and Early Sinemurian times. From the Late Sinemurian (=Lotharingian) upwards, the long-term sea-level rise continued since: (a) progressive overstepping of all basin margins resulted in the widening of epicontinental seas, in which silts and shales are the dominant lithologies, (b) drowning of remnant highs, (c) development of kerogenous rocks during transgressive phases of higher order cycles (Fleet et al., 1987; Bessereau et al., 1994; Hanzo and Espitalié, 1994; Morton, 1993), (d) derestriction and flooding of the area between Greenland and Norway and (e) faunal exchanges between Boreal and Tethyan faunas across the Paris Basin (especially during the Tragophylloceras ibex ammonite Zone of the Lower Pliensbachian).

However, carbonate platforms that extended northward during the early phase of the Liassic transgression (i.e. Hettangian and Sinemurian), shifted southward during Pliensbachian and Toarcian times. Late Liassic platforms are only present in few places of the Northern margins along the Tethys, in Provence...
for example, but they are widely developed on its southern margins. This evolution suggests the long-term Liassic transgression and long-term climatic evolution had no strict causal link each other.

**Peak Transgression**

The Toarcian peak transgression is characterized by sediment starvation, together with the development of carbonaceous shales in offshore environments. It is the main kerogenous source rock of UK-Germany-Paris Basins (Schistes Carton in the Paris Basin, TOC up to 8% and IH up to 700–800; Hanzo and Espitalié, 1994; Paper shales in UK, Jenkyns, 1988; Wignall, 1991; Posidonien Schiffer in Germany). It is considered as a world-wide anoxic event (Jenkyns, 1980, 1985). Ferruginous oolites and/or manganiferous or phosphate crusts may develop on the raised edge of inherited tilted blocks (Jenkyns et al., 1991; Mettraux et al., 1989). The most condensed sections and the maximum extension of carbonaceous open marine shales are dated from the *Harpoceras falciferum* ammonite Subzone of the *Harpoceras serpentimum* Zone. In some areas, such as the Digne sub-Alpine basin, it may reach the *Hildoceras lusitanicum* “horizon” of the lower *Hildoceras bifrons* Zone (Graiciansky et al., 1993).

Within the Tethyan realm, there are no known aggradational shelfal environments that are time-equivalent to this starved episode. On the margins of North Sea basins, such as on the Horda Platform, Lower Toarcian shelfal environments are thin and do not extend far seaward. This suggests that the hinterland never supplied a significant amount of terrigenous material to the basins at that time and that carbonate platforms were not able to produce significant amounts of carbonate debris as well, a major departure from previous more aggradational, Hettangian and Late Sinemurian, tectonically controlled peak transgressions.

**Regressive Phase**

Following the Toarcian peak transgression the overall regressive phase dated as Late Toarcian to Late Aalenian is a common characteristic of all European basins. During early stages of this long-term regression, shaley sediments dominated over sands and carbonates in all latitudes, leading to the infilling of basin margins (Drake Formation in the North Sea; Steel, 1993; Marnes supérieures Formation and Marnes à Bifrons Formation in the Paris Basin; Mouterde et al., 1980; Alum shales Formation in the Yorkshire Basin; Powell, 1984; lower
inum Zone, Aalenian). Carbonate platforms on the eastern margin of the Paris Basin prograde seaward until the lowermost Bajocian (top LST Bj1, Hyperlioceras discites Zone).

Within the Digne or Serre Ponçon subbasins of the External French Alps (Graciansky et al., 1993) where the successions are very thick (up to 1500 m) and continuous and in more northern areas, the most seaward forestopping limestone and marlstone units are dated as uppermost Aalenian age (top Ludwiglia murchisonae ammonite Zone). The overlying calcareous units are stepping landward (Graciansky et al., 1993).

The effects on the long term accommodation of that unconformity can be documented from geohistory analysis as the result of a widespread uplift. This is probably in relation to the North Sea thermal event (Underhill and Partington, 1993, 1994), even though the uplift is documented hundreds of kilometers away from the North Sea (Fig. 2).

**Transgressive Phase**

The North Sea cycle transgressive phase is marked mainly by a Bajocian to Kimmeridgian episode of encroachment of marine deposits onto basin margins and structural highs. Within the North Sea area, siliciclastic shelfal environments progressively retreated landward on basins margins, leaving seaward shaley hemipelagic sedimentation (Heather Shales Formation). These are successively, from the top of the Brent delta system to the complete drowning at the peak transgression: the Tarbert Sandstone (Late Bajocian), the Krossfjord Sandstone (Bathonian), the Fensfjord Sandstone (Callovian) and the Sognefjord Sandstone (Late Bajocian).

Carbonate platforms dominate at that time in the Jura, Paris and London Basins, contrasting with the mainly siliciclastic facies of the underlying major transgressive/regressive cycle (Fig. 6). These carbonates, which were temperate deposits dominated by crinoidal limestones in Bajocian time, were progressively replaced by tropical deposits dominated by coral reefs from Bathonian through Kimmeridgian times (foramol facies replaced by chlorozoan facies association; Lee and Buller, 1972). The climatic evolution could have been concomitant with an overall sea-level rise. This had an important effect on the large-scale stratal patterns, due to the chlorozoan facies association capability to keep up and balance any increase in accommodation space (Fig. 7).

The increase of faunal exchanges between Arctic, Boreal and Tethyan biostratigraphical provinces and the increase of potential kerogenic source rocks in basin settings are the main consequences of this overall marine transgression. Callovian Tethyan ammonites may be found as far north as Greenland and Callovian Boreal ammonites may be found as south as Iberia (Marchand, 1984; Dardeau et al., 1994). Similar remarks can be done for the Kimmeridgian ammonites during the *Pictonia baylei* through *Aulacostephanus eudoxus* biozones (Hantzpergue, 1993).

**Peak Transgression**

The age of the peak transgression is still a matter of debate depending on the scale and the location in the basin or on the shelf. Evidence from the sea-level curve of Haq et al. (1987, 1988) suggests that the long-term sea-level rise had reached a maximum in the *Pectinatitides (Virgatospininctoides) wheatleyen- sis* zone (lower Volgian, or middle Tithonian sensu gallico or Upper Kimmeridgian, *sensu anglico*). Their key argument comes from the existence of organic-rich Kimmeridge clays (Draupne Shales in the North Sea), ranging from *Aulacostephanus eudoxus* (Late Kimmeridgian sensu gallico) through *Pectinatitides (Arkellites) hudlestoni* (Lower Volgian) ammonite Zones (Wignall, 1991a, 1991b; Herbin et al., 1994, 1995). When considering shelfal successions surrounding North Sea basins (Rawson and Riley, 1982; Doré et al., 1987; Gallois, 1988) such as the London-Paris Basin (Debrand-Passard et al., 1980), the Saxony Basin and also the Lusitanian basins in westernmost Europe (Leinfelder and Wilson, this volume), the maximum landward extent of open and deep marine lithologies (*Exogyra* and ammonide bearing marls) are dated generally from the uppermost part the *Aulacostephanus eudoxus* Zones (Hantzpergue, 1985, 1993).
This major flooding is followed on all shelves by a rapid regression dated as Portlandian (= Lower Volgian in the North Sea). This regression ended with complete exposure. Structural highs within North Sea basins were drowned by the same *Aulacostephanus eudoxus* flooding (Underhill and Partington, 1993b). In addition, this major flooding yields the most diversified and abundant faunal associations (Partington et al., 1993).

Even though the North Sea Kimmeridge Clay Formation comprises typical plankton-derived organic matter, as observed during every major transgression (Wignall, 1991a, b; Herbin et al. 1994, 1995), these carbonaceous shales interfinger during Volgian deposition with regressive sand-prone shelf-derived deposits (mass flows, turbidites, fault aprons). However, the Volgian organic shales have a different geochemical and mineralogical signature from true Kimmeridgian ones, showing the superimposed effect of terrigenous supply and incipient regression. Such observations document the partition of the Kimmeridge Clay Formation into transgressive “hot” shales and regressive “hot” shales (Clark et al., 1993).

The persistence of anoxia on the sea floor during the Volgian regression can be related to local submarine physiographic conditions, strongly enhanced by the Kimmeridgian extensional faults along the basin margins, that probably reduced oceanic circulation (Posamentier and James, 1993). In fact, the time distribution of the organic matter within the Kimmeridge Clay Formation is cyclic and corresponds to higher frequency (third-order) transgressive pulses. Examples are given in Yorkshire and in Boulonnais (northern France) by Herbin et al. (1994, 1995) and Proust et al. (1993).

**Regressive Phase**

The regressive phase is marked by rapid shoaling on European shelves during the Lower Volgian in the UK (Cope et al., 1980), the Paris Basin (Debrand-Passard, 1980), the Tethyan margins (Enay, 1984) and in Iberia (Aurell and Melenderz, 1993; Jimenez de Cisneros and Vera, 1993). The shelves show the widespread development of: (a) Portlandian facies, mainly low energy calcareous mudstones (formerly nanno-oozes; Busson et al., 1993), (b) Purbeckian facies, mainly lacustrine to evaporitic, confined lithologies, (c) Wealden facies, mainly continental terrigenous clastics, interfingering with the upper part of the Purbeckian (Allen and Wimbledon, 1991), (d) a hiatus with karstic development, which helps defining the unconformity bounding the major North Sea transgressive/regressive cycle. Within the basins, the regression is recorded by the stacking of thick masses of gravity flow, storm-induced deposits that were derived from the adjacent shelves. In the North Sea, gravitational deposits interfere with the Lower Volgian source rocks. They were derived from the erosion of Triassic to Jurassic sandstones along the adjacent emerged tilted blocks (Vollset and Doré, 1984; Dalland et al., 1988). Formations such as the Mun-
nin Sandstones in the North Viking Graben, the Šgiåth and the Piper Sandstones in the Witch Ground Graben (UK North Sea; O’Driscoll et al., 1990) and the Brae conglomeratic fans in the South Viking Graben (Chery, 1993) are examples. In basinal areas belonging to the Tethyan margin, the gravitational deposits form massive units interfingered with hemipelagic calcareous mudstones. These deposits are dated as Lower Volgian by calpionellids and ammonite fauna, as their North Sea equivalents (Tithonian facies, Dromart et al., 1993; Ogg et al., 1991). These gravitational deposits were derived from the adjacent carbonate platforms and may result from the erosion of previous Jurassic strata and other Late Jurassic beds (down to the Oxfordian). The onset of the gravitational sedimentation in the Boreal and Tethyan areas is precisely concomitant with the onset of the Portlandian facies on shelfal areas.

NORTH ATLANTIC CYCLE: MAJOR TRANSGRESSIVE/REGRESSIVE CYCLE 4

Main Characteristics

The lower boundary is dated as Late Berriasian (138 Ma). The upper boundary is dated as Paleocene (lowest Thanetian, 60 Ma). The peak transgression is dated as Lower Turonian; Watinoceras coloradoense (Europe) equivalent to Vas coceras birchbyi (USA) ammonite Zones and Withineilla archaeocretacea foraminiferal Zone (92.3 Ma). The duration is 78 Ma.

Lowermost Unconformity

The lowermost unconformity is referred to as the so-called late Cimmerian unconformity (Stille, 1924; Ziegler, 1978). Erosional truncation and subaerial exposures and karstification document this unconformity on all basin margins (Debrand- Passard, 1980; Aurell and Melendez, 1993; Detraz and Steinhauser, 1988; Deville, 1990; Stromenger et al., 1991; Allen and Wimeldon, 1991). These features are linked with a long-term relative sea-level fall (Haq et al., 1987, 1988) which is associated with global plate rearrangement and in particular, a renewal of the intraplate tectonic activity in Europe. The reorganization of the main depocentres between Jurassic and Cretaceous times records that change of the strain field. Within North Sea basins, this discontinuity is known as the Base Cretaceous Unconformity (BCU) where it may cut down as deep as Lower Jurassic strata (Johnson 1975). Nevertheless, a further, Late Barremian, unconformity, is often superimposed onto the BCU and intersects by places Jurassic strata. This is also known in Svalbard (Steel and Worsley, 1984).

In the London and Paris Basins, transgressive Valanganian Wealden facies overly unconformably Purbeckian and Portlandian strata and other Late Jurassic beds (down to the Oxfordian) at least on the structural highs and basin margins. These were truncated by the Late Cimmerian unconformity. The truncation is linked with a basin-scale subaerial exposure event, which lead to the development of a deep karst on the underlying Upper Jurassic limestones. In the Netherlands, gas-bearing sandstones of Valanginian age accumulated directly on a Zechstein or Triassic palaeorelief (Cottençon et al., 1975). In a similar manner the peri-Tethyan carbonate platforms such as the Jura, Moesia, Apulia, Yugoslav karst were submitted to pervasive karstification.

Dating such widespread unconformities is always questionable. In basinal areas where the stratigraphical successions are the most continuous and the age control can be good, the unconformity is not expressed as an obvious break. In southeastern France basins, it can be picked as the turning point between abundant platform derived gravity deposits and mainly shaley hemipelagic layers. From this point of view, the renewed abundance and diversity of dinoflagellate cyst species are indicative of the onset of the next long-term transgressive phase (Monteil, 1993). The same trend is recorded by the overall kaolinite/illite ratio decrease (Deconinck, 1993). Sedimentological, paleontological and mineralogical evidences indicate a Late Barrian age (Berriasella picteti ammonite Zone) for the unconformity. In the London/Paris Basin where the truncation is well expressed, a mid- to late Berriasian age can be assumed from paleontological evidence to the Purbeckian and Wealden equivalent below the unconformity (Fyfe et al. 1981; Strauss et al., 1993; Sigogneau-Russel and Esom, 1994). In the North Sea, the youngest strata recovered below the BCU are Late Volgian, Surites (Bojarkia) stenomphalus ammonite Zone equivalent from dinoflagellate cyst associations (Casey et al., 1993; Par tington et al., 1993). In other places, the gap is generally too long to allow more accurate dating.

Transgressive Phase

The transgressive phase of cycle 4 corresponds to the worldwide Early Cretaceous sea-level rise (Hancock and Kauffman, 1979) which culminated in early Turonian time and was the latest phase of the Pangean encroachment half-cycle (Duval et al., this volume; Haq et al., 1987, 1988; Jenkyns, 1985). However, the development of this long-term transgression was not continuous. Prior to late Barremian times, tectonically-controlled regressive pulses (Lower Valanganian and Lower Bar remian) interfered with clastics deposition in all basins and created widespread erosional unconformities.

For instance, in the Barents Sea (Svalbard), late Barremian sands with dinosaur foot prints rest unconformably over Kimmeridge clays (Steel and Worsley, 1984). In the northern North Sea, the Asgard Sandstone (Mid- to Late Barremian) overlies unconformably the Lower Cretaceous transgressive shales and the BCU (Skibeli et al., 1994, 1995). The Paris basin was also emergent during Barremian times. The maximum seaward propagation of Lower Cretaceous carbonate platforms on the northern Tethyan margin was reached by Late Barremian time (Jacquin et al., 1991).

The Apulian platform, isolated within the Tethyan ocean, also underwent a succession of subaerial exposures in the middle-late Barremian times (D’Argenio et al., 1987, 1991). The characteristics of the Cretaceous transgression appear from the Late Barremian times and onwards, with a tremendous increase in the diversity and abundance of planktonic foraminifers. This coincides in time with the onset of a green house period (Fischer, 1982), the globalization of oceanic anoxic events (Jenkyns, 1980; Robaszynski, 1989) and the widespread development of smectite-rich deposits (Thiry and Jacquin, 1993). The Cretaceous transgression also can be characterized by high-amplitude, high-frequency (2nd-order) flooding events, that progressively covered increasing areas of Europe, Africa and America and were probably controlled by eustasy, as suggested by their synchronicity worldwide.
In the North Sea, the Lower to «middle» Cretaceous transgression is indicated by a general, basin-scale, onlap of the inherited Rhyanzian and Late Jurassic structures. The change from the previous organic-rich Draupne Shale to the Lower Cretaceous transgressive deposits is reflected by the regional termination of kerogenous shales, due to the reoxygenation of sea waters (Ziegler, 1988). In the Tethyan basins, including the Central Atlantic Tethys, the change from the Volgan regression to the Lower Cretaceous transgression is marked by a progressive increase of organic matter content within the deep-sea sediments (Graciansky et al., 1984). This change was induced, by (1) oxygen depletion in the bottom sea waters in the context of sealed sub-basins and by (2) the development of an oxygen minimum layer at mid-depth following the relative sea-level rise (Graciansky et al., 1984; Arthur et al., 1987; Schlanger et al., 1987). This anoxia is also recorded in small embayments in the Tethys, such as the Vocontian trough in southeastern France, where TOC values may reach 3% and HI 300 (Breheret, 1994).

**Peak Transgression**

The Lower Turonian peak transgression is the time of maximum landward extent of open-marine facies towards the continental areas for all Mesozoic and Cenozoic times (Duval and Cramez, this volume). Turonian layers onlap onto older rocks, including Hercynian or Caledonian basements, around the Paris and London Basins and the North Sea (Juignet, 1980; Nederlandse A.M.B.V. and Rijks Geol. D., 1980). The peak transgression is known as a starred interval from Svalbard to Italy and to the Betics in Southern Spain (Graciansky et al., 1984) and even in the small Vocontian embayment of the Tethys in SE France (Crumière et al., 1990). It corresponds to a well-known oceanic anoxic event (Jenkyns, 1980) associated with relative starvation in the deep oceans (such as the Atlantic) and well preserved and abundant organic matter, when water depth and local physiographic conditions allowed. In the northern North Sea or in the Voring basin (Mid Norway), the peak transgression is a good regional through-going reflector with no special enrichment of organic matter, in probably related to the high sedimentation rates during the Upper Cretaceous deposition (Isaksen and Tonstad, 1989). In the chalk of the Southern North Sea, the peak transgression can be a good kerogenous interval (Plenus marls; Hart & Bigg, 1981). Thin organic–matter-rich layers have also been described within the white chalk of the onshore and offshore northern France and southern UK provided that water depth was sufficient (Graciansky et al., 1984). Where the carbonaceous layers are missing, the peak transgression is often recorded as a hardground incrusted with authigenic minerals.

**Regressive Phase**

The Late Cretaceous regressive phase is marked by a reorganization of the spreading system of the whole Atlantic and Arctic domains and coincides with the early Alpine orogenic cycle (Triumpy, 1960, 1980). Stratigraphic features, such as the level of erosion along major unconformities, the direction of progradation and/or retrogradation and the possibility to accommodate huge thicknesses of sediments, were controlled by these tectonic parameters.

Within North Sea basins, tremendous thicknesses of Late Cretaceous hemipelagic sediments accumulated (Shetland Group, Deegan and Scull, 1977), reaching up to several kilometers in the mid-Norway basins. Facies are mainly siliciclastic-dominated in mid-Norway (and more northward) and chalk-dominated in the southern North Sea (Isaksen and Tonstad, 1989). The pattern of the successions is very uniform, dominated by continuous fine-grained sediments, as a consequence of slow lithospheric cooling and linked long-term subsidence (McKenzie, 1978; Sclater and Christie, 1980). There are two major exceptions to this continuous sedimentation: a late Turonian phase of progradation of shelfal environments from the Norwegian continent and progressive infill of basins during the uppermost Cretaceous-lowermost Paleocene. The late Turonian regressive pulse is associated with a major, temporary, downward shift of coastal onlap. Intense tectonic activity, including block faulting and the local inversion of the previous structures, also is reported at that time. The uppermost Cretaceous infilling relates to the onset of rifiting in the North Atlantic region (Gage and Doré, 1986; Ziegler, 1988) and to the major Paleocene sea-level fall (Haq et al., 1988).

Within the London-Paris Basin and the southern North Sea, chalk was the dominant lithology. In the Alps, the consequences of the Late Cretaceous thrusting (Tricart, 1984) is the synorogenic production of huge masses of clastics deposited at the front of advancing thrust sheets in successive forelands. Such deformations started during Late Turonian times (Pyrenean phase, Olivet et al., 1984) and are also recorded as a major regressive pulses on surrounding peri-Tethyan shelves.

Within the Tethyan realm, the peri-Adriatic carbonate platforms were progressively dislocated into relatively independent blocks. Some of the blocks of the Apulian promontory (Monte Gargano area, D’Argenio and Mindszenty, 1991; Graziano, 1992) were emerged during the Late Turonian phase of compression. Most of them were subjected to karstification, bauxitization and/or erosion by the end of the Late Cretaceous deposition (Vecsei et al., this volume; Accordi and Carbone, 1988). This also happened in the Dalmate zone, the High Karst and pre–karstic units of the ex-Yugoslavia (Aubouin et al., 1970). In other units, the Late Cretaceous platform carbonates grade upwards into the hemipelagic Scaglia Formation (hemipelagic calcareous mudstones) and then to detrital flysch-type deposits. This occurred at varying ages depending on the location (Boselli et al., 1993), a consequence of the progressive break up of the Apulian platform recording the incipient Alpine orogeny.

The motion of Iberia relative to Europe during Late Cretaceous times induced the accumulation of synorogenic deep-water flyschs in the rapidly narrowing Pyrenean troughs (Debroas, 1990; Boillot et al., 1984). The eastern and northeastern part of the Iberian block itself was occupied by extensive carbonate platforms of Late Cretaceous age. They have been overlain by a regressive and/or emerged series dated as Late Santonian or Campanian through Paleocene age (Azema et al., 1974; Garcia-Hernandez et al., 1980; Gräfe and Weidman, 1993, this volume; Floquet, this volume). This evolution is again indicative of the end of the transgressive/regressive n°4.

**Uppermost Unconformity**

The North Atlantic cycle ended with an overall regression, which was probably induced by regional lithospheric defor-
CONCLUSION

Major transgressive/regressive cycle analysis has been applied to the Mesozoic succession of several basins from the Western European Craton. These cycles primarily record long-term change in shelfal accommodation, independently of the nature and thickness of the sediments of the depositional environments and of structural settings. It is not the nature and/or the expression of the cycle boundary which defines the cycle, but its large-scale facies and stratal stacking pattern. This approach differs from Embry’s (1993) transgressive/regressive cycle analysis for which a hierarchical system reflects the different nature of the boundary characteristics. Our approach compares well with the Sloss sequences of the cratonic North America Interior. (1963)

Four principal points characterize major transgressive/regressive cycles, as exemplified on the Paris Basin Mesozoic section (Fig. 8):

- Each starts with a major step within the overall Pangean continental encroachment (Duval et al., this volume) over the European Craton. The Eastern Tethyan Cycle (1) starts with the onset of the post-Hercynian transgression, following the upper Permian-lowermost Triassic phase of rifting. The Ligurian Cycle (2) starts with the widespread Rhaetian-Liassic transgression, following the Norian restriction (Fig. 2). The North Sea Cycle (3) begins with the Bajocian transgression, following an upper Toarcian thermal event (Fig. 7). The North Atlantic Cycle (4) was initiated with Lower Cretaceous transgression and continental encroachment and lead to major (Early Turonian) Pangean flooding (Duval et al., this volume). It follows the global sea-level lowstand around the Jurassic-Cretaceous boundary.

- Each ends with a major phase in disintegration of post-Hercynian platforms and craton. These phases coincide with the onset of the major rifting that affected the European Craton: Eastern Tethys, Western Tethys (i.e. Ligurian Tethys, Lemoine et al., 1986), North Sea (Ziegler, 1988), North Atlantic between Newfoundland-Gibraltar and Charlie-Gibbs fracture-zones including the Bay of Biscay (Olivet et al., 1984). Their characteristics, regarding the timing of the major boundaries and the direction of movement landward or sea-
ward of the shoreline, are not fully dependent of the succes-
sion of syn- rift and post-rift (or syn-drift) phases.
• Their corresponding accommodation curve, which can be drawn from the principles of backstripping analysis, shows a concave upward pattern. This is indicative of increasing ac-
commodation during transgressive phase and decreasing dur-
ing regressive phase. Such a pattern, together with its ampli-
tude and the wavelength (duration), documents and quantifies the process which is cyclic but aperiodic. Such characteristics match well with those of the McKenzie’s stretching model (1978), where syn-rift crustal extension and post-rift thermal cooling phase, amplified by the isostatic effect of sediments, give similar patterns.
• Their record concerns the whole European craton, which im-
plies that local tectonic subsidence is not the only leading factor which controls the long-term fluctuations of accom-
modation space. The global nature of major transgressive/ regressive cycle boundaries, as perceived by Stille (1924), is indicative of their eustatic origin. Observations and models converge to show the effects at distance of local tectonic events on both the intracratonic stress field and on the global sea-level variations (Cloetingh, 1988). Eustasy and (orogeny or) epeirogeny can no longer be viewed as mutually exclusive hypotheses (Gurnis, 1992) for explaining stratigraphy within intracratic and marginal sedimentary basins (Vail et al., 1991; Leinfelder and Seyfried, 1993). Thus the major trans-
gressive/regressive cycles, as well as the major European-scale unconformities, can be considered clearly as of tectono-
eustatic origin.

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TRANSGRESSIVE/REGRESSIVE (SECOND ORDER) FACIES CYCLES:
THE EFFECTS OF TECTONO-EUSTASY

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ABSTRACT: Transgressive/regressive facies cycle analysis combines the approaches of sequence stratigraphy at outcrop/core/well-log scales and seismic stratigraphy at seismic scales (large-scale stratal pattern and termination), to determine the facies stacking pattern and the partitioning of sediments following long-term changes in shelfal accommodation. Thus, it is an interdependent approach with the main purpose being to build a hierarchy of stratigraphic cycles.

The building blocks of transgressive/regressive (T/R) facies cycles are 3rd-order depositional sequences. Four types of 3rd-order depositional sequences may develop within a 2nd-order transgressive/regressive facies cycle: infilling and forestepping during the regressive phase and advancing and backstepping during the transgressive phase. These four types of sequences do not occur systematically together within a second-order cycle. Four end-members of T/R cycles can be defined depending on (1) the capability of sediment deposition to keep up with relative sea-level rises; (2) the rates at which accommodation space changes. The four end-members will include (1) T/R cycle with or without advancing sequences and (2) T/R cycles with or without forestepping sequences.

About 18 T/R cycles have been found within the Western European Mesozoic stratigraphic successions. At the craton scale, some of the characteristic surfaces and events are very synchronous. This synchrony suggests a tectono-eustatic control. Cycles which are not synchronous within a basin usually result from variations in local sea-floor subsidence/uplift. This can be seen particularly in the syn-rift and syn-compressional successions.

Both the type and occurrence of 3rd-order sequences (in respect to stratigraphy, depositional environments, reservoirs, source rocks and facies) depends of the type of 2nd-order cycle to which they belong. A full understanding of these characteristics observed in the data is essential to the analysis of the stratigraphic signature of a basin.

INTRODUCTION

The depositional (3rd-order) sequence is the basic stratigraphic unit of sequence stratigraphy (Mitchum et al., 1977, Vail et al., 1977, Vail 1987, Van Wagoner et al., 1987, 1990). Third-order depositional sequence are defined on the basis of stratal geometry and physical relationship, using objective stratal and facies criteria, whose scales are always compatible with those of outcrop and well log. According to sequence stratigraphic concepts, the distribution of these stratigraphic features primarily depends on shelfal accommodation changes. Therefore, to create new space for sediment that has filled the shelf or the platform to sea level, a relative sea-level rise is necessary. On the contrary, a relative sea-level fall is necessary to create a significant exposure surface. Unfortunately, the low resolution of seismic data generally does not allow the complete recognition of all the constituent elements of depositional sequences, except in areas with a high rate of sediment supply and tectonic subsidence. The seismic stratigraphic methodology (Vail et al., 1977) contributed to the solution of such problems by using the patterns of both the reflectors and their termination (i.e. onlap, toplap etc.). Thus, the resulting seismic stratigraphic framework depends more on the nature of the major bounding surfaces, than on the facies and stratal stacking patterns between these surfaces. The transgressive/regressive facies cycle (2nd-order) analysis combines both approaches, by first examining the seismic scale (large-scale stratal pattern and termination) and second by utilizing the outcrop/core and well-log scale to determine the facies stacking pattern and the partitioning of sediments following changes in shelfal accommodation. The transgressive/regressive facies analysis is an interdependent approach whose main purpose is to build the hierarchy of stratigraphic cycles (Fig. 1). This is important because each type of cycle (from the higher to the lower frequency) has a different predictive facies model, therefore understanding the hierarchy of the stratigraphic cycles allows proper application of the different predictive models.

The transgressive/regressive facies analysis is not independent of the sequence stratigraphic approach, until we integrate the depositional (3rd-order) sequences within the context of the longer term (2nd-order) accommodation cycles. In that context, the term transgressive defines stratal packages characterized by aggradationally-stacked or retrogradationally-stacked depositional sequences. The term regressive defines stratal packages characterized by progradationally-stacked depositional sequences. The definition of second-order cycles is independent of the nature, erosional or not, of their boundaries. Thus it differs from the definition of the 2nd-order super-sequences of the chart given by Haq et al. (1987, 1988), or the definition of second-order transgressive/regressive cycles of Embry (1993); both are based on a hierarchy of cycle boundaries, ignoring the long-term evolution of accommodation space. Our second-order cycles correspond best with the definition of post-rift megasequences (Steel, 1993), which consist of successively stacked basinward-stepping and then of landward-stepping progradational “tongues” but are genetically linked with a passive margin evolution.

Second-order transgressive/regressive cycles have a time duration ranging from 3 to 30 million years. Geohistory analysis shows that they are caused by 2nd-order relative sea-level changes, mostly created by variations in the long-term (1st-order) subsidence evolution (Fig. 2). Detailed backstripping analysis for the Jurassic interval in the North Sea, for example, shows a close relationship between the pattern of transgressive/regressive cycles and changes in accommodation (Fig. 3). It is clear that each regressive phase corresponds to a decrease in the rate of accommodation and each transgressive phase to an increasing rate of accommodation. Both the amplitude (n × 10^2 m) and the duration of these accommodation cycles (3 to
Infilling sequences

Infilling sequences develop in the early phase of an overall facies regression, either by aggradation or by progradation onto starved, generally shale-prone deposits of the previous 2nd-order peak transgression. The relatively small amount of progradation related to the infilling sequence is a consequence of the accommodation space still being created but with a decreasing rate over a long time. The 3rd-order infilling sequences may pinch-out by downlap towards the basin, just above the peak transgression surface.

Through time, 3rd-order infilling sequences extend further into the basin by overall progradation. Commonly, the thickness of shelf and basin sections are similar with neither a major erosional surface nor any particular enhancement of highstand versus lowstand systems tracts. Such geometries are not dependent on the nature of the sediment, carbonate or siliciclastics. Lowstand turbidites, slumped beds, megabreccias and other gravity deposits are seldom formed during that stage.

Forestepping Sequences

Forestepping sequences form large prograding intervals, merging landward into an erosional unconformity surface. Such a surface forms in response to major basinward shifts of the coastal onlap. Forestepping sequences typically occur during the later phase of overall facies regression. Landward, they pinch out by onlap, either below the storm wave-base or at the fair-weather wave-base (offlap break of previous sequences), depending on the hydrodynamic conditions and the depositional

30 my) fit with the intraplate stress model proposed by S. Cloetingh et al. (1986 a, b), in which the lithospheric stress field is correlatable with and may control the change of the sea-level.

ORGANIZATION OF TRANSGRESSIVE-REGRESSIVE FACIES CYCLES

Four types of 3rd-order depositional sequences may develop within a 2nd-order transgressive/regressive facies cycles (Jacquini et al., 1992). They are called (a) infilling and forestepping during the regressive phase and (b) aggrading and backstepping during the transgressive phase (Fig. 4). We insist on the point that the 3rd-order sequences are the building blocks of the 2nd-order facies cycles. So that each phase—infilling, forestepping, aggrading and backstepping—include several 3rd-order depositional sequences which characteristics is depending on their position within the 2nd-order facies cycle.

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profile. The unconformity surface that develops landward, corresponds to a 3rd-order sequence boundary, enhanced by low rates of tectonic subsidence, uplift of the basin margin and faulting (Jacquin et al., 1991). The maximum progradation (i.e., the maximum regression) is younger in age than such unconformities (Steel, 1993). For example, the maximum progradation within the Brent Group can be dated of the Upper Carixian (Amundsen and Burton cycles deposition). This uplift caused a major facies downward shift and induced the rapid progradation of the Brent delta (the lower part, Broom, Rannoch and Etive Formations within R6). The onset of the Bajocian transgression (T7) is observed in the lower part of the aggradational Ness Formation, when the accommodation becomes positive (change at 177.3, from long-term fall with negative values to long-term rise with positive values). Most of the aggradational alluvial plain deposits (Ness Formation) are deposited during a period of increasing accommodation (from 177.3 to 173.9 my). The Tarbert unconformity (onset of extensional tectonic in the northern North Sea) perturbs that trend. Unc.: unconformity; Mid Cim. Unc.: Mid Cimmerian unconformity.

The maximum regression occurs at the top lowstand of the last forest stepping sequence (Fig. 5). The last forest stepping sequence in the shallow-marine setting illustrates the lowest aggradational potential of the regressive phase. More precisely, this point is reached at the turning point when the ratio of progradation/aggradation for each 3rd-order depositional sequence evolves from increasing values (forest stepping phase) to decreasing values (Fig. 7, Jacquin and Schlager, 1993). In the deep-marine settings, this point generally coincides with the last sequence of a stack of 3rd-order sequences displaying particularly thick lowstand deposits (Graciansky et al., 1993, this volume).

**Aggradating Sequences**

Aggradating sequences develop in the early phase of the transgressive period (keep-up stage in carbonate environments). They form thick, widespread, low-energy, lagoonal deposits that build up during transgressive and highstand systems tracts. They may comprise evaporites (Carnian salts in the Paris basin, Goggin et al., this volume); mudstones (Late Barremian Urgonian, Arnaud and Arnaud-Vanneau, 1990, Jacquin et al., 1993); alluvial plain deposits, including peat and coal (Bajocian Ness Formation of the North Sea; Helland-Hansen et al., 1992); and stacked braided fluvialite systems (Late Triassic Lunde Formation, Steel, 1993). On seismic lines, they are recorded as widespread, generally low-amplitude reflectors (Fig. 7).

The most restricted sediments can be deposited at this stage, forming great thicknesses which indicates an increasing shelfal accommodation and not a maximum regression, as frequently misinterpreted from the presence of restricted facies only. It is during this stage that carbonate platforms are best developed, because of the increasing accommodation space in which carbonates growth can keep up. This relationship prevents lateral
**Thierry Jacquin and Pierre-Charles De Graciansky**

Thin but widespread highstand deposits

**AGGRADING SEQUENCE**

Thick shoaling upward highstand deposits

**BACKSTEPPING SEQUENCE**

Thin but widespread highstand deposits

**INFILLING SEQUENCE**

Moderately progradational lowstand deposits

**FORESTEPPING SEQUENCE**

Thick progradational lowstand deposits

**FIG. 4.**—The four types of depositional sequences that may be recorded within a second-order cycle. Each of these stages (Infilling and forestepping during the regressive phase, aggrading and backstepping during the transgressive phase) may show several 3rd-order depositional sequences. Note that tectonically enhanced unconformities (T) develop some distance below the level of maximum progradation (forestepping sequences). Backstepping sequences typically thin upward to a surface of drowning (D) where several 3rd-order depositional sequences may be merged by sediment starvation. LST: lowstand systems tract; TST: transgressive systems tract; HST: highstand systems tract.

**FIG. 5.**—Cross section of the Urgonian platform margin (Lower Barremian) in the Vercors area (SE France). Lower Barremian forestepping sequences (B0 to top lowstand of B3) consist of highly prograding bioclastic rimmed-shelf deposits. The northern part of the cross section shows the development of large prograding complexes (mainly lowstand, but also highstand). On the southern part (south of the Mont Aiguille fault zone), these forestepping sequences are represented by stack, bioclastic slope fan and gravity deposits with several internal erosional surfaces indicative of slope failure. The turning point between the overall progradation and aggradation at the top lowstand of B3 is indicated by a tongue of more open-marine facies interfingering landward within the bioclastic shelfal deposits. B0 to B5 refer to 3rd order depositional sequences dated as Barremian.
outgrowth and promotes upbuilding and consequently rimmed shelf profiles (Jacquin and Schlager, 1993).

Siliciclastic settings may also show rapid and widespread alluvial plain aggradation during this stage if the terrigenous sediment supply is high enough. Maximum regression, maximum progradation and maximum aggradation occur at different times and places. The maximum regression does not coincide always with the shallowest facies recorded. They all depend on the balance between sediment production and shelfal accommodation. High rates of sedimentation can fill all the space created and can maintain sub-aerial exposure conditions even during periods of long-term rise of relative sea-level, as in the aggradational interval of the 2nd-order transgressive phase.

**Backstepping Sequences**

Backstepping sequences are characteristic of the final stage of the transgressive phase. Here the sediments are not produced fast enough to fill in all the space being created, thus the depositional environments backstep. When approaching the 2nd-order peak transgression, the third-order lowstands deposits are generally not well developed. The most common are rather perched lowstand systems tracts or shelf margin wedges (Vail, 1987; Vail et al., 1991; Van Wagoner et al., 1990). The effects of downwarp shift at sequence boundaries may be subdued, limiting subaerial erosion and bypassing conditions.

The transgressive and highstand systems tracts on the outer parts of shelves and platforms are mainly composed of relatively thin, high-energy facies. Backstepping sandstones, lag deposits and ravinement surfaces are one of the common features of that interval with rapid relative sea-level rise and associated wave base erosion. In the Paris Basin, the Callovian bioclastic sandstones, preceeding the lowermost Oxfordian 2nd-order peak transgression (Fig. 8) were deposited during such a backstepping phase and are known through out Europe (Marchand, 1984). These deposits yield one of the major hydrocarbon reservoirs of the Paris basin (Villeperdue field, Duvall, 1992; Fitzgeral and Mousset, 1987).

Backstepping sequences thin upward to a drowning surface (Fig. 7, look at particularly the Kimmeridgian and Callovian backstepping sequences). This sequence is not necessarily synchronous over the entire platform depending on lateral changes of sediment supply. Above the drowning surface, a condensed interval commonly develops where several 3rd-order depositional sequences may be merged by sediment starvation (Fig. 9). The timing of the drowning surface is dependent on the balance between sediment production and shelfal accommodation. The drowning surface is not necessarily at peak transgression (Fig. 8). As a consequence, backstepping sequences are poorly resolved on seismic profiles due to their thinness. They generally coincide with a strong reflection amplitude generated at the interface between overlying marine condensed sections and underlying layers.

In areas where the platform underwent an earlier drowning, a subtidal shaley package of the aggradational lowstand systems tract at sequence boundaries may develop between the drowning unconformity and the peak transgression (Fig. 7, Kimmeridgian transgressive phase). Backstepping sequences thin basinalward because of the process of sediment starvation that occurs when depositional environments backstep. They commonly display thin basin members with high TOC values within the shaley highstand condensed sections.

The overall retrogradation of the depositional environments following an aggradational phase of growth often induces a change in depositional profiles. This is particularly noticeable in carbonate settings, due to the ability of carbonates to keep up with rapid rises of relative sea-level and to form rimmed...
Fig. 7.—Well log and seismic signatures of Jurassic transgressive/regressive cycles in the Paris Basin. B.: Berriasian; H.: Hauterivian; Barremian: A.: Aptian; Alb.: Albian; C.: Cenomanian.
shelves during the aggradational phase. Conversely, during the backstepping phase, the landward retreat of depositional environments may produce various morphologies: distally steepened ramps, isolated raised rims or empty lagoon (i.e., empty bucket geometry, Schlager, 1992; Jacquin and Schlager, 1993).

Peak Transgression

The peak transgression is placed at the maximum flooding surface of the backstepping sequence displaying the maximum shelfal accommodation potential. It can be traced from the deepest interval on the outershelf to the most aggradational interval on the platform. Second-order peak transgressions are among the best correlatable surfaces within marine series, where they are resolved as through-going high-amplitude reflectors (Fig. 7) sometimes associated with large-scale downlap termination. Good potential source rocks associated with sediment starvation may be found at that time (Loutit et al., 1988).

A Particular Case: The Healing Phase Wedge

During transgressive phases, the landward migration of the shoreline is able to remove sediments from the substrate, resulting in successive ravinement surfaces. This allows the deposition of backstepping washover sands landward, transgressive lag deposits seaward (Swift, 1968, 1975) and, in particular conditions, a sigmoidal-shaped sedimentary wedge further seaward that has been called the “healing phase deposit” (Posamentier and Allen, 1993 a,b). This process has been described basically at parasequences scale within 3rd-order transgressive systems tract.

A similar organization (Fig. 10) can be shown at the 2nd-order scale when strong differential subsidence occurs between shelf and basin settings (with increasing subsidence towards the basin). In proximal locations (Fig. 11A), extremely condensed, high-energy deposits (such as ferruginous oolites or phosphatic-glaucolithic sands) are amalgamated with numerous internal erosional surfaces (classical backstepping sequences). The only preserved sediments correspond to successive 3rd-order maximum flooding surfaces. In distal locations (Fig. 11B), these condensed deposits interfinger with terrigenous silty shales. These shales form wedge-shaped units onlapping onto the basin margins (the “healing phase” wedge). They may comprise organic-rich layers especially during time equivalents of transgressive lags of the backstepping sequences. Their thickness can reach those of the preceding shoreface deposits. Suspension is the main mechanism of transport, but turbiditic and/or storm-derived sandstones with good reservoirs characteristics can also be found. These healing phase wedges, that are time of the 2nd-order backstepping phase, including the peak transgression, compare well with the parasequence scale “healing phase unit” defined by Posamentier and Allen, 1993.

The tectonic setting, rather than the backstepping phase, is probably the main factor inducing the development of the 2nd-order “healing phase”. Sediments derived from ravinement processes during overall transgression accumulate where the accommodation potential is the highest. Therefore, the preservation of the 2nd-order healing phase is more likely to occur when the subsidence on the slope or on the basin margin is much more rapid than on the shelf itself.

Both the Late-Middle Jurassic Black Shales (Terres Noires) and the Mid-Cretaceous Blue Marls (Marnes bleues) in the Subalpine Basin (SE France) are good examples of healing phase wedges (Fig. 10). They both developed at the onset of the postrift thermal cooling subsidence, associated with Tethyan

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**Fig. 8.**—Schematic organization of Callovian backstepping sequences in the southeastern Paris Basin (adapted from Floquet et al., 1989 and Garcia, 1993). U. Cal: Upper Callovian; L. Oxf: Lower Oxfordian; M: marl; L. limestone; Ph: Pholadomya. Bt3 to Bt5 and Ca0 to Ca2 refer to 3rd-order depositional sequences dated as Bathonian and Callovian, respectively.
breakup (Terres Noires) and North Atlantic breakup (Marnes bleues).

EFFECTS OF SEDIMENT PRODUCTION AND ACCOMMODATION CHANGE

Four end-member types of transgressive/regressive facies cycles can be defined depending on: (1) the capability of sediment deposition to keep up with relative sea-level rises, thus controlling the development of aggrading sequences and (2) the rate at which accommodation space changes, controlling the development of forestepping sequences and associated tectonically enhanced sequence boundaries during the regressive phase (Jacquin et al., 1993). Various 2nd-order cycles give a different prediction on both the type and occurrence of 3rd-order sequences in respect to stratigraphy, depositional environments, source rocks and facies. These four types (Fig. 12) can be shortly described as: (1) with or without an aggrading phase depending on the sedimentary supply and (2) with or without a forestepping phase depending on the tectonic regime.

Transgressive/Regressive Cycles with Aggrading Sequences

Such transgressive/regressive cycles imply high rates of sediment supply in order to fill all the available space being created...
during the transgressive phase. They are characteristic of south Tethyan-type carbonate platforms (Jacquin and Vail, 1995). They may show tremendously thick aggradational buildups due to the high growth potential of the Tethyan carbonate factories, as long as they remain healthy (Schlager, 1992). Cretaceous Apulian platforms (D’Argenio et al., 1987, 1991) or Triassic buildups from the Dolomites of the Southern Alps (Gianolla et al., this volume; De Zanche et al., 1993) are examples. In siliciclastic settings, similar conditions are reached when uplift on basin margins and wet climatic conditions supply abundant sediment to the shelfal areas. The Triassic succession in the North Sea is a good example (Steel, 1993).

The accumulation potential of both terrestrial organic matter (leading to coal measures) and evaporites can also be high enough to form thick aggradational sequences during 2nd-order transgressions. This can be exemplified respectively by the Carboniferous paralic basins of western Europe (Courel, 1989) and the Keuper evaporitic succession of the Germanic Triassic in both Germany and Paris basins (Courel et al., 1994; Bourquin and Guillocheau, 1993; Courel et al., this volume; Goggin and Jacquin, this volume).

**Infilling-Forestepping-Aggradning-Backstepping.**

It is the most complete cycle with aggrading and backstepping sequences during the transgressive phase, infilling and forestepping sequences during the regressive phase. It has already been reviewed in the previous section of this paper, related to the organization of transgressive/regressive facies cycles.

**Infilling-Aggradning-Backstepping.**

Recognized as nearly symmetrical cycles, without forestepping sequences and any major unconformities, they form during periods of moderate to low tectonic activity, such as the post-rift period of passive margin’s evolution. The lack of forestepping sequence is the consequence of the overall relative sea-level rises throughout the whole second-order cycle without any long-term fall during the regressive phase. Decreasing and increasing rates of relative sea-level rise create respectively the regressive and transgressive phases.

The Comblanchien carbonate platform in the Bathonian-Callovian of the Paris Basin (Rat et al., 1986; Thierry, 1980) developed during one cycle of infilling-aggrading-backstepping-type (Fig. 8). This carbonate platform is made up of a relatively conformable succession of genetically linked depositional environments, with a stratal pattern indicative of infilling sequences (with moderate progradation) during the regressive phase and aggrading followed by backstepping sequences during the transgressive phase. During the regressive phase, facies of the infilling sequences evolve from offshore marls (Marnes à Acuminata, late Bajocian) to high-energy tidal oolitic grainstones (Oolite Blanche, late Bathonian). Although these high-energy grainstones extend a great distance basinward from previous shelf edges, they do not form forestepping sequences, because they don’t merge landward towards an unconformity. However, on structural highs in neighboring areas such as the Burgundy High and the Jura fold belt, mid-Bathonian oolitic forestepping sequences develop laterally into a major unconformity in response to a regional extensional tectonic episode. The vertical change from prodagadational infilling to aggradational is marked by the change from grainstones (Oolite Blanche) to mudstones with frequent subaerial exposure surfaces (Comblanchien). This is followed by the Callovian backstepping sequences (Fig. 8) which indicate a return to high-energy tidal environments (Javaux, 1992).
Transgressive-Regressive Cycle without Aggrading Sequences

The absence of aggrading sequences within a long-term transgressive episode can be related to several causes: (1) the rate of long-term rise of relative sea-level exceeds the rate of the sediment production so that depositional environments move landward (or backstep) to a shallower higher position on the shelf; (2) the particularly rapid rise of the relative sea-level, induced either by tectonism or by eustasy, may inhibit the sediment supply. These cycles are characteristic of North Tethyan carbonate platforms such as the Lower Jurassic units of the German and London-Paris Basin where the production potential is much lower than in the Southern Tethyan counterparts; and (3) when adverse conditions, such as temperature lowering or the arrival of turbidite waters or clastic terrigenous sediments, affect the ability of carbonate platforms to produce and export sediments (Jacquin and Vail, 1995).

Second-order cycles, without aggrading sequences, are the most common in siliciclastic settings. This is due to the fact that siliciclastic depositional environments lack in-situ sediment factories (as reeval carbonate platform rims) and are more likely to move landward or seaward in response to fluctuations of relative sea-level, instead of stacking upwards.

Infilling-Backstepping.—

This transgressive/regressive cycle is the most symmetrical of the four end-members. The seaward-stepping units (infilling sequences) comprise prograding sequences characterized by progradation/aggradation ratios of less than 10, which means a moderate amount of progradation with respect to aggradation. The successive marine sandstone “tongues” derived from the nearshore prograding wedges pinch out seaward into marine shales. The landward stepping units (backstepping sequences) can be very thin, as a result of repetitive ravinement processes. Backstepping sequences are frequently condensed in the basin, as a consequence of sediment starvation. The late Pliensbachian (Lower Liassic) succession of northwestern European basins (cycle 5) is an example: Mid Norway basins (Tilje cycle, Dalland et al., 1988), Northern North Sea Basins (Cook Cycle, Doré et al. 1985; Steel, 1993), Yorkshire basin (Staithes Sandstones cycle, Ivimey-Cook et al., 1983; Powell, 1984), Paris Basin (Calcaires à Gryphées and Grès médio liasiques cycles, Graciansky et al., this volume), Eastern Aquitaine (Barre à Pectens cycle, Rey and Cubaynes, 1991), Holland (Aalburg cycles, Nederlandsee, A.M., 1980).

Infilling-FORESTEEPING-Backstepping.—

The infilling-forestepping-backstepping-type of 2nd-order cycle is the most asymmetrical of the four end members. Shoreface sediments from forestepping sequences may extend tens of kilometers out into the basin. They merge towards major tectonically enhanced (3rd-order) sequence boundaries landward. Submarine erosional surfaces also develop at that stage on top of the intrabasinal swells. Backstepping sequences are thin and cover extremely wide surfaces similarly to the previously described type of sequences. The high degree of asymmetry of this type of 2nd-order cycle with the enhancement of 3rd-order relative sea-level falls and associated progradational deposits indicate a strong tectonic control on the long-term evolution of accommodation. Such characteristics are found during synrift phases of basin evolution where the pattern of tectonic subsidence is decreasing nearly at an exponential rate (Figs. 2, 3). The most typical cycle of that sort in northwestern Europe is the N°4 (Lower Pliensbachian): Northern North Sea basins (Amundsen cycle, Steel, 1993), Yorkshire basin (Red Scar Cycle, Ivinney-Cook and Donovan, 1983; Powell, 1984), Paris and Aquitaine basins (Calcaire à Davoei cycle, Guillocheau, 1991; Graciansky et al., this volume), Subalpine basin in southeastern France (Carixian cycle, Graciansky et al., 1993).

Conclusions

The different types of 2nd-order transgressive/regressive cycles are defined on the basis of stratigraphic and physical relationships, using objective stratal and facies criteria, independently of the frequency of occurrence, the nature of cycle boundaries and depositional processes. Such characteristics are only dependent on the rate at which long-term accommodation is created. The physical characteristics of the transgressive/regressive cycles, such as the duration and amplitude of accommodation, the wavelength of the linked deformation and the lateral extent of the amount of seaward and landward stepping, are the main features that determine the stratigraphic signature of individual basin or subbasins. In consequence, these features can be quantified using conventional backstripping analysis and measuring physical parameters such as the aggradation/progradation ratio.

About 18 transgressive/regressive cycles have been found within the Western European stratigraphic successions (Gianolla et al., Graciansky et al. and Jacquin et al., this volume). They are relatively synchronous at the craton scale. This synchronicity suggests a tectono-eustatic control. Cycles that are not synchronous, even within a single basin, result from variations in local sea-floor subsidence/uplift. This can be particularly seen in the synrift and syn-compressional successions.

Long-term accommodation changes affect the pattern and facies of third-order depositional sequences, that are the building blocks of the transgressive/regressive facies cycles. Four types of third-order depositional sequences, infilling, forestepping, aggrading, backstepping, can be documented as a direct consequence of these long-term accommodation changes. These four types of depositional sequences have been documented in both siliciclastic and carbonate settings and in a broad range of structural settings. A full understanding of these four types of sequences is essential to avoid the classical confusion between second-order T/R cycles and third-order depositional sequences whose limits and system tracts could have been influenced by local conditions. Most of the criticism of seismic stratigraphy and sequence stratigraphy rightly focuses on the inadequacy of seismic lines to resolve stratigraphic features and on the frequent lack of objective unconformity-bounding depositional sequences. The understanding of the hierarchy of second- and third-order stratigraphic cycles solves that problem. It shows that seismic-scale unconformities mainly develop at a second-order scale and should not be confused with sequence boundaries. Similarly (for example) seismic scale “lowstand deposits” generally coincides with a set of third-order forestepping sequences and should not be interpreted as a single lowstand. This has important consequences for exploration, as forestepping sequences and associated unconformities may yield major reser-
voirs zones. Moreover a good understanding of that hierarchy of stratigraphic cycles allows the interpreter to better integrate the available tools (seismic lines, well data, cores . . .) with the stratigraphic framework.

The four types of 3rd-order sequences do not occur systematically together. Their presence or their absence has also a predictive potential. It reflects the stratigraphic signature of the basin history, including long-term changes of the sediment supply, as expressed by the presence or absence of aggrading and/or backstepping sequences and long-term changes of the tectonic subsidence, as expressed by the degree of asymmetry of transgressive/regressive cycles.

This approach should be a very powerful tool, especially for frontiers areas, where well data is sparse. Understanding the nature and the distribution of third-order depositional sequences within their second-order framework from mature areas is critical for prediction in frontier areas of potential distribution of forestepping sequences with major lowstand reservoirs and of starved backstepping sequences with major source rocks.

REFERENCES


STRATIGRAPHIC CYCLES AND MAJOR MARINE SOURCE ROCKS

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ABSTRACT: Four types of stratigraphic cycles with a time duration greater than 10,000 years are recognized in the sedimentary record. In order of decreasing time duration and scale they are: (i) continental encroachment cycles, (ii) regression-transgression cycles, (iii) sequence cycles and (iv) parasequence cycles.

The recognition and understanding of the architecture of the continental encroachment cycles, subcycles and/or the regression-transgression cycles and the location of the major downlap surfaces are important steps in the study of petroleum systems. They allow explorationists to locate the most likely marine source rocks. On seismic data, continental encroachment cycle and subcycle interpretations are used, particularly in the proximal part of sedimentary basins, where the encroachment is relatively easy to recognize. As an alternative, in the intermediate parts of the basins, where the offlap-breaks are usually identifiable, regression-transgression cycle interpretations can also be used to locate potential marine source rocks. In this paper, applications of the continental encroachment cycle and subcycle concept in locating potential marine source rocks using seismic data are presented, together with comments on the stratigraphic distribution of major potential marine source rocks.

INTRODUCTION

In the field, sedimentary cycles have been recognized for centuries (e.g., Steno, 1669, de Mailliet, 1748, Lavoisier, 1789, Lyell 1830, Suess, 1888, etc.), and it has been suggested that sea-level rise and fall was the main cause of the cyclicity in sedimentary rocks. In this century, geologists (e.g., Lemoine, 1911; Graubau, 1936; Burollet, 1956, Sloss, 1962, etc.) have recognized the component effect of tectonics and sea-level variations on sedimentary cycles. The advent of the plate tectonic paradigm provided the foundation for the concepts of eustasy and relative sea-level changes which are the basis of seismic stratigraphy, which was introduced in 1977.

In summary, the eustatic curve (Haq et al., 1987) depicts the global mean sea-level variations* during the Mesozoic and Cenozoic and is composed of various eustatic curves with cycles of different periods or time duration. According to the duration of each eustatic cycle component, cycles of four orders can be recognized:

a. 1st Order, with a duration greater than 50 my,
b. 2nd Order, with a duration of between 3 and 50 my,
c. 3rd Order, with a duration of between 0.5 and 3 my,
d. 4th Order, with a duration of between 0.01 and 0.5 my.

As each component cycle of the eustatic curve induces a stratigraphic cycle, some geologists tend to rank the stratigraphic cycles in terms of eustatic time orders. We avoid such a ranking for two main reasons. Firstly, a stratigraphic cycle (i.e., the rocks) does not record the corresponding geological time of an eustatic cycle due to erosion and non-deposition (Ager, 1984). Secondly, different types of stratigraphic cycles (e.g., continental encroachment subcycle and regression-transgression) can be associated with eustatic cycles of the same order (2nd order); also, the same type of stratigraphic cycle (e.g., sequence cycles) can be induced by eustatic cycles of different orders (3rd or 4th order).

TYPES AND HIERARCHY OF THE STRATIGRAPHIC CYCLES

In order of decreasing time duration and scale, four major stratigraphic cycles are recognized in sedimentary records (field, seismic, electrical log and core data):

1. Continental encroachment cycles (Fig. 1) are defined on the basis of features showing onlapping against cratons. They are produced by eustatic cycles with a time duration greater than 50 my, (i.e. 1st order eustatic cycles), induced by continental breakup and subsequent aggregation of the supercontinents. They are bounded by major tectonically enhanced unconformities and are composed of a backstepping transgressive phase overlain by a foresteppeing regressive phase. A major downlap surface separates the transgressive and regressive phases. These phases are reflected by smooth long term shoreline displacements.

The continental encroachment cycles can be subdivided into different encroachment subcycles (Fig. 2) using significant downward shifts of encroachment associated with 2nd-order eustatic cycles. These subcycles are bounded by major erosional unconformities in the proximal parts of basins, resulting from the downward shift of onlap, and are bounded by correlative paraconformities in the distal parts of basins.

2. Regression-transgression cycles (Fig. 1) are defined on the basis of long-term displacements of the shoreline and are associated with 2nd order eustatic cycles. These eustatic cycles are interpreted to be the result of changes in the rate of regional tectonic subsidence and/or changes in the rate of

*Mean sea level avoids reference to local variations induced by gravity anomalies
sea-floor spreading (Vail, et al., 1984). These regression/transgression cycles are bounded by significant downlap surfaces in distal and intermediate parts of basins, where hiatuses due to non-deposition are common. In proximal parts of a basin, where hiatuses are insignificant, these cycles are bounded by up-dip correlative surfaces. Regressive and transgressive facies are separated either by unconformities, in proximal parts of basins, or, in distal parts, by first flooding surfaces on top of lowstand deposits.

Due to the large range of 2nd-order eustatic cycles (3 to 50 my), regression-transgression stratigraphic cycles can be subdivided, using the same criteria into subcycles depending on the resolution of offlap breaks on the data available.

3. Sequences cycles (Fig. 1) are defined on the basis of shelfal accommodation changes and are bounded by unconformities induced by relative sea-level falls associated with 3rd-order eustatic cycles (i.e., those with a time duration of between 0.5 to 3 my), which are assumed to be caused by glacial events (Vail et al., 1977). However, in certain areas, sequence cycles are associated with eustatic cycles of 4th-order.

In terms of the presence of various systems tracts, the sequence cycles can be complete or incomplete. They are complete (Fig. 1) in areas with high or normal sedimentation rate (i.e., where all the available shelfal space is filled). In this case, we can recognize from bottom to top the following depositional systems tracts:
STRATIGRAPHIC CYCLES AND MAJOR MARINE SOURCE ROCKS

The sequence cycles are incomplete (Fig. 1) in areas with low sedimentation rates, where only part of the shelfal accommodation is filled. Several types of incomplete sequence cycles can be recognized in the field or on the seismic lines, such as: flooding and forestepping, bypass-forestepping, forced regression, channel-fill and overbanks, backstepping and forestepping, etc.

4. Parasequence cycles (Fig. 1) are intervals bounded by flooding surfaces or their correlative conformities developed in association with 4th-order eustatic cycles (0.01–0.5 my) or higher.

These four types of stratigraphic cycles can be recognized in the field. However, on conventional seismic data, only the continental encroachment and the regression/transgression cycles are usually interpretable. The sequence cycles can only be identified in certain basins with high rates of sediment accumulation, such as the Gulf of Mexico or the Mahakam delta.

There are two major continental encroachment cycles in the Phanerozoic (Fig. 3) which are associated with 1st-order eustatic cycles created by the changes in ocean basin volume induced by the breakup and subsequent gathering of the proto-Pangea and the Pangea supercontinents.

a. lowstand systems tract, including basin floor fans, slope fans, and prograding wedge.
b. transgressive systems tract, and
c. highstand systems tract.

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a. lowstand systems tract, including basin floor fans, slope fans, and prograding wedge.
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Fig. 4.—Major organic-rich intervals in the Phanerozoic are associated with downlap surfaces of continental encroachment subcycles. Their correlation with the percentage of the world’s original petroleum reserves generated by source rocks of stratigraphic intervals proposed by Ulmishek and Klemme (1990), is quite good despite hydrocarbon loss during the aggregation of the supercontinent Pangea. The highest percentage area is associated with the major downlap surface of the continental encroachment cycles (Ordovician-Silurian and middle-upper Cretaceous).

The older cycle (Fischer, 1984) started in uppermost Proterozoic time and extended to the end of Permian time. The Proterozoic was a time of slow encroachment with regression, whereas Cambrian time was a period of extensive encroachment with transgression. A eustatic high was reached during Ordovician-Silurian time and from Silurian to Permian time, there was a gradual restriction of the marine domain.

The younger cycle (Fischer, 1984) started in the Triassic and extends to the Present time. The Triassic Period was a time of slow encroachment of sediments onto the craton, while the Jurassic and the Early Cretaceous Periods were times of extensive encroachment. Early Turonian time is believed to have been the time of the maximum eustatic high, while Late Cretaceous and Cenozoic time was characterized by a gradual restriction of sediments to the continental margins and basinal areas.

The maximum marine transgression within the continental encroachment cycles occurred at the Ordovician-Silurian boundary in the older cycle and near the late Cretaceous time in the younger cycle. Each continental encroachment cycle (Fig. 3) shows a smooth long-term landward displacement of the shoreline followed by a seaward displacement. Sedimentary packages deposited during the seaward displacement of the shoreline have mainly forestepping progradational geometries and comprise what is termed the regressive phase. The transgressive phase thickens landward to a maximum and then pinches out against the craton. The regressive phase reaches maximum thickness seaward and becomes condensed in the more distal parts of the basin. The surface between the transgressive and regressive phases is marked by a major downlap surface which represents the eustatic high and a period of starved sedimentation. These geological conditions combined with coeval upwelling currents and anoxic environments are favorable for the development and preservation of major source rocks. By recognizing the major downlap surface of continental encroachments cycles we can locate the most likely marine source rocks (Fig. 4).

The seismic line from southern offshore Angola (Fig. 5) illustrates the post-Pangea continental cycle. Several sedimentary basins can be recognized (rift type, cratonic and divergent margin). Despite strong salt tectonic deformation which creates large rafts in the transgressive phase and huge depocenters in the regressive phase, the downlap surface between these two phases still can be easily identified on the eastern part of the line. Also, the contrast between the sub-parallel geometry of the transgressive phase versus the progradational geometry of the regressive phase helps us to locate the major downlap surface and the associated marine source beds.

The geochemical logs (TOC%* and Tmax**) of wells A and B (Fig. 6) clearly illustrate the presence of a major organic

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**TOC:** Total Organic Carbon

**Tmax:** temperature of the maximum of hydrocarbon generation from pyrolysis
interval between the transgressive and regressive phases. As indicated by Tmax values (less than 430°), these potential source rocks are immature in the area of the section. However, in the northern part of offshore Angola the sediments associated with this major downlap surface have been buried under the Congo deltaic complex and the associated organic matter has reached maturation. These mature marine source rocks have been responsible for the generation of hydrocarbons.

Another seismic example of the post-Pangea continental encroachment cycle is shown in Figure 7. It comes from the eastern offshore Venezuela, where the stacking of a Triassic rift-type basin, a Lower-Cretaceous cratonic basin and Mesozo-Cenozoic divergent margin is recognized above a Paleozoic or Precambrian substratum (Bally, 1980). The major downlap surface (in white on the seismic line) separating the subparallel geometry of the transgressive phase from the progradational geometry of the regressive phase cannot be missed if the progradations of the Upper Cretaceous-Lower Paleogene strata are carefully interpreted. The transgressive phase is indicated by the arrow pointing to the left and the regressive phase by the arrow pointing to the right, where significant Pliocene growth faults are recognized. The vergence of the arrows does not only show the direction of the smooth long-term displacement of the shoreline, but also the direction of thickening of the respective phases, particularly that of the seaward thickening of the regressive phase. Since the maximum of thickening is located near the right lower corner of the line, landward the combination of backstepping and pinchout geometries reduces the thickness of the transgressive phase against the craton. The major organic-rich level is associated with the early Late Cretaceous downlap surface between the transgressive and regressive phases.

CONTINENTAL ENCRoACHMENT SUBCYCLE AND SECONDARY MARINE SOURCE ROCKS

During a continental encroachment cycle, onlap against the cratons does not always show continuous landward and upward movement (positive aggradation). Occasionally, the onlapping shows major shifts seaward and downward (negative aggradation). During these large downward shifts of onlapping, the coastal plain and the upper slope are exposed, due to major eustatic sea-level falls, and a pronounced erosional unconformity is formed. Like the definition of a continental encroachment cycle, a continental encroachment subcycle is defined as being between two consecutive downward shifts of onlap (i.e., two significant erosional unconformities within a continental encroachment cycle, (Fig. 8). Each subcycle is developed in association with a 2nd-order eustatic cycle (3–50 my).

Inside each encroachment subcycle, the most significant downlap surface (Fig. 8) separates a retrogradational subphase from an overlying progradational subphase. The basinward ex-
CONTINENTAL ENCROACHMENT CYCLE: MAJOR MARINE ORGANIC INTERVALS

<table>
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OFFSHORE ANGOLA

TOC % T max
420˚ 440˚

Pliocene
Miocene
Oligocene
Eocene
Middle Cretaceous
Aptian salt
Pre-salt
Volcanics

FIG. 6.—The major organic interval in the post-Pangea continental encroachment cycle in the southern Angola offshore is located between the transgressive and regressive phases of the cycle and is of middle Cretaceous age. The TOC reaches 10% in well A and 7% in well B. However, as indicated by the Tmax log, this organic matter is immature.

CONTINENTAL ENCROACHMENT CYCLE: MAJOR MARINE ORGANIC INTERVALS

West

Projected well
(long distance correlation)

East

Offshore Venezuela

Forestepping regressive phase

Major downlap surface

Backstepping transgressive phase

Fig. 7.—The continental encroachment cycle interpretation of a seismic line from the eastern offshore of Venezuela indicates the probability of potential source rocks associated with the major downlap surface of the Meso-Cenozoic continental encroachment cycle.
CONTINENTAL ENCROACHMENT CYCLE AND SUBCYCLE

Break-up and gathering of supercontinents

Changes in rate of tectonic subsidence

Supercontinent

Major downward shifts in continental encroachment

Transgressive phase

Regressive phase

first order

break-up and gathering of supercontinents

second order

3-50 My

Transgressive phase

Regressive phase

transgressive subphase

regressive subphase

transgressive subphase

regressive subphase

regressive subphase

transgressive subphase

transgressive subphase

regressive subphase

regressive subphase

regressive subphase

transgressive subphase

Regenerative phase

major downlap surface

Transgressive phase

FIG. 8.—Continental encroachment cycles can be divided in subcycles using the major erosional unconformities created by significant downward shifts of onlapping. Within each encroachment subcycle a retrogradational transgressive subphase and a progradational regressive subphase can normally be recognized. The downlap surface between these subphases represents the most likely location of organic-rich marine sediments.

West

Continental encroachment subcycles interpretation

East

North Sea

Major organic-rich intervals: Middle Miocene, Lower Turonian, Kimmeridgian

Fig. 9.—Three major levels of potential source rocks are identified in the North Sea in association with the downlap surfaces of the continental encroachment subcycles: the middle-Miocene, the lower Turonian and Kimmeridgian. The Kimmeridgian marine shales, deposited during the base Jurassic-middle Berriasian encroachment subcycle, are by far the richest in organic matter.
Major organic-rich intervals: Late Devonian and Early Silurian

FIG. 10.—The continental encroachment subcycle interpretation of the seismic lines of the southern Algeria suggests two major levels of potential source rocks for the Protopangea continental encroachment subcycle. They are associated with the downlap surface of the base Silurian-base middle Devonian and base middle Devonian-base middle Carboniferous encroachment subcycles.

tent of this downlap surface varies from one subcycle to another and depends on the construction of each subcycle (Fig. 8). The major downlap surface of the continental encroachment cycle coincides with the downlap surface of only one of the subcycles. Downlap surfaces of the rest of subcycles are therefore considered to be secondary downlap surfaces (Fig. 8).

In petroleum basins around the world, seismic interpretations using stratigraphic cycles and geochemical studies show a correlation between the marine source rocks and the downlap surfaces of the continental encroachment cycles and subcycles.

During the Phanerozoic, two major downlap surfaces at Cambro-Ordovician and early Late Cretaceous times, respectively of the Proto-Pangea and Pangea continental encroachment cycles, are associated with major organic-rich marine source rocks. Secondary downlap surfaces within the encroachment subcycles are associated with the rest of the important marine source rocks. The world’s percentage distribution of original petroleum reserves according to their source rock stratigraphic interval (Ulmichek, 1991) shows a good correlation with the downlap surfaces of the encroachment subcycles.

As an example, a seismic interpretation of the Meso-Cenozoic continental encroachment subcycles from the North Sea is shown in Figure 9. Four encroachment subcycles are recognized and are characterized by downward shifts of the onlapping. They are bounded by the base Triassic, base Jurassic, mid Berriasian and mid Oligocene unconformities. The downlap surface within the first subcycle (base Triassic-base Jurassic) is not discernible, whereas the others, although slightly masked by the isostatic uplift, are easily recognized. Their ages are respectively Kimmeridgian, lower Turonian and middle Miocene, and they separate the transgressive phase from the regressive phase within each subcycle. The lower Turonian downlap surface is easier to recognize in terms of geometrical relationships. It separates the retrogradational and progradational subphases within the Berriasian-mid Oligocene subcycle and also the transgressive phase from the regressive phase of the continental encroachment cycle. Other downlap surfaces are less visible (secondary downlap surfaces). The sediments related to these downlap surfaces are potential source rocks, particularly those within the Jurassic-mid Berriasian subcycle (i.e., the Kimmeridgian clays which are the marine source rocks of hydrocarbons produced in northern North Sea). The sediments associated with the downlap surface of the mid Berriasian-mid Oligocene subcycle (lower Turonian) have here an abnormally low organic content, and they have not been sufficiently buried to reach maturation. However, in other parts of the world, such as the Gulf of Mexico, Lake Maracaibo (Venezuela), Cabinda (Angola), they are excellent source rocks.

Paleozoic continental encroachment subcycles (Fig. 4) and the location of the likely potential source rocks in South Algeria are shown on the subcycle interpretation of the seismic line illustrated in Figure 10. Below the Hercynian unconformity,
two significant downlap surfaces are recognizable. The first one is of Late Devonian age within the base Middle Devonian-mid Carboniferous encroachment subcycle. The second one is of Early Silurian age within the base Silurian-base Middle Devonian encroachment subcycle. The downlap surface within the encroachment subcycle bounded by the late Proterozoic and base Middle Ordovician unconformities is obscured by noise on the seismic data, while that within the mid Car-boniferous-base Triassic subcycle has been eroded by the Hercynian orogeny. These downlap surfaces (Early Silurian and Late Devonian) are associated with the best source rocks in the Paleozoic, particularly Silurian shales, which have generated most hydrocarbons in Algeria.

CONCLUSIONS

In petroleum exploration, the recognition and mapping of continental encroachment cycles and subcycles strongly enhance the probability of locating the all important petroleum generative subsystem without which all the other necessary requisites for a play, (i.e., reservoir, trap and seal) become irrelevant. The focus is therefore on 1st-order and 2nd-order stratigraphic cycles. In a more practical way, one can say that the tools of choice in the exercise are: (i) a proper identification of encroachment cycles and subcycles, using all geometrical aids offered by seismic (onlap, downlap offlap breaks, etc.), (ii) recognition and mapping of major downlap surfaces. One can also note that, whereas seismic stratigraphy is generally more focused on finding reservoirs, the described methodology contributes to identifying regional seals, which is of course another major issue, and (iii) finally the fields of application are mainly new areas and areas with poor stratigraphic control. Therefore, this approach may become critical for the long-term strategy of an industry faced with an ever-increasing maturity of established petroleum provinces.

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Sequence and biostratigraphic analysis of the margin of the Apulian carbonate platform in the Montagna della Maiella (central Italy) reveal a platform margin evolution that is controlled by long-term sea-level changes, tectonism and changing platform morphology. The Upper Cretaceous to Miocene strata can be subdivided into six supersequences that are separated by deeply incised truncation surfaces. Biostratigraphy documents a major hiatus for all but one of these boundaries. The supersequences reflect distinct stages of platform development, thus the depositional systems remained the same within each supersequence but changed across the supersequence boundaries.

The Apulian platform grew on a passive margin of the Jurassic-Cretaceous (Neo-) Tethys. During the early platform history, subsidence rates decreased exponentially with time and controlled the long-term aggradation potential of the platform. The generally decreasing total subsidence rates permitted the basin in front of the platform to be filled up by the Late Campanian strata (Supersequence [SS] 1), resulting in a change from aggradation to progradation. This enabled slope carbonates of Late Campanian to Late Eocene age (SS 2 to 4 and lower part of SS 5) and finally shallow-water platform carbonates of Late Eocene to Late Miocene age (upper part of SS 5 to SS 6) to prograde basinwards.

The supersequence boundaries are to a large extent controlled by long-term (2nd-order) eustatic sea-level changes, but climate and tectonism influenced their duration and expression. Climate, initially tropical to subtropical but temperate in Miocene time, and the respective evolution of flora and fauna were major controls on sequence architecture but did not significantly influence the formation of the supersequence boundaries. The tectonic movements related to Alpine orogeny and foreland basin development were not able to completely obliterate the long-term eustatic signal but greatly enhanced the boundaries, although the exact amount of this influence cannot be assessed.

Platform morphology was very influential on sequence architecture. From at least Early Cretaceous to Late Campanian time, the presence of a steep escarpment resulted in detached sequences, consisting of an onlapping basinal part and an aggrading part on the platform top, separated by a bypass slope. In Late Campanian to Oligocene time, a distally steepened slope profile was deeply incised, most pronounced along the platform margin and the upper slope, during 2nd-order sea-level lowstands. Sea-level fluctuations along the gently inclined Miocene shelf resulted in deposition of deepening-upward sequences under conditions of low carbonate productivity.

**Abstract:** Sequence and biostratigraphic analysis of the margin of the Apulian carbonate platform in the Montagna della Maiella (central Italy) reveals a platform margin evolution that is controlled by long-term sea-level changes, tectonism and changing platform morphology. The Upper Cretaceous to Miocene strata can be subdivided into six supersequences that are separated by deeply incised truncation surfaces. Biostratigraphy documents a major hiatus for all but one of these boundaries. The supersequences reflect distinct stages of platform development, thus the depositional systems remained the same within each supersequence but changed across the supersequence boundaries.

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**Sequence Stratigraphy and the Evolution of Carbonate Platforms**

In the last few years, our understanding of the evolution of carbonate platforms has been substantially advanced by the application of sequence stratigraphy to their investigation. Studies of both reflection seismic sections (e.g., Eberli and Ginsburg, 1987, 1988, 1989; Playford et al., 1989; Sarg, 1988; Schlager, 1989, 1991) and of large-scale outcrops (e.g., Bosellini, 1984; Franseen et al., 1989; Ravenne et al., 1988; Sarg, 1988; Simo, 1989; Souquet et al., 1989; Arnaud-Vannaud and Arnaud, 1990; Jacquin et al., 1991) include pure carbonate depositional settings and facies that are well preserved in the outcrop.

**Cretaceous to Miocene Sequence Stratigraphy and Evolution of the Maiella Carbonate Platform Margin, Italy**

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**GEOLOGICAL FRAMEWORK**

The Mesozoic to mid-Tertiary platform margin of Montagna della Maiella is exposed in the north-south-trending frontal anticline of the youngest and most externally outcropping thrust sheet of the southern Apennines (Figs. 1, 2). Deep valleys and an eroded fault scarp expose wide and spectacular outcrops of the platform margin. Biogeographic evidence, both with respect to Cretaceous rudistid assemblages (Accordi et al., 1987) and Early Tertiary larger benthonic foraminifera (Pignatti, 1990), suggests that this margin was part of a much larger Jurassic to mid-Tertiary carbonate platform, the Apulian platform (Accordi and Carbone, 1992). Most of the Apulian platform was not involved in Tertiary thrusting and is still part of the unfolded foreland of the southern Apennines. Much of it is covered by younger Tertiary sediments of the Apenninic foredeep and by...
During Late Triassic to Early Jurassic times, areas that were to become the continental margins of the Tethys were affected by crustal extension that eventually led to the opening of a small ocean basin. As a result, large parts of the megabank were drowned during the early to middle Liassic, and only a few isolated carbonate platforms, separated by deeper basins, persisted (Bernoulli and Jenkyns, 1974). Deep wells in the northern part of the Maiella anticline show that the Marche-Umbria basin to the north of the Maiella platform margin also originated during Liassic rifting (cf. Crescenti et al., 1969). A strong similarity exists between these platforms and the still active ones of the Bahamian archipelago, not only in terms of facies, size and shape, time-space relationships and subsidence rates (Bernoulli, 1972; D’Argenio et al., 1975), but also in terms of internal architecture (cf. Eberli and Ginsburg, 1989).

In the Montagna della Maiella, the platform was bordered by a steep, non-depositional escarpment incised by submarine valleys during much of the Cretaceous (Fig. 3; Crescenti et al., 1969; Accarie, 1988). The escarpment shows pronounced analogies with buried escarpments of the Great Bahama Bank (cf. Eberli and Ginsburg, 1989) and those below the west Florida shelf (cf. Mullins and Hine, 1989). The locally scalloped morphology of the escarpment, outcrops in the escarpment wall of Lower Cretaceous horizontally bedded, lagoonal to supratidal carbonates deposited in an internal platform environment, as well as lithic megabreccias intercalated with the adjacent basal deposits, show that considerable submarine erosion and gravitational collapse shaped the escarpment in Cretaceous time. Most probably, the escarpment was inherited from Early Jurassic rifting (cf. Bice and Stewart, 1990).

The “middle” Cretaceous was a time of major crisis for the peri-Adriatic carbonate platforms. Along the Maiella margin and elsewhere, subaerial exposure and bauxite formation occurred during this interval (Crescenti, 1970; D’Argenio, 1970; D’Argenio and Mindszenti, 1992). In other areas such as the Caribbean and the western Atlantic, carbonate platforms are interpreted to have “drowned” during the same time interval (Bryant et al., 1969; Schlager, 1989; Winker and Buffler, 1988). This suggests that eustatic sea-level change was not the only important controlling factor and that tectonic (e.g., intraplate stresses) and environmental changes played a role. The Maiella platform was flooded again during the Middle to Late Cenomanian (Accarie, 1988; Accarie and Delamette, 1991). During Late Cretaceous platform aggradation, the steep escarpment was maintained, and in the adjacent basin to the north, lithic breccias, bioclastic turbidites and pelagic limestones were deposited. In Late Campanian time, the basin immediately adjacent to the platform was filled by these deeper marine sediments onlapping the escarpment, and bioclastic limestones prograded over the basal deposits, forming a distally steepened ramp (Accarie, 1988; Eberli et al., 1993). This stage of platform evolution was ended by a major relative sea-level fall, documented by a prominent regional truncation surface and subaerial exposure formed during the latest Maastrichtian and part of the Danian. Analogous unconformities, though less precisely dated, are present on other central Italian platforms (Accordi and Carbone, 1988).

Because of nondeposition and/or erosion, only parts of the Tertiary are recorded by sediments on the peri-Adriatic platforms (e.g., Colacicchi, 1987; Accordi and Carbone, 1988). The Maiella margin shows a relatively complete early Tertiary record along its slope, whereas long hiatuses are present on the

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**FIG. 1.**—Location of Montagna della Maiella and relative position of the carbonate platforms of Apulia (A) and Southern Limestone Apennines (B). The Apulian platform is largely autochthonous (Apulia, Monte Gargano) and partly covered by foreland deposits and allochthonous units of the Southern Apennines. The platform of the Southern Limestone Apennines is thrust onto basinal Mesozoic sequences and Tertiary flysch, whereas the Apulian platform is only marginally involved in Tertiary thrusting in the Montagna della Maiella (from Eberli et al., 1993, reprinted by permission).
CRETACEOUS TO MIOCENE SEQUENCE STRATIGRAPHY—EVOLUTION OF MAIELLA CARBONATE PLATFORM MARGIN

Fig. 2.—A. Simplified geological map of Montagna della Maiella, after Catenacci et al. (1970; from Eberli et al., 1993, reprinted by permission). B. Locations of Figures 4, 6, 7 and 11. B: Blockhaus, CD: Colle Daniele, MB: Monte del Belvedere, MF: Monte Focalone, RF: Rava del Ferro.

platform top: here Paleocene to Bartonian deposits occur only as relatively thin erosional relics. During the Middle/Late Eocene, the platform was extensively flooded, and the establishment of coralgal buildups above the Upper Eocene bioclastic limestones was followed by progradation of the shallow platform over the former slope. Much of the Oligocene time is represented by a long hiatus both on the platform and on the slope, interrupted only by a short episode of deposition around the time of the Early/Late Oligocene boundary. As over large parts of the southern Apennines (Carannante et al., 1988), aggradation resumed in the Miocene during several poorly dated episodes. The Miocene platform carbonates are relatively thin, they were deposited on a gently inclined shelf under temperate climatic conditions.

In central Italy, carbonate platform evolution was ended by clastic sedimentation in the foredeep of the advancing Apenninic orogen and by sea-level lowering during the Messinian salinity crisis (cf. Hsu et al., 1978). Thrusting began in the Montagna della Maiella in the Pliocene (De Giuli et al., 1987; Ghisetti and Vezzani, 1983). During the formation of the sedimentary décollement nappes (Bally et al., 1986), platform margins typically became the site of tectonic decoupling and thrusting; only in the Montagna della Maiella, where the former platform margin is perpendicular to the trend of the tectonic units is the platform margin entirely preserved.

METHODS OF SEQUENCE ANALYSIS
Depositional Geometry

In the Montagna della Maiella, outcrops of the platform margin that reach several kilometers in width and 2 km in height (Fig. 2) permit examination of large-scale geometries that are comparable to the ones observed on reflection seismic sections. Each valley side was photographed in a panorama view. The panoramic photomosaics were analyzed like seismic sections (i.e., the sediments were subdivided into unconformity-bounded sequences according to the procedure described by Mitchum et al., 1977 and Vail, 1988). Sections measured in the
ADAM VECSEI, DIETHARD G. K. SANDERS, DANIEL BERNOULLI, GREGOR P. EBERLI, AND JOHANNES S. PIGNATTI

FIG. 3.—A. Schematic stratigraphic platform to basin cross section, based on measured sections, showing supersequences (SS), depositional geometries and lithostratigraphy of the Maiella platform margin. The Lower Cretaceous platform (KL) is bound by a steep escarpment and unconformably overlain by Upper Cretaceous shallow-water carbonates (SS 1). Onlapping basin sediments of SS 1 bury the escarpment. After Late Campanian time, sequences are physically continuous from the platform interior onto the low-angle slope (SS 2). A major unconformity separates the Cretaceous from the Tertiary section (SS 2/3 boundary). During Paleocene through Middle Eocene time (SS 3 and 4), the former Cretaceous platform was repeatedly flooded, but most shallow-water carbonates from this interval were subsequently eroded. During the Late Eocene and Early Oligocene (SS 5), reefs prograded over the slope. The Miocene carbonates (SS 6) were deposited on a gently inclined shelf that covered both the former shallow-water platform and slope areas. Formations are from Crescenti et al. (1969) as redefined by Vecsei (1991). Many formations designate shallow-water platform facies (brick signature), or basinal and slope facies units that are part of one of several supersequences (shaded or white); only two formations coincide with supersequences (Orfento Formation = SS 2, Bolognano Formation = SS 6). The Gessoso Solfifera Formation, deposited during the Messinian salinity crisis of the Mediterranean, consists of breccias, evaporites and shales. B. Chronostratigraphic chart of the Maiella platform margin.

field provide the sedimentologic and stratigraphic information for sequence analysis.

Biostratigraphy

The biostratigraphic subdivisions used are shown in Table 1. The scheme of age units follows more or less the time-stratigraphic units of Haq et al. (1988). The Upper Cretaceous sediments are dated by globotruncanids determined in thin section, with the addition in the Campanian and Maastrichtian strata of larger foraminifera (mostly orbitoids and siderolitids; van Gorsel, 1978; Neumann, 1980). Cretaceous zonation and chronostratigraphic correlation are after Haq et al. (1988).
TABLE 1.—CORRELATION SCHEME OF DANIAN TO AQUIтанIAN CHRONOSTRATIGRAPHY AND BIOSTRATIGRAPHIC ZONATIONS.

<table>
<thead>
<tr>
<th>Time in m. y.</th>
<th>SERIES</th>
<th>PLANKTONIC FORAMINIFERAL ZONES</th>
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<td>40</td>
<td>LOWER</td>
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Chronostratigraphic subdivisions follow Haq et al. (1988). Planktonic foraminiferal zonations are those adopted in Haq et al. (1988): (1) after Blow (1969) and Berggren (1972); (2) after Stainforth et al. (1975). Larger foraminiferal zones adopted are those of Hottinger (1960), Schaub (1981), De Mulder (1975), and Drooger and Laagland (1986).
In Paleocene to Eocene strata, the planktonic foraminifera were zoned, interpreted chronostatigraphically and correlated with the nannofossil zonation of Haq et al. (1988); this zonation can be correlated easily with the zonation for planktonic foraminifera of Blow (1969), Stainforth et al. (1975; Table 1) and Bolli et al. (1985). However, the determination of planktonic foraminifera in thin section is usually difficult and sometimes unreliable. The presence of larger foraminifera, mostly nummulitids (Schaub, 1981), but in a few cases other groups like the alveolinids (Hottinger, 1960), makes a parallel zonation possible (cf. Pomeler, 1973; Hillebrandt, 1980). These larger foraminifera were important in age dating and in determining stages and their boundaries. The Oligocene and Lower Miocene (Aquitanian) strata are poor in diagnostic planktonic foraminifera but are dated with the help of larger foraminifera, particularly lepidocyclinids and to a lesser degree also miogypsinids (Vervloet, 1966; Lange, 1968; De Mulder, 1975; Schüttenhelm, 1976; Drooger and Laagland, 1986). For the correlation of the Tertiary larger foraminiferal zonations with the planktonic foraminiferal zones of Haq et al. (1988) we have used a newly elaborated scheme (Table 1).

The strata of the Maiella platform margin are divided into supersequences with parts on the shallow-water platform and parts on the slope and/or in the basin. Each supersequence is a genetic unit and corresponds in all cases but one to a distinct phase of platform evolution characterized by a distinct depositional system and typical facies associations.

The characteristics of the supersequence boundaries of the Maiella platform margin are summarized in Table 2. All supersequence boundaries are deeply incised truncation surfaces, although the depth of truncation varies between and along the boundaries. Most of the supersequence boundaries can be physically traced and biostratigraphically correlated over the entire Montagna della Maiella. In addition, biostratigraphy shows that all boundaries except one are associated with a major hiatus (Table 2). Biostratigraphic data are also in accord with the postulate that time lines do not cut but might merge along sequence boundaries (Vail et al., 1977). Along only one boundary (SS 1/2) is the hiatus not long enough to be defined by planktonic foraminiferal biostratigraphy.

The 2nd-order supersequences are composed of several 3rd-order depositional sequences formed during shorter sea-level cycles (cf. Vail et al., 1977; Van Wagener et al., 1988). This paper concentrates on the evolution of the 2nd-order supersequences. The complete data sets, further interpretative details and the more complete descriptions of the 3rd-order sequences are found in Vecsei (1991) and Sanders (1994).

**EVOLUTION AND SUPERSEQUENCES OF THE MAIELLA PLATFORM MARGIN**

The following sections describe the succession of supersequences with particular emphasis on supersequence boundaries characteristics and the depositional systems active in each supersequence. The depositional systems change from one supersequence to another, but remain unchanged within each supersequence, reflecting distinct stages of platform evolution. Six supersequences are present between Upper Cretaceous and Upper Miocene deposits. Except for thin “middle” Oligocene sediments, all deposits of the Maiella platform margin can be clearly attributed to one of the supersequences.

**Upper Jurassic to Upper Albian Succession**

Upper Jurassic to Upper Albian carbonates form the substratum of the Upper Cretaceous to mid-Tertiary supersequences and comprise the oldest deposits outcropping in the Montagna della Maiella. These sediments are of two areally separated facies: carbonate platform in the south and basinal in the north (Figs. 3, 4). The carbonate platform succession is part of the Morrone di Pacentro Formation of Crescenti et al. (1969; Fig. 3). Although the platform-basin transition is not exposed in the

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**Table 2.—CHARACTERISTICS OF THE SUPERSEQUENCE BOUNDARIES ON THE MAIELLA PLATFORM MARGIN.**

<table>
<thead>
<tr>
<th>Supersequence</th>
<th>Platform Characteristics</th>
<th>Slope or basin Characteristics</th>
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<tbody>
<tr>
<td>SS 6</td>
<td>Geometry at base</td>
<td>deep truncation, locally channels</td>
</tr>
<tr>
<td>H hiatus at base</td>
<td>Late Cretaceous-Langhian (c. 50 my)</td>
<td>Early Ripelian-late Chattian/early Aquitanian (c. 9 my)*&lt;br&gt;(none)</td>
</tr>
<tr>
<td>SS 5</td>
<td>Evidence for exposure at top</td>
<td>microkarst, soils with Microcsubbotinae</td>
</tr>
<tr>
<td>Geometry at base</td>
<td>Late Maastrichtian-early Priabonian (c. 28 my)</td>
<td>within Bartonian (&lt;2 my)</td>
</tr>
<tr>
<td>H hiatus at base</td>
<td>microkarst, soils with Microcsubbotinae</td>
<td>(none)</td>
</tr>
<tr>
<td>SS 4</td>
<td>Evidence for exposure at top</td>
<td>truncation, channels</td>
</tr>
<tr>
<td>Geometry at base</td>
<td>Ypresian p.p.-Maastrichtian p.p. (&gt;12 my)</td>
<td>within Ypresian (M. subbotinae zone, &gt;0.5 my)</td>
</tr>
<tr>
<td>H hiatus at base</td>
<td>Ypresian p.p.-Campanian (&gt;20 my)</td>
<td>(none)</td>
</tr>
<tr>
<td>SS 3</td>
<td>Evidence for exposure at top</td>
<td>microkarst, soils with Microcsubbotinae</td>
</tr>
<tr>
<td>Geometry at base</td>
<td>up to c. 50 m truncation, many channels</td>
<td>(none)</td>
</tr>
<tr>
<td>H hiatus at base</td>
<td>?Ypresian p.p.-Campanian (?&gt;20 my)</td>
<td>?same as on platform</td>
</tr>
<tr>
<td>SS 2</td>
<td>Evidence for exposure at top and within SS</td>
<td>meteoric dia genesis indicated by secondary porosity, silicification</td>
</tr>
<tr>
<td>Geometry at base</td>
<td>deep truncation (≤100 m)</td>
<td>within Late Campanian (G. calcarata zone, &lt;1.5 my)</td>
</tr>
<tr>
<td>H hiatus at base</td>
<td>?mid-Campanian-Maastrichtian p.p. (&lt;10 my)</td>
<td>(?± straight (little exposed)</td>
</tr>
<tr>
<td>SS 1</td>
<td>Evidence for exposure at top</td>
<td>karst, caliche</td>
</tr>
<tr>
<td>Geometry at base</td>
<td>up to &gt;100 m truncation</td>
<td>?none****</td>
</tr>
<tr>
<td>H hiatus at base</td>
<td>Middle Albian-Late Cenomanian*** (&gt;4 my)</td>
<td>(none)</td>
</tr>
<tr>
<td>KL</td>
<td>Evidence for exposure at top</td>
<td>several 10s of meters truncation, deep karst and bauxite</td>
</tr>
</tbody>
</table>

---

*Geometries observed along the base, time spans of hiatuses determined along the base, and evidence for subaerial exposure observed at the top and within the supersequences (SS) and their substratum (KL). The maximal time span of the hiatuses on the platform and their minimal time span on the slope are given.

**from Moussavian and Vecsei (1995)

***from Accarie (1988) and Accarie and Delamette (1991)

****from Accarie and Deconinck (1989)
CRETACEOUS TO MIOCENE SEQUENCE STRATIGRAPHY—EVOLUTION OF MAIELLA CARBONATE PLATFORM MARGIN

Fig. 4.—Panoramic view of the central Montagna della Maiella, from Blockhaus. Deep valleys expose the more or less horizontally bedded Cretaceous shallow-water platform (KL = Lower Cretaceous, and SS 1) and the 35° inclined escarpment that borders the platform to the north. The northern flank of Cima delle Murelle is possibly an exhumed part of the escarpment. The lower part of the escarpment is onlapped by Cenomanian to Upper Campanian basin deposits, including megabreccias of Supersequence (SS) 1. Slope carbonates of SS 2 (Upper Campanian to uppermost Maastrichtian) form a wedge thickening basinwards; slope deposits of SS 2 also onlap and bury the escarpment and submarine valleys, and finally overstep the platform top, which is almost flat here. Slope deposits of SS 3 and 4 fill wide channels incised into the top of SS 2, and are locally eroded below SS 5. A thick series of prograding bioclastic limestones with a few reef knolls forms the lower, Priabonian part of SS 5 that buried the previous relief. Below Pesco Falcone, coralgal reefs were reestablished during the early Rupelian stage, and prograded laterally and basinward over the slope (after Vecsei, 1991).

Lower Cretaceous strata, platform and basin already must have been separated by a precursor of the steep, erosional escarpment limiting the platform to the north.

In the southern and central Montagna della Maiella, the pre-“middle” Cretaceous carbonate platform deposits are as old as Late Jurassic age (Crescenti et al., 1969) and are mostly composed of cyclically bedded shallow subtidal to supratidal limestones with few intercalations of rudist biostromes near the platform margin. Their exposed thickness is estimated at 1400 m (Crescenti et al., 1969; Accarie et al., 1986; Accarie, 1988). In the escarpment, internal platform facies are outcropping, indicating subsequent erosion of the platform margin. Thus only part of the depositional system of the platform is preserved.

Time-equivalent basinal deposits occur only at a few localities in the northern Maiella, where up to 50 m of pelagic limestones, bioclastic turbidites, breccias and pebbly mudstones of Late Albian age are exposed (Accarie and Deconinck, 1989; observations by D. Bernoulli, 1994). In a drill hole approximately 20 km north of the escarpment, Lower Cretaceous pelagic limestones (Maiolica Formation), with interbedded black shales in the Aptian-Albian interval (Marne a Fucoidi Formation), are the basinal equivalents (Crescenti et al., 1969).

Shallow-water carbonate production was interrupted in the “middle” Cretaceous by long-lasting subaerial exposure. Locally deep karstic cavities formed and were filled by breccias (Fig. 5A) and partly by pisolithic bauxite. In some places, an angular unconformity is observed. It is not clear, however, whether this is due to local tilting caused by intraplate karstification or whether it reflects tectonic tilting during the “middle” Cretaceous.

Supersequence 1 (Middle Cenomanian to Upper Campanian Substages)

In the Middle to Late Cenomanian, the platform was flooded again and shallow-water carbonates were deposited above the supersequence boundary between the underlying platform carbonates and SS 1 (the mid-Cretaceous unconformity). Accarie (1988) and Accarie and Delamette (1991) have documented a hiatus of variable duration along this boundary that maximally spans the Middle Albian to Late Cenomanian interval and minimally comprises Late Albian to Early Cenomanian time.

Until Late Campanian time, the shallow platform was separated from the basin by a steep submarine escarpment (Crescenti et al., 1969; Accarie, 1988). At times, this escarpment may have reached a height of 1000 m. Along the north face of Cima della Murelle, the escarpment is inclined about 35°, which approximately corresponds to its slope in the Late Cretaceous Period. Along the undulatory escarpment surface, the cores of Upper Cretaceous rudist biostromes are exposed. This facies distribution shows that the escarpment was continuously shaped by mass wasting and submarine erosion but that it retained its steep angle during aggradation as long as it was not buried by the onlapping sediments (Figs. 3, 4). When the escarpment was
eventually buried during latest Cretaceous time, headward erosion had created a system of submarine valleys isolating part of the platform.

Different cycle stacking patterns are observed on the platform interior and along the platform margin, respectively. On the platform interior, a succession of meter-scale cycles was deposited. The peritidal cycles in the lower part of this succession are arranged in an overall thinning- and shoaling-upward stacking pattern. The upper part of the succession is composed of thicker cycles with grainstones and rudist-oyster biostromes, indicating a deepening of facies. These younger cycles are stacked in a retrogradational stacking pattern that is followed up-section by a progradational pattern. These systematic thickness variations suggest a subdivision of SS 1 on the platform interior into two sets of sequences.

Along the platform margin, cycles dominated by bioclastic sand bodies were deposited in a 2.5- to 3-km-wide belt behind the platform edge. Bioclastic sand bodies make up the lower part of these cycles, while the upper part consists of rudist biostromes (Figs. 5B, C) and of inter- to supratidal limestones. The rudist assemblages are typical for high-energy platform environments (Accordi et al., 1987). The lower part of the external platform succession, consisting of cycles arranged in a progradational pattern, is again separated from the upper part of the
succession by a surface along which deepening of facies occurs and which is correlated with the analogous surface observed on the internal platform. The bioclastic sand bodies consist of northward (i.e., offbank-dipping sigmoid to oblique) master-beds (Fig. 5C). These are composed of a coarse biodetrital lower part and a bi-directionally cross-laminated bioclastic-oobioctlastic upper part. Both truncation and downlap surfaces are present within the sand bodies and, together with the bi-directional cross-lamination, suggest progradation under the influence of storms and tides.

Excess sediment produced on the platform and lithified clasts reworked from its margins bypassed the escarpment. In Cenomanian time, a wedge-shaped talus consisting of breccias and coarse bioclastic sand was deposited along the foot of the escarpment (Accarie, 1988; Valle dell’Inferno Formation). This wedge and the remaining escarpment are onlapped by a unit of megabreccias, turbidites and pelagic limestones (Fig. 3; Tre Grotte Formation). The clast-supported megabreccia beds, up to 50 m thick, are either single beds or, more often, amalgamated (Fig. 5D). Their chaotic internal organization and the scarcity of matrix suggest deposition by rock avalanche and, possibly, by debris flow. The megabreccias are intercalated with pelagic limestones and calcarenites (Fig. 5D). The pelagic background sediment probably is a mixture of coccolith ooze and winnowed bank-top derived carbonate lutum (i.e., periplatform ooze). The calcarenites consist mainly of biogenic debris; they are identified as turbidites, partly reworked by contour currents, based on grading and parallel or cross-laminations.

The megabreccias fill shallow channels that truncate the underlying beds. Their composition is dominated by clasts of lithified platform limestones. They are interpreted as the products of platform margin erosion and collapse probably during relative sea-level lowstands, whereas the pelagic limestones and intercalated bioclastic turbidites might represent in large part unconsolidated sediment exported from the active, producing platform during relative sea-level rises and highstands. Based on these criteria, seven 3rd-order sequences have been distinguished in the basin within SS 1.

The depositional system of SS 1 ended with the formation of the upper supersequence boundary, along which local subaerial exposure occurred (Eberli et al., 1993; Mutti, 1995; Fig. 6). At the platform edge, the boundary runs along the interface between the shallow-water platform of SS 1 and the onlapping upper slope carbonates of SS 2 (e.g., at Monte Rotondo on the right side of Fig. 4). The boundary of SS 1/2 can be traced on the slope over most of the northern Maiella.

Supersequence 2 (Upper Campanian to Uppermost Maastrichtian Substages)

On the slope, deposition of SS 2 started in Late Campanian time above the SS 1/2 boundary, which is a slight truncation surface changing downslope to a conformity. Thus, this is the only supersequence boundary across which no hiatus could be proved biostratigraphically on the slope; pelagic limestones immediately below and above the boundary were dated as the Late Campanian Globotruncanita calcarata zone. On the platform, the SS 2/3 boundary is a surface characterized by strong truncation, but the associated hiatus could not be determined exactly.

In Late Campanian to Early Maastrichtian time, the depositional pattern changed from aggradation to progradation; the former basin and the intra-platform valleys were largely filled, and a wedge of carbonate sands and breccias (Orfento Formation, SS 2, Fig. 3) was deposited on the former platform margin and upper slope. Within this wedge, lithic turbidites and mass-flow deposits are over lain by bioclastic sand waves. These sand waves prograded towards the basin over a gently inclined, distally steepened ramp that extended above the former basin and gradually encroached onto the shallow-water platform (Figs. 4, 7). In situ rudistid buildups are preserved only locally along the platform margin that, during the course of Maastrichtian time, was also covered by offbank transported carbonate sands.

The carbonate sands of SS 2 onlap the eroded top of the platform along the basal boundary of SS 2 (Fig. 7). In other places, the carbonate sands show a downlap onto the underlying carbonate platform. The slope deposits of SS 2 are bioclastic
limestones made up mostly of rudist debris. Other constituents are larger benthonic foraminifera, coral, red algae, and intraclasts. The sand waves are up to 20 m high and a few hundreds of meters wide (Fig. 8A). They prograded unidirectionally downslope, probably under the influence of ebb-dominated tidal currents. Above the former platform, the sand waves cannot be traced as they laterally merge to form sand bodies cross-bedded on the scale of several meters. A few thin beds of pelagic limestone are intercalated with these deposits on the lower slope. Additional facies are breccias with shallow-water limestone clasts on the lower and upper parts on the slope (Fig. 8B) and breccias with redeposited rudists on the upper slope. These breccias occur above truncation surfaces that subdivide SS 2 into four 3rd-order depositional sequences with a similar internal organization (Figs. 4, 7), and have been interpreted as lowstand breccias.

The youngest beds of SS 2 are Late Maastrichtian age (late G. gansseri or A. mayaroensis zone). Subaerial exposure during formation of the supersequence boundary is suggested by the development of important secondary porosity and silification of SS 2 limestones, probably caused by the circulation of meteoric waters (cf. Knauth, 1979; Mutti, 1995).

**Supersequence 3 (Danian to “Middle” or Upper Ypresian Substages)**

The SS 2/3 boundary is a truncation surface that in places reaches to a depth of 50 m and more (Monte del Belvedere, Fig. 7). Limestones of SS 3 fill the deep erosional relief along the top of SS 2. In addition to the large-scale erosion, the sediments of SS 2 are truncated in many places by shallow channels filled by lithic breccias of ?Early Paleocene and Thanetian age (Figs. 7, 9A).

SS 3 marks a turning point in the evolution of the platform margin. During the earliest Tertiary Period, deposition on the Maiella platform margin was minimal, and if sediments were deposited, they were eroded shortly thereafter. A shallow-water platform of late Thanetian age was reworked, and its products redeposited as channelized gravity flow deposits. These deposits are intercalated with pelagic and bioclastic limestones that mark a rapid deepening over large parts of the platform margin. Thus nondeposition and erosion were the dominant processes on the platform and its upper slope.

No persistent shallow-water platform was established between latest Maastrichtian and middle to late Ypresian times. However, upper Danian-lower Thanetian coralgal reefs displaced downslope as slide blocks are preserved at the base of SS 3 along the northern flank of Valle delle Tre Grotte (Mousnavian and Vecsei, 1995). Together with the dating of the top of SS 2 as late 'G. gansseri' or early A. mayaroensis zone, these reef sediments and the 'Lower Paleocene lithic breccias constrain the age of the supersequence boundary as latest Maastrichtian (near the base or within the A. mayaroensis zone) to Danian. In the Thanetian stage, extensive reefs must have existed further up the platform, but they were eroded during late
Thanetian time and are found only as large volumes of clasts in channel fills.

Channel fills are nearly the only sedimentary record of this time interval on the platform and on the upper slope, which also displays erosion and local nondeposition. The channels are complex, often amalgamated, and younger channels frequently cut older ones. Channel width and depth generally increase upsection, with the width reaching several hundreds of meters. Calcirudite beds in the channels become laterally more extensive and intercalations of calcarenites and pelagic limestones become more common as channel size increases. The calcirudite and calcarenite beds are turbidites and other gravity flow deposits containing lithic and bioclastic sand and gravel transported downslope. The lithoclasts are indurated shallow-water limestone clasts, mostly eroded from older sediments within SS 3, but near the basal supersequence boundary, clasts eroded from the underlying SS 2 are also abundant. Small benthonic foraminifera, green and red algae, and larger foraminifera are from the underlying SS 2 are also abundant. Small benthonic foraminifera, green and red algae, and larger foraminifera are the main bioclastic components.

There is no evidence of subaerial exposure along the top of SS 3 either on the slope or the platform. However, clasts with Microcodium SS 3 either on the slope or the platform. However, clasts with *Microcodium* and microkarst redeposited onto the slope indicate subaerial exposure of the platform top.

**Supersequence 4 (“Middle” or Upper Ypresian to Bartonian Substages)**

Sedimentation resumed on the lower slope in middle to late Ypresian time above a surface characterized by slight truncation where underlain by sediments of SS 3. The minimum duration of the hiatus along this part of the basal supersequence boundary cannot be determined exactly by biostratigraphy; however, it lies within the interval between the *M. formosa formosa/G. aragonensis* zones of the middle Ypresian time and the *A. pentacamerata* zone of the late Ypresian time. On the platform, times of nondeposition and erosion prevailed during middle or late Ypresian to Bartonian stages, but are more difficult to date.

A shallow-water platform was reestablished briefly but was eroded subsequently. Thus, SS 3 and 4 are laterally discontinuous on the former platform top and on the upper slope and at places the SS boundaries 2/3 and 3/4, and 3/4 and 4/5, respectively, merge (Fig. 3). As a result, SS 4 may directly overlie SS 2.

On the platform in the southernmost Montagna della Maiella, grainstones rich in alveolinids occur (Bally, 1954). The grainstones form large-scale cross-bedded to massive, up to 10-m-thick bodies that locally show a convex upper relief. Aside from alveolinids, they are made up of miliolids, bryozoans, red algae, larger benthonic foraminifera (gyspiniids, soritids) and encrusting foraminifera. They are, on a scale of meters, interbedded with plankton-bearing hemipelagic limestones that contain graded bioclastic beds.

Farther north on the former platform, channels similar in geometry to the channels described in SS 3 were filled with litho- and bioclastic components. Sediment funneled from the platform through these channels bypassed the uppermost part of the slope and was deposited further downslope as a succession of bioclastic and lithic turbidites up to 40 m thick intercalated with pelagic limestones (Fig. 9B). Reconstruction of the slope reveals a concave-upward upper surface of this pile of slope sediments (Fig. 3), suggesting deposition on a submarine fan. Truncation surfaces along the base of the lithic turbidites allow identification of at least four 3rd-order sequences within this supersequence. The bioclastic turbidites are mostly composed of larger foraminifera, red algae, small benthonic foraminifera and echinoids, typical for a deeper shelf or a temperate climate zone. Organisms characterizing tropical or subtropical shallow photic conditions are conspicuously rare. The lithic breccias contain clasts eroded from older platform deposits, including hemipelagic and shallow-water limestones of SS 3, 4 and in places also of SS 2. Deep channels and slumps are relatively frequent in the slope sediments of SS 4 (Fig. 9B).

Subaerial exposure of the platform after deposition of SS 4 is indicated only by clasts redeposited onto the slope that contain *Microcodium* or microkarst. On the lower slope, this boundary can be observed in only a few places where it is not sufficiently exposed for analyzing its exact geometry and characteristics.

**Supersequence 5 (Bartonian to Lower Oligocene Stages)**

The basal boundary of SS 5 is associated with deep erosional truncation on the platform and the slope. As a result, SS 5 lies directly on SS 2 (e.g., along most of its outcrop in Fig. 4) over large areas. Thick channelized lithic breccias with characteristic nummulites in the matrix were deposited above the truncation
ADAM VECSEI, DIETHARD G. K. SANDERS, DANIEL BERNOULLI, GREGOR P. EBERLI, AND JOHANNES S. PIGNATTI

FIG. 9.—Facies of Supersequences (SS) 3 and 4. (A) Along the lower boundary of SS 3, a channel cut into the bioclastic sands of SS 2 (arrows) is filled by a breccia with lower Tertiary platform limestone clasts. Note the flat top typical for many channel fills. Lower Valle del Orfento. (B) Part of the lower slope, where two 3rd-order sequences in SS 4 are separated by a low-angle truncation surface (SB, bottom marked by line of arrows) overlain by lithic breccia beds. The major part of the sequences consists of pelagic limestones and bioclastic turbidites. Note channels (C) with steep and with gently inclined walls (line of arrows, lower right) and slumps (S) with contorted bedding. Lower Valle del Orfento.

With progradation of the shallow-water areas, larger marginal coralgal reefs were established over most of the pre-existing topographic highs and partially also on the shallowest part of the slope (Figs. 12A, B). Subsequently these reefs prograded at least 4 km basinwards over the gently inclined upper slope (Fig. 4 on both sides of Pesco Falcone).

Coeval limestones on the lower slope strongly reflect the change in platform evolution: depositional rates drastically increased in SS 5 due to shedding from the active platform, and the amount of bioclastic material (about 95%) in the lower slope limestones greatly exceeds lithoclastic and pelagic deposition. The turbidite sequences deposited on the lower slope are divided into six 3rd-order depositional sequences (Fig. 11). Most of the sequences are composed of a coarsening and thickening upward cycle (Fig. 12C), interpreted to be the result of increasing shedding during 3rd-order sea-level highstand progradation of the shallow-water platform. Components of the bioclastic beds are mainly large and small foraminifera, green and red algae, bivalves and crinoids. These components reflect the production of organisms in the photic zone. Lithic breccias on the slope are also different from those of the underlying SS 2 and 3; they are exclusively eroded from the penecontemporaneous reefs along the platform margin, and older constituents are lacking. These breccias are interpreted as the lowstand deposits of the 3rd-order sequences.

The local occurrence of Microcodium in the uppermost reefs of SS 5 suggests subaerial exposure and diagenesis (cf. Klappa, 1980; Esteban and Klappa, 1983). However, Microcodium occurs also along exposure horizons deeper in SS 5. Clasts with microkarst eroded from SS 5 during formation of the boundary also indicate subaerial exposure.

“Middle” Oligocene Limestones of Uncertain Supersequence Affinity

Limestones of “middle” Oligocene age, exposed in a few small and isolated outcrops in the northwestern Montagna della Maiella, cannot be allocated to a specific supersequence with certainty, nor are there sufficient biostratigraphic data to define surface that forms the basal supersequence boundary. On the lower part of the slope, these breccias are sometimes deformed by slumping. Lateral tracing of the breccias allows definition of the lower boundary of SS 5 in places where large-scale outcrops are missing. On the lower slope, the hiatus along this boundary is dated with nummulites and includes a time interval within the Bartonian. Lithic breccias on the slope are also different from those of the underlying SS 2 and 3; they are exclusively eroded from the penecontemporaneous reefs along the platform margin, and older constituents are lacking. These breccias are interpreted as the lowstand deposits of the 3rd-order sequences.

SS 5 is on the order of 180 m thick over both earlier platform and slope areas (Figs. 3, 11). This indicates that significant accommodation space was created on the morphological high above the previous platform.

In earliest Priabonian time, slope carbonates were deposited in the entire Montagna della Maiella (e.g., at Monte Rotondo and Monte Focalone on Fig. 4). Shallow-water deposition was restricted to the south beyond the exposed area, where a platform must have become reestablished in the late Bartonian to early Rupelian. This Tertiary platform was rimmed by a gently inclined slope (right side of Fig. 4, and its continuation down-slope on Fig. 11).

During the early Priabonian, sediments of the uppermost slope prograded basinwards. They consist mainly of tempestites with local small patch reefs (below Pesco Falcone on Fig. 4).

“Middle” Oligocene Limestones of Uncertain Supersequence Affinity

Limestones of “middle” Oligocene age, exposed in a few small and isolated outcrops in the northwestern Montagna della Maiella, cannot be allocated to a specific supersequence with certainty, nor are there sufficient biostratigraphic data to define

![Fig. 10.—The duration of the hiatus along the basal boundary of Supersequence (SS) 5 increases up-slope (from northwest to southeast) from an undetermined interval within the Bartonian in section S. Croce, where SS 5 is underlain by SS 4; through intervals spanning Early Paleocene or Thanetian to Bartonian time at Rava Cupa and Pesco Falcone, where SS 3 underlies SS 5 and SS 4 is missing, to the ?Maastrichtian-Priabonian interval at Monte Rotondo, where SS 5 directly overlies SS 2. The basal supersequence boundary everywhere contains an isochron in the Bartonian. Numerical ages are according to Haq et al. (1988; from Vecsei, 1991).]
CRETACEOUS TO MIocene SEQUENCE STRATIGRAPHY—EVOLUTION OF MAIELLA CARBONATE PLATFORM MARGIN

Fig. 11.—Panoramic view of the middle part of Valle del Orfento. The slope series of Supersequence (SS) 4 is here reduced to an amalgamated stack of thick lithic breccias. SS 5 (Bartonian to lower Rupelian) is on this part of the slope subdivided into six sequences bounded by truncation surfaces. The thin unit of uncertain supersequence allocation (upper Rupelian to lower Chattian) is separated from SS 5 by another truncation surface. The lower three sequences of SS 6 are exposed; each consists of a lower bioclastic facies unit and an upper pelagic marly limestone unit. The boundaries of these sequences are channelized truncation surfaces in this section.

Upper Rupelian to lower Chattian patch reefs are located a few kilometers basinward of the lower Rupelian reefs of SS 5 (Fig. 3). Their composition is very similar to that of the lower Rupelian patch reefs, except for the occurrence of small lepidocyclinids instead of lower Rupelian nummulitids. Hermatypic corals indicate that the Apulian platform was in the subtropical climate zone during "middle" Oligocene time. The small size and the isolation of the reef outcrop do not permit determination of the depositional geometry along its base.

However, there must be a relatively long hiatus along the interface between SS 5 and the "middle" Oligocene strata as indicated by the marked faunal change.

On the slope, above the truncation surface along the top of SS 5, a few meters of bioclastic turbidites similar to those in the upper part of SS 5 are present. Their bioclasts, derived from a shallow-water eurytopic zone, include small lepidocyclinids also indicating a late Rupelian/early Chattian age.

Supersequence 6 (Uppermost Chattian/Aquitanian to ?Lower Messinian Substages)

The basal boundary of SS 6 is a truncation surface recognizable across the whole Montagna della Maiella. Erosion was
most pronounced in the southwestern Maiella where it cut down to Upper Cretaceous or Lower Tertiary strata (Bally, 1954; Crescenti et al., 1969; Catenacci et al., 1982).

During Miocene time, the depositional system of the Maiella platform was different from that of the previous supersequences. The low-angle slope of the early Tertiary Period evolved during latest Chattian to Early Miocene time into a gently inclined carbonate shelf. This change indicates that the slope had previously been filled and that the relief between platform and slope was greatly reduced. As a result, SS 6 overlies SS 5 along the former slope and platform margin areas; the corresponding hiatus spans almost the entire Oligocene and large parts of the Priabonian time, except where “middle” Oligocene reefs and turbidites (of uncertain supersequence allocation) are intercalated between SS 5 and 6. In the southwestern Maiella, the Miocene (Langhian) deposits of SS 6 overlie the Upper Cretaceous carbonate platform deposits (SS 1) along an angular unconformity (Catenacci et al., 1982; Accarie, 1988).

The Miocene succession is subdivided into four vertically stacked 3rd-order sequences, dominated by cross-bedded grainstones (Fig. 11). In the lower two sequences (Fig. 13), the grainstones contain mainly transported and broken benthonic foraminifera and bryozoans, whereas red algal rhodoliths dominate in the upper two sequences. Both small and large benthonic foraminifera are abundant in the lower part of all four sequences. These biota indicate deposition in relatively shallow water, probably within or slightly below the euphotic zone, during 3rd-order sea-level lowstands and transgression. The car-
CRETACEOUS TO MIocene SEQUENCE STRATIGRAPHY—EVOLUTION OF MAIELLA CARBONATE PLATFORM MARGIN

Fig. 13.—Facies of Supersequence (SS) 6. At the north side of Valle S. Bartolomeo, the two lower 3rd-order sequences within SS 6 are preserved between the truncation surfaces of the lower and upper supersequence boundaries (SSB, big arrows). The lower facies units of both sequences, forming cliffs, consist of cross-bedded bioclastic limestones. Foresets and cross-bedding are visible in the lower sequence. The beds are arranged into sand waves, bounded by master bedding surfaces that can be followed along the width of the outcrop. The upper facies units of both sequences, weathering back, are composed of pelagic marly limestones.

Bonate production rate of these organisms was relatively low; consequently the shelf “drowned” during the sea-level rises and highstands, and marly pelagic carbonates with abundant planktonic foraminifera were deposited. These pelagic units thin and eventually disappear on the upper part of the shelf because of non-deposition or subsequent erosion. The faunal and floral associations of SS 6 show that the Maiella had left the subtropical climate zone and entered the temperate zone (cf. Carannante et al., 1988).

EXPOSURE VERSUS SUBMARINE EROSION ALONG THE SUPERSEQUENCE BOUNDARIES

Van Wagoner et al. (1988) and Vail (1988) postulated that the landward portions of sequence boundaries are truncation surfaces formed during subaerial exposure, whereas submarine erosional processes would be active along the boundaries below sea level. Strong truncation and erosion of the platform (Table 2) clearly show that the supersequence boundaries of the Maiella platform margin formed during periods of major base level lowering that, at least in part, are connected to relative sea-level lowstands.

Clear evidence for subaerial exposure exists only along two supersequence boundaries on the Maiella platform margin, but there are good indications for exposure along all supersequence boundaries (Table 2). The mid-Cretaceous unconformity is associated with karstification and bauxite formation. Karst morphology and associated caliche also document subaerial exposure along the platform portion of the boundary of SS 1/2 (Mutti, 1995). There is evidence for subaerial exposure during the formation of the SS 2/3 boundary by the development of secondary porosity and local silicification in the slope sediments of SS 2. Boundaries of SS 3/4 and 4/5 are deep erosional truncations on the platform top, and up-slope subaerial exposure is only indicated in clasts redeposited onto the slope that contain Microcodium or microkarst. The local occurrence of Microcodium in SS 5 indicates the existence of paleosols and thus of subaerial exposure that probably occurred during the formation of sequence boundaries in SS 5 and between SS 5 and 6.

The scarcity of preserved karstification and related features along some of the supersequence boundaries is striking. We speculate that truncation during sea-level lowstands and the subsequent flooding were so vigorous that most deposits related to subaerial exposure were eroded. The great depth of truncation observed along several supersequence boundaries and the abundance of lithoclasts with signatures of subaerial diagenesis, which were eroded from the platform and redeposited in the breccias overlying the supersequence boundaries, are in line with this interpretation. However, the truncation at the platform margin and the upper slope along SS boundaries 3/4 and 4/5 may be submarine.

The processes responsible for submarine erosion along the slope portions of the supersequence boundaries are difficult to reconstruct. Shanmugam (1988) proposed that submarine erosion along sequence boundaries is due first to mass movements. Indeed, there are slumps and debris flow deposits along the boundaries of SS 2/3 and 4/5. However, there is no preferential occurrence of slumps and other resedimented units along these boundaries, as they are also present within the supersequences. Shallow channels at the platform margin, the upper slope and, although less commonly, also on the lower slope have been observed at most of the supersequence boundaries. Their origin may be associated with erosion by mass movement processes, although tidal currents may also have played a role.
These observations show that the criteria postulated for the recognition of supersequence boundaries in seismic sections by Van Wagoner et al. (1988) and Vail (1988) can be applied also in good outcrop, even though the lateral extent of the observable geometries may be inferior to that seen on many seismic sections.

CONTROLS ON PLATFORM EVOLUTION

In order to understand the main factors that have influenced the evolution of the Maiella carbonate platform margin, we analyze the subsidence and aggradation of the platform, determine the history of relative sea level and possible eustatic influence, and finally take into account the effect of climate and the evolution of the shallow marine fauna and flora (Fig. 14).

Subsidence and Aggradation History

Subsidence is of particular importance because it controls the long-term aggradation potential of the platform. Although the pre-Upper Jurassic substratum of the Maiella platform is unknown, its history may be similar to that of the other peri-Adriatic platforms that formed on the same passive margin. Where the subsidence history of this margin can be established with some confidence, it shows an exponentially decaying thermal subsidence following initial rifting subsidence (Winterer and Bosellini, 1981; Bertotti, 1991).

Figure 15 shows an aggradation curve for the Maiella platform determined from the thickness of the supersequences and their substratum, measured in biostratigraphically well-defined time intervals. The rates of aggradation generally decreased with time from the Late Jurassic to the Late Cretaceous Period, whereas there was only discontinuous aggradation during the Tertiary Period. A smoothed and averaged curve fitted to the aggradation curve is assumed to be a crude approximation of the total subsidence (i.e., the sum of tectonic subsidence and isostatic adjustment of sediment and water load during the Jurassic-Cretaceous interval). Total (and tectonic) subsidence rates decreased approximately exponentially with time during the Late Jurassic and the Cretaceous time, indicating that thermally induced post-rift subsidence of the continental margin was the main controlling factor (cf. McKenzie, 1978). This decrease of subsidence rates allowed sediment production rates to exceed the subsidence rates significantly during Late Cretaceous time. Consequently, the basin in front of the platform was filled with redeposited sediments and peri-platform and pelagic ooze by the Late Campanian. This in turn allowed progradation of slope carbonates over basinal deposits from Late Campanian onwards and finally progradation of the shallow-water platform over the slope carbonates from the Priabonian stage onwards.

In the central Mediterranean area, the change from divergent to convergent movements between Adria and Eurasia started as early as Late Jurassic time with subduction and obduction of ophiolites along the eastern margin of Adria (e.g., Dercourt et al., 1986). From the “middle” Cretaceous onwards, Alpine orogeny appears to have influenced the subsidence history of...
large parts of the Adria microplate. A possible slight tilting of the Maiella margin and emergence of large parts of the Apulian and South Apenninic platforms during the Albian-Early Turonian interval might be the first indications of convergence in this area. These movements might reflect crustal deformation due to intraplate stresses (cf. Cloetingh, 1991), connected with continent/continent collision along the northern Adriatic margin in the east Alpine/Carpathian area (Eberli, 1991). In the Maiella, the long phases of erosion during Paleocene to Bartonian time (SS 3 and 4) as well as the Priabonian to early Rupelian pulse of aggradation (SS 5), suggest tectonic enhancement of the sealevel signal. We speculate that intraplate stresses related to the beginning of collision in the western Mediterranean area are responsible for uplift and subsidence that resulted in prolonged emergence and flooding of the platform.

In the Miocene Epoch, the prograding emplacement of the Apenninic nappes onto the Adriatic foreland must have influenced uplift (foreland bulge) and subsidence (foreland basin) of the Maiella platform margin before its incorporation into the fold and thrust belt (cf. De Giuli et al., 1987). Vail et al. (1991) have shown that total subsidence curves typically take a convex-upward shape under the influence of foreland tectonic pulses, with phases of increased subsidence during nappe advancement. In the Miocene time, aggradation and total subsidence rates are on the order of several tens of meters in 0.5 to a few millions of years (i.e., similar to the amplitudes of 2nd-order and of major 3rd-order sea-level changes; cf. Haq et al., 1988). Unfortunately, Miocene time resolution is not good enough to permit the determination of a more detailed subsidence curve that could give a better record of the effects of foreland tectonics.

Relative and Eustatic Sea-Level Changes

The history of major fluctuations of relative sea-level along the Maiella platform margin during Cretaceous to Miocene time can be tentatively reconstructed from the alternating phases of platform aggradation, indicating relative sea-level highstands and phases of platform exposure or erosion (during the formation of the supersequence boundaries), suggesting phases of relative sea-level lowstands (Fig. 16).

In the case of an isolated, flat-topped and, during much of its history, steep-flanked carbonate platform like the Apulian platform, the amplitude of relative sea-level changes cannot be determined by coastal onlap. Therefore, determination of relative sea-level changes is here restricted to the qualitative representation of the periods during which 2nd-order supersequences, 3rd-order sequences and their boundaries formed. Flooding is indicated by deposition on the platform top. However, only a relatively small amount of time is represented by the Upper Cretaceous to Paleogene limestones on the platform top due both to erosion after deposition and to non-deposition. Even where shallow-water limestones have been preserved, their biostratigraphic resolution is generally poor. Therefore basinal and slope sequences, datable with much greater precision, are used to date sea-level lowstands, rises and highstands (cf. Vail et al., 1977; Haq et al., 1987, 1988). Sea-level lowstands are indicated by regional truncation at 2nd-order supersequence and 3rd-order sequence boundaries, and additionally by platform-derived lithified clasts in breccias and turbidities found in the basin and on the slope.

The fluctuations of relative sea level are caused by the combined effect of eustasy and tectonism. In order to constrain the eustatic effects, we compare the phases of aggradation with the onlap chart of Haq et al. (1987, 1988). We assume that at least the major 2nd-order sea-level excursions recorded on Haq et al.’s (1988) curve reflect eustatic events, although their ages may be offset in time by tectonism (cf. Christie-Blick, 1991). We also recognize the problems regarding the 3rd-order sea-level curve (cf. Kendall and Lerche, 1988; Cloetingh, 1991; Miall, 1991). Given its limitations, the comparison (Fig. 16) shows that:

1. The brackets containing the ages of the mid-Cretaceous unconformity (substratum/SS 1), the boundaries of SS 2/3 (latest Maastrichtian to Danian), 3/4 (middle to late Ypresian) and 4/5 (Bartonian), each contain the age of a supersequence boundary as proposed by Haq et al. (1988).
2. The age of the boundary of SS 1/2 (Late Campanian) is the same as that of a sea-level lowstand ranked between the 2nd and 3rd orders by Haq et al. (1988).
3. The basal boundary of SS 6 falls within the long-lasting period of eustatic sea-level lowstand during the Chattian stage on the Haq et al. (1988) curve. Its age is possibly only slightly younger than the deepest Phanerozoic sea-level lowstand at the Rupelian/ Chattian boundary.
4. Four supersequence boundaries proposed by Haq et al. (1988) to have occurred in Late Cretaceous to Miocene time could not be recognized on the Maiella platform margin.

Fig. 15.—The aggradation curve of the platform (full line) shows that the length of aggradational phases generally decreased with time from Late Jurassic to Miocene time. In Late Jurassic and Cretaceous time, platform aggradation was not interrupted for biostratigraphically resolveable periods except during the "middle" Cretaceous tectonic event associated with platform emersion. The length of the intervals of nonaggradation and/or erosion (horizontal segments) and of minimal sedimentation increases from the Late Jurassic to the Tertiary time. A smoothed and averaged curve fitted to the aggradation curve (dashed line) is a crude approximation of the total subsidence curve; it shows an exponential decay in Late Jurassic and Cretaceous time. In Paleogene and in Miocene time, subsidence was no longer governed primarily by the thermal subsidence of the passive continental margin, but was increasingly influenced by intraplate stresses and by the evolution of the foredeep of the Dinaric and Apenninic orogens. KL = Upper Jurassic and Lower Cretaceous substratum, SS = Supersequences, minimal duration of major hiatuses along the SS boundaries are ruled.
FIG. 16.—Times of platform aggradation in the Maiella platform margin in comparison with the “eustatic” curve proposed by Haq et al. (1988). Thick lines indicate 2nd-order supersequence boundaries; their ages may lie anywhere inside the intervals represented by brackets on the right side. The minimal durations of hiatuses along supersequence boundaries are ruled. Boundaries of 3rd-order sequences, not detailed in this text, are shown by small horizontal lines. The age brackets of some of the Maiella supersequence boundaries in-
clude a sequence boundary of Haq et al. (1988).

5. In the Maiella, 3rd-order sequences can only be recognized in part of the succession, and age resolution is not refined enough for their correlation with the onlap curve of Haq et al. (1988).

The ages of several of the supersequence boundaries estab-
lished on the Maiella platform margin correlate relatively well with ages proposed by Haq et al. (1988) for their global super-
sequence boundaries. Assuming that the curve of Haq et al. (1988) documents the eustatic signal, this could suggest an eus-
tatic cause of the boundaries. In this case, tectonic overprint could not eliminate the 2nd-order signal. Caution has to be taken when comparing the ages of Haq et al. (1988) with our data. Many of our ages are derived from benthonic foraminiferal stratigraphy, and there is an uncertainty of correlation with the planktonic foraminiferal zonation. Haq et al. (1988) estimated their uncertainty of numeric dating at ± 1.4 my in Early Tertiary and ± 0.6 my in Miocene time, but no estimates are given for the individual boundaries. Considering the resolution of our biostratigraphic data set, resulting in long age brackets and shorter but still significant brackets on the dates on the curve of Haq et al. (1988), a worst case scenario could be drawn in which none of the boundaries would match. Nevertheless, the maximally possible deviations between the ages in the Maiella and on the “global” chart are significantly smaller than the length of the supersequences, suggesting that the superse-
quencies were largely deposited during the 2nd-order sea-level highstands and that their boundaries formed during 2nd-order sea-level lowstands. However, we expect that in the Maiella platform margin the global signals are offset in time by the effects of regional tectonics.

Climate and Evolution of Fauna and Flora

Changes in the biological associations greatly influenced sed-
imentation on carbonate platforms (James, 1983). The biotic changes on the Maiella platform margin are observed across supersequence boundaries, probably because they represent longer periods of time, whereas no equally important ecological changes appear within the supersequences.

Plate kinematic reconstruction show that during Cretaceous time, the peri-Adriatic carbonate platforms were located in low latitudes, 10° to 30° N (Scotese et al., 1989). A tropical to sub-
tropical climate is also confirmed by the abundance of rudists associated with hermatypic corals in the Lower Cretaceous sub-
stratum, SS 1 and 2 (Accordi et al., 1987). Clay mineral assem-
blages, dominated by smectites (Accarie and Deconinck, 1989), suggest a subtropical, warm and only seasonally humid climate. Humid intervals are indicated by extensive bauxite horizons, particularly in “middle” Cretaceous time (Crescenti, 1969; D’Argenio, 1969; D’Argenio and Mindszenti, 1992). On the Maiella, the last rudists occur in beds deposited during the latest Maastrichtian (late G. gansseri or early A. mayaroensis zone, i.e., approximately 1 to 2 my before the Cretaceous/Tertiary boundary; Fig. 14). The disappearance of rudists coincides with a major sea-level lowstand documented by the SS 2/3 boundary. This age of the extinction is compatible with Kauffman’s (1984) conclusion that the extinction of low-latitude rudists occurred 1 to 2 my before the Cretaceous/Tertiary boundary. Kauffman (1984) speculates that sea-level lowering and drawback of the sea from the platforms and shelves was one of the main factors
for the extinction of the rudistids, whereas Stanley (1984) argues that a massive temperature drop around the time of the Cretaceous/Tertiary boundary would have been of greater importance.

The existence of coralal reefs in the Paleocene (SS 3), in upper Priabonian to lower Rupelian strata (SS 5) and again in upper Rupelian to lower Chattian (unit of uncertain supersequence allocation), indicates that the platform remained within the subtropical climate zone, for much of the Early Tertiary period, at least, until “middle” Oligocene time. The extensive cooling event around the time of the Eocene/Oligocene boundary (Öberhänsli and Hsu, 1986; Frakes, 1986) is possibly reflected by changes in the benthonic foraminiferal associations of the Maiella platform margin (Pignatti, 1990).

An abrupt change is observed in uppermost Oligocene to Miocene biotic associations right above the basal boundary of SS 6, where the fauna and flora are strikingly different from those of the “middle” Oligocene Epoch. Bryozoans and red algae were abundant, whereas no corals were found. This biotic change indicates that the Maiella had probably entered the temperate climate zone, where it remained throughout Miocene time (cf. Carannante et al., 1988). The benthonic foraminifers did not suffer greatly from this climatic deterioration; they show largely continuous evolutionary trends across the boundary of SS 5/6.

We assume four different causes for the biotic changes on the Maiella platform margin: (1) mass extinctions of platform organisms (e.g., around the time of the Cretaceous/Tertiary boundary), (2) the slow drift of the Adriatic microplate to cooler climate zones (VandenBerg et al., 1978) with different ecological associations, (3) the latitudinal shift of the climate zones (e.g., by the Miocene cooling of the northern hemisphere; Vincent and Berger, 1985), and (4) variations in the distribution of seaways (e.g., the closure of Tethyan seaways; Ricou et al., 1986), which influenced the biota through regional climatic changes and geographical isolation. However, such changes are more difficult to document on the Maiella platform margin.

CONCLUSIONS

From Late Cretaceous to Miocene time, the Maiella carbonate platform margin evolved from an aggrading platform separated from an adjacent deep basin by a steep escarpment (Late Cretaceous to Late Campanian, SS 1) to a distally steepened slope (Late Campanian to “middle” Oligocene, SS 2 to 5) and finally to a gently inclined shelf (Miocene, SS 6; Fig. 14). The Upper Cretaceous to Miocene carbonate sediments of the Maiella platform margin are divided into six supersequences, separated by deeply incised truncation surfaces that are interpreted to have formed during major sea-level lowstands.

Each supersequence is characterized by a distinct depositional system that drastically changed across most supersequence boundaries. The evolution of the Maiella platform margin was interrupted and the platform partially eroded during the relatively long periods of important 2nd-order sea-level lowstands. Platform growth resumed when the platform was again flooded, with the depositional system adapted to the new controls exerted by changed topography, the constraints of higher-order sea-level fluctuations, climate and the evolving fauna and flora. The combined effect of all these changes facilitates the recognition of the supersequences in the field.

Subaerial and submarine erosion contributed to the formation of the supersequence boundaries, although the products of subaerial processes are rarely preserved and might have been largely eroded during the same sea-level lowstands or the subsequent flooding of the platform. The hiatus along each supersequence boundary increases in duration from the basin towards the platform due to more intense proximal exposure and erosion. The duration of Cretaceous to the Miocene hiatuses along the supersequence boundaries generally increased in response to the decreasing total subsidence rates. Thus, the criteria postulated for recognizing sequence boundaries in seismic sections by Van Wagoner et al. (1988) and Vail (1988) could be applied also in good outcrop sections.

Total subsidence of the platform, approximated by platform aggradation, decreased more or less exponentially with time, probably from the rifting period, but certainly from the Late Jurassic to the beginning of the Tertiary. However, in detail, the subsidence pattern was determined by phases of uplift, probably related to intraplate stresses. In Oligocene and Miocene time, uplift and subsidence resulted from loading of the lithosphere by the thrust nappes of the advancing Dinaric and Apenninic orogen. Tectonic uplift enhanced the expression of sea-level lowerings and is responsible for extensive erosion along the supersequence boundaries. Loading of the lithosphere and subsidence of the foredeep helped to preserve the Upper Tertiary carbonate shelf.

Sedimentation rates began to exceed the total subsidence rates significantly during Late Cretaceous time, which allowed the basin in front of the platform to be filled by the Late Campanian. As a consequence, the slope prograded over the basinal carbonates from Late Campanian time on, and the shallow-water platform prograded over the slope in the late Priabonian to early Rupelian.

Assuming our chronostratigraphic correlation is correct, the ages of 2nd-order sea-level lowstands recognized in the Maiella are within the range of the ages of 2nd-order eustatic sea-level lowstands proposed by Haq et al. (1988). Therefore, they could suggest synchrony of the formation of these supersequence boundaries with those proposed by Haq et al. (1988), implying dominant eustatic control of sea level on the long-term evolution of the Maiella platform margin despite tectonic overprint. However, considering the age brackets in our data as well as the uncertainties in Haq et al.’s (1988) curve, the supersequence boundaries can all be older or younger than the ages given by these authors. The mid-Cretaceous supersequence boundary is an exception in that uplift and possibly also slight tectonic tilting of the platform were important causes of its formation.

Climate was an important controlling factor during the evolution of the peri-Adriatic carbonate platforms, in that its low latitude position during most of the period from the Cretaceous to the “middle” Oligocene allowed high rates of carbonate production. This evolution was severely interrupted during the latest Maastrichtian, when the rudistids disappeared from the Maiella platform margin during the time of a major sea-level fall. In the Miocene Epoch temperate climatic conditions were established. Corals and other fast-growing organisms are missing in Miocene strata, and accumulation rates remained relatively low. These faunal and floral changes did not influence the formation of supersequence boundaries, but strongly influenced the internal architecture of the sequences.
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OXYGEN ISOTOPE SYNTHESIS: A CRETACEOUS ICE-HOUSE?

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INTRODUCTION

Carbonate precipitated organically or inorganically in the ocean records the sea-water isotopic composition. The δ18O composition of marine calcite is dependent on diagenesis, water temperature, salinity and ice volume. Increased ice-volumes during glacial periods correspond to heavy or more positive δ18O values and decreased ice-volumes during inter-glacial periods correspond to light or more negative δ18O values. By observing downcore variations in the δ18O of diagenetically unaltered calcite, variations in ice volume can be inferred. This is only a rough estimate because a significant component of the δ18O record is expected to have been caused by global cooling and decreasing ocean temperatures through Cenozoic time (Savin, 1977). Therefore, oxygen isotope records are among the most widely used proxy indicators of glaciation and sea-level fluctuation. A number of records were generated from Deep Sea Drilling Project (DSDP)/Ocean Drilling Program (ODP) sites (e.g., Shackleton and Kennett, 1975; Matthews and Poore, 1980; Miller et al., 1987, 1991a; Prentice and Matthews, 1988). Abreu and Haddad (this volume) and Abreu and Anderson (in press) compiled deep-water oxygen isotope data sets and produced a composite smoothed isotope record for the Cenozoic to compare to eustatic curves derived from sequence stratigraphy.

Matthews and Poore (1980) suggested ice build-up in Antarctica during the Cretaceous based on a generalized oxygen isotope record. They assumed constant tropical sea surface temperatures since the Cretaceous to evaluate ice-volume changes. Their approach implied significant ice volumes at least since the late Eocene and possibly for much of the Cretaceous. However, geological evidence for widespread glaciation on the continent at that time is very limited, since no lower Cretaceous exposures exist on Antarctica and upper Cretaceous strata are restricted to the Antarctic Peninsula. Thus, the continent’s Cretaceous glacial and climatic setting is essentially unknown (e.g., Abreu and Anderson, in press).

Several studies indicate warm climatic conditions, generally ice free, during Cretaceous time based on different proxies: terrestrial plants (e.g., Parrish and Spicer, 1988), marine fossils (e.g., Gordon, 1973), increase in volcanism (e.g., Larson, 1991), increase in atmospheric CO2 (e.g., Arthur and Dean, 1986), paleogeographic reconstructions (e.g., Barron, 1983), black shales (e.g., Arthur and Schlanger, 1979) and stable isotopes (e.g., Douglas and Savin, 1975; Huber et al., 1995). It is generally accepted that the climate remained warm and equable throughout the Cenomanian to Campanian stages, coinciding with high sea level (Hays and Pittman, 1973; Haq et al., 1987). However, some recent stable isotope studies suggest relatively cooler climatic conditions in high latitudes during the lower Cretaceous (Sellwood et al., 1994; Price et al., 1996; Stoll and Schrag, 1996) and Maastrichtian (Barrera, 1994; Huber et al., 1995).

DATA SETS AND METHODS

The Cretaceous isotope record (Fig. 1) is based on benthonic foraminifera from DSDP/ODP sites and on bulk rock from outcrops. The benthonic isotope record indicates changes in deep water δ18O and the bulk rock δ18O record represents an average of calcareous microfossils (nannofossils and benthonic and planktonic foraminifera) and fine-grained calcite. The upper Aptian and Albian record is based on the benthonic foraminifera Gavelinella spp. (upper bathyal paleowater depth) from DSDP sites 392 and 511 (Fassell and Bralower, in press). The Cenomanian to lower Campanian record is based on mixed benthonic foraminifera from DSDP Site 511 (Huber et al., 1995). The Campanian and Maastrichtian record is based on Gavelinella beccariformis and Gavelinella spp. from ODP Site 690 (Barrera, in press), respectively. The isotope records for the sites 690 and 511 were kept separate in Figures 1 and 2 because the isotopes were derived from disparate foraminifera assemblages.

The upper Albian to lower Campanian bulk rock record (Jennkyns et al., 1994) is from a smoothed oxygen isotope record (7 points least square method) from England (English chalk) and Italy (Gubbio). The upper Campanian to Maastrichtian record...
VITOR S. ABREU, JAN HARDENBOL, GEOFFREY A. HADDAD, GERALD R. BAUM, ANDRE W. DROXLER, AND PETER R. VAIL

CRETACEOUS OXYGEN ISOTOPES

Smoothed Oxygen isotope curve based on bulk rock from outcrops in England, Italy, and Tunisia.

Oxygen isotope curve based on benthic foraminifera from DSDP/ODP sites.

<table>
<thead>
<tr>
<th>TIME (Ma)</th>
<th>CHRONO-STRATIGRAPHY</th>
</tr>
</thead>
<tbody>
<tr>
<td>-4</td>
<td>MAST.</td>
</tr>
<tr>
<td>-3</td>
<td>CAMPANIAN</td>
</tr>
<tr>
<td>-2</td>
<td>T.T.R.</td>
</tr>
<tr>
<td>-1</td>
<td>CENOM.</td>
</tr>
<tr>
<td>0</td>
<td>M.B. B.</td>
</tr>
<tr>
<td>1</td>
<td>APTIAN</td>
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<thead>
<tr>
<th>CRETACEOUS OXYGEN ISOTOPES</th>
<th>EUSTATIC CURVE AND SUPERCYCLES SETS</th>
</tr>
</thead>
<tbody>
<tr>
<td>DSDP Site 392 - Gavelinella spp.</td>
<td>CYCLE Pi</td>
</tr>
<tr>
<td>DSDP Site 511 - mixed benthics</td>
<td>CYCLE Ni</td>
</tr>
<tr>
<td>DSDP Site 511 - Gavelinella spp.</td>
<td>CYCLE UK</td>
</tr>
<tr>
<td>ODP Site 690 - Gavelinella beccariformis</td>
<td>CYCLE UK</td>
</tr>
</tbody>
</table>

**LEGEND**

-2 -1 0 1-2 -1-3-4

18O (‰)

**TIME (Ma)**

95 90 80 75 70 85 105 100 110 115

**CRETACEOUS CENOZOIC**

APTIAN ALBIAN CENOM. TUR. CON. S. CAMPANIAN MAAST. PALEOCENE EOCENE OLIGOCENE MIOCENE PLIO.

**CHRONO-STRATIGRAPHY**

**OXYGEN ISOTOPES**

δ18O (‰)

**EUSTATIC CURVE AND SUPERCYCLES SETS**

200 (meters)

**Fig. 1.—** Correlation between the oxygen isotope record from outcrops and from DSDP/ODP sites for the Cretaceous. The outcrop record is based on bulk rock and the deep water record is based on benthonic foraminifera.

is based on powder bulk rock from central Tunisia (Abreu et al., in prep.).

We used the composite smoothed isotope record of Abreu and Haddad (this volume) and Abreu and Anderson (in press) for the Cenozoic (Fig. 2). They defined positive isotope events in the deep-sea benthonic record through the identification of the event in a reference site with a good sampling rate and fairly complete sedimentary section and its correlation with other sites. The chronostratigraphic position of the isotope events was defined through first-order correlation with magnetostratigraphy and in some cases biostratigraphy. The chronostratigraphic position, oxygen isotope value and amplitude of each event was defined in sites with well represented stratigraphic sections. Positive events present in only one site but absent in other sites were not considered for the composite record. Abreu and Haddad (this volume) and Abreu and Anderson (in press) attempted to use, as far as possible, sites with isotope records based on the benthonic foraminifera *Cibicidoides*, situated in mid-latitudes and intermediate paleowater depths (between 500 and 2,000 m). The value and estimated amplitude of each positive δ18O event interpreted from DSDP/ODP sites was plotted

**Fig. 2.—** Correlation between the oxygen isotope record from the upper Aptian to the present and the eustatic curve of Haq et al. (1987).
through time and is presented in graphic form (Fig. 2). Abreu and Anderson (in press) proposed 7 middle and late Eocene positive isotope events, based mostly on sites 689 and 690. These sites are located in high latitudes and present the best resolution for the benthonic foraminifera record in the Eocene. The isotope values showed in Site 690 are similar to those at the equatorial Site 865 (Bralower et al., 1995). The standard isotope records from sites 522 (Miller et al., 1988) and 608 (Miller et al., 1991b) are the basis for the Oligocene and Miocene records respectively. The isotope events defined by Miller et al. (1991a) and Abreu and Haddad (this volume) were used for the Oligocene and Miocene records. The Pliocene-Pleistocene isotope records, based on sites 502 (Oppo et al., 1995) and 704 (Hodell and Venz, 1992), were smoothed to keep the 100-ky and longer cycles. The filtered isotope record of Haddad and Vail (1992) was used for the Pleistocene to the Recent sections, which is a filtered version (low-pass, 1/66 ky – 1/45 ky) of the stacked benthonic isotope records from sites 607 and 677 (Raymo et al., 1990). The procedure established by Abreu and Haddad (this volume) was followed to convert the isotope record of each site and to the time scale of Berggren et al. (1995) and Haddad (this volume) was followed to convert the isotope record of each site and to the time scale of Berggren et al. (1995) for the Cenozoic and Gradstein et al. (1994) for the Cretaceous.

**Isotope Events and Trends**

The Cretaceous composite DSDP/ODP isotope record shows a positive interval characterized by about 0.5% near the Aptian/Albian boundary (Fig. 1). A trend towards lighter values began in the lower Albian and continued to the lower Turonian. From upper Turonian to Maastrichtian, the isotope record shows a gradual trend towards heavier values, reaching about 0.5% during the uppermost Maastrichtian.

The Cretaceous bulk rock isotope record shows a period of light δ¹⁸O values during the Cenomanian and Turonian stages (about ~3%), with negative events near the Cenomanian/Turonian boundary and in the upper Turonian (Fig. 2). The data continues to show light values during the Coniacian and Santonian stages (about ~2.5%), with a shift towards heavier values in the upper Santonian. The Tunisia record shows positive events during the upper Campanian stage, near the lower/upper Maastrichtian boundary, two events during the upper Maastrichtian and an event near the Cretaceous/Tertiary boundary. Two significant light isotope events occur during the upper Campanian.

There is a trend during Paleocene time towards lighter isotope values that persists during lower Eocene. The most negative oxygen isotope values for the entire Cenozoic column (about ~0.5%) occurred during the lower Eocene. A pronounced positive shift of the oxygen isotopes occurs in the lower Lutetian Stage (base on the middle Eocene). The isotope values in the upper Lutetian Stage reached 1%. Another significant positive shift of the isotopes occurs during the Bartonian Stage. The isotope values in the upper Eocene reach 2% with high-amplitude (of about 1%) fluctuations. The overall Eocene isotope record suggests cooling of bottom waters and/or an increase in ice volume towards the upper Eocene. The top of the Rupelian and the base of the Chattian stages are marked by the heaviest isotope values in Cenozoic time. The lighter lower Miocene oxygen isotope values indicate a period of relatively warmer bottom waters and/or smaller ice volumes. The pronounced positive shift during the middle Miocene is related to a period of major ice build-up in Antarctica (Shackleton and Kennett, 1975; Savin et al., 1975). From the upper Pliocene to the Present, the isotope record shows a continuous trend towards heavier values with major high-frequency fluctuations.

**Low-Frequency Cycles and Glacioeustasy**

The δ¹⁸O record reveals three long-term cycles defined by positive-maximum excursions from the Cretaceous to Cenozoic (Fig. 2). We propose the names: Uki (upper Cretaceous isotope cycle), Pi (Paleogene isotope cycle) and Ni (Neogene isotope cycle). The oldest cycle (Uki) extends from the Aptian/Albian boundary to the uppermost part of the Maastrichtian, with the lightest values near the Cenomanian/Turonian boundary. The middle cycle (Pi) spans the uppermost Maastrichtian and most of the Paleogene, until near the base of the Chattian (base of the upper Oligocene). The lightest middle values in the second cycle occur in the lower Eocene. The youngest cycle (Ni) begins near the base of the upper Oligocene, reaching the most negative values at the uppermost portion of the lower Miocene. After the lower Miocene, the δ¹⁸O values increase in steps until the upper Pleistocene. These three low-frequency cycles (Fig. 2) show a good correlation to the Haq et al. (1987) second-order supercycle sets Upper Zuni A (UZA), Tejas B (TB) and Tejas A (TA). There is also a strong correlation between the long-term sea-level curve (Haq et al., 1987) and the low-frequency δ¹⁸O positive-negative cycles, except for the Pi cycle, which shows a high-amplitude negative shift during the lower Eocene period which does not correspond to a significant sea-level rise in the sequence stratigraphic record (Haq et al., 1987).

Correlation between δ¹⁸O events and sequence boundaries has been the subject of several articles (i.e., Miller et al., 1987, 1991a; Williams, 1988; Haddad and Vail, 1992; Wright and Miller, 1992; Abreu and Savini, 1994; Baum et al., 1994; Pekar and Miller, 1996; Browning et al., 1996; Abreu and Haddad, this volume). Abreu and Anderson (in press) show a strong correlation between positive shifts in the oxygen isotopes from the middle Eocene to the Present and the third-order sequence stratigraphic record of Haq et al. (1987) and Hardenbol et al. (this volume). Abreu and Anderson (in press) integrated the deep-sea record for glaciation (stable isotopes, ice-rafted debris, deep-sea hiatus) in the Southern Ocean with the terrestrial and continental margin sedimentary records and presented additional evidence for glaciation in Antarctica since the middle Eocene. The most compelling evidence for glaciation in East Antarctica during the Eocene comes from ODP Leg 119 drill sites Prydz Bay. Site 742 in Prydz Bay recovered middle/upper Eocene massive diamictons, interpreted as water-lain till (Hambrey et al., 1991; Barron et al., 1991). The occurrence of till corresponds to the first significant occurrence of ice-rafter detritus at Leg 119 Site 738 on the Kerguelen Plateau (Ehrmann, 1991). On the Wilkes Land continental margin (East Antarctica), the oldest major unconformity on the shelf, interpreted to be a glacial erosion surface, is inferred to be middle/late Eocene in age, based on extrapolation to DSDP Site 269 on the adjacent continental rise (Eittreim et al., 1995). Glacial-marine sediments with mid-Eocene dinoflagellates (Hannah, 1994) were also recovered at CIROS-1 in western Ross Sea. However, there
is no evidence that the West Antarctica Ice Sheet advanced across the continental shelf during Eocene time (Cooper et al., 1991). Middle Eocene glacial deposits occur in small outcrops on King George Island (South Shetland Islands), recording an episode of glaciation in the northern Antarctic Peninsula region (Krakow Glaciation, Birkenmajer, 1991).

The deep-water benthonic isotope record (Fig. 2) for the upper Aptian/Albian and upper Campanian/Maastrichtian shows $\delta^{18}O$ values (of about 0.5%) similar to those in the middle Eocene, which suggests perhaps some continental ice in Antarctica may have been present as early as the Early Cretaceous.

Figure 3 shows a comparison between the Upper Cretaceous and Cenozoic sequence stratigraphic terms eustatic cycles of Vail (Haq et al., 1987) for the upper Cretaceous and Cenozoic, the short-term sequence stratigraphic (Hardenbol et al., this volume) and isotopic (Abreu and Haddad, this volume, and Abreu and Anderson, in press) cycles, and the long-term isotopic cycles (this work). All data is presented in the time scale of Berggren et al. (1995). The long-term isotope record (Fig. 4) was built using an envelope of the lightest values in the smoothed composite oxygen isotope record (Abreu and Haddad, this volume, Abreu and Anderson, in press, and this work). Tentatively, the isotope value of 3.5% was used to adjust the isotope record horizontal scale to the zero meter point of the eustatic curve horizontal scale, representing modern value for oxygen isotopes (benthonic foraminifera) in deep-water settings (e.g., Dwyer et al., 1995). The horizontal scale for the isotope record and the eustatic curve were calibrated using the Pleistocene calibration of 0.11% $\delta^{18}O$ variation per 10 m of sea-level change determined by Fairbanks and Matthews (1978) and Fairbanks (1989). In fact, the Fairbanks calibration is based on the comparison between sea-level and $\delta^{18}O$ changes from the Last Glacial Maximum (ca. 18 ky BP) to present. The calibration yields reasonable sea-level variations for individual glacial-interglacial cycles. The isotopic variation observed from the Cenomanian to present (almost 5%) is probably strongly influenced by cooling of the global oceans (e.g., Savin, 1977). In general, the long-term eustatic curve (Haq et al., 1987) and the low-frequency isotope cycles show a similar trend, with high sea-level during the Cenomanian/Turonian, early Eocene and early Miocene coinciding with light oxygen isotope values. However, there are some significant differences. For example, the high sea-level indicated by the eustatic curve during the Rupelian stage is not confirmed by the isotope record. There is also disagreement concerning a sea-level high during Campanian times showed by the eustatic curve (Haq et al., 1987). In figure 1, the low-resolution deep-sea record based on benthonic foraminifera shows a continuous trend to lighter values from the Maastrichtian to the Cenomanian, although the higher resolution bulk-rock isotope record suggests a shift towards lighter values in the Campanian.

CONCLUSIONS

There is a positive correlation between the stable isotope record presented here for the Cretaceous and Cenozoic (Fig. 2) and Vail’s sea-level curve based on the coastal onlap record (Haq et al., 1987). However, the most negative Cenozoic values (cycle Pi-lower Eocene) show a weak correspondence with significant high sea-level in the Haq et al. (1987) chart. Acceptance of continental ice during the Cretaceous and Paleogene is still highly controversial among stratigraphers. Recent publications have addressed the controversy and a consensus has begun to emerge for the existence of continental ice as early as the Eo-

![Diagram](chart.png)
The composite oxygen isotope record presented here seems to indicate the possibility that continental ice may have existed as early as the Aptian. Data supporting the current consensus in the scientific community on the absence of continental ice in the Cretaceous are not conclusive and may need to be challenged.

ACKNOWLEDGMENTS

The authors would like to express their gratitude to Petrobras and Rice University for their support. Thanks to Dr. Albert Bally for reviewing the manuscript and to Gabor Vakarcs for discussions. Thanks also to Dr. Enriqueta Barrera whose review greatly improved the original work. We are also grateful to Emoke Vakarcs for the preparation of the Tunisia samples for isotope analyses.

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**Fig. 4.**—Long- and short-term eustatic curves (Haq et al., 1987) compared with the long-term isotope cycles (this work) and short-term sequence stratigraphic (Hardenbol et al., this volume) and isotopic (Abreu and Haddad, this volume, and Abreu and Anderson, in press) cycles.


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PART II
CENOZOIC ERA
Neogene deposits discussed in the six papers submitted for this volume are situated in the North Sea basin, Pannonian Basin, Piedmont Basin and the Central Mediterranean Basin (Sicily). Contributions for the North Sea Basin discuss the Paleogene and Neogene deposits of the eastern subsurface offshore Denmark and its southern rim in Belgium. Studies of the Pannonian Basin (Hungary) and Piedmont Basin (Italy) address Oligocene through middle Miocene deposits. Two papers on the Central Mediterranean Basin in Sicily describe the Plio-Pleistocene record. Neogene Alpine compressive tectonics led to a general uplift of Europe except for the present offshore North Sea basin. Neogene deposits in northwestern Europe are generally thin, stratigraphically incomplete and often marginal to non-marine. In the Alpine/Carpathian realm, localized tectonic events formed isolated basins such as the Piedmont and Pannonian Basins. The Mediterranean area was undergoing major tectonic rearrangements in the Neogene.

Stratigraphic calibration of proposed sequence records to the integrated Berggren et al. (1995) time scale could only be achieved with confidence for lower to middle Miocene and Plio-Pleistocene deposits laid down under open-marine conditions in mid latitude settings in the Pannonian, Piedmont and Mediterranean Basins. The incomplete record for the upper Miocene could not be reliably calibrated to the Berggren et al. (1995) time scale and the Haq et al. (1987) record for the upper Miocene is retained on the Cenozoic chronostratigraphic chart.

**Miocene Sequences**

The stratigraphic position of Neogene sequence boundaries Ch 4/Aq 1 to Ser 3 on the new chart are based on sequences identified from reflection seismic records in the Pannonian Basin calibrated to Vakarcs et al. (this volume). Vakarcs et al. (this volume) identify nine sequences in the lower and middle Miocene of the Pannonian Basin; two more (in the Burdigalian) than were identified by Haq et al. (1987). Gnaccolini et al. (this volume) describe seven sequences (B5 to C6) in the lower and middle Miocene of the Piedmont Basin in Italy. Gnaccolini et al. (this volume) do not identify sequences Aq 2, Bur 2 and Bur 3. The stratigraphic position of the lower sequence boundaries of Gnaccolini et al. that can be calibrated to biozones agree well with the findings of Vakarcs et al. (this volume). There is also good agreement with Haq et al. (1987) considering the revised biozonal interpretation. The identified sequences correspond well with the oxygen isotope events of Abbreu and Haddad (this volume) and thus support a glacio-eustatic mechanism for the Neogene sequences.

Analysis of sequences in the eastern North Sea Basin by Michelsen et al. (this volume) suggest six sequences (5.2 to 6.3) in the Aquitanian to Burdigalian interval. The observed frequency of sequences appears to agree with the findings of Vakarcs et al. (this volume) although available biostratigraphic data from the subsurface do not permit conclusive stratigraphic calibration. Along the southern border of the North Sea Basin (Belgium), Miocene deposits are poorly developed (Vandenberghhe et al., this volume). Of the three sequences that are developed, the youngest can be calibrated with sequence 6.2 of the eastern North Sea, C1 of the Piedmont Basin and Bur 4 of the Pannonian Basin.

In a comprehensive review of northwest European Tertiary basins (Vinken, 1988), three major Neogene regressive transgressive cycles are described. The second cycle (cycle 8, sensu Vinken, 1988) in the middle and upper Miocene of the North Sea, begins at the base of the Nordland Group of the Norwegian-Danish Basin and the south Viking Graben which corresponds to the “Reinbek marker” in Germany. This cycle, characterized by a prominent seismic and gamma-ray marker, represents a major surface caused by a sudden and significant rise in sea level. Mollusc and otolith biostratigraphy (Vinken, 1988) calibrate this rise tentatively to the base of calcareous nannofossil zone NN5 which coincides with the major flooding at the base of the Langhian situated between sequence boundaries Bur 5/Lan 1 and Lan 2/Ser 1. The description by Michelsen et al. (this volume) of the sequence boundary between units 6 and 7 in the eastern North Sea appears to correspond to the base of cycle 8 of Vinken (1988).

**Pliocene and Pleistocene Sequences**

The study by Catalano et al. (this volume) on the central Mediterranean Sea near Sicily, integrates outcrop and seismic sections with Mediterranean ODP and Atlantic DSDP records. Close to thirty bioevents are used to correlate Pliocene and Pleistocene sequences and calibrate the stratigraphy to the as-
Mesozoic - Cenozoic Sequence Chronostratigraphy of European Basins

HARDENBOL, J., THIERRY, J., FARLEY, M.B., JACQUIN, T., DE GRACIANSKY, P.C., & VAIL, P.R.

NEOGENE SEQUENCE CHRONOSTRATIGRAPHY

HARDENBOL, J., THIERRY, J., FARLEY, M.B., JACQUIN, T., DE GRACIANSKY, P.C., & VAIL, P.R.

This work calibrated to the bio-chronostratigraphic framework of Berggren et al. (1995). Sequences are compared to Haq et al. (1987), 1987, 1988.
The new data from tectonic basins in Europe have not changed the Haq et al. (1987) sequence interpretation significantly, which suggests that the sequences are not related to the specific tectonic history of the basins but rather represent variation in sea level.

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INTRODUCTION TO THE PALEogene

Jack E. Neal and Jan Hardenbol

The stratigraphic position of sequences on the new Cenozoic chronostratigraphic chart is based on information from papers submitted for publication in this volume. Much of this data was presented at poster sessions of the Sequence Stratigraphy of European Basins Meeting in Dijon, France in 1992. Of the eleven papers addressing Cenozoic sequence stratigraphy in this volume, four papers include sequence information for both Neogene and Paleogene, three papers address sequence stratigraphy in the Neogene and four papers deal with the Paleogene.

Paleogene sedimentary basins described in the papers submitted for this volume occur in compressional, extensional, and passive margin settings and are filled with sediments deposited in paleoenvironments that varied from non-marine to bathyal depths. Despite the variability, deposits of this age can be easily correlated due to an integrated biochronostratigraphic calibration that encompasses every type of available stratigraphic marker.

In general, studies on Paleogene sequence stratigraphy fell into two major geographical areas: the greater North Sea basin, and the greater Mediterranean basin. Much of the detail in the new European Basins Cycle Chart comes from the greater North Sea basin. This province has been intensely studied for biostratigraphy, lithostratigraphy and sequence stratigraphy for much of the last 25 years. The International Geologic Correlation Program (IGCP) #124 (Vinken et al., 1988) gathered stratigraphic information from many different sources to erect a stratigraphic calibration that contributed to the present studies. Subsequent work has refined that framework even more and now combines physical (sequence) stratigraphy with the more traditional biostratigraphic and lithostratigraphic approach. The publication of “A Revised Cenozoic Geochronology and Chronostratigraphy” (Berggren et al., 1995) provided a state of the art temporal framework.

The greater North Sea basin was essentially a passive margin intra-cratonic seaway with estuarine branches across northwest Europe connected to the mostly shelfal to deep sea fan deposition of Denmark and the central North Sea. Here, a high-frequency (eustatic?) changing sea-level signal should be most easily correlated. Studies in the greater Mediterranean basin come from much more tectonically active basins with less biostratigraphic detail. We have therefore chosen to rely on northwest Europe for our standard, but calibrate the southern European studies for comparison.

Final positioning of most Paleogene sequences on the cycle chart comes from correlation between four key areas: the central North Sea, Denmark, Belgium and the London-Hampshire basin of southeastern England. These areas were chosen for their high quality data sets, long history of work and the fact that together they record complete sequences of a shelf to basin transition. The central North Sea is covered by Neal et al. (this volume), using a seismic, well log and biostratigraphy data set to demonstrate a hierarchy of stratigraphic cycles. Graphic correlation biostratigraphy (Stein et al., 1995) provided a framework to integrate several biostratigraphic disciplines and to identify times of deposition and sediment starvation. This methodology shows that the central North Sea depocenter records mainly lowstand deposition, which ties to unconformities in northwest European outcrop sections. Conversely, highstand deposits in northwestern Europe correlate with periods of sediment starvation in the central North Sea.

Lowstand-dominated sequences of the central North Sea are correlated to the biostratigraphic framework of onshore basins through deposits in Denmark. Michelsen et al. (this volume) present a sequence stratigraphic framework for the entire Cenozoic section that is based on a rigorous biostratigraphic framework and an extensive seismic data grid with several wells. The Paleogene portion of this framework identifies few sequences, calibrated however, to a detailed biostratigraphy based on an extensive published record (e.g., Heilmann-Clausen, 1985; Heilmann-Clausen et al., 1985; Nielsen et al., 1986; Thomesen and Heilmann-Clausen, 1985). With this framework in place, we then looked at the shallow water sections of Belgium and southeastern England. Dinoflagellate (Eaton, 1976; Powell, 1992; Jolley, 1992; DeConinck, 1990; Heilmann-Clausen, 1994, 1985) and nanofossil (Aubry, 1983, 1986; Steurbaut and Nolf, 1986; Thomesen and Heilmann-Clausen, 1985) stratigraphy is used to calibrate sequences documented by Vandenberghe et al. (this volume) in Belgium and many workers in England (Plint, 1988; Knox, 1989; Knox et al., 1994; King, 1981, 1991). The resulting sequence framework is shown in Figure 1 compared to various lithostratigraphic units in northwestern European basins.

The new Paleogene sequence chart differs from the Haq et al. (1988) chart over the same interval. Ten additional Paleogene sequences are documented in Europe with recent studies, mainly in the Danian and upper Thanetian. This documentation is only possible by studying the most complete section for each age and creating a composite “standard” section. For example, new Danian-aged sequences (Da-1, Da-2 and Da-4) were documented by detailed study of the Danian limestone in Denmark (Thomesen, 1990). This section is hiatal in most of the rest of Europe. Also, many of the Thanetian sequence additions come from calibrating the UK record to Belgium and correlating new sequences in one location (e.g. Knox et al., 1994) that were previously unrecognized at the other (Neal, 1996). The correlation of these new events rely on the resolution of the data and may vary by author. Some differences exist, for example, between the most detailed frameworks to date in northwestern Europe. Depending on the author preference, the same lithologic unit may be lumped into a composite sequence, dismissed as a parasequence, or correlated as a high frequency sequence. An example of this preference is demonstrated in the Thanetian where Vandenberghe et al. (this volume) chose to consolidate units that Neal et al. (this volume) divide (Fig. 1). These same units are further divided by Knox (1996) into more high frequency sequences than Neal et al. (this volume) recognize. Each
Fig. 1.—Summary figure of Paleogene sequences in northwest Europe as documented in recent papers and works of this volume, compared to the bio-chronostratigraphic framework of Berggren et al. (1995) and sequences of Haq et al. (1988). See text for details of the construction of the new sequence summary.
paper recognizes essentially the same stratigraphy, but may interpret the same surface as either unimportant, parasequence bounding, or even full sequence bounding. Correlation and calibration require the exceptional available biostratigraphy from integrated biostratigraphic disciplines, and if complete stratigraphic descriptions are made, a dialog is possible. The summary figure of Paleogene sequences (Fig. 1) represents the resolution of sequences documented in the papers collected in this volume. Future refinements will likely come with more research.

Re-calibration and positioning of Oligocene sequences was based on the stratigraphic record of the Pannonian basin in Hungary (Vakarc et al., this volume). The position of these sequences is based on biostratigraphic criteria. A new Oligocene sequence in Hungary (Ru-2) can be calibrated with a sequence in offshore Denmark (sequence 4.2 of Michelsen et al., this volume). Abreu and Haddad, (this volume) show a good agreement between oxygen isotope event records from several DSDP and ODP sites and a well in the Campos basin offshore Brazil and the sequences identified by Vakarc et al., (this volume) as well as those of Haq et al., (1988)

Gnaccolini et al. (this volume) describes six sequence boundaries in the Oligocene of the Piedmont Basin in Italy. Two sequences in the lower Oligocene (A1 and A2) tentatively calibrated with planktonic foraminifera zones P20 and P21 may or may not be synchronous with Ru 2 and Ru 3. Four sequence boundaries in the upper Oligocene (B1 to B4) compare stratigraphically well with the new chart. Sequence boundary B1 calibrated to the upper part of P21 correlates well with the Ru 4/Ch 1 sequence boundary of the new chart (= 30 my sequence of Haq et al., 1988). Sequence boundary B2 a few meters below the P21/P22 compares stratigraphically very well with sequence Ch 2 of the new chart. Sequence boundary B3 falls within P22 as does sequence Ch 3 of the new chart. The final sequence boundary in the Oligocene B4, characterized by a “latest Oligocene Globigerina ciperoensis bloom” still in zone P22, compares reasonably well with sequence boundary Ch 4/Ag 1 of the new chart.

Paleogene sequence stratigraphic studies of southern Europe encounter difficulties not seen in their northern counterparts. The active margin nature of the Spanish basins (Alicante region, Basque region, and Pyrenean region) makes inter-regional correlation more difficult and biostratigraphic resolution is not as precise as in the northern European basins. Relying mainly on planktonic foraminifera for calibration, Pujalte et al. (this volume) and Geel et al. (this volume) present correlations to Paleogene sequences of the Haq et al., (1988) chart. In Figure 1 the sequences of Geel et al. (this volume) and Pujalte et al. (this volume) are re-calibrated to the new sequence chart. Pujalte et al. (this volume, 1995) can be correlated with some NW European sequences (ex. Th 4 = DS-T/Y), but others have no equivalence (ex. DS-P2). The same can be said about Geel et al. (this volume), where some Lutetian and early Eocene cycles correlate within biostratigraphic resolution, but others do not (ex. P15 sequence). Since these frameworks were correlated to the Haq et al. (1988) chart, a re-calibration with the revised biochronostratigraphy of Berggren et al. (1995) is required, to verify if the southern European sequence frameworks are synchronous with northern Europe.

The final Paleogene paper of this volume is by Luterbacher, who cautions the reader to be aware of the limits of biostratigraphic resolution. Using a previously published framework from the Pyrenees (Luterbacher et al., 1991), this paper discusses calibration needs down to a single taxa as it evolves and concludes that one should not put too much faith in “global” calibrations.

In conclusion, the Paleogene sequence stratigraphy presented in the papers of this volume represent the latest thinking in correlations today as far as the Paleocene, lower Eocene and Oligocene is concerned. This does not mean that tomorrow a new, more detailed framework will not emerge. Sequences in the middle and upper Eocene on the new chart are merely recalibrations of sequences from Haq et al. (1988) to the Berggren et al. (1995) stratigraphic framework. The current partial revision of the Haq et al. (1988) Paleogene portion of the Cenozoic chart shows 36 sequences, an increase of 10 sequences, which were documented in less than a decade! The bottom line is improved calibration and chronostratigraphic resolution will lead to a more detailed framework. The level of detail presented here can now serve as a test case for calibration within and outside of northern Europe.

REFERENCES


HEILMANN-CLAUSEN, C., 1985, Dinoflagellate stratigraphy of the Uppermost Danian to Ypresian in the Viborg 1 borehole, Central Jylland, Denmark: Danmarks Geologiske Undersøgelse, Series a, v. 7.


ABSTRACT: The Cenozoic evolution of the epicontinental North Sea Basin is described on the basis of sequence stratigraphy, comprising analyses of seismic sections, petrophysical logs and biostratigraphic studies of foraminifera, dinoflagellates and calcareous nanofossils. Stratigraphic, palaeogeographic and palaeoecological information from the Danish onshore area is integrated in the study.

The deposits are subdivided into 21 sequences, which group into seven informal major units. The sequence boundaries are identified by differences in seismic facies and by seismic onlap, toplap and truncation features. The maximum flooding surfaces are placed at internal downlap surfaces which correlate with high values on the gamma-ray logs.

The source of sediments and the direction of sediment transport changed several times during Cenozoic deposition. Transport was mainly from the north during the Late Paleocene and Early Eocene, from the west during the Middle and Late Eocene and from the north and northeast during the Oligocene to Quaternary times. The depocenters of the seven major units are generally located marginally, apparently adjoining the source areas. There is only minor evidence for changes in subsidence rates in the basin. A constant rate is assumed from Paleocene to mid Middle Miocene time. For the remaining part of the Cenozoic time an increased rate is indicated.

A tentative relative sea-level curve for the North Sea Basin is proposed. The overall trends of the curve are broadly comparable with the global sea-level curve of Haq et al. (1988). Discrepancies may be caused by differences in the biostratigraphic calibrations. The most pronounced Oligocene sea-level fall is dated as latest Oligocene.

INTRODUCTION

Cenozoic deposits in the central part of the North Sea Basin reach a thickness of more than 3000 m, representing most of the erathem. Records are available in a large number of wells and seismic surveys. The stratigraphy, however, is rather poorly known, especially in the post-Eocene deposits.

The purpose of the present study is to establish a sequence stratigraphic scheme of the Cenozoic deposits in the eastern North Sea, including the Danish North Sea sector and the adjacent parts of the Norwegian, German and Dutch sectors (Figs. 1, 2). The stratigraphic section comprises the siliciclastic sedimentary rocks covering the Danian Limestone (i.e., the Upper Paleocene to Quaternary deposits). However, the uppermost few hundreds of meters are poorly illustrated by the available data. Deposits in the eastern part of the basin are exposed in Denmark, and correlation to the Danish sections provides important additional stratigraphical, palaeoecological and sedimentological information.

PREVIOUS STUDIES

Previous stratigraphic studies in the North Sea Basin mainly concern lithostratigraphic and biostratigraphic problems. Rhys (1974) published a lithostratigraphic nomenclature for the southern North Sea but did not subdivide the Cenozoic succession into lithostratigraphic units. The first lithostratigraphic scheme for the Norwegian and northern UK sectors was established by Deegan and Scull (1977). Isaksen and Tonstad (1989) revised this scheme for the Norwegian North Sea sector. They described several Paleocene formations, overlying the Danian Ekofisk Formation. The younger Cenozoic section was subdivided into two groups, separated by high gamma-ray readings, representing the so-called Mid Miocene unconformity. NAM and RGD (1980) established a lithostratigraphic scheme of the offshore Netherlands region. The Cenozoic siliciclastic succession of the Danish Central Trough was subdivided into seven informal lithostratigraphic units by Kristoffersen and Bang (1982). Nielsen et al. (1986) mapped the Cenozoic stratigraphic units in the central North Sea. Most of the stratigraphic data from the North Sea region has been compiled in a comprehensive paper edited by Vinken (1988).

Vail et al. (1977a,b) studied the Cenozoic succession in several seismic sections from the North Sea. They recognized several sea-level fluctuations and proposed the first relative sea-level curve for the North Sea. Clausen (1991) mapped a large number of seismic sequences in the Danish Central Trough and suggested a preliminary curve of coastal onlap. Stewart (1987)
recognized 10 seismic stratigraphic sequences in the lower Paleogene deposits northwest of our study area. Galloway et al. (1993) subdivided the Cenozoic deposits of the northern and central North Sea into a large number of so-called tectonosequences. Several recent sequence stratigraphy studies concern the Paleogene deposits in the North Sea, in particular Neal et al. (1994), Mudge and Bujak (1994, 1996), Jordt et al. (1995) and Neal (1996).


The present paper was submitted in the spring of 1993. Since then several papers referring to it have been published. They include Michelsen (1994), Danielsen et al. (1995), Jordt et al. (1995), Michelsen et al. (1995), Sørensen and Michelsen (1995), Thomsen and Danielsen (1995), Michelsen and Danielsen (1996).

**GEOLOGICAL SETTING**

During Cenozoic time, the North Sea region constituted a large epicontinental sag basin with a north-south axis above the older Central Trough structure (Nielsen et al., 1986). The basin was bordered by the elevated areas of the Fennoscandian Shield to the northeast, Central Europe to the south, and the British Isles to the west (Fig. 1).

The sag basin developed in Late Cretaceous time. Major older structures, such as the Central Trough, the Ringkøbing Fyn High, the Norwegian-Danish Basin and the Northwest German Basin, became less active and underwent regional subsidence (Glennie, 1990).

Halokinesis of the Zechstein salt influenced Cenozoic depositional patterns in most of the North Sea Basin (Korstgaard et al., 1993). In the Central Trough area and along the Fjerritslev Fault, early Tertiary sedimentation was locally affected by tectonic inversion (Clausen and Korstgård, 1993).

**METHODOLOGY**

Approximately 20,000 km of seismic sections, petrophysical logs from 76 wells and biostratigraphic samples from 14 wells were analyzed for this study (Fig. 2). The majority of the seismic sections are in half scale. Gamma-ray, sonic and resistivity logs cover the main part of the Cenozoic succession, whereas the neutron-density logs cover only the lower part. Synthetic seismograms and velocity logs are available for most wells. The biostratigraphic and lithologic studies are based on cuttings samples, occasional sidewall cores and a few conventional cores.

The stratigraphic analysis follows principally the procedures described by Vail (1987). Stratigraphic surfaces are identified on the basis of an integrated analysis of seismic sections, log data and biostratigraphy. The geometry of the sequences is mapped on the basis of seismic sections. The sedimentary facies is interpreted from log analyses and lithologic descriptions of well samples. Biostratigraphy provides the chronostratigraphic control of the sequence stratigraphic surfaces.

Depths calculated from seismic section on the basis of synthetic seismograms are estimated to be correct within a few meters. In wells either lacking or with poor velocity information, the time-depth conversion is based on a constant velocity function, and the calculated depths are regarded to be correct within an interval of ±20 m.

Boundaries identified on seismic sections are tied to log intervals, the lengths of which are determined by the seismic resolution and the reliability of the time-depth conversions. A more precise location of the boundary within this log interval is evaluated in accordance with the methods described by Van Wagoner et al. (1990).

**BIOSTRATIGRAPHY**

The geologic ages of the sequence stratigraphic surfaces are assessed by biostratigraphic analyses of foraminifera, calcareous nannofossils, and dinoflagellates (Fig. 3, Tables 1A, 1B). These studies are partly overlapping, so that interpretations based on one fossil group may be compared with interpretations based on the others.

Foraminifera were studied in the Upper Eocene to Quaternary sections. The stratigraphic framework is the zonations of King (1983, 1989) (Fig. 3). NSB zones (North Sea Benthic) represent the most important zonation. NSA zones (North Sea Agglutinated) are used where the dominating part of the fauna...
is arenaceous. Planktic species are long ranging in the North Sea and the NSP zones (North Sea Planktic) are considered only supplementary. The ages of some of the zones of King (1983, 1989) have been reinterpreted in the light of more recently published literature.

Dinoflagellate analyses were carried out primarily for Paleogene strata (Fig. 3). The zonations of Heilmann-Clausen (1985, 1988) and Köthe (1990) are used. In Miocene units, where no zonal scheme is used, the ages of particular events are interpreted on the basis of published information from various regions.

Calcareaous nannofossils were studied in Upper Eocene to Middle Miocene units (Fig. 3). We used the standard NP and NN zonation of Martini (1971) with a modification in the NP 23–24 interval (see below).

**CHRONOSTRATIGRAPHY**

Presently, only three Cenozoic boundaries have been precisely defined, namely the Eocene-Oligocene, Oligocene-Miocene and the Pliocene-Pleistocene boundaries. As regards other boundaries, we generally follow the practice used by most Cenozoic stratigraphers for the last 25 years. These workers have developed a chronostratigraphic scale based on successive planktic biozones (mainly planktic foraminifera and calcareaous nannofossil zones) that are calibrated with palaeomagnetic anomalies. Most of these zones have been established in lower or middle latitude areas and their identification is less certain at higher latitudes, such as the North Sea, where many key fossils are absent or have different ranges.

Globally used zones identified in the North Sea are essentially limited to those based on calcareaous nannofossils (Fig. 3). However, only some of the NP and NN zones can be safely identified by their marker species. In other intervals, the key fossils are absent, and correlation is based on substitute markers. Often calcareaous nannofossils are entirely absent; in these cases we have inferred the ages by means of the other two fossil groups.

Most of the Paleogene stages were originally defined in NW Europe. Where possible, we have identified these stage by a direct comparison with the type areas.

Upper Paleocene and Eocene chronostratigraphic interpretations are mainly based on dinoflagellates. Nannofossil zonation in Upper Paleocene strata of northwest Europe allow a correlation to some of the North Sea dinoflagellate zones. The Selian limestone is placed as originally defined in Denmark (i.e., at the base of the greensand unconformably overlying the Danish limestone). This boundary coincides with the boundary between the Viborg dinoflagellate zones 1 and 2 (Heilmann-Clausen, 1994) and it is considered to be near the NP4-NP5 boundary (Thomsen and Heilmann-Clausen, 1985; Thomsen, 1994) (Fig. 3).

The Viborg zone 4 assemblage is comparable to that in the lower part of the Thanet Beds in Southern England. These strata are assigned to the NP 6, NP 7 and NP 8 zones (Knox et al. 1994). Viborg zone 5 is correlated with NP 9 in Belgium (Heilmann-Clausen, 1985).

The boundary between Paleocene and Eocene units is taken at the uppermost occurrence of the dinoflagellate *Apectodinium augustum*, following most North Sea biostratigraphers (e.g.,

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**Fig. 3.—**Chronostratigraphic interpretation of the biostratigraphic zones.
### Table 1A—Biostratigraphy and Chronostratigraphy of Sequence Stratigraphic Surfaces

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**Legend:**
- **NSP:** Number of stratigraphic points
- **NSB:** Number of sequence boundaries
- **NSA:** Number of successions
- **NP:** Number of points
- **Dino:** Dinosaurian age
- **SB:** Sequence boundary
- **MFS:** Maximum flooding surface
- **Hiatus:** Disconformity

**Notes:**
- Late Cenozoic uplift and erosion
- Position of SB and MFS uncertain

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**Sequence boundary (SB) — — — — — — Maximum flooding surface (MFS)
The *A. augustum* zone (defined by the range of *Apectodinium augustum*) equates Viborg zone 6 and lowermost part of Viborg zone 7 (see Fig. 3). The *D. oebisfeldensis* zone equates the upper part of Viborg zone 7 (above the top of *A. augustum*). The ‘nt 2a middle’ palynozone of Jacque and Thouvenin (1975) is a facies zone occurring within the *D. oebisfeldensis* Zone.

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The Eocene-Oligocene boundary is defined in Italy. It falls in the NP 21 zone and little below the uppermost occurrence of the dinoflagellate *Aresospheridium diktyplopus* (Brinkhuis and Biffi, 1993). In the present study, the highest occurrence of *A. diktyplopus* was also found in the NP 21 zone. The Eocene-Oligocene boundary in the North Sea Basin seems, therefore, to occur just below the top of *A. diktyplopus* (top of zone D12).

Calcareous nannofossils are generally abundant in the Oligocene units of the North Sea Basin. However, the marker species selected by Martini (1971) are absent. Instead Martini and Müller (1975, 1986) and Müller (1978) suggested that the first occurrence of *Cyclicargolithus abisectus* could be used as a substitute marker for the lower boundary of NP 24 at high latitudes and the first occurrence of *Pontosphaera enormis* as a marker for NP 25. *C. abisectus* shows some very characteristic size variations that allow a subdivision of the strata usually referred to NP 24 into two subzones, a lower NP 24a (small forms) and an upper NP 24b (large forms, max. diam. >13 μm, E. Thomsen, in prep.) (Fig. 3). Comparison with strata in Germany shows that the NP 24a/NP 24b boundary roughly correlates with the Rupelian/ Chattian boundary and that NP 24b corresponds, approximately, to Chattian A. However, correlation with the international NP zonation suggests that NP 24b correlates with NP 24 *sensu stricto* and that NP 24a may correlate with the upper part of NP 23 (see also Firth, 1989). The Oligocene/Miocene boundary is placed at the last occurrence of *Dictycocccites bisecta*.

The top of foraminiferal zone NSB 8b in the Upper Oligocene is based on the last common occurrence of *Elphidium subnodosum* (King, 1989). This species is extremely rare in the study area, and the zone was only recognized in the A/12-1 well in the Dutch sector. NSB 8b is, therefore, included in NSB 8c in the present study.

No foraminiferal events are known from the North Sea Basin which can be safely used to identify the Plio-Pleistocene boundary in cuttings samples (i.e., by first downhole occurrence). However, two events are available in the vicinity of the boundary. These are the uppermost occurrences of planktic species *Neogloboquadrina atlantica* (in the Upper Pliocene) and of benthic species *Elphidiella hannai*. The latter species is placed in the Lower Pleistocene in traditional northwest European stratigraphy, but its position relative to the international boundary is uncertain; it may be either in Upper Pliocene or Lower Pleistocene strata.

**CENOZOIC SEQUENCE STRATIGRAPHY**

Twenty-one sequences were defined within the respective depocenters. In some cases, it was difficult to tie the sequences around in larger areas, probably because they locally are beyond seismic resolution. The sequences are, therefore, grouped into seven major sequence stratigraphic units. The boundaries of these units are easily recognized on the seismic sections in the entire study area (Michelsen et al., 1995) (Fig. 4). The term “major sequence stratigraphic unit” is used informally. It must not be confused with terms such as megasequence or supersequence, suggested as formal chronostratigraphic terms by Haq et al. (1988). The general features of the units and the sequences are shown in Figure 5. Units 4–7 are truncated by late Cenozoic erosion in the northeastern part of the study area, and the proximal parts of the units are missing.

**Sequence stratigraphic analyses**

Two types of surfaces have been interpreted in this study: sequence boundaries and maximum flooding surfaces.

The sequence boundaries are identified primarily in the sequence depocenters by seismic onlap, toplap and truncation features. In a basinward direction away from the depocenters, the boundary can often be followed as a strong seismic reflection within a concordant reflection pattern.

The overall gamma-ray log pattern of our sequences is composed of an upward increasing trend succeeded by an upward decreasing trend, separated by a narrow stratigraphic interval with high gamma-ray values. The interval with high gamma-ray values correlates with an internal downlap surface on the seismic section, and it is usually synchronous within the biostratigraphic resolution (Michelsen and Danielsen, 1996). The gamma-ray peaks are sometimes associated with sediments rich in glauconite or organic matter. We interpret high gamma-ray intervals and the corresponding seismic downlap surface as a maximum flooding surface or condensed section (sensu Dixon and Dietrich, 1988; Loutit et al., 1988; Galloway, 1989).

The log and seismic reflection patterns of the sequences show much variation both within and between sequences. These differences, however, are usually subtle and difficult to interpret in terms of changes in the depositional environment. Transgressive and lowstand deposits proved to be especially difficult to distinguish. Below, we will describe some of these variations and indicate how they are interpreted.

An interval with low or constant to upward decreasing gamma-ray values is often intersected between the lower sequence boundary and the upward increasing trend. In the proximal part of the sequence, the interval may attain a blocky pattern with distinctly low values. In the distal part, it is often associated with downlaps on the lower sequence boundary and a very high position for the maximum flooding surface (Michelsen and Danielsen, 1996).

The intervals with decreasing or distinctly low gamma-ray values are interpreted as representing prograding lowstand deposits. Increasing gamma-ray values in the overlying interval indicate upward fining deposits, which may represent transgressive conditions. The interval above the maximum flooding surface is often characterized by an aggrading to prograding seismic reflection pattern. Decreasing gamma-ray values here
FIG. 4.—Northeast-southwest seismic section (RTD81-22) showing principal features in the distribution of the major sequence stratigraphic units 1–2, 3, 4, 5, 6, and 7. The concordant reflections of units 1 and 2 and the overlying chaotic pattern of sequence 3 are illustrated. Note the prograding to aggrading character of units 4, 5, and 6 and the overall aggrading character of unit 7. The horizontal scale is strongly reduced.

indicate upward coarsening deposits, which may represent highstand conditions.

We were rarely able to identify the bounding surface between possible lowstand and transgressive deposits on the seismic sections, and a systems tract interpretation in accordance with the concepts of sequence stratigraphy (Van Wagoner et al., 1987, 1988, 1990; Posamentier et al., 1988; Posamentier and Vail, 1988) was therefore not attempted. Interpretation was also hampered by the fact that the proximal parts of the sequences mostly are located outside of the study area or have been removed by erosion.

Major Sequence Stratigraphic Unit 1

Distribution and general features.—

Unit 1 is recognized in the entire study area. Its thickness varies between approximately 300 m in the depocenter in the northern part of the study area and less than 50 m in the distal area to the southeast (Fig. 6).

The depocenter and the proximal part of the unit are located north of our study area. In the study area, the unit constitutes a lenticular shaped, southwestward prograding clinoform. The seismic reflections show, generally, an oblique progradational pattern in the downlap area and a concordant subparallel reflection pattern in the landward direction.

The unit is subdivided into two sequences 1.1 and 1.2 on the basis of log interpretation. The unit is mostly very thin and it is generally difficult to identify internal structures on seismic sections (Figs. 5, 7).

Surfaces.—

The lower boundary of sequence 1.1 is placed at the base of an interval with upward decreasing sonic velocities and increasing gamma-ray values (Figs. 7, 8). On seismic sections, the boundary is normally characterized by a strong, low frequency, continuous reflection. Downlaps and onlaps on the lower boundary are seen in the northeastern area close to the depocenter. The boundary is located at the top of the Danian limestone or Upper Cretaceous chalk. Erosional features in the limestone are widespread on the margins of the Ringkøbing-Fyn High.

The lower boundary of sequence 1.2 is placed below an interval with high gamma-ray readings (Figs. 7, 8). The maximum flooding surface is in both sequences indicated by distinct gamma-ray peaks (Fig. 8).

Facies and depositional environment.—

The inclination of the internal seismic reflections suggests that sediment transport was mainly from the north and northeast (Fig. 6).

The gamma-ray log pattern of sequence 1.1 is normally composed of an upward increasing trend succeeded by an upward decreasing trend (Fig. 8). The lower part corresponds to the North Sea Marl (Kristoffersen and Bang, 1982). The interval with high gamma-ray values, which apparently represents the maximum flooding surface, consists of non-calcareous, silty, dark grey clay. The decrease in gamma-ray values in the upper part of the sequence is caused by the presence of a very fine grained smectite-rich, greenish or reddish clay.

Sandy deposits, below the gamma-ray highs in the southern part of the Norwegian sector and in the adjacent part of the Danish sector (e.g. Cleo-1 and Elna-1) on both seismic sections and logs, are interpreted as submarine fan deposits.

The gamma-ray log of sequence 1.2 is composed of two parts. The lower part is characterized by high gamma-ray values
FIG. 5.—Sequence stratigraphic interpretation of the Cenozoic section in the northeastern North Sea outlined by seismic section RTD81-22 (sp 4250-8300), and by log profiles of the Cleo-1, Elna-1, L-1, Ibenholt-1 and Inez-1 wells. For location see Figure 2.
FIG. 5.—Continued
ending upwards with a distinct peak. The upper part is distinguished by very low gamma-ray values and, in addition, high sonic log readings (Fig. 8). The deposits in the lower part of the sequence consist of laminated, organic-rich, silty clays. The deposits indicate a change to an anoxic bottom environment, possibly due to a silled basin (Ziegler, 1982). The clay is identified as the Sele Formation (Deegan and Scull, 1977). The maximum flooding surface is identified by the distinct gamma-ray peak (Fig. 8).

The upper part of sequence 1.2 corresponds to the main volcanic ash phase in the North Sea, the Balder Formation (Deegan and Scull, 1977). The deposits indicate a return of moderately oxic conditions.

Thus, within the study area, unit 1 consists of fine-grained distal deposits. The succession is thin and condensed. Apart from a minor, but regional hiatus at the boundary between the two sequences, the succession seems fairly complete. The lower parts of both sequences were deposited in a sublittoral environment. During the deposition of sequence 1.1, water depth increased to bathyal conditions. The increase in water depth was apparently less marked during the deposition of sequence 1.2.

Sequence 1.1.—

Maximum thickness is about 60 m in the northwestern part of the study area. The age of the sequence is Late Paleocene (Tables 1A, 1B). The lower boundary is identified below samples of dinoflagellate zones Viborg 2–3. On the basis of data from Heck and Prins (1987) the boundary is interpreted to occur in the uppermost part of NP 4. The maximum flooding surface is below the uppermost occurrence of the dinoflagellate Isabelidinium? viborgense.

Sequence 1.2.—

Maximum thickness is 238 m in the 9/11-1 well. The age of the sequence is latest Paleocene to earliest Eocene (Tables 1A, 1B). The lower boundary is located between dinoflagellate zones Viborg 5 and 6 in the Mona-1 well. The maximum flooding surface is in zone Viborg 7, above the uppermost occurrence of Apectodinium augustum.

Major Sequence Stratigraphic Unit 2

Distribution and general features.—

Unit 2 is recognized in the entire study area, except for the R-1 and S-1 wells on the Ringkøbing-Fyn High. The proximal part and most of the depocenter are situated north and west of the study area. Within the study area, two separate depositional maxima are recognized; a northern maximum in the Danish-Norwegian Basin with a greatest thickness of 250 m and a western maximum with a greatest thickness of 150 m (Fig. 9). The unit is locally below seismic resolution. Only one sequence has been recognized.

In the northern depositional maximum, the internal seismic reflections form a sigmoidal and weakly progradational clinoform. Onlaps and downlaps are recognized north and south of the depositional maximum, respectively (Fig. 9). In general, the seismic reflections seem to be concordant with the lower boundary, giving the stratal pattern a draped configuration (Fig. 10).

Surfaces.—

In the northern depositional maximum the boundary is defined by downlaps, onlaps, toplaps (Fig. 7) and erosional truncations. A few onlaps are also identified in the western depositional maximum. In most of the study area the boundary is characterized by a strong, nearly continuous, seismic reflection. In the Danish Central Trough area, the seismic reflections below the boundary are continuous with high amplitudes. The reflections above show low continuity and low amplitude.

On logs, the lower boundary is placed at an abrupt upward decrease in the sonic velocities (Fig. 8). The decrease reflects the transition from the main ash phase of unit 1 to the fine-grained clay of unit 2.

The maximum flooding surface is recognized by high gamma-ray peaks (Fig. 10). It usually occurs in the lower part of the sequence, and lowstand deposits are only interpreted in the Norwegian 3/5-1 and Dutch A/18-1 wells, where the surface is situated relatively high.

Facies and depositional environment.—

The direction and position of the downlaps indicate sediment transport from the north in the northern depositional maximum (Fig. 9). In the western maximum, sediment transport from the west or southwest is assumed.

The lower part of the unit consists of a reddish, very fine-grained clay. The gamma-ray peaks at the maximum flooding
FIG. 7.—Seismic section (CGT-8103B) showing onlaps, downlaps and top-laps in sequences 1.1 and 1.2. The sequences are correlated with logs from the 9/11-1 well. A thick layer of sand above the maximum flooding surface in sequence 1.2 in the 9/11-1 well is interpreted as “a basin-margin deposit” (Isaksen and Tonstad, 1989). A sequence boundary may be present below the sand. For location see Figure 2.

surface probably reflect glauconite beds. The upper part consists of a uniform very fine-grained greenish clay. The maximum flooding surface is correlated biostratigraphically to the R6-lower L2 beds of the Røsnæs Clay and Lillebælt Clay Formations in Denmark (Heilmann-Clausen et al., 1985). Black, sapropelic clay bands are usually present within this interval in the onshore area.

The very fine clay, the low rate of sedimentation (as evidenced by biostratigraphy) and the dominance of pelagic fossils indicate a depositional environment in the bathyal zone. The deposits are condensed and the study area was clearly situated in a distal position. The reddish and greenish colors suggest variable degrees of oxygen surplus in the bottom waters.

Age.—

The age of the unit is Early Eocene to earliest Middle Eocene. In the 10/5-1 well the lower boundary apparently coincides with the Viborg 7—W. astra zone boundary (Tables 1A, 1B). The maximum flooding surface occurs within or at the top of the A. diktyoplokus zone.

Major Sequence Stratigraphic Unit 3

Distribution and general features.—

The depocenter is located along the eastern margin of the Mid North Sea High with an almost north-south orientation. The maximum thickness is about 800 m in the Danish Central Trough area (Fig. 11). The unit is thin or absent in the eastern part of the study area. Only one sequence has been recognized, but indications from onshore Denmark and from the Norwegian sector (Fig. 12) suggest that the unit might consist of more than one transgressive/regressive cycle.

In general, the sequence can be characterized as a lenticular body with the internal seismic reflections inclining slightly to the east (Fig. 13). In the upper part, however, the reflections are parallel with the upper boundary.

Surfaces.—

The lower boundary is identified by onlaps in the depocenter and downlaps to the east (Figs. 5, 13). The boundary is elsewhere recognized by a change in seismic configuration across the boundary. The position of the boundary on logs is determined by conversion from the seismic sections.

A maximum flooding surface is indicated by a minor gamma-ray high in the lower part of the unit (Fig. 12). In the Norwegian area small seismic units with double-directed downlaps on the lower boundary may be lowstand units.

Facies and depositional environment.—

The location of the depocenter and the general inclination of the internal seismic reflections indicate a main direction of sediment transport from the west (Fig. 11). In the Dutch area, however, the downlap directions suggest south-southwest directions.

FIG. 8.—Gamma-ray and interval transit time logs from the Cleo-1, Elna-1, L-1 and Ibenholt-1 wells illustrate the log signature of sequences 1.1, 1.2 and 2. For location see Figure 2.
FIG. 9.—Isopach map of major sequence stratigraphic unit 2. Onlap and downlap areas and direction of sediment transport are indicated.

Analyses of cuttings reveal a distinct facies change from the central to the eastern and southern parts of the basin. In the central North Sea, the deposits consist of non-calcareous fine clays. To the east and south, the content of calcium carbonate increases, especially in the upper part which may be developed as a highly calcareous marl. The increase in the content of calcium carbonate is reflected in lower gamma-ray values.

The fine-grained pelagic and hemipelagic deposits indicate a deep, low-energy depositional environment. The change in the content of carbonate suggests a gradient with a slightly shallower setting to the east and south as compared to the central part of the basin. In the central North Sea, bottom water was apparently slightly acidic (King, 1983). The depositional environment here is interpreted to have been upper bathyal. The calcareous sediments in the eastern and southern periphery may have been deposited in the lower part of the sublittoral zone.

Age.—

The age of the unit is earliest Middle to latest Eocene. The lower boundary is near the D. pachydermum-W.'articulata-ovalis' zone boundary based on the Mona-1 well (Tables 1A, 1B). In the less complete A/18-1 well, it is situated 5 m above the W.'articulata-ovalis' zone. The discrepancy may be due to uncertainty in the conversion from sample depths to log depths. The maximum flooding surface occurs either in the P. geminatum or in the lower part of the A. arcuatum zone.

FIG. 10.—Major sequence stratigraphic unit 2 shown on a north-south seismic section (CGT81-03B). The sequence is correlated with the 9/11-1 well. The maximum flooding surface is identified as a high gamma-ray peak. For location see Figure 2.
better be characterized as an oblique prograding configuration identified by toplaps against the upper boundary. South and west of the depocenter the reflections are mainly subparallel.

Gamma-ray readings are slightly higher than in unit 3 (Fig. 5). Values fluctuate considerably, but a general upward increase is discernible. Wells located in the proximal part of the sequence (e.g., K-1, F-1 and Inez 1) show much lower average values than wells located in the depocenter and in the distal part (e.g., Elna-1, Cleo-1 and Mona-1).

Unit 4 is subdivided into four sequences: 4.1, 4.2, 4.3 and 4.4, with sequence 4.1 positioned closest to the Fennoscandian Shield and 4.2, 4.3, and 4.4 gradually displaced towards the center of the basin.

Surfaces.—

The sequence boundaries are, in general, defined by onlaps in the proximal part of the depocenters and downlaps in the distal part (Fig. 15A,B). Outside the depocenters, the boundaries are characterized by continuous concordant reflections (Figs. 5). Toplaps and evidence for erosion of the underlying sequences have been observed in or around the depocenters at all sequence boundaries except for the lower boundary of sequence 4.1. At the lower boundary of sequence 4.2, erosional truncation has also been observed in the distal part of the sequence. The lower boundary of sequence 4.1 (i.e., the boundary between units 3 and 4) is, furthermore, characterized by a change in the prevailing reflection amplitude in the distal part of the sequence.

The lower boundary of sequence 4.1 is marked by a distinct upward increase in gamma-ray values, (Figs. 5, 16). An upward increase also characterizes the lower boundary of sequence 4.2 in distally located wells. In the proximally located Inez-1 well, however, a distinct shift to lower values is seen (Fig. 5). The lower boundaries of sequences 4.3 and 4.4, in contrast, are marked by an upward decrease in gamma-ray readings in most wells.

The maximum flooding surface of each sequence is placed at an interval with relatively high gamma-ray values (Fig. 16), which correlates with an internal seismic downlap surface.
Facies and depositional environment.—

The overall lithology is an olive grey to brownish grey, silty clay. The higher gamma-ray values as compared to those of unit 3 are apparently associated with a higher content of mica and illite (Danielsen, 1989) and may be related to a different source of sediments and a closer position to the denudation area. The internal downlap pattern indicates sediment transport from the north and northeast (Fig. 14) as opposed to the west in unit 3 below. The gamma-ray log suggests an increase in the content of silt and sand in the proximal part of the unit in the northeastern part of the study area. This interpretation is confirmed by the inspection of cuttings samples. They reveal the presence of thick sand bodies in the Inez-1, F-1 and K-1 wells (Fig. 5).

The large prograding bodies of quartz sand in the proximal part of the unit are interpreted as lowstand units. They were apparently deposited in a near-shore upper sublittoral environment. The homogeneous olive-grey to brownish silty clay in the depocenter and in the distal part was probably deposited in the lower sublittoral to upper bathyal zones. Calcareous fossils are absent in the distal part of sequences 4.1 and 4.2 indicating a somewhat reduced circulation in the deepest part of the basin. The abundance of large calcareous nanofossils and the considerable increase in the areal extent of calcareous benthic foraminifera in sequences 4.3 and 4.4 indicate a gradual improvement in the water circulation. The content of total organic carbon (TOC) is moderately low. In the Lulu-1 well located distally to the depocenter, it varies between 1.48 and 1.84% (Danielsen, 1989). This may suggest open marine conditions with rather well-oxygenated bottom water.

Sequence 4.1.—

Maximum thickness in the depocenter is about 510 m. Seismic reflections below the maximum flooding surface are mainly aggrading. In the depocenter, prograding reflections downlap on the aggrading package (Fig. 15B). Progradation is towards the south. A thick interval of upwards decreasing gamma-ray values in the proximal part of the sequence indicates the presence of lowstand deposits (Fig. 5).

The age of the sequence is earliest Oligocene. The lower boundary is located closely beneath the top of the D12 zone (top of *Areosphaeridium diktyoplokus*) in NP 21 and in NSA...
FIG. 15(A,B).—Sequences 4.1, 4.2, 4.3, and 4.4 shown on the seismic section RTD81-22. The sequence boundaries are characterized by onlaps and downlaps; in sequence 4.1 also by toplaps. Note the change from a concordant seismic reflection pattern below the lower boundary of sequence 4.1 to a prograding pattern above. For location see Figure 2.

Sequence 4.2.—

Maximum thickness is about 300 m. The reflections below the maximum flooding surface are mainly parallel with a high degree of continuity (Fig. 15B). Reflections above the surface aggrade in the proximal part of the depocenter and prograde in the distal part. A few internal downlaps distally in the upper part indicate that the sequence might be composite.

Constant to slightly upward decreasing gamma-ray values in the Central Trough area and a blocky pattern with low gamma-ray values in the proximal part are interpreted as lowstand deposits (Fig. 5).

The age of the sequence is Early Oligocene. The lower boundary is located in the NP 23, NSB 7a, NSA 7 and D14na zones (Tables 1A, 1B). The maximum flooding surface occurs within NP 23 or at the NP 23/24a boundary.

Sequence 4.3.—

Maximum thickness is about 450 m. The seismic reflections of sequence 4.3 generally show a parallel stacking pattern with downlaps south and west of the depocenter (Fig. 15A). Internal truncations are present in the Ibenholt-1 area.

Slightly prograding log trends suggest that lowstand deposits may be present in the depocenter and the distal part of the sequence (Fig. 5).

The age of the sequence is late Early to early Late Oligocene. The lower boundary is located in the NP 24a, NSB 7b, NSA 7 and D14na/nb zones (Tables 1A, 1B). The maximum flooding surface occurs within NP 24b.

Sequence 4.4.—

Maximum thickness is about 400 m. The internal reflections aggrade in the depocenter and are parallel to the upper boundary in the downlap area close to the Ringkøbing-Fyn High (Fig. 15A). Small seismic units with double directed downlaps on the lower boundary south of the L-1 well may be lowstand

6b (Tables 1A, 1B). The maximum flooding surface is in NP 22.

Fig. 16.—Seismic section (DK1-0515) showing the sequences of units 4 and 5. The sequences are correlated with the L-1 well. For location see Figure 2.
units. Upward decreasing gamma-ray values suggest that lowstand deposits also may be present in the neighboring Elna-1 and Ibenholt-1 wells (Fig. 5).

The age of the sequence is Late Oligocene. The lower boundary is located in the NP 24b, NSB 8a and D15 zones (early Chattian) in the 2/2-1 well (Tables 1A, 1B). The maximum flooding surface occurs within NP 25.

**Major Sequence Stratigraphic Unit 5**

**Distribution and general features.—**

Major sequence stratigraphic unit 5 is identified in most of the study area. Its depocenter is located on the northwestern margin of the Ringkøbing-Fyn High basinwards of the depocenter of unit 4 (Fig. 17). The maximum thickness is approximately 500 m. The unit thins rapidly in the southern part of the Danish sector, and it has not been observed in the Dutch A/18-1 well.

Unit 5 is associated with a pronounced basinward shift of coastal onlap (Fig. 5). The unit is lenticular with downlaps to the west and southwest and onlaps to the east and northeast. In the depocenter, the reflection pattern is mainly aggrading. West of the depocenter, the pattern becomes prograding in the upper part.

Unit 5 is subdivided into four sequences: 5.1, 5.2, 5.3, and 5.4. The depocenter of sequence 5.1 is restricted to a small area on the northern margin of the Ringkøbing Fyn High basinwards of the depocenter of sequence 4.4. The depocenters of 5.2 and 5.3 are located slightly landwards of that of 5.1. The depocenter of 5.4 is located basinwards of the depocenter of sequence 5.3.

**Surfaces.—**

The boundaries between unit 4 and 5 and between sequences 5.1 and 5.2 are identified by onlaps and downlaps in the depocenter. On gamma-ray logs, they correlate with a small upward decrease and increase, respectively, in the readings in the basinal area (Fig. 5). The lower unit boundary is, furthermore, marked by toplaps and by erosion proximal to the depocenter. The boundary between sequences 5.2 and 5.3, and 5.3 and 5.4 have only been defined on logs. They are placed at the base of intervals with low gamma-ray readings (Figs. 5, 16).

The maximum flooding surface is in all sequences placed at the top of an interval with increasing gamma-ray readings (Figs. 5, 16). In sequences 5.1, 5.2 and 5.3, it is generally located in the middle or upper part of the sequence. In sequence 5.4, it is located in the lower part.

**Facies and depositional environment.—**

Unit 5 consists mainly of grey-brown to almost black, silty clay with varying amounts of sand and mica and a general high content of organic matter. The downlap directions of the seismic reflections suggest sediment transport from the northeast and east.

The high content of silt and sand combined with a significant amount of planktic foraminifera suggest that most of the unit was deposited in the sublittoral zone. A lateral depth gradient is indicated by a change in the proportion of agglutinated to calcareous benthic foraminifera. Calcareous forms dominate in the proximal part of the unit (Ibenholt-1 and Inez-1 wells), while agglutinated forms dominate in the distal part (Mona-1 and 2/2-1). The depositional environment probably ranged from littoral in the proximal part to lower sublittoral in the distal part.

The fossil assemblages indicate a fully marine depositional environment. The amount of organic matter increases upwards in most wells. In the Lulu-1 well, the TOC content increases from 1.5% at the base of sequence 5.1 to 4.1% in sequence 5.4 (Danielsen, 1989). The sediment color in sequence 5.4 is almost black. Thus, the oxygen level of the bottom waters probably changed from aerobic to dysaerobic.

The increasing areal extent and onlap of the four sequences indicate successively higher relative sea levels. On the other hand, gamma-ray values generally decrease upwards in spite of the higher content of organic matter. This is undoubtedly due to a general increase in the content of silt and sand and may possibly indicate a slight shallowing in the depositional environment.

**Sequence 5.1.—**

Maximum thickness is about 140 m. The reflections onlap to the east and downlap to the west. The lowermost part of the sequence shows upward decreasing gamma-ray values in some wells and is interpreted to be a lowstand unit (Fig. 5). This explanation is supported by the high sand and silt content in the corresponding cuttings samples in the L-1 and Elna-1 wells. A small lenticular unit with a hum-
mocky reflection pattern and downlaps in both directions near the Cleo-1 well is also interpreted to be part of a lowstand unit.

The age of the sequence is Latest Oligocene. The lower boundary is placed in the NP 25 zone and in the lower part of the NSB 8c zone in most wells (Tables 1A, 1B). This is in agreement with a dinoflagellate dating from the Mona-1 well, which indicates a Chattian age. However, in the A/12-1 well the boundary is located in the NSB 8b zone (see Chronostratigraphy). The maximum flooding surface occurs within NP 25.

**Sequence 5.2.—**

The maximum thickness is 152 m in the L-1 well. The sequence is dominated by parallel seismic reflections of low continuity (Fig. 16). The internal reflections downlap to the west.

An interval with low gamma-ray values and sandy deposits in the lower part of the Elna-1, L-1 and Ibenholt-1 wells may possibly represent a lowstand unit (Fig. 5).

The age of the sequence is Latest Oligocene to Early Miocene (Tables 1A, 1B). The lower boundary is located in the NSB 8c zone in the Mona-1, L-1, Ibenholt-1 and R-1 wells. The maximum flooding surface is in NSB 8c or NSB 9. The dinoflagellates suggest an Aquitanian age.

**Sequence 5.3.—**

The maximum thickness is 107 m in the L-1 well. The internal seismic reflections downlap in the western part of the depocenter. East of the depocenter, the sequence is dominated by onlap of parallel reflections.

Low gamma-ray values in the lowermost part of the sequence (Fig. 5) correlate with a high silt and sand content in the cuttings samples and may possibly reflect lowstand conditions. The interval above the maximum flooding surface contains abundant quartz grains in the proximally located Ibenholt-1 well.

The age of the sequence is Early Miocene (Tables 1A, 1B). The entire sequence is located in the NSP 10, NSB 9 and NSA 10 zones. The dinoflagellate analyses indicate an age not younger than Aquitanian.

**Sequence 5.4.—**

The maximum thickness is 172 m in the Elna-1 well. The sequence is characterized by upward increasing gamma-ray values in the lower part and constant or decreasing values in the upper part (Fig. 5).

The age of the sequence is Early Miocene (Tables 1A, 1B). The lower boundary is located in the NSB 9 zone, and in the Aquitanian on the basis of dinoflagellates. Calcareous nanofossils, indicating NN 1–2, were recorded above the maximum flooding surface.

**Major Sequence Stratigraphic Unit 6**

**Distribution and general features.—**

Unit 6 occurs over most of the study area. The depocenter is located on the southwestern part of the Ringkøbing-Fyn High south of the depocenter of unit 5 (Fig. 18). The maximum thickness is about 550 m.

Unit 6 is subdivided into three sequences (6.1, 6.2 and 6.3) on the basis of gamma-ray logs from Danish wells. Each sequence consists of a thin basal interval with a distinct gamma-ray low followed by an upward increasing trend and finally an upward decreasing trend.

Two different seismic reflection packages can be recognized (Fig. 19). The lower package consists of a thin almost transparent unit overlain by a thick unit dominated by an aggrading pattern. The upper package is characterized by a sigmoidal prograding pattern with onlaps to the northeast and downlaps to the southwest.

The three sequences are recognized on a few seismic sections in the area around the Mona-1 well (Fig. 5). Sequence 6.1 corresponds roughly to the lower package on the seismic sections; sequences 6.2 and 6.3 to the upper prograding package. Measurements of thickness variations in a large number of wells clearly indicate that the depocenters of the three sequences gradually moved in a basinward direction from the northeast to southwest.

The average gamma-ray level in sequence 6 is lower than in the underlying sequence 5.

**Surfaces.—**

The lower boundary is identified by onlaps and by toplaps (Fig. 19). South of the Ringkøbing-Fyn High the reflections are mainly concordant with the lower boundary. On the gamma-ray logs the boundary is located at the base of a minor upward decrease.
The maximum flooding surfaces are mostly located in the lower to middle part of the sequences. In the Dutch wells, however, the maximum flooding surface of sequence 6.1 almost coincides with the upper sequence boundary.

**Facies and depositional environment.**

The general inclination of the internal reflections indicates sediment transport from the north or northeast. The sediment consists of a dark, greenish grey to dark brown, silty clay. The fossil assemblages suggest a fully marine depositional environment. The relatively low gamma-ray values as compared to unit 5 and the upward decreasing readings probably reflect higher amounts of silt and sand and a gradual shallowing of the depositional environment. The distinct gamma-ray lows in the lower part of the sequences in many wells may perhaps be attributed to lowstand deposits.

The content of planktic foraminifera and the proportion of agglutinated to calcareous benthic foraminifera suggest an average water depth ranging from middle sublittoral in the proximally located Ibenholt-1 well to uppermost bathyal in the distally located Mona-1 well. The content of organic matter indicates that the bottom water, periodically, suffered from a reduced oxygen content, especially during the deposition of the sequences 6.1 and 6.3. The amount of TOC varies between 2.5% and 3.4% in the Lulu-1 well (Danielsen, 1989). The highest values are measured in the lower and upper parts of the unit.

**Sequence 6.1, 6.2 and 6.3.**

The age of sequence 6.1 is Early Miocene (Tables 1A, 1B). The lower boundary is located in the NSB 9 zone and the nannofossil NN 1–2 zones (between NN 1–2 and NN 3–4 in well 2/2-1). The maximum flooding surface is in NSB 10 in most wells.

The age of sequence 6.2 is Early Miocene (Tables 1A, 1B). The entire sequence is located in the NSB 10 zone. A NN 1–2 zonal assignment in Ibenholt-1 is based on a poor assemblage and is therefore uncertain.

The age of sequence 6.3 is latest Early Miocene to earliest Middle Miocene (Tables 1A, 1B). The lower boundary is located in the NSB 10 and NSP 11 zones in most wells. Dinoflagellate analyses refer the boundary to late Early Miocene. Identification of NN 3–4 above the boundary in Mona-1 and Ibenholt-1 supports this interpretation. The maximum flooding surface is in NSB 10 or NSB 11 and between samples of NN 3–4 and NN 5–6.

**Major Sequence Stratigraphic Unit 7**

**Distribution and general features.**

Unit 7 is present in most of the study area. The depocenter is located in the Central Trough area, west and northwest of the depocenters of units 4, 5, and 6 (Fig. 20). Maximum thickness is about 1600 m.

Unit 7 is a lenticular prograding clinoform (Fig. 21). The onl laps migrate eastwards in the Norwegian-Danish Basin and on the Ringkøbing-Fyn High. The downlaps migrate westwards across the Central Trough. The gamma-ray pattern shows a slight upward decreasing trend in most of the unit, but the readings fluctuate considerably. One or two distinctive peaks in the lowermost part separate a thin basal interval with upward increasing values from the main section.

Unit 7 is subdivided into six preliminary sequences, 7.1–7.6. The sequence boundaries are identified on a limited number of seismic sections and subsequently converted to the gamma-ray logs. Further studies may result in a more detailed subdivision.

**Surfaces.**

The lower boundary is characterized by onlaps in the Norwegian-Danish Basin and on the Ringkøbing-Fyn High and by downlaps in the Central Trough and German and Dutch areas (Fig. 21). Possible erosional truncations of the underlying unit are seen in the northern part of the study area. On the gamma-ray logs, the boundary is placed at the base of an interval with upward increasing values ending with one or two very prominent peaks (Figs. 5, 21).

The upper boundary is preliminarily placed at the sea floor. It is not a valid sequence boundary, but the quality of the present data is rather poor in the uppermost part of the section. Further studies on new seismic data may identify a major sequence stratigraphic boundary between sequences 7.4 and 7.5, roughly corresponding to the Pliocene-Pleistocene boundary.

**Environment and facies interpretation.**

The deposits of the thin basal interval below the gamma-ray peaks consist of black, organic-rich, smectitic clays rather simi-
Fig. 20.—Isopach map of unit 7. Onlap and downlap areas and the direction of sediment transport are indicated. Please note that the contoured TWT-values include the water column.

Fig. 21.—Seismic section (RTD81-22) showing the four lowermost sequences of unit 7. The sequences are correlated with the Mona-1 well. The high gamma-ray peak lowermost in sequence 7.1 correlates with the seismic downlap surface at or near the base of the unit. The downlap surface is interpreted as the maximum flooding surface of sequence 7.1. A gamma-ray high slightly above the maximum flooding surface indicates that sequence 7.1 is composed of more than one cycle. Notice that the horizontal scale is strongly reduced. For location see Figure 2.
grading conditions (Fig. 21). The biostratigraphic evidence indicates a high rate of sedimentation except for the lowermost part below the gamma-ray peaks.

Sequences 7.1, 7.2, 7.3, 7.4, 7.5 and 7.6.

All sequences are present in the Dutch Central Trough area. To the north, in the Danish Central Trough, only the lowermost four sequences have been identified (Fig. 21). Sequences 7.1 and 7.2 are dominated by aggrading reflection patterns but with an overall progradation toward the west. Sequence 7.3 is lenticular and composed of a series of westward outbuilding units with a prograding pattern. The overlying sequences are mainly aggrading.

The depocenters of sequences 7.1 and 7.2 are located in the northern part of the study area. The depocenters of the remaining sequences are all located in the northern part of the Dutch sector.

The age of sequence 7.1 is early Middle Miocene to Late Miocene (Table 1A, 1B). The lower boundary is located in NSB 11 from the A/18-1 and the 2/2-I wells, where the succession is most complete. The calcareous nanofossils indicate that the boundary occurs in NN 5–6. In most other wells, the boundary is marked by a hiatus. In Mona-1, dinoflagellates indicate early Serravallian for the basal strata, while foraminifera refer the boundary to NSB 12 and NSP 12. These two zones are not concurrent according to King (1989), and the planktic foraminifera dating is considered the more reliable. On balance, the results suggest that the lower boundary is located in the lower Middle Miocene (close to the Langhian-Serravallian boundary). The maximum flooding surface occurs in NN 7–8 in one well; foraminifera datings mainly indicate NSB 12.

The age of sequences 7.2, 7.3 and 7.4 is Pliocene (Tables 1A, 1B). The lower boundaries are located in NSB 13b, NSB14 and NSB 15, respectively. The age of sequences 7.5 and 7.6 is Quaternary. Both sequences are located in the NSB 16b zone.

### BASIN EVOLUTION

#### Position of Depocenters and Source of Sediments

The isopach maps of the seven major sequence stratigraphic units show that the position of the depocenters changed several times during Cenozoic time (Fig. 22). The depocenters are generally located marginally, probably adjoining source areas, and it appears that the moves are caused by changes in the direction of sediment transport and in the source of sediments. Two major shifts in the direction of sediment transport are recognized, one during Early Eocene time and another at the Eocene-Oligocene boundary.

Late Paleocene and Early Eocene sediment transport was from the north or northeast. During Early Eocene time sediment input gradually shifted to the west. The shift is possibly reflected in the Danish Lillebælt Clay Formation, where a shift from mixed to smectite-dominated clays is recognized between the early Middle Eocene L4 and L5 beds (Heilmann-Clausen et al., 1985). Sediment transport from the west prevailed during the remaining part of the Eocene age.

The Eocene-Oligocene boundary marks an important change in the depositional pattern of the North Sea. The western source of sediment in unit 3 was replaced by a mainly northeastern source in unit 4 (Fig. 22). The shift is paralleled by a change in the type of sediment. The Upper Paleocene and Eocene deposits are dominated by very fine-grained clays and marls. From the base of the Oligocene Series the amount of mica, silt and sand increases significantly. The Middle Miocene to Recent deposits represent the latest infilling phase of the Cenozoic Basin, and the depocenter is located centrally in the North Sea.

The change in depositional pattern between units 3 and 4 is most probably caused by rising of Scandinavia and the adjacent areas of the North Sea. This event is generally considered to be of late Cenozoic age (e.g., Doré, 1992; Michelsen and Nielsen, 1993; Jensen and Schmidt, 1993); our results suggest that it was initiated close to the Eocene-Oligocene boundary.

### Subsidence rate

Rates of subsidence and sedimentation are difficult to estimate from the present data. Mapping shows that approximately half of the Cenozoic section in the central part of the North Sea belongs to unit 7 (Middle Miocene to Recent). The overall aggradational to progradational seismic reflection pattern of this unit, which downlaps to the west, indicates a pronounced subsidence rate in the central North Sea (Fig. 4). We therefore propose that both sediment supply and subsidence rate increased during earliest Middle Miocene time. This increase may be related to a higher rate of exposure in the Scandinavian region.

### SEA-LEVEL CHANGES IN THE EASTERN NORTH SEA

A tentative relative sea-level curve for the eastern North Sea Basin is constructed on the basis of the sequence stratigraphical and palaeoecological interpretations (Fig. 23).

Biostratigraphic correlations allow a detailed comparison with the Danish onshore area and lithological and palaeoecological data from onshore outcrops and cored boreholes provide important additional information.

The first relative sea-level curve of the Cenozoic North Sea was proposed by Vail et al. (1977a,b). A seismic stratigraphic study by Stewart (1987) provides a comparison with interpreted early Paleogene sea-level changes of the western North Sea.

#### Interpretation of Sea-Level Changes

Sedimentation of Danian limestones was followed by a regional regression (Gramann and Kockel, 1988). Deep erosion of Danian limestones is known from Danish localities and from the North Sea, where erosional valleys have been observed on the Ringkøbing-Fyn High. The sea flooded the region at the beginning of Late Paleocene time.

Sequence 1.1 (Late Paleocene) shows a lithological succession similar to the contemporaneous onshore Danish deposits. These are interpreted to reflect gradually deeper and more sediment-starved conditions (Heilmann-Clausen, 1985). Deposits equivalent to the lower part of sequence 1.1 have a very restricted distribution in Northwest Europe; they are mainly confined to onshore Denmark and the North Sea. The Danish deposits (Lellinge Greensand and Kerteminde Marl) apparently represent lowstand and transgressive conditions. A pronounced onlap into northwest European subbasins is witnessed by the wide distribution of deposits equivalent to the upper part of sequence 1.1 (e.g., the Thanet Sands and the French Thanetien
deposits; see Bignot, 1987; Knox, 1994). The upper part of sequence 1.1 is time-equivalent with the Danish Æbelø and Holmehus Formations (Fig. 23). The Holmehus Formation is a condensed deep water deposit formed in an oxygenated environment. It represents the most pelagic environment of the North Sea Basin during Paleocene deposition.


Sequence 1.2 (Paleocene and earliest Eocene) base is time-equivalent with the boundary between the Danish Holmehus and Ølst Formations (Fig. 23). This boundary is interpreted to coincide with a rapid and major sea-level fall (Heilmann-Clausen, 1995). An unconformity, which is time-equivalent with the lower boundary of sequence 1.2, is well-known throughout the Northwest European Tertiary Basin (e.g., the hiatus between the Thanet Beds and the Woolwich-Reading Beds in England), and it is interpreted as resulting from a major and rapid relative sea-level fall (Knox, 1994, 1996).

Marine deposits coeval with sequence 1.2 have a more confined distribution in the North Sea Basin than those of sequence 1.1. Coastal and non-marine deposits, on the other hand, are more widespread (see Bignot, 1987). This indicates that the relative sea-level rise of sequence 1.2 was smaller than that of sequence 1.1.

A relatively wide geographic distribution of the upper part of sequence 1.2 (Balder Formation) and its onshore equivalents as compared to that of the lower part (Sele Formation) indicates a transgression (Knox and Harland, 1979; Heilmann-Clausen, 1995).

The ash-bearing basal London Clay of East Anglia correlates with the upper part of sequence 1.2 and with the Fur Formation in Denmark. Jolley and Spinner (1991) interpreted two transgressive-regressive cycles in the East Anglian deposits. Stewart (1987) also interpreted two cycles in his time-equivalent sequences (7–9) in the northwestern North Sea. We have not been able to recognize a corresponding subdivision in our study area, which is located more distally and observations from the Danish onshore area are not conclusive on this point.

Sequence 2 (Early Eocene to earliest Middle Eocene) is lithologically similar to the contemporaneous Røsnæs Clay Formation and the lower part of the Lillebælt Clay Formation in Denmark (Fig. 23). Except for northwest Jylland, the lower boundary of the Røsnæs Clay marks an unconformity associated with a regional glauconitic horizon, which is interpreted to reflect a major transgressive event (Heilmann-Clausen, 1995). The Røsnæs Clay and the Lillebælt Clay were deposited in a deep pelagic, sediment-starved environment. Deposits coeval to most of the Røsnæs Clay are markedly transgressive in
Fig. 23.—Stratigraphic scheme of the Cenozoic succession in the eastern North Sea with a correlation to the Danish onshore area. The relative sea-level curve is constructed on the basis of the sequence stratigraphic interpretation and the Danish onshore data. The curve is juxtaposed with the eustatic sea-level curve of Haq et al. (1988). The lithostratigraphic subdivision of the Danish onshore formations follows Buchardt-Larsen and Heilmann-Clausen (1988). The age interpretation is mainly based on the following publications: King (1983, 1989), Thomsen and Heilmann-Clausen (1985), Heilmann-Clausen (1985, 1988), Perch-Nielsen (1971), Thiede et al. (1980), UUleberg (1987), Schnetler and Beyer (1987, 1990), Larsen and Dinesen (1959), Rasmussen (1961), and Kristoffersen (1972). The three unspecified dinoflagellate zones in the lower Ypresian include the W. astra, W. meckelfeldensis and E. ursulae zones. For the informal calcareous nannoplankton zones NP 24a and NP 24b, see the chapter “Chronostratigraphy”. Unpublished information has been used in the interpretation of some of the onshore formations: A minor, but apparently regional, hiatus in the upper part of the Søvind Marl Formation coincides with the base of dinoflagellate subzone.
D12nc and is within calcareous nannoplankton zone NP19/20 (C. Heilmann-Clausen and E. Thomsen, in prep.). The calcareous nannofossils refer the Viborg Formation to NP21 (Mikkelsen, 1975; E. Thomsen, in prep.). The dinoflagellates, including A. diktyoplokus, refer the formation to the D12 zone (Koethe, 1990; C. Heilmann-Clausen, in prep.). The glauconitic basal part of the Branden Clay contains benthic foraminifera (Ulleberg, 1987) which we correlate to NSB 7b, and calcareous nannofossils which we refer to NP 24a. The remaining part of the Branden Clay is correlated to NSB 8a and NP 24b. The stratigraphic relationships between the Vejle Fjord Formation and the Klintinghoved Formation are uncertain. They seem to overlap. A borehole in the type area of the Klintinghoved Formation includes foraminifera of zones NSB 8c and NSB 9 (G. V. Laursen and F. N. Kristoffersen, in prep.). NSB 12b, 12c, and 13 have been identified in the Gram Formation in a borehole near the type locality. The glauconitic basal beds are barren. The underlying Hodde Clay contains a fauna that may be referred to NSB 11 (G. V. Laursen and F. N. Kristoffersen, in prep.).
the London, Belgian and Paris Basins (King, 1981) indicating a relative sea level at least as high as that of sequence 1.1.

The presence of lowstand units and the restricted number of erosional features at the top of sequence 2 suggest a moderate relative drop in sea-level at the transition to sequence 3. The major part of sequence 3 (Middle-Late Eocene) is confined to the western and southern part of the study area (Fig. 11). The sequence is thin or absent in the eastern and northern part. The uppermost section, however, onlaps onto the Ringkøbing-Fyn High and into the Norwegian Basin.

The relative sea-level of sequence 3 is difficult to interpret, but it appears to have been moderately high with an increase in the latest phase. A few onshore data suggest that the more widespread uppermost part might constitute a separate sea-level cycle (shown as a hatched line in our relative sea-level curve in Fig. 23). Well logs from the Norwegian sector also suggest the presence of two cycles (Fig. 12).

Sequence 3 is time-equivalent with the upper part of the Danish Lillebælt Clay and Sovind Marl Formations. The calcareous strata in the topmost Lillebælt Clay and in the Sovind Marl Formations (Thiede et al., 1980) appear to represent the prograding unit. Sequence 3 is thin or absent in the northeastern part of the study area, where thick Oligocene deposits often rest directly on Lower Eocene clays. We are, at present, not able to give a satisfying explanation for this distributional pattern. It is possibly related to late Eocene-early Oligocene uplift in the Scandinavian region.

Major sequence stratigraphic unit 4 (Oligocene) covers a large part of the study area (Fig. 14). The unit is subdivided into four sequences. The sigmoidal pattern of the prograding units of sequences 4.1 and 4.2 is interpreted to reflect increasing accommodation space (Fig. 15A,B), and the sequences apparently represent progressively higher sea levels. The oblique progradation pattern of sequences 4.3 and 4.4, on the other hand, indicates almost constant accommodation space. This suggests that the sea during the deposition of these two sequences reached approximately the same maximum level as it did during the deposition of sequence 4.2.

Truncation of reflectors and valley erosion at the top of sequence 4.1 and a small downward shift of onlap suggest a moderate sea-level drop at the transition between sequences 4.1 and 4.2.

Several observations suggest a considerable sea-level drop between sequences 4.2 and 4.3. The lower part of sequence 4.1 is time-equivalent with the Danish Viborg Formation, while the upper part of sequence 4.3 is equivalent with the Branden Clay Formation. Thus, most of sequence 4.1, all of sequence 4.2 and the lower part of sequence 4.3 fall in the hiatus between the Viborg Formation and the Branden Clay (Fig. 23). A similar hiatus is also present on the offshore part of the Ringkøbing-Fyn High. Sequence 4.3 is, furthermore, associated with a clear downward shift in onlaps (Fig. 15). Altogether, it appears that relative sea level dropped significantly between sequences 4.2 and 4.3.

The glauconitic basal beds of the Branden Clay are characterized by a strong influx of planktic foraminifera (Ulleberg, 1987) and calcareous nanofossils, indicating the introduction of oceanic water into the area. This may suggest a pronounced sea-level rise during the deposition of sequence 4.3.

Sequence 4.4 is time-equivalent with the onshore hiatus above the Branden Clay. The dominance of concordant seismic reflections uppermost in sequence 4.3, combined with a moderate downward shift in onlap, indicate a relatively minor sea-level drop at the transition from sequence 4.3 to 4.4.

Major sequence stratigraphic unit 5 (latest Oligocene-Early Miocene) is subdivided into four sequences. The pronounced basinward shift of onlaps at the lower boundary of unit 5 and the more basinward location of the downlap areas as compared to that of unit 4, indicate a major sea-level drop in the latest Oligocene (Figs. 4, 5). A lens-shaped seismic unit interpreted as a lowstand unit is located basinswards of the depocenter of unit 4.

Correlation to Danish localities provides further information on this major Late Oligocene sea-level fall. The onshore Branden Clay is interpreted as an open marine shelf deposit. It partly correlates with sequence 4.3, as described above. The overlying Vejle Fjord Formation was deposited in a much shallower, coastal to lagoonal environment (Fig. 23). The formation includes fresh, unstable weathering products of crystalline rocks, indicating that the source was the old metamorphic rocks of Fennoscandia (Larsen and Dinesen, 1959). It is regarded as the proximal parts of sequence 5.1 and possibly 5.2. The hiatus between the Branden Clay and the Vejle Fjord Formation, therefore, appears to be associated with a major relative sea-level fall, parallelling the shift from sequence 4.4 to 5.1 in the North Sea.

There is no evidence of a significant increase in the rate of elevation of the Scandinavian region during this period. We therefore presume that the sea-level fall was controlled primarily by a change in the eustatic sea level.

Sequences 5.1–5.4 gradually onlap and finally overstep the top of unit 4, indicating an overall increase in relative sea level during this time period (Fig. 5).

Major sequence stratigraphic unit 6 (late Early-early Middle Miocene) is subdivided into three sequences. The depocenters and downlap areas are located basinswards of the underlying unit 5. The apparent absence of erosion at the top of unit 5 indicates a minor sea-level drop prior to deposition of unit 6. A thin seismic unit with distinct low gamma-ray values at the base of sequence 6.1 is interpreted as a lowstand unit.

Cycles that may be compared to the transgressive-regressive cycles of unit 6 have been described from Denmark (Rasmussen, 1961), but the exact chronostratigraphic correlation between the North Sea cycles and the onshore deposits is uncertain. However, the available data suggest that relative sea level reached approximately the same level in each of the three sequences.

The domination of shallow marine and non-marine deposits in the uppermost Oligocene to Middle Miocene onshore formations indicates that overall relative sea level, after the late Oligocene drop, remained low and never reached its former level.

Unit 7 (Middle Miocene to Recent) is subdivided into six sequences indicating repeated sea-level fluctuations. Log patterns suggest that sequence 7.1 consists of more than one cycle, and future work may result in a more detailed subdivision of the unit (Figs. 21, 23). The depocenters are located centrally in the North Sea.
Sequences 7.1 and 7.2 show an aggrading to slightly prograding seismic reflection pattern (Fig. 21), signifying a rapid rise in the relative sea level. Data from time-equivalent Danish Hodde and Gram formations are ambiguous on this point. The planktic foraminifera increase in numbers upwards through the Hodde formation, and they are common in the overlying Gram formation (Kristoffersen, 1972). This indicates an increase in oceanic influence. The benthic faunas, on the other hand, suggest a shallower environment for the Gram formation.

Sequence 7.3 comprises a series of small sigmoid prograding sedimentary bodies, building out westwards across the Central Trough region. This may indicate a slower rise in the relative sea level.

The late Pliocene sequence 7.4 is mainly aggradational comparable to the pattern of sequences 7.1 and 7.2, suggesting an increase in the rise of the relative sea level.

The aggrading to prograding character of unit 7 with a prevalent parallel reflection pattern, indicates a generally high relative sea level (Fig. 4). Combined with a high subsidence rate, this may suggest a general lowering of the eustatic sea level.

Comparison with the Haq et al. (1988) Sea-Level Curve

The relative sea-level changes presented above are broadly comparable with the major trends in the eustatic sea-level curve of Haq et al. (1988).

The Late Paleocene-Early Eocene curves show several corresponding trends, although the precise timing of the individual events differs somewhat (Fig. 23). The discrepancies may be due to uncertainty in the chronostratigraphy of biozones and to different definitions of the Paleocene-Eocene boundary (see Chronostratigraphy).

The latest Eocene increase in relative sea level is mainly based on data from the Danish onshore area. The global sea-level curve of Haq et al. (1988) indicates a general decrease during this time interval.

The pronounced Oligocene sea-level fall placed by Haq et al. (1988) in the mid Oligocene is in the present study dated to latest Oligocene. The fall was already identified and dated by Vail et al. (1977a,b). According to Vail et al. (1977b, Fig. 5) the drop seems to be best documented by seismic sections from the North Sea. The sections illustrated by Vail et al. (1977a, Figs. 11 and 12) bear strong resemblance to our section RTD 81-22 (see Fig. 4), and it seems clear that the boundary between our sequences 4.4 and 5.1 corresponds to the boundary identified by Vail et al. (1977a,b) as representing the great Oligocene sea-level drop. This boundary is referred to the top of supercycle TA4 in the curve of Haq et al. (1988). The discrepancies between the two sea-level curves may be explained by differences in the chronostratigraphic interpretation of biostratigraphic data.

Lower Miocene sea-level changes correlate generally well with the Haq et al. (1988) cycles (Fig. 23). However, our proposal for a post-mid Miocene relative sea-level curve is very tentative. The proximal part of the sequences is absent due to erosion, and sediment transport and subsidence rates have yet to be estimated.

CONCLUSIONS

The Cenozoic deposits of the eastern North Sea are subdivided into 21 sequences (Fig. 23). Some of the sequence boundaries are recognized regionally defining seven “major sequence stratigraphic units”.

Units 1–3 comprise mainly distal and hemipelagic sediments, deposited in the lower sublittoral to bathyal zones. Units 4–7 are characterized by an overall prograding pattern. They were deposited in the sublittoral to upper bathyal zones.

Sequence boundaries and maximum flooding surfaces are the only sequence stratigraphic surfaces identified. The sequences of units 4–7 seem to include prograding lowstand deposits, which are probably related to a ramp margin deposition such as that described by Posamentier et al. (1992). Transgressive and lowstand deposits are often difficult to separate, and it has not been possible to identify system tracts in accordance with the definitions of Van Wagoner et al. (1988).

The direction of sediment transport changed several times during Cenozoic time (Fig. 21). Transport from the north during Late Paleocene and Early Eocene time was followed by transport from the west during the middle and Late Eocene.

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INTRODUCTION

The Tertiary deposits in Belgium are marine shelf to coastal deposits formed in the southern part of the North Sea Basin. Lithologically they vary from calcareous deposits at the beginning of the Paleogene, almost indistinguishable from the underlying Maastrichtian chalks, to marls, clays and sands towards the top. In northern Belgium, these deposits reach thicknesses of several hundreds of meters. Stratigraphically they cover almost the whole Tertiary, albeit with many important hiatus intervals.

The stratigraphy in the region was established quite firmly from outcrops, already in the former century. International stage names such as Ypresian and Rupelian are derived from well studied outcrop areas in this region, and several more regionally used stage names were defined in Belgium such as Montian, Landenian, Bruxellian, Ledian, Wemmelian and Tongrian. As these chronostratigraphic names have become obsolete, the names of their type localities are now reserved for lithostratigraphic units: groups, formations and members (Marechal, 1991). In fact this is even more appropriate as these stratigraphic units were originally individualized on the basis of lithological characteristics and much less on grounds of stratigraphically significant fossils.

The stratigraphic ordering of the different Tertiary strata has traditionally been based on a sedimentological interpretation of the alternation of marine and continental strata (Gullentops et al., 1988) for which even a theoretical model was elaborated by Rutot (1883). This alternation was believed to originate from the repetition of transgression-regression cycles. Already in Rutot’s time, it was obvious from observations that superimposed on the major transgression-regression cycles were minor oscillations, but their importance was minimised, and they were described as ‘récurrences’. Obviously, the observed transgressions and regressions are only the recorded expression of a causal mechanism, which the geologist wants to unravel. The basic driving forces for regressions and transgressions as understood now are: tectono- or glacioeustasy, local tectonics and the rate of sediment supply. Obviously both last factors are often related to each other. By comparing particular cycles and their boundaries to a worldwide stratigraphic inventory chart of cycles as observed worldwide in tectonically independent basins (Haq et al., 1987), a eustatic driving force can be identified. On the other hand, a regional stratigraphic analysis like the one presented here is a contribution to the make-up of such a chart and the validation of the established chart.

The present study reviews all available stratigraphic and sedimentological information and interprets it in sequence stratigraphic terms. This stratigraphic approach allows a better understanding of the local tectonic history as was demonstrated by Demyttenaere (1989) in his analysis of the post-Paleozoic geological history of northeastern Belgium.

DANIAN TO EARLY THANETIAN SECTIONS

Mons Basin, Southwest Belgium

In the southern part of Belgium, the Tertiary deposits are generally very thin except in the Mons Basin. The Mons Basin is a very local subsidence basin, originating from the dissolution of anhydrite in the underlying Lower Carboniferous period (Delmer, 1977). The early Tertiary sedimentation in the Mons Basin was in part controlled by tectonic activity (Dupuis and Vandycke, 1989).

The basal part of the Tertiary in the Mons Basin consists of a porous marine calcarenite, the ‘Tuffeau de Ciply’, of which the lithology is almost undistinguishable from the underlying Cretaceous chalks. This calcarenite unit has a transgressive geometry. Its base, known as the ‘Poudingue de la Malogne’, consists of small phosphatic pebbles and fossil fragments, both reworked from the underlying Cretaceous sediments. The base of the pebble horizon is erosive (Marlière, 1964).

The marine fossiliferous Mons Limestones (Calcaire de Mons), stratotype of the former Montian stage, is only locally known from boreholes. Although its geometric position on top of the Ciply Calcarenites has not really been proved yet, Marlière (1957, 1964) recognised their vertical succession on the basis of ostracod data. Other calcareous facies have been reported (e.g., these of Cuesmes, Ghlin and Obourg, including the Mons Limestone), but the number of observations is too limited to be useful in a sequence stratigraphic context (Dupuis, pers. commun.).

The ‘Tuffeau de Ciply’ is a shallow marine deposit, although its foraminiferal content indicates a deeper environment than
the littoral Mons Limestone (Moorkens, 1982). Marlèire (1957) suggested on lithological grounds, that the top of the Montian limestones represents the transition towards a lagoonal environment. Godfriaux and Marlèire (1971) stressed that neither sedimentological nor faunistic evidence indicated sedimentary breaks in the marine strata. Therefore, these marine calcareous facies have to be considered as a single sedimentary cycle.

The base of the Mons Limestone is reported to belong to calcareous nannoplankton interval NP3-4 and to ostracod zone K3 (Vinken, 1988), the underlying Ciply calcarenite to planktonic foraminiferal zone P1 (Hooyberghs, 1983) and to ostracod zone K1 (Vinken, 1988). Moorkens (1984) reported a Middle Danian age for the start of the transgression of the Ciply Calcarenites, but the fossils in the basal Malogne Gravel are definitely older than the fossils in the Ciply Calcarenites (Marlèire, 1964).

The marine deposits are overlain by lacustrine and continental clays and limestones containing lignites. In a borehole at Hainin, and possibly also in the Ghlin area, two lacustrine sedimentary cycles could be demonstrated by Godfriaux and Robaszynski (1974). These authors, however, did not present a genetic interpretation.

The fossil content of these continental deposits does not very well constrain their stratigraphic position. They fall within pollen zone SP1 (Vinken, 1988). Ostracods do occur but do not allow stratigraphic precision. The mammalian fauna is the oldest Tertiary association in Europe (Godfriaux and Thaler, 1972) and is considered (Vinken, 1988, p. 398) as Middle Paleocene age, which in his stratigraphic summary table (Vinken, 1988, Fig. 267) is listed as Thanetian time. In all other figures of Vinken’s summary volume of IGCP 124, the continental deposits (‘continental Montian’) are placed right on top of the marine deposits and still classified within Danian time (see also Godfriaux and Robaszynski, 1974; Robaszynski, 1978).

Undoubtedly, the traditional view of a marine-continental cycle, has led to this representation, rather than firm paleontological evidence. It is very well possible that a hiatus exists between the marine and the continental strata and that the base of the continental deposits is cutting into the underlying marine strata (Fig. 1). A hiatus exists on top of the continental deposits.

Campine Area, North Belgium

In north Belgium, the base of the Tertiary section, consists of marine porous chalks, the Vroenhoven tuffeau, very similar to the ‘Tuffeau de Cîpy’ in the Mons Basin. The same lithological unit is also found in the south Netherlands where it is called the Houthem tuffeau. Although its lithology is almost indistinguishable from the underlying Maastrichtian chalks, its Danian age is well established even in boreholes from the Campine area (Gullentops, 1954). Recent paleontological and geophysical well log investigations have shown the presence of marine Danian chalks over a large area in northern Belgium (Felder et al., 1985).

In a classical section along the Albert Canal in north Belgium, the base of the Tertiary overlies a Maastrichtian hardground and the lowermost meter of the Danian chalk contains a higher amount of glauconite (P1-Bed in Vinken, 1988). Verbeek (1986) has examined in detail the nannoplankton of the section and concluded that the lower glauconitic chalk falls within NP1, whereas the upper whiter chalks have an NP3 or even NP4 floral content (see also Vinken, 1988). Although the NP1 marker, Biantholithus sparus, was not found, it remains impossible to place the flora in another zone (Verbeek, pers. commun.). Verbeek (1986) also refers to investigations in Denmark and Sweden where a similar hiatus has been found at the NP2 level. Recently, thin K/T boundary clay layers were found in the Dutch Limburg area overlying a fossil hash layer yielding latest Maastrichtian belemnites (Jagt, 1993). Moorkens (1982, Table 3) ranks the top of the Maastrichtian stratotype chalks in the area within the Globotruncanita contusa (or G. mayaroensis or R. varians) zone. Based on data in Felder et al. (1985), it is suggested that the precise biostratigraphic age of the top Maastrichtian strata in the area is somewhat variable, probably due to differential erosion resulting from vertical tectonic movements.

On top of the porous Vroenhoven chalk occurs a marine calcarenite unit known as the Maasmechelen Calcarenite. This calcarenite, which is only known from boreholes, has a much higher sand and clay content than the underlying porous chalks. Unfortunately, the nature of the base of the unit is poorly known due to the lack of cores. The foraminiferal assemblage of the Maasmechelen Calcarenite is intermediate between the assemblages of the underlying chalks and the overlying marine assemblage of the Heers Formation (Moorkens, 1982). Therefore, Moorkens (1982) suggested the possibility of a lateral equivalence of this calcarenite with the continental deposits overlying the ‘Tuffeau de Cîpy’ in the Mons Basin.

The Maasmechelen Calcarenite was considered to belong to planktonic foraminiferal zone P3 and most probably P3A, because of the presence of Morozovella angulata (Hooyberghs, 1983, pers. commun.). Later, contradictory data have been published (Vinken, 1988) with the calcarenite being attributed to foraminiferal zone NPF1b as well as to zone NPF2. The underlying Vroenhoven chalks were reported to belong to zone NPF1a (Vinken, 1988). No calcareous nannofossils have been recorded in the Maasmechelen Calcarenite. However, its position in between the Vroenhoven chalks and the Heers Formation suggests a correlation with the NP4-NP6 interval. It can be concluded that the interpretation of the plankton and the nannoplankton data points to an Early Thanetian age (sensu Haq et al., 1987) for the Maasmechelen Calcarenite.

On top of the Maasmechelen Calcarenite occur continental deposits. Where the Maasmechelen calcarenite is absent, these continental deposits rest directly on the Vroenhoven tuffeau. They consist of red and black clays, lignitic clays and sands, that are rich in brackish water molluscs. Several descriptive terms have been used to name these deposits, ranging from ‘infra-Heersien’ to ‘Couches à Cyrènes du Paléocène du Limbourg’ (Halet, 1932). The most formal stratigraphic name, however, has been the Eysdenian (Stainier, 1931). The regional distribution of the continental deposits is much more limited than the regional occurrence of the underlying marine porous chalk or tuffeau. The contact with the underlying marine chalk, which was observed in the mine shaft at Eysden (Halet, 1932, Fig. 1), is clearly erosive.

Biostratigraphic information on these continental deposits is poor, although according to Vincent (1930), their mollusc fauna is much more related to ‘Heersien’ (Early Thanetian) than to ‘Montien’ (Danian) mollusc faunas. Their pollen content, in-
indicating zone SP1 (Vinken, 1988), does not really constrain the stratigraphic position any further.

**Sequence Stratigraphic Interpretation of Danian-Early Thanetian sections**

The similarity, both in lithology and fossil content of the ‘Tuffeau de Ciply’ and the ‘Vroenhoven Tuffeau’, allows these porous marine chalks to be considered as one unit, although they are now geographically isolated. This isolation results from tectonic uplift, which took place before the deposition of the Heers Formation, as can be derived from the geometry of the strata (Fig. 2).

Geophysical well logs show the porous chalks to have a very homogeneous lithology. In the Mons Basin, the basal conglomerate (Malogne) and the geometric position of the different units point to a transgressive start, whereas the fossil content shows towards the top the shallowing of the basin. Therefore, the porous chalks are considered as an individual sequence, falling within the NP3-NP4 interval.

The underlying glauconitic chalk in the Campine area is interpreted as a transgressive unit because of its marked glauconite content, but it is separated from the previously defined sequence by a an important time hiatus. Therefore it is interpreted as a truncated part of a previous sequence within NP1.

The marine Maasmehelen Calcarenite, being different in lithology and of slightly younger age could well represent an individual sequence. However, its nature is poorly documented. Biostratigraphy places this calcarenite at the very base of the Thanetian.

Contrary to the traditional close linking of the marine and continental deposits into a single Dano-Montian cycle, or the Danian to Middle Paleocene cycle 1 of Vinken (1988), it is suggested that a more important time gap exists between the marine and the continental deposits. Paleontological evidence in the Campine area places the continental deposits well after the beginning of the Thanetian, and in all probability, it can be assumed that the similar continental deposits in the Mons Basin are of the same age. This time gap is in accordance with the stratigraphic table by Cavelier in Schuler et al. (1992). The erosive nature of the base of the continental deposits, for which there is evidence in the Campine area and which is equally well possible for geometrical reasons in the Mons Basin, indicates that the continental deposits represent the filling of an erosional depression cut at low sea-level position. Therefore the continental deposits are considered as the sedimentary response to a rising sea level, and they are thought to represent a new sequence. This sequence is very incomplete, probably due to the renewed and substantial uplift of the Brabant Massif in the central part of Belgium.

This sequence stratigraphic interpretation is represented in Figure 3 together with the sea-level controlled sequence chart of Haq et al. (1987). The hiatuses are thought to reflect the epeirogenetic tectonic uplift signature of the Brabant Massif, a Caledonian block linked up with the London Massif, and making repetitive positive and negative vertical movements since Hercynian times. The change of fully marine deposits at the base of the Danian to continental deposits in Early Thanetian time can be understood as a second order sequence caused by the increased tectonic uplift of the Brabant Massif in Central Belgium. The uplift culminated in the complete erosion of the strata in central Belgium before the deposition of the Heers Formation (Fig. 2). It is possible that the distribution of the continental deposits left after this erosion beneath the Heers unconformity, of limited extent in comparison with the underlying marine deposits, reflects the area of deepest incision during the time of low sea level. Only in this area could the later infilling continental deposits be preserved from the erosion.
FIG. 2.—Schematic cross section from the southern Brabant area to the Campine area.
THANETIAN TO EARLY YPRESIAN SECTIONS

Central and North Belgium

In central and north Belgium, the tectonically induced Early Thanetian unconformity is overlain by the Gelinden Marls which generally have a basal glauconitic sand (Orp), obviously transgressive in this particular geometry (Fig. 2). These Orp Sands and the overlying Gelinden Marls constitute the Heers Formation. The erosion resulting from the pre-Heers tectonic uplift is documented by the abundant reworked Cretaceous nanoplankton found in both members of the Heers Formation. The transition from the Orp Sands to the Gelinden Marls is gradual as can be observed in several outcrops.

Compared to the glauconitic Orp Sands, the Gelinden Marls reflect a deepening of the depositional environment. The many undeformed land-derived leaf impressions in the Gelinden Marls show on the one hand the nearness of the coast but on the other hand sufficient water depth to prevent seafloor turbulence. The marls are made up of a series of cycles about one meter thick, with slight variations in clay and fossil content (Gullentops, 1963).

The top of the Gelinden Marls is intensely burrowed and represents a marine abrasion surface (Gullentops, 1963). The eroded nature of this top is further documented by an abrupt shift in well-log response at the top of the marls (Fig. 4). Although the carbonate content slightly increases towards the top, the lithology remains generally very constant. Nevertheless the detailed microfossil and sediment analysis has shown that regressive conditions with only very weak marine influence must have occurred after the transgressive lower part of the Gelinden Marls (Schumacker-Lambry, 1978). The shallow depositional environment explains the poor content in planktonic forams. The available nanoplankton points to a NP6 or NP7 zone (Vincken, 1988) (see note, Fig. 8). The palynological content of the Gelinden Marls is similar to that of the Thanetian II zone of the Paris Basin (Schumacker-Lambry and Roche, 1973).

In the outcrop area, the marine abrasion surface is overlain by a porous siliceous carbonate sediment called the ‘Tuffeau de Lincent’. The base of this tuffeau is enriched in glauconite and contains some pebbles. The detrital quartz content is below 10%, and most of the sediment is made up of the calcareous remains of foraminiferids, molluscs and echinids. The high porosity, up to 25%, is due to the dissolution of siliceous sponges (De Geyter, 1981). As the sediments overlie an abrasion surface, they obviously must be transgressive. Their transgressive nature is confirmed by the presence of pebbles and glauconite and also by the more widespread extension of the deposits as compared to the underlying Heers Formation (Figs. 2, 5, 6). They mark the start of a new cycle which corresponds to the Landen Group (formerly Landenian). This Landen cycle is another classical example of a marine sediment package overlain by continental deposits.

Although hard to demonstrate, it is possible that a tectonic uplift may have occurred in the time span between the deposition of the Heers cycle and the overlying new cycle initiated by the deposition of the Lincent Tuffeau.

In the subsurface of the Campine area, the cycle starts very often with a few meters of calcareous clay, described by Gulink and Hacquaert (1954, p. 463) as a ‘tuffaceous glauconitic clay with preserved sponge spicules’. When present, this basal layer is easily recognised on geophysical well logs (Laga and Vandormael, 1990, annotated on gamma-ray logs as calcareous Landen Clay on Fig. 2). The basal layer is overlain by a thick clay unit which shows a characteristic well-log pattern recognised in many wells (Fig. 4). Higher up in the section, siltstones occur, and again porosity tends to develop from dissolved spicules. The fine grained sediment is overlain by a homogeneous glauconitic fine marine sand locally known under the names of Hoegaarden and Racour Sand (= top Landen Sand on Fig. 4). This sand unit occurs both in the Campine subsurface and in the outcrop area where it overlies the ‘Tuffeau de Lincent’ (Figs. 2, 5). The transition to the underlying clays is gradual (De Geyter, 1991), but the boundary with the overlying continental Landen deposits is always sharply expressed. The gradual transition from the underlying clays in the Campine sub-
surface or from the ‘tuffeau’ in the outcrop area towards the overlying top Landen Sand can also be observed systematically on well logs (Fig. 4). Towards the top the glauconite content decreases (De Geyter, 1991) and cross bedding tends to develop as do thin pure clay intercalations, which according to outcrop observations are sometimes ripped up. This points to a shallowing of the water and hence to a regression.

The different units of the Landen cycle described previously, from the Lincent Tuffeau up to the Racour and Hoegaarden Sands, have rich in-situ nannoplankton assemblages belonging to NP8 (Vinken, 1988). Foraminiferal zone P5 has been recorded in the Lincent Tuffeau (Hooyberghs, 1983). However, this identification of the zone P5 is based on the presence of *Globorotalia* or *Acarinina soldadoensis* which is now known to occur in zones P4b and P4c.

**Mons Basin**

In the Mons Basin, the up to 40 m-thick Bertaimont Formation, also known as the Paternostre Formation, unconformably overlies Senonian chalks, Maastrichtian chalks, Danian limestones and the Early Thanetian continental deposits (Hainin...
FIG. 5.—Schematic cross section from the Belgian Coast (Knokke-Zeebrugge area) over the Brussels area to Tongeren.
FIG. 6.—Schematic cross section from the Brussels area (Halle-Anderlecht) through the Waasland and the Antwerp area to the boundary with the Netherlands.
Formation). This Bertaimont Formation is composed of three lithological units (Fig. 1) which all three contain to a more or lesser degree pebbles, lime and glauconite. Its transgressive nature was clearly demonstrated, the different steps in the transgression being characterised by pebble beds (Marlière, 1969). This formation is overlain by glauconitic and clayey sands containing silicified horizons (Tuffeau d’Angre et de Chercq) which in central and north Belgium contain pebbles at the base. The fine glauconitic sands at the top of the series described in central and north Belgium (Racour or Hoegaarden Sands) also occur in the Mons Basin where they were originally described as the Grandglise Sands and Sandstones. In the Mons Basin, as in central Belgium, the transition from ‘tuffeau’ to sands is gradual (Maréchal and Laga, 1988).

Traditionally the lower unit, the Bertaimont or Paternostre Formation, is correlated with the Heersian of central and north Belgium whereas the ‘Tuffeau d’Angre et de Chercq’ together with the overlying Grandglise Sands are correlated with the marine Landen cycle from central and north Belgium (Marlière, 1969; Robaszynski, 1978). The nannoplankton assemblage in the ‘Tuffeau d’Angre et de Chercq’ and the Grandglise Sands is indeed referred to zone NP8 (De Coninck et al., 1981). However, also in the lower part of the Bertaimont Formation, nannoplankton zone NP8 was identified (Steurbaut, unpubl. data, quoted in Maréchal and Laga, 1988) preventing a correlation with the Heers Formation. Probably the relatively thick, gravel-bearing Bertaimont Formation represents a local tectonically influenced sediment package, as the Mons Basin was still actively subsiding at that time as documented by the lateral shift of the depocenters (Marlière, 1969).

In the Erquelinnes area, 20 km SE of Mons, an individual clayey sand layer was identified, the Bois Gilles Sand, above the fine glauconitic marine sands of the Landen cycle (Grandglise) and below the continental Landen deposits (De Coninck et al., 1981). A complete paleontological investigation has shown that the Bois Gilles Sands are very shallow marine sands located at the transition of Deflandrea speciosa and Apectodinium homomorphum dinoflagellate zones and within NP9. Typical NP9 nannoplankton species were recorded together with one reworked NP8 species. In the North Sea Basin, the appearance of A. homomorphum is best used as the basis of the A. hyperacantha zone which most zonations situate at the start of NP9 (De Coninck, pers. commun.), although Cavelier and Pomerol (1986) situate this base in the top of NP8. In the Bois Gilles Sands, A. homomorphum appears already together with NP9 nannoplankton. De Coninck et al. (1981) suggest that the gravel separating the Grandglise and the Bois Gilles Sands probably does not represent a major break but only a sedimentary rearrangement during the regression. It should be noted, however, that Dupuis and Steurbaut (1987) have observed at different localities in Haute-Normandie (NW France) two Thanetian sand units with similar facies but different ages namely the Criel Sands belonging to NP9 and the Dieppe Sands to NP8. No dinoflagellate data are available from these two sand units.

**Continental Landen Deposits and the Transition to the Ypresian Clays**

Both in the Mons Basin and in north Belgium, continental deposits (continental Landenian) overlie the marine Landen deposits described above (Gulinck, 1973).

In both areas, the facies are characterised by lignite, lignitic clays and fluviatile sands which point to a NNW oriented transport direction. Vertebrate fossils and upright silicified trunks have been recorded. Silification caused by rising and evaporating silica-rich pore waters formed a widespread massive quartzitic layer at the top. Brackish faunas occur but are subordinate unlike in the more western Flanders area where brackish lagoonal deposits dominate. Field observations have shown that the fluviatile continental deposits deeply erode into the underlying deposits (Leriche, 1928). The eroding nature of the base of the continental deposits is confirmed by seismic records (Demyttenaere, 1989). All authors have always considered the fluviatile facies and the lagoonal facies as part of one palaeogeographical configuration.

The boundary between the marine and fluviatile continental Landen deposits corresponds to the limit between pollen zones SP2 and 3 (Vinken, 1988). The youngest marine unit underlying the fluviatile sands is the Bois-Gilles Sands unit. As discussed previously, these marine sands were deposited in the beginning of the NP9 zone around the appearance of Apectodinium homomorphum (De Coninck et al., 1981).

To the west, in the Flanders area, the continental deposits (Tienen Formation) are characterised by abundant brackish water molluscs intercalated in a complex of mainly lignitic sands and clays (Gulinck, 1973). This facies in fact very much resembles the ‘Sparnacian facies’ (faluns with Cyrena) recognised in NW France. Although Feugueur (1955) has suggested a correlation with the ‘Thanétien of the Ile-de-France’ (Sables de Brachex) based on molluscs and mammalia, Roche (1973, Tableau IV) has shown by sporomorphs that the flora of the lagoonal Landen deposits in Belgium is very similar to the ‘Early Sparnacian’ flora in the clay and lignite deposits of the Paris area. Micropaleontological investigation of these deposits in two cored wells, Knokke and Kallo, (Dupuis et al., 1990) has shown that these deposits belong to the A. hyperacantha zone, including the top of the zone which is straddling the NP9-NP10 boundary (this volume). These authors conclude that the Sparnacian facies, considered to be pro parte Ypresian in age in the Paris Basin, is diachronous from the southern North Sea Basin to the Paris Basin.

Thanetian deposits are lacking along the axis of the Brabant Massif in eastern central Belgium (Gulinck, 1973). But as younger Ypresian clay deposits do cover this area, important tectonic uplift and erosion must have taken place in the time interval between the deposition of the continental Landen deposits and the start of the deposition of the Ypresian clays.

Leriche (1928) assumed an uplift of the Artois axis during the deposition of the continental Landen deposits based on the occurrence of the fluviatile facies over the Artois area contrasted to the lagoonal deposits to the north in Flanders and to the south in France. Dupuis et al. (1984) have shown that the Weald-Artois uplift ceased to influence the sediment distribution from the Wetzeliella meckelfeldensis dinoflagellate zone onwards.

In the Knokke well, the plastic clays with brackish molluscs are overlain by an almost 10-m-thick sand unit, containing dinoflagellates, particles of peat debris and some glauconite grains. These characteristics are compatible with a littoral environment as suggested by Geets and De Geyter (1990) based on grain size properties.
FIG. 7.—Sequence identification on well data of Ypresian clays in Belgium using grain-size trend identification within biostratigraphic constraints. The geophysical logs are gamma-ray logs (Vandenberghe et al., 1988b), the grain-size logs are taken from Geets (1988), the clay pit data from Steurbaut (1987). The nannoplankton data are from Steurbaut (1988a), the dinoflagellate data from De Coninck (1980, 1988) and the dinoflagellate correlation data (horizontal
The plastic clays with brackish molluscs were deposited shortly after the first appearance of *Apectodinium parvum* and *homomorphum*, the overlying sands at the time of the *Apectodinium acme*. The presence of numerous fragments of pyritized centric diatoms suggests a correlation with the volcanic ash layers observed in the North Sea Basin (Dupuis et al., 1990). The sands are overlain by a thin clay unit of only a few meters thickness with a particular geophysical log signature when compared to all other logs of the Ypresian Clays (Fig. 7). The clay is slightly more silty than the clays above, contains frequent glauconite and especially degraded volcanic ash particles (King, 1990). In the routine practice of borehole description, this clay is interpreted as the base of the Ieper Group Clays, but obviously the detailed characteristics individualize the unit (unit X of King, 1990). This is all the more true as the clay above starts with a silty horizon containing large glauconites.
Deflandrea oebisfeldensis and from the clays above. It was deposited at the time of the dinoflagellate content of unit X is distinct from the sands below and from the clays above. It was deposited at the time of the Distal landa oebisfeldensis acme (Dupuis et al., 1990) which is normally associated with the D5b zone (Vinken, 1988). Knox (1990) has correlated this acme with the lower part of the nanofossil zone NP10. It was suggested by De Coninck (1991) that the acme probably represents here the very base of the W. astra zone and hence indeed the base of the NP10 zone. The ranging of the D. oebisfeldensis acme zone within NP10 was also concluded by Gradstein et al., (1992). According to Knox (1984) and Schmitz et al., (1996), the first occurrence of W. astra, which defines the base of the W. astra zone, is believed to lie within the middle of nanofossil zone NP10.

Sequence Stratigraphic Interpretation

The Orp Sands and Gelinden Marls (Heers Formation) represent a sequence that starts with transgressive glauconitic sands and ends with a regressive facies at the top of the Gelinden Marl. The section is truncated at the top by marine abrasion. Biostratigraphic control is poor but indicates NP6-NP7 (see note Figure 8).

A second sequence (Figs. 8) is fully developed in the Campine subsurface including, according to the well logs, a relatively thin transgressive unit, with probably a low-stand deposit at the very base (Fig. 4), and a thicker highstand clay deposit, which coarsens upwards into the topmost marine sand. It is possible that the shift towards a sandy grain size for these marine sands might be sufficient to qualify it as a sequence boundary. In the outcrop area, the second cycle starts with a transgressive glauconite bed with scattered pebbles and ends with regressive sands at the top. In the Mons Basin, different transgressive steps can be recorded in the marly and glauconitic sediments, formerly attributed to the lower cycle (Heers). These transgressive sediments are overlain by the same porous limestones (tuffeau) and sands as in the outcrop area of central Belgium. This second sequence falls within NP8 (see note Figure 8).

In the Mons Basin occurs a thin sand layer, the Bois Gilles Sand of Early NP9 age, with a transgressive pebble bed at the base and truncated at the top. Although this sand is observed only at one locality in Belgium and De Coninck et al. (1981) suggested that only a minor sedimentary rearrangement was involved, the present authors think that it may well represent a third sequence similar to a sand found in northern France (Figs. 8, 9).

The widespread and marked erosional base of the continental deposits is thought to have occurred during a sea-level lowstand, helped by local tectonic uplift, and the infilling is thought to represent the result of the next sea-level rise. Initially the incisions are filled with fluvial uplift, and the lagoonal clays show a coarsening upwards trend (Fig. 7) suggesting, indeed, that the continental Landen deposits represent a lowstand systems deposit. The overlying unit X, which shows a fining upwards grain size trend (Fig. 7) represents a transgressive systems tract with a sharp transgressive surface at its base. The sequence is incomplete due to the tectonic uplift preceding the later Ypresian subsidence and transgression. Indeed, as discussed, in the eastern part of the Brabant Massif, mapping shows that Landen deposits were removed before the deposition of the Ieper Group clays (Gulinck, 1965). This tectonic pulse is documented in the Knokke well by the presence of the small flint pebbles at the base of the overlying sequence, which represents the start of the classical Ieper Group clays (Fig. 7).

The biostratigraphic position of this sequence containing the continental Landen and the unit X, corresponds to the Apectodinium hyperacantha zone including the D. oebisfeldensis acme which is probably in the very base of the W. astra zone (i.e., from within NP9 till the base of NP10). It also corresponds to pollen zone SP3.

YPRESIAN TO EARLY LUTETIAN

The Ieper Group Sediments

During Early and Middle Ypresian times, clayey deposits, traditionally called Ieper (Ypres) Clay and roughly corresponding to the London Clay in England, accumulated in Belgium. Lithostratigraphically these clays and sandy clays in Belgium are formally included in the Ieper (= Ypres) Group (Figs. 2, 5, 6).

The lithological variations occurring in the Ypresian sediments, and also in the former ‘Panselian’ sediments were unraveled after detailed paleontological investigations (Steurbaut and Nolf, 1986, Steurbaut, 1987). This work has served as the basis for a formal lithostratigraphy. The nomenclature of the different formations and members in the Ieper Group is represented on Figures 2, 5, and 6.

Grain size analysis (Geets, 1988) and especially well-log analysis (Vandenbergh et al., 1988b) throughout the whole section have shown the succession of different fining and coarsening upwards tracts that could be correlated fairly well over
TERTIARY SEQUENCE STRATIGRAPHY AT THE SOUTHERN BORDER OF THE NORTH SEA BASIN IN BELGIUM

FIG. 9.—Sequence stratigraphic interpretation of the Late Thanetian and of the Ypresian clays in Belgium compared to the standard sequence chart (Haq et al., 1987). Calcareous microfossil biostratigraphy and magnetostratigraphy are taken from Haq et al. (1987). Dinoflagellate stratigraphy is taken from De Coninck (1980, 1988), Vinken (1988), Cavelier and Pomerol (1986), Dupuis and al. (1990). The magnetostratigraphy of the Belgian section is from Ali et al. (1993). Arrows as in Figure 3. Sequence boundary numbers as in Figure 7.
the whole area of occurrence. These grain-size trends are the basis for unraveling the sequences but do not correspond exactly to the lithostratigraphic units in use.

Within the clays, the grain-size couples of fining and subsequently coarsening upwards trends are interpreted as representing, respectively, the transgressive and the highstand sequence tracts.

Sequences are designated as sequence 1–2, sequence 2–3, and so forth, which is referring each time to the cycle between boundaries 1 and 2, 2 and 3. These boundaries are indicated as encircled numbers or as numbers preceded by SB on Figures 9 and 7. It is not advisable for the moment to use the existing lithostratigraphic terminology for the annotation of the sequences. Indeed, the lithostratigraphic units are defined on the basis of the properties of the main mass of the units and less on the characteristics of the boundaries, while sequence boundaries are identified precisely at the turning points of grain size trends.

Transgressive base of the Ieper Group and sequence 1–2

The base of the Ieper Group represents without doubt a new transgressive event after the tectonic uplift at the end of the Thanetian time. This tectonic pulse is also shown in the sudden increase of chlorite, illite and also kaolinite at the base of the clays (Mercier et al., 1985; Mercier and Dupuis, 1988). At the same time, the base of the Ieper Group marks the start of a relatively long period of marked subsidence, as testified by the thick clays deposited during the main part of the Ypresian time interval.

In northern Belgium, the transgressive base falls within the Pseudomasia trinema zone, which is approximately equivalent to the Wetzeliella astra zone. Towards the south, the base of the transgression becomes slightly younger and corresponds already to the beginning of the Wetzeliella meckelfeldensis zone (De Coninck, 1988). These dinoflagellates define the D6 zone, correlated with NP10 in Vinken (1988).

This transgressive event at the base of the Ieper clays, undoubtedly representing a sequence boundary (boundary, SB1 in Fig. 7), corresponds in the Knokke well to the pebble layer above the silty clays with the volcanic glass shards (unit X). The clay immediately above this sequence boundary contains the occurrence of W. lobisca, the first appearance of which is situated at the top of the W. astra zone (De Coninck, pers. commun.).

The basal member of the Ieper Group, just above sequence boundary 1, known as the Mont-Héribu Clay Member, is slightly different, containing more glauconite and a much higher silt content than the overlying clays. Its thickness ranges from a few meters to about 10 to 15 meters. Its grain size is rapidly fining upwards (Geets, 1988) as can also be observed on well logs (Vandenbergh et al., 1988b; Fig. 7).

The lowermost two fining-coarsening upwards cycles are developed mainly in a very heavy clay unit, corresponding approximately to the Orchies (also named St.-Maur) Clay, the upper one also partly in the sandy Roubaix Clay, depending on the position in the sedimentary basin (Fig. 5). There is evidence for the existence of two sequences in that heavy clay interval on the basis of the pronounced trends in grain-size or in gamma-ray response in several boreholes (sequences 1–2 and 2–3 in Fig. 7).

As discussed previously, the base of the first cycle (sequence 1–2) falls within the NP10 interval. The boundary between the first two cycles (boundary 2 in Fig. 7) is stratigraphically documented in the Kallo well as follows. Grain size data, taken from Geets (1988), allow the identification of the cycle boundary 2 around 358 m depth although a grain size jump at 341 m could also be considered as a boundary (Fig. 7). Locating boundary 2 at 358 m is consistent with the FAD’s of several dinoflagellates (Hochuli, pers. commun.). The nanoplankton at 360 m belongs to zone NP11 (Steurbaut, 1988a). Also, the dinoflagellates show the appearance of several new species at about 350 m depth. The first recorded appearance of Dracodium simile (Wetzeliella similis), marking the base of the D7 zone which seems to correlate with the base of zone NP11 (Vinken, 1988; Gradstein et al., 1992), occurs at 345 m depth (De Coninck, 1975, 1980). The Phthanoperidinium crenulatum zone (De Coninck, 1980) starts at about 350 m depth. The first occurrence of Acrarina broedermanni recorded at 358.5 m depth (Willems, 1980, 1988, Plate 10, Fig. 4) is reported to fall within the P7 Morozovella formosa formosa zone (Tomkun and Lutterbacher, 1985). The base of this zone is known to be slightly older than the first occurrence of D. simile and to lie within NP11 (Berggren et al., 1985, Fig. 3). Note, however, that the first appearance of Dracodium simile in the stratigraphic table of Vinken (1988) corresponds to the base of zone D7 and hence in this author’s table to the base of zone NP11 (Fig. 9). Steurbaut estimates in general the maximum clay thickness belonging to NP10 at 18 m (Steurbaut, 1988a). In the North Sea Basin, the first major planktonic influx of foraminiferids and nanofossils is known to coincide with the base of the division B1 (sensu King, 1981). This base of division B1 corresponds to the base of magnetochron C24BN (Ali et al., 1993, Fig. 15). In the Kallo well, this bio-event, which coincides with the base of the benthonic foraminifera association IV of Willems (1982), lies at about 335 m depth and therefore clearly above the boundary 2 (see Tielt and Knokke wells, Fig. 7; see also Fig. 11 in King, 1988). This means that the boundary 2 and the sequence 1–2 are to be situated in the magnetochron C24BR (Fig. 9). This conclusion is in accordance with the discussion of the nanoplankton and the dinoflagellate microfossils in the Kallo well from which it can be concluded that sequence boundary 2 is situated in the lower part of NP11 (Fig. 9). This is however in apparent contradiction with the level of the recorded presence of Acrarina broedermanni in the Kallo well which would require, accepting the relative position of the zones P7 and NP11 and the magnetostratigraphy as given in the Haq et al. (1987) stratigraphic chart, a position of boundary 2 in the middle part of NP11 and in chron C24BN.

Sequences 2–3 and 3–4

Sequence 2–3 is fully developed within NP11 as the boundary between NP11 and NP12 was found just below an easily recognisable silty and fossiliferous layer, outcropping in the basal part of the clay pit Marke and which could be correlated with a turn around in the grain size trend interpreted as the start of already sequence 4–5 (boundary 4 in Fig. 7).

Sequences 1–2 and 2–3 are systematically found on well logs and are confirmed by grain-size analysis in several boreholes (Fig. 7). The probable position of boundary 3 in the Kallo well,
as shown on Figure 7 corresponds to an oyster layer. This oyster layer occurs at a turning point of the grain-size trend and has also been recognised in the Kester and Mont-Panisel boreholes. It should be noted on Figure 7 that the dinoflagellate correlation number one between the Knokke and Kallo wells (Hochuli, pers. commun.) occurs around boundary 3, however in Knokke above it and in Kallo below it. In the Tielt well, for which only a grain-size log is available, it is tempting to put the boundary 3 at the thin glauconite bed described in the cores of the borehole but the grain size trend seems to locate the boundary slightly deeper (Fig. 7). According to Steurbaut (1988a), the glauconite bed falls within his nannoplankton zone IIIa1, which means approaching the top of NP11. Presently, the base of the NP12 zone is put slightly higher, just below boundary 4 (Fig. 7). The sequence 3–4 is lithologically more differentiated than the underlying sequences (see e.g., the Gistel and Marke wells in Fig. 7). This sequence consists of alternating clays, fine sands and silts in which several glauconite beds can be recognised. The precise meaning of the glauconitic layer in the Tielt well, as well as of the glauconite layer just under the base of the Marke clay pit (Steurbaut, 1987), is not yet well understood.

Sequence 2–3 is developed in the Orchies (also known as St. Maur) Clay and partly in the Roubaix (also known as Moen) Sandy Clay, sequence 3–4 entirely in the Roubaix Sandy Clay.

Sequence 4–5

Sequence 4–5 contains the other part of the Roubaix (also known as Moen) Sandy Clay, the heavy Aalbeke Clay and the lower part of the Kortemark Silt (Fig. 5).

As discussed above, sequence 4–5 starts slightly above the NP11-12 boundary, and it is developed fully in the NP12 zone. The best documentation of this cycle is the field evidence of the Marke clay pit and the grain-size analysis of the nearby Marke borehole (Steurbaut, 1987; Fig. 7). The position of the NP11-12 boundary in the Marke clay pit as represented on Figure 7 is situated under the silty fossiliferous layer which contains shell remains coated with glauconite. This glauconitic silty layer corresponds to a turn around in the grain size trend and is therefore interpreted as the transgressive base of the sequence 4–5. Indeed, King (1988) noted in the London Clay that towards the continent glauconite levels change into pebble layers. These levels form the transgressive bases of cyclic coarsening upwards units. The highest sea-level deposits within cycle 4–5 correspond to the base of the Aalbeke lithostratigraphical unit. Phosphate nodules are known to be associated with this clay layer. The end of this sequence is marked by the sudden shift to more sandy deposits. The identification of the stratigraphic position of the sequence 4–5 in the Kallo and Knokke wells is confirmed by dinoflagellate data from Hochuli (pers. commun.; Fig. 7).

Sequence 5–6

The base of sequence 5–6, at least as located in Figure 7, is a fine-grained, laminated sandstone layer of about half a meter thickness filling in an irregular erosive base as can be observed in the Kortemark clay pit. Especially in the fourth and fifth sequences (4–5 and 5–6), grain-size variations apparently increase, and it becomes difficult to decide unambiguously between third- and fourth-order variations as the time control is not fine enough. Therefore on Figure 7, boundary 5 is only tentatively indicated, and in addition two systematically occurring higher order cycles are indicated by dashed trend lines. The location of boundary 5 can be justified on the basis of the thickness of the cycles 4–5 and 5–6 which are comparable to the other cycles identified in the Ieper clays.

Dinoflagellate data (Hochuli, pers. commun.) confirm the correlation of sequence 5–6 between the Kallo and Knokke wells. In the Knokke area, certainly the most basinwards location of all Belgian data points, cycle 5–6 has a thickly developed lowstand wedge.

The top of sequence 5–6 is also marked by the sudden appearance of coarse grained sediments.

The precise location of the base of this new sequence within the Egem sands, containing nannofossils of the NP12 zone, however is debatable. In the Egem claypit which can be correlated with the Tielt well (Fig. 7), there occurs a thin clay layer (Egemarkel Clay) under the Egem Sand facies bounded above (SB6' in Fig. 7) and below (SB6 in Fig. 7) by two erosional surfaces, both of which could be qualified as sequence boundaries (Fig. 7). The lower one, however, seems to be more important as it is associated with lots of rolled fish teeth and phosphatic particles. The Egemarkel Clay is also rich in planktonic microfossils. The upper boundary is very similar in appearance to the base of the sand layer in the Kortemark pit forming the base of the fifth sequence (5–6). Although the Egemarkel Clay is also present in the Kallo well (Fig. 7), its limited thickness suggests to us that a higher frequency cycle might be involved rather than a supplementary sequence.

In fact, a close inspection of the log trends in Figure 7 shows the presence of higher frequency oscillations. They probably represent parasequences; also thin lowstand systems wedges may be more common than interpreted here. However, a more detailed interpretation could be more than mere speculation if additional detailed biostratigraphic data become available.

The five sequences in the clays are shown on Figure 9 with their biostratigraphic and magnetostratigraphic constraints and compared to the standard sequence chart of Haq et al. (1987). The number of sequences identified in that time interval using the grain-size trend criterion, corresponds to the number of sequences in the chart of Haq et al. (1987). Some biostratigraphic discrepancies exist, however. This is the case for the SB1 boundary which occurs in NP10 whilst the corresponding 54.5 boundary on the Haq et al. (1987) chart occurs in the NP9 zone. The SB5 and SB6 boundaries occur in the NP12 zone whilst the corresponding 50.5 and 50 boundaries on the Haq et al. (1987) chart are situated in NP13. The same holds for the magnetostratigraphy as recently determined for clay sections in Belgium (Ali et al., 1993). For example, the C22-C23 boundary is identified in Belgium in sediments containing NP12 nannofossils although in the Haq et al. (1987) chart this boundary is placed in the NP13 zone.

The next sequence above SB6, including the Egem sands, will be discussed together with the Panisel lithostratigraphic units.

Paniselian Lithostratigraphic Units

In the traditional Belgian stratigraphy, the Paniselian stage name was reserved for generally glauconite-rich sands and
clays occurring between the Egem Sands (Yd in the traditional stratigraphy) and the Middle Lutetian Lede Sands (Fig. 10). Curiously enough, the Paniselian deposits, named after the Mont-Panisel near Mons in the Mons Basin, are found only to the west of the Zenne valley whilst to the east the Brussels Sand facies (formerly Bruxellien) occurs between Ypresian clays and the same Lede Sands (Gulinck and Hacquaert, 1954; Hotyat-Mayne, 1959). Although it was shown by Steurbaut and Nolf (1986) that some Panisel Sands, including those in the type locality, were in fact equivalent in time to the Egem Sands in their type locality, a model relating the Panisel facies to the Brussels Sands facies is still lacking.

Traditionally, two cycles have been recognized in the ‘Panisélien’, annotated on the Belgian geological maps as P1 and P2 (Fig. 5). The lower one (P1) comprises an occasional basal pebble layer (P1a), a lower glauconite sand (P1b), overlain by a thicker glauconitic clay deposit namely the Pittem Sandy Clay (P1c) and a top regressive sand, Vlierzele Sands (P1d). In these top marine sediments (Vlierzele Sands), there occurs continental deposits, the Aalterbrugge Complex with silicified wood and lignites. The upper cycle consists of coarse glauconite-rich sands with many shells and shell beds, namely the Aalter Sands (P2). At the base of the lower cycle, there occurs a few-meter-thick heavy clay layer, the Merelbeke Clay (P1m). This clay layer probably develops at the expense of the lower part of the Pittem Clay (Fig. 5). In the type locality area (Egem, Pittem), the base of the clay is developed as a cemented sandstone layer, rich in shell remains and particularly rich in large authigenic glauconite grains. The glauconitic sandstone layer occurs also below the Merelbeke Clay (P1m). The sand-clay complex of Kwatrecht, described by De Moor and Geets (1973) in the neighborhood of Gent between the Merelbeke Clay and the Egem Sand, is considered as part of the Egem Sands.

As discussed before, the base of the Egemkapel Clay or the Egem Sands is considered as the boundary of a sequence because of the sudden shift in grain size (Fig. 7) (Geets, 1988). The thickness of the Egem Sand package varies between a few meters in most outcrops and wells to 25 m in the Tielt area (Geets, 1988). The sedimentary features that can be observed in the Egem (Ampe) pit are: fine grained glauconitic sand, with many shells and especially towards the top Nummulite concentrations, tidal clay layers and storm-cut erosion. Obviously, the depositional depth was very shallow, but nevertheless a surprisingly large variety of microfacies is preserved, and therefore no strong sorting action can be assumed. The thickness distribution combined with the sedimentary facies suggest the filling of a broad bay, protected from strong coastal marine sorting currents and wave action. If the Egem Sands were part of a transgressive systems tract, more sorting and a more homogeneous facies would be expected. Grain-size data and geophysical logs (Figs. 7, 11) of the Egem Sands in the area of their thickest development show clearly a coarsening upwards trend, but the grain size becomes almost constant towards the top. Therefore, the Egem Sands are considered as a lowstand wedge deposit, at the onset of a slow rise in sea level.

The transgressive, glauconite, shell-rich sandstone layer (P1b) that covers the Egem Sand represents a sudden flooding of the area, without much erosion, as indicated by the absence of pebbles and marked the start of the transgressive systems tract. Although not much erosion has taken place at the start of the transgression, Islam (1982) identified a slight hiatus, based on dinoflagellates, between the Egem Sands and the base of the clays (Pittem) just above the stone band. He correlates the hiatus to eroded surfaces in the London and Hampshire Basins in England. Laterally, the sandstone layer is probably replaced by a thicker sand package, the Hyon Sands (the Panisel Sands at the Mont Panisel, Dupuis et al., 1988) which underlies the Merelbeke Clay in the Brabant area (Fig. 5). The precise sedimentological and biostratigraphical correlation of these sands with the classical Panisel succession has not yet been resolved.

The base of the Merelbeke Clay is characterised by the appearance of several new dinoflagellate species (De Coninck,
FIG. 11.—Geophysical well logs of the Brugge and Maldegem boreholes in the “Paniselian” stratigraphic interval.

Geophysical well logs of the Merelbeke clay from the Brugge area (Lie Sun Fan and Laga, 1986; De Breuck et al., 1989) show that the transgression quickly reached its maximum in the Merelbeke Clay, after which the basin was filled in with prograding sediments forming the upper part of the Merelbeke Clay and the coarsening upwards Pittem sandy clays (Fig. 11). The relatively high sea level in the Merelbeke Clay is also demonstrated by the influx of planktonic forms, including pteropods and radiolarians.

On Figure 11, the Vlierzele Sands are distinctive both from the underlying Pittem unit and from the overlying Aalter Sands, which constitute the classic second Panisel cycle (P2). Certainly, more data are needed to confirm the grain-size trend, but the Maldegem well log shows a coarsening upward trend. In outcrop, a lower fine-grained sand unit could be differentiated from the more typical cross bedded tidal Vlierzele Sands (Houthuys and Gullentops, 1988a). According to these authors, the Vlierzele Sand facies was formed as a series of parallel tidal ridges, progressing by offlap sedimentation over a quiet, intensively burrowed, sandy sea bottom in a regressive marine environment.

The lower boundary of the Vlierzele cycle is erosive. This is suggested by its geometric position relative to the underlying Panisel units (Fig. 4). The erosive nature, however, was most convincingly demonstrated by Henriet et al. (1989), De Batist et al. (1989) and De Batist (1989) in sparker seismic investigations just offshore of the Belgian coast. These authors demonstrated the presence of a deep erosional cut and interpreted the fill as Vlierzele Sands.

In a few areas in Flanders, continental-type deposits containing lignite and silicified trees (the Aalterbrugge Complex) occur within the top of the Vlierzele Sands, demonstrating the very littoral character of the Vlierzele Sands at the top (Gulinck and Hacquaert, 1954). A section based on a series of closely spaced wells near Gent (De Moor and Geets, 1973) suggests that some erosion might have accompanied the development of the continental complex, as was already suggested by Hacquaert (1939) based on the presence of intraformational pebbles at the lignite level.

The overlying Aalter Sands, formerly considered as the second Panisel cycle (P2), are very rich in glauconite and shells. They have a more limited extension than the lower Panisel cycle (Gulinck, 1965). Several lithologic members have been recognized (Maréchal and Laga, 1988), but a complete sedimentologic study is still lacking; therefore, the exact facies relationships are not yet well understood. In a study of the Aalter Sand section, between the Vlierzele Sands and the overlying Lede Sands, Steurbaut and Nolf (1989) could identify at least three distinct transgressive cycles.

Sequence Stratigraphic Interpretation of the Panisel Stratigraphic Unit

A first sequence is bounded by the base of the Egem Sands and the top of the Pittem Clays (Fig. 10). The Egem Sands have an erosive base and represent the infilling of a bay-like depression, created by erosion. They represent the lowstand prograding wedge of the sequence. The erosive base of the Egemkapel
Clay might be the true sequence base rather than the Egem Sand base. It is possible that two different sequence boundaries are involved, although in the present state of knowledge, the reduced thickness of the Egemkapel Clay would favour an inclusion into the Egem Sand lowstand systems tract. The overlying glauconitic shelly sandstone represents the base of the transgressive systems tract. A main flooding surface occurs in the Merelbeke Clay or in the base of the Pittem Clay, when the Merelbeke facies is not developed. The main part of the Pittem clay is coarsening upwards and is therefore interpreted as the prograding highstand systems tract.

The boundary between NP12 and NP13 approximately lies between the Egem Sands and the Pittem Clay (Steurbaut, 1987). The reported NP13 association at the top of the Egem Sands in the Kallo well (Steurbaut, 1988a) was based on the interpretation of rare fragments of *Tribrachiatus orthostylus* as reworked. Recent investigations however have shown that these fragments represent *in situ* occurrences, and therefore, the top of the Egem Sands should be incorporated still in zone NP12.

The erosive base of the Vlierzele Sands is interpreted as cut during a sea-level lowstand while the sands are interpreted as the lowstand wedge filling the space when sea-level rise resumed slowly. The presence of strong directional tidal currents in the Vlierzele Sands is in accordance with a low sea level focusing the currents in topographic lows which were created by erosion at the time of the lowest sea level. The coarsening upwards grain-size trend is thought to confirm the progradation of sediments during lowstand filling.

The lignitic Aalterbrugge complex is interpreted as indicating that the Vlierzele Sands had almost completely filled the space available and that thereafter continental conditions prevailed. This might have been helped by the ongoing tectonic uplift of the area. De Coninck (1988) has also reported a hiatus between the Vlierzele and the Aalter Sands based on dinoflagellates. Wetland conditions, evidenced by the start of peat growth in the area, were probably also announcing a transgression. A minor transgressive pulse might have eroded the coastal vegetation and transported continental debris to the sea.

The onset of the major flooding brought the glauconite-rich Aalter Sands over the area, representing a transgressive system tract. Geophysical log trends (Fig. 11) seem to suggest also the presence of a highstand system tract. The geographical change in the position of the sedimentation area of the overlying Aalter Sands compared to the Vlierzele Sands suggests that some epeiric sedimentation proceeded the deposition of the overlying Aalter Sands. Howver, the time span of deposition of Brussels Sands is relatively long, as shown by the planktonic foraminiferid associations (the base is correlated with zone P9, the bulk of the sands contains fauna corresponding to P10 and NP14) (Hooyberghs, 1992; Steurbaut, 1988a). With the P9 forams, there occurs nanoplankton of the middle of zone NP14.

Comparing the Brussels Sands to the Panisel units in the west, their geometric configuration is strikingly similar. The deep erosional cut at the base of the Vlierzele Sands in the North Sea and at the base of the Brussels Sand in the west are comparable. It would be hard to imagine that deep cuts, only 100 km apart and both filled with sediments belonging to NP13–14 nanoplankton zones, are not contemporaneous. It is assumed that they were cut during the same period of low sea-level position. Taking into account our analysis of the sequences in the ‘Paniselian’ units, as discussed above, and the P9 biozonation at the very base of the Brussels Sands, sea-level fall involved in the cutting of the bases of the Brussels and the Vlierzele Sand must have occurred late in the P9 zone and in the later part of the NP13 zone, around the NP13-14 boundary.
Traditionally, the fauna of the Aalter Sands has been closely associated with the fauna of the Brussels Sand, although differences were recognised (Gulinck and Hacquaert, 1954). Steurbaut (1988b) distinguished subzones within NP14 and considered the Aalter Sand one subzone older than the Brussels Sands; however, he thought that the top of one member of the Aalter Sands seemed to be as young as the Brussels Sands.

In the Mont Cassel and the Mont de Récollets, just south of the border in northern France, the occurrence of Brussels Sands on top of the Aalter Sands and below the Lede Sands is well known. The former presence of Brussels Sands was also noticed along the north-east coast of Belgium (Depret and Willems, 1983; Vandenberghe et al., 1990). Gulinck and Hacquaert (1954) already put forward the idea of the occurrence of a widespread thin bed of Brussels Sand, with shells and nummulites, in the area west of the Zenne valley. This thin bed was almost completely removed before the deposition of the Lede or Wemmel Sands. In this area normally only ‘Paniselian’ facies occur.

The sections from Mont-des-Récollets (Fig. 12) show in the basal part of the quartz sands, described as the Brussels Sand, a very coarse quartz sand overlying a perforated surface probably representing a new transgressive surface and sequence boundary. We interpret the Brussels Sands on top of Aalter Sands as the remnants of a generally eroded sequence.

This interpretation of the relationship between Brussels Sands and the Panisel units in sequence stratigraphic terms, taking into account the biostratigraphic constraints, is represented in Figure 10. It implies that the sequence boundary between the Aalter sequence and the overlying Brussels Sand sequence of northern France must be present somewhere in the mass of the main body of the Brussels Sands in the Brabant area and in the sandmass filling in the offshore valley cut. The complex sedimentary build up of the Brussels Sand in the Brabant area (Fig. 5) prevents a precise location of this hidden sequence boundary, but at the same time it delivers several candidate boundaries. The offshore fill at the L1-Y5 boundary, in De Batist’s (1989) terminology, or the erosive boundary between Y5 and Y4 could be suitable candidates. According to the interpretation of De Batist (1989), the L1 (Fig. 10) corresponds to the start of the second Panisel cycle, whereas, Y5 together with Y4, the main filling unit, are interpreted as Vlierzele Sands. No direct observations through wells are available, however, and therefore the correlation with the classical Panisel succession on land remains tentative.

**LUTETIAN, BARTONIAN AND PRIABONIAN**

**Lede Sands**

The Lede Sands are transgressive over the underlying deposits of both the Paniselian units and the Brussels Sands (Fig. 5). The base of the sands consists of some pebbles, a concentration of fish teeth, reworked and perforated sandstones of the underlying sand formations and reworked fossils. The basal few meters of the deposit are locally very coarse grained; formerly, a particular stratigraphic stage name was reserved for this coarse basal part, namely the Laekenian. Its top was defined by a second pebble layer recurring above the basal gravel of the sand package. Fobe (1986), however, describes several coarser layers in the Lede Sands. The sands are generally carbonate rich and contain many shells, including *Nummulites variolatus*. There are also several carbonate cemented stonelayers in the sands, some being obviously cemented shell accumulations. The sands never exceed about 10 m in thickness and on well logs no obvious grain-size trend is detectable. On top of the Lede Sands, a basal gravel marks the beginning of the next transgressive cycle.

Biostratigraphic investigations conclude that the nanoplankton present belongs to the NP15 zone and the planktonic foraminifera to the P10 zone (Hooyberghs, 1983). Although the overlying Wemmel Sands still belong to zone NP15, they are distinctly younger than the Lede Sands as they contain planktonic foraminifera belonging to zone P12 (Hooyberghs, 1983). The Lede Sands contain a typical shallow-water Miliolid-rich foraminiferal association, while the Wemmel Sands contain an open marine microfauna (Drooger, 1969). Besides the base of the Wemmel Sands has eroded the underlying Lede Sands in the area of the residual hills in southwestern Flanders and northern France (Gulinck and Hacquaert, 1954). Therefore, the Lede Sands are thought to represent an individual but not fully developed sequence. Its boundaries are tectonically enhanced due to ongoing tectonic uplift of the area which will finally separate the Paris Basin from the Belgian part of the North Sea Basin. Consequently, the basal coarse part of the Lede Sands is thought to represent a relatively long period of coastal conditions during which rising sea level was in balance with the uplifting Brabant Massif axis before rising sea level finally could flood the area.

**Wemmel Sands and Asse Clays**

Tectonic uplift resumed after the deposition of the Lede Sands and continued in the eastern part of Belgium until very late in Eocene time, but in north and northwest Belgium, a period of subsidence started towards the end of Lutetian time. In this subsiding area, alternating fine glauconitic sands and clays have been deposited.

The lowermost unit is the Wemmel Sand. At the base of these sands, a basal gravel layer is composed of reworked fossils and sandstones from earlier Tertiary formations. The sand is fine grained, glauconitic, slightly clayey and contains abundant fossils (i.e., *Nummulites wemmelensis* (*orbignyi*)). The fossil content diminishes away from the type area. To the south of Brussels, the facies becomes somewhat coarser grained, and it is interpreted as a more coastal deposit (Gulinck and Hacquaert, 1954). To the west, the sands become more glauconitic and the bio- and lithofacies change (Schuler et al., 1992). The clay content increases upwards, and both grain-size analysis (Jacobs et al., 1991) and well logs show a fining upward trend in the Wemmel Sands (Fig. 13). The transition to the overlying Asse Clays is gradual and characterized by an increase in clay and glauconite content. Higher up in the clay, both the sand content and the glauconite disappear. Sometimes the base of the clay is particularly rich in authigenic glauconite (Bande Noire). Radiometric age dating of the Bande Noire glaucony gives an age of 41 Ma based on K-Ar and 41.8 Ma based on Rb-Sr (Keppens et al., 1978). As both methods give similar ages, the age is considered to be reliable (Vinken, 1988). This age is slightly younger than the age given in the stratigraphic chart for that sequence (Haq et al., 1987). The fauna of the Wemmel Sands and the sandy parts of the Asse Clay is very similar even to the extent that some authors have considered an occasional coarse
FIG. 12.—Composite section of the Mont-des-Récollets, Cassel NW France (Nolf and Steurbaut, 1990; Steurbaut and Nolf, in prep.).
The grain-size log and the geophysical well logs in the Asse-Ursel Clay show that sediments reach the finest grain size in the lower part of the heavy clay and gradually coarsen upward; at the top of the clay, a marked shift in grain size brings a new sand layer over the clay. This shift in grain size is considered as the base of a new sequence (Fig. 13). Therefore, the Wemmel Sands and the Asse-Ursel Clays are interpreted as an individual sequence. The Wemmel Sands and the glauconitic lower part of the Asse Clay are considered to be a transgressive systems tract, while the coarsening upward sequence is considered to be a highstand systems tract (Fig. 13). The Bande Noire, or more generally the glauconite-rich base of the clay and also the level of the rapid shift of the sandy Asse Clay to the heavy Ursel Clay (Fig. 14) are considered to be major flooding surfaces. De Coninck (1987) reports that the microfossil content, in comparison to the Hampshire Basin and the Paris Basin, suggests a hiatus between the Wemmel and the Asse units.

Detailed nannoplankton studies have shown that the Wemmel Sands and the sandy base of the Asse Clay are within zone NP15 whilst the main body of the clay (Ursel Clay) already belongs to zone NP16 (Steurbaut, 1986). According to Hooyberghs (1983) both the Wemmel and the Asse units belong to planktonic foraminifera zone P12. Poorly preserved pteropods at the base of the Ursel Clay suggest the zone GP11 (King, 1990); although, there is an apparent misfit with the identified nannoplankton, using the correlation table of Vinken (1988).

The biostratigraphical constraints on the Wemmel-Asse-Ursel sequence are shown on Figure 14.

**Kallo Complex (Formations of Maldegem and Zelzate)**

Several fine-grained and glauconitic sand units systematically alternate with clay layers between the Asse Clay and the younger Rupelian Boom Clay. Their succession can easily be recognized in boreholes and could therefore be individually labeled, a (argile) for clay and s (sable) for sand. This succession is described as the Kallo Complex (Gulinck, 1969).

Well-log analysis in the area confirms a consistent log pattern in this complex (Fig. 15). On top of the Wemmel-Asse-Ursel sequence occurs a sequence with Asse or Onderdijke Sands (s1), the Zomergem Clay (a2) and probably the base of the Buisputten Sand (s2) (Figs. 13, 15). As mentioned earlier, the base of the sequence is marked by a sudden grain-size shift which is observed on grain-size analysis logs and on geophysical well logs and which is also reported in borehole descriptions. The transgressive system tract and the highstand tract are interpreted from the grain size trend (Figs. 13, 14). Biostratigraphic control is only available from the Mol borehole where the Onderdale Sand unit contains NP16 nannoplankton. Otherwise, the sand and clay do not contain calcareous microfossils.

This lack of biostratigraphic control is also the case for the overlying sequence, comprising the main body of the Buisputten Sand (s2) and the Onderdijke Clay (a3). The presence of the grain-size trend couple, of fining upwards at the base followed by a coarsening upwards trend in the upper part of the sequence, is also used in this case to identify the sequence (Figs. 13, 14). The NP18 nannoplankton content in the base of the Bassevelde Sands (s3), overlying the Buisputten-Onderdijke sequence, constrains the stratigraphic position of the latter sequence. The top of the Onderdijke (a3) Clay is burrowed and traces of peat occur between the clay and the Bassevelde Sands (Gulinck, 1969). De Coninck (1987) also found a dinoflagellate hiatus between the Onderdijke Clay and the Bassevelde Sand.

The Bassevelde Sands (s3) contain rich nannoplankton associations ranging from NP18 at the base to NP21 at the top (Schuler et al., 1992). The base of the Bassevelde Sands is sharply delineated by the sudden appearance of sand (Jacobs and Sevens, 1988). The well logs in the area systematically show a coarsening upward trend at the base of the sands (Fig. 15) confirmed by the grain-size analysis (Fig. 13). Above this lower coarsening upward trend, a fining upward trend occurs ending in the Watervliet Clay on top of which again more silty sediment occurs (Wintham Silt). The top of the Watervliet Clay is a strongly burrowed surface. Therefore, it is interpreted as the top of a sequence, and the Wintham Silt is thought to form the base of a new sequence (Steurbaut, 1992). At this level, a dinoflagellate hiatus also was reported by De Coninck (1987). Using the available information, the Bassevelde-Watervliet sediment package can be interpreted in two ways. A first interpretation considers the coarsening upward tract at the base as a
lowstand prograding deposit, followed by a transgressive systems tract and a short highstand systems tract at the top of the Watervliet Clay. A second interpretation considers the lower coarsening upward tract as a highstand deposit after a very rapid transgression and hence two sequences would be present in the Bassevelde Sands. The long time span, ranging from NP18 until NP21, favors this second interpretation. In addition, a rapid transgression at the base also can be expected by the combination of a sea-level rise and the resumption of subsidence after an uplift pulse. It should be noticed, however, that the time range of the NP21 zone as compared to the planktonic foraminifera zones, is much more limited on the Haq et al. (1987) time scale than was concluded in a recent study of the Eocene-Oligocene transition in central Italy (Brinkhuis and Biffi, 1993; Fig. 14).

The sequences in the Kallo Complex are shown with their biostratigraphic constraints on Figure 14. From this analysis, it is very probable that the burrowed top of the a3 Onderdijke Clay with the peat remnants is a tectonically enhanced sequence boundary. A similar sequence boundary at the burrowed top of the a4 Watervliet Clay might also reflect a tectonic uplift pulse.

Transition to the Rupelian Boom Clay

The fine-grained Rupelian Boom Clay becomes coarser downward (Fig. 16). At its base in the Waasland area, the development of a very silty base over several meters thickness can be observed in the field (Vandenberghhe and Van Echelpoel, 1987). The silty base (Belsele-Waas Clay) was calibrated to well logs at Reet (Van Echelpoel, 1991; Schuler et al., 1992). The silt is underlain by fine glauconitic sands which show up clearly on resistivity logs. The top of the sand, especially the burrows and shells, have been cemented by phosphates, at least locally. The phosphates are now slightly reworked and occasionally form a thin pebble layer at the base of the Boom Clay in the Waasland area. The sand underlying the Boom Clay, the Ruisbroek Sand, is coarser than the silty base of the Boom Clay, but it is more clayey than the underlying Bassevelde Sands. An interpretation of the grain-size trend in the Wintham Silt and the Ruisbroek Sand needs to distinguish cycles of different orders as established by Steurbaut (1992). The well logs of the Ruisbroek Sand and the Wintham Silt in the type area are apparently very consistent (Fig. 15). Steurbaut (1992) has interpreted the lower Wintham Silt as a lowstand deposit and the Ruisbroek Sands, containing glauconite at the base, as the start.
of a transgressive systems tract. However, an alternative interpretation is possible. The general grain-size trend in the Ruisbroek Sands can be interpreted as a coarsening upward trend reflecting the prograding of the sediment. This trend is disturbed by a higher order relative sea-level rise (Fig. 15). In this interpretation, the Wintham and Ruisbroek sediments are the lowstand deposits and the phosphate bed corresponds to a major transgressive step, causing some wave cutting erosion. This slight transgressive erosion explains the break up of the phosphate and the slight hiatus as observed at that level by Steurbaut (1992). The lowstand nature of the Ruisbroek Sands is further corroborated by the abundant presence of reworked Cretaceous and Jurassic microfossils as compared to the Boom Clay above the transgressive surface phosphate bed and especially by the occurrence of reworked freshwater microfossils which are absent in the overlying Boom Clay. The abundant presence of reworked microfossils in the Ruisbroek Sands as compared to the underlying Bassevelde Sands corroborates a supposed slight tectonic uplift between the deposition of both units (Fig. 14).

The biostratigraphy allows a good stratigraphic positioning of the Wintham-Ruisbroek package. The top of the Wintham Silt is situated in the lower part of NP22 whilst the top meter of the Ruisbroek Sand contains NP23 (Steurbaut, 1992).

**Tongeren Deposits**

A formerly classic stage of Belgian Tertiary stratigraphy was the Tongrian deposits. It perfectly fits the concept of a lower marine part and an upper continental part of the Tongrian stage (Gullentops, 1987). Unfortunately, the typical lithostratigraphic units of both parts are only well known in the Leuven-Tongeren outcrop area in central and east Belgium. To the north, in the subsurface, these lithological units are no longer recognized as such because in that area the time equivalent sediments have another more basinward facies.

The marine Tongeren deposits in the classic outcrop area are made up of two units, a lower clay-rich, very fine-grained, bioturbated Grimmertingen Sand and an upper, fine-grained, glauconitic Neerrepen Sand. From extensive grain-size investigations, Winkelmolen (1972) concludes that the Grimmertingen Sands are shelf sediments, deposited well below wave influence and transported by general and tidal currents. The Neerrepen Sands, on the other hand, are somewhat coarser, with more pronounced bedding, less burrowing and containing tidal structures. The Neerrepen Sands are deposited in a shallow off-shore zone in front of the beach. The geophysical well log of the Sint-Huibrechts-Hern cored bore hole (Laga, 1987) which is representative for the marine Tongeren deposits shows (Fig. 17) a fining upward trend at the base, representing the rapid submergence of the area, followed by a longer trend of coarsening upward grain size, the highstand systems tract, ending in the shallow Neerrepen Sands. The transgressive nature of the base of the marine Tongeren Sands is undisputed as it occurs over a major unconformity (Fig. 5), has a pebble horizon of southern provenance at the base and grain size and shape properties at the base inherited from the local underlying sediments as demonstrated by Winkelmolen (1972). The filling of the basin during the highstand ended in some areas with a tidal regressive deposit at the top (the Valkenburg deposit) (Buurman and Langereer, 1975). To the north of Leuven (Kesselberg), Grimmern...
Fig. 16.—Correlation between Rupelian sections in the type area, the Campine area and the Leuven-Tongeren area. Note: the presence of two third-order sequences in the Rupelian Boom Clay of the type area as shown by the grain-size trend enveloping the high-frequency grain-size variations. Layers annotated S are septaria layers.
Sedimentary characteristics and the fossil content of the green custrine in contrast to the normal marine environment of pe-
lowest Oligocene strata, like similar clays from other areas (e.g., grounds, Porrenga (1968) demonstrates that green clays in the top of the Henis Clay, gullies are present. On geochemical tent shows both brackish-water and also freshwater conditions together as the continental Tongeren deposits. The fossil con-
clays (Glaise verte or Henis Clay) were deposited and grouped sands and marls (Boutersem, Oude Biezen) and green very fine vertebrate remains (the Hoogbutsel horizon).

In the Henis Clay, several sandy horizons occur with tidal gul-
(2054; Calembert and Gulinck, 1954). In the Henis Clay, several sandy horizons occur with tidal gul-
ies (Janssen et al., 1976). Also in the Oude Biezen Marls, on top of the Henis Clay, gullies are present. On geochemical grounds, Porrenga (1968) demonstrates that green clays in the lowest Oligocene strata, like similar clays from other areas (e.g., the ‘illite ferrifé’ type clays) are lagoonal or hypersaline la-
custrine in contrast to the normal marine environment of pe-
loidal glauconite formation. Gullentops (1963) interpretes the sedimentary characteristics and the fossil content of the green Henis Clay as indicative of purely fluviatile influences at the base and brackish-water conditions to even a connection with the sea at the top. The fauna of the overlying Oude-Biezen Marls is indicative of a still greater marine influence (Gullentops, 1956).

From the above description of the sediments, it is clear that after continental conditions with soil formation a new trans-
gression started and installed a coastal-lagoonal geography over the area. On top of these coastal-lagoonal sediments, the marine Berg Sands have been deposited with a flint gravel at the base which contains reworked elements of the Oude-Biezen Sands and Marls. These Berg Sands are traditionally considered as the start of the major Rupelian transgression.

In the area between Leuven and Tienen, a particular facies has developed in the continental Tongeren unit, the Kerkom Sands. A detailed description has already been given by Van den Broeck (1893). Grain-size studies, mineral content and sed-
imentary structures of the Kerkom Sands are generally inter-
preted as fluviatile under tidal influence (Van den Broeck, 1893; De Raaf and Boersma, 1971; Gullentops, 1990). Towards the top of the sands, the fluviatile nature is replaced by a more shore-type facies. The Kerkom Sands form a band in an east-

top of this impregnated Kerkom Sand was eroded almost im-
mediately after the impregnation as shown by the erosional con-
tact of the stained sands with the overlying pure white sands which contain pebbles and blocks of the stained impregnated sands at the base. Such erosion requires that the stained sand was cemented shortly after it was impregnated, probably through evaporation of the lighter fractions as the oil was seep-
ging out of the sands. The eroded and stained surface is bur-
rowed. After erosion, white sands are brought in the area. They are completely burrowed and therefore thought to be at least brackish and probably marine sands. At the base, clay layers of tidal origin are preserved and abundant black flattened flint peb-
bles are found in the area. The erosional contact is irregular and cliffs on the order of more than one meter are preserved in the stained top of the Kerkom Sands. On top of the white sands, the more clayey marine Berg Sands are found with a wave cut erosional surface at the base, containing black flint pebbles and occasionally small pebbles or streaks of stained sand, derived from the flattening of the irregular cliff topography over which the sea transgressed (Fig. 16).

Biostratigraphy and Sequence Stratigraphic Interpretation of the Tongeren Deposits

The marine Tongeren deposits are interpreted as forming one sequence. A basal gravel represents a transgressive surface. The
transgression was rapid, as indicated by the relatively short fin-
ing upward section, the transgressive system tract. The coars-
ening upward section represents the highstand systems tract,
including a facies shift from the Grimbertingen to the Neer-
rep Sands representing an increased rate of relative sea-level
drop. The regression ends with the formation of tidal deposits
(the Valkenburg deposits) and finally continental conditions
are installed as witnessed by the soil on the top of the marine Tong-
eren Sands (Fig. 14).

Martini (1970) reports nannoplankton belonging to zone NP21
in the Grimbertingen Sand. This zonation has been con-
ﬁrmed both from the basal part and the top of the Grimmer-
the spores and pollen content of the Neerrep and the Grim-
bertingen Sands are very similar and belong to the Wetzeliel-
laceae zone W13 (Ludian) in contrast to the Henis Clays in
which the zone W14 is found (Stampian).

In the Leuven-Tienen area, on top of the soil, there occurs a
tetrapod fauna which constitutes one of the ﬁrst assemblages
known after Stehlin’s Grande Coupure, at which time the nature
of the European terrestrial fauna changed dramatically (Vinken,
1988). According to Cavelier and Pomerol (1986), the Grande
coupure corresponds to the P17/P18 boundary, which lies
within zone NP21. According to Pomerol (1973), it also cor-
responds to the boundary between the Eocene fauna of Mon-
martre and the Oligocene fauna of Rozon in the Paris Basin.
The pollen of the Grimbertingen and Neerrep Sands indicate
zone SP6, whereas the Henis Clay pollen association already
belongs to zone SP7 (Vinken, 1988). Nannoplankton investi-
gations in the Oude-Biezen Marl show that these sediments are
attributable to the NP22 zone (Stearbaut, 1986). Based on ﬁsh
otoliths, Gaemers (1984) also correlates the Oude-Biezen Marls
with the Ruisbroek Sands (confusingly named Bassevelde
Sands in his paper). In the Henis Clay and at the base of the
Ruisbroek Sands, the dinoflagellate marker Geridocysta cono-
pea occurs (De Coninck, pers. commun.). This together with
the spore content allows a correlation with the ‘argile verte’
of the Paris Basin which is of Early Stampian age and is developed
under a Sannoisian facies (Châteauneuf, 1980).

The Berg Sands, transgressing over the continental Tongeren
deposits, contain NP23 nannoplankton (Stearbaut, 1986).
Hooyberghs (1983) recognized the planktonic foraminifer zone P17
in the Ruisbroek Sands of the Mol well (which he erroneously
named Berg Sands, although the evidence for this attribution is
poor). The basal silts (Belsele-Waas) in the Waasland area are
attributable to planktonic foraminifer zone P18 (Hooyberghs, 1983)
and the basal layer of the Boom Clay overlying the Berg Sand
between Leuven and Tienen to interval P19/20 (Hooyberghs,
pers. commun.).

Using these different biostratigraphic constraints, the marine
Tongeren deposits, together with the Bassevelde Sands (at least
with the eventual youngest sequence of the Bassevelde Sand),
form the last sequence of Eocene deposition (Fig. 14). Hence,
the low-angular unconformity at the base of the marine Tong-
eren (Fig. 5) continues into the subsurface either between the
two Bassevelde sequences or between the Bassevelde Sands
and the Onderdijk Clay (a3). The latter would be the case if
the Bassevelde Sands are interpreted as only one sequence. To-
wards the east, progressively older units of the Kallo Complex
are subcropping under the unconformity. This unconformity is

observable on seismic sections (Demyttenaere, 1988; De Batist
et al., 1992).

On top of the Tongeren-Bassevelde sequence, the next se-
quence starts with a variety of facies depending on its paleo-
geographical position. The marine Ruisbroek Sands, with the
Wintham Silts at the base, correspond in time with the conti-
nental and transitional-marine Tongeren deposits whilst the
Berg Sands in the Leuven-Tongeren area contain the same
planktonic microfossils as the basal silts of the Boom Clay (Bel-
sole-Waas) in the Waasland area (Figs. 14, 16).

The incision of the base of the Kerkom Sands into the under-
lying marine Tongeren deposits occurs during the lowest
position of the relative sea level. During the late lowstand, when
sea level started to rise slowly, the Wintham Silt and the Ruis-
broek Sands were deposited in the southern marine realm whilst
over the continent, more to the east and the south, lagoonal
deposits and the continental Tongeren deposits, started to form
in response to rising sea level. Simultaneously, the Kerkom ex-
stuary started to ﬁll up (Kerkom estuary mouth sands). The trans-
gressive pulse, eroding the oil impregnated top of the Kerkom
Sands into still preserved cliffs, might correspond to the pulse
in sea-level rise documented by the clayey intercalation in the
Ruisbroek prograding sand (Fig. 16) and the change from Henis
Clay to Oude Biezen Sand Marls. The ravinement by wave
erosion, forming the base of the coastal Berg Sands, is the start
of the transgressive systems tract and corresponds to the phos-
phate bed at the base of the transgressive Boom Clay (Belsele-
Waas member), overlying the Ruisbroek lowstand systems
tract, deeper in the basin. The very sharp base of the Boom
Clay, overlying the Berg Sands, is another major ﬂooding sur-
face documenting the backstepping nature of the transgression.
It probably corresponds to a major grain size shift deeper in the
basin (Fig. 16).

The sequence boundary identiﬁed at the base of the Ruis-
broek Sands, the erosional base of the Kerkom Sands and the
base of the lagoonal deposits of the continental Tongeren de-
posits all correspond to the same sequence boundary at the base
of Rupelian deposits, documented on the standard chart (Haq
et al., 1987; Fig. 14).

The possible time equivalence of the continental Tongeren
deposits with the base of the Rupelian deposits in their type
area was already discussed by Gullentops (1956) and the equiv-
ance was proposed by Batjes (1958), Drooger (1964, 1969),
Roche and Schuler (1980) and pro parte by Gullentops (1990).

Oligocene

Rupelian deposits

As previously discussed, the continental Tongeren deposits
and the marine Ruisbroek Sands are late lowstand deposits at
the beginning of Rupelian deposition. The transgressive surface
at the base of the transgressive systems tract is expressed as a
slightly reworked phosphate bed at the base of a shelf clay
(Boom Clay) in northern deposits and as a pronounced pebble
bed at the wave cut base of a coastal sand (Berg Sand) in the
south. The phosphate enrichment is probably linked to the up-
wellling of nutrient-rich, deeper marine waters during trans-
gression (Föllmi and Garrison, 1991).

The variations of the grain size in the Boom Clay sections of
the Rupelian stratotype area were studied by Vandenberge
The most striking feature of the clay deposit is the banding caused by alternating layers with thicknesses in the order of tens of centimeters, representing variations in organic-matter content, carbonate content but mostly clay and silt content (Fig. 16). Using power-spectral analysis, Van Echelpoel (1991) and Van Echelpoel and Weedon (1990) were able to demonstrate that the grain-size banding was linked to 100 and 41 Ka orbital Milankovitch cycles. However (Fig. 16), grain-size data show another obvious lower frequency trend, that is interpreted as representing two 3rd-order eustatic sequences. The fining upward trend above the transgressive surface, with the phosphate pebbles, is marked. This fining upward leads to a finest grain size occurring in a characteristic pinkish horizon, representing the two 3rd-order eustatic sequences. The abundance of the nannoplankton species *Braarudosphaera biglowi* in these coarse layers points to very shallow water sedimentation at that time. The transgressive fining upward trend ends at a septaria level. The septaria contain an appreciable amount of siderite and also exhibit bioturbation by worms. Above this septaria level, the clay grain size starts again to coarsen upward (Fig. 16). To the east in the Campine area, the top part of the clay is even replaced by a clayey glauconitic sand, the Eigenbilzen Sands.

The two grain-size cycles are interpreted as two eustatic sequences. Their existence can be proven in the subsurface of north Belgium through well-log correlation. Biostratigraphy shows that the two sequences have developed in zone NP23 (Steurbaut, 1992). As discussed above, the Ruisbroek lowstand sands at the base of the oldest sequence developed during NP22 time. No nannoplankton belonging to NP24 has been found in the type area of the Boom Clay, although it was confusingly reported in earlier literature (Bramlette and Wilcox, 1967; Baumann and Roth, 1969). The Boom Clay contains pollen and spores of zone SP7, as do the Berg Sands and the Henis Clay (Vinken, 1988). Planktonic foraminifera of the zones P18, P19, P20 and P21 were identified by Hooijerghs (1983) and the P18/P19 and P20/P21 boundaries could be precisely located in the section (Fig. 16).

Towards east Belgium, the clay section consists of a lower, marly *Nucula comta* Clay or the Klein Spouwen Clay, which is overlain by a fine sand unit, the Kernels Sands, which in turn is overlain by Boom Clay and Eigenbilzen Sands at the top. The Kernels Sands are known to be clayey at the base and the transition to the underlying *Nucula comta* clay is known to be gradual in some cases (Kruissink et al., 1978).

The section in the Leuven area, where Boom Clay is overlying the Berg Sands, can be correlated with the type area (Antwerp-Boom) using a particular septaria horizon (S20 on Fig. 16), the presence of P19/P20 planktonic forams at the very base of the Boom Clay in the Leuven area and the dinoflagellate assemblages in the Boom Clay just overlying the Berg Sands in the Leuven area which correspond to assemblages between S10 and S20 in the Boom Clay of the Rupel area (Stover and Hardenbol, 1993). Both the molluscs (Janssen, 1984) and the fish otoliths (Gaemers, 1985) suggest a correlation of the lowermost part of the Boom Clay overlying the Berg Sands in the Leuven area with the upper part of the *Nucula comta* Clay to the east. No obvious grain-size trends in the Boom Clay of the Leuven area exist and precise biostratigraphic control is lacking.

A dinoflagellate study by Stover and Hardenbol (1993) shows that the base of lowermost Rupelian sequence (base Ruisbroek Sands sensu strictu) occurs at least three dinoflagellate zones above the proposed Eo-Oligocene boundary at Massigno, Italy and one zone above the top of the Priabonian type section.

The Milankovitch-related higher order cycles are not only expressed in gradual grain-size variations but also in a regular alternation of black vegetal land-derived organic matter occurring each time at the top of a silt horizon and the base of the overlying claybed (Vandenbergh, 1978; Vandenbergh and Van Echelpoel, 1987). Obviously, the spectral analysis of the organic-matter content also shows the same orbital cyclicity (Van Echelpoel, 1991). These higher order cycles are interpreted as parasequences. Two interpretation models are possible. Either, parasequence boundaries are located where the change in grain size is largest and the lowstand part is identified (interpretation 1 in Fig. 18), or the parasequence boundaries are located at the maximum of coarse particles and only transgressive and highstand system tracts are present (interpretation 2 in Fig. 18). The relationship between grain-size evolution and organic matter can be well explained in terms of relative sea-level fluctuations (Fig. 18). Indeed, during the transgression, the coastal areas are encroached; consequently, the coastal vegetation is destroyed, and the organic detritus is transported into the basin. This may explain the time lapse between the coarsening of the bottom and the influx of organic detritus into the sediment. When sea level is approaching its highest relative level, in the middle of the heavy clay layers, no more vegetation is destroyed and the influx of land-derived organic detritus in the basin is coming to an end.

**Chattian deposits**

Consolidation characteristics and deep erosion gullies on the top of the Boom Clay in the Southern Bight of the North Sea

### MODEL PARASEQUENCE SECTION

**INTERPRETATION 1.**

- % Core.
- % > 32μm

**INTERPRETATION 2.**

- % Core.
- % > 32μm

![Fig. 18.—Two alternative interpretations of the parasequences in the Boom Clay in terms of relative sea-level fluctuations based on the variation of the grain-size and the organic matter content.](image-url)
(Henriet et al., 1989) and geological estimates of post-Rupelian sediment overloading show that a very considerable amount of Rupelian Boom Clay was eroded before the deposition of the Neogene glauconitic sands overlying the Boom Clay (Schittekat et al., 1983). In the Antwerp area, reworked Chattian foraminifera are found at the base of Neogene sands overlying the Rupelian Clays (Vandenbergh et al., 1988a; Vandenbergh and Laga, 1986). Therefore, the erosion must have taken place early in Chattian time.

To the east, Chattian deposits, namely glauconitic Voort Sands and Veldhoven Clays, are found and in considerable thickness. The western boundary of their present occurrence is one of the Roer Valley Graben faults. On seismic sections, an unconformity exists between the Chattian Voort Sand and the Rupelian Boom Clay, but deeper in the graben, sedimentation might have been going on continuously from Rupelian to Chattian time (Demyttenaere, 1989).

Whereas in the graben area subsidence resumed during Late Oligocene time, apparently the Late Rupelian to Early Chattian time in the rest of the country was marked by an important uplift of the Brabant and Ardennes areas. For the Brabant area, this tectonic uplift is corroborated by an important pre-Chattian erosional event and is probably also shown by the increasing thickness towards the top of the Boom Clay of the silt-clay couples, which are linked to Milankovitch orbital parameters and therefore represent equal time intervals (Fig. 16).

In the Ardennes, a Late Oligocene uplift is suggested by the stratigraphy of the large karst inlets in Paleozoic carbonates of the Condroz area. At the base of the karst infill, marine fine sands are preserved. Although they lack characteristic fossils or minerals, their particular fine grain-size and their geometrical position strongly suggest an equivalence with the marine Tongrian Sands, Late Priaonian in age. Generally, the sands are considered Oligocene age (Calembert and Gulincck, 1954; Gulincck, 1966) and traditionally, in Belgium, the Tongrian was considered as Early Oligocene age. The Boncelles Sands near Liège are even attributed to Chattian time, although the paleontological proof for it is lacking. In the marine transgressive Boncelles Sands, several pulses could be identified (Thorez et al., 1973). Some of these sands are now preserved at the bottom of huge karstic depressions, sometimes up to 100 m deep (Gulincck, 1963; Soyer, 1978). At the top, the karstic depressions are filled with palustrine and fluvialitic sediments. The pollen and spores show these continental deposits to have massively developed since the Middle Miocene (Russo Ermolli, 1991). Hence it is very well possible that the uplift of the Ardennes and the Condroz area, responsible for the lowering of the watertable and the start of the karst development, took place in the Chattian time.

Biostratigraphical information on the Voort Sands shows the presence of planktonic foraminifera zone P22 (Hooyberghs, 1983), NP24-25 nannoplankton zones, Gadidae otolith zone 6 (Vinken, 1988). The absence of P21 places the Voort and Veldhoven deposits at the end of Chattian deposition; although, oto
lith zone 6 represents the base of Chattian (Vinken, 1988). The Voort Sands evolving into the Veldhoven Clays are interpreted as one sequence, but further sedimentological data is lacking to interpret these Chattian sediments in more detail (Fig. 19).

**MOIocene DEPOSITS**

The occurrence of marine Neogene deposits in Belgium is limited to north Belgium. The Late Oligocene tectonic uplift had pushed the sea northwards. In the Roermond graben area in the northwest of the country, Neogene deposits are considerably thicker due to continuing subsidence.

Marine Neogene deposits in north Belgium consist dominantly of glauconitic and often shell-rich sands. The sands are mostly shallow-marine deposits. The lithostratigraphy of the Neogene deposits in Belgium was established by De Meuter and Laga (1976). The geometrical relationships of the different deposits are represented on Figures 2 and 6.

Neogene sedimentation resumes with the deposition of glauconitic sediments. The biostratigraphic positioning of the Neogene sand units in Belgium is difficult because of the very shallow sedimentation conditions and also by the boreal nature of the microfossils which are difficult to correlate with standard biozonation scales from warmer waters.

**Antwerp area**

The earliest deposits in the Antwerp area, the Edegem Sands, have been reported to contain reworked Oligocene planktonic foraminiferids and indigenous forms of zone N4 (Hooyberghs, 1983). Calcareous nannoplankton however indicates the younger zone NN3 (Martini and Müller, 1973) or interval NN2-NN3 (Vinken, 1988). Gadidae otoliths belong to the zone 11 (Vinken, 1988), whilst the benthic forams indicate zone B7 (Vinken, 1988; Doppert et al., 1979). Grain-size data published for the Edegem section in Antwerp show a marked coarsening upward trend (Bastin, 1966).

The Edegem Sands are overlain by Kiel Sands, which rarely contain fossils and only occasionally have a basal gravel layer. Radiometric data in the Neogene glauconitic sands show a clear break between the Kiel Sands and the overlying Antwerp Sands (Keppens and Pasteels, 1982). Therefore, the Kiel Sands are considered as a unit separate from the Antwerp Sands and probably also from the Edegem Sands.

The Antwerp Sands are very glauconitic dark green sands and contain phosphate pebbles at the base as well as higher up in the sands (see also Harsveldt, 1973). Several mollusc beds are present, and the grain size is variable. The sedimentary facies points to very shallow near shore depositional conditions (Bastin, 1966). Burrowing of bivalves has been observed in hard ground type beds within the sands (Dijkstra and Janssen, 1988). Biostratigraphy shows the Antwerp Sands to be slightly younger than the Edegem Sands. Planktonic forams may range from N6 to N9 (Hooyberghs, 1983) or may represent NPF13 and maybe even the base of NPF14 (Vinken, 1988). The nanoplankton both in the Antwerp Sands and in their somewhat clayey lateral equivalent the Zonderschot Sands (Huyghebaert and Nolf, 1979) belongs to zone NN4 (Vinken, 1988). The Gadidae otolith content situates the Antwerp Sands in zone 12. Benthonic foraminiferas associations, although different from the Edegem Sands association, are still within the same zone B7 (Doppert et al., 1979; Vinken, 1988).

Radiometric data for these glauconitic units are discussed in Vinken (1988). The obtained ages are considered to be reliable if the ages measured with potassium-argon and with rubidium-strontium are comparable. The Edegem and Kiel sediments...
contain reworked glauconites. The Antwerp Sands contain authigenic glauconites with ages between 18.5 and 21.5 Ma. Also, the Zonderschot glauconites are thought to give reliable ages of 15.5 Ma.

On top of the Edgem, Kiel, Zonderschot and Antwerp Sands, the Deurne Sands contain top Miocene fossils. Therefore, the top of the Antwerp Sands represents a significant hiatus.

In terms of sequences, the three sand units probably each represent an incomplete sequence. The Edgem Sands have a transgressive pebble layer at the base, and the coarsening upwards grain size suggests a highstand systems tract. The pebble layer at the base of the Kiel Sands, found in one exposure, is considered as an indicator of a new sequence. The abundance of phosphate and the neofomed glauconite in the Antwerp Sands is probably also linked to a new transgression. The sequences are incomplete because tectonic uplift was continuing and competing with the fluctuating sea level. All the sediments have been deposited during the later part of Early Miocene time and are separated from underlying and overlying sediments by rather long hiatusal intervals (Fig. 19).

The Limburg area

The Miocene deposits overlaying the Oligocene sediments in the Limburg area are another example of a classical marine-continental sedimentary and stratigraphic cycle (Fig. 2). A transgressive basal gravel with reworked Oligocene components and phosphate pebbles (Elsloo Gravel) is overlain by the shelly glauconitic Houthalen Sands (Gulinck, 1970). These sands are in turn overlain by continental yellow Genk or white Oprgrimbie Sands containing lignites. Sections of the classical glass sand exploitation pit at Oprgrimbie can be found in Gul-linck (1961), Gullentops (1963, 1972) and Gullentops et al. (1988).

Within the white sands at Oprgrimbie, a ridge and channel system can be observed; the top is cemented into a silcrete (Gullentops et al., 1988), very similar to the quartzites observed in the brown coal pits in the Niederrhein graben between Aachen and Kölén. Presumably, the silcrete formed under subaerial conditions, marking the end of a cycle. Above the silcrete, there exists a lignitic zone of a few meters thickness. Whether the lignite represents a soil (Gullentops, 1972) or an influx of floated wood (Gullentops et al., 1988), in both cases a new rise in the water table and the installment of swamp conditions is indicated by the presence a thick lignite. Therefore, the lignites might represent the base of the next sequence. The overlying sands have a gravelly base and the grain-size analysis in the overlying sand shows a fining upward trend followed by a coarsening upward trend (Gullentops, 1972). This sand unit contains a thinner lignite horizon toward the top, and a leached soil formed on the top of the unit. This is overlain by a pebble layer which is overlain by marine sands. It represents the third sequence identified in the section.

The marine Houthalen Sands which belong to the lower sequence have a benthic foraminiferid association which is comparable to that of the Edgem Sands but different from that of the Antwerp Sands (De Meuter and Laga, 1976; Doppert et al., 1979). The planktonic foraminiferid associations studied by Hooyberghs (1983) are intermediate between these of the Edgem and Antwerp Sands (N5-6) whilst the nannoplankton indicates zones NN2-3 (Vinken, 1988). According to Martini and Müller (1973), the nannoplankton of the Edgem and the Houthalen Sands are almost identical. The otoliths suggest Zone 12, as in the Antwerp Sands.

No biostratigraphic data are available from the lignites in the second sequence. It is logical, however, to correlate these lignites, located at the rim of the Roermond graben, with the maximal northwestward extension of the Rhine graben lignites occurring within biozone NN4 (Hager, 1981). Biostratigraphical data on the third (upper) sequence are totally lacking.

The Miocene sequences of both the Antwerp and the Limburg area are represented in their biostratigraphic context on Figure 19. Although an exact correlation between both areas is not yet possible, the figure shows that during most of the Miocene time, tectonic uplift was keeping the area well above sea level except for the Burdigalian time from which very shallow marine and also continental uplifts are preserved. These approximately contemporaneous sediments document in both areas three sequences.

The Deurne, Diest and Dessel Sands

On geological maps of north Belgium (Tavernier and de Heinzelin, 1962; Gulinck, 1962), the distinctive geometrical nature of the Diest Sands, crossing the general strike direction of the other Tertiary deposits, is obvious.

This feature is caused by the deep erosional base of the Diest Sands (Fig. 2). The erosion before the deposition of the Diest Sands has locally eroded almost 100 m of Boom Clay. Demyttenaere (1988) has mapped the base of the erosion gully and shown its relationship with the subsiding Rhine graben. Gullentops (1963) and Gullentops et al. (1988) have suggested that the opening of the English Channel at that time could be responsible for strongly eroding coast-parallel tidal currents. However, the currents in the Channel today, although indeed strongly winnowing the bottom sediments, are not causing such deep erosion. Channels under North Sea sandbanks, as described by Trentesaux et al. (1993) are paleovalleys cut during glacial base level fall (Liu et al., 1993). Gullentops compares the Diest Sands to the Flemish sandbanks (Gullentops, 1957; Gullentops et al., 1988). It is assumed that deep erosion could only take place when the erosion base for the gully was relatively lowered, either by a sea-level drop or by an uplift of north Belgium or through both. The sands infilling the gully are very glauconitic and occur in large-scale foresets downlapping directly over the erosive base of the gully. Therefore, it is assumed that the transgression into the gully occurred rapidly and that initially during the transgression some erosion continued leaving a coarse basal lag deposit. Once sea level reached its highest level, the sediments infilled the gully progressively towards the northeast. Outside the gully, to the north, the facies of the Deurne Sands was deposited, rich in calcareous fossils in sharp contrast to the almost complete absence of fossils in the Diest Sands. In the deeper parts of the gully, a finer facies, the Dessel Sands occur, which does contain microfossils.

Benthonic foraminifera of the Deurne and Dessel Sands can be attributed to the zone B9 (Doppert et al., 1979; Vinken, 1988). Lagaaïj (1952) based on bryozoa, de Heinzelin (1955) using molluscs and Glibert (1962) using all invertebrates have
Fig. 19.—Biostratigraphy and sequence stratigraphic interpretation of Chattian and Miocene deposits compared with standard sequences (Haq et al., 1987). Arrows as in Figure 3. The correlation of standard planktonic foraminifera and nannoplankton zones with the biostratigraphic data from Vinken (1988) is based on the nannoplankton zones (up to the NN4–5 boundary) and on the stage boundaries for the Middle and Late Miocene interval.
all claimed a Late Miocene rather than a Pliocene age for the Diest and Deurne Sands. This view is still held by Belgian stratigraphers and in Vinken (1988), the Deurne Sands molluscs range in the Late Miocene zone BM21A. Martini and Müller (1973) show a similar range for the nanoplankton of the Deurne and Diest Sands (Late Miocene). De Meuter and Laga (1970) have observed a difference between the coiling ratios of *Globigerina pachyderma* in the Deurne Sands and in the overlying definitely Pliocene sands. The planktonic foraminifera of the Deurne Sands indicate zone NPF15 which can be correlated with interval N16–17 of Blow (Vinken, 1988).

The overlying Kattendijk Sands contain a younger fauna and have a transgressive basal gravel (De Meuter and Laga, 1976) and are therefore belonging to a younger sequence than the Diest-Deurne-Dessel sequence. The stratigraphic position of the Diest sequence is given in Figure 19.

**Pliocene Deposits**

The Pliocene sediments of the Antwerp area can be subdivided on the basis of their mutual geometric relationships and the presence of base gravels (Fig. 6). The faunal content, especially benthic foraminifera (Doppert et al., 1979), molluscs and fish remains (Vinken, 1988) confirm the divisions and allow some time stratigraphical ranking, although microfossils suitable for wide correlation are absent or not yet studied (e.g., dinoflagellates). All sediments are very shallow marine deposits and contain variable proportions of glauconite and molluscs. Each unit probably represents a pulse in relative sea level in a fairly uniformly subsiding area.

The Kattendijk and Kasterlee units have a basal gravel lag. The Kattendijk base gravel contains phosphatic nodules. Both units are correlated with each other based on their geometrical relationship (Gulinck, 1962). The Kasterlee Sands do not contain fossils. The Kattendijk Sands belong to NPF16 (N18–19) plankton foraminiferid zone, the benthic foraminiferid zone B10, the benthic mollusc zone BM21C and the otolith zone 17 (Vinken, 1988; Fig. 20).

The thin but shell-rich Luchtbal Sand unit contains benthic foraminifera associations of the zone B11, benthic mollusc associations of zone BM22A and otoliths of zone 18 (Vinken, 1988; Fig. 20).

The other members of the Lillo Formation (Fig. 6), the clayey Oorderen Sands with pebbles at the base, Kruisschans Sand and the Merksem-Zandvliet Sands (only differing by the carbonate content), have the same benthic foraminifera (B12) but can be further differentiated by the mollusc and fish remains. The Oorderen Sands contain mollusc zone BM22B and otolith zone 19, while the Merksem Sands contain the mollusc zone BM22C (Vinken, 1988). The different sand units were grouped into one Lillo Formation, together with the Luchtbal Sands, because of the apparently gradual transitions without pebbles and the common presence of numerous shells in the whole section. At the base of the Merksem Sands, an occasional gravel was observed (Tavernier and de Heinzelin, 1962). The benthic foraminiferal assemblage, rich in species at the bottom of the formation, becomes gradually poor in species with increasing domination of euryhaline conditions.

The Oorderen Sands, however, are transgressive with respect to the underlying Luchtbal Sands (Fig. 6). The eastern facies, Poederlee Sands, Neeroeteren Sands and Mol Sands (Fig. 2), are probably time equivalent to at least part of the Lillo Formation. Moving eastward, the more estuarine and fluvialite these facies become. The sands contain no stratigraphically use-
ful fossils. The Merksem Sands have been considered the lateral equivalents of the Poederlee Sands (Tavernier and de Heinzelin, 1962), although the sedimentological similarity between the Poederlee and the Kasterlee Sands would suggest a lateral equivalence between the Poederlee Sands and the lower part of the Lillo Formation. Within the Mol Sands, a lignite layer has been analyzed for its flora content which shows a Reuverian age, the regional Late Pliocene stage in the Netherlands (Tavernier, 1954). The Poederlee Sands contain reworked bryozoans from the underlying Pliocene Sands. In the Mol Sands, the petrology of the quartz grains also indicates the arrival of cold continental phases (Tavernier, 1954). Indeed in Belgium and also in the Netherlands, deposits younger than 2.5 Ma (van Staalfuinen et al., 1979) are strongly influenced by glacially related changing climate conditions and marine deposits become the exception.

The different Pliocene lithological units are represented within their biostratigraphic constraints on Figure 20 together with the standard sequences recognised elsewhere.

CONCLUSIONS

The recognition of sequences in outcrops and boreholes in Cenozoic units of Belgium at the southern rim of the North Sea Basin is facilitated by marked changes in lithofacies. The Cenozoic succession in Belgium consists of nearshore sediments deposited in a slowly subsiding basin. With exception of the Ieper Clay and the Boom Clay, most lithological units consist of relatively thin units separated by unconformities. This depositional setting favors the development of low-frequency sequences. The number of sequences identified and their stratigraphic position differs slightly from the global model (Haq et al., 1987) because of the availability of additional data. The stratigraphic position of the sequences identified in Belgium is determined mostly on the basis of calcareous nannofossils and thus well calibrated with the Haq et al. (1987) timescale.

Even though stratigraphic geology had a long tradition in Belgium, the application of sequence stratigraphic concepts has added to the understanding of cyclicity in the depositional record. A comparison of the stratigraphic interpretation in this study with the interpretation described in the IGCP 124 report (Vinken, 1988) demonstrates the progress. More than thirty sequences are identified in the present study of the Belgian Basin. In Vinken (1988, Fig. 35A), the northwest European Tertiary Basin is described by less than half this number of cycles. Detailed sequence stratigraphic analysis has increased our understanding of the effects of tectonic activity in the Artois, Brabant and Ardennes areas as well.

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SEQUENCES AND SYSTEMS TRACTS CALIBRATED BY HIGH-RESOLUTION BIO-CHRONOSTRATIGRAPHY: THE CENTRAL MEDITERRANEAN Plio-Pleistocene Record

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ABSTRACT: Large-scale correlations and sequence stratigraphic analyses have been carried out in the central Mediterranean region, a tectonically active area crossing the extensional margin of the southern Tyrrhenian, the compressional front of the Siculo-Maghrebian Tertiary chain and the North African foreland. The Plio-Pleistocene marine record has been subdivided in sequences and systems tracts on the basis of both original data and correlations. We provide seismic, well-log and outcrop data supporting the occurrence of regional unconformities of constant age, related to glacio-eustatic oscillations. Evidence of transgressive-regressive facies cycles having different orders of duration, major erosional truncations and basin starvation events contributed to the construction of a new sea-level cycle chart based on the available Mediterranean high-resolution biochronology and magnetostratigraphy. We largely used the deep-sea correlative conformities of sequence boundaries in order to improve the age calibration of the cycle chart. The chart, based on a new Plio-Pleistocene time scale, can resolve boundary ages up to 5th-order paracycles based on correlations to the high-frequency oscillations of the deep-sea record. Outcrop evidence of correlations between individual parasequences and 41 ky astronomical and climatic oscillations of the deep-sea record is supported by high-resolution biochronology. A comparison with the Mediterranean Plio-Pleistocene sequence chart confirms the general validity of the Gulf of Mexico cycle chart of Wornardt and Vail (1991), except for minor differences in age and number of 4th-order sequences.

The sequence stratigraphic subdivisions are recognizable even in active sectors where stratigraphic analysis separates the eustatic from the tectonic component; from this perspective, our experience support the regional synchrony of sequences and systems tracts occurring in the studied interval independently of local tectonic factors.

INTRODUCTION

Sequence stratigraphy is a modern methodology for subdividing, correlating and mapping sedimentary rocks, defining units that evolve in response to changes in shelfal accommodation (Vail et al., 1990). Vail et al. (1977) demonstrated that primary reflections follow geologic time lines, including large-scale stratigraphic discontinuities. These reflections had been used for subdividing the rock record in depositional sequences that are in cyclic genetically-related packages bounded by substantially synchronous surfaces. These studies resulted in a set of cycle charts for the Mesozoic and Cenozoic strata, showing the global chronostratigraphic distribution of unconformities that were controlled by changes of sea level (Haq et al., 1985).

A more detailed sea-level cycle chart of the Plio-Pleistocene interval was successively developed by Wornardt and Vail (1991), based on a dataset from the Gulf of Mexico. This chart reported 3rd- and 4th-order sequence cycles, and adopted bioevent ages and biochronozones based on the Berggren et al. (1985) absolute time scale. Since the biostratigraphic resolution in the interval was very high, Wornardt and Vail (1991) also interpreted a number of species abundance/diversity curves to establish more precisely the ages of the sequence stratigraphic surfaces. Haddad and Vail (1992) correlated the relative sea-level oscillations of the Gulf of Mexico with the isotopic curves recorded in some ODP wells in the north Atlantic and south Pacific Oceans. However, they concluded that the Plio-Pleistocene obliquity and precessional cycles of the δ18O records have frequencies too high to be compared with the 3rd- and 4th-order sea-level oscillation curves, even if a low-pass filtered curve of the isotopic record, representing long-term changes in ice volume, may be used to better constrain the ages of sequence stratigraphic surfaces.

Within the research project on Sequence Stratigraphy of European Basins, our research group began a scientific collaboration with the Rice University of Houston in order to (a) apply the sequence stratigraphic techniques to the Plio-Pleistocene rock record of Sicily both on seismic lines and outcrops and (b) to understand to what extent the resulting data could fit with the Gulf of Mexico cycle chart. Joining field studies of the Plio-Pleistocene outcropping successions with a modern integrated plankton biostratigraphic dataset, Catalano et al. (1992a, 1993a) presented the first central Mediterranean chronostratigraphic chart of 3rd- and 4th-order depositional sequences, with reference to the Berggren et al. (1985) time scale.

In this paper, we present an updated chronostratigraphic chart of the depositional sequences of the Plio-Pleistocene marine record of the central Mediterranean, using the descriptive support of outcrop and seismic lines. The chart can be a good predictive reference tool in regional basin analysis, particularly where the biostratigraphy is poorly documented.

Sequence stratigraphic surfaces have been dated and correlated as accurately as 5th-order paracycles, and their absolute ages referred to the more recent time scales provided by Shackleton et al. (1990), Hilgen (1991a) and Cande and Kent (1992). The construction of a sea-level cycle chart for the Mediterranean Plio-Pleistocene units integrated seismic, outcrop, and well-log data with other data from the peri-Tyrrhenian region and Mediterranean ODP and Atlantic DSDP record.

Another important aspect that we aim to point out is the regional synchronity of sequences and systems tracts found in the area, independently of the local tectonics. Correlations, stratigraphic analyses and sequence stratigraphic interpretations have been performed in a setting of tectonically active satellite and foreland basins. Following the indications of recent sequence stratigraphic guide lines (Vail et al., 1990; Posamentier and Allen, 1993), we attempted to separate the effects of eustasy and local factors as components of individual stratigraphic signatures of the basin fill. In this paper, we wish to outline some
examples illustrating that local factors, such as sediment supply and tectonics, mainly control the stacking pattern, basin shapes and thickness of the fill, leaving unaltered the age and occurrence of the sequence stratigraphic surfaces.

**THE PLEISTOCENE RECORD OF THE CENTRAL MEDITERRANEAN Geologic Setting**

Sicily and its surrounding areas are a segment of the peri-Mediterranean late Cenozoic orogenic belt, developing from the Apennines in Italy to the Maghrebides in North Africa across the Siculio-Tunisian platform (Fig. 1). The chain and its submerged extension are located between the Sardo-Balearic block and the Pelagian foreland; part of the belt collapsed with the opening of the central south Tyrrenian sea. The eastern margin of the Sardinia block, Kabilian-Calabrian units and the Sicilian-Maghrebian units are the three structural elements of the chain, geometrically arranged in a thrust pile verging toward the east and southeast (Fig. 1). The late Pliocene-Pleistocene Gela foredeep, present both in offshore southern Sicily and in the mainland subsurface, underlies the thrust front of the chain between Sciacca and Gela (Figs. 1, 2A), where thin-skinned allochthons of mostly Miocene-lower Pleistocene rocks are covered by the late Pleistocene strata. The foreland area is exposed on land in the Hylbean Platform and underlies Plio-Pleistocene successions in offshore southern Sicily (Sicily Channel).

In the study area, three different type of basins formed during the Plio-Pleistocene interval in the frame of both compressional and extensional tectonic events (Fig. 2B): (a) hinterland basins located along the Tyrrenian margin; (b) satellite basins located along the belt in central and southern Sicily, and passively transported above the thrust sheets; and (c) foreland basins, located near the southeastern margin (Sicily Channel).

![Tectonic map of the Central Mediterranean area](image-url)

**Mediterrenean Pliocene and Pleistocene Stratigraphy**

**Biostratigraphy.**—

**Planktonic foraminifera biostratigraphy.**—Several planktonic foraminifera biostratigraphic schemes have been proposed in the last years for the Pliocene and Pleistocene Mediterranean sequences (Colalongo and Sartoni, 1979; Spaak, 1983; Cita 1973, 1975a; among others). In this paper, we use the biostratigraphic scheme of Cita (1975a) emended by Sprovieri (1992) that is commonly adopted for the pelagic sediments of the Mediterranean (Table 1). It includes 6 biozones (from M P1 1 to M P1 6) and 2 sub-zones (corresponding to both M P1 4 and M P1 5) for the Pliocene and 2 biozones for the Pleistocene strata.

**Calcareous nannofossils biostratigraphy.**—Calcareous nannofossil biostratigraphic schemes for the Pliocene and Pleistocene Mediterranean sequences have been proposed by Bukry (1973, 1975), Schmidt (1973), Mueller (1978), Ellis (1979), Raffi and Rio (1979), and Driever (1988). In this paper, we use the more detailed scheme of Raffi and Rio (1979) derived from the study of the sequences of the Leg 13-Site 132; this scheme was emended by Rio et al. (1990) from Leg 107-Site 653A in the Tyrrenian basin. The correlation of this regional scheme (Table 1) is reported together with the standard zonations of Martini (1971) and Okada and Bukry (1980). Several of the stratigraphic markers adopted here are not included in the standard zonations. For instance, the Early Pliocene scheme of Martini (1971) and Okada and Bukry (1980) is based on ceratolitid events which are rare to absent in the Mediterranean region; consequently, a low biostratigraphic resolution could be possible there. The introduction of other biostratigraphic events is necessary in the Mediterranean record to obtain a more detailed biostratigraphic subdivision; for instance, the Pleistocene evolution of the *Gephyrocapsa* complex (Gartner, 1977; Rio, 1982) allows a very detailed biostratigraphic subdivision. In the proposed regional scheme (Table 1), the biostratigraphic resolution is comparable with the standard zonations in the Pliocene interval, and it is far greater in the Pleistocene interval. Only the NN 19 biozone is recognized in the Martini (1971) standard zonation in the early-middle Pleistocene.

**Integrated calcareous plankton biostratigraphy.**—For the purpose of this study, we adopted an integrated biostratigraphic framework in order to better constrain the age of the geological and climatic events occurring in the Pliocene-Pleistocene Mediterranean record. Foraminifera and calcareous plankton zonal schemes allow a stratigraphic resolution of about 0.5 my in the approximately 3.6 my Pliocene time interval. The integration of the two zonal schemes allows a greater resolution, with intervals of about 0.3 my. Based only on planktonic foraminifera, a very low resolution can be obtained in Pleistocene strata, but time intervals of about 0.2 my can be detected by the integration with the calcareous nannofossil biozones.

**Chronostratigraphy.**—

**Pliocene chronostratigraphy.**—The Pliocene three-fold scheme of Ruggieri and Selli (1950) and the two-fold scheme of Berggren et al. (1985) have been recently reviewed and discussed by Cita et al. (in press), who proposed a three-fold subdivision based on well-established stratotype boundaries. This scheme was adopted here. The basal unit (Zanclean) spans the interval between the Pliocene base, defined at the base of the “Trubi”
sequence outcropping at Eraclea Minoa, southern Sicily (Hilgen, 1991a, b), and the extinction level of *Globorotalia puncticulata*. The second unit is the Piacenzian Stage. According to the Piacenzian stratotype section (Barbieri, 1967), the top of this chronostratigraphic unit is within the *Discaster pentaradiatus* Zone. Nevertheless, Barbieri (1967) extended the Piacenzian Stage to include sediments up to the Plio-Pleistocene boundary. Rio et al. (1994) questioned this usage and proposed the introduction of a new stage, named “Gelasian”, to indicate the rock bodies comprised between the top of the Piacenzian stratotype section and the Pliocene-Pleistocene boundary.

**Early Pleistocene chronostratigraphy.**—In this paper, we adopted the scheme proposed by Ruggieri et al. (1984) and Rio et al. (1991). According to these authors, only one stage (Selinuntian) covers all the early Pleistocene chronostratigraphic interval. It can be subdivided into three substages, Santernian, Emilian and Sicilian. The recognition of the boundaries of these substages, both in deep- and shallow-water setting, is essentially based on calcareous nannofossil markers, as reported in Table 1. The chronostratigraphic subdivision of the Middle-Late Pleistocene is left as still undefined.

Table 1 also reports, for comparison, another chronostratigraphic subdivision of the Pleistocene according to the recent proposal of Cita and Castradori (1995).

**Biochronology.**—

Recent magnetobiostratigraphic studies in Pliocene and early Pleistocene Mediterranean successions allowed the correlation of calcareous planktonic biostratigraphic events to the Geomagnetic Polarity Time Scale (Channell et al., 1988; Channell et al., 1990, 1992; Hilgen and Langereis, 1988; Zachariasse et al., 1989; 1990). A biochronologic framework was therefore originally proposed using the time scale of Berggren et al. (1985). More recently, cyclostratigraphic results from both lithological signals (Hilgen, 1991a, b) and planktonic foraminiferal relative abundance fluctuations (Sprovieri, 1993) have allowed a new and more accurate estimation of the absolute age of chronostratigraphic and biostratigraphic boundaries. A direct comparison between these new ages and those proposed by the Geomagnetic Polarity Time Scale of Berggren et al. (1985), extensively used in the literature, is also offered (Table 1). In this paper we adopted the biochronostratigraphic framework of Sprovieri (1993), obtained from new paleomagnetic and cyclostratigraphic data available in the Mediterranean, to take into account the ages of every biostratigraphic event recognized in the studied outcrops and wells, for which magnetostratigraphy and cyclostratigraphy were often not available. The assignment of these ages allowed a precise bio-chronological estimation of several levels. Consequently, we obtained precise absolute ages for both physical surfaces and depositional events, as well as the length of time involved in hiatuses.

**Sequence Stratigraphy**

The regional analyses or correlations of stratal pattern and facies distributions in more than 30 sites led us to recognize in
### Table 1.—Calcereous planktonic biostratigraphic scheme for the Mediterranean Pliocene—Pleistocene succession.

<table>
<thead>
<tr>
<th>Chrono</th>
<th>System</th>
<th>Stage</th>
<th>Substage</th>
<th>Polarity</th>
<th>Epochs</th>
<th>Foraminifera</th>
<th>Nannofossils</th>
<th>Biostratigraphy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Early Pliocene</td>
<td>GILBERT</td>
<td>Middle Pliocene</td>
<td>Pliacenian</td>
<td>Serravallian</td>
<td>Brunhes</td>
<td>Not zoned</td>
<td>Not zoned</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Late Pliocene</td>
<td>MILOMA</td>
<td>CALABRTIAN</td>
<td>IONIAN</td>
<td>Santorinian</td>
<td>Gauss</td>
<td>Not zoned</td>
<td>Not zoned</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Pleistocene</td>
<td>PLEISTOCENE</td>
<td>Early Pleistocene</td>
<td>Zandian</td>
<td></td>
<td>Matuyama</td>
<td>Not zoned</td>
<td>Not zoned</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>Age of bioevents</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(after Sprovieri, 1993)</td>
</tr>
</tbody>
</table>

#### Events

<table>
<thead>
<tr>
<th>Events</th>
<th>Sh/Hg</th>
<th>CK</th>
<th>Cycle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Top Sicilian</td>
<td>0.89</td>
<td>0.89</td>
<td>mid 22</td>
</tr>
<tr>
<td>Gephyrocapsa sp. 3 FO</td>
<td>0.99</td>
<td>0.98</td>
<td>mid 27</td>
</tr>
<tr>
<td>Gr. truncatulinoides excelsa FO</td>
<td>1.19</td>
<td>1.17</td>
<td>base 35</td>
</tr>
<tr>
<td>Base Small Gephyrocapsa</td>
<td>1.25</td>
<td>1.22</td>
<td>mid 37</td>
</tr>
<tr>
<td>Base Large Gephyrocapsa</td>
<td>1.50</td>
<td>1.47</td>
<td>mid 49</td>
</tr>
<tr>
<td>C. macintyrei LO</td>
<td>1.62</td>
<td>1.59</td>
<td>mid 55</td>
</tr>
<tr>
<td>G. oceanica s.l. FO</td>
<td>1.75</td>
<td>1.72</td>
<td>mid 61</td>
</tr>
<tr>
<td>N. pachyderma left FCO</td>
<td>1.81</td>
<td>1.79</td>
<td>mid 64</td>
</tr>
<tr>
<td>Base Pleistocene</td>
<td>1.83</td>
<td>1.81</td>
<td>mid 65</td>
</tr>
<tr>
<td>D. brouweri LO</td>
<td>1.99</td>
<td>2.02</td>
<td>top 73</td>
</tr>
<tr>
<td>Gr. truncatulinoides FO</td>
<td>2.07</td>
<td>2.10</td>
<td>top 77</td>
</tr>
<tr>
<td>Gr. inflata FO</td>
<td>2.13</td>
<td>2.17</td>
<td>base 80</td>
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<tr>
<td>Gr. bononiensis LO</td>
<td>2.45</td>
<td>2.48</td>
<td>mid 96</td>
</tr>
<tr>
<td>D. pentaradiatus LO</td>
<td>2.51</td>
<td>2.53</td>
<td>mid 99</td>
</tr>
<tr>
<td>D. tamalis LCO</td>
<td>2.82</td>
<td>2.81</td>
<td>mid 115</td>
</tr>
<tr>
<td>N. atlantica left FO</td>
<td>2.83</td>
<td>2.82</td>
<td>top 116</td>
</tr>
<tr>
<td>Top paracme D. tamalis</td>
<td>2.86</td>
<td>2.85</td>
<td>mid 11</td>
</tr>
<tr>
<td>Base glacial regime</td>
<td>2.94</td>
<td>2.92</td>
<td>base 121</td>
</tr>
<tr>
<td>Base paracme D. tamalis</td>
<td>2.99</td>
<td>2.98</td>
<td>top 107</td>
</tr>
<tr>
<td>Sphaeroidinellopsis spp. LO</td>
<td>3.22</td>
<td>3.22</td>
<td>top 97</td>
</tr>
<tr>
<td>Gr. bononiensis FO</td>
<td>3.31</td>
<td>3.30</td>
<td>top 93</td>
</tr>
<tr>
<td>Top paracme D. pentaradiatus</td>
<td>3.56</td>
<td>3.53</td>
<td>base 81</td>
</tr>
<tr>
<td>Gr. punctatula LO</td>
<td>3.57</td>
<td>3.55</td>
<td>base 80</td>
</tr>
<tr>
<td>Sphenolithus spp. LO</td>
<td>3.73</td>
<td>3.67</td>
<td>top 73</td>
</tr>
<tr>
<td>Gr. margaritae LO</td>
<td>3.75</td>
<td>3.69</td>
<td>base 2</td>
</tr>
<tr>
<td>R. pseudoumbilicus LO</td>
<td>3.85</td>
<td>3.77</td>
<td>base 67</td>
</tr>
<tr>
<td>Base paracme D. pentaradiatus</td>
<td>3.90</td>
<td>3.81</td>
<td>mid 5</td>
</tr>
<tr>
<td>Gr. margaritae LCO</td>
<td>3.94</td>
<td>3.84</td>
<td>top 63</td>
</tr>
<tr>
<td>D. asymmetricus FCO</td>
<td>4.11</td>
<td>3.99</td>
<td>base 56</td>
</tr>
<tr>
<td>Gr. punctatula FO</td>
<td>4.52</td>
<td>4.31</td>
<td>base 37</td>
</tr>
<tr>
<td>Gr. margaritae FCO</td>
<td>5.10</td>
<td>4.93</td>
<td>top 10</td>
</tr>
<tr>
<td>Base Pliocene</td>
<td>5.33</td>
<td>5.13</td>
<td>base 1</td>
</tr>
</tbody>
</table>
the Plio-Pleistocene marine record of Sicily and Calabria a number of unconformities interpreted as 3rd- and 4th-order sequence boundaries marked by important lithological variations or erosional truncations (Figs. 3, 4). These surfaces, whose age is constrained by biostratigraphy and available magnetostratigraphy and cyclostratigraphy, occur everywhere synchronously; they were therefore assumed as essentially related to glacio-eustasy. The occurrence of sequence boundaries and condensed sections has also been inferred within continuous pelagic successions, where no facies contrast appears, but only cyclical variations in limestone/marl ratio are evident. In these cases, the sequence stratigraphic surfaces have been traced using correlations with long-term foraminifera abundance fluctuations, oxygen isotope stages and the astronomical record, based on the high-resolution grid of bioevents and available magnetostratigraphy. In the Plio-Pleistocene interval, both high-frequency fluctuations (21–41 ky) and low-frequency fluctuations (100 ky or multiple, generally 200–400 ky up to 1200 ky) are present in the deep-sea record. In the past, these oscillations have been related to Milankovitch effects forced by astronomical factors (e.g., Hays et al., 1976; Shackleton and Opdyke, 1977; Keigwin and Thunell, 1979; Ruddimann et al., 1986); important ice sheets variations are implied by these climatic factors (e.g., Hays et al., 1976; Shackleton and Opdyke, 1977; Keigwin and Thunell, 1979; Ruddimann et al., 1986); important ice sheets variations are implied by these climatic cycles. Mitchum and Van Wagoner (1991) noted that high-frequency cyclicity (4th-order and higher) has long been recognized in carbonate rocks, and the same high-frequency eustatic cyclicity can be recognized in siliciclastic rocks by invoking the same mechanism. The correlations we have attempted (see next paragraph) would appear to indicate that in the Plio-Pleistocene record the long eccentricity oscillations or multiples could represent long period climatic changes that produced global-sea level drops or basin starvation. The study of the relative conformities of sequence boundaries in the pelagic successions allowed precise age evaluations also in the shelfal successions that are often characterized by extended hiatuses.

We also attempted correlations between short-term cyclicity of the pelagic record (in particular the 41-ky average cycles, related to precession and orbital obliquity oscillations) and the occurrence of individual parasequences, which are particularly well preserved and exposed in the late Pliocene—lower Pleistocene marine record of Sicily. Several outcrop studies outlined high-frequency lithological cycles in Mediterranean Plio-Pleistocene pelagic successions (e.g., Hilgen, 1987; Thunell et al., 1991) and suggested its correlation with the oxygen isotope stages recorded in the Mediterranean ODP site 653A (Vergnaud-Grazzini et al., 1990) and the North Atlantic DSDP site 607 and 609 (Raymo et al., 1989; Ruddimann et al., 1986, 1989), as well as with astronomical cycles and foraminifera abundance fluctuations (Hilgen, 1991a,b; Sprovieri, 1992, 1993). According to Ruddimann et al. (1986) and Raymo et al. (1989), in particular, the dominant periodicity between 2.47 Ma (stage 100) and 0.735 Ma was 41 ky, and it was correlated to variations in Northern Hemisphere ice sheets. These oscillations consequently produced short-term glacio-eustatic fluctuations and sea-surface temperature variations, corresponding to the ice volume changes, as also outlined for the Mediterranean record by Lourens et al. (1992).

The Butera and Narbone Pliocene successions (reported later) and the Capodarso sequence (see Vitale, this volume) are good examples of evident correlations between 41-ky cycles, constrained by the ages of bioevents, and parasequences defined by stratal pattern and facies analysis.

**Comparison with other Sea-level Cycle Charts**

The lower resolution of the sea-level cycle chart of Haq et al. (1988) is difficult to correlate with the Mediterranean cycle chart. On the basis of the first sequence-stratigraphic studies of the Plio-Pleistocene of Sicily, Catalano et al. (1992a, b, 1993a, b, c) proposed a sea-level cycle chart supporting the same number and ages of sequence boundaries of the Wornardt and Vail (1991) cycle chart (both based on the Berggren et al. (1985) time scale), outlining a good correlability between Sicily and the Gulf of Mexico.

New data recovered in the central Mediterranean area suggest some discrepancies with the Gulf of Mexico curves. The sea-level cycle chart presented here is based on modern biochronology calibrated with high-frequency cyclostratigraphy and the astronomical record (Shackleton et al., 1990; Hilgen, 1991a, b; Cande and Kent, 1992). These schemes are commonly accepted as valid for the central Mediterranean region. Consequently, the discrepancies with the Wornardt and Vail (1991) cycle chart could be explained by different time scales and biozonations adopted for the Gulf of Mexico successions. In particular, we found small differences in the absolute ages of the lower Pliocene sequence boundaries, a striking similarity in the trend in-

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**Fig. 3.—** Chronostratigraphic chart of the Mediterranean Plio-Pleistocene depositional sequences and studied lithostratigraphic allogroups.

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Fig. 4.—Chronostratigraphic distribution of the main lithofacies associations and interpreted position of the sequence boundaries in a set of Mediterranean Plio-Pleistocene successions, assumed as reference sections for sequence stratigraphic calibration of the record (modified from Catalano et al., 1993a).
terval between 3.0 Ma and 0.8 Ma and one sequence boundary more in the upper Pliocene succession of Sicily not recognized in the Gulf of Mexico (Fig. 5).

The Sequence Stratigraphic Surfaces: Ages and Characters
Lower Pliocene.—

The lower Pliocene record of Sicily, Calabria and surrounding marine areas consists of pelagic deposits known in the region as “Trubi”, a rhythmic succession of 0.5-m-thick limestone-marl couplets. Since a classical sequence stratigraphic subdivision, based on shelf break variations, is not possible, we attempted correlations with the Gulf of Mexico sea-level cycle chart of Wornardt and Vail (1991) in order to point out in the record the possible deep-sea responses to the sea-level drops and maximum flooding events.

The succession of Trubi chalks at Capo Rossello on the southern coast of Sicily, defines the Zanclean neostratotype (Cita and Gartner, 1973). Since the 1970s, a number of micropaleontological studies, focused on the Trubi, defined a standard planktonic foraminiferal biozonation of the Mediterranean deep-sea record (Cita, 1973, 1975a, b). Recent studies on the Trubi succession provided a new nannoplankton biostratigraphy of the Zanclean (Rio et al., 1984) and new data on magnetostratigraphy, paleoclimatology, biochronology and cyclostratigraphy (e.g., Hilgen and Langereis, 1988; Zachariasse et al., 1989, 1990; Hilgen 1991a, b; Langereis and Hilgen, 1991; Sprovieri, 1992, 1993). Hilgen (1991a) proposed a useful correlation within the Capo Rossello composite section between the detailed record of CaCO₃ cycles and the astronomical solution according to Berger and Loutre (1991, 1992). This led to an extension of the astronomically calibrated new time scale of Shackleton et al. (1990) and Hilgen (1991a) down to the Miocene/Pliocene boundary, obtaining new absolute ages for magnetic reversals that are considerably older than those given by conventional time scales.

We attempted a sequence stratigraphic interpretation (Fig. 6) of the Hilgen scheme (1991b) in order to correlate the Mediterranean record with the sea-level cycle chart of Wornardt and Vail (1991). The proposed interpretation shows that the se-
sequence boundaries, or sometimes the cold-type condensed sections, correlate with the minimum peaks on the carbonate curves, generally corresponding to groups of closely spaced sapropel-rich levels (Figs. 6, 7). These cycle boundaries also seem to correlate with the long eccentricity minima (400 ky average cyclicity or multiple). The warm type-condensed sections correlate with the calcium carbonate and eccentricity maxima. Using these types of correlations, we obtained: (1) more precise evaluations of the occurrence of the sequence stratigraphic surfaces in the record, and the inferred relationships with the astronomical cycles; and (2) the absolute ages of the surfaces according to the most recent polarity time scale of Shackleton et al. (1990), Hilgen (1991a, b) and Cande and Kent (1992).

**Middle-upper Pliocene and lower Pleistocene.**

In this time interval of the study area, shelf and slope successions are available for sequence stratigraphic interpretation. The Mediterranean pelagic record can also be correlated with the oxygen isotopic curve of the North Atlantic DSDP site 607-609 record (Raymo et al., 1989; Ruddimann et al., 1989) and those recorded in ODP site 653A of the central Mediterranean (Vergnaud Grazzini et al., 1990; Thunell et al., 1990). Correlations with the isotope record are valid based on both bioevent ages and fluctuations of foraminifera according to Sprovieri (1992, 1993) and Rio et al. (1994). We also took into account some sediment-accumulation curves and foraminifera abundance cycles on both exploration wells and outcrops from Scarrantino (1993). The sequence stratigraphic interpretation of the pelagic record in this time interval is based on data and correlations with sections proposed by Hilgen (1991a, Fig. 3); Channell et al. (1992, Fig. 17); Lourens et al. (1992, Fig. 8); Scarrantino (1993, Fig. 6.1.1; Hilgen et al. (1993, Fig. 1); Di Stefano et al. (1993b, Fig. 2); Sprovieri (1993, Figs 5, 7, 9, 11, 13, 14, 15, 16, 18). With regard to the shelf-slope successions our original contribution in facies and stratal pattern analysis calibrated by biostratigraphy was based on the following land sections (Figs. 2A, 4): Poggioreale, Montevago, Selinunte, Capo Rossello, Monte S. Nicola, Monte Narbone, Butera, Capodarso, Lannari and Ficarazzi. Moreover the Onda, Pina, Marinella and Menfi wells from AGIP Co. have been considered according to recent biostratigraphic revisions (Di Stefano et al., 1993a; Scarrantino, 1993).

The proposed interpretations and correlations of the dataset point out that in the pelagic record:
1. Sequence boundaries and cold-type condensed sections are generally expressed by groups of closely-spaced sapropel-rich layers, or thick shaly intervals;
2. The warm-type condensed sections (maximum flooding surfaces) usually have no lithologic markers, except within the Trubi succession, where thick, carbonate-rich white beds are observed.

The stratigraphic analyses within the shelf successions also point out:
1. The same number of 3rd- and 4th-order sequences recognized on both deep-and shallow-water successions;
2. Well developed 20- to 50-m-thick high-frequency shelf cycles are preserved in the record, particularly between 2.5–2.1 Ma and between 1.6–0.8 Ma.
3. The boundary of a 2nd-order cycle was produced by a peak regression at about 1.4 Ma (Fig. 3).

Middle—Upper Pleistocene.—

The Middle-Upper Pleistocene record, not well preserved on land, represents the thickest part in several offshore basins fills. Consequently, our sequence stratigraphic analysis is mostly based on the interpretation of the seismic lines. Important unconformities are useful marker-reflectors in the studied region, both in the northern and southern offshore of Sicily; they bound well-developed, several hundred meters thick sequences. Stratigraphic geometries and seismic facies distribution analysis pointed out two 3rd-order depositional sequences enclosing eight 4th-order sequences. Since high-quality biostratigraphic controls are not available, the ages of the sequence stratigraphic surfaces are inferred from imprecise correlations to ODP site 653A in the Tyrrenian Sea (Vergnaud-Graziini et al., 1990) and to the Pacific core V28-239 (Shackleton and Opdyke, 1976). Our observations on seismic lines confirm the same number of sealevel oscillations expected from analysis of the deep-sea record in and out of the Mediterranean. Several authors pointed out that the dominant cyclicity in this part of the record is the 100 ky component of the short eccentricity, which reflected the intense oscillations of glacial-interglacial periods of the last 0.8 my (e.g., Ruddiman et al., 1986). Wornardt and Vail (1991) also interpreted the sequences of the last 0.8 my as reflecting sea-level oscillations of the 100-ky periodicity. Taking into account these correlations and observations, we believe that the sequences found in the middle-upper Pleistocene basin fill of the Sicily offshore reflect the same 100-ky frequency.

BASIN ANALYSIS AND SEQUENCE STRATIGRAPHY

Offshore studies

In order to analyse the Plio-Pleistocene sedimentary infill of the Sicily offshore basins, we used data from both wells and seismic lines. A multichannel seismic profile grid of C and G zones from the Italian Minister of Industry provided the regional stratigraphic analysis and aided in the correlation of sequences. In the framework of a scientific collaboration project between the Dipartimento di Geologia e Geodesia of Palermo University, the Geomare Sud of Naples and the Istituto Universitario Navale of Naples, a number of oceanographic cruises were also organized in recent years to acquire high-resolution seismic reflection profiles in the central Mediterranean. These profiles provided detailed data at the scale of high-frequency sequences and parasequences. Well logs from Onda, Pina 1, Orion, Oscar, Pamela, Marinella and Menfi exploration wells from the Sicily mainland and southern offshore calibrate the most interesting sections (Fig. 2A). AGIP also provided cuttings from Menfi, Marinella, Onda and Pina wells; the analyses yielded a new set of high-resolution biostratigraphic data (Scarantino, 1993; Di Stefano et al., 1993a).

The sequence stratigraphic interpretation of the Sicily offshore basin fill was performed in the framework of the fundamental papers on well-log and seismic sequence stratigraphic analysis (e.g., Payton, 1977; Wilgus et al., 1988; Vail and Wornardt, 1990; Van Wagoner et al., 1990; Wornardt and Vail, 1991; Vail et al., 1977, 1990). The subdivision in sequences and systems tracts had a resolution comparable to that provided by the outcrop basin analysis. Some 3rd-order sequence boundaries have not been traced because of the poor resolution of the seismic lines, the characters of lithofacies associations and local tectonic factors. Consequently, we grouped some sequences in higher-order sets. For instance, the Trubi group, consisting of three different 3rd-order sequences, is distinguished on outcrop on the basis of lithologic cycles and foraminifera abundance fluctuations. The same subdivision is clearly impossible on seismics. The Plio-Pleistocene sedimentary succession has been subdivided into six 3rd-order sequences, within which 14 4th-order sequences have been distinguished. All sequences having regional or subregional extension are type 1 sensu Vail et al.
(1977) and their boundaries are associated with erosional truncations.

**The Southern Sicily offshore.**

In this area, we provide a sequence stratigraphic interpretation of the Plio-Pleistocene strata of the Selinunte and Sciacca basins, extending along the southern Sicily offshore (Fig. 2A). The Selinunte offshore basin sedimentary infilling consists mainly of shelf deposits (Fig. 8, northwestern side of the seismic line C529) more than 1000 m thick. The Sciacca offshore basin, a northwest-southeast trending intraslope basin filled by more than 2000 m of clastic carbonate deposits, represents a late Plio-Pleistocene foredeep of the frontal part of the Sicilian chain (Fig. 8, southeastern side of the seismic line C529). The Selinunte and Sciacca basins are separated by a north-south trending structural high which was actively undergoing uplift until early Pleistocene. The sequence boundaries are generally tectonically enhanced and associated with angular unconformities and stratigraphic gaps that extended close to the major anticlines.

Common elements for both basins are the reduced thickness of the Pliocene succession (about 400–450 m) that was deposited under starved conditions and the strong increase in thickness of the Pleistocene succession (up to 2000 m). The Pliocene sequences are characterized by transparent seismic facies, whereas the Pleistocene sequences show numerous internal reflectors with high amplitude and lateral continuity.

The Selinunte offshore basin.—This basin is well described by the seismic line C529 (Fig. 8). It was directly calibrated to the Onda well (Lat. N 37°33′24″E, Long. E of Greenwich 12°49′11″, 232, Fig. 9) which in turn was correlated on land with the Marinella well (see location on Fig. 2A). The well log sequence analysis, carried out from SP and Resistivity logs at scale 1:1000, allowed us to distinguish three 3rd-order sequences and three 4th-order sequences. Their boundaries have been dated on the basis of biostratigraphic analysis carried out from a total of 90 samples (Di Stefano et al., 1993a). The P3 sequence rests unconformably on Messinian evaporites that have been found at a depth of 710–730 m. The biostratigraphic data indicate that the unconformity marks a hiatus corresponding to the Amaurolithus tricorniculatus, Ceratolithus rugosus and to the lower part of the Reticulofenestra pseudoumbilicus Zones. The absence of lowermost Pliocene deposits appears related to the tectonic activity of the growth-anticline located close to the well site (Fig. 8, northwestern side of the seismic line C529). The log analysis of the P3 sequence which consists mainly of Trubi chalks (710-635 m) yielded a number of cyclic variations probably related to major changes of carbonate content. The Trubi succession is represented on seismic lines by two reflectors with low-to medium-amplitude and lateral continuity (a in Fig. 8).

The P4a and P4b sequences consist of a 60-m-thick succession of marls and sandy shales that unconformably overlie the P3 sequence. The boundary between P4a and P4b has been placed, in the Onda well, at a depth of 570 m (Fig. 9). The P4a, P4b and the lower part of the P4c (out in the Onda well) sequences show a prevailing transparent seismic facies interbedded with a few seismic reflectors characterized by high amplitude and lateral continuity (b and c in Fig. 8).

In the Onda well, the Q3 sequence unconformably overlies the P4b sequence. The biostratigraphic analysis documented a stratigraphic gap ranging from the late Pliocene to the early Pleistocene and corresponding to the *D. pentaradiatus*, *Discoaster brouweri*, *Dictyococites productus*, *Calcidiscus macytrei*, *Helicosphaera sellii* and probably the lower part of the Large *Gephyrocapsa* Zones. The lower boundary of the Q3 sequence, located at a depth of 555 m in the well, was enhanced by middle-late Pliocene compressional tectonics (Fig. 8, northwestern side of the seismic line C529). This boundary is represented on line C529 by a reflector with high amplitude and lateral continuity associated with an abrupt change of seismic facies and a basinward shift of the oflap break (d in Fig. 8). The high-amplitude and lateral continuity reflectors of the Q3 sequence correspond, in the Onda well log, to interbedded...
The well-log analysis recognized six 3rd-order sequences and eleven 4th-order sequences (Fig. 11). The 4th-order sequence boundaries have been clearly recognized and correlated on low-and high-resolution seismic lines. The 4th-order sequence boundaries have been clearly recognized and correlated on low-and high-resolution seismic lines. In the 4th-order sequences, the lowstand prograding complex of the Q3 sequence corresponds to a peak of 2nd-order regressive cycle.

The lower boundary, located in the well at a depth of 2255 m, is associated with a stratigraphic gap corresponding to the A. tricorniculatus and partly to the C. rugosus Zones. The upper boundary is associated with a lithologic facies variation revealed both on seismic and well-log curves (Figs. 10, 11).

Upsection, the Trubi chalks are gradually replaced by a 150 m thick succession of marly shales with sandy intercalations. These deposits form the P4 and P5 sequences and are represented by a transparent seismic facies (a in Fig. 10). The biostratigraphic analysis (Di Stefano et al., 1993a) pointed out a sedimentary gap coincident with the D. pentaradiatus zone at the interval between 2090-2075 m related to the P4b upper boundary. Shingled turbidites have been recognized at the base of the lowstand systems tract of Q1 (at a depth of 1925–1970 m). This type of turbidite is probably due to the instability of the depositional slope linked to the growing structures that formed the northwestern shoulder of the basin (Fig. 10). On well logs the shingled turbidites are represented by Christmas tree shaped curves (Fig. 11), on seismic lines by reflectors with high amplitude and lateral continuity interbedded with reflectors with low amplitude and low lateral continuity or transparent seismic facies (b in Fig. 10).

Biostratigraphic control on the Pina well, at a depth of 1850 m, documented an apparent hiatus corresponding to the H. sellii and lower part of Large Gephyrocapsa Zones; the lack of these biozones has been interpreted as the Q3 sequence lower boundary truncating the Q1 sequence (Di Stefano et al., 1993a). On seismic lines, a group of thinned horizons found between Q3 and Q1 sequences has been interpreted as corresponding to the Q2 sequence. The Q3 lower boundary is tectonically enhanced (as shown in Fig. 8, southeastern side of the seismic line C529 and Fig. 10, northwestern side of the seismic line C531). The basal part of the Q3 sequence consists of a 350-m-thick interval of shingled turbidites. Upsection, a thick lowstand prograding complex, formed by four parasequences, is followed by thin transgressive and highstand systems tracts. The top of the lowstand prograding complex of the Q3 sequence corresponds to a peak of 2nd-order regressive cycle.

The Q4 sequence is composed by three 4th-order sequences. Its lower boundary appears enhanced by the uplifting of the Gela Thrust System (Fig. 8, southeastern side of C529). The upper boundary of the Q4 sequence is inferred at 0.5 Ma. Unfortunately, the Pseudoemiliana lacunosa LO event has not been clearly found in the well probably due to the reworking of sediments between 600 and 900 m of depth (c in Fig. 10).

The 4th-order sequence boundaries have been clearly recognized and correlated on low-and high-resolution seismic lines. In the 4th-order sequences, the lowstand systems tracts generally consist of:

1. sheet turbidites of the basin floor fan in the Q4a sequence (type 1 turbidites of Mutti, 1985) represented by high-amplitude and lateral continuity reflectors (d in Fig. 10);
2. channel-levee systems of the slope fan (type 3 of turbidites of Mutti, 1985) characterized, on well logs, by several crescent-shaped log patterns of coarsening/fining upward lobes. Within crescent-shaped lobes, the curves are highly digitated. On seismic sections, channel-levee systems are represented by chaotic (corresponding to the mass transport at the base of channel-levee), hummocky facies and wedge-shaped reliefs (e in Fig. 10).
The lowstand prograding complex, transgressive and highstand systems tracts are represented by groups of reflectors with high amplitude and lateral continuity (f in Fig. 10). The upper boundary of the Q4 sequence is related to a dramatic sea-level fall: upper slope and shelf deposits of the Q5 sequence rest directly on basinal deposits of the Q4 sequence.

Tectono-eustatic 2nd order cycles.—Groups of 3rd- and 4th-order depositional sequences arranged to form transgressive-regressive facies cycles are bounded by major tectonically enhanced unconformities (Vail et al., 1990). Each cycle starts when an increase in subsidence rate occurs causing starved conditions in the basins; transgressive trends are formed in this case (Fig. 12). On seismic lines, the transgression is represented by prevailing transparent seismic facies with interbedded high-amplitude and lateral continuity seismic reflectors (a in Fig. 12). When an uplift occurs at the basin margin or the subsidence rate is slower, the basin evolves to overfilled conditions forming a regressive pattern. On seismic sections they appear as prevailing high-amplitude and lateral continuity seismic reflections and, locally, as chaotic facies (b in Fig. 12).

In the studied basins, we recognized three main tectonic cycles from their stratigraphic signature. In the lower cycle, encompassed by the tectonically enhanced lower boundaries of the P3 and Q3 sequences, the increase in subsidence rate is related to the extensional opening of the Selinunte and Sciacca offshore basins. The slower subsidence rate corresponding to a growing anticline (on the right side of Fig. 8 and on the left side of Fig. 10) causes the end of the cycle. The middle cycle, enclosed between the lower boundaries of the Q3 and Q5 sequences, shows a transgressive trend related to an increase of the Gela thrust system loading while the regressive system is related to the uplift of the nappe (Fig. 8). The upper cycle starts when a younger extensional tectonic event affected the Gela thrust system, increasing the accommodation space.

The Northern extensional margin.—

The Erice basin.—The Erice basin, controlled by strong vertical tectonics, is located on the Tyrrhenian side of the Sicily continental margin, at a depth of 2000 m below sea level (Fig. 13). Previous studies based on seismostratigraphic criteria provided the tectonic and stratigraphic framework of the area. Catalano et al. (1992c) recognized within the Plio-Pleistocene record the same number of seismic sequences on both northern and southern Sicily offshore basins; in the Erice basin record, Agate et al. (1993) pointed out a number of sequence boundaries tectonically enhanced by post-Messinian regional tectonic events.

Since the area has no stratigraphic control by exploration wells, we studied the basin succession using only seismostratigraphic analyses. The sedimentary fill of the Erice basin consists of a layered Plio-Pleistocene succession almost 1 s (twt) thick unconformably overlying a strong reflector with high amplitude and good lateral continuity corresponding to the Messinian horizon. We observed that groups of seismic sequences are encompassed between two major tectonically enhanced sequence boundaries. Each sequence group shows the clear pattern of a transgressive-regressive facies cycle. We recognized three transgressive-regressive facies cycles (Figs. 12A, B; Fig. 13) that correspond, in number and facies trend evolution, to those already described before for the southern Sicily offshore well-controlled basins (Figs. 8–12).

The lower cycle shows, from base to top, a transparent seismic facies (a in Fig. 13) and reflections with medium to high amplitude and lateral continuity. Upward, a lowstand prograding complex has been recognized (b in Fig. 13). The regressive trend, related to basin inversion tectonics (Agate et al., 1993, Fig. 10), is bounded by a tectonically enhanced unconformity (c in Fig. 13). In the middle cycle, the transgressive trend corresponds to the extensional tectonics that increases the accommodation space. The regressive trend is related to the compressional tectonics enhancing the upper boundary of the cycle.
Systems tracts

- Highstand systems tract
- Transgressive systems tract
- Lowstand prograding complex
- Shingled turbidites
- Slope fan

Lithology

- Shale
- Sand
- Shale and sand
- Marl limestone (Tribe Fm.)
- Lacustrine limestone

Fig. 11.—Biostratigraphy and sequence stratigraphic interpretation of the electric and sonic logs of Pina 1 well (well site located on Fig. 2A).
Fig. 12.—Set of both high- and low-resolution seismic reflection profiles from northern and southern Sicily offshore, showing the 2nd-order transgressive-regressive seismic facies cycles (ages of the sequence boundaries in Ma). Location of A and B profiles on Figure 13; Onda and Pina profiles located respectively on Figures 8 and 10.

Fig. 13.—A segment of the S90-16 high-resolution seismic line (sparker 16 kJ) across southern Tyrrhenian margin (for a, b, c, d, e symbols see text). A and B traces correspond to profiles of Figure 12. Trace of the profile on Figure 2.
(Agate et al., 1993, Fig. 9). The upper cycle starts when a younger extensional tectonic event occurs; the subsidence increase is considered responsible for the emplacement of rotational slides (d in Fig. 13) and debris flows (e in Fig. 13).

The correlation of seismic sequences and transgressive-regressive cycles in two different-settled basins implies a common stratigraphic signature in both northern and southern Sicily offshore. Compared with the southern Sicily offshore record that was age-calibrated by wells, the Erice basin is a seismic example showing application of sequence stratigraphy in areas lacking biostratigraphic data. Age prediction of both 2nd-order cycle boundaries and 3rd-order sequences using our sea-level cycle chart (Fig. 3) is possible based on physical stratigraphy.

**Outcrop Studies**

**Pliocene sequences.**—

The Pliocene sequences exposed on land generally fill syntectonic basins located at the top of advancing major thrusts or onlap minor structures forming a roof-thrust complex of Miocene terrains detached from the Mesozoic-Paleogene substratum. Asymmetric Pliocene basins wedging northward are commonly found in western and central Sicily. The subsidence rate of these basins is often very high due to thrust pile loading, but it may be locally exceeded by the uplift close to the hinge of growth structures. In these cases, tectonic unconformities may locally develop with very reduced hiatuses. The intense syn-sedimentary tectonic activity is also revealed by the occurrence of slumps and mass transport movements along the tilting flank of high-amplitude growth-fold systems in deep-sea environments. Clear erosional surfaces truncating the hinges of anticlines may also form in a shelf setting during a time interval corresponding to the deposition of the parasequence set (Vitale, this volume). As described in the following examples, the sequence stratigraphic tool applied to the Pliocene syntectonic basin fill improved the age constraints and outlined the structural evolution.

**The Belice Valley basin.**—Along the Belice Valley, western Sicily, a well exposed, thick Pliocene succession filled a piggyback basin developed above the southvergent Magaggiaro thrust sheet (see Vitale, this volume, Fig. 15). The basin-fill onlaps northward the Poggioreale ridge, a structural culmination consisting of a major fault-propagation anticline (Fig. 14) and related minor folds (Vitale, 1990). Southward, the Belice basin is bounded by the Montevago normal fault, a listric discontinuity showing evidence of synsedimentary growth during the middle Pliocene-early Pleistocene, as revealed by the fan shaped fill divergent toward the fault plane at the south (Fig. 15).

Along the Poggioreale ridge (Fig. 14), the basal Pliocene strata, represented by the Trubi pelagic chalks, unconformably overlie the Messinian evaporites. The Trubi chalks underlie a thick package of terrigenous deposits locally known as the “Marnoso Arenacea del Belice” formation (MAB), which consists of three members. A reduced interval of brownish hemipelagic silty-shales covered by a 100- to 150-m-thick package of fine-grained turbidites composes the lower MAB member. The transition of the Trubi chalks to the basal shales of the MAB formation is considerably younger than in many other known Trubi successions in Sicily (within the MPL 3 biozone). Di Stefano and Vitale (1988) discovered a hiatus between the top of the Trubi strata and the lower MAB member, corresponding to the *R. pseudoumbilicus* calcareous nanofossil biozone. Vitale (1990) provided the first sequence stratigraphic interpretation of the succession, recognizing two Pliocene depositional sequences above the Trubi units, and advocated the role of slumps along the tilting flank of the Poggioreale growth-fold systems in generating a deep-sea tectonic unconformity (see also Vitale, this volume). The lower sequence, formed by the lower and middle MAB members, is syntectonic to the growth of a major anticline formed in the basin during the lower Pliocene and consists of deep-sea deposits. The upper sequence, consisting of the upper MAB member shelf deposits, is post-tectonic and appears only weakly tilted southward. This reconstruction constrained the age of the thrust-related folding event displayed in the area.

Along the Poggioreale ridge, the boundary between the Trubi chalks and the lower MAB member is marked by a paraconformity laterally passing to an angular unconformity near the major anticline hinge to the north (Fig. 14). This surface represents the lower sequence boundary of the P3 sequence, dated at about 4.2 Ma. In other synchronous successions of Sicily, where no facies contrast occurs, this sequence boundary is revealed by a low-frequency peak in the carbonate content correlated to a cold climate event and to long eccentricity maxima in the astronomical record (see also Fig. 6). Consequently, we interpreted the angular unconformity and deep-sea erosion surfaces occurring at the Poggioreale ridge as being related to the local synsedimentary tectonic activity. The local tectonics only enhances an important, large-scale facies change related to glacio-eustasy. The top of the P3 sequence is bounded by an unconformity surface dated at about 3.0 Ma, within MPL 4a and *Diccoaster tamalis* calcareous plankton biozones. Here, this surface marks a sharp contact between the basin-to-slope deposits of the P3 sequence and the lowermost shelf prograding strata of the P4 sequence, pointing out a clear basinward shift of the shelf break. Along the Poggioreale ridge, the 3.0 Ma sequence boundary commonly represents an erosional truncation cutting the hinge of the major anticline exposed to the north; this surface, generated as a consequence of a sea-level drop, is also tectonically enhanced and marks the end of the thrusting event. The overlying shelf deposits (upper MAB member) are not involved in the structures of the substratum (Figs. 14, 15), showing only a minor tilting southward. Elsewhere in Sicily, the same sequence boundary in the pelagic record is represented by a sharp contact between the Trubi chalks and the silty shales of the Narbone Formation.

Southward, the 3rd-order P4 sequence is well exposed. On the basis of the stacking pattern analysis and biostratigraphic data we can distinguish the 4th-order P4.a, P4.b and P4.c sequences (Fig. 15) that may be seen as the systems tracts of the P4 sequence. The P4.a sequence is formed by a set of 5 major cycles; each parasequence consists of 50- to 100-m-thick prograding clinoforms of shallow-water conglomerates, sandstones and calcarenites interbedded with outer shelf silty-shales. The highstand systems tract forms the thickest part of the P4.a sequence and shows a progradation of about 4 km southwards.

Two minor unconformities, exposed at the south of the Belice River incision, are the basal boundaries of the P4.a and P4.c sequences, occurring within the *D. pentaradiatus* Zone; the mudstones of the uppermost one are within the *D. brouweri*
Fig. 14.—Panoramic view of the Poggioreale ridge. The photo illustrates the geometrical relationships between the major growth structure in the north (left) and the syntectonic strata of the lower Pliocene the south (right). The major unconformities, enhanced by important facies changes are related to glacio-eustasy.

Fig. 15.—Geologic composite section (see trace Bs on Fig. 2A) and sequence stratigraphy of the Pliocene strata of the Belice Valley. The P4 sequence is the thicker fill of the piggyback basin growing above the Magaggiaro thrust at south (left side of the profile and Fig. 16). The stratal expansion downward and the fan-shaped arrangement of the parasequences suggest an increase subsidence was locally controlled by syn-sedimentary tectonics.
Zone and consequently correlate with the Capodarso depositional sequence (Figs. 2, 3; see also Vitale, this volume). These sequence boundaries are revealed by sharp contacts at the base of shallow-water grainstone bodies overlying offshore mudstone packages, whereas the internal stacking pattern of the P.4.b and P.4.c sequences is not resolved due to the poor outcrop and the prevalent muddy facies.

We did not find *Gephyrocapsa oceanica* s.l. in the uppermost mudstone levels outcropping at the northeastern flank of the Monte Magaggiaro (Fig. 15), and we consequently concluded that the whole mudstone succession is of Pliocene age; the overlying calcarenites have been referred to the lower Pleistocene only on the basis of the low-resolution stratigraphy from the molluscan fauna.

The Menfi basin.—South of the Belice River Valley, a Plio-Pleistocene succession is exposed in the area of Selinunte (an important archeological site in Sicily) and fills the Menfi basin with a total thickness exceeding one kilometer. The geologic surface data of Vitale (1990) pointed out that the Menfi basin onlaps the southern flank of the Monte Magaggiaro structure, a major ramp anticline mainly composed of Mesozoic carbonates. We propose a sequence stratigraphic reinterpretation of the Plio-Pleistocene basin fill, based on original surface data, AGIP well logs, available seismic lines, and revised biostratigraphy of the Menfi and Marinella wells (Scarantino, 1993).

The profile across the area (Fig. 16) shows the Magaggiaro anticline overthrusting Miocene to lowermost Pliocene terrains forming a thin roof-thrust complex detached by their Mesozoic substrate. The bottom of the succession not involved in the roof-complex is located at a depth of 1280m in the Marinella 2 well and consists of strata pertaining to the P3 sequence (Fig. 17). Log signals suggest the occurrence of hemipelagic shales with thin sandstone units. The P3 sequence is affected only by weak folding, with onlap terminations of basal strata, and is consequently considered partly syntectonic to the last fault-propagation folding of the roof complex. At the surface, the P3 sequence is not present along the profile; an incomplete portion of the lowermost Pliocene Trubi chalks is unconformably truncated at the top by an 80-m-thick package of biocalcarenites tentatively referred to as middle upper Pliocene age. Based on biostratigraphic data from outcrops we attribute the Trubi strata to the P1–P2 sequences. The same Trubi levels are found in the roof-complex at the bottom of the Menfi well; we consequently consider the Miocene-lower Pliocene outcropping succession in the northern side of the basin as also detached from the Mesozoic-Paleogene substratum (northern side of the profile in Fig. 16).

The succession in the Marinella well continues upward with a 300-m-thick package not present at the surface because of onlap terminations of strata above the lower part of the Magaggiaro anticline flank (Figs. 16, 17). These strata, a prevailing muddy facies coarsening upward to sand, are assigned to the sequences P4 and P5 according to the biostratigraphic data from the Marinella well (Fig. 17). The log stratigraphy points out the possible occurrence of minor hiatuses at 2.7 Ma (1105-m depth), at 2.5 Ma approximately (1090-m depth, a few meters above the *Globorotalia bononiensis* LO) and at 2.1 Ma (1075-m depth, below the *Globorotalia inflata* LO, Scarantino, 1993). Unfortunately, the sequence boundaries cannot be traced within the studied line because of the poor seismic resolution; conse-

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**Fig. 16.**—Geologic profile and sequence stratigraphy of the Plio-Pleistocene fill of the Menfi basin (see trace Ms on Fig. 2A), based on surface, wells and seismic data.
sequently, we grouped together the sequences P.4.a, P.4.b, P.4.c and P5, also considering the reduced thickness of the package. These sequences are overlain by the Q1 sequence, consisting of a sandy facies association with minor interbedded shale layers. The Q1 strata are present in the well between 900- and 730-m depth and crop out along the Belice River where they show a prevailing silty facies association. The Q1 sequence covers the biostratigraphic interval of the D. productus Zone and contains in its upper part the G. oceanaica s.1. LO event. The stratal pattern on the profile shows that only the upper part of the sequence is exposed on the surface (Fig. 16).

A 50-m-thick calcarenite interval outcrops along the Belice River incision, dipping southward about 5° and unconformably overlying the Q1 sandy siltstones (Fig. 16). Within this package a set of three parasequences occurs; the H. baltica FO is found in the uppermost parasequence, confirming the Butera basin data (see following paragraph). According to both seismic and outcrop data these calcarenites are the lowstand and transgressive systems tracts of the Q2 sequence. The basal calcarenites (HST and TST) of the Q2 sequence rest within the H. sellii Zone, above the C. macintyrei LO and containing Gephyrocapsa spp. < 5.5 μ. In the seismic lines the base of the Q2 sequence can be correlated with a sharp contact between siltstones and overlying sands located at 730-m depth (Fig. 17). Di Stefano et al. (1993a) inferred the occurrence of an important hiatus at 690 m, located between the C. macintyrei Zone and the Large Gephyrocapsa Zone (1.2 Ma, now aged at about 1.4 Ma according to the new absolute time scales). Outcrop and seismic stratal pattern analysis reveals that no important erosion occurs below this unconformity; consequently we infer that the occurrence of H. sellii Zone could not be detected by Di Stefano et al. (1993a) because of the preservation of cuttings, as well as the high sand content. The Pleistocene stratal pattern of the Menfi basin fill suggests that the hiatuses are minor and that sedimentation was substantially continuous, as supported by the high subsidence and sediment supply rates.

The log analysis led us to identify the Q3 sequence basal boundary (1.4 Ma) at about 500-m depth on the Marinella well and at about 450 m on the Menfi well. Important log signatures in the Menfi log, pointing out sandy units, are found at the base of this sequence. We correlate this boundary to peak 44 of the foraminifera abundance curve of Scarantino (1993, Fig. 6.6.1). Upward, the Q3 sequence consists of prevailing muddy facies with rare and thin calcarenite levels, outcropping along the Belice River valley with generally poor exposures. Its upper boundary is visible at the surface at Selinunte, Casa Catarincchia site. Here, a 15-m-thick inner shelf-transition calcarenite horizon overlies an outer shelf mudstone succession pertaining to the Small Gephyrocapsa Zone. The calcarenite overlies a 70-m-thick offshore mudstone succession; these mudstones are referred to the Globorotalia truncatulinoides excelsa and P. lacunosa Zones (Scarantino, 1993) and terminate the outcropping succession near the present-day coastline. Di Stefano et al. (1991) correlated the Selinunte calcarenite horizon to a calcarenite bed outcropping at Ficarazzi (near Palermo) which defines the top of the Sicilian chronostratigraphic substage (Ruggieri et al., 1984; Rio et al., 1991), and is dated at about 0.8 Ma.

**Lowermost Pleistocene sequences**

In order to establish the precise ages of sequence stratigraphic surfaces in this interval and to verify the correlation potential between high-frequency shelf facies cycles and bioevents, we studied in detail the Monte Narbone succession, located close to Agrigento, and the Butera basin fill exposed close to Gela in southeastern Sicily. These successions provided good control on the occurrence of an important sequence boundary at 1.6 Ma, and its correlation with the cycle 55-54 of the North Atlantic δ¹⁸O formalized record (Ruddimann et al., 1989). Within the lower Pleistocene, our high-resolution stratigraphic data point out the synchronous occurrence of sequence boundaries, independently of local factors. The tectonic evolution of
the studied basins is similar to those observed in central Sicily: superficial thrusting during the deposition of the Trubi forma-
tion and folding gradually decreasing up to the Lower Pleis-
tocene. The lower Pleistocene strata very often fill low-ampli-
tude growth synclines; repeated episodes of
depocenter-directed slumps testify to a synsedimentary tilting
of the flanks of these minor satellite basins.

Monte Narbone.—The Pleistocene succession of Monte Nar-
bone (1°15' Long.-37°15' Lat.; see location on Fig. 2A) lies
over Pliocene strata represented by Trubi chalks at the base and
hemipelagic shales upward. The succession is the fill of a low-
amplitude growth-syncline striking about north 20° east and
outcropping a few kilometers north of Marina di Palma.

A succession of 7 shelf cycles is exposed along the south-
eastern flank of the Monte Narbone (Fig. 18). Each cycle, about
50 m thick, consists of bioturbated grey shales grading upward
to white calcareous mudstones (cycles 1–4) or sandy siltstones
(cycles 5–6), capped by a biocalcarenite (packstone) horizon
(cycle 7). The fossil content and facies analysis reveal that the
shales, often bearing an in-situ molluscan fauna, were deposited
in an outer shelf environment, whereas the sandy siltstones,
having plane-parallel stratification and sharp upper boundary,
were formed in offshore-transition and inner shelf environ-
ments. In the four lowermost cycles, the paleobathymetric var-
iations are less evident, but the occurrence of a cold-type for-
aminifera assemblage is constantly present in the white
calcareous mudstones. Consequently, each cycle marks a shoal-
ing upward trend bounded by sharp flooding surfaces. The
thickness variation trend of the exposed parasequences records
a stacking pattern evolution from aggradation to progradation.

On the basis of these data, we propose a sequence-strati-
graphic interpretation of the Monte Narbone section (Fig. 18).
A lower sequence boundary is at the lowermost shelf cycle
above the hemipelagic shales (base of cycle 1). The occurrence
of cold- and warm-type condensed sections is revealed by in-
situ molluscs (Corbula gibba) and iron/manganese nodules par-
ticularly concentrated at 3–4 boundary cycle and at the base of
cycle 5. An upper sequence boundary is placed at the base of
the 7th cycle, where lower shoreface medium to coarse sandy
strata occur.

Our biostratigraphic analyses resulted in the following data:
1. The D. productus calcareous nannoplankton Zone is within
the basal part of the sampled interval;
2. The first occurrence of left-type Neogloboquadrina pachy-
derma is within the white mudstones at the top of the first
cycle;
3. The G. oceanica s.1. FO event is close to the base of the
sandy levels at the top of the third cycle.
4. The boundary between C. macintyreii and H. sellii zones has
been uncertainly placed within cycle 6.

In accordance with the bioevent age evaluations of Sprovieri
(1993), we correlate the Monte Narbone cycle 1 with stages 65-
64 and cycle 3 with stages 61-60. Consequently, we indicate:
(a) the lower sequence boundary as stage 66, (b) the warm-type
condensed section (maximum flooding surface) as the stage 57,
and (c) a cold-type condensed section as stage 60. Moreover,
the aggradational stratigraph pattern, as well as the lacking of deep
erosional truncations at the base of cycle 7, also allow us to
exclude the occurrence of a very extended hiatus below the
uppermost calcarenite level. Consequently, numbering the up-
permost cycles as their corresponding isotopic stages, we ob-
tained a correlation of the upper sequence boundary to stage 54
(about 1.58 Ma). These results confirmed the interpretation and
the correlation with pelagic successions located elsewhere in
the study area (Fig. 4). Following the field dataset we conclude
that the Monte Narbone succession is a documented outcrop
example of the complete sequence Q1.

Butera.—We studied part of the Q2 sequence, lying over the
previously described 1.58 Ma sequence boundary in the area of
Butera, southeastern Sicily (1°45' Long., 37°10' Lat.). Here the Plio-Pleistocene succession was also deposited in a growing low-amplitude syncline, northeast-southwest striking for more than 10 km at the north of Gela (Fig. 19). The Butera Pleistocene succession unconformably covers a deformed substratum consisting of Miocene shales and evaporites, lower Pliocene Trubi chalks and the middle-upper Pliocene hemipelagic shales. The Pleistocene succession is formed by a 500-m-thick alternation of yellowish shallow-marine sandstones, fossiliferous calcarenitic bars, and bioturbated sandy and silty grey shales, interpreted as a coastal to open shelf facies deposited close to a deltaic influx. The stratal pattern (Fig. 20) documents a set of four major coarsening upward cycles (parasequences). A type cycle consists of thick sandstones bodies prograding towards south south-east over offshore mudstone strata. The stacking pattern suggests the occurrence of two major prograding cycles at the base (lowstand prograding complex) overlay by backstepping and aggradational cycles formed by thick mudstones and thin cross-stratified calcarenites upwards (transgressive and early highstand systems tracts). Each parasequence, on average 100 m thick near the depocenter of the basin, thins towards the flanks down to 20–30 m, where minor erosional surfaces are found (Fig. 20). These minor tectonic unconformities are clearly associated with the growth of the underlying structures and, by their local occurrence, cannot be confused with the sequence boundaries.

The mudstones of the first cycle, representing the lower boundary of the Butera depositional sequence, enclose the *C. macintyrei* LO and contain *G. oceania* s.l. from the outcropping base. The mudstones at the base of the third cycle pertain to the Large *Gephyrocapsa* zone and contain *Hyalinea baltica*. Sprovieri (1993) correlates the *Hyalinea baltica* FO in the Mediterranean with stage 50 of the Atlantic record. Consequently, we can correlate the lower sequence boundary to stages 55–54 and the uppermost cycle outcropping at Butera village to stages 49–48 (Fig. 20).

On the basis of our data, we conclude that the studied lower Pleistocene facies cycles formed in response to high-frequency global sea-level oscillations. Using the absolute age of bioevents in a high-resolution frame, these fluctuations can be correlated to the high-amplitude and low-period climate changes associated with the 41 ky component of the orbital obliquity as also found in the middle and upper Pliocene-lower Pleistocene carbon and oxygen isotopes deep-sea record in the North Atlantic and central Mediterranean (Raymo et al., 1989; Ruddiman et al., 1989; Thunell et al., 1990; Vergnaud-Grazzini et al., 1990).

![Fig. 19.—Geologic section crossing the Butera basin, southeastern Sicily (see location map of Fig. 2A). The Plio-Pleistocene strata fill a low-amplitude and large-wavelength syncline growing on the deformed Miocene substratum of the Gela Thrust system.](image1)

![Fig. 20.—Sequence stratigraphic interpretation of the Butera basin fill.](image2)
CONCLUSIONS

The marine Plio-Pleistocene record of the central Mediterranean area, particularly in Sicily successions, was studied and interpreted using sequence stratigraphy methodologies applied to outcrops, seismic lines and well logs, on land and sea. These studies resulted in a new sea-level cyclical chart calibrated on a high-resolution biochronology and the new absolute age scale according to Shackleton et al. (1990), Hilgen (1991a) and Cande and Kent (1992). The chart outlines the major 2nd-order transgressive-regressive facies cycles, and eight 3rd-order depositional sequences and their systems tracts. Some 4th-order sequences were recognized locally inside the 3rd-order ones and dated using the available high-resolution biochronology. The high resolution degree, permitted by a closely spaced grid of Mediterranean bioevents, was possible also in the shelfal record due to the general stratal expansion. The parasequences, in particular, were correlated with good confidence to the climatic oscillations (41 ky periodicity of the orbital obliquity) of the Atlantic and Mediterranean deep-sea record.

The comparison with the Plio-Pleistocene chronostratigraphic chart of the Gulf of Mexico of Wornardt and Vail (1991) showed a general similarity in the number of sequences. Some discrepancies in ages may be due to the different time scales adopted and biochronozones.

The sequence stratigraphic analysis was carried out in a tectonically active area affected by extensional and compressional events occurring during Plio-Pleistocene deposition. Sequence stratigraphic analysis provides some criteria to discriminate tectonic from eustatic effects in the geological record of this tectonically active study area. High-resolution data from small-scale basin analysis supported evidence of the synchronicity of sequence boundaries and individual parasequences, independently of local tectonics. The geologic dataset, including seismic lines and well data, yielded a detailed chronology of the tectonosedimentary evolution of the basins during the last 5 my. According to our seismic interpretation, the alternating extensional and compressional events outlined in the region yield a large-scale common stratigraphic signature, recorded both on and offshore Sicily. These stratigraphic signatures point out the interaction between eustasy and tectonics, confirming the synchronicity of sequences and systems tracts. Our data indicate that local factors, such as sediment supply and tectonics, controlled mainly the stacking pattern development, basin shapes and thickness of the fill, but they do not seem to alter the age and occurrence of sequence stratigraphic surfaces. Subdividing and correlating the sedimentary record on the basis of the stratal pattern and synchronous unconformity surfaces can be an advantageous operative tool in Mediterranean basin exploration when chronostratigraphic controls are lacking; on both seismic and outcrop studies, the chart may predict the ages of regional unconformity surfaces in the Plio-Pleistocene basin fill, particularly where the biostratigraphy has little or no documentation.

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STACKING PATTERN AND TECTONICS: FIELD EVIDENCE FROM Plio-Pleistocene Growth Folds of Sicily (Central Mediterranean)

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ABSTRACT: Syn-sedimentary deformation occurred during late Neogene time in the foreland and piggyback basins of Sicily, located between the South-Tyrrhenian belt and the North Africa foreland. The geometric relationships of the Plio-Pleistocene syntectonic strata (5.3–0.5 Ma) with structures like thrust-related folds are spectacularly preserved, together with the modifications in the stacking pattern of the parasequences. In some expanded successions, it is evident that significant pulses in the depositional growth of structures occur on the time scale of Milankovitch cycles, producing stratal geometries that may become unpredictable by current models. Due to tectonic compressive stresses, the tectonic subsidence (or uplift) rate may exceed the eustatic rate, generating stratal expansions or local unconformity surfaces, formed at the scale of systems tracts of the 4th order depositional cycles. High-resolution biochronostatigraphy and sequence stratigraphic interpretations allow precise evaluations of the vertical absolute growth and growth rates relative to very short time intervals. In the Caltanissetta area, central Sicily, structures have the amplitude of tens of meters are generated within the duration of Milankovitch cycles (5th order). In central and western Sicily, the available data indicate that the duration order of 0.4–1 my seems to be the necessary time interval to complete the growth of structures having the amplitude of some hundred meters; the resulting growth rates are between 0.5 and 2 m/ky. The total amount of uplift within the studied structures, calculated at the 3rd, 4th, and 5th-order of duration, is not much greater than the thickness of the syntectonic strata that onlap the growth antclines bordering the basins; this explains the evidence that the tectonic unconformities are only locally extended. Since individual growth events are unable to produce regional surfaces of erosion during the considered time intervals, the sequence stratigraphic analysis has been profitably applied in this setting of active compression. Regionally-extended unconformities with constant age on wide areas related to glacio-eustasy can be recognized, dated and used for large scale correlations.

INTRODUCTION

The study of middle and large scale compressive tectonic structures, combined with the detailed analysis of the stratigraphic pattern of related syn- and post-tectonic cover, is a very recent field of investigation, especially with reference to the interpretative processes offered by sequence stratigraphy models. Sequence stratigraphy was largely constructed on the basis of seismic analyses of passive continental margins and cratonic basins (Van Wagoner et al., 1988); consequently, the proposed models assumed rates of subsidence constant for time intervals of the order of millions years. Mutti (1989, 1990) objected that in the sedimentary successions deposited in epi- and perisutural basins, where regional and local tectonic deformations occur during deposition, the assumption that the highest rates of eustatic variations exceed those related to subsidence or uplift should be mostly incorrect; therefore, the sedimentary cyclicity recorded in such sedimentary successions, typically expressed by the 3rd order sequences, are not necessarily related to eustatic variations. On the basis of detailed field studies from Spanish Pyrenees foreland basins, Mutti (1990) proposed that very high rates of tectonic uplift associated with the migration of the tectonic hinge line and related equilibrium point, may alter the eustatic-controlled stacking patterns within depositional systems tracts over short periods of time (0.01–0.5 my).

At the same time, Vail et al. (1990) suggested the possibility to distinguish the stratigraphic signature of tectonics, eustasy, sedimentologic, and climatic processes recorded in sedimentary successions, using the sequence stratigraphic analysis coupled with subsidence and tectono-stratigraphic analysis. The fundamental step to address each effect on the sedimentary record to their appropriate cause is to define time-equivalent rock packages that are genetic intervals lacking of significant hiatuses. Each signature can be linked to a specific process when it is understood how each process affects the accommodation space, thus determining their rates and the duration over which these rates are active. For this purpose, Vail et al. (1990) grouped the known stratigraphic signatures in terms of orders of duration. They recognize that tectonism, resulting from a wide range of processes, has the greatest influence on the variations of the accommodation space: 2nd-order cycles (3–50 my), defined by maximum surface of transgression and tectonically-enhanced unconformities, are believed as the ideal tectono-stratigraphic units for analyzing the geologic history of a region. Active deformation, flexural loading variations and uplift, linked to faulting, folding and thrusting may strongly influence 3rd-order episodic events, that are believed as essentially due to glacio-eustasy. Both in 2nd- and 3rd-order cycles, changes in subsidence rates may be shown on the tectonic subsidence curve constructed at the scale of single sedimentary basins. Vail et al. (1990) stated that only such cycles are visible in tectonic subsidence curves, probably due to lack of the stratigraphic resolution of the well data utilized; cycles in the geologic record that are 3rd to 5th order (0.01–0.08 my) are believed to be glacio-eustatic. The smaller magnitude cycles do not clearly exclude the possible occurrence of tectonically-influenced cycles developed at the same order of duration.

Well documented outcrop examples of stratigraphic units containing facies cycles deposited during periods of active deformation are provided by the recent geologic literature, sometimes with opposite conclusions concerning eustatic versus tectonic controls (Riba, 1976; Steel et al. 1977; Anadon et al., 1986; Nichols, 1987; Puigdefabregas and Soquet, 1986; Mutti e Sgavetti, 1987; Blair and Bilodeau, 1988; Sarewitz, 1988; Dolan, 1989; De Boer et al., 1991; Suppe et al., 1992; Dera­mond et al., 1993; Martinsen et al., 1993; among others). Further improvements in the basic models and concepts of sequence stratigraphy have been proposed recently to address the possible variation occurring in response to local physiography, tectonic and sedimentologic conditions (Posamentier et al., 1992; Weimer, 1992; Posamentier and Allen, 1993; Klein, 1993), with the common conclusion that tectonics and glacio-eustasy are concurrent factors in the development of depositional architectures. In the recent geologic literature, the interaction between tectonics and eustasy has been observed at various scales of time and space; however, concerning the basins located around the collisional systems, the studies focused on the sedimentary processes with shorter-term cycles (3rd to
scale duplex geometry, consisting of major Mesozoic carbonate
of the central Mediterranean region (after Catalano et al., 1993c). Key: 1. Sardinia block; 2. Kabilo-Calabride crystalline units;
3. Apenninic-Maghrebian units; 4. Hyblean-Tunisian foreland platform; 5. Superposed basins; and 6. Plio-Pleistocene basins (grey areas) and trend of the major thrust fronts along the Sicily mainland. The dotted line indicate the present day front of the chain, buried below late Pleistocene-Holocene cover.

5th order, 3-0.01 my) and are generally not able to resolve these
problems on the basis of the available chronostratigraphic data.
In other words, the chronostratigraphic resolution of the
adopted methods or available data are usually significantly
lower than the physical-stratigraphic resolution, and ultimately,
the application of models based on regional analyses of seismic
stratigraphy remains problematic if adopted at the scale of out-
cropping structures and genetically-related stratigraphic units.

This paper aims to illustrate examples of tectonic structures
some hundreds of meters in size, whose kinematic development
may be scanned at the duration of the 4th- and 5th-order cycles
using the Pliocene high-resolution biochronology, stratal pat-
tern and facies analysis, and sequence stratigraphy interpreta-
tive tools. The excellent exposures allow direct observations of
the physical stratigraphy of the syntectonic deposits, giving the
possibility to reconstruct the evolution of the structures at vari-
ous stages of growth. The common growth rate found in the
area ranges between 0.5 and 2 m/ky, with total amount of ver-
tical uplift of some hundred of meters. The stratal geometries
and the available chronostratigraphic constraints allow to ad-
dress the driving forces that influence the development of the
parasequences in a tectonic setting of active compression. Con-
siderations about the interplay of eustasy, climate and tectonics
are proposed relative to the development of the shorter-order
facies cycles (parasequences).

GEOLOGIC SETTING

The study areas are located in western and central Sicily, the
major island of the central Mediterranean (Fig. 1A) linking the
Apennines of southern Italy and the Tunisian-Maghrebian chain
(Catalano et al., 1993b, c). Several Plio-Pleistocene satellite
basins, partly covering the active structures of the belt, outcrop
between the hinterland northern chain emergent along the Tyr-
rhenian margin, and the foreland basins of the southern offshore
(Fig. 1B). A simplified sketch of the belt (Fig. 2) shows a large-
scale duplex geometry, consisting of major Mesozoic carbonate
structures emplaced below a Tertiary roof-thrust complex of
incompetent rock allochthons. This complex of thrusts rep-
resents the shallower and younger tectonic element of the Sicilian
belt; its emplacement is believed to be contemporaneous to the
accretion of Mesozoic duplexes at deeper levels (Catalano et
al., 1993b). In the central and southern part of Sicily, the evi-
dences of synsedimentary tectonics within the lower and middle
Miocene strata are generally obscured or lacking. Between the
latest Miocene and the early Pleistocene, the deformational pro-
cess was still largely active and resulted in both high-amplitude
fold systems developed in the piggyback basins bottom and
emergent thrusts. Between the late Pliocene and the Early Pleis-
tocene, the front of deformation shifted some tens of kilometers
from central Sicily to the southern offshore area, where it is
thought to be still active. The surficial effects of the compres-
sive tectonics, expressed by marked increase of the tectonic
subsidence or uplift, are spectacularly recorded in the coeval
sedimentary cover and enhanced by high rates of sediment sup-
ply and sea-level oscillations. The study areas satisfy the fol-
lowing conditions believed fundamental to better understand
the growing processes of the synsedimentary compressive
structures during very short time interval:

1. The deformational processes are very recent. This enable to
analyse relatively simple tectonic structures and related syn-

Fig. 1.—(A) Tectonic scheme of the central Mediterranean region (after Catalano et al., 1993c). Key: 1. Sardinia block; 2. Kabilo-Calabride crystalline units;
3. Apenninic-Maghrebian units; 4. Hyblean-Tunisian foreland platform; 5. Superposed basins; and 6. Plio-Pleistocene basins (grey areas) and trend of the major thrust fronts along the Sicily mainland. The dotted line indicate the present day front of the chain, buried below late Pleistocene-Holocene cover.

Fig. 2.—Schematic cross section of the western Sicily illustrating the structural setting of the orogenic system. Note the large scale duplex geometry of
the chain and the back-arc extensional system related to the opening of the
Tyrrhenian basin. Key: 1. Major duplexes of Mesozoic carbonates; 2. Tertiary
roof-thrust complex; 3. Late Neogene syntectonic basin fill; and HB, PB and
FB are respectively: hinterland, piggyback and foreland basins. The examples
reported in the following text are referred essentially to piggyback basin fill.
tectonic stratigraphic reference scale that is now the most accurate within the global geologic time scale. Several detailed bio-
chronostratigraphic studies, conducted in the area during the
last ten years, allowed time resolutions on the order of 0.1
to 0.02 my, integrating quantitative planktonic biostrati-
graphy, magnetostratigraphy, and cyclostratigraphy (e.g., Rio et
al., 1984, 1991, 1994; Rio and Sprovieri, 1986; Zachariasse
et al., 1989; Hilgen 1991a, 1991b; Langeris and Hilgen,
1993; Channel et al., 1992; Sprovieri, 1992, 1993; Hilgen
and Langeris, 1993; Di Stefano et al., 1993b).

2. It is possible to analyze well exposed sections with sufficient
lateral continuity which have been subjected to a very recent
strong uplift. This condition is generally not verified along
the front of recent chains, since they are areas that are nor-
mally strongly subsiding due to flexural loading. In Sicily,
the more recent syntectonic sediments (5.3-1.2 Ma), depo-
sited within piggyback and foreland basins, were soon after
involved in the shallower fold and thrust systems of the belt.
Accompanied by normal faulting, they are now uplifted in
some cases more than 1500 m above the sea level with spec-
tacular preservation of the stratigraph pattern.

3. Sequence stratigraphic subdivision for the Plio-Pleistocene
was constructed using the Berggren et al. (1985) time scale
(Catalano et al., 1992a, 1993a; Di Stefano et al., 1993a); this
is updated in Catalano et al. (this volume) with reference to
the time scales of Shackleton et al. (1990) and Hilgen
(1991b). This new chronostratigraphic chart (see Catalano
et al., this volume, Fig. 3) reports the accurate age-evaluation
of 3rd-order depositional sequences and their systems tracts
(4th-order simple sequences). Sequence stratigraphic sur-
faces have been found to have regionally constant ages in
different depositional and tectonic settings which indicate
that they are globally synchronous since they are linked to
glacio-eustatic oscillations (for details, see Catalano et al.,
this volume). The chart may be also used as a field reference
in order to distinguish the effects due to sea-level changes
from those due to the local tectonics. The following exam-
ple, coming from the Belice depositional sequence (4.2-3.0
Ma) and from the Capodarso depositional sequence (2.5-2.1
Ma), illustrate how each variation in the stratigraph pattern
may be addressed to its proper cause within 3rd-, 4th- and
5th-order cycles, even where unconformities and tectonic e iso-
V 2 - sional truncations merge into a complex surface.

STRUCTURES GROWTH AT THE SCALE OF THE 4TH- AND 5TH-ORDER CYCLES

The Caltanissetta Basin

The Caltanissetta basin (Figs. 1A, 3) located in central Sicily,
is a late Neogene depression developed above a Tertiary roof-
thrust complex and filled by Plio-Pleistocene deposits. In this
area, the major duplex structures of Mesozoic carbonates are
known by exploration wells to be present at some thousand of
meters of depth. The simplified section on Figure 2 may be
used to illustrate the structural setting since the geologic lit-
terature lacks deep regional profiles across the area; according
to the sketch, the Caltanissetta basin may be considered a wide
piggyback basin bordered on the south by emergent carbonate
thrusts. The structural analysis of the area between Enna and
Caltanissetta (Fig. 3) has pointed out the occurrence of major

Messinian-lower Pliocene thrusts and associated folds trending
nearly East-West. These structures are covered by middle Pli-
ocene deposits that are involved in a fold system trending
Northeast-Southwest. Both these systems are clearly syntec-
tonic and consist of major asymmetric fault-propagation folds
and other associated minor symmetric folds having amplitudes
in the order of 250–400 meters and a wavelength of a thousand
meters; strata younger than 2 Ma appear unfolded or less
deformed.

The Pliocene succession (Fig. 4A), overlying a group of late
Miocene siliciclastic units, is represented at the base by the
pelagic chalks of the Trubi Formation (5.3-3.2 Ma) deposited
over Sicily during the Mediterranean flooding that followed the
late Messinian desiccation. These chalks pass upward paracon-
formably into the Enna Formation, consisting of hemipelagic
shales at the base and slope to offshore siltstones and sandy
mudstones toward the top. The Enna shales are truncated by
two sets of coastal biocalcareous bodies (Capodarso and Lan-
nari Formations, about 50–80 m thick each) separated by a shelf
mudstone interval (Gerace Formation, about 100 m thick),
(Roda, 1967a, b). The lowermost sandy interval of the Capo-
darso Formation shows a marked progradational pattern, re-
cording the Middle Pliocene southeasterly shift of the coastline.
Recent studies integrating data of facies analysis, physical stra-
graphy and high-resolution biostratigraphy, have provided a
sequence stratigraphic interpretation of this part of the Pliocene
succession (Catalano et al., 1992a, b, 1993a). A complete 4th-
order depositional sequence (here named Capodarso Sequence)
and its relative systems tracts have been recognized (Fig. 4B);
it is defined between the bottom of the Capodarso Formation
and the bottom of the Lannari Formation. In particular, the Ca-
podarso Formation has been interpreted as a lowstand prograd-
ing complex stacked in 6 parasequences. Catalano et al. (1992b)
also dated the unconformity found at the bottom strata of the

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**Fig. 3.**—Geologic map of the Caltanissetta-Enna area (after Bigi et al., 1991). S1: approximative trace of the profile of Figure 7.
Capodarso calcarenites at 2.4 Ma with reference to the Berggren et al. (1985) time scale. This regionally occurring unconformity is approximately the boundary between the D. pentaradiatus and the D. brouweri calcareous nanofossils biozones (Catalano et al., this volume, Fig. 3). The same unconformity is now dated at 2.5 Ma, according to the more recent astronomically-calibrated time scale of Hilgen (1991b) adopted in this paper, and correlated to the isotopic stage 100 of Raymo et al. (1989). The Capodarso Sequence, coincident to the P.4.c sequence of the Plio-Pleistocene cycle chart (Fig. 3), is bounded at the top by the 2.1 Ma unconformity closely placed to the Globorotalia inflata FO bioevent (2.07 Ma according to Sprovieri, 1993).

The Capodarso Sequence: Tectonic and Eustatic Controls

Since the Capodarso strata were deposited in a tectonically active basin, any sequence stratigraphic interpretation must discuss the arguments in favour of a tectonic or eustatic cause for the occurrence of the lowermost sequence boundary. It is important to take into account the possibility that an abrupt tectonic uplift of the syntectonic structures may have produced a regional basinward shift of the paleo-shoreline.

The clearest evidences of active compression affecting the deposition of the Capodarso sequence are: (1) occurrence of Northeast-Southwest trending fold system, whose orientation differs from Messinian-early Pliocene structures (Fig. 3); (2) an angular unconformity of 15°–25°, occurring between the Capodarso and the overlying unfolded Lannari strata (see Monte Capodarso and Monte Carangiaro sections); and (3) the occurrence of local erosional truncations and varitions of the thickness in the strata clearly associated with the folds growth (see Monte Salsello and Monte Juculia sections).

There are other arguments in favour of an eustatic origin of the lower boundary of the Capodarso depositional sequence. The available sequence-stratigraphic data for the central Mediterranean area (Catalano et al., 1992 a, b, 1993a) indicate the occurrence of a pronounced sea-level drop at 2.4 Ma, which may be correlated to the Plio-Pleistocene sea-level cycle chart of the Gulf of Mexico (Wornardt and Vail, 1991) with absolute age calibrated using the Berggren et al. (1985) absolute time scale. According to Raymo et al. (1989), the 2.4 Ma event (chronologically equivalent to 2.5 Ma according to the more recent time scale of Hilgen, 1991b) is probably related to an important climatic change due to the increase in northern hemisphere ice sheet volume forced by orbital variations; prior to 2.4 Ma the variations in ice sheet volume appear to be, on average, 1/4 to 1/2 as large as those of the late Pleistocene, whereas after 2.4 Ma, these variations seem to be about 1/4 with respect to the late Pleistocene ones. According to Posamentier et al. (1992), the seaward shift of facies tracts and shoreline regression in response to sea level lowering, termed “forced regression”, is independent from sediment supply and is commonly associated with zones of sedimentary bypass and subaeiral exposure. These features also are present in the Capodarso Formation where subaeiral exposure, enhanced by the tectonics, eroded the top of major substrate structures.

On the basis of the regional correlations, the Capodarso Sequence is thought to be the sedimentary expression of a 4th-
order glacio-eustatic cycle, with a duration of about 400 ky (2.5-2.1 Ma) comparable to the typical period of long eccentricity oscillations. As illustrated by the following examples, the local tectonic effects on the deposition may be evaluated looking at the modifications in the stacking pattern of parasequences and systems tracts; these observations determine the growth rates and absolute amount of uplift considered on different orders.

**Type-Parasequence in the Monte Capodarso Area**

The detailed examination of the sedimentological and stratigraphic features of the Capodarso sequence is beyond the purpose of this paper. However, the collected field data in the Caltanissetta area of central Sicily suggest some considerations about the parasequences as sedimentary bodies developed on continental shelves with mixed carbonate-siliciclastic sedimentation. The most interesting aspects are related to their cyclical occurrence and to their stacking patterns.

Several contributions or reviews have pointed out the main sedimentological and stratigraphic characters of the coastal sand bodies, as well as the shallowing upward or deepening upward facies trend and their arrangement in forestepping or backstepping sets (e.g., Campbell, 1971; Howard, 1972; Reineck and Singh, 1973; Ghibaudo et al., 1974; Hunter et al., 1979; Clifton, 1981; Mc Cubbin 1981; Elliott, 1986; McCrory and Walker, 1986). Other significant contributions provided data and improved the conceptual framework about the coastal responses to eustasy, subsidence and sediment supply, or interpreted the coastal sedimentary processes in a sequence stratigraphic key (Swift et al., 1971; Swift, 1976; Vail et al. 1977; Plint, 1988; Walker and Eyles, 1988; Dolan, 1989; Walker, 1990; Field and Trincardi, 1991; Posamentier et al., 1992; Weimer, 1992; and reference therein). In the Exxon systematics (e.g., Van Wagoner et al., 1988, 1990), the parasequences are sedimentary bodies some meters or tens of meters thick, bounded by flooding surfaces. According to Mutti (1989, 1990), parasequences, typically gradationally-based and shoaling upwards cycles, form downwip of the equilibrium point, whereas updip typically sharp-based sequences form (the “simple sequences” of the Exxon systematics).

The field observations and data from the Monte Capodarso area seem to confirm the alternative scheme proposed by Mutti (1990, Fig. 3 on p. 636) and the recent improvements to sequence stratigraphic concepts and models provided by Posamentier and Allen (1993). Each parasequence of the Capodarso succession (Fig. 5A) shows at the bottom a downlapping bioclastic grainstone wedge (calcarenite) showing reverse gradation and cross-stratification. This prograding wedge is covered by a set of sandy strata showing landward onlap terminations and a plain-parallel stratification. This set progressively thins landward up to a point of by-pass of the sediment facies, here informally called starting point of the sand wedge (Fig. 5B).

The two different sets of sandy strata, representing short-term regression and transgression respectively (landward forestep and backstep of the facies associations), have been interpreted as nearshore deposits. The sandy wedges are covered with a sharp contact by grey siltstones containing molluscs (in-situ fauna) and interbedded fine sand sheets. The flat flooding surface separating these bodies clearly marks an abrupt deepening of the sedimentary facies from lower shoreface or subtidal to offshore/transition. A relevant feature of the grainstone wedges is the common occurrence of ravinement surfaces of erosion that truncate the toplap terminations of the prograding strata (Fig. 6A). In these cases, a shell lag supported by fine sand or siltstone matrix occurs between the ravinement truncation and the overlying plain parallel bedded calcarenite set (Fig. 6B).

The particles of the calcarenites are formed by medium to coarse detritus of shells, bryozoans, echinoderms, algae, corals, benthic foraminifera and other shallow-water organisms. Also in the finer sands, the quartz content does not exceed the 20%; therefore, the bulk of the prograding grainstone wedge is composed by bioclastic materials derived from coastal environments and gradually downlapping towards the outer shelf. Since prograding sand bodies mainly consists of locally-derived sediments, the fluctuations in bioclastic sediments availability may be also correlated to biogenic productivity variations which accompanies climatic and sea-level high-frequency oscillations (Milankovitch cycles). Consequently, all these parameters may be considered as concurrent factors in the development of the shelfal parasequences. These sedimentary bodies, consisting of a two-phase transgressive-regressive facies cycle (Fig. 5), are generally well preserved within the Plio-Pleistocene shelf successions of the study area; normally, they have very good lateral continuity, allowing useful middle-scale physical correlations.

**Stacking Pattern at the Monte Capodarso Section**

The Capodarso strata filled a 10 Km-wide syncline growing between two major fault-propagation anticlines (Fig. 7). In particular, the core of northwestern anticline bordering the basin, between Monte Capodarso and Monte Pasquasia (Fig. 3), displays some minor horses of the roof-thrust complex. The hinge of the major anticline has been eroded since the time of deposition of the Capodarso lowstand strata, as demonstrated by the large-scale stratral pattern of the syntectonic fill. In addition to syntectonic structures developed at the scale of 4th-order depositional sequences, the Capodarso section shows spectacular
FIG. 6.—(A) Plane parallel set of strata (shoreface deposits), underlain by a matrix-supported shell lag, covers a transgressive surface of erosion (ravinement surface). (B) Sketches illustrating the occurrence of ravinement erosion during a short-term transgressive phase; the prograding packstone s, having truncated toplap terminations, are covered by a shell lag (storm deposits) and by a plain-parallel bedded shallow-water calcarenites, deposited during a short-term transgression.

FIG. 7.—Depth reconstruction of the Capodarso basin between Caltanissetta and Pietraperzia areas (central Sicily; see trace in Fig. 3) at the time of the deposition of the Lannari sequence (late Pliocene). The reconstruction was performed integrating original surface data in the northern sector with subsurface data from Roda (1967b) in the southern zones. Note the difference between the synsedimentary character of the Enna complex, associated with surficial thrusting, and the Capodarso-Lannari complex mainly associated to fault-propagation folding. The central part of the profile is illustrated in Figure 9.
Fig. 8.—Panoramic view of the Monte Capodarso, west-northwest flank (a). The Capodarso Formation represents a lowstand prograding complex developed after the 2.5 Ma relative sea-level drop, spectacularly stacked in parasequences (symbols as in Figs. 4, 9). The facies association consists of a set of shallow-water bioclastic packstone wedges overlying the slope and offshore mudstones of the Enna Formation. The Capodarso packstones laterally pass and evolve upward into offshore bioturbated silstones and sandy mudstones (Geracello Formation). The lateral continuity of these bodies for tens of kilometers, and the occurrence of the parasequences in a constant number over the whole area suggest possible relationships with Milankovitch cyclicity (see discussion for details). In (b) a detail of the base of the Capodarso Formation and the lowermost prograding calcarenite bodies is shown.

Fig. 9.—Geological section of the Monte Capodarso traced perpendicularly to the dip of the major structure present in the substrate (trace located on the central part of Fig. 3). The stacking pattern of each systems tract is largely influenced by strong lateral variations of the subsidence rates, due to the active deformation.

tract), it is evident a relative stillstand occurred showing no significant lateral shift. The flank of the structure probably only rotated around that hinge point, generating an aggradational fan in the syntectonic strata; this pattern points out a strong lateral variation of the subsidence rate, dramatically increasing basinward during that period. Finally, PCd 7 represents a clear landward shift of the shoreline followed upwards by a condensed section and by thick highstand deposits southwards (PCd 8 and 9), preserved only in their outer shelf deposits.

**Stacking Pattern at the Monte Carangiato Section**

The Monte Carangiato outcrop, located in the Pergusa Lake area (Fig. 10) close to the town of Enna about 20 Km to the east of the Monte Capodarso section, provides a second example of growing structures covered by the Capodarso strata. The structural analysis shows a system of mainly symmetric folds trending N40°E with a subvertical axial plane. These folds, involving the Capodarso strata, show an amplitude on the order of 250 m (considered from the base of the Capodarso Formation) and wavelength of 2.5 km. The Geracello and Lannari Formations (Fig. 4) outcrop a kilometer to the southeast (not shown in the profiles) and represent the unfolded post-tectonic deposits. Along the strike of the northern major fault-propagation anticline, it is possible, with some difficulty, to correlate the parasequences of the Monte Carangiato-Scioltabino section with the parasequences outcropping at the Monte Capodarso section. The base of the Capodarso depositional sequence is still controlled using the *D. pentaradiatus* LO bioevent (2.51 Ma, according to Sprovieri, 1993); consequently system tracts and their parasequences are correlated on the basis of both physical stratigraphy and biostratigraphy.

In order to describe the stacking pattern, I have assigned to the parasequences the prefix “PCr” followed by progressive
numbers. The Monte Carangiaro section (Fig. 11A, B) is oblique to the tectonic strike and displays 8 parasequences: the PCR 0 represent an incised valley fill (lowstand systems tract); the parasequence-set composed of PCr 1 to PCr 5 represents the lowstand prograding complex, with varying stratal patterns depending on its relative position to the growing tectonic structures; parasequence PCr 7 represents the transgressive systems tract. Note that the PCr 4 may simulate the base of a transgressive systems tract outcrop a few kilometers to the southeast and are represented by outer shelf silstones and mudstones.

Stacking Pattern at the Monte Salsello-Serieri Valley Section

Along the valley of the Torrente Serieri (Fig. 13), culminating at the foot of the Monte Salsello, a set of parasequences (here labeled PVs) is exposed, from the Capodarso Formation. This outcrop occupies an intermediate position between the Monte Carangiaro and the Monte Capodarso sections (Fig. 10). Two different tectonic trends have been found: the first set of east-west trending structures, is related to tectonic events pre-dating the 2.5 Ma basal unconformity of the Capodarso sequence. At north and northeast of the Monte Salsello, an overturned succession of Late Messinian evaporites overlying the Trubi Formation are interpreted as the truncated limb of a south-vergent ramp anticline associated with a shallow-sited thrust. The second set of structures, northeast-southwest striking, were generated during the deposition of the Capodarso strata, as demonstrated by the occurrence of a local erosional truncations within the lowstand prograding complex. This unconformity, sketched on the profile of Figure 13, is well exposed; a quasi planar set of strata cover the folded and eroded complex of the lowermost parasequences of the Capodarso Sequence. This section shows a minor anticline, developed since deposition of the Enna Formation, that has been partly inverted as a strongly subsiding syncline, allowing the deposition of a thick early lowstand unit. A new tectonic pulse, tilting, uplift and consequent erosion occurred, followed by a new tectonic inversion to the south, with the deposition of the late lowstand southward expanded strata, thin transgressive and, more to the south, thick highstand systems tract deposits.

The Salsello-Serieri section displays a stacking pattern made up as follows. The parasequence PVs 0 fills some small incised valleys in the deformed and eroded substratum. The set composed of PVs 1 to PVs 3 shows the maximum rate of progradation, interpreted as oblique offlap on the basis of the distribution of the offlap break points. The PVs 4 to PVs 6 group shows an evident aggradational fan, with the starting points shifting southward, and interpreted as the maximum increase of the subsidence developed in a growing fold limb. Finally, PVs 7 represents the transgressive systems tract and shows at the base a planar ravinement surface (transgressive surface of erosion). Note the similarities with the stacking pattern displayed by the Monte Capodarso section, especially between the late lowstand systems tracts and the transgressive systems tracts.

Stacking Pattern at the Monte Juculia Section

The Monte Juculia anticline (Figs. 10, 14) allows physical correlations between the strata filling the Serieri and Scioltabino synclines. Here the parasequences associated with a late lowstand and the transgressive systems tracts outcrop. The stacking pattern (Fig. 14), reconstructed on the basis of surface and well data, delineates a succession of different growth episodes of the Juculia anticline which control the development of minor tectonic unconformities as shown by local truncations of the strata (LTU). The first episode of growth predate the deposition of the PVs 3-PCr 3 parasequences: the southerly vergent major fault-propagation anticline, here buried below the filling of the Serieri syncline (NW side of the profile), was strongly uplifted and eroded within the hinge zone (see also Fig. 13). A second episode of folding and uplifting of the Juculia anticline was accompanied by subsidence increase and flank tilting of the Serieri syncline, up to the deposition of the parasequence PVs 7. The uplift of the anticline appears exaggerated by the thickening of silstone strata present in the hinge of the structure.
this indicates that the present-day axial plane of the Juculia anticline was located in a zone where a subsident syncline was growing during the deposition of the early lowstand systems tract. The synsedimentary growth of the Carangiaro anticline (southeastern side of the cross section of Fig. 11B) caused a progressive northwesterly shift of the axis (depocenter line) of the Scioltabino syncline, while the axis of the Juculia anticline (Fig. 14) was remaining about in the same position. The physical correlations between the two flanks of the Juculia anticline, marked in the profile by the thin dashed lines, are made possible by the continuity of the strata existing in the northern zones close to the trace of the section.

Field evidence (see Discussion paragraph) and paleocurrent data (Fig. 10) indicates that the direction of progradation of the
STRUCTURE GROWTH AT THE SCALE OF THE 3rd-ORDER CYCLES

The Belice Piggyback Basin

In order to have a comparison with the vertical growth rate obtained in the Caltanissetta Basin, I examined the relationships between Pliocene sedimentation, eustacy and tectonics of western Sicily, considering that the growth of structures occurred on greater scale. In the Belice River area, the synsedimentary growth of a major fault-propagation fold, occurring mainly in deep-water conditions, is constrained by 3rd-order sequence stratigraphic surfaces, their relative ages given by biostratigraphy, and their absolute ages as inferred by the comparison with the Plio-Pleistocene sequence stratigraphic chart (Catalano et al., this volume).

Recent field studies carried out in western Sicily pointed out that the Plio-Pleistocene strata deposited within piggyback basins, mainly developed over advancing Mesozoic thrust sheets (Vitale, 1990; Di Stefano and Vitale, 1993). The Belice Basin developed since late Messinian time above the Monte Magaggiaro thrust sheet; the main ramp anticline of the mesozoic subcrust outcrops 20 Km southward, close to the southern coast of Sicily (Fig. 15). The Pliocene-type succession is very similar to that previously described for the Caltanissetta area, except for the occurrence of a thick interval of turbidite sandstones and siltstones over the Trubi Formation (Fig. 15C). The turbidites and part of the overlying open shelf siliciclastic and carbonate clastic coastal deposits are locally known as “Marnoso-Arenacea del Belice” Formation (Ruggieri and Torre, 1974) and largely exposed in the northern and central sector of the basin. More recently, Di Stefano and Vitale (1988) provided a biochrononstratigraphic evaluation of the depositional interval of the Marnoso-Arenacea del Belice Formation (from herein shortly...
Fig. 15.—(A) Structural map of the Belice Basin, southwestern Sicily (From Di Stefano and Vitale 1993). The dashed line indicates the trace of the cross section (B). The inset shows the approximate location of the profiles reported in Figure 16. (C) Stratigraphic column of the Upper Miocene-lower Pleistocene succession outcropping in northern sector of the Belice Basin, along the Poggioreale ridge.

MAB); Vitale (1990) pointed out the synsedimentary character of the tectonic structures bordering the Belice Basin, recognizing some syntectonic unconformities.

The Poggioreale Structure

Along the ridge of Poggioreale, a complex set of structures is exposed. They consist mainly of fault-related folds and minor thrust faults probably associated with a deeper-seated thrust involving the Tertiary cover. In the area (Fig. 15), the first evidence of compressive tectonics occurred in latest Miocene time, as the base of the Trubi chalks (lowermost Pliocene) truncated previously formed units including late Messinian evaporites.

On both sides of the Belice River gorge, a few kilometers east of the Poggioreale ruins, two major anticlines are exposed, and separated by a narrow south-southeast verging syncline (Fig. 16). These folds show a maximum wavelength of 1 Km close to Pizzo di Gallo. The structures continue westward to lower wavelength anticline crests separated by tight syncline,
overlying brown siltstones and it is a paraconformity between the Trubi white chalks and the boundary has been dated at Pizzo di Gallo hill (Fig. 17), where a number of the MAB Formation (Figs. 15, 17). The 4.2 Ma sequence syntectonic cover consists of the lower and intermediate members of the MAB Formation (Figs. 15, 17). A thrust-related fold system is present toward the east (cross-section 1), gradually passes westward into a minor surflicial thrust (cross section 2 and 3) involving part of the syntectonic cover. Syntectonic strata consist of the basal part of the depositional sequence Belice 1 (4.2–3.0 Ma, lower member of the MAB Formation). The post-tectonic set is the prograding lowstand complex related to the 3.0 Ma sea-level drop, here represented by the shallow-water packstone s of the upper member of the MAB Formation, truncating landward the structures at Pizzo di Gallo. The section (4) illustrates data and geometrical elements used to constrain the vertical growth rate of the structure.

Fig. 16.—Structural variation along the strike of the Poggioreale structure, illustrated by a set of three parallel profiles (approximate location of the traces in Fig. 15). A thrust-related fold system is present toward the east (cross-section 1), gradually passes westward into a minor surflicial thrust (cross section 2 and 3) involving part of the syntectonic cover. Syntectonic strata consist of the basal part of the depositional sequence Belice 1 (4.2–3.0 Ma, lower member of the MAB Formation). The post-tectonic set is the prograding lowstand complex related to the 3.0 Ma sea-level drop, here represented by the shallow-water packstone s of the upper member of the MAB Formation, truncating landward the structures at Pizzo di Gallo. The section (4) illustrates data and geometrical elements used to constrain the vertical growth rate of the structure.

up to become ramp anticlines having overturned fold forelimbs to the south. This evidence indicates that these are fault-propagation folds, with the displacement of a partly covered thrust decreasing updip and laterally. The continuity of the structure away from the Belice gorge is generally difficult to reconstruct because of the excessive thickness of the Pliocene cover; however, the synsedimentary growth processes during the deposition of the MAB Formation are documented mainly using 3rd-order sequence boundaries present in the syntectonic strata. The sequence stratigraphic interpretation allows recognition of two complete depositional sequences (Belice 1 and Belice 2 in the chronostratigraphic chart of Catalano et al., this volume, Fig. 3) having boundaries approximating the ages of 4.2, 3.0, and 2.7 Ma (see also Catalano et al., this volume); the Pliocene syntectonic cover consists of the lower and intermediate member of the MAB Formation (Figs. 15, 17). The 4.2 Ma sequence boundary has been dated at Pizzo di Gallo hill (Fig. 17), where it is a paraconformity between the Trubi white chalks and the overlying brown siltstones and fine sandstone of the MAB lower member. The lack of the R. pseudouni/bilica biozone (Di Stefano and Vitale, 1988) indicates a hiatus of about 0.25 My. This hiatus within a deep-sea succession can be explained by looking at a second outcrop, located on the northern flank of the Pizzo di Gallo hill, where the vertical limb of the major fold outcrops (Fig. 17). Here an angular unconformity between the Trubi and the lower MAB Formations is present (Fig. 18), but the hiatus has the same length of time. Analysis of the stratigraphic pattern on a local scale suggests a tectonic cause for the unconformity; the pelagic chalk of Trubi, the topmost strata of the growth-fold, are truncated by an unconformable surface of erosion and are still overlain by deep water strata having onlap terminations (Fig. 18). Part of the unconsolidated fine-grained materials of the Trubi Formation were removed by submarine slides and slumps down the tilting fold limb, leaving upslope scar surfaces. The lowermost strata of the MAB Formation contain blocks of Trubi and reworked faunas from older formations, suggesting compressional stresses active mostly during the time interval represented by the missing biozone. The upper member of the MAB Formation, characterized by a prograding lowstand complex of shelfal deposits, overlies an unconformity surface truncating the structure, clearly post-dating the last relevant growth event of the structure (Figs. 16, 17). This uppermost unconformity (3.0 Ma), marked by an abrupt change to shallow-water deposition, was interpreted as due to eustasy, because it was regionally recorded under different paleobathimetric conditions and tectonic settings. It must be noted that the times corresponding to the 4.2 and 3.0 Ma sequence boundaries are regionally marked in the pelagic record of Sicily by a decrease in carbonate content and increase in clay or sand content (Langereis and Hilgen, 1991; Catalano et al., this volume), suggesting an eustatic sea-level drop. Therefore, in this case, the effects produced by a tectonic event concur with the eustatic effects to form a peculiar stratigraphic signature (tectonically-enhanced sequence boundaries). The physical correlation between the main 3rd-order unconformities, the biochronological constraints and the recognition of the major lithofacies associations allows the evaluation of what is the mean growth rate in two different time spans:
1. considering the whole 1.2 my interval, the duration of the Belice sequence, a slow and constant vertical growth is hypothesized; and
2. taking into account the interval of 0.25 my, that is the duration of the hiatus of the basal unconformity, the possibility of a quick and discontinuous uplift is considered.

The available data still indicate that the component of vertical growth of 4.2 Ma was insufficient to produce subaerial exposure of the major anticline. The average paleobathymetry of the Trubi strata is inferred in the order of 700 m by the benthic fauna content; also the lower MAB member shows facies and faunal evidences of persisting deep-water conditions. Considering size and geometry of the structure (last profile of Fig. 16), the estimated average rate of uplift of the structure during the considered interval is calculated between 0.5 and 2 m/ky; this evaluation also may in agreement with higher rates of horizontal displacement.

DISCUSSION

Stacking Pattern and Tectonics

In the study area, the stacking pattern variations point out important correlations between tectonics and the development of parasequences. An indirect evidence for tectonic control can be presented by the generation and degree of preservation of regressive downlapping coastal deposits, believed uncommon without some fundamental conditions. According to Field and Trincardi (1991) these conditions are: high gradient of shelf morphology and expanded thickness, normally occurring in areas of high rates of subsidence and sediment supply. The development of thick parasequences and their high degree of preservation is thought to be related to strong subsidence affecting the syncline where the Capodarso strata were deposited. Field
observations point out an interplay among tectonics, deposition or erosion evident at the scale of parasequences and systems tracts (Fig. 19). Compressive tectonics controlled local changes in subsidence, producing significant variations in the stacking pattern observed across adjacent structures or sectors of an individual structure. Stratal geometries of the northeast-southwest trending fold system fill demonstrate that the growth of the structures occurred following tectonic pulses. These pulses are responsible for small vertical increments of tectonic subsidence during intervals that may be shorter than a single parasequence. The stacking pattern is often comparable to the stratal geometry termed by Riba (1976) as “progressive syntectonic unconformities”; however, examples like the Monte Juculia and the Monte Carangiaro structures (Figs. 11–14) revealed more complex patterns probably related to tectonic inversion. As observed in some cases, fold axes shift laterally for distances on the same order as folds wavelengths, allowing the structural inversion of some synclines. Associated to the structural inversion, other examples have shown that some anticline limbs (e.g., Monte Salsello, Fig. 13) were first subaerially eroded during uplift and then strongly tilted and subsided.

As previously discussed, tectonics appear to be the most significant factor controlling the stacking pattern at the scale of parasequence sets in a setting of surficial active compression. The tectonic uplift may generate local erosional surfaces, but it does not affect the internal stratal geometry and the rhythmic occurrence of individual parasequences. In particular, taking into account the constant number of parasequences within the Capodarso depositional sequence (see also Table 1), I suggest that local tectonics cannot obliterate their natural sedimentary expression as cycles having large-scale extent. Erosion or lateral stratal expansions, and consequently small modifications in the seaward or landward shifting of the shoreline, may affect the normal development of the systems tracts. In a setting of active tectonics, the systems tracts, recognized in different places only on the basis of the stacking pattern, have to be cautiously considered as “local systems tracts” where the chronostratigraphic controls have poor resolution. Since the landward and seaward shifts of the shoreline within the same depositional sequence cycle may be strongly influenced by the local tectonics, the systems tracts due to pure eustasy may differ substantially from those interpreted from the stratal geometry, leading the geologist to uncorrect correlations between not synchronous set of strata. For these reasons, the studied sections system tracts and parasequences have been recognized and correlated on the basis of both physical stratigraphy and biochronology.

The Parasequences: Tectonic Versus Eustatic Control

Although the compressive stress is a very important factor controlling stacking patterns, the rhythmic occurrence of the parasequences is still hard to be explained as due only to the tectonics, simply because cyclical variation of the paleobathymetry constant across different structures or sectors of structures should be required. On the contrary, the total subsidence is generally observed changing from axes of synclines (depo-center lines) to axes of anticlines (hinge lines of uplift). Active tectonics should also influence the downlap direction of individual parasequences, which expresses the shoreline sense of progradation. On the contrary, the Capodarso parasequences

![Diagram of parasequences and tectonic uplift](Fig. 19)
show a wide stratal continuity across different structures, for kilometers along the downlap direction and for tens of kilometers along the paleo coast line. Moreover, the downlap direction is quasi-constant southward, indifferently across each growth structure (see Fig. 10). This evidence suggests that the occurrence of parasequences and systems tracts are synchronous and independent from individual features, being probably related to larger-scale factors (climate, sediment supply, high-frequency sea-level changes).

The solution of the problem of shelfal facies cycles can be aided by correlations with the coeval pelagic record, generally well controlled by an higher resolution biochronology. Modern Mediterranean Plio-Pleistocene stratigraphy improved significantly the resolution power of biochronology and demonstrated that the occurrence of the limestone-marl couplets within the pelagic/hemipelagic successions reflects Milankovitch cycles that are mainly related to climatic oscillations forced by astronomical factors (Hilgen, 1991a, b; Langereis and Hilgen, 1991; Rio et al., 1991; Thunnel et al., 1991; Channel et al., 1992; Sprovieri, 1992; Di Stefano et al., 1993b; Rio et al., 1994). Some of these studies also demonstrated that abundance peaks of some planktonic taxa, synchronously occurring with the lithological variations, can be successfully correlated with the oxygen isotope stages recognized on Atlantic ODP sites. The interval of duration of the pelagic lithological cycles have been carefully evaluated improving the stratigraphic resolution. In particular, the Pliocene record in the northern Atlantic (Raymo et al., 1989) as well as in Sicily (Hilgen, 1991a; B; Langereis and Hilgen, 1991; Sprovieri, 1992) displays dominantly the astronomical precession periodicity in the basal part, until about 2.8 Ma, and obliquity periodicity between 2.8 and 1.6 Ma.

On the basis of the exposed data and similar correlations demonstrated between Pleistocene shelfal parasequences and 41ky cycles (see Catalano et al., this volume), it seems convincing that the occurrence of Capodarso parasequences reflects the 41-ky periodicity of the orbital obliquity controlling high-frequency sea-level changes related to glacial variations. The 21-ky climatic oscillations related to the precession, controlling organic productivity (Lourens et al., 1992), could be still expressed in shallow-water environments controlling locally-derived sediment availability. From this perspective, parasequences could be adopted as important marker tools for improving correlation potential between shelfal and pelagic successions.

**Tectonic Control on the 4th and 3rd Order Cycles**

The most significant stages of the history of the Capodarso sequence may be kinematically depicted for the complete 4th-order cycle (Fig. 20). Local variations in the stacking pattern relative to the early lowstand systems tract along the flanks of the different synsedimentary folds reveal the maximum tectonic activity. The stratigraphic pattern of the late lowstand to highstand systems tracts has shown a common evolution observed in different outcrops, characterized by a southerly subsidence increase. Table 1 summarizes the correlations between the parasequences in different outcrops of the Caltanissetta area, their relative systems tracts, the biochronology and the isotopic stages formalized for this part of the Pliocene record (Raymo et al., 1989), and the tectonic events occurring synchronously during the deposition of the Capodarso sequence.

The tectonic setting where the Capodarso sequence was deposited, modeled in Figure 21, consists of a major fault-propagation anticline bounded southeastward by minor order synclines and anticlines. The generation and growth of major structures is thought to be related to the propagation towards the foreland of shallow-seated thrust system largely active on the surface during deposition of the lowermost part of the Capodarso sequence (2.5-2.1 Ma); the growth of minor folds is probably related with a transpressional "en echelon" system due oblique component of thrusting. In this general context growth rates for the structures are calculated to be 1.0–1.5 m/ky. Rates of 0.5–2 m/ky are obtained from the Belice basin looking at the structures growth during higher orders of duration (1.2 my; these values are very similar to those ones calculated by Suppe et al. (1992) who reported several examples from different thrust belts demonstrating that average growth rates in synsedimentary fault-related folds are commonly found to be of 1–2 mm/y lasting for 1–8 my. These rates, when considered for the duration of 400 ky, may have produced absolute amount of uplift of the fault-propagation anticlines that are comparable with the total thickness of the Capodarso sequence itself. Similar observations are possible in offshore seismic lines of Sicily (See Catalano et al., this volume). These evidences suggest that in foreland basins comparable with southern Sicily the local tectonics cannot produce regionally-extended erosional surface since: (1) the local uplift of the major anticlines is generally balanced by sediment supply and regional subsidence due to flexural loading; and (2) growth rates of individual structures

<table>
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<tr>
<th>Sequence</th>
<th>Systems tracts</th>
<th>Stacking pattern</th>
<th>Capodarso</th>
<th>Suntello</th>
<th>Serrieri</th>
<th>Sicilabino</th>
<th>Caragiano</th>
<th>K- stage</th>
<th>Tectonic evidences</th>
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<td>—</td>
<td>PCd 10</td>
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<td>aggradational offlap</td>
<td>PCd 9</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>82-81</td>
<td>Subsidence</td>
</tr>
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**Table 1.**
FRANCESCO P. VITALE

1. Shoreline position at 2.5 Ma (base of the Capodarso Sequence)

2. Early lowstand s.t. and beginning of the late lowstand s.t.

3. Deposition of the late lowstand s.t. completed

4. Deposition of the transgressive s.t. and highstand s.t. completed, and shoreline position at 2.1 My (base of the Lannari depositional sequence)

FIG. 20.—Kinematic reconstruction of the depositional and tectonic history of the Capodarso sequence. The tectonic climax occurs in the area before the deposition of the first parasequence; the major structure, near east-west trending along the northern flank of the basin, continues afterward to grow as a fault-propagation anticline, whereas the superficial thrusting is probably still active some tens of kilometers to the south.

FIG. 21.—Depositional and structural simplified model of the Capodarso-Carangiaro-Juculìa fold system, reconstructed from surface data.
may locally exceed eustatic rates only for a distance lower than the fold half-wavelength (hundreds of meters to few kilometers).

CONCLUSIONS

In central-southern Sicily, Plio-Pleistocene marine strata overlie synsedimentary tectonic structures growing in piggy-back and foreland basins located between the emergent northern chain and the southern offshore foreland. Field evidence from this setting points out that tectonics, eustasy, sediment supply and climate collaborated to form peculiar stratigraphic signatures at different orders of duration. The 5th-order shelfal parasequences of Capodarso depositional sequence, occurring in a constant number on different outcrops, were formed independently from the local tectonics. Local changes in subsidence rate, clearly controlling stacking pattern variations at the scale of sequences and systems tracts, did not affect the arrangement of the sedimentary facies within individual parasequence.

The stratal geometry of syntectonic deposits and stacking pattern modifications forced by tectonic stresses defined the kinematic evolution at various stages of growth during time intervals constrained by high-resolution biochronostratigraphy. Where relative uplift rates locally exceeded the eustatic rates, surfaces of local unconformity formed inside successions that are part of the 4th-order depositional sequence. Tectonic unconformities occur locally and can always be distinguished by 3rd- and 4th-order sequence boundaries. As illustrated in the Plio-Pleistocene sea-level cycle chart, the regional sequence boundaries are essentially related to glacio-eustasy; these surfaces and their correlatable conformities have been used to infer the rate of vertical growth of the structures since they bound synchronous sets of strata. The estimated absolute ages of these unconformities constrain the time interval needed to complete tectonic uplift. The absolute amount of growth has been evaluated by amplitude variations of a fold during time intervals corresponding to depositional cycles on different orders. Local discontinuities are recognized at the scale of parasequences or sets of parasequences and bound structures subjected to absolute growth of tens of meters. Considering time intervals on the order of 400 ky, maximum absolute growth of some hundreds of meters have been observed, with relative growth rates of 1–1.5 m/ky. The vertical growth of the structures calculated on the order of duration of 1 my, pointed out a total amount of some hundred meters of vertical growth and a mean growth rate of 0.5–2 m/ky.

The available data indicates that absolute amount of growth is sufficiently balanced by syntectonic deposition and regional subsidence, excluding the possibility of a wide-scale extension of tectonic unconformities. Since the individual growth events are unable to produce regional surfaces of erosion, the sequence stratigraphic analysis may be usefully applied to the study of sectors subjected to active compression; regionally extended unconformities, synchronous and independent from local tectonic processes, can be recognized across wide areas on outcrops, well-log and seismic lines. These surfaces can be used for large-scale correlations and also to constrain the ages of the most significative tectonic events. At the same time, the possible relationships of the parasequences with astronomically-forced Milankovitch cycles could open new working perspec-


PLIO-PLEISTOCENE SEQUENCE STRATIGRAPHY AND TECTONICS OF THE GIBRALTAR ARC

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ABSTRACT: In this paper we attempt to show the relationship between geodynamics and sequence stratigraphy, emphasizing the effects of regional and local tectonics and their interaction with eustasy. These factors contributed to the final architecture of an area located in the frontal region of the Gibraltar Arc. The Plio-Pleistocene succession located in the Atlantic margin offshore of northern Morocco represents a well-developed Neogene progradational succession. Flexural loading in response to thrusting and extensional collapse of the Gibraltar Arc accretionary wedge, followed by widespread late orogenic uplift are the major mechanisms that controlled accommodation space along the margin. The main stratigraphic units have been subdivided into sequences and systems tracts and reveal nine Pleistocene fourth-order sequences interpreted as glacio-eustatic cycles.

The role of minor structures, such as diapirs, ramp anticlines and normal faults that were active during sedimentation affects stratigraphic patterns but does not modify the presence and timing of sequence boundaries and maximum flooding surfaces. The final stratigraphic signature of the Pleistocene in the study area is the result of the complex interaction of regional tectonics, glacio-eustasy, local tectonics and sedimentary processes.

REGIONAL SETTING

The Gibraltar Arc is the westernmost orogenic loop of the Alpine-Mediterranean system. It consists of the Betic Cordillera to the north and by the Rif Cordillera to the south (Fig. 1).

The external domain of the Betic and Rif Cordilleras is represented by non-metamorphic sedimentary successions characterized by thin-skinned tectonics. It represents, respectively, the South-Iberian and North-African Atlantic-type passive margin successions incorporated into the folded belt (García-Hernández et al., 1980; Vera 1981; 1983; Martín-Algarra 1987). The internal or Alborán domain, which includes the internal zones of the Betic and Rif Cordilleras, consists of Alpine-type Triassic carbonates, Early Alpine (Cretaceous-Paleogene) polyphase compressional deformation and HP/LT metamorphism (Galindo-Zaldívar et al., 1989; De Jong, 1991). Neogene extension associated with the collapse of the Alborán Sea and Late Neogene inversion and transpressional tectonics severely modified the overall compressional development of the Gibraltar Arc (García-Dueñas et al., 1992; Flinch, 1993). Equivalent structural units of the Betics and the Rif are presently separated by the Alborán Sea (Fig. 1).

The study area is located offshore the external Western Rif (i.e., on the northern Moroccan continental margin) (Fig. 1). This region is occupied by the westward prolongation of the frontal accretionary prism of the Gibraltar Arc (Flinch and Bally, 1991; Flinch, 1993), a unit classically considered as an olistostrome or “melange” viewed as a gravity driven unit (Lajat et al., 1975; Malod and Didon, 1975; Vidal 1977; Malod and Mougenot, 1979). The Prerifaine Nappe is the accretionary wedge of the Rif Cordillera, which is equivalent to the Guadalquivir Allochthon in the Betic Cordillera. This frontal allochthonous units have been classically subdivided into infra-nappe, nappe and supra-nappe units. The infra-nappe unit is characterized by flat well-bedded reflectors locally affected by imbrication that represent the thin Mesozoic cover of the Palaeozoic Moroccan Meseta and the most proximal part of the Atlantic passive margin (Flinch, 1993) (Fig. 2). The structure of the overlying Prerifaine Nappe and the Guadalquivir Allochthon is characterized by imbricates soling out into a basal decollement and extensional faults that offset the top of the prism. Biostratigraphic and structural data show that extensional collapse on the upper part of the wedge is coeval with thrusting along the basal decollement (Flinch and Bally, 1991; Flinch, 1993; Fig. 2). The prism involves strongly deformed Triassic, Cretaceous, Paleogene and Miocene up to Tortonian sediments (Michard, 1976).

During Plio-Pleistocene deposition, the geodynamic scenario of the Gibraltar Arc is characterized by superposition of several tectonic processes. Uplift in the internal domain of the Gibraltar Arc is coeval with back-arc extension in the Alborán Sea and with the emplacement of the accretionary wedge in the front of
the arc. The emplacement of the prism and the subsequent loading onto the respective Iberian and Moroccan Mesetas results in the development of the Betic and Rif foreland basins (Flinch, 1993). Subsidence along the foreland basins is coeval with uplift of the Hercynian massifs (e.g., Moroccan and Iberian Mesetas; Morel et al., 1983; Cailleux et al., 1986; Morel, 1988) and flexural extension. Normal faulting related to flexural extension (Flinch, 1993) affects the Paleozoic basement and the thin Neogene cover along the outer margin of the foreland basin (Michard, 1976).

Post-Pliocene uplift is well-constrained by geologic, geomorphologic and radiometric data. Differentially uplifted beaches can be traced along the Gibraltar Straits and the Alborán Sea indicating uplift until Tyrrhenian time (Cadet et al., 1977, 1978). Marine sediments with _G. margaritae_ are exposed more than 300 m above sea level in the internal Rif, unconformably overlying basement rocks of the Alborán domain (Cirac and Peypouquet, 1983; Morel, 1988). These marine sediments occupied ancient estuaries not far from the present day Mediterranean shoreline. Nowadays, they are exposed along fluvial valleys (i.e., Oued Laou, Oued Martil) within the renewed relief of the Rif Mountains. The study area was tectonically very active during the Neogene and the emplacement of the accretionary wedge took place only in 9 Ma, from the Tortonian to the Uppermost Miocene (Feinberg 1986, Flinch 1993).

**SUPRA-NAPPE UNITS**

The Supra-Nappe succession consists of Upper Miocene to Pleistocene siliciclastic prograding wedges that cover, with onlap and downlap terminations, the accretionary prism. They present a well-defined clinoformal pattern only locally disrupted by diapirs, ramp anticlines and listric normal faults (i.e., Fig. 2). The progradation direction of these sedimentary wedges is towards the west, nearly perpendicular to southward thrust transport direction of the accretionary prism. Planktonic biostratigraphy from wells and outcrops in the Rharb Basin has been tied into offshore seismic lines to estimate the age of the supra-nappe sedimentary units. Several biozones defined by Wernli (1988) in northern Morocco were used to subdivide the supra-nappe succession of the study area.

According to their geometry and stratigraphic pattern, the supra-nappe succession can be subdivided into a number of units (Fig. 2). The geodynamic evolution of the folded belt controls the mechanisms that provide accommodation space and sediment supply. Two major mechanisms provided tectonic subsidence in the study area: flexural loading and extensional collapse of the accretionary prism. The structural evolution of the study area can be subdivided into a number of stages (Fig. 3). In the following we will relate the lithologic and geometric characteristics of each unit to the geodynamic setting.

**Upper Miocene Units**

During Late Miocene to Early Pliocene time, sedimentation was characterized by anoxic pyrite-bearing marls with occasional sand or siltstone beds. Pelagic fauna is represented by the planktonic biozones of _G. menardi, G. miotumida_ and _G. primitiva_ (Fig. 3A).

Supra-nappe gravitational normal faulting and related growth affects the sediments during this stage, suggesting that subsidence, directly related to extensional collapse of the accretionary wedge, provided accommodation space (Fig. 3A). The beginning of the extensional collapse is indicated in the Rharb Basin by coarse sandstones rich in shallow-water faunas redeposited on the bottom of extensional satellite basins. Anoxic conditions associated to topographic lows on top of the nappe (Cirac and
Peypouquet, 1983). The up-dip convergence of reflectors in the extensional satellite basins shows that sedimentation kept pace with the collapse of the extensional basins.

Lower-Middle Pliocene Units

During Lower-Middle Pliocene time, sedimentation remained pelagic. Pyrite-bearing marls with interbedded sandstones and siltstones were deposited during the biozones *G. margaritae* and *G. punctulata* (Fig. 3B). By this time, the accretionary prism was emplaced in the study area (Feinberg, 1986; Flinch, 1993). The emplacement and loading of the accretionary prism (i.e., Prerifaine Nappe) onto the Paleozoic basement and its thin Mesozoic cover led to subsidence of the underlying basement monocline. Thus, accommodation space, was mostly provided by load-induced subsidence, eventhough supra-nappe normal faulting was also active (Fig. 3B). This stage coincides with the main southward thrusting of the allochthonous unit on top of the Moroccan Meseta.

Upper Pliocene Units

Upper Pliocene shallow-water coquinoid sandstones were deposited along the borders of the Rharb Basin as the sea in-
The Pleistocene-Holocene Units

The Pleistocene shows a very well developed progradational pattern downlapping onto previous strata (Fig. 3D). During Pleistocene time the offlap break shifted 30 km in the seaward direction as a result of an important progradation. Fluvial-aluvial sandstones, mudstones and conglomerates filled the onshore Rharb Basin, and coastal-marine and aeolian sandstones were deposited along the Rharb coastal area and Lalla Zara area (Morel, 1988). The deposition of a thick prograding wedge in the study area is coeval with generalized uplift in the Rif Cordillera demonstrated by several authors (Cadet et al., 1977; 1978; Morel et al., 1983). During Pleistocene time, progradation in the offshore shelf margin coincides with slow deposition in the onshore Rharb Basin. Well-log data suggests the merging of several maximum flooding surfaces and sequence boundaries that indicate bypass and very low depositional sedimentary rates (Fig. 4). At the end of the Pleistocene the characteristic oblique progradation becomes aggradational progradation revealing an increase in accommodation space (uppermost part of the prograding wedge in Fig. 2). Massive Pleistocene progradation reflects reduced accommodation space, as well as abundant available sediments derived from the uplifting hinterland of the Rif Cordillera (Fig. 3D). The rocks that are exhumed in the mountain belt are eroded away and deposited in the neighboring areas like the Rharb Plain. The shift of facies belts (Tilloy, 1955a, b) and the shelf stratal patterns (i.e., Fig. 2) not only depended on the eustatic sea level but on the migration of the hinge zone that separates an onshore coastal-alluvial plain with little deposition from an offshore progradational shelf margin (Fig. 5). Pleistocene-Holocene continuous uplift caused basinward or seaward migration of the equilibrium point and bayline position and therefore the sediment distribution along the coastal-alluvial onshore Rharb Plain.

The Holocene has been studied using high-resolution seismic data (Cirac et al., 1993). Backstepping littorignous sandstones and suspension mudstones and clays interpreted as paleo-shorelines, aeolian dunes and merja-type marsh deposits were recognized on the present-day shelf below sea-level (Cirac et al., 1993). These deposits drowned with the Holocene transgression and are represented by Late Holocene transgressive systems tracts. High-resolution seismic data also revealed that the Sebou river valley shifted towards the south during the last Holocene transgression (Cirac et al., 1993).

These units, which are in fact, transgressive-regressive facies cycles, are governed by specific subsidence mechanisms. They affect stratal patterns, such as progradation versus aggradation, and explain the organization and spatial distribution of sequences and systems tracts. Nevertheless, the timing and number of high order sequences and systems tracts cannot be explained by these factors.

LOCAL TECTONICS

“Local tectonics” includes the role of several structures in the geometry and distribution of sequences and systems tracts (the stratal architecture). Figure 2 shows that thrust emergence or half-graben development affects stratal patterns but do not change the position and timing of sequence boundaries and maximum flooding surfaces. The growth of a structure such as a ramp anticline or a diapir, eventually changes the offlap break.
position and the progradation direction of the supra-nappe sediments (Fig. 6). Once the structure stops growing, sediments prograde according to their old progradation direction (Fig. 6). The seismic data used in this study (see NW side of Fig. 7) also shows how angular unconformities on growing anticlines coincide with well defined sequence boundaries recognized in the shelf break area. In deep-water settings, visible marine onlaps of lowstand wedges onto highstand prograding units define the unconformity (Fig. 6). These angular unconformities represent tectonically enhanced sequence boundaries (Vail et al. 1991).

CHARACTERISTICS OF SEQUENCES AND SYSTEMS TRACTS

Based on the available seismic data, the Pleistocene sedimentary succession in the study area has been separated into sequences and systems tracts. The often chaotic and poorly organized seismic expression of the supra-nappe succession sug-
forms and chaotic recomplexes. Prograding complexes are characterized by clino-stacke:

mfl

slope fans (SF) overlain by dowlapong highstand deposits (Fig. 7). Lowstand deposits consist of

mass

transgressive backstepping units and thin or deeply eroded

in the study area. Substantial erosion, associated with canyons

characteristic of late lowstand deposition and are very widespread

mass

fl

ectors (Fig. 11).

Transgressive systems tracts (TST) are only identifiable in the shelf margin and pass basinward into a strong reflector that constitutes the maximum flooding surface (mfs, Fig. 11). Highstand systems tracts (HST) are constituted by reflectors with good lateral continuity and clinoformal pattern.

Fourth-order glacio-eustatic cycles.

The sequence stratigraphic analysis of the seismic data reveals nine Pleistocene sequences (Fig. 9). Due to the lack of offshore well data in the study area, the base of the Pleistocene was correlated with onshore wells located in the Rharb Basin (for location see Fig. 3, Flinch, 1993). The timing of the progradational units is provided by the base of the Pleistocene defined with planktonic biostratigraphy as the contact between G. truncatulinoides and G. inflata biozones (Feinberg, 1986; Wernli, 1988) and the present-day Late Holocene transgressive systems tracts recognized with high-resolution seismic data (Cirac et al., 1993). These two references constitute the time boundaries of the progradational succession of our study. The sequence stratigraphic analysis of the seismic data reveals nine Pleistocene sequences (Fig. 7). The estimated time span, defined by these time boundaries is of 0.1–0.2 my, that corresponds to fourth-order cyclicity (Vail et al. 1991). The number of sequences recognized in the study area are the same as those recognized in the Gulf of Mexico (Wornardt et al., 1998), which suggest a global control. Sequences are probably related to glacio-eustatic cycles, which are characterized by smaller magnitude but higher frequency than tectonically induced transgressive-regressive facies cycles (Vail et al., 1991). Isotopic, geomorphologic and geologic data suggest high-frequency sea-level fluctuations during the Pleistocene (Shackleton and Opdyke, 1973; Bartek et al., 1991, Vail et al., 1991). Changes over the last 0.8–0.9 my have a periodicity of 0.1 my and an average amplitude of over 1.5 per mil (a sea level equivalent of >130 m; Williams, 1988). The order of magnitude of these cycles is similar to several stages based on field studies of outcrops located along the Atlantic Coast of Morocco (Fig. 12).

Onshore expression of Pleistocene sequences

The Atlantic Coast of Morocco is an exceptional area to study sea-level oscillations, since Pliocene to Quaternary continuous uplift has exposed several strand lines, dune ridges and other geomorphologic elements and sediments that can record sea-level fluctuations during the last 5 my (Stearns, 1978). Seaward and landward shift of facies belts (marine-eolian-marsh-

suggests a prograding mud-dominated shelf margin (Flinch, 1993). This contrasts with conventional deltaic progradations (Berr

hill, 1986). The lack of sand-size material in the source area prevented the deposition of coarse clastic basin floor fans in the sense of Vail et al. (1991). A complete depositional sequence is represented by: thick lowstand prograding complexes (LPC) dominated by slumps and mass flow deposits, well-developed transgressive backstepping units and thin or deeply eroded highstand deposits (Fig. 7). Lowstand deposits consist of stacked slope fans (SF) overlain by dowlapong prograding complexes. Prograding complexes are characterized by clinoforms and chaotic reflectors at the toe that represent slump or mass flow deposits (Fig. 8). Mass transport deposits are character-

of late lowstand deposition and are very widespread in the study area. Substantial erosion, associated with canyons

and incised valleys, can attain more than 150 m of relief. Strike sections show erosional surfaces related to sequence boundaries (sb) filled with chaotic facies characterized by complex reflection termination patterns (Fig. 9, 10). Incised canyons are common in the slope area, they crosscut thin transgressive units and occasional highstand deposits, resulting in amalgamation of lowstand units (Fig. 9). Lowstand prograding complexes (LPC) are identified by marine onlap onto highstand wedges (Fig. 11). Often lowstand prograding complexes are characterized by a transparent seismic character that contrasts with the well-imaged slope fans (SF) composed of channel-levee complexes (Figs. 9, 11). Backstepping transgressive systems tracts (TST) are well represented by nearly horizontal reflectors (Fig. 11). Transgressive systems tracts (TST) are only identifiable in the shelf margin and pass basinward into a strong reflector that constitutes the maximum flooding surface (mfs, Fig. 11). Highstand systems tracts (HST) are constituted by reflectors with good lateral continuity and clinoformal pattern.


Figs. 10.—Dip section of a slope fan and a lowstand prograding complex onlapping onto a previous highstand prograding wedge. Legend: sb: sequence boundary, mfs: maximum flooding surface, LPC: lowstand prograding complex, SF: slope fan, TST: transgressive systems tracts, HST: highstand systems tracts.
**Fig. 11.**—Strike view of a slope fan that shows well developed channel-levee complexes. Legend: sb: sequence boundary, mfs: maximum flooding surface, LPC: lowstand prograding complex, SF: slope fan, TST: transgressive systems tracts, HST: highstand systems tracts.

**Fig. 12.**—Ages, lithology and faunal content of the marine and continental Plio-Pleistocene local stages of northern Morocco.
fluvial) or uplifted marine terraces (ancient strand lines) constitute reliable sea-level markers. These deposits are particularly well exposed in the coastal and marginal zones of the Rharb Plain (Biberson, 1970; Stearns, 1978; Weisrock and Fontugne, 1991). The Rharb Plain is a coastal-alluvial plain close to sea level located in the front of the Rif orogenic belt. The shoreline area separates differentially uplifted onshore coastal-fluvial-alluvial deposits from offshore shelf-margin siliciclastic wedges and constitutes a hinge zone. The offshore sequences recognized on seismic lines have an expression onshore in the coastal-alluvial plain (Fig. 5). Figure 12 shows a compilation of marine and continental stages for the Pleistocene of Morocco based on several authors (Biberson, 1965; Stearns, 1978; Texier et al., 1985; Raynal et al., 1986; Weisrock and Fontugne 1991; Wengler and Vernet, 1992). Figure 13 shows the estimated correlation of the onshore marine and continental stages with the fourth-order offshore seismic sequences defined in this work.

Pleistocene-Holocene sea-level fluctuations are represented in the coasts-alluvial Rharb Plain by shifting of facies belts. Shallow brackish-water lakes, referred to locally as “Merjas”, developed on the landward side of the aeolian dune ridges due to an older sand dune belt related to former higher sea level (Fig. 5). Transgressions are evidenced by landward shift of the near shore dunes belt (i.e. “grès de Rabat”, “grès de Mamora” and “sables beiges”) and shallow-water marsh-type deposits (“grès des merjas”) and pisolithic sandstone transgressing onto
alluvial plain red siltstones, conglomerates and sandstones (Tilloy, 1995a, b).

The study area provides an interesting example of interaction between tectonics and eustasy. Most of the stratigraphic models developed to analyze the role of tectonics and sea-level changes on the final architecture of the sedimentary units have been defined in foreland basins or passive-margin settings (Posamentier and Allen, 1993). The Pliocene section of the study area is basically affected by thrusting and extensional collapse on the upper plate of an accretionary prism. In this structural setting, a well developed shelf break occurs which contrasts with the ramp-type setting characteristic of foreland basins (Van Wagoner et al., 1988; Posamentier and Allen, 1993). The Pleistocene progradational succession of the study area is strongly controlled by the late orogenic uplift of the Rif Cordillera. The cause of this uplift is not well understood. In the study area, subsidence increases basinward or seaward like in the cratonic side of a foreland basin or in a passive-margin setting (Posamentier and Allen, 1993).

The Plio-Pleistocene sequences and systems tracts of the study area represent fourth-order eustatic cycles like the ones defined in the Gulf of Mexico (Wormard and Vail 1990). These sequences constitute the high frequency element of the eustatic curve and are correlative with other regions of the world. The major offlap break shifts and stratal patterns within these sequences are directly related to regional tectonic events, for example extensional collapse, thrust-sheet loading and generalized uplift. Local tectonic effects like thrust emergence or half-graben growth control stratal geometry but do not affect the timing of fourth order sequences.

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OLIGOCENE—MIDDLE MIOCENE DEPOSITIONAL SEQUENCES OF THE CENTRAL PARATETHYS AND THEIR CORRELATION WITH REGIONAL STAGES

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ABSTRACT: Detailed sequence stratigraphic analysis allowed the interpretation of seventeen depositional sequences in the Oligocene through middle Miocene succession of the Pannonian Basin (Hungary), the largest basin of the central Paratethyan area (central/eastern Europe). Depositional sequences were identified based on the analysis of published geological descriptions of outcrops and study of 3,000 km of 2D reflection seismic profiles and 45 hydrocarbon exploration wells. Eight depositional sequence boundaries coincided with regional stage boundaries; and additional nine sequence boundaries were identified within the regional stages. The sequences were stratigraphically positioned on the basis of calcareous nannofossil data from 26 wells. Within the constraints of the biostratigraphic resolution in this interval, the stratigraphic position of the sequences correlates well with the previous records of depositional sequences (Haq et al., 1988). Three sequences, one in the Rupelian Stage and two in the Burdigalian Stage, were not identified by Haq et al. (1988). The examined regional stages correlate within the Paratethyan region, from Switzerland to the Caspian Sea and show a direct correlation with the standard stages. Stage boundaries typically correlate with episodic closures of connections between the European epicontinental seas from Oligocene through middle Miocene time. These closures are interpreted to result from short-term glacio-eustatic falls that overprint longer-term local tectonics.

Depositional sequences are believed to result from glacioeustasy superimposed on a tectonic signal. The results, obtained in this study, and compared with oxygen isotope records (Abreu and Haddad, this volume), show a close agreement between the number and the stratigraphic position of oxygen isotope events and sequences. This supports that the major driving mechanism of depositional sequence boundary formation is glacioeustasy rather than a local or regional tectonic mechanism, and the identified sequences in this study may thus be global in nature.

INTRODUCTION

Accurate stratigraphic correlation is essential for a successful sequence calibration. Scientific effort of the Regional Committee on Mediterranean Neogene Stratigraphy (RCMNS) through the last twenty years has improved dramatically our understanding of the Cenozoic evolution of the European epicontinental seas (Steininger and Nevesskaya, 1975; Steininger et al., 1985; Fig. 1) and has resulted in the development of the Mediterranean regional stage concepts (Fig. 2). The repeated isolation of this area from the global sea (Laskarev, 1924; Senes and Marinescu, 1974; Steininger et al., 1976; Rögl and Steininger, 1983; Báldi, 1983, 1986) motivated the RCMNS to develop a regional stage concept. By the late of 1980’s, the Paratethyan and Mediterranean regional stages were well defined, although some minor boundary problems remained (Steininger et al., 1990).

Benthic foraminifera and/or mollusk fauna indicate that the regional stages within the central Paratethyan realm (central/eastern Europe) originally were identified in shelf environments (Steininger and Nevesskaya, 1975). These regional stages were defined as transgressive-regressive facies cycles (Steininger and Nevesskaya, 1975; Steininger et al., 1985). At the basin margin, unconformities occur at the base of the regional stages. These unconformities originally were interpreted as the result of orogenic cycles (Hámor and Szentgyörgyi, 1981; Committee of Hungarian Stratigraphy, 1983; Bérczi et al., 1988). Within the basins, however, these unconformities become conformable (Bérczi et al., 1988). Results of the DSDP/ODP program allowed more accurate dating of the regional stages by planktonic, paleomagnetic and radiometric data and the correlation between the regional stages and standard stages improved significantly (Steininger et al., 1990, 1996).

The realization that standard stages often represent unconformity bounded depositional units has led to an initiative to define a “Global Stratotype Section and Point” (GSSP) for each stage boundary (e.g., Premoli Silva et al., 1988; Steininger, 1994). This effort, still in progress, will attempt to define a point in a section with continuous deposition across the stage boundary as the “Global Stratotype Section and Point” and will express the stratigraphic position of that point relative to as many stratigraphic disciplines as feasible. Since a qualifying section probably will be defined in deeper water deposits, the position of the GSSP relative to identifiable sequence boundaries will remain elusive.

The principal goal of the present study is to identify Oligocene through middle Miocene depositional sequence boundaries within the Paratethyan area. The Paratethyan regional stages are bounded by well-defined unconformities that can be traced from western to eastern Europe. The stratigraphic posi-
tions of the regional stage boundaries are determined with ade-
quate biostratigraphic control and thus offer an opportunity to
calibrate the chronostratigraphic position of Oligocene through
middle Miocene sequence boundaries with the standard chrono-
stratigraphic system. Our findings support the correlation be-
tween the regional and global stages proposed by Steininger et
al. (1990, 1996).

A secondary objective is to identify major driving mecha-
nisms creating the sequences. Do third-order sequences reflect
changes of glacioeustasy as noted by Vail et al. (1977), Vail
(1987), Jervey (1988) and Posamentier and Vail (1988)? There
is evidence of continental glaciation at least since the late Eo-
cene (Miller et al., 1991). Oxygen isotope analyses can provide
an independent tool to determine glacio-eustatic sea-level falls
and can be used as an indirect indicator to identify sequence
boundaries for most of the Cenozoic (Miller et al., 1987, 1991;
Abreu and Savini, 1994; Abreu and Haddad, this volume). Al-
ternatively, third-order sea-level fluctuations could reflect
changes in the pronounced intra-plate stresses of Neogene time
(Cloetingh et al., 1985).

Based on available published biostratigraphic and lithostrat-
igraphic data and reflection seismic and well-log data from Pan-
nonian Basin (Hungary; Fig. 2.), we propose an integrated
framework for Oligocene through middle Miocene regional
stages, global stages and sequence boundaries.

Data Set and Methodology
The study area includes the central Pannonian Basin, in
northeastern Hungary. A number of data sets were used in this
sequence stratigraphic work. All available reference sections
for regional and global stages were compared to and completed
with the available field-observation data. Detailed correlation
of 3,000 km of 2D reflection seismic profiles and 45 hydrocar-
on exploration wells from the Pannonian Basin allowed identi-
fication of depositional sequences and offered an opportunity
to compare these results with published data. Nannoplankton
data, available from 26 wells, provided age control of the iden-
tified sequences. Oxygen isotopic data from the DSDP/ODP
sites generated and compiled by Miller et al. (1987, 1991) were
compared with data from PETROBRAS (Abreu and Haddad,
this volume).

The sequence stratigraphic model relied on the methodology
and terminology of Vail et al. (1977), Vail (1987), Van Wagoner
Well-log sequence stratigraphy and detailed systems tracts inter-
pretations are based on techniques described by Vail and
Wornardt (1990), Van Wagoner et al. (1990) and Mitchum and

Geologic Setting
The Pannonian Basin is an integral part of the Alpine moun-
tain belts of eastern central Europe. It overlies the Mesozoic

thrust sheets of the eastern Alps, Carpathians, and Dinarides (Fig. 3). The middle Miocene to present section of the Pannonian Basin is superimposed on a middle Eocene to early Miocene retroarc flexural basin complex, as a result of back-arc extension (Tari et al., 1993; Fig. 3). The Paleogene basin evolution primarily was driven by compressional tectonics; the Neogene evolution was driven by extensional tectonics (e.g., Tari et al., 1992a). Tectonic regimes produced transgressive-regressive facies cycles, which reflect the individual tectonic evolution of the Pannonian Basin (Fig. 4).

In the early phase of the Hungarian Paleogene Basin (HPB), a flexural basin developed along the inner side of the western Carpathian arc. Flexure was controlled by an antigoritic thrust system that developed in response to the subducting European plate (Tari et al., 1993). The thrust load resulted in a deep, "flysch" basin during late Eocene and early Oligocene time. The second phase of the HPB (late Oligocene to early Miocene) was characteristic of a "molasse stage" (Tari et al., 1993). Due to the gradual cessation of thrusting in the adjacent folded belt, the basin shallowed. During the early Miocene, this process culminated in isostatic uplift, characterized by fluvial and/or shallow marine sedimentation (Szántó and Tari, 1993).

During the late early Miocene to Pliocene deposition, the Pannonian Basin was one of several Mediterranean back-arc basins (Bally and Snelson, 1980; Horváth and Berckhemer, 1982; Royden, 1988) resulting from the collision of the African promontory, Apulia, with the European plate. The basin occurs on the concave side of an A-subduction arc ("Pannonian-type basins", sensu Bally and Snelson, 1980). Tari et al. (1992b) suggest that gravitational collapse of the intra-Carpathians domain combined with subduction zone roll-back drove the Neogene back-arc extension. In contrast to the western Mediterranean basins, back-arc extension in the Pannonian Basin did not lead to the opening of an oceanic basin (Fig. 1; Bally and Snelson, 1980).

The Paratethys Concept

Uplift of the Alpine-Caucasian mountain range essentially separated the central and eastern European Cenozoic epicontinental seas (Fig. 1) from the global ocean after the earliest part of lower Oligocene time (Báldi, 1983, 1984, 1986; Fig. 5). Evidence for separation of the Paratethyan basins comes from the increasing endemism of aquatic biota (Steininger and Rögl, 1979). The different fauna and tectonic evolution motivated Laskarev (1924) to distinguish a northern Paratethyan bioprovince and a southern Mediterranean Tethyan bioprovince.

The Paratethys includes the major late Cenozoic basins from the Alps to the Caspian Sea (e.g., Rögl and Steininger, 1983; Steininger et al., 1988). As a result of changing seaway connections through the Paratethys, different ecosystems evolved. Based on these differences, Senes (1969) subdivided the Paratethys into three sub-bioregions: the western Paratethys (Swiss Molasse Basin), the central Paratethys (Pannonian Basin, Alpine-Carpathian foredeep and the intramontaine basins in central Europe) and the eastern Paratethys (Ponto-Caspian area from Black Sea to the Aral Lake; Fig. 1). Periods of endemism occurred when the connections between the Paratethyan seas closed. Endemism complicates the biostratigraphic correlation between the Paratethyan sedimentary successions and the Tethyan standard stages.

Regional Stages

Oligocene through middle Miocene deposits in the central and eastern Paratethyan basins consist of a number of shallow-to marginal-marine depositional units separated by unconformities. These depositional units resulted in a number of regional stages characterized by unique faunal assemblages (Committee of Mediterranean Neogene Stratigraphy, 1971; Steininger and Neveskaya, 1975). During the last two decades, more than twenty different stage schemes have been proposed, with different nomenclature, ages, and durations for each stage (e.g., Rögl et al., 1978, 1993; Rögl and Steininger, 1983; Steininger et al., 1985, 1988, 1990).

We adopted the stage scheme proposed by Báldi (1983, 1986), Steininger et al. (1990, 1996), and Vakarcs et al. (1994) for the European Neogene stages (Fig. 2).

The Kiscellian Stage was defined by Báldi (1968, 1983) based on faunal differences between late Eocene and early Oligocene deposits. The base of the Kiscellian Stage correlates with the base of Rupelian (Báldi, 1983, 1986; Nagymurosy and Báldi-Beke, 1988).

The Eggenburgian Stage (Steininger and Senes, 1968) contains the FADs of “Loibersdorf-type” mollusks *Chlamys gigas* and *Pitar lilacionides* and also the benthic foraminifera, *Elphidium ortenburgense* and *Uvigerina posthantkeni*. The Egerian/Eggenburgian boundary and the Aquitanian/Burdigalian boundary are coeval (Rögl et al., 1979; Rögl and Steininger, 1983; Steininger et al., 1990).

The Ottnangian Stage (Cicha and Senes, 1968; Papp et al., 1971) contains the FADs of euhaline Ottnangian type mollusk fauna, including *Chlamis albina*, *Pecten (F.) hermannensis* and *Pecten fotensis* (Steininger and Nevesskaya, 1975).

The Karpatian Stage (Cicha and Tejkal, 1959) contains the FADs of benthic foraminifera *Cyclammina karpatica*, *Cibicides slovenicus*, *Heterostegina granulatatata praeformis* and *Uvigerina parkeri*, ostracodes *Aurila angulata*, and the mollusks *Chlamys scabrella*, *Chlamys macrotis* and *Aurila div. spec.* (Steininger and Neveskaya, 1975). The Karpatian/Badenian boundary and the Burdigalian/Langhian boundary are coeval (Rögl et al., 1978; Steininger et al., 1990).

The Badenian Stage (Papp, 1968; Cicha and Senes, 1968) contains the FADs of benthic foraminifera *Uvigerina macrocarinata* and *Heterostegina praecos*, ostracodes *Aurila div. spec.* and vertebrates (Steininger and Neveskaya, 1975). Benthic foraminifera, the most commonly used for microbiostratigraphy zonation, is that of Grill (1943), which subdivides the Badenian Stage into lower, middle, and upper substages. Rögl et al. (1978) considered the lower/middle Badenian boundary coeval with the Langhian/Serravallian boundary. The middle Badenian and the upper Badenian substages yield inconsistent ages; this study uses upper Badenian.

The Sarmatian Stage (Suess, 1866) contains the FADs of endemic faunas with mollusks (*Ervilia podolica*, *Cerastoderma vindobonense*, *Mactra vitaliana eichwaldi*, *Pirenella picta*), ostracods, foraminifera and vertebrates (Steininger and Neveskaya, 1975).

The Pannonian Stage (Telegdy-Roth, 1879) contains the FADs of endemic faunas with mollusk faunas (*Melanopsis*, *Lymnocardiidae*), ostracods, foraminifera and vertebrates (Steininger and Neveskaya, 1975).

Senes and Marinescu (1974) subdivided the Paratethyan deposits from the first endemism in the lower Kiscellian Stage to the end of the Pannonian Stage. They proposed the Eo-, Meso- and Neoparatethyan intervals (Fig. 4). The Eoparatethys includes most of the Kiscellian Stage and all of the Egerian, Eg-
FIG. 4.—Chrono- and lithostratigraphy of the Hungarian Paleogene Basin and the Neogene Pannonian Basin from Oligocene to Middle Miocene. Correlation and chronostratigraphic position are based on the Berggren et al. (1995) time scale and isotopic evidence described by Abreu and Haddad (this volume).
Paratethyan Depositional Units

At the end of Eocene time, faunal assemblages were uniform throughout Europe. In early Oligocene time (near the NP22/NP23 boundary), the marine connection with the Mediterranean (Fig. 1) closed (Figs. 4, 5). Closure resulted in the first endemism within the Paratethys (Báldi, 1983). Later in early Oligocene time, a marine connection developed again between the Mediterranean and the central Paratethys through the Slovenian corridor (Báldi, 1983; Figs. 1, 5). Prior to the Eggenburgian Stage in the early Miocene, the marine connection between the central Paratethys and both the Atlantic region and the Mediterranean closed (Figs. 1, 5). The subsequent Eggenburgian transgression opened a new seaway, connecting the central and western Paratethys with the western Mediterranean across the Rhone and Swiss Molasse basins (Rögl et al., 1978; Figs. 1, 5). At the end of the Eggenburgian Stage, all marine connections closed again and large areas within the central Paratethys became subaerially exposed (Nagymarosy and Müller, 1988; Tari et al., 1993). During the Ottnangian, transgression, seaways reopened to the western and eastern Paratethys and to the Mediterranean, through the Slovenian corridor (Rögl and Steininger, 1983).

During Mesoparatethys, prior to the Ottnangian/Karpatian boundary, connection with the western Paratethys ceased and the eastern Paratethyan seaway temporarily closed (Rögl et al., 1978). However, the Karpatian transgression resulted in a well-defined Mediterranean connection again through the Slovenian corridor (Rögl and Steininger, 1983).

The Mediterranean connection remained open at the beginning of the Badenian (Steininger et al., 1988). During the early Badenian transgression, the eastern Paratethyan and the Indo-
Pacific connection opened once again (Steininger et al., 1985) but temporarily closed at the end of the early Badenian. This connection reopened as a result of the late Badenian transgression (Rögl and Steininger, 1983).

In the Neoparateethys, at the Badenian/Sarmatian boundary, the Mediterranean marine connections with the central Paratethys disappeared, however, an Indo-Pacific seaway remained open between the eastern part of the central Paratethys and the eastern Paratethys (Steininger et al., 1988). At the beginning of the Pannonian Stage, all marine connections with the central Paratethys disappeared (Steininger et al., 1988). A river connection between the Pannonian Basin and the Black Sea existed from Pannonian to Messinian stages through the Carpathian foredeep basin. From Messinian to the Pliocene time, a river connection existed with the Mediterranean (Hsu et al., 1978; Tari, 1994). The present Danube river has flowed into the Black Sea since the Pleistocene (Hsu et al., 1978).

DEFINITION OF SEQUENCE BOUNDARIES IN THE PANNONIAN BASIN

The first-order chronostratigraphic calibration of the sequence boundaries described here is based on biostratigraphic data (planktonic foraminifera and calcareous nannofossils). The sequence boundaries were correlated with the Haq et al. (1988) onlap curve and the isotopic record from the early Oligocene to middle Miocene section (Abreu and Haddad, this volume). The age conversion between the different time scales (Berggren et al., 1985, 1995) is based on magnetostratigraphy. The calibration with Haq et al. (1988) chart is based on the position relative to stages. To avoid miscorrelation due to the utilization of different time scales, all events described in this work are calibrated to the Berggren et al. (1995) time scale (Fig. 6).

Figure 7 summarizes the positions of the sequences discussed below. The stratigraphic positions of the sequences are expressed relative to magneto- and biostratigraphic criteria. The sequences are identified by their positions within standard stages. For example, the second Rupelian sequence boundary is called Ru-2. When sequence boundaries approximate regional stage boundaries, the sequence boundaries are given prefixes for both stages (for example: Pr-4/Ru-1). At present, the boundary stratotype for the Eocene/Oligocene boundary at Massignano (Italy) does not reveal whether the sequence boundary is in the Eocene/Priabonian or Oligocene/Rupelian.

Figures 8 and 9 illustrate the sequence interpretation for Wells 1, 3, 4 and 2. Wells 1, 3 and 4 are located on seismic lines A and B (Figs. 10, 11 and 12) located in the northern part of the Hungarian Pannonian Basin (Fig. 1).

Five major unconformities can be identified on seismic lines A and B (Figs. 10, 12): top of basement, Eocene/Oligocene boundary, lower/upper Oligocene boundary, Ottnangian/Karpatian boundary in the upper part of the lower Miocene and Sarmatian/Pannonian boundary.

Available biostratigraphic control and the seismic reflectors constrain the chronostratigraphic chart (Fig. 11). The chart illustrates the extent of hiatuses in the rock record as a result of non-deposition and erosion. This chart also shows that some unconformities represent an amalgamation of more than one erosional event.

**Pr-4/Ru-1**

**Field Reference.**—

Báldi (1983, 1986) demonstrated that the first horizon of the upper bathyal lower Kiscellian Tard Clay unconformably overlies the neritic upper Priabonian Szépvölgy Limestone. In basin environments, the lower neritic-upper bathyal Tard Clay (30- to 500-m water depth) conformably overlies the bathyal Buda Marl (200- to 1,000-m water depth; Báldi, 1983).

**Seismic Signature.**—

On seismic lines A and B, reflections representing the lower Oligocene Tard Clay onlap older reflections of the upper Eocene Buda Marl (Fig. 10 at 6, 11, 12 and 13 km; Fig. 12 at 7–9 km) or reflections of the Mesozoic basement (Fig. 10 at 15 km).

**Interpretation.**—

The unconformity between the late Priabonian Szépvölgy Limestone and lower Kiscellian Tard Clay is considered a sequence boundary (Pr-4/Ru-1). Báldi (1983, 1986) suggested that the dramatic decrease in carbonate content from 35–40% in the Buda Marl to 10% in the Tard Clay, coupled with the
Fig. 7.—Chronostratigraphic and biostratigraphic position and ages of observed sequence boundaries. The sequences are identified by their position within the global stages (e.g., the second Rupelian sequence boundary is called Ru-2).

Age.—

The Eocene/Oligocene boundary stratotype at Massignano in Italy is calibrated close to the top of Chron C13r and is radiometrically dated at 33.7 Ma (Premoli Silva et al., 1988; Berggren et al., 1995; Figs. 4, 7). The series boundary correlates just above the LAD (last appearance datum) of Hantkenina spp. (low in P18) and to the middle part of NP21 nannoplankton chronozone (Premoli Silva et al., 1988; Berggren et al., 1995; Figs. 4, 7).

Correlation.—

This sequence boundary correlates with sequence boundary TA4.4 (36.0 Ma) of Haq et al. (1988) at the Eocene/Oligocene boundary (33.8 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as ORi-1 at 33.7 Ma, (Abreu and Haddad, this volume; Fig. 4).

Seismic Signature.—

On seismic line A, reflections representing the lower Kiscelian Tard Clay onlap older reflections (Fig. 10 at 3–6 and 10 km). Within the Tard Clay, reflections truncate below younger reflections on Figure 10 at 3 km and on Figure 12 at 9–10 km.

Interpretation.—

The lower/middle member boundary of the Tard Clay is considered a sequence boundary (Ru-2). In the lower part of the NP23 nannoplankton chronozone (Báldi-Beke, 1977), the first appearance of the endemic Ergenica cimlanica mollusk fauna (Báldi, 1983, 1986) reflects the initial period of endemism of the central Paratethys. The abrupt appearance of reworked Eocene fossils and older limestone intercalations within the basin reflects extensive erosion in the neritic environment (infra-Oligocene denudation, Telegdy-Roth, 1927). Based on seismic lines A and B (Figs. 10, 12), we correlate the sequence boundary Ru-2 with the sequence boundary identified in wells 1, 3 and 4 within the Tard Clay (Fig. 8). In wells 1 and 4, this sequence boundary occurs in the lower part of NP21 nannoplankton chronozone. In well 3, the sequence boundary is a composite boundary that consists of the Ru-2 and Ru-3 sequence boundaries.

Age.—

The Ru-2 sequence boundary corresponds to the P18/P19 planktonic foraminifera chronozone (Báldi, 1986) and to the decreasing depositional depth from middle bathyal in the Buda Marl to upper bathyal in the Tard Clay, is consistent with a relative sea-level fall. Báldi (1983, 1986) postulated an abrupt climatic cooling at this boundary. The sequence boundary identified on seismic lines A and B (Figs. 10, 12) correlates with the sequence boundary interpreted in wells 1, 3 and 4 at the boundary of Buda Marl and Tard Clay (Fig. 8). In well 1 the sequence boundary is between the NP21 and NP22 nannoplankton zones. In well 3, the sequence boundary is a composite boundary that consists of the Pr-4/Ru-1 and Ru-2 sequence boundaries. Lakatos et al. (1992) also described a sequence boundary dated by the top of NP21 nannoplankton zone.

Field Reference.—

The neritic Hárhegy Sandstone unconformably overlies the lower member of the Tard Clay (Fig. 4; Báldi, 1983). Varga (1982) recognized graded, several mm- to 10-cm-thick late Eocene limestone intercalations in the middle member of the Tard Clay in a basin setting. In the same stratigraphic position of the Tard Clay, Báldi (1983) described turbiditic sandstones. From industry wells, Csiky (1961) and Lakatos et al. (1992) described sandstone beds at the base of the middle member of the Tard Clay ("Lattorian Sandstone").
OLIGOCENE—MIDDLE MIocene STAGES AND DEPOSITIONAL SEQUENCES

FIG. 8.—Well-log cross section from the Hungarian Paleogene Basin. Lakatos et al. (1992), published the well-log and biostratigraphic data. Location of profile is shown in Figure 3. Nannoplankton chronozone data from cores provides age control for wells. See seismic interpretation on Figures 10 and 12. All sequence boundaries described in the text are presented. Due to differential erosion, some sequence boundaries are composites of two or more sequence boundaries in the same position.

lower part of the NP23 (Báldi-Beke, 1977) nannoplankton chronozone (Figs. 4, 7). It is calibrated to the middle part of Chron C12r and is dated at 32.0 Ma (Berggren et al., 1995; Figs. 4, 7).

Correlation.—

The sequence boundary seems to correlate with an isotopic event referred to as ORi-2 at 31.9 Ma (Abreu and Haddad, this volume; Fig. 4). There is no sequence boundary at this position on the Haq et al. (1988) chart.

Ru-3

Field Reference.—

The lower part of the middle to upper bathyal Kiscell Clay (200- to 1,000-m water depth) is a non-stratified and non-lam-
boundary occurs on seismic lines A and B (Figs. 10, 12) and correlates with the sequence boundary interpreted in wells 1, 3 and 4 within the Kiscell Clay (Fig. 8). In wells 1 and 4, this sequence boundary is in the lower part of NP24 nannoplankton chronozone.

Age.—

The Ru-3 sequence boundary corresponds to the P20/P21 planktonic foraminifera chronozone boundary and to the lower part of the NP24 nannoplankton chronozone (Báldi, 1986). It is calibrated to the top of Chron C11n.1n and is dated at 29.4 Ma (Berggren et al., 1995; Figs. 4, 7).

Correlation.—

This sequence boundary seems to correlate with the sequence boundary TA4.5 (33.0 Ma) of Haq et al. (1988) in the early Oligocene (31.2 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as ORi-3 at 29.4 Ma (Abreu and Haddad, this volume; Fig. 4).

Ru-4/Ch-1

Field Reference.—

In deeper water environments, the middle to upper neritic Kiscellian Kiscell Clay is conformably overlain by the upper bathyal Egerian Szécsény Siltstone (200- to 400-m water depth; Báldi, 1983, 1986). At the base of the Egerian Eger Formation, the Szőlőske Gravel Member unconformably overlies the middle-upper bathyal Kiscellian Kiscell Clay. Tari and Sztanó (1992) and Tari et al. (1992b) interpreted it to be a canyon fill related to a sea-level fall.

Seismic Signature.—

On seismic line A reflections representing the Kiscellian Kiscell Clay truncate below reflections of the Egerian Szécsény Siltstone (Fig. 10 at 2, 10, 11 and 17 km). Reflections representing the Szécsény Siltstone onlap the oldest reflections of Kiscell Clay (Fig. 10 at 5–7 and 22–23 km).

Interpretation.—

The Kiscellian/Egerian boundary coincides with a basinwide regressive event, which is correlated with the major middle Oligocene eustatic sea-level fall (Báldi, 1973, 1983, 1986; Tari et al., 1992b, 1993). This stage boundary is a well defined sequence boundary, with a marked basinward shift of facies (Báldi, 1983, 1986; Tari et al., 1992b, 1993). The sequence boundary identified on seismic lines A and B (Figs. 10, 12) correlates with the sequence boundary interpreted in wells 1, 2, 3 and 4 at the boundary of Kiscell Clay and Szécsény Siltstone or Eger Formation (Figs. 8, 9). In these data, the sequence boundary lies in the middle part of NP24 nannoplankton chronozone. The sequence boundary is a composite boundary which consists of the Ru-4/Ch-1 and Ch-2 sequence boundaries in wells 3 and 4.

Age.—

The Kiscellian/Egerian boundary correlates with the P21A/ P21B planktonic foraminifera chronozone boundary and to the middle part of the NP24 nannoplankton chronozone (Báldi, 1986; Figs. 4, 7). It is calibrated to the top of Chron C10n.1r and is dated at 28.5 Ma (Berggren et al., 1995; Fig. 4).

Correlation.—

This sequence boundary seems to correlate with the sequence boundary TB1.1 (30.0 Ma) of Haq et al. (1988) at the early Oligocene/late Oligocene boundary (28.4 Ma; Hardenbol et al., this volume) and with an isotopic event referred to as OCi-1 at 28.4 Ma (Abreu and Haddad, this volume; Fig. 4).

Ch-2

Seismic Signature.—

On seismic line A, within the lower Egerian Eger Sandstone younger reflections onlap older reflections (Fig. 10 at 4–6 km).

Interpretation.—

The sequence boundary identified on seismic line A (Fig. 10) correlates with the sequence boundary interpreted within the Eger Formation in wells 2, 3, and 4, and within the Szécsény Formation in well-1 (Figs. 8, 9). In wells 2, 3, and 4, this se-
quence boundary occurs at the base of NP25 nannoplankton chronozone. In wells 3 and 4 the sequence boundary forms a composite boundary that consists of the Ru-4/Ch-1 and Ch-2 sequence boundaries. Well-1 lacks biostratigraphic age control within this interval, therefore the correlation is tentative there.

**Age.**

The Ch-2 sequence boundary age originates from linear interpolation of seismic, well-log and chronostratigraphic data (Figs. 8, 10, 11). It corresponds to the upper part of P21 planktonic foraminifera chronozone (high in P21) and to the NP24/ NP25 nannoplankton chronozone boundary. It is calibrated to the middle part of Chron C9n and is dated at 27.5 Ma (Berggren et al., 1995; Figs. 4, 7).

**Correlation.**

This sequence boundary seems to correlate with the sequence boundary TB1.2 (28.5 Ma) of Haq et al. (1988) in the late Oligocene (27.1 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as OCi-2 at 27.3 Ma (Abreu and Haddad, this volume; Fig. 4).

**Ch-3**

**Seismic Signature.**

On seismic line A, within the lower Egerian Eger Formation, younger reflection onlap older reflection (Fig. 10 at 2 km) and older reflection truncate below younger reflection (Fig. 10 at 7 km).

**Interpretation.**

Sequence boundary was interpreted on seismic line A (Fig. 10) and correlated with the sequence boundary identified within the Eger Formation in wells 2, 3 and 4, and within the Szécsény Formation in well-1 (Figs. 8, 9). In wells 2, 3 and 4, this sequence boundary is at the middle part of NP25 nannoplankton chronozone. Well-1 lacks biostratigraphic age control within this interval, therefore the correlation is tentative there.

**Age.**

The Ch-3 sequence boundary age originates from linear interpolation of seismic, well-log and chronostratigraphic data (Figs. 8, 10, 11). The Ch-3 sequence boundary corresponds to the middle part of P22 planktonic foraminifera chronozone and to middle part of NP25 nannoplankton chronozone. It is calibrated to the lower part of Chron C7r and is dated at 25.4 Ma (Berggren et al., 1995; Figs. 4, 7).

**Correlation.**

This sequence boundary seems to correlate with the sequence boundary TB1.3 (26.5 Ma) of Haq et al. (1988) in the late Oligocene (25.4 Ma; Hardenbol et al., this volume) and with an isotopic event referred to as OCi-3 at 25.2 Ma (Abreu and Haddad, this volume; Fig. 4).

**Ch-4/Aq-1**

**Field Reference.**

The upper neritic late Egerian Bretka Limestone unconformably overlies Triassic carbonates (Báldi, 1986). Based on mol-lusk and large foraminifera (*Miogypsina gunteri*) correlation, the Bretka Formation is coeval with the lower Aquitanian (Báldi, 1983). An unconformity occurs between the neritic Törökbalint Sandstone and the first pebbly and sandy beds of the Budafok Sandstone (Báldi, 1983, 1986).

**Seismic Signature.**

On seismic line A, within the Eger Formation older reflection truncates below younger reflection (Fig. 10 at 7 km).

**Interpretation.**

Unconformities occur: (1) between the late Egerian Bretka Limestone and the Triassic carbonates and (2) between the Törökbalint Sandstone and Budafok Sandstone (considered the sequence boundary, Aq-1). This sequence boundary identified on seismic line A (Fig. 10) correlates with the sequence boundary identified within the Eger Formation in wells 2, 3 and 4 and within the Szécsény Formation in well-1 (Figs. 8, 9). In wells 2 and 3, this sequence boundary is in the lower part of NN1 nannoplankton chronozone. Well-1 lacks biostratigraphic age control within this interval, therefore the correlation is tentative there.

**Age.**

The Chattian/Aquitanian boundary (Oligocene/Miocene) corresponds to the FAD of *Globorotalia kugleri* (P22/N4A planktonic foraminifera chronozone boundary) and in the lower part of NN1 nannoplankton chronozone. It is calibrated to the top of Chron C6Cn.2r and is dated at of 23.8 Ma (Berggren et al., 1995; Figs. 4, 7).

**Correlation.**

This sequence boundary seems to correlate with the sequence boundary TB1.4 (25.5 Ma) of Haq et al. (1988) prior the Oligocene/Miocene boundary (24.4 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MAi-1 at 23.8 Ma (Abreu and Haddad, this volume; Fig. 4).

**Aq-2**

**Seismic Signature.**

On seismic line A, within the Eger Formation, younger reflection onlap older reflection (Fig. 10 at 4 km).

**Interpretation.**

Sequence boundary interpreted on seismic line A (Fig. 10) correlates with the sequence boundary interpreted within the Eger Formation in wells 2, 3 and 4 and within the Szécsény Formation in well-1 (Figs. 8, 9). In well-3, this sequence boundary is in the lower part of NN2 nannoplankton chronozone. Well-1 lacks biostratigraphic age control within this interval; therefore, the correlation is tentative there.

**Age.**

The Aq-2 sequence boundary age originates from linear interpolation of seismic, well-log and chronostratigraphic data (Figs. 8, 10, 11). The Aq-2 sequence boundary corresponds to the middle part of N4B planktonic foraminifera biochronozone and to the lower part of NN2 nannoplankton biochronozone. It
Fig. 10.—Uninterpreted and interpreted seismic line A from the Hungarian Paleogene Basin. Seismic and well data published by Lakatos et al. (1992). Location of profile is on Figure 3. Nannoplankton chronozone data from cores provided age control. See interpretation of wells 3 and 4 on Figure 8. Line A includes the interpreted sequence boundaries except the Bur-4 and Lan-1. The profile shows the two major tectonic phases of the Pannonian Basin. Compressional-style tectonics dominated during Kiscellian and Egerian time; a late stage isostatic uplift occurred during Eggenburgian and Ottnangian time. Sequence boundaries...
Ru-2, Ru-3 and Ch-3 represent tectonically enhanced sequence boundaries as a result of repeated folding. Because of the change of tectonic style from compressional to extensional, the late Eggenburgian and Karpatian strata are eroded. The extensional style commenced in the Karpatian Stage (Tari et al., 1992a) and resulted in volcanic sedimentation during the early Badenian Stage (strata between Bur-2—Ser-1/Ser-2).
is calibrated to the middle part of Chron C6Aa2r.1n and is dated at 22.2 Ma (Berggren et al., 1995; Figs. 4, 7).

**Correlation.**

This sequence boundary seems to correlate with the sequence boundary TB1.5 (22 Ma) of Haq et al. (1988) in the upper part of Aquitanian (21.4 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MAi-2 at 22.2 Ma (Abreu and Haddad, this volume; Fig. 4).

**Aq-3/Bur-1**

**Field Reference.**

Sztanó and Tari (1993) observed a non depositional hiatus between the neritic Eggenburgian Pétervására Sandstone and the upper bathyal to lower neritic Egerian (60- to 300-m water depth; Báldi, 1973) Szécsény Siltstone. An unconformity was also described between the late Egerian Törökbálint Sandstone and the Eggenburgian Budafok Sandstone (Báldi, 1973, 1986; Vass et al., 1979).

**Seismic Signature.**

On seismic line A, at the base of the Pétervására Sandstone, younger reflections onlap older reflections (Fig. 10 at 1–2 km).

**Interpretation.**

The unconformity between the Eggenburgian and Egerian Formations is considered a sequence boundary. This sequence boundary interpreted on seismic line A (Fig. 10) correlates with the sequence boundary identified at the base of the Pétervására Sandstone in wells 2 and 3 (Figs. 8, 9). In well-3, this sequence boundary is in the middle part of NN2 nannoplankton chronzone. Well-1 lacks biostratigraphic age control within this interval, therefore the correlation is tentative there. Tari et al. (1992a), and Sztanó and Tari (1993) also interpreted a sequence boundary at the Egerian/Eggenburgian boundary.

**Age.**

The Aquitanian/Burdigalian boundary corresponds to the FAD of *Globigerinoides altiaperturus* (in the lower-middle part of N5 planktonic foraminifera chronozone; Montanari et al., 1991) and to the middle part of NN2 nannoplankton chronzone. It was calibrated to the top of Chron C6Aa.1n and is dated at 20.5 Ma. Checking the original data sets of these descriptions (Berggren et al., 1983; Pujol, 1983), we found the FAD of *Globigerinoides altiaperturus* slightly below the top of Chron C6Aa.2n with an estimated age of 21.1 Ma (Fig. 4).

**Correlation.**

This sequence boundary seems to correlate with the sequence boundary TB2.1 (21.0 Ma) of Haq et al. (1988) at the base of Burdigalian (20.5 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MAi-3 at 21.1 Ma (Abreu and Haddad, this volume; Fig. 4).
HIATUS AND/OR NONDEPOSITION

Field Reference.—
Within the neritic Pétervására Formation, the upper neritic Ilonavölgy Member (coarse gravel sand, conglomerate) unconformably overlies the middle neritic glauconitic Pétervására Sandstone (Báldi, 1983, 1986). The Darno Conglomerate Formation, coeval with the Ilonavölgy Member (Báldi, 1983), unconformably overlies Paleozoic and Mesozoic rocks and shows a transgressive-regressive cycle (Báldi, 1986).

Seismic Signature.—
On seismic line A, within the Pétervására Sandstone, younger reflections onlap older reflections (Fig. 10 at 1 km) and older reflection truncates below younger reflection (Fig. 10 at 13 km).

Interpretation.—
The unconformity between the Pétervására Sandstone and the Ilonavölgy Member and between the Darno Conglomerate and the Mesozoic basement is considered a sequence boundary. The sequence boundary identified on seismic line A (Fig. 10) correlates with the sequence boundary interpreted within the Pétervására Sandstone in wells 2 and 3 (Figs. 8, 9). In these wells, the sequence boundary is in the upper part of NN2 nannoplankton chronozone. In wells 1 and 4, the sequence boundary is a composite boundary which consists of the Aq-2—Bur-4, and the Aq-2—Lan-1 sequence boundaries.

Age.—
The Bur-2 sequence boundary age originates from linear interpolation of seismic, well-log and chronostratigraphic data (Figs. 8, 10, 11). The Bur-2 sequence boundary corresponds to the upper part of N5 planktonic foraminifera biochronozone and to the upper part of NN2 nannoplankton biochronozone. It is calibrated to the middle part of the Chron C6n and has an estimated age of 19.5 Ma (Figs. 4, 7).

Correlation.—
The sequence boundary seems to correlate with an isotopic event referred to as MBi-1 at 19.4 Ma (Abreu and Haddad, this volume). There is no sequence boundary described on the Haq et al. (1988) chart at this position (Fig. 4).

Field Reference.—
The Committee of Hungarian Stratigraphy (1983) described unconformities: (1) between the continental Eggenburgian Csata Gravel Formation and the lower neritic Karpatian Garáb Siltstone Formation, (2) between the marine neritic Eggenburgian Budafök Sandstone Formation and the Ottnangian deltaic-fluvial pebbly and sandy Szászvár Formation.

Seismic Signature.—
On seismic line A, older reflection truncates below younger reflection (Fig. 10 at 1 km).
The unconformity between the Eggenburgian and Ottangian Formations is considered a sequence boundary. This sequence boundary identified on seismic line A (Fig. 10) correlates with the sequence boundary interpreted in wells 2 and 3 (Figs. 8 and 9). In these wells, this sequence boundary is in the middle part of NN3 nannoplankton chronozone. In wells 1 and 4, the sequence boundary is a composite boundary which consists of the Aq-2—Bur-4 and the Aq-2—Lan-1 sequence boundaries.

Age.—

The Eggenburgian/Ottnangian boundary corresponds to the lower part of N6 planktonic foraminifera and to the middle part of NN3 nannoplankton biocronozone (Nagymarosy and Müller, 1988; Steininger et al., 1990). It was calibrated to the lower part of Chron C5En and is dated at 18.7 Ma (Fig. 7). This correlation is different than Steininger et al. (1996), who calibrated this boundary with the base of mammal zone 4 (MN-4) with an estimated age of 18.0 Ma.

Correlation.—

The sequence boundary seems to correlate with an isotopic event referred to as MBi-2 at 18.0 Ma (Abreu and Haddad, this volume). There is no sequence boundary described on the Haq et al. (1988) chart at this position (Fig. 4).

Field Reference.—

The Committee of Hungarian Stratigraphy (1983) described unconformities: (1) between the paralic Ottangian Salgótarján Coal Formation and neritic Karpatian Egyházasgerge Sandstone Formation and (2) between the continental Ottangian Szászvár Formation and the littoral Karpatian Budafa Conglomerate Formation.

Seismic Signature.—

On seismic line B, younger reflections onlap older reflections (Fig. 12 at 10 km) and older reflection truncates below younger reflection (Fig. 12 at 9 km) at the base of the Karpatian strata.
Interpretation.—

The unconformity between the Ottnangian and Karpatian Formations is considered a sequence boundary. This sequence boundary identified on seismic line B (Fig. 12) correlates with the sequence boundary interpreted in well 1 (Fig. 8). In this well, this sequence boundary is at the middle part of NN4 nanoplankton chronozone. In wells 2, 3 and 4 (Figs. 8 and 9), the sequence boundary is a composite boundary which consists of the Bur-3—Lan-1 and the Aq-2—Lan-1 sequence boundaries. At the base of Karpatian units, the first appearance of the endemic *Oncophora (Rhezekia)* mollusk fauna (Báldi-Beke and Nagymarosy, 1979; Bérczi et al., 1988) reflects the third period of endemism of the central Paratethys (Fig. 5). This fauna shows a similarity with the Rupelian *Ergenica* fauna (Báldi, 1986) and is a good indicator of endemism.

Age.—

The Ottnangian/Karpatian regional stage boundary corresponds to the N6/N7 planktonic foraminifera chronozone boundary (Steininger et al., 1996) and to the middle part of NN4 nanoplankton biochronozone (Nagymarosy and Müller, 1988). It is calibrated to the top of Chron C5Dn and is dated at 17.3 Ma (Berggren et al., 1995; Figs. 4, 7).

Correlation.—

This sequence boundary seems to correlate with the sequence boundary TB2.2 (17.5 Ma) of Haq et al. (1988) at the top of Burdigalian (17.3 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MBSi-3 at 17.2 Ma (Abreu and Haddad, this volume; Fig. 4).

Field Reference.—

The Committee of Hungarian Stratigraphy (1983) and Bérczi et al. (1988) described unconformities: (1) between the upper neritic Leitha Limestone Formation and upper neritic Fertőrákos Limestone Formation, (2) between the lower neritic Tekeres Siltstone Formation and the upper neritic Fertőrákos Limestone Formation and (3) between the upper bathyal-lower neritic Báden Clay and the upper neritic Fertőrákos Limestone Formation.

Seismic signature.—

On seismic line B, younger reflections representing the Badenian Makó Formation onlap older reflections (Fig. 12 at 2–4 and 8–9 km).

Interpretation.—

The unconformity between the Badenian Formations is considered a sequence boundary. This sequence boundary, identified on seismic line B (Fig. 12), correlates with the sequence boundary interpreted in wells 1, 2, 3 and 4 (Figs. 8, 9). In well-1, this sequence boundary is in the middle part of NN5 nanoplankton chronozone. In wells 2, 3 and 4, the sequence boundary is a composite boundary which consists of the Lan-2/Ser-1—Ser-3 and the Lan-2/Ser-1—Ser-2.

Age.—

The Karpatian/Badenian boundary corresponds to the N7/N8 planktonic foraminifera chronozone boundary and to the upper part of NN4 nanoplankton biochronozone. It is calibrated to the middle part of Chron C5Chn.2n and is dated at 16.4 Ma (Berggren et al., 1995; Figs. 4, 7).

Correlation.—

This sequence boundary seems to correlate with the sequence boundary TB2.3 (16.5 Ma) of Haq et al. (1988) near the Burdigalian/Langhian boundary (16.4 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MLSi-1 at 16.4 Ma (Abreu and Haddad, this volume; Fig. 4).
Ser-2

Field Reference.—

The Committee of Hungarian Stratigraphy (1983) and Bérczi et al. (1988) described unconformities: (1) between the upper neritic Badenian Fertórákos Limestone Formation and the continental Sarmatian Sajóvölgy Formation and (2) between the neritic Badenian Ebes Limestone Formation and shallow brackish Sarmatian Dombegyháza Formation.

Seismic Signature.—

On seismic line A, older reflections representing the early Badenian Mátra Volcanic Formation truncate below younger reflections of the Sarmatian Hajduszobszló Formation (Fig. 10 at 10, 11, 21 and 22 km). On seismic line B, younger reflections representing the Sarmatian Hajduszobsló Formation onlap older reflections (Fig. 12 at 1, 3, 5 and 9–10 km).

Interpretation.—

We interpret the unconformity between the Badenian and Sarmatian Formations as a sequence boundary on seismic line B (Fig. 12) and in wells 1, 2, 3 and 4 (Figs. 8, 9). In well-1, this sequence boundary is at the base of NN6 nannoplankton chronozone. In wells 2, 3 and 4, the sequence boundary is a composite boundary which consists of the Lan-2/Ser-1-Ser-3, and the Lan-1/Ser-1-Ser-2.

Age.—

The Badenian/Sarmatian Stage boundary corresponds to the middle part of N10 planktonic foraminifera chronozone (Steininger et al., 1996) and to the NN5/NN6 nannoplankton chronozone boundary. It is calibrated to the middle part of Chron C5Ar1r and is dated at 13.6 Ma (Berggren et al., 1995; Figs. 4, 7).

Correlation.—

This sequence boundary seems to correlate with the sequence boundary TB2.5 (13.8 Ma) of Haq et al. (1988) near the base of Serravallian (14.1 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MSi-2 at 13.6 Ma (Abreu and Haddad, this volume; Fig. 4).

Ser-3

Field Reference.—

The Committee of Hungarian Stratigraphy (1983) and Bérczi et al. (1988) described unconformities: (1) between the shallow brackish Sarmatian Hajduszobsló and the shallow brackish Pannonian Békés Conglomerate Formation, (2) between the shallow brackish Tinnye Formation and the shallow brackish Pannonian Békés Conglomerate Formation and (3) between the upper bathyal Sarmatian Dombegyháza Formation and the lower neritic Pannonian Tótkomlós Claymarl Formation.

Seismic Signature.—

On seismic lines A and B, younger reflections representing the Pannonian strata onlap older reflections (Fig. 10 at 7, 15–16 km; Fig. 12 at 1, 2, 6, 7 and 8 km). Also, older reflections truncate below younger reflections (Fig. 10 at 2–5 km; Fig. 12 at 5–7 km).

Interpretation.—

The unconformity between the Sarmatian and Pannonian Formations is considered a sequence boundary and identified on seismic lines A and B (Figs. 10 and 12) and in wells 1, 2, 3 and 4 (Figs. 8, 9). In well-1, this sequence boundary is in the middle part of NN6 nannoplankton chronozone. In well-2, the sequence boundary is a composite boundary which consists of the Lan-2/Ser-1-Ser-3.

Age.—

At the Sarmatian/Pannonian boundary, the Pannonian Basin finally became a brackish lake which complicates the direct biostratigraphic correlation with biochronozones defined in open marine settings. Rögl et al. (1978), Rögl and Steininger (1983) and Steininger et al. (1985, 1990) calibrated this boundary with the boundary between mammal biozones MN8 and MN9, dated radiometrically between 10.5 Ma to 12.3 Ma which correlates approximately with the Serravallian/Tortonian boundary. Vakarcs et al. (1992, 1994), on the basis of seismic and well data, identified a sequence boundary at the Sarmatian/Pannonian boundary and correlated it with the TB2.6 sequence boundary of Haq et al. (1988) chart. Based on magnetostratigraphical data, this sequence boundary was dated at 12.7 Ma (Vakarcs et al., 1994). This boundary corresponds to the N10/N11 planktonic foraminifera chronozone boundary and to the middle part of NN6 nannoplankton biochronozone. It is calibrated to the top of Chron C5Ar.r, which has an estimated age of 12.7 Ma (Berggren et al., 1995; Figs. 4, 7).

Correlation.—

This sequence boundary correlates with the sequence boundary TB2.6 (12.5 Ma) of Haq et al. (1988) in the middle part of Serravallian (13.2 Ma, Hardenbol et al., this volume) and with an isotopic event referred to as MSi-3 at 12.7 Ma (Abreu and Haddad, this volume; Fig. 4).

DISCUSSION

Regional Stages vs. Standard Stages

The sequence boundaries described in this paper support the correlation between the regional and the standard stages previously proposed by Báldi (1986), Nagymarosy and Müller (1988), Steininger et al. (1990, 1996), and Vakarcs et al. (1994). Traditionally, the type sections of the central Paratethys regional stages were described in shelf or upper slope settings (e.g., Steininger and Neveskaya, 1975). Most of the biostratigraphic control initially described from these regional stages is based on benthic foraminifera and mollusks. These types of fauna are sensitive to environmental changes and therefore are not the most appropriate groups for regional or global correlation. As a result of these limitations, the correlation between regional and standard stages remained tentative.

More recently, the calibration of regional stages with planktonic microfossil zonations has advanced greatly. As a result, the correlation between regional and standard stages improved significantly (Steininger et al., 1990, 1996). According to Báldi (1983, 1986), and Steininger et al. (1990, 1996) there is direct correlation between the upper boundaries of the following stages: Kiscellian and Rupelian, Egerian and Aquitanian and
Karpatian and Burdigalian (Fig. 2). Because of the improved correlation between standard and regional stages, a standard stage terminology can be used in the Paratethyan basins.

The age control for the standard stages is based on integrated zonations of planktonic foraminifera and calcareous nannofossils, calibrated to the magnetostratigraphic record of the South Atlantic oceans (Berggren et al., 1995, and selected radiometric dates). Position 3 (basin) marks the deep-sea record in Figure 13. If the conditions are suitable (i.e., moderate sedimentation rate, no deepwater erosion or carbonate dissolution, etc.), this position provides a potentially continuous rock record.

**Regional Stages vs. Depositional Sequences**

From lower Oligocene through middle Miocene time, the central Paratethys was connected to the global ocean except for six short periods of isolation (Figs. 1, 5). In five cases, these periods of isolation and endemism coincide with regional stage boundaries: Egerian/Eggenburgian (Aq-3/Bur-1), Eggenburgian/Ottnangian (Bur-3), Ottnangian/Karpatian (Bur-4), Badenian/Sarmatian (Ser-2) and Sarmatian/Pannonian (Ser-3) (Fig. 4). In all cases, these boundaries are represented by unconformities in shallow environments that become conformable in deeper parts of the basins (Committee of Hungarian Stratigraphy, 1983; Bérczi et al., 1988; Vakarcs et al., 1994). On seismic lines, these stage boundaries are marked by regional unconformities, interpreted as sequence boundaries (Aq-3/Bur-1, Bur-3, Bur-5/Lan-1, Lan-2/Ser-1 and Ser-3).

The Kiscellian/Egerian (Ru-4/Ch-1) and Ottnangian/Karpatian (Bur-4) regional stage boundaries are the most obvious unconformities seen on seismic lines A and B from the Pannonian Basin (Figs. 10, 12). The Kiscellian/Egerian boundary corresponds to the end of the period of major thrusting in the retroarc foredeep basin (Tari et al., 1993). The Ottnangian/Karpatian boundary marks the change in the tectonic style of Pannonian Basin from compressional to extensional (Tari et al., 1992a). We interpret these regional stage boundaries as tectonically enhanced sequence boundaries (Ru-4/Ch-1 and Bur-4).

**Stages and Depositional Sequences**

Stages are chronostratigraphic units that represent all rocks deposited during a defined unit of time. The type section of a stage, or stratotype, in conjunction with stage boundary stratotype designations, defines its chronostratigraphic dimensions. In general, stages are defined in shelfal areas of basins. Stages...
represent units of marine to marginal marine deposits separated by unconformities. Regional stages defined in shallow-marine deposits of the Paratethyan basins represent at least, one or often two or more, depositional sequences separated by unconformities. These depositional sequences can be traced on seismic profiles toward the deeper portions of the basins, where sedimentation is becomes continuous and the bounding surfaces are conformable (Fig. 13). The chronostratigraphic position of sequences bounding the stages should, in theory, be determined at the place were the unconformities become conformable. However, in reality this position is hard to identify and is rarely accessible for detailed chronostratigraphic analysis.

There are several possibilities for the chronostratigraphic position of the stage boundary. It can be placed at the top of the underlying stage, at the base of the overlying stage or somewhere in between. For the chronostratigraphic position of the sequence boundary only the third option is possible. Chronostratigraphic placement of sequence boundary occurs below the upper stage and above the lower stage. To express this uncertainty, sequences near stage boundaries are given prefixes for both stages.

The base of stratotype of the Badenian Stage, a well documented transgressive surface, was described in a neritic environment (Steininger and Neveskaya, 1975). In sequence stratigraphic terms, the chronostratigraphic position of this flooding surface is located near the beginning of the transgressive systems tract, related to Position 1 at time “c” in Figure 13. The Badenian Stage overlies the Karpatian Stage, also identified within a neritic environment (Steininger and Neveskaya, 1975). The reference section of the Karpatian Stage shows a transgressive/regressive facies cycle with an unconformity at the top of the Karpatian section. We interpret this unconformity as Bur-5/Lan-1 sequence boundary, related to Position 1 at time “a” in Figure 13.

In other words, an unconformity occurs at the top of the Karpatian reference section, and a transgressive surface lies at the base of the Badenian reference section (Position 1 at time “a” and “c”, Fig. 13). The Badenian Stage was dated by the first appearance of N8 plankton foraminifera chronozone (Position 1 at time “c”, Fig. 13; Steininger et al., 1990). The top of Karpatian section is placed in the N7 planktonic foraminifera chronozone (Position 1 at time “a”, Fig. 13; Steininger et al., 1990). An age difference exists between the top of Karpatian and the base of Badenian reference section; however, the Karpatian/Badenian Stage boundary is placed as the N7/N8 planktonic foraminifera chronozone boundary. As we discussed earlier, the planktonic foraminifera and the nannoplankton zones were correlated to the deep water DSDP/ODP legs in Position 3 (Fig. 13; Berggren et al., 1995). This boundary lies between time “a” and time “c” and should be close to time “b”, the sequence boundary in Position 3 of Figure 13.

**Sequence Boundaries vs. Glacioeustasy**

Abreu and Haddad (this volume) described 16 isotopic events (Figs. 4, 14) interpreted as glacioeustatic sea-level falls during Oligocene and middle Miocene time. Their interpretations are based on the deep-sea oxygen isotope record (Miller et al., 1991) and the isotope record of an industry well (Well A) in the Campos Basin, offshore SE Brazil (Abreu and Savini, 1994). The isotopic events of Abreu and Haddad (this volume) and the ones described by Miller et al. (1991) show a good correlation with the sequence boundaries proposed in this work and in the Haq et al. (1988) chart.

The correlation between isotopic events and sequence boundaries is based on the premise that the isotopic curve is an approximation of the eustatic curve. The inflection point of the isotopic curve, not the isotopic peak, would be the best position for correlation with the sequence boundaries. However, for practical purposes, the isotopic peak is more commonly used in correlations with the onlap curves (Miller et al., 1991; Abreu and Savini, 1994; Abreu and Haddad, this volume). The time between the inflection point and the peak occurrence in an isotope curve is normally smaller than the resolution of the stratigraphic age control (chronozones and biozones). Moreover, in
most of the cases, the inflection point of an isotopic peak is difficult to determine, while the peak is typically sharp, easier to recognize and more adequate for stratigraphic correlations.

The isotope events are used as an independent and indirect indicator of possible sequence boundaries (Miller et al., 1991; Abreu and Savini, 1994; Abreu and Haddad, this volume). The events provide a global mechanism (glacioeustasy) to generate sequence boundaries at least from Oligocene time onward (Miller et al., 1991; Wright and Miller, 1993; Abreu and Haddad, this volume). We observed a good correlation between stage boundaries, sequence boundaries and oxygen isotope events. These correlations support glacioeustasy as the controlling factor for depositional sequences and global stages, since early Oligocene time (Miller et al., 1991; Abreu and Haddad, this volume).

Glacioeustasy vs. Orogenic Cycles

Well-defined unconformities were observed in the lower Oligocene to middle Miocene strata of the Pannonian Basin at the basin margins, explained by the effect of “Stille-style” (1924) orogenic cycles (e.g., Hámor and Szentgyörgyi, 1981; Bércezi et al., 1988). However, within the basins, these unconformities become conformable (e.g., Nagymarosy, 1981; Bércezi et al., 1988).

A sequence boundary can be developed due to a relative sea-level fall (Vail et al., 1977; Posamentier and Vail, 1988). The relative sea-level fall is a function of eustasy and tectonic subsidence (Posamentier and Vail, 1988). To prove the global behavior of sequences, we must assess whether or not the driving mechanism is global. Tectonism is local and can vary from basin to basin. However, eustasy reflects continental glaciation at least since the upper Eocene (Miller et al., 1991).

As discussed earlier, the Pannonian Basin has two major tectonic phases producing long term transgressive/regressive facies cycles; a compressional phase during Paleogene and lower Miocene deposition, and an extensional period during most of the remainder of Neogene time. A major unconformity occurs near the Ottnangian/Karpatian stage boundary, marking the end of the compressional regime and the beginning of extensional processes. These long-term cycles are different as compared with other basins or to the Haq et al. (1988) second-order cycles.

Although the Pannonian Basin is marked by long term tectonic regimes, all the sequences described by Haq et al. (1988) are present in the Pannonian Basin. The Ru-3, Bur-2 and Bur-3 events are new compared with the Haq et al. (1988) chart. More significant is the correlation between the sequences and the isotope (glacioeustatic) events described by Miller et al. (1991) and Abreu and Haddad (this volume). Such correlation implies that a global (eustatic) mechanism has generated these unconformities (sequence boundaries). These higher frequency oscillations caused by glacioeustatic fluctuations would effect the entire globe.

CONCLUSIONS

Within the central Paratethys, seventeen sequence boundaries were identified from Oligocene through middle Miocene strata. The seven regional stages of the same time period are bounded by unconformities that show a strong correlation with eight sequence boundaries. Four regional stage boundaries correlate with global stage boundaries; these also may represent sequence boundaries. The episodic closure of connections between the European epicontinental seas from Oligocene through middle Miocene time are the result of short-term glacio-eustatic falls and correlate with regional stage boundaries (sequence boundaries). The identified sequence boundaries also correlate well with those of Haq et al. (1988), but we have found three additional sequence boundaries. The Pannonian Basin has two long-term tectonic phases that resulted in second-order transgressive/regressive cycles. The short-term glacio-eustatic falls overprint long-term local tectonics. The identified third-order sequence boundaries show a direct correlation with oxygen isotope events, reflecting the global continental glaciation, at least from upper Eocene time. These facts provide a convincing argument that the sequences described here are also global in nature.

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OLIGOCENE—MIDDLE MIocene STAGES AND DEPOSITIONAL SEQUENCES


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The general characteristics of the depositional sequences recognized in the “Langhe” region (Fig. 3) permit a three-fold subdivision: group A consists of sequences formed by continental and coastal conglomerates, shallow-marine sandstones and open-marine mudstones; group B comprises depositional sequences with turbiditic sandstones in the lower part and homogeneous biogenic marine mudstones with high plankton/benthos ratios (hemipelagites) in the upper part; group C consists mostly of turbidite systems in the central part of the basin, which change to marine mudstones towards the margin areas.

Nannofossils and planktonic foraminifers provided the necessary tool for dating the depositional sequences. About 240 samples have been studied and correlated to the standard sequence of Oligo-Miocene biostratigraphic events as well as to the zones recognized by Blow (1969) and Iaccarino (1985) for the planktonic foraminifers and Okada and Bukry (1980) for the nannofossils.

**Group A Sequences**

Two depositional sequences can be recognized in group A (the Molare Formation and Rocchetta Formation pro parte of Sheet 81 Ceva, Geologic Map of Italy): a lower one made up of continental to coastal deposits and an upper one with continental to open-marine deposits. These sequences are particularly well-developed in two areas:

In the Bagnasco-Seacchello area, the lower group A sequence consists of proximal alluvial fan conglomerates, with a maximum thickness of 130 m, passing upward to lacustrine sandstones and mudstones up to 70 m thick.

This sequence unconformably lies on a pre-Cenozoic basement and is overlain by a second sequence which begins with conglomerates which pass upward into alternating conglomerates and sandstones organized in repetitive cycles, for a total thickness of over 400 m. The succession represents an alluvial fan depositional system (intermediate and distal). It is overlain by a transgressive unit consisting of better sorted conglomerates alternating with marine fossil-bearing sandstones forming an interval of several tens of meters thick, which locally directly overlies the pre-Cenozoic basement. This sand interval is over-
lain by open-marine mudstones representing the upper part of the sequence. In the Spigno Monferrato area, the lower sequence overlies a pre-Cenozoic substrate and has erosional contact. It is about 20 m thick and consists of poorly sorted, coarse conglomerates, grading upward into finer grained, well sorted conglomerates. The sequence is truncated at the top by an erosional surface. The overlying sequence consists of continental conglomerates and breccias followed by an up to 20 m thick transgressive succession (biolithites, fossiliferous conglomerates and sandstones) and finally by open-marine mudstones, about 150 m thick.

Biostratigraphic data.—Biostratigraphic data are available only for the upper hemipelagic mudstones, sampled near Spigno Monferrato, Mombaldone and Millesimo. The observed faunas and the sporadic occurrence of *Globigerina ampliapertura* can be ascribed to Zone P20. At the top of group A sequences, the first occurrence of *Paragloborotalia opima opima* is recorded, indicating an age in Zone P21. This is confirmed by the occurrence of nannofossils typical of Zone CP19A (Okada and Bukry, 1980): *Cyclicargolithus floridanus*, *C. abisectus*, *Dictyococcites bisectus* and *Reticulofenestra lockeri* predominant with subordinate *Sphenolithus ciperoensis* and *S. distentus*.
SEQUENCE STRATIGRAPHY OF THE “LANGHE” OLIGO-MIOCENE SUCCESSION, TERTIARY PIEDMONT BASIN

Group B Sequences

Group B consists of six depositional sequences (B1 to B6) characterized by turbiditic sandstones and subordinate amounts of reworked conglomerates in the lower part and by hemipelagic mudstones in the upper one. The turbiditic bodies tend progressively to change their geometry from lens-shaped to tabular up-section. The maximum thickness of group B sequences is over 1,000 m. A complex and irregular basin physiography is made clear by paleocurrent analyses (Fig. 4; Gnaccolini, 1968; Cazzola et al., 1981; Cazzola and Fornaciari, 1990), which show conspicuous variability over short distances.

Part of the Rocchetta Formation and the Monesiglio Formation of Sheet 81, Ceva, of the Geologic Map of Italy belong to this group.

Sequence B1.—

In the upper Bormida River valley, its lower boundary is an erosional surface deeply cut into the mudstones and locally the conglomerates of the underlying group A sequences. Sometimes, it may even reach the pre-Cenozoic basement. The base of the sequence consists of matrix-supported conglomerates with muddy matrix and interspersed clasts up to a meter in size. Resedimented, locally very coarse conglomerates follow, grading to coarse-grained, graded sandstones. The succession is up to 170 m thick near Biestro; whereas to the north, between Millesimo and Bormida, its thickness decreases abruptly down to a few tens of meters. These deposits form a lens-shaped body, representing the filling of a lower part of a submarine canyon. Sequence B1 terminates with hemipelagic mudstones (about 60 m thick) intercalated with very thin-bedded turbiditic sandstones.
Near Mombaldone, sequence B1 consists of a 40-m-thick turbiditic sandstone body that forms a gentle anticline. The widespread cover hinders the observation of the lower boundary. The upper one is marked by an erosional surface overlain by onlapping mudstones of the following sequence (Figs. 5, 6).

**Biostratigraphic Data.**—The mudstones (Spigno Monferrato, Millesimo) yielded associations typical of the upper part of Zone P21, that are characterized by the occurrence of Paragloborotalia opima opima, Catapsydrax dissimilis, Globoquadrina tripartita and Globigerina venezuelana. The most common nannofossils are Dictyococcites bisectus, Cyclicargolithus abisectus, C. floridanus, Reticulofenestra lockeri and rare Helicosphaera recta; zonal marker are lacking. This association can be tentatively referred to CP19B.

**Sequence B2.—**

In the Bormida di Millesimo River valley, this sequence begins with a lens-shaped body of turbiditic sandstones, about 65 m thick (“Cengio Turbidite System”, Cazzola et al., 1981). Sandstones seem to fill and onlap a depression created by the gentle flexure of the mudstones at the top of sequence B1 (Fig. 7). The sequence ends with mudstones and subordinate very thin-bedded turbiditic sandstones (30 m thick).

In the Bormida di Spigno River valley, near Mombaldone, sequence B2 consists of mudstones, about 100 m thick, overlying the erosional surface at the top of sequence B1. Between Mombaldone and Spigno Monferrato, the lower part of sequence B2 is marked by a huge lens of poorly sorted ophiolitic breccias (with clasts up to 1 m in diameter), about 45 m thick and with a visible lateral extent of more than 80 m (Fig. 8). These breccias fill a canyon deeply cut into the underlying B1 mudstones.

**Biostratigraphic Data.**—The entire sequence was sampled near Mombaldone, where it consists exclusively of mudstones. Other samples were taken at Spigno Monferrato and in the Bormida di Millesimo valley for control. At the sequence B2 base, Paragloborotalia opima opima still occurs allowing attribution to Zone P21. A few meters above this, the first occurrences of Globigerinoides primordius and Globoquadrina praedehiscens are recorded, indicating the P22 Zone.

**Sequence B3.—**

In the Bormida di Millesimo River valley, the B3 sequence, up to 300 m thick, starts with turbiditic sandstones. The upper part is represented by mudstones. Near Mombaldone, in the Bormida di Spigno River valley, turbiditic sandstones at the base of the B3 sequence onlap the gently flexed mudstones at the top of the underlying B2 sequence (Fig. 9).

**Biostratigraphic Data.**—Samples collected in the Bormida di Spigno and Bormida di Millesimo valleys show the same characteristics as the upper part of B2. Globoquadrina praedehiscens, Catapsydrax, large Globigerinids and Globigerinoides primordius are the dominant forms, which still indicate P22.

**Sequence B4.—**

The lower part of sequence B4 consists of the “Noceto Turbidite System” of Cazzola and Fornaciari (1990) with a maximum thickness of about 350 m. The upper part is a muddy succession 30 to 100 m thick.

**Biostratigraphic Data.**—Globigerinae ciperoensis gr. are very common associated with faunas similar to sequence B3, thus still representing P22 Zone. The bloom of Globigerina ciperoensis gr. is considered a significant biostratigraphic event (latest Oligocene in age) in the whole TPB. The flora association yielded by sequence B4 can be tentatively referred to Zone CP19B.

**Sequence B5.—**

Sequence B5 may be clearly distinguished from B6 only west of the Bormida di Spigno River. It consists of basal turbiditic sandstones, up to 150 m thick, with some intercalated biocalcareite graded beds, succeeded by 40 m thick mudstones.

**Biostratigraphic Data.**—The succession of biostratigraphic events in sequence B5 is as follows:

- occurrence of Globorotalia gr. kugleri (recorded only in a section sampled near Spigno);
- first occurrence of Globoquadrina dehiscens;
- differentiation of Globigerinoides (G. des immaturus, G. des trilobus, G. des sacculifer, G. des quadrilobatus). These data extend the range of sequence B5 from the top of Zone P22 to N5.

**Sequence B6.—**

Sequence B6 in the Bormida di Millesimo River valley consists of the turbiditic sandstones of the “Arbi-Gottasecca sandstone body” (about 25 m thick; Gelati and Gnaccolini, 1980) overlain by 130 m of mudstones. In the eastern part of the studied area, near Mombaldone, sequence B6 begins with glau-
conitic arenites and slumped biocalcarenites, overlain by mudstones with a total thickness of 80 m. Further east, near Visone, sequence B6 lies on sequence B4 with a major angular unconformity. Here sequence B6 begins with platform biocalcarenites that are up to 20 m thick. They are overlain by glauconitic sandstones grading upward into mudstones 15–20 m thick.

Biostratigraphic Data.—The base of sequence B6 records a bloom of calcareous nannoplankton species *Helicosphaera am-
Fig. 7.—The boundary between B1 and B2 sequences at Cengio, Bormida di Millesimo River valley (m—mudstones, t—turbiditic sandstones; sandstone maximum thickness 5 m).

Fig. 8.—A coarse ophiolite breccia (ob; 45 m thick) fills a canyon deeply incised in the underlying mudstones (m). Boundary between B1 and B2 sequences, near Spigno Monferrato.
FIG. 9.—The boundary between B2 and B3 sequences near Mombaldone. The turbiditic sandstones (t) of the B3 sequence onlap the gently folded mudstones (m) of the underlying sequence. The sandstones layer marked by an asterisk is 2 m thick.

*pliaperta, H. carteri, H. minuta, Reticulofenestra pseudoumbilica* (CN1C Zone) near Mombaldone. In the mid sequence the first occurrence of *G. des altiaperturus* (indicating the Aquitanian-Burdigalian boundary according to Iaccarino, 1985) and the first occurrence of *Sphenolithus belemnos* (base of CN2 Zone) are recorded. Thus, sequence B6 is probably latest Aquitanian-early Burdigalian age.

It is noteworthy that group B sequences may be represented solely by mudstones with intercalated very thin-bedded turbidites. This is due to lateral thinning at the end of turbidite bodies. For example, this thinning can be observed between Montechiaro d’Acqui and Cartosio, where group B mudstones are about 150 m thick and onlap the underlying, faulted and tilted conglomerates of group A (Fig. 10). Deposition of these mudstones probably occurred on highs or slopes located at the sides of the basin, and they are the lateral equivalent of more than 1,000 m of sediment deposited within the depocenters.

Group C Sequences

Group C sequences are over 2,200 m thick and comprise the Cortemilia, Paroldo, Cassinasco, Cessole, Murazzano and Léguio Formations (Sheet 81, Ceva, of the Geologic Map of Italy). In the central part of the basin, group C sequences mostly consist of turbidite systems with sandstone/mudstone ratios from >1 to 1. Towards the margins of the basin, the upper part of the sequences, or even the whole sequence, may consist of mudstone. Group C sequences (Late Burdigalian-Early Tortonian) show regular, basin-wide turbidite systems that are mostly tabular in geometry. Paleocurrent directions are relatively uniform: from west to east for sequence C1; from south-west to north-east for the overlying sequences (Fig. 4). This indicates a regular and uniform physiography for the bottom of the basin.

Sequence C1.—

Sequence C1 is up to 800 m thick and consist of a basin plain turbidite system, characterized by alternating sandstones and mudstones with sand/mud ratios about or less than one (Fig. 11). This succession crops out over the entire study area and continues to the east. In the “Langhe” region, the C1 sequence corresponds to the Cortemilia Formation. Between the Belbo stream and Tanaro River valleys, this sequence laterally changes to mudstones, which can be interpreted as slope deposits (mapped as Paroldo Marl in the Geologic Map of Italy, Sheet 81). The transition is relatively rapid, especially in the lower part of the sequence, since it can be observed at Monsiglio, in the Bormida di Millesimo River valley.

The lower boundary of sequence C1 has been placed where turbiditic sandstones begin to rythmically alternate with mudstones. The upper boundary, at least in the depocenter, is marked by an increased sandstone/mudstone ratio to a succession mostly consisting of turbiditic sandstones. Locally, as near
Fig. 10.—Mudstones belonging to group B sequences unconformably overlie the block-faulted conglomerates of the group A, between Montechiaro d’Acqui and Cartosio.

Fig. 11.—Rhythmically alternating sandstones and mudstones of sequence C1 between Serole and Cortemilia. Scale bar ≈ 1 m.

Gorzegno (Bormida di Millesimo River valley), the upper boundary is located at the base of a 13m-thick pebbly mudstones containing sandstone olistoliths up to 5–6 m in size.

Biostratigraphic Data.—The base of sequence B1 has been sampled in the Bormida, Belbo and Tanaro valleys. The first occurrence of *Globigerinoides bisphericus* is recorded in all these sections. This event is correlatable with the upper part of the *Globigerinoides trilobus* Zone of Iaccarino (1985), corresponding to the upper part of Blow’s (1969) Zone N7 (late Burdigalian age). The nannofossil association supports this dating, being characteristic of the CN3 Zone, with *Helicosphaera ampliaperta* and *Sphenolithus heteromorphus*.

Sequence C2.—

In the Bormida di Millesimo River valley, south-west of Cessole, sequence C2 consists of turbiditic sandstones, sometimes amalgamated or alternating with subordinate mudstone layers, which form a system of basin-wide lobes. In this area, the se-
sequence thickness is probably more than 600 m; however, its upper boundary cannot be traced with certainty.

In the northeastern margin of the study area, near Cessole, sequence C2 is about 400 m thick and is dominated by mudstones. These deposits can be interpreted as sediments deposited on a turbidite devoid slope. Sequence C2 corresponds to part of the Murazzano and Cassinasco Formations and to the Cessole Formation (Sheet 81, Geologic Map of Italy).

Biostratigraphic Data.—Sequence C2 is Langhian in age and, in the Cessole section, is the Langhian stratotype. According to the correlation proposed by Haq et al. (1988), sequence C2 should be latest Burdigalian to earliest Serravallian age. Sampling was carried out at Cessole and Gorzegno, in the Bormida di Millesimo River valley, near Mombacaro and in the Tanaro River valley. At the base of the sequence, the occurrence of the genus Praeorbulina is recorded; towards the top, Orbulina suturalis is found. Therefore, the sequence can be ascribed to the Praeorbulina glomerosa and Orbulina suturalis zones of Iacccarino (1985), equivalent to Zone N8 and the lower part of Zone N9 of Blow (1969). The base of the sequence is characterized by the disappearance of the nannofossil Helicosphaera ampliaperta (top of the CN3 Zone), while the rest of succession yields Sphenolithus heteromorphus, indicative of the CN4 Zone.

Sequence C3.—

The beginning of sequence B3 is clear near Cessole (Bormida di Millesimo River valley) because of the occurrence of conspicuous turbidite deposits (Fig. 12) onto the mudstones belonging to sequence C2. Elsewhere, the lower boundary is difficult to identify, because of the monotonous turbidite succession.

Sequence C3 is about 500 m thick and always consists of dominant sandstones in the lower part (basin-wide lobes). The upper part mostly consists of mudstones, which can be recognized throughout the study area. The C3 sequence makes up part of the Murazzano, Cassinasco and Lequio Formations (Sheet 81, Geologic Map of Italy).

Biostratigraphic Data.—The mudstones yield Orbulina univer-

sae indicating a Serravallian age according to all the biostratigraphic zonations considered. Nannofossils support this age. Sphenolithus heteromorphus is still present along with predominant long-ranging forms as Cyclicargolithus floridanus, Helicosphaera carteri, Coccolithus pelagicus, Reticulofenestra pseudoumbilica.

Sequences C4, C5 and C6.—Sequence C4, which is up to 140 m thick, is clearly recognizable in the Belbo stream valley (western side), where it consists of basal turbiditic sandstones overlain by mudstones. The sudden renewal of turbidite sedimentation marks the beginning of sequence C5. The C5 turbidite succession is up to 130 m thick and makes up the entire C5 sequence. It exhibits a clear upper boundary. In fact, sequence C6 begins with an over 15m-thick unit consisting of pebbly mudstones containing sandstone/mudstone olistoliths, up to 20 m in length (Fig. 13).

Sequence C6 consists mainly of turbiditic sandstones (over 90 m thick) in its lower part. They are overlain by mudstones. The upper sequence boundary has not been taken into consideration in this paper.

Fig. 12.—Thick turbidite sandstone layers in the lower part of sequence C3 near Cassinasco. Scale bar = 1 m.
Biostratigraphic Data.—Mudstones of the upper part of sequences C4 and C5 were sampled near Somano and on the western slope of the Belbo stream valley. The only diagnostic planktonic foraminiferal is *Orbulina universa*. Nannofossil associations are rich but are characterized by long-ranging forms (*Cyclicargolithus floridanus*, *Reticulofenestra pseudoumbilica*, *Helicospæra carteri*, *Coccolithus pelagicus*). The only event recorded in these successions is the first occurrence of *Reticulofenestra pseudoumbilica* var. *gelida*, the stratigraphic significance of which is, however, still undetermined.

Two sections were studied in sequence C6 near Albaretto della Torre and Somano. In the upper muddy part, *Globorotalia* gr. *menardii* gradually increases reaching layers which surely belong to Blow’s (1969) Zone N15. This is indicative of the Late Serravallian-Early Tortonian age of the sequence. Nannofossil associations do not allow further precision.

Environmental Indicators

Benthic foraminifers were also considered, mostly as bathymetric indicators. In the upper part of group A and in group B, the most common forms are: *Cibicidoides*, *Anomalinoides*, *Gyroidinoides*, *Lenticulina*, *Stilostomella*, *Heterolepa*, *Uvigerina*, *Planulina* and *Karrieriella*. The plankton/benthos ratio varies between 2 and 6. In the mudstones of group C, *Lenticulinae* are very common with a great number of species, followed by *Nodosariids* and *Uvigerinaceae*, with a mean plankton/benthos ratio of 8.

Similar associations are considered indicative of the upper bathyal zone (Van Morkhoven et al., 1986) with depth range between 200 and 600 m. The presence of *Uvigerina* in the whole succession is particularly important because it may exclude depths shallower than 200 m.

The sporadic presence of shelf foraminifers (rare *Miliolids* in all the sequences; *Elphidium* in group C), always associated with bathyal forms, emphasizes the importance of resedimentation phenomena in the mudstones.

**SYNSEDIMENTARY TECTONICS**

Evidence of tectonic control on sequence boundaries and on both lateral and vertical evolution of facies within the sequences can be observed in several locations and different stratigraphic positions:

1. sequence boundary marked by angular unconformity;
2. sequence boundary characterized by the onlap of turbiditic sandstones or of mudstones on gently folded deposits of the underlying sequence; and
3. the lack, within the sequence, of the vertical and lateral facies distribution typical of eustatically controlled sequences (cf. Posamentier and Vail; 1988; Van Wagoner et al., 1990; Vail et al., 1991).

Clear angular unconformities can be observed at the boundary between sequences B1 and B2 and at the lower boundary of sequence B6. The first case is particularly well exposed near Mombaldone, in the wide meander of the Bormida di Spigno River. Here, the mudstones of B2 directly and unconformably overlie the sandstones of B1 (Figs. 5, 6), which are gently folded and intensely fractured. The contact is marked by an erosional surface, cutting into the sandstones with about 30 m of relief.

A clear angular unconformity can be observed also at the contact between the biocalcarenites at the base of sequence B6 and the underlying mudstones, near Visone, in the easternmost part of the study area. In the Bormida di Spigno valley, resembed biocalcarenites, at the base of sequence B6, seal a block faulting structure affecting the underlying sequences.

Turbiditic sandstones onlap gently folded mudstones and mark the boundaries between sequences B1 and B2 near Cengio (Fig. 7), in the Bormida di Millesimo River valley, and between B2 and B3 near Mombaldone (Bormida di Spigno River valley; Fig. 9).

C1, C2 and C5 sequences seem to be very different from the model for facies evolution proposed for eustatically controlled sequences. In fact, especially at the depocenters, these sequences consist only of turbidite systems (Fig. 11) as can be observed in the following areas: between Pezzolo Valle Uzione and Cortemilia and between Monesiglio and Torre Bormida; on the left side of the Bormida di Millesimo River valley, between Torre Bormida and Cravanzana; between Cerreto and Lequio Berria on the left side of the Belbo stream valley. In this turbiditic succession the sequence boundaries generally are marked by a sudden increase of the sandstones/mudstones ratio and locally also by the occurrence of widespread pebbly mudstones intervals containing olistoliths.

The differences between the facies of sequences C1, C2 and C5 and the model could be the result of synsedimentary tectonic activity. Nevertheless, it could be also related to the occurrence of hiatuses at the bottom and at the top of each turbidite succession, as will be illustrated in the following sections.

**DEPOSITIONAL SEQUENCES IN THE “LANGHE” REGION: LOCAL OR GLOBAL?**

The tentative age range for the studied sequences is shown in Figure 14. From this figure, it appears that some of the sequences (B5, B6, C1, C2 and possibly B1) can be correlated with the 3rd-order global cycles (Haq et al., 1988) as follows:

- sequence B5 is likely to correspond to the 3rd-order cycle 1.4 (+1.5?) (Supercycle TB1);
- sequence B6 correlates to cycle 2.1 (Supercycle TB2);
• sequence C1 can be related to the 3rd-order cycle 2.2 (Supercycle TB2);
• sequence C2 can be related to cycle 2.3 (Supercycle TB2); and
• sequence B1 is tentatively correlated to the cycle 1.1 (Supercycle TB1).

For sequences B2–B3–B4 and C3–C4–C5, biostratigraphic data do not allow an accurate age determination of their boundaries, but these sequences also show frequencies comparable to 3rd-order global cycles. These data could suggest a relation to glacioeustatic sea-level oscillation (cf. Bartek et al., 1991). Nevertheless, taking into account the ubiquitous evidence of synsedimentary tectonic activity, we believe that tectonics played an important role, at least in enhancing the effect of eustatic sea-level falls.

DEPOSITIONAL SEQUENCES AND SYSTEMS TRACTS

The internal organization of sequences recognized in the Oligo-Miocene deposits of the “Langhe” region varies from the bottom to the top of the succession (Fig. 3).

In group A, the lower sequence is wholly represented by alluvial and lacustrine deposits, locally overlain by coastal sediments. In the upper sequence, continental deposits are followed by shallow-marine sandstones passing upward, abruptly, to hemipelagic mudstones. This passage records the rapid relative sea-level rise characteristic of the transgressive systems tract (“top lowstand surface” or “top shelf-margin surface” of Vail et al., 1991). The mudstones at the top of the sequence thus represent both the transgressive systems tract (TST) and the highstand systems tract (HST).

The other sequences recognized (group B and group C) consist of open marine deposits, with the exception of the most marginal part of sequence B6. Group B and Group C sequences are organized according to the following two types.

The first type is characterized, in its lower part, by turbidite systems and, in its upper part, by mudstones with intercalated thin turbiditic sandstone layers. This type comprises group B sequences and part of those of group C (C3, C4, C6).

The second type consists exclusively of turbidites in the depocentre of the basin, while mudstones were deposited on the slopes. Sequences C1, C2 and C5 follow this type.

As for the first type, turbidite systems dominate the lowstand systems tract (LST), while mudstones may extend from the top of the LST into the HST. This situation is analogous to that described for divergent margins by Vail and coworkers (1991). However, the characteristics of deep-sea sandy depositional systems described by these authors are poorly understood; we refer the reader to the critical analysis by Mutti (1992), who infers that LST sandstone deposits in divergent margin settings may consist of contourites and not turbidites.

The internal organization of the second type can be interpreted in two different ways:

1. If the turbidite succession in the depocenter is not bound by hiatuses, turbidity currents predominate. Sands transported by these currents often bypass the slope; whereas on the slope, muds from diluted turbidity current tails and from grain-to-grain hemipelagic settling prevailed. In this case, the internal organization of the sequence would clearly be
different from the classical model (cf. Posamentier and Vail, 1988; Van Wagoner et al., 1990).

2. If the turbidite succession in the depocenter is bound by hiatuses, it could represent only the LST. Deposits connected to the following systems tracks would be missing in the central part of the basin, but would be represented in the marginal zone by mudstone units.

The biostratigraphic data does not distinguish between these two hypotheses. As physical evidence of hiatuses is lacking, the first one can be accepted, provisionally.

CONCLUDING REMARKS

In the “Langhe” region, which is a part of the episutural Tertiary Piedmont Basin, fourteen depositional sequences (mainly Early Oligocene-Early Torrionian) were recognized. Some of these sequences can be correlated with the 3rd-order global cycles of Haq et al. (1988) occurring in conjunction with intense synsedimentary tectonics.

Group A sequences (Early Oligocene; locally also Late Eocene?) mark the beginning phase and subsequent deepening of the basin, whose later evolution is represented by the sequences of group B and C. Group B sequences (Late Oligocene-Early Burdigalian) are characterized by the progressive tendency of turbidite sandstone bodies to change geometry from lens-shaped to tabular. The complex bottom physiography is indicated by the paleocurrent variability. Group C sequences (Late Burdigalian-Early Torrionian) show more regular, mostly tabular basin-wide turbidite systems. Paleocurrent directions is relatively uniform: from west to east for sequence C1 and from south-west to north-east for the following sequences. The change in geometry of the turbidites from the group B sequences to the group C sequences may indicate a marked change in the regional geodynamic regime from a dominant extensional synsedimentary regime to a compressional one, with an increase in subsidence rate and terrigenous input.

The evidence of synsedimentary tectonic (occurrence of angular unconformities, sequences made up wholly by turbidites, synsedimentary faults) allow the identification of tectonic phases. By correlation to the regional tectonic activity, these phases may represent the acme of deformation at the 3rd-order cycle scale. The relation between changes in geometry of the sedimentary basin and deep crustal deformation in the study area (subcrustal indentation, “crocodile pattern”), between Alpine and Apenninic crusts) has been shown by Biella et al. (1992).

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REFERENCES

GLACIOEUSTATIC FLUCTUATIONS: THE MECHANISM LINKING STABLE ISOTOPE EVENTS AND SEQUENCE STRATIGRAPHY FROM THE EARLY OLIGOCENE TO MIDDLE MIOCENE

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ABSTRACT: One of the most difficult challenges of sequence stratigraphy is the establishment of synchrony between events observed in widely separated depositional basins. Problems arise primarily because the chronostatigraphic control in most passive margins is not adequate to constrain the ages of sequence boundaries to better than plus or minus a few million years. This resolution is often insufficient for the correlation of third-order sequences. Furthermore, unless a common mechanism affecting eustasy is assumed, such as variations in the volume of ice on the planet, there is no a priori reason to expect that sequences of similar age in widely separated basins are indeed synchronous.

The stable oxygen isotope composition (\(\delta^{18}O\)) of marine carbonates is an independent proxy for ice volume (sea level) which has been under utilized in sequence stratigraphic analyses. This is somewhat surprising given the number of studies that have established a good relationship between foraminifera \(\delta^{18}O\) and ice volume in Pliocene to Pleistocene units. This paper builds on the work of Miller et al. (1987, 1991) and Abreu and Savini (1994) in identifying major Oligocene to middle Miocene isotope events and correlating them to sequence stratigraphic events. Identification of isotope events is based on \(\delta^{18}O\) data from DSDP sites 522, 529, 563, and 608, and ODP Site 747, drilled in abyssal water depths in the Atlantic and Indian oceans. These isotope records were used by Miller et al. (1991) to define Oligocene and Miocene oxygen isotope zones. In addition to the DSDP/ODP sites, we present oxygen and carbon isotope data from Petrobras Well A drilled in bathyal water depths in the Campos Basin on the Brazilian passive continental margin. Detailed biostratigraphy indicates that this well contains a fairly complete Oligocene to middle Miocene record.

Ages of common isotope events in DSDP and ODP sites and Well A correspond remarkably well with the ages of Oligocene to middle Miocene sequence boundaries identified by Hardenbol et al. (this volume) and Vakarcs et al. (this volume) and correlated to the new time scale of Berggren et al. (1995). Because of the good correlation between the isotope and sequence stratigraphic records, we reconfirm that ice-volume change is the common mechanism driving both oxygen isotope and sea-level fluctuations from Oligocene to present time. We propose four previously unidentified early Oligocene to middle Miocene heavy oxygen isotope events that correlate with sequence boundaries identified in the Pannonian Basin (Vakarcs et al., this volume) and presented in the new cycle chart of Hardenbol et al. (this volume). Additionally, we suggest new chronostratigraphic positions for most of the heavy oxygen isotope zonal boundaries observed previously by Miller et al. (1991). We also present the chronostratigraphic positions for minimum ice-volume events (maximum flooding surfaces) determined from the isotopic record.

INTRODUCTION

Vail et al. (1977) defined a depositional sequence as a stratigraphic unit composed of a relatively conformable succession of genetically related strata bounded at the top and base by unconformities or their correlatable conformities defined as sequence boundaries. Vail (1987) later called the depositional sequence a third-order sequence (duration of 0.5 to 3 my) within a hierarchy of first-order (duration >50 my) to sixth-order (duration 0.01 to 0.03 my) cyclicity. Variation of glacioeustasy was proposed as the driving mechanism for third-order depositional sequences throughout Phanerozoic time (Vail, 1987). Assuming a common mechanism for the development of depositional sequences permitted the correlation of third-order sequences in widely separated depositional basins and led to the age assignments shown for third-order sequence boundaries and maximum flooding surfaces in the cycle chart of Haq et al. (1988).

Vail’s assumption of a glacioeustatic mechanism for third-order sequences has met considerable skepticism. At the time when third-order sequences were proposed, most geologists believed that the Cretaceous and Cenozoic times were essentially ice free until middle Miocene time. However, a revolution in our knowledge of polar glaciations took place during the 1970s when long pelagic sediment cores with unprecedented stratigraphic continuity were recovered and studied. Using stable oxygen isotope measurements of foraminifers, Emiliani (1955) and Shackleton and Opdyke (1973) demonstrated that there have been many more Pleistocene glaciations than the four identified in continental records from North America and Europe.

Stable isotopic measurements of longer deep-sea records recovered by the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Program (ODP) have yielded paleoclimate information that has revised our ideas about Tertiary glaciations. Shackleton and Kennett (1975) and Savin and Savini (1975) suggested that the \(\delta^{18}O\) enrichment in foraminifera at approximately 14 Ma indicates the initiation of the Antarctic ice sheet. However, many investigators contend that significant ice development had already occurred by this time (e.g., Matthews and Poore, 1980; Miller et al., 1987, 1991; Prentice and Matthews, 1988; Denton et al., 1991; Woodruff and Savin, 1991; Barron et al., 1991). Continued isotopic research, complemented by sedimentologic and seismic studies of the Antarctic margin and Southern Ocean, now suggests the presence of ice sheets on East Antarctica at least since early Oligocene and probably as early as late middle Eocene time (Poore and Matthews, 1984; Prentice and Matthews, 1988; Hambrey et al., 1989; Miller et al., 1991; Barron et al., 1991; Ehrmann and Mackensen, 1992; Denton et al., 1991; Wright and Miller, 1992). Although the importance of glaciations as a driving mechanism for third-order eustatic fluctuations is questionable during much of the Phanerozoic time, there is little doubt that glaciations have been important during the “ice house world” (Fischer, 1984) extending from at least the early Oligocene to Recent time.

Stable isotopes can provide a measure of sea-level (ice volume) change during the early Oligocene to middle Miocene time that is independent of sequence stratigraphic analysis. Increases in ice volume inferred from stable isotope records can

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be used to indirectly define sequence boundaries. The most significant contribution for understanding the relationship between ice volume and the origin of Oligocene to Miocene sequence boundaries is the work of Miller et al. (1987, 1991). Miller et al. (1991) recognized twelve δ18O maxima during early Oligocene to upper Miocene time and defined them as isotope zone boundaries representing significant Antarctic glaciations. These δ18O maxima were based on analyses of benthic foraminifer tests picked from DSDP and ODP pelagic sediment cores. Independently, Abreu and Savini (1994) identified eight early Oligocene to early Miocene isotopic events and correlated them to sea-level falls. Well cuttings of pelagic marls deposited in upper bathyal paleoenvironments of the Campos Basin (Brazilian continental margin) were analyzed for oxygen and carbon isotopes. Abreu and Savini (1994) correlated the isotope events with the ones proposed by Miller et al. (1987). In this paper, we correlated the events defined by Abreu and Savini (1994) with the isotope zones of Miller et al. (1991). The correlation between the isotope events determined in the DSDP/ODP isotope record (Miller et al., 1991) and the ones determined in an industry well in Campos Basin (Abreu and Savini, 1994) is remarkable (Table 1).

We also identified δ18O minima and demonstrated the good correlation between the δ18O maxima with sequence boundaries and δ18O minima with maximum flooding surfaces shown on the cycle charts of Haq et al. (1988) and Hardenbol et al. (this volume), and sequences in the Pannonian Basin of Hungary identified by Vakarcs et al. (this volume).

The results of our study are consistent with previous works suggesting that variations of polar ice volume were the driving mechanism for early Oligocene to middle Miocene sea-level fluctuations. Given this common mechanism for stable isotopic fluctuations and third-order sequences identified in Haq et al. (1988), Vakarcs et al. (this volume) and Hardenbol et al. (this volume), it is reasonable to assume synchronism between similar aged sequence stratigraphic events from the early Oligocene to middle Miocene time. We test this assumption by combining careful and detailed analyses of existing biostratigraphic and magnetostratigraphic data from DSDP/ODP sites and an industry well to assign ages to Oligocene to middle Miocene sequence boundaries and maximum flooding surfaces and define δ18O events.


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### Notation and Terminology

It is difficult to determine the absolute abundance of such minor isotopes as 16O and 13C, or the absolute value of the isotopic ratios 18O/16O and 13C/12C, with sufficient accuracy for most paleoclimatological applications. However, it is possible to measure directly and precisely by mass spectrometer the difference between isotopic ratios of two different samples. For this reason, isotope abundance is reported as differences between isotopic ratios of a sample relative to a standard. Notation for the relative difference in isotopic ratios for oxygen is given by the equation:

$$\delta^{18}O = \left( \frac{^{18}O}{^{16}O}_{\text{sample}} - \frac{^{18}O}{^{16}O}_{\text{standard}} \right) \times 1,000$$

The standard used to report both oxygen and carbon isotopic measurements of carbonate samples is CO2 released from the dissolution of Belemnita americana calcite from the Cretaceous Pee Dee Formation in South Carolina (PDB, Urey et al., 1951). The PDB standard was depleted in the 1950s, so isotope laboratories use standards with known isotopic offsets from the PDB standard. Values for δ18O and δ13C are reported in parts per thousand (ppm) or per mil (‰) deviation from the standard.

Some of the terminology commonly used to describe stable isotope data includes such words as ‘light’, ‘heavy’, ‘more positive’, and ‘more negative’. The terms ‘light’ and ‘more negative’ indicate that there is relatively more of the light isotope (e.g., 18O or 13C) in the sample than in the standard and the δ18O or δ13C values are relatively low. The terms ‘heavy’ and ‘more positive’ mean there is relatively more of the heavy isotope (e.g., 18O or 13C) present in the sample than in the standard and the δ18O or δ13C values are relatively high. These terms will be used throughout the remainder of this paper.

### Oxygen Isotopes and Ice Volume

Oxygen has three stable isotopes, 16O, 17O, and 18O, of which 16O and 18O are by far the most common. The relative proportion of any one of these isotopes in a given oxygen-bearing compound is dependent on a number of factors that influence their partitioning or fractionation. We are concerned mostly with two types of isotopic fractionation: (1) fractionation that occurs when water changes state and (2) fractionation between oxygen isotopes in CaCO3 and the water from which it precipitates.

An important oxygen isotopic fractionation occurs between the liquid and vapor phase of water during evaporation. The process of evaporation discriminates against the heavy 18O isotope leaving the liquid phase relatively enriched in 18O (heavier or more positive δ18O) and producing water vapor that is relatively depleted in 18O (lighter or more negative δ18O). Prolonged evaporation of an unreplenished reservoir of water will cause the reservoir to become progressively enriched in 18O. This is the case for sea water during the growth of large ice sheets (Fig. 1). Thus, glacial ice is depleted in the heavy isotope 18O and
the marine ocean reservoir is enriched in $^{18}$O. When ice caps melt, isotopically light water returns to the ocean (Fig. 1). Given the approximately 1 ky mixing time of the ocean (Gordon, 1975), the $\delta^{18}$O value of sea-water changes is homogenized. Therefore, when large ice sheets grow, sea-water $\delta^{18}$O becomes heavier, and when ice sheets recede, sea-water $\delta^{18}$O becomes lighter.

Carbonate precipitated organically or inorganically in the ocean records the sea-water isotopic composition. If the ocean is enriched in $^{18}$O because of the growth of large ice sheets (e.g., the Last Glacial Maximum), the $\delta^{18}$O of calcite is heavy. If the ocean is depleted in $^{18}$O because of the melting of major ice sheets (e.g., the Holocene), the $\delta^{18}$O of calcite is light. By observing downcore variations in the $\delta^{18}$O of diagenetically un-

altered calcite, variations in ice volume can be inferred. Caution must be taken in making this interpretation because the $\delta^{18}$O composition of calcite is also dependent on water temperature and salinity.

The first Quaternary $\delta^{18}$O record measured on planktic foraminifera was interpreted to indicate variations in sea-surface temperature with a minor contribution from ice-volume change (Emiliani, 1955). Shackleton (1967) determined that most of the isotopic signal in late Quaternary isotope records is from ice-volume changes and that the contribution from temperature and salinity was secondary. Averaged globally, ice-volume change accounts for approximately two-thirds of the $\delta^{18}$O decrease from the Last Glacial Maximum to present. Ideally, the best record of ice-volume variations is derived from calcite precipitated in a marine environment where temperature and salinity are expected to have remained relatively stable, such as abyssal and bathyal environments below the thermocline.

**Carbon Isotopes**

Heavy isotope $^{13}$C is discriminated against during photosynthesis. Therefore, all organic matter is enriched in the light isotope $^{12}$C. The discrimination of $^{13}$C translates into a $\delta^{13}$C in plants that is approximately 20 ppm lower than the $\delta^{13}$C of carbon in CO$_2$ (Craig, 1953). Just as the growth or decay of continental ice alters the sea-water $^{18}$O/$^{16}$O ratio, any growth or depletion of the organic carbon reservoir will alter the $^{13}$C/$^{12}$C ratio in sea water.

During interglacial highstands, a large amount of organic carbon (light $\delta^{13}$C) is stored in the terrestrial biomass (mostly trees), in soils and in sediments deposited on the flooded shelves. Carbonate sediments deposited in the bathyal environment during sea-level highstands record the $\delta^{13}$C of the sea water from which they precipitate. During glacial lowstands, the shelf sediments are exposed and eroded. Most of the isotopically light carbon from eroded soils and shelf sediments is transferred to the ocean carbon reservoir (Broecker and Peng, 1982). Large negative $\delta^{13}$C events measured in sediments deposited at bathyal water depths could represent the sudden transfer of light carbon from the continental shelf to the bathyal environment during intervals of falling sea level and shelf exposure. However, it is important to note that the deep-sea isotope record often shows a positive shift of the $\delta^{13}$C during periods of major ice growth, concomitant with a positive shift of the $\delta^{18}$O. Ventilation of the bottom water masses during glacial periods can provide an explanation for the covariance of the carbon and oxygen isotopes in the deep ocean.

**CRITERIA FOR WELL SELECTION**

**DSDP/ODP Sites**

Most DSDP and ODP holes have been drilled in lower bathyal and abyssal water depths, the best locations to recover a complete and undisturbed sedimentary record. DSDP/ODP sites generally contain excellent bio- and magnetostratigraphies. However, deep-sea cores do not always yield optimum sedimentary records. They are usually characterized by low sedimentation rates which decrease their stratigraphic resolution. Furthermore, many DSDP/ODP sites were drilled in water depths below the calcite lysocline. Calcite dissolution can se-
lection remove carbonate microfossil tests resulting in loss of biostratigraphic data, a decrease in the accuracy of stratigraphic correlations and alteration of stable isotopic values (e.g., Wu and Berger, 1991). Magnetostratigraphy is the most powerful conventional tool for stratigraphic correlations. However, the magnetostratigraphic record of DSDP/ODP holes often contains zones of uncertain polarity that decrease their stratigraphic resolution. Moreover, bottom currents can create hiatuses in the sedimentary column due to the increase of bottom-current velocity during glacial periods (Wright and Miller, 1992). These erosional episodes in the deep ocean often coincide with erosional periods in neritic settings caused by glacioeustatic falls (Wright and Miller, 1992). Therefore, although the deep-sea record represents the best possible stratigraphic sections, correlations between DSDP/ODP sites are not always straightforward.

Data from DSDP/ODP sites 522, 529, 563, 608 and 747 (Fig. 2 and Table 2) were generated and compiled by Miller et al. (1988), Miller et al. (1991), Wright and Miller (1992) and Wright et al. (1992), to develop a composite Oligocene to Miocene oxygen isotope record and to define Oligocene and Miocene isotope zones. Although these are the best available deep-sea records, they include some with the stratigraphic problems discussed above such as carbonate dissolution intervals, non-polarity zones, and hiatuses. The best available Oligocene to Miocene deep-sea records (Miller et al., 1991) are compared to our data from Petrobras Well A, to the sequences of Vakarcs et al. (this volume), and to the cycle chart of Hardenbol et al. (this volume).

**Petrobras Well A**

Well A was drilled by Petrobras on the upper slope of the Campos Basin, located in the southeastern portion of the Brazilian continental shelf (Fig. 2). Detailed biostratigraphic analysis indicates that this well contains a fairly continuous lower Oligocene to middle Miocene section of interbedded marls and shales. Age control is based on calcareous nannofossil last appearance data (LAD’s, Fig. 3), which were correlated with the international biostratigraphic framework (Blow, 1969; Martini, 1971). Paleowater depth estimates suggest upper bathyal depths during the Oligocene and Miocene times.

We suggest that the bathyal environment can be the best location to study the relationship between seismic sequence- and δ¹⁸O-stratigraphies. From the perspective of seismic sequence stratigraphy, a sequence boundary is defined as an onlap surface with truncation of the underlying reflectors. In the inner and middle shelf environment, using conventional seismic data, one can normally observe only parallel reflectors making it very difficult to identify sequence boundaries (Fig. 4). Sequence boundaries are more easily recognized in the outer shelf/upper slope environment where reflector terminations are more common (Fig. 4). Furthermore, neritic sections often lack good biostratigraphic or magnetostratigraphic control (Miller et al., 1991), while the upper bathyal environment has more pelagic input for better biostratigraphic control. Moreover, the neritic setting tends to be more sandy, whereas shale and marls, lithologies more favorable for stratigraphic correlation, are more common in bathyal depths.

Conceptually, the stratigraphic record in neritic settings shows unconformities, encompassing long time periods, which tend to become conformable in bathyal settings. However, unconformities generated by oceanic currents can affect the bathyal settings in different continental margins, such as the Florida escarpment of the Gulf coast (Mullins et al., 1987) and on the New Jersey margin along the east coast of North America (Mountain and Tucholke, 1985). Unlike the unconformities on the shelf, slope unconformities formed by the action of oceanic currents are not widespread along all continental margins. For example, many segments of the Brazilian margin show no significant evidence of erosion by marine currents. Advantages of the bathyal environment over the abyssal environment are that it is generally located above the calcite lysocline (good micro- and nannofossil preservation) and the bathyal setting has a greater sedimentation rate than the abyssal setting, thus providing a better represented sedimentary section. However, slope settings are often characterized by sedimentary reworking related to gravity processes, such as slumps, debris flow and gravity flow, that can limit the use of slope sediments in testing the cause of sea-level changes.

Cutting samples, the most common sample type available in industry wells, also have limitation for their usage in stratigraphic correlations. Cavings and imprecision in the sample depth are usual problems in using cutting samples. However, the consistent good correlation between biostratigraphic data and seismic reflections observed in more than 200 industry wells in the Campos Basin indicates the reliability of this sampling procedure.

**ANALYTICAL METHODS: PETROBRAS WELL A**

The carbon and oxygen isotopic analyses of Petrobras Well A were made on carbonate fragments picked from cuttings at a sample interval of 3 to 9 m (Azevedo et al., 1993). A total of 305 samples was powdered and then reacted with 100% H₂PO₄. Oxygen and carbon isotopic ratios were determined for CO₂ liberated during this reaction using a Finnigan MAT 252 mass spectrometer housed in the Petrobras Research Center (CENPES). Isotope values are expressed relative to the PDB standard.

Whole rock analysis is uncommon in paleoclimate studies which in most cases involve isotopic measurements on single species of foraminifera. Assumptions prevail that mixing carbonates from different sources causes the δ¹⁸O signal to have little stratigraphic value. Nevertheless, in certain depositional environments, the whole rock δ¹⁸O record resembles closely the record derived from careful analyses of single foraminiferal species (Shackleton et al., 1993). Whole rock analyses are far less time consuming than single foraminiferal analyses and have the added advantage of not requiring the expertise of a micropaleontologist.

**DETAILED STRATIGRAPHIC CORRELATION: STRATEGIES AND METHODS**

Identification of Isotopic Events

The correlation between δ¹⁸O events and sequence boundaries is not direct. To a first approximation, the “ice house” δ¹⁸O curve can be considered to be similar to the eustatic curve, as discussed previously. The inflection points of δ¹⁸O curves, not the isotopic peaks, should correspond to sequence boundaries and maximum flooding surfaces. However, for practical pur-
poses, isotopic peaks have been used in correlation with the sequence boundaries shown in the Haq et al. (1988) chart (Miller et al., 1991; Abreu and Savini, 1994). In most cases, the inflection point of an isotope curve is difficult to determine, whereas the peak is normally sharp and easy to recognize. This makes the peak more adequate for stratigraphic correlation. Correlation based on the peak isotopic values builds-in a slight lag between the time of sequence boundary development, corresponding to the maximum rate of eustatic fall and the time of peak glaciation corresponding to the lowest eustatic position. However, the age difference between the isotopic inflection point (true sequence boundary) and the isotopic peak is normally less than the resolution of the stratigraphic age control given by chronozones and biozones.

The strategy used to define isotope zone boundaries in Miller et al. (1991) and Wright and Miller (1992) is based on the simultaneous increase in the oxygen isotopic values of both benthic foraminifera and western equatorial (non-upwelling) planktonic foraminifera at DSDP/ODP sites located in abyssal paleowater depths. Matthews and Poore (1980) were the first to suggest that surface temperature in the western equatorial region should be relatively stable through time and, therefore, that the δ¹⁸O changes in planktonic foraminifera from this region would indicate ice-volume changes. Miller et al. (1991) and Wright and Miller (1992) defined Oligocene and Miocene isotope zone boundaries using a first-order correlation with magnetochronostratigraphy and, in some cases, biochronostratigraphy. All ages were based on the time scale of Berggren et al. (1985).

The proposed Petrobrás Well A isotope events correlate well with glacioeustatic events defined by Miller et al. (1991) and Wright and Miller (1992), and they show a good relationship with global sea-level lowering events (Haq et al., 1988). Moreover, new isotope events are proposed (using the time scale of Berggren et al., 1995): OR-2 (31 Ma), OC-3 (25.2 Ma), MB-3 (18.0 Ma) and MS-1 (14.6 Ma), comparing favorably with other studies (Pekar and Miller, 1996; Miller et al., 1996). First-order correlation of the isotopic events described in this study was based on biostratigraphy (Fig. 3) and the second-order correlation on magnetostratigraphy (time scale of Haq et al., 1988).

**Chronostratigraphic Position of Isotopic Events**

We consider a time scale as a stratigraphic framework model in which the numerical age works as a reference level. More important than the absolute age itself is the relative position of a determined event in relation to physical data, such as bio- or magnetozones. To avoid miscorrelation due to the utilization of different time scales, all events described in this work are referred to the Berggren et al. (1995) time scale. The age conversion between the different time scales (Berggren et al., 1985; Haq et al., 1988; Berggren et al., 1995) was done through magnetostratigraphy. The isotopic record of sites 522 and 529 (South Atlantic, Miller et al., 1991), 563 and 608 (North Atlantic, Miller et al., 1991) and 747 (Indian Ocean, Wright and Miller, 1992) were used to establish the chronostratigraphic position of each event. For most of these events, we suggest a chronostratigraphic position slightly different from the ones proposed by Miller et al. (1991) and Wright and Miller (1992). The new positions were determined by correlation of the same event through different wells and using the chronzone determination in the well with the best sedimentary record at that level (Fig. 5). We also suggest new isotopic events based on DSDP/ODP records. Figure 6 shows the correlation between DSDP/ODP and Well A isotope records.
The summary of isotope events here described (Tab. 3) and their correlation with sequence boundaries defined by Haq et al. (1988), Vakarcs et al. (this volume) and Hardenbol et al. (this volume) is presented in Table 4. The ages are based on the Berggren et al. (1995) time scale.

Magnetochronology is based on the characteristic binary sequence of normal and reversed polarity of the geomagnetic field (Berggren et al., 1985). To describe the chronostratigraphic position of an event, we refer to the specific position of that event related to the normal or reverse polarity band within a certain chron. For example, the position C10.2n refers to the younger boundary of the second normal polarity band within Chron C10.

Perhaps one of the most important contributions of the sequence stratigraphic approach developed by P. Vail and his colleagues at Exxon was to highlight the importance of the series and stages in stratigraphic correlations, instead of formation names that by definition have only regional significance. Following this precedent, we name each one of the stable isotope events according to the stage in which the event occurs and its relative position within the stage. For example, the ORi-1 event represents the oldest isotopic event described within the Rupelian stage.

**Oligocene Time**

Isotope event ORi-1 corresponds to the Priabonian/Rupelian boundary (Eocene/Oligocene boundary). Miller et al. (1991)
defined isotope zone Oi-1 in DSDP Site 522 as the first isotopic event for the Oligocene near the base of magnetozone C13n, and Abreu and Savini (1994) defined an isotopic event at 33.8 Ma in Well A. We correlate this isotopic event with the 36 Ma (TA4.4) sequence boundary of Haq et al. (1988) at the Eocene/Oligocene boundary (Miller et al., 1991; Abreu and Savini, 1994). We further correlate this early Rupelian isotope event with sequence boundary Ru-1 (Vakarcs et al., this volume; and Hardenbol et al., this volume).

ORi-2 occurs in the middle part of the Rupelian Stage (lower Oligocene). Well A data suggests an isotopic event at 31.9 Ma, corresponding to an event in Site 522 at the middle part of the chronozone C12r. This isotopic event correlates with the 33 Ma (TA4.5) sequence boundary of Haq et al. (1988) and also with sequence boundary Ru-2 (Vakarcs et al., this volume; Hardenbol et al., this volume).

ORi-3 occurs in the upper part of the Rupelian Stage. Well A isotope data indicates an event occurring at 29.4 Ma, which corresponds to a positive event in Site 522 at the top of chronozone C11r (isotope zone Oi2 of Miller et al., 1991). We correlate this event to sequence boundary Ru-3 (Vakarcs et al., this volume; Hardenbol et al., this volume).

OCi-1 marks the boundary between the Rupelian and Chattian stages. The DSDP isotopic data suggests an isotopic event at the top of chronozone C10n.1n at 28.4 Ma (Site 529), and Abreu and Savini (1994) identified the heaviest $\delta^{18}$O event at 28.6 Ma in Well A. This event correlates with the 30 Ma (TB1.1) sequence boundary of the Haq et al. (1988) chart (Abreu and Savini, 1994). Note that the 30 Ma event (Haq et al., 1988) has a new age of 28.5 Ma, if converted to the Berggren et al. (1995) time scale (Table 4). We correlated this event to the sequence boundary Ch-1 (Vakarcs et al., this volume; Hardenbol et al., this volume).

OCi-2 occurs in the lower Chattian Stage (upper Oligocene). Miller et al. (1991) defined an isotopic event at the top of magnetozone C9n (isotope zone Oi-2b), and Abreu and Savini (1994) defined an isotopic event at 27.3 Ma in Well A. This isotopic event appears to correlate with the 28.4 Ma (TB1.2) sequence boundary of the Haq et al. (1988) chart (Miller et al., 1991; Abreu and Savini, 1994) and also with sequence boundary Ch-2 (Vakarcs et al., this volume; Hardenbol et al., this volume).

OCi-3 occurs in the middle part of the Chattian Stage. Isotope data from Well A indicates an event occurring at 25.2 Ma, which corresponds to a positive event in Site 529 at the top of chronozone C7r (25.2). We correlate this isotopic event with the 26.5 Ma (TB1.3) sequence boundary of the Haq et al. (1988) chart and also with sequence boundary Ch-3 (Vakarcs et al., this volume; Hardenbol et al., this volume).

Miocene Time

MAi-1 corresponds to the boundary between the Chattian and Aquitanian stages (Oligocene/Miocene boundary). Miller
Fig. 6.—Isotope record of the studied well sites. Correlation and chronostratigraphic positions, based on the Berggren et al. (1995) time scale, of the isotope events defined by Miller et al. (1991) and the ones proposed in this study. Correlation of the isotope record with the cycle chart of Hardenbol et al. (this volume). The position and relative magnitude of the Maximum Flooding Surfaces (MFS) is proposed in this study based on the Oxygen isotope record.

Table 3.—Correlation between the heavy oxygen isotope events identified by Miller et al. (1991) and the ones proposed in this study.

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(*) Berggren et al. (1995)
et al. (1991) defined the first isotopic event for the Miocene near the top of magnetozone C6Cn (isotope zone Mi-1). Isotope data from Well A shows a $\delta^{18}O$ event at 23.8 Ma that correlates to an isotopic event in Site 529 in the lower portion of chron C6Cn.2n (23.8 Ma). We correlate this isotopic event with the 25.5 Ma (TB1.4) sequence boundary of the Haq et al. (1988) chart and also with sequence boundary Aq-1 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MAi-2 occurs in the middle part of the Aquitanian Stage (lower Miocene). Abreu and Savini (1994) defined an isotopic event observed in Well A at 22.2 Ma, corresponding to an event in Site 608 at the top of chronzone C6AAr.1r (22.2 Ma) and correlated this isotopic event with the 22 Ma (TB1.5) sequence boundary (Haq et al., 1988). This event correlates with sequence boundary Aq-2 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MAi-3 corresponds to the boundary between the Aquitanian and Burdigalian stages. Miller et al. (1991) defined an isotopic event in the magnetozone C6Ar (isotope zone Mi-1a). Abreu and Savini (1994) defined an isotopic event in Well A at 21.1 Ma, corresponding to an event in Site 608 in the top of chronzone C6An.2n (21.1 Ma). We correlate this event with the 21 Ma (TB2.1) sequence boundary in the Haq et al. (1988) chart and also with sequence boundary Bur-1 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MBi-1 occurs in the lower part of the Burdigalian Stage (lower Miocene). Site 747 (Indian Ocean) isotope data displays a heavy $\delta^{18}O$ event occurring in the middle of chronzone C6n at 19.7 Ma. This event was observed by Wright and Miller (1992) and designated the Mi-1aa event. However, the Mi-1aa event was not defined as a true isotopic zone boundary. The event was not observed previously by Miller et al. (1991) and may be missing from the Site 608 and 563 isotopic records due to erosion probably caused by deep-sea currents in the Atlantic. More recently, Miller and Sugarsman (1995) identified a sequence boundary in the New Jersey subsurface at $\sim$20 Ma that they believe may correlate to this isotope event. This event is also observed in the Well A isotope record at 19.4 Ma. We correlate this event with sequence boundary Bur-2 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MBi-2 occurs in the middle part of the Burdigalian Stage. Well A isotopic data indicates an event at 18.0 Ma, corresponding to an event in Site 608 in the middle part of chronzone C5Dr (18 Ma). We correlate this isotopic event with sequence boundary Bur-3 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MBi-3 occurs in the middle part of the Burdigalian Stage. Miller et al. (1991) defined an isotopic event at the chronzone C6Dr (isotope zone Mi-1b). Abreu and Savini (1994) defined an isotopic event at 17.2 Ma in Well A that correlates with a positive event in Site 747 in the chron C5Cr (17.2 Ma). This isotopic event seems to correlate with the 17.5 Ma (TB2.2) sequence boundary of the Haq et al. (1988) chart (Miller et al., 1991; Abreu and Savini, 1994) and also with sequence boundary Bur-4 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MLi-1 marks the boundary between the Burdigalian and Langhian stages (lower/middle Miocene boundary). Miller et al. (1991) defined an isotopic event at the top of chronzone C5Br (isotope zone Mi-2). Abreu and Savini (1994) defined an isotopic event at 16.4 Ma that correlates with an event in Site 608 in the chron C5Br (16.0 Ma). This isotopic event seems to correlate with the 16.5 Ma (TB2.3) sequence boundary observed in the Haq et al. (1988) chart (Miller et al., 1991; Abreu and Savini, 1994), and also correlates with sequence boundary Lan-1 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MSi-1 marks the boundary between the Langhian and Serravallian stages. The DSDP isotopic data suggests an isotopic event occurring near the base of chronzone C5Adn at 14.6 Ma (Site 747). This isotopic event correlates with the sequence boundary of 15.5 Ma (TB2.4) of the global chart (Haq et al., 1988) and also with sequence boundary Ser-1 (Vakarcs et al., this volume; Hardenbol et al., this volume).
MSl-2 occurs in the lower Serravallian Stage. Miller et al. (1991) defined an isotopic event in the middle of chronzone C5ABr (isotope zone Mi-3) which corresponds to an age of 13.6 Ma. This isotopic event correlates with the sequence boundary of 13.8 Ma (TB2.5) of the Haq et al. (1988) chart (Miller et al., 1991) and also with sequence boundary Ser-2 (Vakarcs et al., this volume; Hardenbol et al., this volume).

MSl-3 occurs in the middle part of the Serravallian Stage. Miller et al. (1991) defined an isotopic event near the base of chronzone C5Ar (isotope zone Mi-4). We suggest the chron C5Ar.2n (12.7 Ma) in Site 608 to define the chronostratigraphic chronozone C5Ar (isotope zone Mi-4). We suggest the chron C5Ar.2n (12.7 Ma) in Site 608 to define the chronostratigraphic chronozone C5Ar (isotope zone Mi-4).

Isotope Events and Maximum Flooding Surfaces

We present a tentative correlation between the negative δ18O events and maximum flooding surfaces in the early Oligocene to middle Miocene section, based on DSDP/ODP and Petrobrás Well A isotope records. The δ18O events were then compared to maximum flooding surface ages on the onlap curve of Haq et al. (1988). The results are summarized in Table 5. Magnetozyones were used to convert ages from the Haq et al. (1988) to the Berggren et al. (1995) time scales. Remember that a light δ18O peak represents a sea-level high in the eustatic curve, whereas a maximum flooding surface represents the inflection point during a sea-level rise. Although the correlation between negative δ18O events and maximum flooding surfaces is not direct, we present the correlation between these two records as a first approximation of the sea-level highs in the eustatic curve during the “ice house world”.

DISCUSSION

Global Correlation

The global nature of sequence boundaries is the subject of great criticism and skepticism in the scientific community (e.g., Miall, 1994). Most of the criticism is related to the stratigraphic correlations between different basins. The duration of third-order sequences (0.5 to 3 my) is at the limit of biostratigraphic resolution. Moreover, one can often find more than one sequence boundary within the duration of a biozone. The combination of more than one biostratigraphic method can increase the accuracy in such correlations, although the resolution in this ideal situation (500 ky to 1 my) is still not adequate to determine the synchronism of third-order sequences. Better resolution may be achieved through magnetostratigraphy. However, good magnetic records are restricted mostly to deep-water sites that often contain non-polarity zones that can decrease their stratigraphic resolution.

Due to these limitations in the resolution of the most common stratigraphic tools for age control in marine sedimentary records, the assignment of ages to the stratigraphic record is estimated by linear interpolation between fixed data that are normally more than 1 my apart. In performing this interpolation, one assumes an improbable constant sedimentation rate during a given time interval. These time intervals can encompass short-term hiatuses and facies changes reflecting sedimentologic processes with different depositional behaviors. The resolution of the best stratigraphic methods is not good enough to establish the synchronism of similar-aged events on a global scale. To demonstrate the global scope of events with similar age, it is necessary to postulate a global mechanism that can generate them. Vail et al. (1977) suggested glacioeustasy as the driving mechanism for third-order sequences. Cloetingh (1988) suggested that intraplate stress could generate unconformities within the resolution of third-order sequences, even in passive margin settings. Although the driving mechanism for third-order sequences in Phanerozoic time is not clear, the good correlation between the isotopic and seismic events in our study strongly suggests that glacioeustasy is the mechanism responsible for the generation of unconformities in continental margins during the “ice house world”, and that the sequences proposed by Haq et al. (1988), Vakarcs et al. (this volume) and Hardenbol et al. (this volume) are indeed global.

Frequency of Sea-Level Fluctuations

The number of sequence boundaries has increased in the work of Vakarcs et al. (this volume) and Hardenbol et al. (this volume) compared with the ones proposed by Haq et al. (1988). One major difference between these studies is that the Haq et al. (1988) study was based mostly on seismic data and the Vakarcs et al. (this volume) and Hardenbol et al. (this volume) studies were based mostly on outcrop and well-log interpretation. Sixteen heavy isotopic events have been identified from the early Oligocene to middle Miocene, representing 21 my and corresponding to an average of one event each ~1.3 my. The temporal resolution of the isotope data and the sequence stratigraphic record of the Pannonian Basin is about the same, although several smaller, poorly represented events recognized in the isotope record are not observed in the sequence stratigraphic record.

One can foresee an increase in the number of early Oligocene to middle Miocene isotope events related to ice growth when more complete sections with higher sedimentation rates are studied. Improved isotope records will likely emerge from on-
going studies of Ocean Drilling Program cores recovered recently from the eastern equatorial Pacific (Leg 138), from the New Jersey continental margin (Leg 150, Northwest Atlantic), from the Ceará Rise (Leg 154, Southwest Atlantic), from Bahamas (Leg 166, Caribbean) and from industry wells drilled in slope settings along the Brazilian continental margin.

There is poor agreement between the number of Antarctic glaciations identified and described from the Antarctic terrestrial and shelf sedimentary records (e.g., Bart and Anderson, 1995) and the number of potential glacial events indicated in the glacial record ranges from the poorly age constrained, low-resolution, direct sedimentary evidence to the recent compilation of more than one glacial period. Therefore, the continental record of glaciations in King George Island (e.g., Melville glaciation, Late Oligocene to Early Miocene in age, Abreu et al., 1992; Legru and Polonez glaciations, Oligocene in age, Birkenmajer, 1987) encompasses periods of time of a few million years. The Ciros-1 well drilled in the Ross Sea, one of the best sedimentary records for Oligocene Antarctic glaciations (Barrett et al., 1988), indicates glaciations of longer duration than the ones inferred from the oxygen isotope record. Each of the glacial sedimentary records may represent, in fact, an amalgamation of more than one glacial period. Therefore, the Oligocene to Miocene glacial record ranges from the poorly age constrained, low-resolution, direct sedimentary evidence around Antarctica to the well age constrained, high-resolution, indirect isotopic evidence in pelagic settings.

Amplitude of sea-level fluctuations

Several authors used the Pleistocene calibration of 0.11% δ18O variation per 10 m of sea-level change determined by Fairbanks and Matthews (1978) and Fairbanks (1989) to estimate the magnitude of Cenozoic sea-level variations (e.g., Miller et al., 1987; Williams, 1988; Miller et al., 1991; Haddad et al., 1993; Abreu and Savini, 1994). This is only a rough estimate because a significant component of the δ18O record is expected to have been caused by global cooling and decreasing ocean temperatures through Cenozoic time (Savin, 1977; Miller et al., 1987). Abreu and Savini (1994) compared the timing and magnitude of sea-level changes estimated by the δ18O record with the timing and magnitude of changes in coastal onlap estimated for global third-order events (Haq et al., 1988). In all cases, the inferred change in sea level, using the sequence boundary patterns, is larger than the magnitude predicted by the δ18O signal. An explanation for this difference is that depositional sequences are the result of relative sea-level change (eustasy plus subsidence) while the isotopic signal is related more to eustasy (ice volume).

Figure 7 shows a summary of all the isotopic events discussed and presented in this paper. The timing and estimated magnitude of positive and negative δ18O events interpreted from DSDP/ODP and Petrobras Well A isotope records are presented in a graphic form to simulate an early Oligocene to middle Miocene sea-level curve. The “sea-level” curve (Fig. 7) is compared with the global sea-level curve of Haq et al. (1988). We do not suggest that the absolute magnitude of the sea-level changes is valid, however, we believe that the relative changes are meaningful. In other words, small isotopic changes probably represent small sea-level changes and large isotopic changes indicate large sea-level changes (see Tables 5, 6).

In a first look, the curves display a similar overall trend of falling sea level towards the late Miocene time and higher magnitude of sea-level variations during the Oligocene time. The major difference between these curves is the higher number of oscillations and the higher sea level observed from the isotope record than in the late Oligocene global sea-level curve (Haq et al., 1988). The isotope record is also characterized by lower amplitude sea-level oscillations than the ones present in the global curve (Haq et al., 1988). Another important difference is the magnitude of the 28.5 Ma sea-level fall (formerly 30 Ma in Haq et al., 1988) inferred from ODP and DSDP sites versus the sea-level fall inferred in Well A and the Haq et al. (1988) chart. The ODP and DSDP sites show no pronounced heavy δ18O event at this time, whereas both the Well A and Haq et al. (1988) records suggest that the 28.5 Ma (30 Ma of the Haq et al., 1988 chart) sequence boundary corresponds to the largest sea-level fall of the entire Oligocene and lower Miocene time. The Well A δ18O record provides a possible independent verification of the large 28.5 Ma sea-level fall suggested by Vail et al. (1977) and Haq et al. (1988).

Figure 8 shows a more extensive comparison between the isotope record and the eustatic curve of Haq et al. (1988) for the last 34 my. The isotope events here described and the isotope zones proposed by Miller et al. (1991) were used for the Oligocene and Miocene sections. The Pliocene isotope record was smoothed to keep approximately the 100-ky cycle and longer cycles. We have used the filtered isotope curve of Hadad and Vail (1992) for the Pleistocene to the Recent sections which is a filtered version (low-pass, 1/66 ky — 1/45 ky) of the stacked benthic isotopic records from Sites 607 and 677 (Raymo et al., 1990). Sites 552 (Keigwin, 1987), 704 (Hoddell and Venz, 1992) and 502 (Oppo et al., 1995) were also used for the Plio-Pleistocene composite smoothed isotope record. Therefore, the composite isotope “sea-level” curve is a smoothed version of the above records (Fig. 8).

The composite sea-level curve based on sequence stratigraphy associates the curves from Haq et al. (1988) for the Oligocene and Miocene sections and the sea-level curve from Mitchum et al. (1994) for the Pliocene to the Recent sections. Although estimates of sea-level change made from seismic and isotope data have limitations, the correlation between these two different records is remarkable. The good agreement between the age and magnitude of sea-level fluctuations inferred from two independent sources of data lends strong support to the hypothesis of ice-volume change as the driving mechanism for eustatic variations during the “ice-house world”.

Relevance to Hydrocarbon-Exploration

Paleoceanographic analysis has been underutilized as an exploration tool. The application of the techniques described in this paper could help exploration in passive margin offshore basins, mainly in Tertiary sequences. Tertiary turbidite deposits have been considered important oil targets, mostly in deep water. Exploration efforts in deep-water Tertiary turbidites are expected to be intensified in the near future, as is already occurring in the Gulf of Mexico and on the Brazilian margin. Tertiary deep water turbidites have completely changed oil exploration in Brazil during the middle 80s. The significant increase in the Brazilian oil reserves from 1 bbl, in early 70s, to the present...
day 10 bbl was due to the discovery of giant sandy basin floor fan-related deposits correlated to the major sea-level falls in the Oligocene and lower Miocene times.

Detailed study of glacio-eustatic variations, based on the integrated analysis of isotope-, seismic- and biostratigraphies, is expected to yield a more accurate chronostratigraphy of erosive horizons and possible associated turbidite deposits of potential economic interest. This method has been successfully applied to oil exploration in Plio-Pleistocene sections of the Gulf of Mexico (e.g., Trainor and Williams, 1987) and has the potential to be applied in the Oligocene/Miocene marine sections (Azvedo et al., 1993; Abreu and Savini, 1994).

CONCLUSIONS

Careful, detailed study of DSDP and ODP $\delta^{18}O$ records and the $\delta^{18}O$ and $\delta^{13}C$ record from Petrobrás Well A has led us to propose four additional early Oligocene to middle Miocene isotope events that were not identified in the works of Miller et al. (1991) and Abreu and Savini (1994). These isotope events include (using the time scale of Berggren et al., 1995): ORi-2 (31 Ma), OCI-3 (25.2 Ma), MBi-3 (18.0 Ma) and MSi-1 (14.6 Ma). In addition, we suggest new chronostratigraphic positions for many of the heavy oxygen isotope zonal boundaries observed previously by Miller et al. (1991) and Wright and Miller (1992). These isotope events show good correlation with sequence boundaries identified by Vakarc et al. (this volume) and Hardenbol et al. (this volume). Furthermore, we present chronostratigraphic positions for minimum ice-volume events (maximum flooding surfaces) determined from DSDP/ODP and Well A $\delta^{18}O$ records. These light $\delta^{18}O$ events are in good agreement with maximum flooding surfaces observed in the Haq et al. (1988) chart. Based on the $\delta^{18}O$ record, we propose ages and
amplitudes for the early Oligocene to middle Miocene maximum flooding surfaces.

Future works aiming to demonstrate the relationship between sequence boundaries and isotopic events need to be done by developing both records from one basin along a passive margin setting. Continuous cored sections with good bio- and magnetostratigraphies, high resolution and/or high quality seismic lines, showing as complete as possible sedimentary sections, are prerequisites for such study. New reference sections have to be chosen in regions with the potential to fulfill these prerequisites, to develop a stratigraphic framework as complete as possible, and to better understand the cause of eustatic changes and erosions along passive margins.

Our results support a glacioeustatic cause for third-order sequences since the early Oligocene. Assuming that ice-volume change has driven sea-level change, the frequency and synchronism of early Oligocene to middle Miocene sequences described from widely separated passive margin settings is not
only reasonable but probable. This permits global sequence stratigraphic correlations for Oligocene and Miocene passive-margin sequences.

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GLACIOEUSTASY: THE MECHANISM LINKING ISOTOPE EVENT AND SEQUENCE STRATIGRAPHY


WRIGHT, J. D. AND MILLER, K. G., 1992, Miocene stable isotope stratigraphy, site 747, Kerguelen plateau: Proceedings of the Ocean Drilling Program Scientific Results, v. 120.


ABSTRACT: A sequence stratigraphic analysis of subsurface data from the Paleogene strata of the central North Sea has documented a stratigraphic framework of 18–20 “third-order” depositional sequences nested within 5 “second-order” major regression/transgression facies (RTF) cycles. Additional sequences have been documented through correlation of the subsurface deposits to outcrop sections of northwest Europe. This paper will document only those sequences observable in the subsurface data and characterize these sequences within a low and high order framework. The order of a cycle is based on observations concerning its constituents and its impact on the depositional systems of the basin, not strictly on its duration. Integration of the composite standard biostratigraphic method enabled the construction of a consistent chronostratigraphy based on the correlation of hiatal intervals identified with graphic correlation data terraces. An ideal relationship of graphic correlation terraces within a sequence stratigraphic model is diagrammed, providing the theoretical basis for the correlations presented. A depositional model is also proposed as a variant of the classic Vail model, considering the effect of depositional profile and sediment supply in the preservation and distribution of sequence stratigraphy and sequence stratigraphy provide an opportunity for comparison between different methods and data resolutions.

INTRODUCTION

The central North Sea Paleogene section is a good place to test sequence stratigraphic concepts and correlations since many years of hydrocarbon exploration has built an extensive seismic, well-log, and biostratigraphic database. Numerous companies exploring the region have constructed internal and proprietary stratigraphic frameworks that hinder correlations as time equivalent deposits can have many different lithostratigraphic names. The stratigraphic framework presented here is based on sequence stratigraphic interpretation of seismic data and well logs, calibrated with graphic correlation of all available biostratigraphic datums (not just biostratigraphic zones). We document 18–20 “third-order” depositional sequences within 5 “second-order” major regression/transgression packages and observe the impact of these nested low- and high-frequency relative changes in sea level on the central North Sea Paleogene depositional system. We can also correlate the subsurface deposits to a sequence stratigraphic framework for northwest Europe and resolve an even more complete relative sea level history for the region (Neal, 1996; Neal and Hardenbol, this volume).

PREVIOUS STUDIES

Cycles of relative change in sea level can occur at many different time scales, from supercontinent break-ups (Pitman, 1978; Sloss, 1963) to local inundations caused by tectonic activity (Gawthorpe et al., 1994). Depositional sequence stratigraphy (Vail, 1987; Van Wagoner et al., 1988, 1990; Vail et al., 1991) developed from seismic stratigraphy (Mitchum et al., 1977). As seismic data resolution improved and outcrop sequence stratigraphy reach a wider audience, a trend towards recognition of higher frequency signals is emerging. Papers by Mitchum and Van Wagoner (1991) and Posamentier et al. (1992a) highlight sequences at much higher frequency than suggested by the eustatic curve of Haq et al. (1988). In basin analysis, recognition of the different “orders” of stratigraphic cycles and their interaction complicates the identification of a unique basinwide correlation. Reservoir studies are conducted with dense data control and can be unconcerned with the regional consequences of recognizing the “global signal” for correlation. Basin analysis work must place observations at any one location into a stratigraphic framework for all other locations in the basin. Constructing a stratigraphic framework of nested stratigraphic cycles is one way to identify a more complete basinwide sea-level history.

The stratigraphy of the lower Paleogene section of the central North Sea has evolved with the development of higher quality seismic data and numerous well penetrations since the first basal studies of Parker (1975) and lithostratigraphic nomenclature of Deegan and Scull (1977). Rochow (1981) published the first seismic stratigraphic study of the North Sea “Paleogene” strata, relating seismic units to the lithostratigraphic division of Deegan and Scull (1977). Stewart (1987) followed with a 10-sequence stratigraphic framework of the region, incorporating seismic and well data, tied to a biostratigraphic framework. Stewart’s framework was related to tectonics by Milton et al. (1990), expanding the idea of nested high- and low-frequency sea-level change cycles.

Recently, a series of papers were published in a volume edited by Parker (1993) that cover a range of time scales and stratigraphic tools. Galloway et al. (1993) proposed a tectonostratigraphic framework for the entire Cenozoic North Sea Basin that displays the lowest resolution but most regional aspect of these studies. Timbrell (1993) focused on the Lower Eocene units only and the stratigraphic complexity of the Balder and Frigg sections. Morton et al. (1993) used heavy mineral analysis along with key marker biostratigraphy to subdivide the section, and Vining et al. (1993) emphasized palynofacies with respect to sea-level changes. Den Hartog Jager et al. (1993) and Armentrout et al. (1993) presented regional frameworks on a scale similar to our study, emphasizing the integration of seismic and well data. Other regional studies on this same geologic interval focus more on seismic geometries related to tectonics (Jones and Milton, 1994) and biostratigraphic markers related to well-log picks (Mudge and Bujak, 1994). The standard North Sea lithostratigraphy has also been recently revised, with more em-
phasis placed on chronostratigraphic correlation (Mudge and Copestate, 1992; Knox and Holloway, 1992). The most recently published North Sea sequence stratigraphic frameworks (Knox et al., 1996) will be discussed in the regional correlation section.

GEOLOGIC SETTING

The Paleogene central North Sea was part of an intracratonic seaway that covered the area occupied by the present-day North Sea and embayments up to 300 km inland of the current northern European coast (Ziegler, 1990). The basin was formed by a series of rifting events throughout the Mesozoic Era that were subsequently blanketed by a layer of pelagic chalks during Upper Cretaceous time. During Paleogene deposition, the mainland of Britain was uplifted due to a combination of rift activity tied to the spreading of the North Atlantic (Ziegler, 1990) and the development of a ‘hot spot’ responsible for the Thulean volcanics found in the Scottish Highlands and Isle of Skye (Knox and Morton, 1988; Lewis et al., 1992; Knox, 1996). This uplift initiated deposition of clastic turbidites to the west of the Scottish Highlands into the Faeroe-Shetland Basin (Mitchell et al., 1993) and reworked chalk turbidites to the east of the High-lands into the North Sea Basin (Johnson, 1987). Following the Lower Paleocene chalk turbidites, siliciclastic submarine fans, gravity flow aprons, and extensive deltaic wedges were deposited into the central North Sea basin during the remainder of Paleogene time. The change from chalk to siliciclastic deposition was due to the removal by erosion of chalks capping the Devonian Old Red Sandstone (Ziegler, 1990). This sandstone provenance provided a coarse-grained source area for the submarine fan reservoirs of numerous Paleogene fields. The amount of uplift based on mass balance calculations from the entire Paleogene fan section reported by Den Hartog Jager et al. (1993) is estimated to be less than 1 km. Nadin and Kusznir (1996) estimate 255–375 m of Paleocene uplift in the Outer Moray Firth area.

Paleogeographic Setting

The estimated amount of uplift in the provenance area has important implications in the distribution of systems tracts (Brown and Fisher, 1977; Van Wagoner et al., 1988) within Paleogene depositional sequences. When interpreting systems tract distribution, one of the most important controls is the depositional profile or margin morphology. On the southern coast of the North Sea, very low-angle dips of an intracratonic seaway ramp produce a complicated depositional history due to wavebase erosion during relative changes in sea level (e.g., Plint, 1988). In the central North Sea, a shelf-slope break developed, providing the necessary gradient (<0.5°; Stow, 1986) to trigger turbidity currents that funnel coarse clastics into the basin during times of relative low sea level (Mitchum, 1985; Posamentier and Vail, 1988; Posamentier et al., 1991; Mutti, 1992).

On seismic profiles, it is usually easy to distinguish a single prograding deltaic unit that distally downlaps the turbidite package within each depositional sequence. This prograding package onlaps landward as it builds out into the basin, creating an extensive coastal plain. The relatively small amount of vertical uplift produced a gentle shelf gradient, creating the potential for rapid increase in shelfal accommodation for sediment, once the prograding unit ceases to keep up with rising relative sea level. The effect of this depositional profile on sediment delivery to the basin is critical when considering a sequence stratigraphic depositional model as discussed below.

MATERIAL AND METHODOLOGY

A 7400-km seismic grid (Fig. 1) was interpreted with approximately 100 well logs tied by synthetic seismograms supplement seismic facies analysis with geologic information. Of those wells, 40 have detailed biostratigraphic reports and are part of the 150 wells with detailed biostratigraphy that were compiled to build a North Sea composite standard from Amoco’s global composite standard (Stein et al., 1995). Interpretation of the biostratigraphy for individual well was accomplished by graphic correlation (Shaw, 1964; Miller, 1977), plotting data from a particular well against the composite standard of fossils for the entire basin. The construction of a composite standard and its precise application is complex. The methodology and the way it was used to document a sequence stratigraphic framework throughout NW Europe is summarized here and explained in more detail elsewhere (Neal et al., 1994, 1995; Neal, 1996).
Application of Graphic Correlation

The composite standard used in this study (Stein et al., 1995) developed from an iterative process of integration and emendation of all biostratigraphic datums encountered (Fig. 2). Biostratigraphic datums from each well are graphed against the composite standard (minus the well itself to avoid a circular correlation), which represents the most complete ideal fossil succession. This process leads to a scatter of data points in a time (composite standard units) versus depth (from the well being graphed) plot. A biostratigrapher constructs a line of correlation (LOC) through the points. The LOC is interpretive, giving the biostratigrapher freedom to weight key index fossils or lessen the influence of fossils that are clearly out of the normal succession (Shaw, 1964). Sloping LOCs indicate deposition through time, whereas LOC terraces indicate nondenposition, erosion, faulting or any other type of break in rock accumulation resolvable with the biostratigraphy at the well location.

Sequence stratigraphy has a predictable impact on a graphic correlation LOC (Neal et al., 1995). A foundation of overlapping LOC terraces forms the backbone of our stratigraphic framework, however, all depositional sequences identified from well logs and seismic data are not presently resolvable with graphic correlation. Biostratigraphic resolution, sediment supply and the effect of high and low frequency sea level changes are factors that control the effectiveness and resolution of this model (Neal et al., 1995). We have supplemented the graphic correlation framework with certain key fossil markers that help tie to other frameworks.

Correlation Specifics

Depositional sequences observed from central North Sea subsurface data have been given lithostratigraphic names from the lithostratigraphic frameworks of Deegan and Scull (1977), Isaksen and Tonstad (1989), Mudge and Copestake (1992), and the Paleo-Services/Chevron framework of the Mid-Upper Eocene (Newton and Flanagan, 1993). Our central North Sea sequences are distinguishable with biostratigraphy and from physical surfaces seen in well logs and in seismic data. They do not represent all the sequence cycles recorded in the North Sea basin. More sequences have been observed in outcrop sections from northwest European basins, but are not recognized within the resolution of the subsurface data base.

Figure 3 illustrates the correlation of depositional sequences used in the Paleocene to Lower Eocene section to lithostratigraphic units from type wells of Mudge and Copestake’s (1992) and Isaksen and Tonstad’s (1987) frameworks. Also displayed here is a correlation to depositional sequences published by Stewart (1987), using his well sections and maps as tie points. The diachronality of Stewart’s (1987) sequences illustrates the need for a framework founded on interpretation of biostratigraphic and well log data, that is then tied to seismic data.

The depositional sequences of this study do not always correlate with their lithostratigraphic equivalents. Sequences are identified by their biostratigraphy and position on a graphic correlation LOC, related to surfaces on seismic data and well logs. As such, they are chronostratigraphic units and the same sequence may contain more than one lithostratigraphic unit (see discussion by Bertram and Milton, 1988). Correlation of lithostratigraphy with sequence stratigraphy must be done with extreme caution. Lithostratigraphy is based on objective criteria with strict rules of application (e.g. Whittaker et al., 1991).

Describing a type log signature for any sequence is problematic as illustrated by the “Upper Sele” sequence (Fig. 3). This sequence typically has acme occurrences of the dinocysts, Cerodinium wardenense at its base and Deflandrea oebisfeldensis at its top. This sequence is often also associated with a downhole influx of large leiospheres, however, an acme occurrence of Pterospermella, grouped with the leiospheres influx by Knox and Holloway (1992), is found in an older sequence. In type wells for each lithostratigraphic framework, the Upper Sele se-
Fig. 3.—Correlation of depositional sequences, key biostratigraphic markers, and hiatal intervals with namesake lithostratigraphic units from recently published frameworks. Abbreviations for lithostratigraphic units: E = Ekofisk Fm., M = Maureen Fm., A = Andrew mbr., G = Glamis mbr., Bal = Balmoral mbr., L = Lista Fm., S = Sele Fm., F = Forties mbr., D = Dornoch mbr., Be = Beauty Fm., B = Balder Fm., Fg = Frigg Fm., Hr = Hermod Fm., Hm = Heimdal Fm. Also shown are Stewart’s (1987) sequences: S = type section for the sequence.
quence markers may be found together or separated, in the Balder Formation (tuffaceous claystone), Sele Formation (dark grey clay), Hermod Formation (deep water sandstone), or in their coaly shelfal equivalent, the Beauty Formation.

A 30-m “error bar” between the reported fossil depth occurrence and the well log horizon was used by Armentrout et al. (1993), to account for operator consistency and seismic data resolution. This margin of error facilitates the correlation a sequence boundary between well log, seismic and biostratigraphic data. Generally, we tried to achieve precise agreement between at least two of the three disciplines used in this study (log interpretation, paleontology, and seismic interpretation). The error bar mentioned by Armentrout et al. (1993) is one way to reconcile minor discrepancies in reported fossils with well log and seismic interpretation. When sequences are below seismic data resolution or occur within a poor seismic data zone, well log interpretation and biostratigraphy are the only tools that allow correlation with any confidence.

**Well Examples of Graphic Correlation Using the Composite Standard**

The composite standard for the central North Sea has been largely built and utilized in the deep-marine environment where correlation of turbidite packages relies on good biostratigraphy. A sequence stratigraphic interpretation involves a series of steps that begins with interpretation of the biostratigraphic data, followed by picking sequences on the well logs that are tied to seismic data with synthetic seismograms, and finally correlated on seismic data between well control points. Construction of a composite standard begins with a search through the data set for sections that appear to have the most complete biostratigraphic succession. The best section serves as a reference section for the basin. In this study, the Norwegian well, N16/1-1 (Fig. 4) was chosen based on the fact that it contained a nearly complete succession of biostratigraphic zones used in the well report by Robertson Research International (1987). This data is shown in Appendix Table 1 of this paper and can be used to reconstruct and reproduce the composite standard used in this study. When biostratigraphic data from well N16/1-1 was first graphed as the standard reference section, the graph had a single LOC terrace at 7260 ft. Following iterative rounds of graphic correlation, this well is now shown (Fig. 4) to contain 9 to possibly 11 LOC terraces (Stein et al., 1995).

The sequence stratigraphic interpretation of this well follows from the model described in Neal et al. (1995). The Robertson Research well report (1987) estimates paleobathymetries in this well were never shallower than upper bathyal, and the LOC terraces represent sediment starvation and possibly submarine erosion. The solid LOC (Fig. 4) represents a graphic correlation based on biostratigraphic markers that regularly occur in proper succession. These markers include the first downhole occurrence (FDO) of *Globorotalia pseudobulloides*, *Globorotalia compressa*, and the acme occurrence of *Areoligera cf. senonensis* sensu RRI. (now *gippingensis*; Jolley, 1992). The stippled LOC highlights some possible departures from the solid line where sequence stratigraphic analysis of the well log suggests alternative terraces. The base of the Andrew sequence is picked at the base of a limestone, interpreted to be a debris flow, at 8680-ft well depth. A FDO of *Alisocysta reticulata* occurs near this contact. Although the well report indicates the occurrence to be reworked, it appears to fall in proper succession (on the LOC). A series of markers starting with *Isabelidinium? viborgense* at 8550 ft seem to define a line segment up to the acme occurrence of *Palaeoperidinium pyrophorum* at 8300 ft. In order for this segment to represent the true LOC, a cluster of data points at 8400 ft containing *G. pseudobulloides* and *G. compressa* must be interpreted as reworked and the well report does question whether these markers are indeed in situ. Unfortunately, this well is not tied to seismic data in our study area, so we cannot rule out either interpretation at this time.

Another departure from the solid LOC is observed at 7860 ft, where the base of the Upper Balmoral sequence is picked at the base of a 200-ft-thick sand. The regional hiatus between the Upper and Lower Balmoral sequences is usually associated with the acme of *A. cf. senonensis* (*gippingensis*); however, that marker is picked in a cored interval (8000 ft) within the Lower Balmoral sand and the solid LOC reflects that control point. The stippled LOC is influenced by interpretation of the well log and honors a FDO of *Pseudobolivina* sp. 1 RRI that normally occurs within the Lower Balmoral sequence. Above this fossil, the stippled LOC terraces over to the solid LOC, which is controlled by markers such as *Alisocysta margarita* and the reappearance downhole of agglutinated foraminifera like *Spirelectamminia spectabilis* and *Rhizammina/Bathyssiphon* in Paleocene strata. These departures illustrate the possibilities of a checks and balances system between the sequence stratigraphic and graphic correlation interpretations.

**Shelfal Interpretation**

A different depositional setting is considered in the next example as shelfal environments are encountered. Figure 5 shows the gamma-ray curve, lithology, graphic correlation and sequence stratigraphic interpretation of a well through the Upper Paleocene shelf. Sequence and systems tract picks come from well-log interpretation of depositional environment (Vail, 1987; Vail and Wornardt, 1990) and basinward shifts in facies. Well-log patterns alone are not diagnostic though, as the sand at a depth of 3500 ft has a log pattern similar to other sands below 4600 ft. Critical to the interpretation of this section is the paleobathymetry column estimating the depositional environments of these sediments using benthic forams and other indicators. Paleobathymetry estimates places the upper sand in the transitional marine to inner shelf environment while the lower sands were deposited in an upper bathyal water depth. The lower sands represent fan deposition during falling sea level and the upper sand is interpreted as incised valley fill during rising sea level. The paleobathymetry alone can track low-frequency sea-level cycles with peak floodings around 4500 ft and 2700 ft.

The sequence stratigraphic interpretation of this well identifies many more depositional sequences and systems tracts than are resolvable with graphic correlation. Individual fossil markers are key to the correlation, as is the facies interpretation of the well log. Seismic data interpretation can strengthen correlation of shelfal sequences especially when well-log and biostratigraphic data are inconclusive. The correlation of Middle and Upper Sele sequences benefits from seismic data interpretation between wells 15/6-1 and 15/13-2 (Fig. 6). Well 15/13-
FIG. 4.—Graphic correlation and depositional sequences of basinal well N16/1-1. Paleobathymetric estimates are reported as upper-to-lower bathyal. Biostratigraphic data is courtesy of Simon Petroleum Technologies (Robertson Research) with numbered tops listed in Appendix Table 1.

Key
+ - - = biomarker top
--- = conservative LOC honoring the most possible key points
--- = A more interpretive LOC recognizing additional offsets with the aid of sequence stratigraphic analysis

---sand, ---silt, ---tuff, ---limestone.
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<th>taxa #</th>
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Fig. 5.—Sequence stratigraphic interpretation of UK 15/6-1, a well with shallow water section. Systems tract interpretations are based on log character and paleobathymetry: LLST = lower lowstand (bypass sediments), ULST = upper lowstand (lowstand prograding and slumps), TST = transgressive systems tract, IVF = incised valley fill, HST = highstand systems tract. Paleobathymetry estimates and well data courtesy of S. P. T. Ltd. (Robertson Research).
2 has also been interpreted for depositional sequences, and both wells were tied to seismic data with synthetic seismograms. The Sele sequence boundaries are identified at physical surfaces in the wells and at reflector discontinuities in seismic data (Fig. 6). The Sele sequences fall between two overlapping graphic correlation terraces, which means we must rely on other tools to further divide the section. In the 15/6-1 well, both sequence boundaries are picked at the base of transgressive sands, indicated by upward increasing gamma-ray values. In well 15/13-2, a slope sand rests on the Middle Sele sequence boundary. This sand is interpreted as a lowstand deposit. Transgressive and highstand deposits that may occur further updip are represented in this well by condensed section mudstones (1360–1380 m; Fig. 6). A thin carbonate layer marks the Upper Sele sequence boundary in 15/13-2 (1360 m) with the overlying silty section representing a lowstand deposit. The Upper Sele sequence has a highstand progradational wedge (see Fig. 5 at ~2900–3100 ft), seen on seismic data, that thins rapidly toward the shelf break (Fig. 6) and is thin or absent down-dip. Graphic correlation brackets these sequences within a similar time frame, but well-log and seismic data interpretation better resolves the stratigraphy in this location.

The seismic line in Figure 6 provides evidence for the interpretation of these prograding units as lowstand deposits. The base of the incisive canyon at the left of the line (Fig. 6) occurs at the same depth as the first toplap in the eastern prograding package. Toplaps in deltaic units are considered to represent baselevel in true dip sections (Mitchum et al., 1977, Jones and Milton, 1994), and by calculating from seismic conversion the depth at the top of a prograding package to the depth of the first toplap in a subsequent progradation, a minimum estimate of relative sea-level fall is obtained. The physical relationship of deposits in this example (Fig. 6) suggests that relative sea level fell at least 150 m between the Upper Balmoral and Upper Forties Sequences.

Central North Sea Paleogene Depositional Model

A notable feature of the depositional sequences seen in the previous examples is the lack of significant transgressive and highstand systems tract deposition. This situation is consistent with published models (Vail, 1987; Posamentier et al., 1991) for the bathyal environment, seen in Figure 4. However, the shelfal deposits of the UK15/6-1 well (Fig. 5) show only minor transgressive and highstand systems tracts, which rapidly thin basinward to below seismic resolution (Fig. 6). Explaining this situation requires a variation of the Vail (1987) sequence stratigraphic depositional model similar to the one proposed by Armentrout et al. (1993), where transgressive and highstand deposits are landward of the lowstand shelf break.

Depositional sequences in the subsurface study area consist mainly of a shelfal bypass package and a prograding deltaic package. The delta downlaps directly onto the shelfal bypass sediments at the base of the slope. In deeper parts of the basin, bypass packages are separated by thin hemipelagic muds where biostratigraphic data terraces are expected and observed (Fig. 4). These terraces have been shown to correlate with shelfal deposits in northwest Europe, suggesting a partitioning of sediments between the shelf and deep basin tied to relative changes in sea level (Neal et al., 1994, 1995; Neal, 1996). The depositional model presented here considers the effect of the depositional profile on shelfal accommodation and changing sediment supply as relative sea level fluctuates.

Highstand or Lowstand Prograding

Highstand progradation occurs as relative sea-level rise decreases. The coastline regresses due to sediment influx filling shelfal accommodation faster than it is created (Jervey, 1988; Posamentier et al., 1988). Theoretically, parasequence stacking patterns and clinoform geometries will be aggradational to progradational (Van Wagoner et al., 1990). In North Sea well data, this situation is difficult to observe. However, downlap onto the maximum flooding surface should be apparent with the maximum flooding surface capping all lowstand deposits. In North Sea data, the clearest downlap surface comes directly over the bypass package, below the prograding delta. This presents three possible systems tract interpretations for this section: (1) the prograding delta is highstand progradation and the lowstand delta is thin or absent (e.g., Vining et al., 1993), (2) the prograding package represents the lowstand prograding, transgressive and highstand systems tracts combined due to overpouring sediment supply from the uplift of the British mainland (e.g., Jones and Milton, 1994), or (3) the prograding is a lowstand delta with transgressive and highstand deposits thin or absent in this location.

We interpret the prograding packages to be lowstand based on their internal stratal patterns, relationship to canyon incision (see Fig. 6), the presence of graphic correlation terraces on the shelf and well log interpretation indicating deepening over the progradation. With this interpretation, a depositional model through one cycle of sea level change is illustrated in Figure 7 (A-D).

Deposition Through a Cycle of Relative Change in Sea Level

Identifying highstand deposits is difficult due to uplift and erosion of older sequences into younger deposits. The diagram of a highstand shelf is hypothetical (Fig. 7A) and considers the effect of a reduced sediment provenance area in the production of sediment, and its delivery to the coast. The main source area for North Sea Paleogene deposits is the Scottish Highlands (Ziegler, 1990) with an estimated 1 km of uplift (Den Hartog Jager, et al., 1993). This area is small compared with adjacent basins to the west (Rockall-Faeroe) and east (North Sea). Present-day highstand deltas are building out onto extensive shelves in the Gulf of Mexico (Hamilton and Anderson, 1993) and Mediterranean Sea (Rhone delta, Posamentier et al., 1992b). These deltas have large sediment source areas compared to the Scottish Highlands, which split its potential sediment between two basins. Therefore, sediment supply during highstand periods in the central North Sea is believed to be low.

When sea level begins to fall in this setting, conditions exist for “forced regression” deposits to appear according to Posamentier et al. (1992b). Galloway et al. (1993, p. 38) describe shelf erosion and bypass unconformities in the North Sea following a model proposed by Plint (1988), where wave base erosion removes potential “forced regression” deposits as sea level falls. The North Sea depositional profile acts like a ramp on the shelf, but rapidly deepens across a shelf edge with slopes reaching 5–7°. Sediments swept from the shelf during a fall in
FIG. 6.—Correlation of depositional sequences in a shelfal setting by graphic correlation, well log and seismic data interpretation. Only the LOC for each well is plotted on the same CSU scale (410–1060). The base Upper Forties terrace just overlaps in both wells, the brevity of overlap due to differences in systems tracts encounter by the respective wells. The Upper and Middle Sele sequences also do not precisely overlap on the graphic correlation plots, but they are bracketed by overlapping terraces above and below. Here seismic and well data are necessary to verify the correlation. Lithostratigraphic units are from Robertson Research well reports.
sea level pass into the basin as turbidites (Fig. 7B), forming sandy fans (point source) or gravity flow aprons (line source). Sediment shed from the source area is added to sediment ponded in the hightstand coastal zone, increasing the total sediment input to the basin and shelf.

The shelfal bypass sediments come into the basin through the fall of sea level into its early rise. The early rise of sea level marks the first re-establishment of a lowstand shelf; however, this shelf is unstable and prone to slumping due to the steep slope inherited from the previous depositional sequence. The effect of this slumping and the bypass fill of the slope front is to reduce the profile gradient enough to allow stable progradation to occur. The delta will build out as long as sediment flux into the basin is greater than the rate at which shelfal accommodation is created due to increasing relative rise in sea level (Posamentier et al., 1988). As the rate of relative rise in sea level comes into equilibrium with sediment flux, the delta will aggrade. During this aggradation, the lowstand coastal plain will onlap landward and broaden (Fig. 7C).

This broad, flat coastal plain creates the potential for a rapid transgression once the rate of relative rise exceeds sediment flux. A small relative rise in sea level pushes the coastline far landward, lowering sediment supply to the shelf (Fig. 7D). Once flooded, the shelf may be swept by storm- and fairweather wave base (Plint, 1988) explaining the lack of thick transgressive and highstand deposits.

PALEOGENE STRATIGRAPHIC CYCLES AND DEPOSITIONAL SYSTEMS

This study documents five regressive/transgressive facies (second-order) cycles, the lower four corresponding to similar units mentioned by Den Hartog Jager et al. (1993, p. 61). Overprinting depositional sequence (third-order) cycles are also documented and regionally correlated. These cycles are recognized from their physical expression in the rocks rather than their absolute time duration (sensu Vail et al., 1991). How our cycles compare with published results of other studies will be discussed below.

Depositional sequence cycles reflect a high-frequency sea-level signal that controls distribution of depositional systems and lithofacies. The facies characteristics of individual sequence cycles are linked to the stacking of sequences within a regressive/transgressive facies (R/TF) cycle. Sequences occurring within a regressive phase will contain significant submarine fan deposits, whereas sequences within transgressive phase may be completely sediment starved in the basin and be recognizable through correlation to outcrop sections around northwest Europe. Sequence cycles may be eustatic, however, they have only been documented to be correlative for the central North Sea Basin and its attached embayments of northwest Europe (Neal, 1996; Neal et al., 1994).

R/TF cycles reflect major reorganizations of depositional systems and are bracketed by major condensed (hiatal?) intervals. R/TF cycles contain multiple depositional sequences that are
best resolved during the regressive phase. R/TF cycles have different magnitudes and are believed to reflect periods of uplift and subsidence of the basin rim and sediment provenance areas. In the central North Sea, it is easiest to identify a new cycle by its renewed regression, but outcrop studies around northwest Europe usually distinguish cycles based on renewed transgression over a major unconformity (Vail and Hardenbol, 1979). For this reason, a complete study of both subsurface and outcrop sections is required to ensure that sequences are not miscorrelated from area to the other. R/TF cycles are bounded by peak floodings, which are easier to identify than peak regressions for reasons discussed below.
Fig. 10.—Stratigraphic cross section through the Witch Ground Graben area of the Central North Sea (see Fig. 1 for location). A composite graphic correlation display of the three logs with detailed biostratigraphic reports highlights the major transgressions that bracket multi-sequence regressive pulses into R/TF cycles. Paleobathymetry of the Lower Paleocene pulse is bathyal, while the Upper Paleocene shallows to inner shelf and transitional. The next two regressions are deposited in outer shelf/upper bathyal depths, with UK 15/6-1 having an inner shelf Upper Middle Eocene section (see text for discussion). Paleobathymetry estimates are from S. P. T. Ltd. (Robertson Research).
Regressive/Transgressive Facies and Sequence Cycles from Well and Seismic Data

An example displaying the nesting and similarities of sequence and R/TF cycles is seen from seismic data and in a well-log cross section from the Outer Moray Firth area (Figs. 8, 9, 10). The Lower Paleocene regression consists of sandy turbidites in bathyal paleowater depths at this location (Fig. 8). A Lower Paleocene shelf occurs 20 km to the west of this section (Jones and Milton, 1994). The regressive phase of the Upper Paleocene cycle forms a multi-sequence shelf that progrades far into the basin. Depositional sequence interpretations are based on interpreted basinward shifts of facies from well logs and tied to mappable seismic reflection discontinuities with synthetic seismograms. Above the rapid progradation, there occurs a high-amplitude seismic interval (Figs. 8, 9), interpreted as coaly coastal plain section that aggrades rather than progrades. This section also has more than one depositional sequence. The depositional systems are responding to a turnaround in the R/TF cycle and the lithofacies are characteristic of a transgressive system of stacked coals (Cross, 1988; Milton et al., 1990) rather than...
than the regressive characteristic of rapidly prograding deltas. Above the coaly section is a throughgoing seismic reflector that can be traced to wells that identify a graphic correlation terrace (Fig. 10). This reflector drapes over topographic relief in the coaly sections. Milton et al. (1990) interpret topographic relief in the top coal reflector to represent differential compaction of the underlying strata. The mudstone drape and compaction sag represent a period of sediment starvation in the basin.

The next regression marks another R/TF cycle of lesser magnitude than its predecessor. The small prograding unit downlapping the draping seismic reflector does not reach the previous shelf edge but is more extensive further north (Timbrell, 1993). This regression is important as it produces the submarine Frigg fan with 7 Tcf of gas in place (Heritier et al., 1980). The top of this regressive unit is traced to wells that display another major graphic correlation terrace (Fig. 10). The top of this unit is a basinwide downlap surface that marks a period of quiescence before the next major regression occurs.

The Mid-to-Upper Eocene regression is very sandy on the shelf and upper slope. It has been linked to renewed uplift in the Shetland Platform (Jones and Milton, 1994; Galloway et al., 1993). Graphic correlation indicates that this unit was deposited very rapidly and is capped by one of the longest graphic correlation terraces of the study (Fig. 10). The Upper Eocene terrace will be discussed below, but for now it will simply be described as a stratigraphic break between the Mid-to-Upper Eocene cycle and the overlying Lower Oligocene cycle that thickens to the east. The top of the Lower Oligocene section is a major flooding that signals another reorganization of the basin with subsequent cycles having a larger input from Norway (Jordt et al., 1995; Galloway et al., 1993).

These examples illustrate how higher order cycles can be recognized as data resolution and stratigraphic observations improve. Jones and Milton (1994) have interpreted the Paleogene section as comprising only two second-order cycles. Figure 8 shows how a two-fold division of this section can be easily observed. The sedimentation cycle of bypass, progradation and transgression occurs at a minimum of three different time scales within this example alone. Each cycle can be subsequently split into additional divisions with the sole limiting factor being data resolution. The list of mechanisms that may have caused a stratigraphic cycle changes as the duration of the cycles grows shorter (Van Wagoner et al., 1990). Once divisions are at the scale where autocyclic processes dominate, then the basin-wide correlative nature of events becomes suspect. We have settled for a division of events in the fourth to third order (sensu Vail et al., 1991), which we call sequence cycles, superimposed on R/TF cycles of approximately 3–13 m.y.
Lower Paleocene Cycle

The Lower Paleocene (Paleogene 1; Neal, 1996) second-order cycle formed in response to uplift and tilting of the Scottish Highlands and Shetland Platform (Parker, 1975; Stewart, 1987; Milton et al., 1990). It begins with the Cretaceous/Tertiary boundary in this study, although evidence from chalk reworking events suggests that uplift began in the Upper Cretaceous time (Kennedy, 1987). The upper boundary of this cycle is a data terrace associated with an acme occurrence of *A. gippingensis*. O’Connor and Walker (1993) discuss how this biostratigraphic marker relates to the lithostratigraphic divisions of Lista and Sele Formations (Mudge and Copestake, 1992), finding that it occurs in the upper portion of the Lista, not precisely at the Lista/Sele boundary (see Fig. 3). This interval was recognized by Stewart (1987) as a second transgressive phase within the Paleocene, although its precision varied (Fig. 3—Sequence 6). Subsequent revisions of Stewart’s (1987) framework (Milton et al., 1990; Jones and Milton, 1994) have recognized only one major regressive phase in the Paleocene, comprising the Andrew and Forties cycles of Den Hartog Jager et al. (1993), which approximate the Lower and Upper Paleocene cycles of this study (Paleogene 1, 2; Neal, 1996). Knox (1996) further subdivides the Lower Paleogene R/TF cycle into the lithostratigraphic-based Ekofisk and Montrose cycles.

The Lower Paleocene R/TF cycle may be subdivided into four (and possibly five) sequence cycles in the subsurface data of this study (Fig. 3). Biostratigraphic calibration of the upper boundary data terrace to northwest Europe identifies an extra sequence not resolved in the study area (Neal, 1996; Neal et al., 1994). The Ekofisk sequence may comprise two sequences, although this division is difficult to distinguish where Ekofisk sands do not occur. These two sequences are thickest in the study area over pre-existing Mesozoic grabens that had not filled with Cretaceous chalk. In these localities, the sequences may be identified seismically, but outside of the depocenters it is often difficult to distinguish the top Cretaceous horizon from the top Ekofisk Formation Danian chalk (Johnson, 1987).

Above the chalk, siliciclastics come into the basin as uplift continues and erosion progresses down to Devonian sandstones (Ziegler, 1990). Figure 11 illustrates the distribution of clastic fans and seismic expression of the Lower Paleocene peak regressive shelf in the Moray Firth area from a seismic line published by Jones and Milton (1994). The Maureen sequence, whith key dinocysts markers *S. inornata* at its base and *A. reticulata* at its top, fills in the Moray Firth Basin with a wedge of sediment that has depositional dip slopes of 2°–3°. Locally, mounded fan deposits occur at the eastern edge of UK Quad 13 (Fig. 11). Further east, this sequence becomes thin and marly with occasional blocks of chalk debris until reaching the eastern edge of the junction between the South Viking and Central Grabs (Fleming area). Here, locally well-developed sandy fans are found. If the source area was restricted to northern Britain,
NESTED STRATIGRAPHIC CYCLES AND DEPOSITIONAL SYSTEMS OF THE PALEOGENE CENTRAL NORTH SEA

Fig. 15.—Seismic expression and map distribution of some delta lobes and coals within the transgressive Balder sequence with a time structure map of the overlying Top Balder Reflector. Seismic data courtesy of Schlumberger/GECO-Prakla and Nopec.

these sands bypassed almost 140 km of basin floor before being deposited against submarine topography according to the onlap model proposed by Scott and Tillman (1981). Maintaining a turbidity current requires a slope of at least 0.5° (Stow, 1986), which fits Rochow’s (1981) model of basin floor tilting and deepening in response to uplift of the Shetland Platform.

Thulean volcanics in western Scotland result from ‘hot spot’ activity of a plume presently situated under Iceland (White, 1988). The peak volcanic activity occurred around 60 Ma (Murray et al., 1988), which coincides with peak uplift of northern England from fission tract analysis (Green, 1989). The timing of this peak uplift correlates with the base of the Andrew sequence from biostratigraphic ties. The Andrew sequence has a more extensive fan system than the Maureen sequence (Fig. 11), which reflects this peak uplift and erosion of more clastic sources. The Andrew sequence, containing the key dinocyst *P. pyrophorum*, and the FDO of the radiolarian, *Cenodiscus* (Fig. 3). This unit was identified as Sequence 3 by Stewart (1987), but Figure 11 shows the distribution of the underlying Andrew fan, which closely resembles a map for Stewart’s Sequence 5 (1987, Fig. 11).

**Balmoral Tuffite/Glamis Member**

Within the Lower Balmoral sequence, volcaniclastic sands are ascribed to an eruption phase in the Thulean provenance (Knox and Morton, 1988, Phase 1). These sands are found in the Outer Moray Firth area and have been designated the Glamis Member of the Lista Formation by Mudge and Copstick (1992) and Balmoral Tuffite by Knox and Holloway (1992). The Balmoral Tuffite has been designated as a separate depositional sequence in some sequence stratigraphic frameworks of this section (Stewart, 1987; Milton et al., 1990). The low gamma-ray response and high sonic velocities make this unit stand out in well logs and on seismic sections. Biostratigraphically, this unit is age-equivalent with Lower Balmoral sediments that do not contain the volcanics (i.e., above the acme of *P. pyrophorum* and below the acme of *A. gippingensis*). Knox and Morton (1988) ascribe the top of their Phase 1 to nannofossil zone NP6, but we correlate this unit to NP7 for reasons discussed below. Seismically, the tuff reflector is discontinuous and appears to form large, laterally onlapping lobes (Fig. 12). Picking the base of a single lobe as
FIG. 16.—Thick package of Balder sequence delta lobes in strike view, highlighting the discontinuous nature of coals capping the lobes. Also shown is an expanded view of the line illustrating an onlapping wedge of Lower Eocene (Frigg Undiff.) sediments, infilling a sag of the Balder sequence over a Mesozoic Graben. The Balder package shows that multiple depositional shifts may be autocyclic or allocyclic. Seismic data courtesy of Schlumberger/GECO-Prakla and Nopec.
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FIG. 17.—Time thickness map of the Lower Eocene (Frigg) regression shelf and some bypass units on a time structure map of the Top Balder discontinuity. Displayed seismic data illustrates the subsidence sag of Balder sequence coals and subsequent onlapping fill of Frigg-equivalent sediments that are then downlapped by the Upper-Middle Eocene regression (see text for further discussion). Seismic data courtesy of Schlumberger/GECO-Prakla and Nopec.

a sequence boundary may correlate to the top of a different lobe. We agree with Knox and Holloway (1992) that this unit should be recognized for its stratigraphic position, but that it does not constitute a separate sequence. The Lower Balmoral sequence may be divisible into more sequences (e.g. Den Hartog Jager et al., 1993; Knox, 1996), but our dataset was unable to resolve and correlate any more sequences.

Upper Paleocene Cycle

The Upper Paleocene to Lowermost Eocene (Paleogene 2; Neal, 1996) sediments record the best developed R/TF cycle in the study area. Nearly complete depositional sequences are found with well-developed fans, prograding deltas and even some transgressive and highstand deposits. This R/TF cycle also contains the Lista/Sele Formation boundary that marks a time of oxygen-depleted bottom water conditions thought to be related to an isolation of the North Sea basin from open marine circulation due to tectonic uplift (O'Connor and Walker, 1993; Knox and Holloway, 1992). The upper sequence in the Upper Paleocene cycle contains tuffaceous claystones of the Balder Formation. This tuff is found throughout Northwest Europe and is linked to explosive volcanism in the Thulean provenance that has been timed with a collapse of Paleocene uplift of Scotland (Knox and Morton, 1988; Milton et al., 1990; Knox and Harland, 1979). The oldest depositional sequence of this R/TF cycle is the Upper Balmoral sequence. The namesake of this sequence is the Balmoral Member of Mudge and Copestake (1992) in well 21/10-1, but here lithostratigraphy and sequence stratigraphy diverge as the Upper Balmoral sequence is younger than the Balmoral lithostratigraphic unit in wells 14/25-1 and 15/26-1 (Fig. 3). This sequence occurs above a graphic correlation terrace that often contains the acme of A. gippingensis, with the upper terrace boundary near the Paleocene reappear-
below that the Lower Dornoch Sandstone of 14/30-1 (Knox and Holloway, 1992, p. 105) is not equivalent to the same named unit in well 14/25-1. These inconsistencies highlight correlation problems between lithostratigraphy and chronostratigraphy.

**Upper Forties Sequence Deposition**

The Upper Forties sequence rests on the Upper Balmoral sequence and is marked by an influx of *Apectodinium* spp. especially *A. augustum*, which appears almost exclusively within this sequence. The Upper Forties sequence marks the peak Paleocene regression; although in some locations, the Middle Sele sequence shelf progrades slightly further into the basin. As discussed in Figure 6, the relative fall of sea level for this sequence is estimated at 150 m. The distribution of this fan has been mapped by Armstrong et al. (1987), Knox and Holloway (1992) and Stewart (1987). While precise boundaries and thicknesses may differ, general agreement can be reached on the map distribution of the Forties fan.

Figure 13 charts the distribution of mapping units within the Upper Forties sequence. The Upper Forties Map Unit A (UFA) is mapped on a strong seismically reflector discontinuity within the Upper Forties sequence (Fig. 13). Depositional strike lines across the Upper Forties sequence shelf demonstrate this unit to consist of both discordant and chaotic internal reflections. The Upper Forties Map Unit B (UFB) is purely progradational and onlaps in a landward (west) direction. The UFB also onlaps UFA, demonstrated on seismic data (Figs. 6, 13), after UFA feeds directly through a deep incision in northwestern UK Quad 15. The chaotic reflectors of UFA are interpreted to represent slumped clinoform, concordant reflectors are found in strike lines that cross UFA where dip lines indicate progradation. The Upper Forties map units, along with the main Upper Forties fan deposition document a transition from the lower lowstand bypass sedimentation to the stable progradation and onlap of the upper lowstand. UFA represents the attempts of an early lowstand shelf to re-establish stable progradation, but the delta encounters depositional slopes too great for stability in some areas and slumping occurs. The slumped deposits and prograding remnants form a depositional slope more amenable for stable progradation of UFB. This transitional unit (UFA) illustrates the need for caution when applying the precise definitions of “slope fan” and “lowstand prograding” from published sequence stratigraphic models (Vail, 1987). It also provides another example of the diachronality of the top fan reflector that was recognized by Kolla and Perlmutter (1993).

**Middle Sele Sequence and the Aggradational Phase**

The Upper Forties and Upper Balmoral sequences of the Upper Paleocene cycle can be described as the regressive phase in that they rapidly build out into the basin following the major transgression at the top of the Lower Paleocene cycle. The next three sequences (Middle Sele, Upper Sele, and Balder) represent aggradational and transgressive phases of the Upper Paleocene second-order cycle. These Upper Sele and Balder sequences have a more aggradational stacking pattern compared with the Upper Forties and Upper Balmoral sequences. The Middle Sele sequence occupies a transitional position between the transgressive and regressive phases.

The Middle Sele sequence is positioned stratigraphically above a graphic correlation terrace that often contains the FDO of *A. augustum*. The best biostratigraphic marker for this sequence is an acme occurrence of the acritarch, *Pterospermella*. The Middle Sele contains Hermod (Isaksen and Tonstad, 1989), Cromarty (Murdo and Copestake, 1992) and Flugga basal sands, which were mapped by Knox and Holloway (1992) as well as Beauty coals in shelfal wells. The progradational shelf of the Middle Sele sequence, which normally falls within the upper part of the Dornoch Formation has been mapped from seismic data (Fig. 14). Although technically more regressive than the Upper Forties shelf, the Middle Sele progradation is more aggradational and only locally progrades further into the basin. These localities are closely linked to Mesozoic structural elements with depocenters occurring to the north and south of Halibut Horst and a third to the south where subsidence-related faulting at the intersection of Moray Firth Basin and the Central Graben occurs (Fig. 14).

The Upper Sele sequence is thin throughout the study area. The sequence is unresolved with the present graphic correlation; however, a biostratigraphic signature of an influx of large leiospheres and an acme occurrence of *C. wardenseni* distinguish this unit. The complications of a type log signature for this sequence are discussed above; therefore, the sequence is just noted here for its position within the aggradation-transgression phase of the Upper Paleocene R/TF cycle.

**Balder Sequence Depositional System**

The Balder sequence is very complicated as third order sea level changes interfere with the increasing second-order rise. The complexity of this sequence was demonstrated by Timbrell (1993) for the Beryl Embayment region to the north of our seismic coverage. In the basinal portion of the central North Sea, this sequence is recognized within a tuffaceous claystone with low gamma ray values and a strong seismic reflection. This sequence seems to closely tie with the B2 subdivision of Knox and Holloway (1992). Biostratigraphically, this sequence is marked by an acme of *D. oebisfeldensis* at its base and an acme of *Inaperturepellollenites* spp. at its top. The radiolarian, *Coscinodiscus* also has an acme near the top of this sequence, and an important hiatus is associated with this level. The difficulty in correlation occurs when the shelfal deposits of this sequence are calibrated to basinal observations.

The high-amplitude seismic reflectors of the Beauty coal-bearing formation have a complex pattern of erosion that we cannot correlate regionally with any certainty. When viewed with strike sections (Figs. 15, 16), thick deposits of this high-

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amplitude seismic facies across Halibut Horst display distinct lobe geometries. In wells that penetrate these thick (Fig. 16), multiple coal horizons are encountered within an overall sandy section. The discontinuous coal reflectors cap sediment lobes that onlap each other and backstep overall. Elliot (1986, p. 124) describes a tide-dominated deltaic sequence saying, “The entire delta plain probably comprises a sheet-like complex of small-scale, erosive-based sequences which pass upward from point bar sand-silts into mangrove swamp facies, with localized clay plugs representing the infilled channel.” If the discontinuous coal reflectors are accepted as the remnants of the mangrove swamp facies, then Elliot’s description closely resembles the well and seismic facies encountered in the Balder sequence shelf. An interpretation of a tide-dominated shelf appears reasonable for this setting as the present-day North Sea is known to one of the stronger tidal regimes in the world due to its nearness to tidal resonance (Johnson and Baldwin, 1986). Figure 15 shows the map distribution of numerous delta lobes identified on seismic. Note especially the backstepped relationship of these deposits to the Upper Paleocene peak regressive shelf, highlighting the transgressive nature of this sequence (see also Fig. 9).

Lower Eocene (Frigg) Cycles

Figure 9 highlighted the small magnitude Lower Eocene R/TF cycle (Paleogene 3; Neal, 1996). Evidence for this cycle’s shelf is problematic in most of the seismic data in our study area. Often the Lower Eocene R/TF cycle is either condensed, eroded or indistinguishable from coastal plain deposits of the upper sequences of the Upper Paleocene cycle. Figure 16 displays an important sediment wedge onlapping the top of the Balder sequence in a location that has implications for the time represented by these deposits. This onlapping wedge is located over a Mesozoic graben. If the Balder sequence shelf is assumed to have been deposited at, or near baselevel (evidenced by widespread coals), then the local accommodation created for this wedge could have developed from subsidence related to differential compaction of sediment in the underlying graben compared to lack of compaction over the horsts (Milton et al., 1990). With a compaction differential of 4 cm/k, the section over the graben compared to that over the horst, at least 1.5 my must pass before enough accommodation is created for the 60 (+) m of sediment to fill.

Biostratigraphically, this wedge is not assigned an age in the Shell 14/24-2 well that penetrates it; however, the wedge sediments do overlie the PT 21 zone that has been correlated to the FDO of Coscinodiscus spp. in nearby wells. The sediment overlying this wedge is designated PT 22/23, which covers the entire Lower and Middle Eocene (Schroeder, 1993; Den Hartog Jager et al., 1993). The overlying package has been correlated to other wells with more detailed biostratigraphy by seismic data and is calibrated to the base Middle Eocene sequence. The onlapping wedge must be deposited during Lower Eocene time by process of elimination. When traced 5 km to the west, this wedge has pinched out and the Upper-Middle Eocene sediments lie almost directly on Upper Paleocene section, similar to the pitchout relationship shown in Figure 8.

Figure 17 shows a seismic example of a sediment package filling one of these lows and the map distribution and thickness of the Lower Eocene shelf and bypass sediments. Key to recognizing this interval is the observation of a regional downlap surface above the coal reflectors. This surface represents the Top Frigg graphic correlation terrace, which is normally associated with the FDO of Eotoninysa ursulae. In wells north of our seismic coverage area, three to four depositional sequences can be observed between the key markers for the Balder and Top Frigg graphic correlation terraces (Mudge and Bujak, 1994). Although smaller in magnitude than the other R/TF cycles, the Lower Eocene cycle still contains multiple third-order cycles and represents a reorganization of the depositional system.

Upper-Middle Eocene and Lower Oligocene Cycles

The Upper-Middle Eocene cycle (Paleogene 4; Neal, 1996) has as great a magnitude as the Upper Paleocene cycle, however, it is not nearly as well understood. Seismic data through the Middle Eocene sediments is of poor quality due to disruption of reflectors by swarms of small faults related to dewatering of the sediment (Claussen and Korstgaard, 1993). Depositional models for this section are still evolving along with a refinement in the biostratigraphic framework. The stratigraphy presented by Newton and Flanagan (1993) with the biostratigraphic framework of Paleo Services closely resembles that observed in our study. For this reason, the lithologic units, Caran, Nauchlan, and Brioc have been adopted and appear in the summary framework below with key fossil markers and data terraces that we have found to be consistent.

The depositional systems of this cycle are poorly understood and published work has focused almost exclusively on depositional models for oil-bearing stringer sands in the slope environment (Newton and Flanagan, 1993; Harding et al., 1990). The shelfal system of this cycle, particularly the Nauchlan sequence, is very thick (over 300 m) and progrades rapidly over the Lower Eocene and Upper Paleocene drowned shelves. Clinoforms on seismic indicate a paleobathymetry around 300 m that was filled almost entirely with sand in the Witch Ground Graben (Fig. 10). South of Halibut Horst, this sequence is siltier (Fig. 18) with the best sands captured in transgressive deposits above the prograding package.

The top of the Upper Middle Eocene cycle is a major stratigraphic break. The actual Eocene/Oligocene boundary is marked by the LAD of Aresphaeridium diktyoplokus (Vinken, 1988; Brinkhuis and Biffi, 1993). The boundary is a major marine hiatus in the basin (Harding et al., 1990), but on the shelf, the LAD of A. diktyoplokus ranges within transgressive sands that often contain fossils of the entire Upper Eocene. The result is the longest graphic correlation terrace recorded in the North Sea Paleogene section (Fig. 10). Examination of this age in outcrops of northwest Europe reveal a similar phenomena as the transgressive Bassevelde sand of Belgium (Vandenbergh et al., this volume) is reported to contain nanofossils of NP 18–21 zones (Fig. 19). The upper boundary of the Upper-Middle Eocene cycle is thought to represent a slow transgression with decreased sediment supply before regression reoccurs during Lower Oligocene time.

The Lower Oligocene R/T F cycle (Paleogene 5; Neal, 1996) marks a shift in sediment delivery to the North Sea from Britain to Norway (Jordt et al., 1995; Galloway et al., 1993). Biostratigraphic control for this section is poor since few well reports log samples within these sediments. In the few wells that do contain biostratigraphy, this cycle appears to be bracketed by
A. diktyoplokus at its base and A. arcuatum at its top. In shelfal wells, two depositional sequences can be distinguished, which expand into silty thickens in the basin. Figure 18 shows many of the features described within the nested stratigraphic cycles above and serves as a good overview.

DISCUSSION AND REGIONAL CORRELATION

Knox (1996), Powell et al. (1996) and Mudge and Bujak (1996) present Paleogene-aged sequence stratigraphic frameworks that differ slightly in resolution and emphasis. Mudge and Bujak (1996) and Powell et al. (1996) both emphasize biostratigraphic calibration. Knox (1996) provides a detailed physical stratigraphy calibrated to biostratigraphy. Mudge and Bujak (1996) focus on central North Sea sequences that they could map from well control, while Powell et al. (1996) concentrates on outcrop work in the London-Hampshire Basin to tie facies evidence for sea-level change with the biostratigraphic framework. The succession of key dinocyst markers is very similar for these papers, the Composite Standard presented above (Fig. 5, Appendix 1), and the new Paleogene cycle chart (this volume), but the onshore calibration of Powell et al. (1996) and Knox (1996) differ from the correlation presented in this paper, apparently due to the recalculation of dinocyst markers to nannofossil zones.

The sequence stratigraphic framework we present hinges on calibrations from the deep basin to onshore outcrops. This tie was accomplished through calibration of the North Sea Composite Standard (Fig. 5, Appendix 1) with key boreholes in Denmark and Germany, namely the Viborg 1 (Heilmann-Clausen, 1985) and Wursterheide (Heilmann-Clausen and Costa, 1989) boreholes (Neal et al., 1994; Neal, 1996). Critical to the correlation of North Sea lowstand fans with onshore shallow marine highstand deposits is the calibration of the largely nannofossil-based outcrop framework (Aubry, 1983; Aubry et al., 1988; Berggren and Aubry, 1996) to the more dinoflagellate-based subsurface framework (e.g., Mudge and Bujak, 1996). Our study used Powell’s (1992) calibration as the best available tie, which is reflected in the correlations presented in this work and in Michelsen et al. (this volume). The dinoflagellate-nannofossil calibration of Powell (1992) relied heavily on the results of IGCP 124 (Costa and Manum, 1988) and the same Viborg 1 borehole (Heilmann-Clausen, 1985) so critical to our work (Neal et al., 1994). Since 1992, revised calibrations have come forward, particularly Heilmann-Clausen’s (1994) revision of Powell’s (1992) framework. The recalibration is particularly important in the Selandian/Thanetian section.

A key point of contention is that our framework correlates the Andrew sequence (capped by the top acme occurrence of P. pyrophorum and containing the LAD of I.? viborgense) to NP5-7 (pars.), encompassing the basal Orembsby Clay and Pegwell Marls into this sequence’s highstand. Also contentious is the correlation of the Lower Balmoral sequence (capped by the top acme occurrence of A. gippingensis and containing the Balmoral Tuffite) to NP7-8 and the Reculver Silts. Both Knox (1996) and Powell et al. (1996) calibrate the Lower Balmoral to the Pegwell Marls, with Knox (1996) having an extra sequence for the basal Orembsy Clay that he correlates with our Andrew sequence.

Powell et al. (1996) imply that superabundant A. senonensis reported from the Pegwell Marls is equivalent to the Areoligera acme separating Upper from Lower Balmoral sandstones. There is some confusion though, since the North Sea marker is now renamed A. gippingensis (Jolley, 1992; Knox and Holloway, 1992), a marker reported to have its top acme within the Reculver Silts at the BGS Borehole 217, Orembsy A borehole, and Flatford Lock/Deadham Mill sections (Jolley, 1992b). Jolley’s (1992b) Pegwell Bay sampling shows the peak abundance of A. gippingensis to be low within the Pegwell Marl only at the Pegwell section (also Powell et al., 1996). If A. gippingensis is indeed the key marker, then our correlation of Lower Balmoral to Reculver Silt is correct on palynological evidence. The red mudstone reported by Jolley (1992b) and cited by Knox (1996) as lithostratigraphic evidence for a correlation of Pegwell Marls to Lower Balmoral sandstones must be viewed with some caution. Large paleogeographic differences exist between the shallow shelf Pegwell Marl setting and the deep marine Lower Balmoral fan environment. It is unclear how the same red mudstone could have been deposited synchronously in both depositional settings.

The stratigraphic position of the Balmoral Tuffite is also worthy of discussion. Knox and Morton (1988) place the top of pyroclastic Phase 1 in upper NP6, making it equivalent to the Pegwell marls also dated upper NP6 (Knox, 1996; Knox et al., 1994). All stratigraphic frameworks agree that the Balmoral Tuffite occurs above the top acme of P. pyrophorum (Neal, 1996; Knox, 1996; Knox and Holloway, 1996; Mudge and Bujak, 1996). The LAD of P. pyrophorum in onshore UK sections is reported by Jolley (1992b) within the lowermost Thanetian. The revised biochronostratigraphic chart (this volume) records the true LAD of this marker near the top of the Thanetian, suggesting the observed LAD in outcrop may be environmentally depressed and the acme not even recorded (e.g., Armentrout et al., 1991; Armentrout and Clements, 1991; Powell, 1992, p. 223). This point requires more research, but we note that Berggren and Aubry (1996, p. 328) calibrate the top of Phase 1 to NP7, in agreement with our framework, supported by volcanic extrusives dated 57.2 ± 0.8 and 57.6 ± 0.6 Ma (NP6-8, Berggren et al., 1995) from a cored sill in BGS borehole 88/5 from The Minch in western Scotland (Ritchie and Hitchen, 1996). Additionally, a point of confusion regards the lithology description of BGS borehole 217, which is recorded with a tephra layer in the base Reculver Silt equivalent bed in Jolley (1992b, p. 216). His Figure 7 would strongly support a correlation of the Lower Balmoral sequence to the Reculver...
Silt equivalent. The same tephra layer is not recorded by Knox et al. (1994).

Regarding the calibration of the LADs of \( P. \) pyrophorum and \( I. \) viborgensis to nannofossil zones, Powell (1992) calibrated his Ppy zone to NP6–NP7. This was based on ties to the results of IGCP 124 (Costa and Manum, 1988; Powell, 1992, p. 173). Zone Ppy was defined at its base by the first appearance datum (FAD) of \( A. \) margarita and its top by the FAD of \( D. \) denticulata. Also at the top, Powell (1992) noted the last appearance datum (LAD) of \( P. \) pyrophorum, which Jolley (1992b) observed in the basal Thanet Sand Formation. Powell (1992) placed the LAD of \( I. \) viborgensis within the Ppy zone and into NP6. For reasons not documented, Heilmann-Clausen (1994) redefines this dinoflagellate-nannofossil calibration, placing the top of the Ppy zone at the NP5-6 boundary (instead of Powell’s NP7-8), moving the LAD of \( I. \) viborgensis down to NP5 instead of NP6. Based on this recalibration, not direct nannofossil evidence, Knox (1996, p. 219) moved the basal Orembsy Clay (sequence OTh-1) into NP5 because it contained the reported LAD of \( P. \) pyrophorum (Jolley, 1992b).

We agree with Knox’s (1996) correlation of the Orembsy Clay to the Andrew sequence, but believe his OTh-2 sequence (Pegwell Marl) to correlate within the Andrew sequence as well. As discussed above, we are unclear as to why the top \( P. \) pyrophorum was moved from NP7-8 to NP5-6, and consider this top to be environmentally depressed in the Thanet outcrops. This interpretation frees us to correlate the Lower Balmoral sequence to the Reculver Silt, supported by reasons given above. The same situation may apply to the top acme of \( A. \) gippingensis in the Pegwell section. We agree with Knox (1996) that his OTh-4 sequence is sediment starved in the central North Sea basin and we correlate this sequence within a ubiquitous sediment starvation data terrace in our graphic correlation plots (Lista Terrace). This terrace usually encompasses the top acme occurrence of \( A. \) gippingensis, signifying its tie to the highstand Reculver Silts (Powell et al., 1996), and separates the Upper and Lower Balmoral sequences. We consider the Upper Balmoral to be the lowstand fan of Th-4 (Powell et al., 1996) and OTh-5/Lmb-1 (Knox (1996). The strength of the Lista terrace is explained by the condensation of sedimentation recorded in a highstand (Reculver Silts) and complete sequence (OTh-4 and Th-3) onshore (Neal and Hardenbol, this volume). The Upper Balmoral sequence captures the lithostratigraphic unit of the same name (Knox and Holloway, 1992) and represents regression following the Thanetan transgression.

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FIG. 19.—(Continued) and Steurbaut & Nolf (1986). ‘Paris Basin hiatuses from Pomerol (1989). ‘Danish seismic sequences from Michelsen et al. (this volume). Figure key: ↔ New (O & SS) = position of a sequence boundary from outcrop and subsurface data not picked in Haq et al. (1988). ↔ New (O) = position of a sequence boundary from outcrop data not picked in Haq et al. (1988). Correlation of the frameworks from Den Hartog Jager et al. (1993), and Armentrout et al. (1993) are based on biostratigraphic calibration points. The Haq et al. (1988) eustatic curve is modified based on a correlation of nannofossil zone differences between Haq et al. (1988) and Aubry et al. (1988).


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Eocene Tectono-Sedimentary Patterns in the Alicante Region (Southeastern Spain)

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ABSTRACT: During the Eocene, the Alicante region formed part of the southern passive margin of the Iberian continent. Shelf deposits are found in the northwest and north, slope sediments in the south and southeast parts of the region. On the platforms 14 Eocene 3rd-order cycles can be recognized. They are separated by erosional, karstic or dolomitized surfaces, or paleosoils. The coeval slope deposits contain both lowstand wedges (mass flows and calciturbidites) and late-highstand wedges (calciturbidites). Lithoclasts and loose foraminifera in these mass flows and turbidites were used to identify adjacent platforms and their sequence stratigraphic history. Three sequence boundaries are coincident with tectonic events (during the middle Middle Eocene, at the Middle/Late Eocene boundary, and during the late Late Eocene). These tectonic events, together with differential subsidence during cycle deposition and correlation with plate-tectonic events point to tectonic control of the majority of the south Spanish cycles. Glacio-eustatic fall may have made an important contribution in producing the sequence boundary at the Middle/Late Eocene boundary. The sequence boundary at the Eocene/Oligocene boundary may be explained exclusively by glacio-eustasy.

INTRODUCTION

Changes in lithofacies patterns and development of depositional sequences are primarily controlled by eustatic fluctuations, tectonic events, sediment supply and climatic changes. Long-term tectonic subsidence creates the vertical space in which sediments are deposited and preserved. The cause of the short-term sea-level changes is still a matter of debate between advocates of a glacio-eustatic mechanism (e.g., Vail et al., 1977; Haq et al., 1988) and those who argue for a tectonic cause of the short-term cycles (e.g., Embry, 1989). Cloetingh (1991) suggested that there might be no controversy at all and that the discussion should evolve into a careful search for the changing roles and interplay of the two mechanisms through geologic time.

This study focuses on the Eocene carbonate platform and slope deposits of the Alicante region (Fig. 1, 2). In the Alicante area, Triassic to Miocene deposits are found in outcrop. Undergraduate and graduate students and staff of both Universities of Amsterdam carried out stratigraphic field work between 1985 and 1988. We made no attempt to compile the data on the Paleogene at the time because our attention was focused on younger deposits (Geel et al., 1992). In 1989, three selected Paleogene sections (Ibi on the platform, Relleu on the slope and the Pantanet on the toe of slope-basin floor, Figs. 2, 3) were visited during a Sequence Stratigraphy Workshop under the guidance of P.R. Vail. Though the chosen sections proved to be very well suited to demonstrate the principles of sequence stratigraphy, it was apparent at the same time that additional field and laboratory work was needed to allow sequence stratigraphic analysis and interpretation of the entire platform and slope deposits of the area. Additional material was sampled and analyzed in 1990 and 1991 by the first author. Subsequent visits to the above sections by all four authors led to fruitful discussions that laid the foundation of the present paper.

Facies analysis and correlation of stratigraphic sections of the Eocene platform and slope deposits of the Alicante area resulted in the recognition of 14 3rd-order cycles separated by sequence boundaries. We will show that plate reorganization played an important role in their generation, but that in the Late Eocene glacio-eustasy may have made a significant contribution.

Throughout this paper, the time scale of Haq et al. (1988) is used. The data from the literature have been adjusted to this scale. However, to avoid circular reasoning, the sequence boundary ages of the Alicante region have not been adjusted to the numbers on Haq et al.’s global chart, but have been given a range in time indicated by the planktonic foraminiferal zones obtained from the Alicante samples. The notation of a P, followed by a number (e.g., P9) in the text refers to the planktonic foraminiferal zones of Berggren (1972). We use the systems tract terminology and abbreviations of Haq et al. (1988).

REGIONAL SETTING

The Alicante area is situated in the easternmost part of the Betic Cordilleras of southern Spain. The Betic Cordilleras can be subdivided in the Prebetic, Intermediate and Subbetic (together the External Zone) and the Internal Zone (Fig. 1). The Internal Zone is considered to consist of a stack of allochthonous units (e.g., Egeler and Simon, 1969; Andreieux et al., 1971; Torres-Roldan, 1979). The External Zone represents the former passive continental margin of southern Iberia during the Me-
sozoic and Paleogene, strongly tectonized during a Burdigalian paroxysmal event due to the collision of Iberia with the allochthonous complexes of the Internal Zone (García-Hernández et al., 1980; Hermes, 1985; Geel et al., 1992). Because our study area forms part of the Prebetic (the northern subdivision of the External Zone), the Eocene deposits show the overall facies distribution that characterizes former passive-margin deposits until the Early Miocene: continental, lagoonal or shallow marine in the north and northwest, and more open marine in the southeast and south (Geel et al., 1992). Platform carbonates are exposed between Onil-Ibi and Alcoy, in the Carrasqueta range, in the Aitana mountains (Almela et al., 1973; Martinez et al., 1977; Colodrón et al., 1980) and to the southwest of Penaguila (not shown on the maps of these authors; see Fig. 2). Concurrent slope deposits (marls, turbidites, and mass flows) can be found in the quadrangle between Jijona-Benifallim-Relleu-Aguas de Busot (Figs. 2, 4). Due to intense Miocene tectonics and burial under younger deposits, a direct relationship between Eocene platform and slope deposits is observed only locally.

METHODS AND DATING

Identification of platform cycles is based on field observations aided by microfacies analysis. In the field, cycle boundaries are recognized by karstic surfaces, paleosoil (Microcodium), evaporites (dolomitization or gypsum), erosional truncation, and change in bedding pattern. In thin section, we observe a sudden change in environment at the boundary, e.g., from shallow open platform (deduced from the presence of Assilina and Discocyclina) to deeper platform (characterized by the influx of planktonic foraminifera), or from a more restricted platform interior (limestones with a predominance of Alveolina) to more open-shelf conditions (Assilina-limestones). Within the cycles, transgressive- and highstand systems tracts are discriminated partly in the field by a change from thinning and fining-up bedding patterns to thickening and coarsening-up patterns, and partly by study of thin sections, in which we see a gradual deepening of the environment followed by gradual or sudden changes toward shallowing or increasingly restricted conditions (deduced from a change in biota, as described above).

Dating the marls of the deeper shelf and upper-middle slope in the Alicante region posed a problem because the marls are either deeply weathered (leached fauna) or, if fresh, often contain a poor or contaminated (reworked) fauna. The same problem is undoubtedly the cause for erroneous dating in earlier publications (e.g., Colodrón et al., 1980). The biostratigraphy based on planktonic foraminifera (see below) is the result of repetitive sampling and careful searching for the youngest elements, yielding the point age datum (a biozone) in one out of several samples in a section.

The following biostratigraphical ages could be ascertained, from bottom to top:

P6b: faunas, containing Pseudohastigerina sp., Globigerina velascoensis, Acaninina soldadoensis, Morozovella aequa, M. marginodentata and M. formosa gracilis, indicate a latest Paleocene-earliest Eocene age. However, the absence of Morozovella velascoensis warrants an assignment to the earliest Eocene (Pantanet, Onil, Ibi).

P7: the presence of Morozovella formosa gracilis, M. formosa formosa and M. aragonensis indicates the M. formosa Zone (Pantanet).

P7/8: samples from Onil/Ibi, containing both Morozovella formosa formosa and Acaninina broedermanni, indicate either
EOCENE TECTONO-SEDIMENTARY PATTERNS IN THE ALICANTE REGION (SOUTHEASTERN SPAIN)

FIG. 3.—Stratigraphic sections of the Alicante region. (A.) Approximately from north (left) to south (right) along an eastern traverse. (B.) From northwest (left) to southeast (right) along a western traverse.
Zone P7 or Zone P8 depending on the range of these species (Toumarkine and Luterbacher, 1985; Berggren and Miller, 1988).

**P8**: the assignment of samples to the *Morozovella aragonensis* Zone (Pantanet) is based on the presence of *Morozovella aragonensis*, *Acarinina pentacamerata*, *A. soldadoensis*, *A. broedermanni*, and *Globigerina inaequispira*, and the absence of *Planorotalites palmerae* in rich samples.

**P8/9**: a relatively poor sample from the Puerto de Benifallim yielded *Acarinina soldadoensis* and *Morozovella caucasica*, indicating either the top of P8 or P9.

**P9**: marls are assigned a latest Early Eocene age in the Carrasqueta and Torremanzanas sections based on the presence of *Planorotalites palmerae* associated with e.g., *Acarinina pentacamerata*, *A. soldadoensis*, and *Morozovella caucasica* or, Cirer and Ibi sections, on the assemblage of *Acarinina bul-
brooki, A. pentacamerata, A. soldadoensis, Subbotina frontosa, Morozovella caucasica.

P9/10: in the Pantanet section, the marls directly below the Late Eocene turbidites contain e.g., Morozovella aragonensis, Acarinina pentacamerata, A. bullbrooki, Subbotina frontosa, and Globigerina semini. In the absence of Acarinina soldadoensis and Hankenina, this assemblage corresponds to either the upper P9 or the lower P10 Zone.

P10/11: lower Middle Eocene age assignments are based on the association of e.g., Morozovella aragonensis, Hankenina spp.; Globigerinatheka subconglobata, and Acarinina brodermanni (Benifallim, Torremanzanas).

P11: southeast of Torremanzanas (not shown on the column of Fig. 3, but indicated on the chart of Fig. 5B), marls and turbidites could be assigned more specifically to P11 given the concurrent ranges of Morozovella aragonensis and M. lehneri.

P12: assignment to the Morozovella lehneri Zone is based on the disappearance of Morozovella aragonensis and the presence of Morozovella lehneri, M. spinulosa, Globigerapsis kugleri, Globigerinatheka index, and Hankenina dumblei (Benifallim, Torremanzanas).

P13: though the nominate taxon (Orbulinoides beckmanni) has not been found in our study area, samples from Torremanzanas could be assigned to this Zone given the concurrent ranges of Globigerapsis kugleri and Hankenina alabamensis.

P13/14: late Middle Eocene age assignments in Relleu are based on the presence of Morozovella spp., and Acarinina spp., associated with Globigerina corpulenta and Turborotalia cerroazulensis pomerolii.

P14: latest Middle Eocene ages (Benifallim and Torremanzanas) are deduced from the assemblage of Morozovella spinulosa, M. lehneri, Acarinina spp., and Truncorotaloides rohri, with either Turborotalia cerroazulensis or Subbotina gortanii.

P15: early Late Eocene age assignments are based on the disappearance of Morozovella and Acarinina species and on either the association of Globigerinatheka index, Gltka. barri, Gltka. subconglobata laterbacheri, and Turborotalia cerroazulensis cocciens (Torremanzanas) or on the presence of Globigerinatheka seminimoluta (Benifallim and north of Relleu).

P16/17: late Late Eocene ages are deduced from either the assemblage Turborotalia cerroazulensis, Hankenina alabamensis, Globigerina ampliapertura, and G. pseudoamlapertura (Cirer, Torremanzanas), or on the association of Turborotalia cerroazulensis cocciens, Hankenina alabamensis, and Cribrohankenina sp. (Pantanet).

Dating of the platform carbonates was first based on identification, to the species level, of larger foraminifera (Nummulites, Assilina, Alveolina) from the Lower Eocene (Onil) and to the genus level in thin sections from the Middle-Upper Eocene deposits (Ibi, Carrasqueta, Penaguila, Aitana), where loose foraminifera were not available. The loose foraminifera enabled us to differentiate several levels within the Lower Eocene:

Lower Cuisian: characterized by Nummulites burdigalensis, N. pernotus, N. praelucasi, Assilina placentula, and Alveolina oblonga. The lower Cuisian corresponds to nannoplankton Zone NP12 (Kapelllos and Schaub, 1973) which can be correlated with planktonic foraminiferal Zones P7-lower P8 (Bolli et al., 1985).

Upper lower Cuisian: indicated by Nummulites nitidus and N. planulatus. The upper lower Cuisian can be correlated with the boundary between nannoplankton Zones NP12 and NP13. This boundary falls within planktonic foraminiferal Zone P8 (Bolli et al., 1985).

Lower middle Cuisian: containing Nummulites campesinus, N. burdigalensis, N. praelucasi, and N. partachi. The lower middle Cuisian corresponds to the lower part of nannoplankton Zone NP13 or the upper part of planktonic foraminiferal Zone P8 (Bolli et al., 1985).

Upper middle Cuisian: indicated by the presence of Nummulites burdigalensis, N. partachi, and N. tenuilamellatus. The upper middle Cuisian can be correlated with the upper part of nannoplankton Zone NP13 corresponding to the boundary of planktonic foraminiferal Zones P8 and P9 (Bolli et al., 1985).

Upper Cuisian: characterized by Nummulites friulanus, N. campesinus, and N. rotularius. The upper Cuisian corresponds to the lower part of nannoplankton Zone NP14 or planktonic foraminiferal Zone P9 (Bolli et al., 1985).

Uppermost Cuisian-lowermost Lutetian: with Nummulites praeolorioli, and N. sp. aff. N. gallensis. The uppermost Cuisian-lowermost Lutetian corresponds to the middle and upper part of nannoplankton Zone NP14, and thus to planktonic foraminiferal Zone P10 (Bolli et al., 1985).

In thin section only a subdivision into Middle and Late Eocene could be made based on the disappearance of Alveolina and Assilina and the appearance of Heterostegina, Champanina, Baculogypsinoidea and Neovalveolina at this boundary.

The next step used to differentiate the platform carbonates was to correlate them with the slope deposits of Torremanzanas. Eocene marls, turbidites, and mass flows are imbricated along a major thrust fault southwest of this village. Though no undisturbed section was found, it was possible to restore the original superposition from the Lower-Middle Eocene boundary upward (Fig. 3B).

The clasts in the mass flows of Torremanzanas, being the erosional products of the platform, record the nature of the platform build up during the preceding highstand. Near Torremanzanas, only upper Middle (P14) and Upper Eocene (P15, P16/17) mass flows could be detected, because earlier erosional products were transported to the southwest outside the study area (see below). Consequently, the analysis of the mass flows of Torremanzanas could only be used to differentiate the late Middle and Late Eocene platform cycles.

Thin-section study of turbidites on the slope near Relleu and in the Pantanet revealed that they contain a high amount of...
loose platform-interior grains (Everts, 1991) and showed that a distinction could be made between several lowstand phases (characterized by erosional products of an older platform) alternating with transgressive and/or highstand phases (distinguished by redeposited coeval shelf- and platform-interior grains).

RESULTS

Paleogeography

Following a Paleocene block-faulting phase at 60 Ma (De Ruig et al., 1991), the Early Eocene (Fig. 4A) was characterized by onlap and filling of the newly formed swells and troughs. Tectonic activity was waning but it still affected the facies distribution. The main sediment type, on both shelf and upper slope, was a brownish green marly clay. Near Onil, however, a carbonate platform started to build on the downthrown part of the northwest-southeast tilted block of Ibi-Onil-Carrasqueta. Near the end of the Early Eocene, the updip part of the block (Carrasqueta) became submerged, but the central high between Jijona-Torremanzanas and Aguas de Busot remained emergent. At the same time, quartz detritus spread over the shelf onto the upper slope.

In the early Middle Eocene (P10–11, Fig. 4B) the Onil platform expanded to include most of the tilted block of Onil-Ibi-Carrasqueta. Onlap progressed over the central highs, but condensed sedimentation continued near Rellue. The end of the early Middle Eocene (P12) shows a stratigraphic turning point: deep erosion (up to 120 m) and the development of a quartz-rich prograding shelf fan near Ibi, the toes of which reached the subsiding southwestern part of the Carrasqueta. Until this time, the erosional products of the Onil-Ibi platform were discharged to the area of Agost, some 15 km southwest of our study area (not shown on the maps of Fig. 4). After this turning point, these products were transported to the southeast, building a fan system between Jijona and Torremanzanas, and probably separated from the Carrasqueta by a growth fault. This paleogeographic reorientation was the result of a second major block-faulting event between about 44 and 42 Ma (P12).

During the late Middle Eocene (P13–14, Fig. 4C), a carbonate platform started to build on the shelf in the north, covering the area between Alcoy, Penaguila and the north Aitana. The carbonate platform in the northwest (Onil-Ibi-Carrasqueta) continued to exist. In the area of Rellue, normal sedimentation of green marls began after a long period of condensed sedimentation. At the end of the Middle Eocene, a third tectonic phase caused tilting of blocks towards the southwest, resulting in erosion of the updip parts. However, overall paleogeography did not change.

The early Late Eocene (P15, Fig. 4D) was a period of quiescence. A carbonate platform covered large parts of the shelf in the northwest (Ibi-Carrasqueta) and north (Penaguila-Aitana), overlapping former block boundaries and prograding over the slope. Isolated turbidite fans evolved in the mid-slope area.

In the late Late Eocene (P16/17, Fig. 4E), the carbonate platform and the slope broke up during a fourth, more severe block-faulting event. The platforms of Onil-Ibi, the northeastern part of the Carrasqueta, and the north-Aitana were uplifted, whereas the southwestern part of the Carrasqueta, the Penaguila, and the south-Aitana subsided. The incident was more compressional than tensional, given the development of cleavage near Penaguila and the expulsion somewhere of Triassic gypsum as evidenced by the influx of red idiomorphic quartz, a common accessory of Triassic gypsum, in the coeval mass flows of Torremanzanas. Near Rellue, giant flute casts are found at the base of the upper Upper Eocene midslope turbidite fan, indicating abnormally strong down-slope currents. Mass flows and turbidites reached further downslope, to the site of the Pantanet, and cut deeply into older strata because Middle Eocene sediments are largely removed.

After this event, carbonate-platform sedimentation in the study area was limited to Ibi-Carrasqueta and the north-Aitana. However, notwithstanding the severity of the tectonic impact, the overall facies distribution of continental, lagoonal, or shallow-marine facies in the northwest and north, and deeper marine in the southeast and south, was not disturbed and continued to prevail during the Oligocene.

Sequence Stratigraphy

Early Eocene (P6b–P9).—Lowermost Lower to upper Lower Eocene sections are found only on the Onil-Ibi platform in the northwest and in the Pantanet section, downslope, in the southeast. In the Carrasqueta, near Torremanzanas, to the southwest of Penaguila (Puerto de Benifallim), and between Penaguila and Rellue, Eocene sedimentation did not start until the late Early Eocene (P8/9 or P9) (Figs. 3, 5A, 5B).

On the Onil-Ibi platform, four Lower Eocene sequences have been distinguished, preceded by a less clear one of earliest Eocene (P6b-P7/8) age. The first clear cycle starts with limestones containing an open-shelf fauna of Assilina and Discocyclina (TST) lying upon limestones with a more platform-interior signature (Alveolina-limestones). After a marly interval (maximum flooding), the cycle is completed by a HST of thickening-up limestones with, again, a more restricted fauna, and covered by a karstic surface. The next two cycles both contain an open-shelf fauna at the base and a more restricted fauna at the top. They, too, are separated by a karstic surface. Since they show only a thickening-up bedding pattern and a shallowing-up microfacies, they are thought to represent highstand phases. The karstic surfaces, thus sequence boundaries, are within the range of P8. The fourth cycle starts with a marly interval (maximum flooding), succeeded by a HST composed of a thickening- and shallowing-up set of limestone beds with an open-shelf fauna at the base and a platform-interior faunal content at the top and, upward-increasing amount of quartz. This fourth cycle is late Early Eocene (P9) in age.

In the Pantanet section (basin floor to toe of slope), the Lower Eocene deposits are represented by grey marls containing an interval of thickening-up pelagic limestones (in P7) and a red marly interval (in P8). The limestone interval is thought to be the equivalent of the prograding HST of the lowermost Onil-Ibi cycle. The red interval corresponds to the maximum flooding of the second or third Onil-Ibi cycle and may be interpreted as the condensed basinal counterpart (sensu Loutit et al., 1988) of one of these.

The upper Lower Eocene (P9) quartzose cycle of Onil-Ibi corresponds to coeval glauconite- and quartz-rich calcarenites and green marls on the upper slope (near Torremanzanas and
between Penaguila and RellEU) and to green marls on the basin floor to toe of slope (in the Pantanet). Apart from the quartz, the calcarenites contain a high proportion of loose platform-interior biota. They are considered to be the result of late-highstand shedding.

**Early Middle Eocene (P9/10–P11).**—For this time, only the Onil-Ibi platform carbonates are available for sequence stratigraphical analysis, but rather poorly so, because only one of the cycles of Onil could be dated (P10; see Methods and dating). The slope deposits of this time are represented by monotonous whitish and green marls, containing locally some turbiditic intercalations. They are onlapping on and over former highs.

In the Onil-Ibi platform sections, four cycles have been distinguished. The first cycle, at the Early-Middle Eocene boundary, starts with transgressive marls (maximum flooding) and is completed by thickening- and shallowing-up Assilina-Disco-cyclina limestones containing a restricted fauna in the top (HST). The ensuing three cycles can be studied best in the Ibi section. They show an overall upward increase in dolomitic interbeds and an increasingly restricted fauna content. The first cycle displays some low-angle crossbedded limestones with an open-shelf fauna (Assilina, Discocyclina, some Alveolina) (TST) at the base, succeeded by massive limestones, containing more Alveolina and no Assilina, and showing dolomitization at the top (HST). The next sequence consists of limestones and dolomitic limestones with an open-shelf fauna (Assilina) overlain by dolomites with thinner intercalations of limestones containing exclusively platform-interior biota (HST).

As stated before, only one of the above cycles could be dated, that is the third cycle above the quartz-rich P9-cycle which correlates to P10.

**Late Middle Eocene (P12–P14).**—After the mid-Middle Eocene paleogeographic change between 44 and 42 Ma, platform deposits developed only on the Ibi-Onil-Carrasqueta platform in the northwest, but also on the Penaguila-Aitana platform in the north (Fig. 4C). Coeval upper slope deposits (marls, limestones, turbidites and mass flows) are encountered near Torremanzanas and to the south of Penaguila. On the midslope and toe of slope (near RellEU and east of the Pantanet, respectively), the upper Middle Eocene deposits are represented by greenish marls. Large parts were then removed by subsequent submarine erosion, especially downslope. As a consequence of the yet more widespread occurrence of upper Middle Eocene deposits, as compared with those of the earlier time slices, sequences could be correlated over a larger part of the study area (Figs. 3, 5A, 5B).

Two complete sequences, comprising lowstand wedges, transgressive systems tracts, and highstand systems tracts could be recognized on the platform and upper slope. The base of the first sequence is characterized by deep scouring (up to 120 m in depth) in preceding platform-interior carbonates of the Ibi platform, and by erosion of Carrasqueta limestones. The erosional gaps are filled in by a fan system prograding far over the Onil-Ibi-Carrasqueta platform. The fan is made up of calcarenites containing an upward-increasing amount of quartz and Cretaceous pebbles, and is covered in Ibi by fossil soil (Micro-codium). On the upper slope, between Torremanzanas and Penaguila, brown quartz-rich, very thin turbidites and marls (P12) are found. The prograding fan system and corresponding upper slope turbidites were initially interpreted as a platform-margin lowstand wedge indicating a type II sequence boundary. However, the deep erosion on the platform is at variance with a minor relative sea-level drop. Mid-Middle Eocene block-faulting probably caused temporarily strong local uplift and erosion, after which the platform again subsided. Consequently, though the prograding fan is deposited on the platform and should be denominated a “platform-margin wedge”, it cannot be taken to indicate a type II sequence boundary. An ensuing TST is represented in Ibi by thin platy miliolid-limestones, in the Carrasqueta by thinning and fining-up limestone beds topped by marls, and on the upper slope of Torremanzanas by a grey marly interval. In Ibi the sequence is concluded by thick-bedded dolomites followed by pebbly calcarenites with a restricted microfauna (HST). On the Carrasqueta, the HST is formed by a thickening-up and coarsening-up series of calcarenites containing an open-shelf larger-foraminiferal fauna and coral fragments. This occurrence suggests the existence of coral reefs between the restricted platform of Ibi and the open-shelf environment of the Carrasqueta. On the upper slope, thin limestone intercalations (P13) are thought to be the correlative HST.

The second sequence starts with a P14-lowstand wedge on the upper slope (near Torremanzanas and to the south of Penaguila), built of mass flows and turbidites. In the mass flows the co-occurrence of clasts of coral reef, platform interior, and open shelf points to the erosion of a preceding rimmed platform and the withdrawal of the sea beyond the platform margin (type I sequence boundary). Renewed platform buildup (deepening, TST), followed by prograding (shallowing, HST), can be seen on both the Ibi-Carrasqueta and the Penaguila platforms: thin beds with a less restricted microfauna are overlain by pebbly calcarenites of very restricted signature (Ibi); thinning-up and fining-up calcarenites and marls are succeeded by a coarsening-and thickening-up series of open-shelf calcarenites (Carrasqueta); backstepping deeper-shelf limestones and marls suddenly are capped by shallow open-shelf limestones (Penaguila). There is no evidence for reef building during this time (P14). After the completion of the P14-sequence, a block-faulting phase induced breakup of the Ibi-Carrasqueta and Penaguila platforms.

**Early Late Eocene (P15).**—Platform deposits of this age are widespread in the northwest and north parts of the study area. Upper-slope sediments are found near Torremanzanas and to the south of the Penaguila-Aitana platform; patchy turbidite fans cover the midslope area near RellEU and to the west of RellEU. All P15-sediments are thought to represent one single sequence comprising lowstand, transgressive and highstand deposits (Figs. 3, 5A, 5B).

Lowstand mass flows and turbidites containing clasts derived from a deep carbonate shelf and/or upper slope are found near Torremanzanas. Notwithstanding the preceding blockfaulting and erosional phase, platform-interior clasts are not found on the slope. The erosional products were evidently trapped high on the platform and later, during transgression, incorporated into onlapping beds. Between Penaguila and RellEU, silty fine-
grained blue nodular limestones are found, downlapping southward onto Lower Eocene sediments. They are thought to represent a late stage of lowstand.

The arrival of some turbidites with a large amount of loose open-shelf larger foraminifera on the upper slope suggests renewed platform buildup. On the platforms, the TST is represented in seaward sections by massive limestones with an open-shelf foraminiferal content and a large quantity of Cretaceous pebbles (Penaguila) or by fine-grained calcarenites resting with an angular unconformity on Middle Eocene limestones (Carrasqueta). In the most landward section of Ibi, the TST is typified by massive pebbly calcarenites with a mixed restricted/open-shelf fauna.

The highstand phase on the platforms shows a more uniform picture: after maximum flooding, coral reefs (in the Carrasqueta and near Penaguila) separated a restricted platform interior (Ibi, Alcoy) from an open shelf (southwest of Penaguila). High productivity forced the platform-interior sediments to prograde seaward, given the reef debris with a high incidence of restricted fauna (miliolids) on top of both the Carrasqueta and Penaguila reefs.

Lack of accommodation space due to high productivity on an extensive platform eventually led to progradation over the shelf edge onto the upper and middle slope as is suggested by turbidite fan lobes containing a high amount of loose platform-interior grains near Torremanzanas and Relleu.

**Late Late Eocene (P16/17).—**The dispersal of upper Upper Eocene facies is governed by strong synsedimentary tectonics. Shallow platform carbonates are limited to Ibi, the north Carrasqueta, and the northern part of the Aitana, whereas deep-shelf deposits occur in the south Carrasqueta and near Penaguila. Slope sediments deposited during this time are widespread: upper slope deposits occur between Torremanzanas and Penaguila, midslope near Relleu, and toe of slope in the Pantanet. In spite of platform break-up and later Oligocene and/or Miocene erosion, it still is possible to recognize the presence of two sequences (Figs. 3, 5A, 5B).

The first sequence is the most widely preserved and recognized and comprises lowstand, transgressive, and highstand deposits. On the upper slope of Torremanzanas the cycle begins with lowstand mass flows and turbidites containing limeclasts of a variety of facies (platform interior, shallow open shelf, deep shelf) and red Triassic quartz. Near Relleu, thick turbiditic beds containing platform-interior clasts and reworked Middle Eocene larger foraminifera rest with an erosive contact marked by gigantic flute casts upon lower Upper Eocene turbidites. They are thought to represent the correlative midslope lowstand wedge.

The transgressive systems tract is represented on the subsiding block of the Carrasqueta by either a basal conglomerate overlain by a fining-up and thinning-up calcarenite-marl alternation or by marls. On the less subsiding block of Ibi, massive

![Fig. 5.—A. Chronostratigraphic chart, showing correlation of Eocene cycles from north (left) to south (right) along an eastern traverse in the Alicante region. Interpretation of the sections of Figure 3A. The two uppermost cycles correspond both to P16/P17; they may have occurred both during P16 or during P17, or one can be assigned to P16 and one to P17.](image-url)
calcarenites with a less restricted signature overlie the more restricted limestones of the preceding sequence, truncating the top of the latter. In the north, near Penaguila, where Oligo/Miocene erosion is more severe, remnants of Upper Eocene deep-shelf deposits are found. They are separated from the preceding lower Upper Eocene shallow carbonates by an angular unconformity.

The highstand deposits of the northwestern (Ibi-Carrasqueta) platform on the landward side (Ibi) comprise calcarenites with a restricted fauna topped by a karstic surface and an alternation of marls and coarse calcarenites on the subsiding Carrasqueta. The latter contain an open-shelf fauna and coral reef debris. The coral debris suggests the existence of a rimmed platform during a highstand to the northwest of the Carrasqueta. Equivalent highstand deposits are not known from the northern (Penaguila) platform. However, on a small downthrown block south of Penaguila (Puerto de Benifallim), a remnant of the shelf-slope break is preserved on which an upper Upper Eocene prograding series of pelagic limestones can be seen on top of a thinning and fining-up package of limestones. Progradation beyond the shelf edge during highstand is far less than during the early Late Eocene: only thin packages of coarse calcarenites are found in the platform section of Ibi. They lie on top of a karstic surface and contain a mixed restricted/open-shelf fauna and some coral fragments. This thin deposit is interpreted as a remnant of a transgressive surface. On the still subsiding block of the Carrasqueta this sequence is represented by a thinning and fining-up series of calcarenites and marls with an open-shelf fauna (TST), a marly interval (maximum flooding), and a series of fine calcarenites and marls devoid of larger foraminifera (HST). The upper slope deposits of Torremanzanas show a rather erratic but overall thinning-up bedding pattern in very fine-grained turbidites and an upward increase of marly intervals. Whether these deposits represent lowstand, transgressive or highstand phases is not altogether clear. The fine-grained nature of these deposits may reflect lack of production of coarse, erosional detritus and the absence of a productive shallow larger-foram shelf in their source area. The first may be due to rapid lithification of the remnants of the shallow shelf near Ibi; the second to rapid subsidence of the outer platforms of the Carrasqueta and Penaguila. Another mechanism to explain the fine-grained texture and overall fining-up near Torremanzanas may be that the source area was cut off tectonically which led to bypass of coarse detritus and larger foraminifera to other depocentres.

The second upper Upper Eocene sequence is far from complete on the platform, but is widespread on the upper, middle and lowermost slope. Only some decimeters of coarse calcarenites are found in the platform section of Ibi. They lie on top of a karstic surface and contain a mixed restricted/open-shelf fauna and some coral fragments. This thin deposit is interpreted as a remnant of a transgressive surface. On the still subsiding block of the Carrasqueta this sequence is represented by a thinning and fining-up series of calcarenites and marls with an open-shelf fauna (TST), a marly interval (maximum flooding), and a series of fine calcarenites and marls devoid of larger foraminifera (HST). The upper slope deposits of Torremanzanas show a rather erratic but overall thinning-up bedding pattern in very fine-grained turbidites and an upward increase of marly intervals. Whether these deposits represent lowstand, transgressive or highstand phases is not altogether clear. The fine-grained nature of these deposits may reflect lack of production of coarse, erosional detritus and the absence of a productive shallow larger-foram shelf in their source area. The first may be due to rapid lithification of the remnants of the shallow shelf near Ibi; the second to rapid subsidence of the outer platforms of the Carrasqueta and Penaguila. Another mechanism to explain the fine-grained texture and overall fining-up near Torremanzanas may be that the source area was cut off tectonically which led to bypass of coarse detritus and larger foraminifera to other depocentres.

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**Fig. 5.**—B. Chronostratigraphic chart, showing correlation of Eocene cycles from northwest (left) to southeast (right) along a western traverse in the Alicante region. Interpretation of the sections of Figure 3B (for legend see Figure 5A). The positions of the second (P7/8), sixth (P9/10) and seventh (P10) cycle on the chart are not directly age controlled, but indirectly by their occurrence between dated cycles (for the P16/P17 cycles, see text of Figure 5A).
The midslope (Relleu) and toe-of-slope (Pantanet) turbidites, on the other hand, are coarse grained. Lowstand deposition is indicated in both Relleu and in the Pantanet by scouring and deposition of thick beds containing reworked material (even coral fragments, found somewhat to the northwest of the Pantanet section). The commencement of platform build-up (in the Aitana?) is reflected by turbidites with a high amount of free open-shelf foraminifera. After a marly interval, the sequence is concluded by turbidites containing a large quantity of loose platform-interior fossils.

CONCLUDING REMARKS

On fig. 6 the south-Spanish platform sequences are tentatively compared with the Haq et al. (1988) global chart. In the Alicante area, five cycles and sequence boundaries could be recognized in the Lower Eocene deposits, that is, one sequence boundary between 52.5 and about 51 Ma (P7-lower P8), three between ca. 51 and 50 Ma (P8), and one between ca. 50 and 49 Ma (P9). The Haq et al. (1988) chart shows five sequence boundaries, at 52, 51.5, 50.5, 50, and 49.5 Ma in this interval. Although our findings do not allow a precise dating of the Alicante boundaries, they are not at variance with the global chart.

In the Middle Eocene Alicante deposits, six cycles and sequence boundaries have been distinguished: two sequence boundaries between 49 and 46 Ma (P 10) above the one at the P9/P10 boundary, one between ca. 44 and 42 Ma (P12), one between 42 and 41 Ma (about boundary P13/P14), and one between 40 and 39 Ma (P14/P15 boundary). Again, our data are not exact enough to confirm or to challenge the ages on the Haq et al. (1988) chart. However, it may be remarked that though the global chart also shows six sequence boundaries in the Middle Eocene, at 48.5, 46.5, 44, 42.5, 40.5 and 39.5 Ma respectively, our data suggest the existence in Alicante of two cycles in P10 (above the one at the P9/P10 boundary) and only one in P12.

The Late Eocene deposits of Alicante can be subdivided into three cycles with three sequence boundaries between about 38 and 36 Ma (between P15 and P16/P17, within P16/P17, and at the Eocene/Oligocene boundary, respectively). The Haq et al. (1988) chart shows three sequences with boundaries at 38, 37, and 36 Ma for the Late Eocene.

The chronostratigraphic charts and columns (Figs. 3, 5A, 5B) further reveal a) that at least three Alicante sequence boundaries are coincident with tectonic events (between 44 and 42 Ma, between 40 and 39 Ma, and between 38 and 36 Ma, respectively); b) that due to tectonics, most sections contain only part of the total sequences, with the Ibi section the relatively most complete one; and c) where sequences could be correlated, their thicknesses are highly variable between sections (e.g., the P15-
sequence is about 130 meters in thickness in Penaguila, some 40 meters in Ibi, and 20 meters in the Carrasqueta), reflecting differential subsidence between discrete tectonic events.

**DISCUSSION**

The detailed analysis of stratigraphic data from Eocene deposits in southeastern Spain described in this paper demonstrates a) the presence of 14 3rd-order sequences bounded by unconformities and b) the strong influence of tectonics (differential subsidence and block-faulting) during sedimentation. The problem of the causes of sea-level changes is the subject of a major controversy. Short-term cycles in sea-level change have been generally interpreted in terms of glacio-eustasy (Vail et al., 1977) the last fifteen years, partly due to the lack of a tectonic mechanism. Meanwhile it has been shown that short-term changes in relative sea level can equally well be caused by rapid, stress-induced vertical movements of the lithosphere (Cloetingh et al., 1985; Kooi, 1991). Undoubtedly, both eustasy and tectonics have contributed to the record of short-term changes in sea level. Therefore, careful analysis is required to separate effects of eustatic sea-level changes from stress-induced short-term motions of the lithosphere, as both mechanisms produce rather similar short-term distortions from long-term patterns of subsidence (Cloetingh, 1991). Embry (1989) developed a list of stratigraphic criteria to differentiate between tectonics and eustasy. However, recognition of one or more of his criteria merely proves the presence of a tectonic mechanism. The simple fact that tectonic mechanisms operate during a certain time does not prove that glacio-eustasy does not exist. To separate the relative contribution of tectonics and eustasy in the generation of the southeast Spanish 3rd-order sequences, we will first survey the tectonic context of southeastern Spain and next the possibility of glacio-eustatic changes during the Eocene.

**Plate-Tectonic Context.**—Iberia was part of Africa from the Late Cretaceous until about 42 Ma. The main boundary between Africa (including Iberia) and Eurasia was situated up to that time in the Bay of Biscay. Africa (Iberia) started to move northward at about 60 Ma, which resulted in strong compressive deformation across the Pyrenees, culminating in the Middle Eocene. This climax resulted from initiation of spreading in the Norwegian-Greenland Sea around 55 Ma, which caused an additional northwest-southeast compressive component in the African-Eurasian collision (review in De Jong, 1991). At the same time, a short-lived change occurred in the motion of Africa relative to Eurasia: from a smooth path to the northeast up to 55 Ma, a small kink in the direction occurred between 55 Ma and 46 Ma, after which the strike-slip movement along the plate boundary in the Bay of Biscay changed from dextral to sinistral (Srivastava et al., 1990). Iberia started to move as an independent plate, thus caught between two massive plates, Eurasia to the north and Africa to the south, after reactivation of the Azores-Gibraltar-Fracture-Zone at about 42 Ma, until it became part of the Eurasian plate in the early Miocene (24 Ma). At the Middle-Late Eocene boundary (39.5 Ma) the Eurasian-Iberian plate boundary jumped to a more southern position in the Bay of Biscay (Srivastava et al., 1990).

In the Pyrenees, the African-Eurasian collision created distinct tectonic phases in the Paleocene and Eocene. The Paleocene, middle Eocene and latest Eocene tectorogenic phases in particular seem to have been major West-European-wide events (Plaziat, 1981; Schwan, 1985; Ziegler, 1990). Since stress orientation data indicate the possibility of propagation of stresses away from the plate boundaries over large distances into the plate interiors, where they affect the vertical motions within sedimentary basins (Cloetingh, 1991), a comparison with Pyrenean tectonic phases seems warranted. Accordingly, uplift and slope-faulting in the External Zone of southern Spain around 60 Ma can be explained by far-field transmission of compressive stress during the onset of Pyrenean collision (Kenter et al., 1990; De Ruig et al., 1991).

On Fig. 6, data on Eocene plate-tectonic reorganizations (after Srivastava et al., 1990) and Pyrenean tectono-sedimentary cycles (after Puigdefàbregas and Souquet, 1986) are given, together with the Eocene Alicante cycles and the global cycle chart of Haq et al. (1988). In the Early Eocene (54-49 Ma) continuing compression in the Pyrenees is evident and three tectono-sedimentary cycles have been recognized (TE2-TE4). Correlation of TE2 and TE3 with the Alicante cycles is not altogether clear, but the P9 (50-49 Ma) cycle of Alicante is coincident with cycle TE4 of the Pyrenees, which is thought to be controlled by both sea level fall and nappe emplacement. During the Lutetian (49-42 Ma) strong tectonics (deep thrusting and basin-centre displacement) control the Pyrenean cycle TE5. They are evidently induced by the 46 Ma change in strike-slip from dextral to sinistral in the Bay of Biscay. At this time three sequence boundaries have been recognized in the Alicante area. In both the Pyrenees and in Alicante the unconformities have not been dated with more precision than “Lutetian” in the Pyrenees or early Middle Eocene (P9/10, P10, ?P11) in Alicante. Near the Lutetian-Bartonian boundary (about 42 Ma), the tectonic regime changed in the Pyrenees (TE5-TE6 transition) at the time when Iberia began to move as an independent plate. In the Alicante area, this event is coincident with a tectonically induced paleogeographic reorganization and a sequence boundary between 44 and 42 Ma (intra P12). During the Pyrenean TE6 cycle (late Middle and Late Eocene; Bartonian-Priabonian; 42-36 Ma), the Iberian-Eurasian plate boundary jumps to a more southern position at 39.5 Ma (the Middle-Late Eocene boundary). At the same time, the intraplate stress direction in the European plate and in Iberia rotated from northwest-southeast to northnortheast-southsouthwest (Lepvrier and Martinez-Garcia, 1990; Le Pichon et al., 1988). Strong continuing compression culminated in the major intra-Priabonian Pyrenean phase (38 Ma). Both the 39.5 Ma and the 38 Ma events are expressed in the southeast Spanish record by sequence boundaries and distinct block-faulting phases between 40 and 39 Ma and between 38 and 36 Ma, of which that in the late Late Eocene is the most severe. At the end of the Pyrenean Te6 cycle (boundary Eocene/Oligocene; 36 Ma) extensive evaporites indicate the final retreat of the sea from the south Pyrenean basin, as was also recorded in the Aquitanian basin. However, compression across the Pyrenees continued until earliest Miocene times, when Iberia became part of Eurasia (Srivastava et al., 1990). In the Alicante area, the Eocene/Oligocene boundary is also characterized by a major regressive phase (topmost sequence boundary).

From the above review of events it may be concluded that most, if not all sequences in southeastern Spain might well be
generated by buildup of compressional stresses during collision, interrupted by phases of stress relaxation associated with plate reorganizations. 

Eocene Glacio-Eustasy.—Glacio-eustasy requires global control of waxing and waning ice sheets. Prior to the initiation of Northern Hemisphere glaciations, the Antarctic ice sheets would seem to be the most plausible source for global sea-level fluctuations during the Cenozoic. Isotope studies have been used to argue that Antarctica was ice-free during the Paleocene and early Eocene, but that minor ice growth events occurred in the earliest Middle Eocene, during the Middle Eocene, and in the late Middle Eocene. Extensive ice-sheet growth would have occurred at the beginning and end of the Late Eocene (review in Barron et al., 1991). The sediment record from Antarctica yields evidence for minor and local glacial events in the lower Middle Eocene and Middle Eocene. East Antarctic ice sheets probably reached sea level and advanced partly onto the shelf for the first time at 40 and 39 Ma (Barron et al., 1991). The Late Eocene age of this event, however, remains equivocal, for Miller (1992) pointed out that available data require only an earliest Oligocene or older age. A major increase in ice growth is recorded at the Eocene/Oligocene boundary (Barron et al., 1991). According to Berggren and Prothero (1992) and Miller (1992), the oldest unequivocal evidence for the existence of large continental ice sheets on Antarctica dates from the earliest Oligocene. These finds are consistent with: a) the recordings of progressive cooling during the Middle and Late Eocene and a further drop in temperature in the earliest Oligocene in the Southern Oceans (Kennett and Barker, 1990; Diester-Haas, 1991); b) faunal changes indicating cooling episodes throughout the Middle Eocene to early Oligocene in the Atlantic, Pacific and Indian Oceans (Keller, 1983; Hallock et al., 1991), with a major biotic turnover at the Middle/Late Eocene boundary (Boersma et al., 1987; Aubry, 1992; Keller et al., 1992), and c) the occurrence of a global cooling event formerly thought to occur exactly at the Eocene/Oligocene boundary (Cavelier et al., 1981; Plaziat, 1981; Vianey-Liaud, 1991) but now believed to be earliest Oligocene in age (Berggren and Prothero, 1992). Consequently, significant global control on sea level by waxing and waning ice sheets may have started during the Late Eocene with a first peak in the earliest Oligocene.

In the southeastern-Spanish relative sea-level records, the major cooling event and possible introduction of glacio-eustasy at the Middle/Late Eocene boundary coincides with a sequence boundary also controlled by a block-faulting event (Fig. 6). The columns of Fig. 3 reveal that this is a boundary that can be traced over the entire study area. The ensuing widespread transgressive event is also striking. It is a matter of taste to term this a "tectonically enhanced" or "eustatically enhanced" unconformity. The sequence boundary at the top of the Eocene is less clearly tectonically controlled and may be induced entirely by a glacio-eustatic fall (Fig. 6).

The coincidence of tectonics (plate reorganization) and glaciation around the Middle/Late Eocene boundary is not a casual one. It is the time that the Eurasian-Iberian plate boundary jumped to a more southern position, but also the time of the separation of the Antarctic and Australian landmass, thus opening the Antarctic-Antarctic Basin. It has been argued that opening of the latter basin led to increased precipitation on the cold Antarctic continent and to the growth of an extensive ice sheet during the Late Eocene (Bartek in Barron et al., 1991; Bartek et al., 1992).

The events at the Middle/Late Eocene and Eocene/Oligocene boundaries had a major impact on the larger foraminiferal shelf fauna. During the Late Paleocene to Middle Eocene the fauna is rather conservative at the generic level, but across the Middle/Late Eocene boundary genera such as Assilina and Alveolina disappear to be replaced by Pellatispira, Spiroclypeus, Heterostegina and Neodolevalina (Beckmann et al., 1981; Hallock et al., 1991). The same holds for the Eocene/Oligocene boundary, where discocyclinids are replaced by lepidocyclinids, and plilar Neummulites and Pellatispira became extinct (e.g., Beckmann et al., 1981; Cavelier et al., 1981; Poignant and Lorenz, 1985; Barbin, 1988). Evidently, contrauction and isolation of populations during significant lowering of sea level, together with strong ocean circulation and mixing, producing more nutrient-rich surface waters, enhanced the extinction of old, highly specialized taxa and the generation of new species (cf., Vermeij, 1980; Hallam, 1985; Hallock et al., 1991). The new species spread in an ensuing expansion phase during sea level rise (cf. Hallam, 1985; Cubaynes et al., 1990).

CONCLUSIONS

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EOCENE TECTONO-SEDIMENTARY PATTERNS IN THE ALICANTE REGION (SOUTHEASTERN SPAIN)


ABSTRACT: Eight depositional sequences are recognized within marine Paleogene strata of the Tremp Basin (Southern Pyrenean Foreland Basin). Geometry and distribution of the depositional systems are mainly controlled by tectonics and relative sea-level changes. Carbonate systems dominate during sea-level highstands, siliciclastic systems during sea-level lowstands.

Biostratigraphic boundaries inherently tend to correspond to sequence stratigraphic surfaces (condensed sections, intervals of non-deposition). Biostratigraphic correlations provide only a rather loose framework and should not be overstretched. In marine Paleogene units of the Southern Pyrenees, their resolution is generally not sufficient to identify individual global eustatic 3rd order cycles without ambiguities.

INTRODUCTION

A sequence stratigraphic analysis of marine Paleogene strata of the Tremp Basin in the central southern Pyrenees is based on a combination of facies geology, biostratigraphy and paleoecology. This paper is a brief summary, for more details reference is made to the papers by Betzler (1988), Eichenseer (1988), Eichenseer and Luterbacher (1992), Luterbacher et al. (1991) and Van den Hurk (1990). Our group has mainly concentrated on the shallow-marine to coastal deposits of the “Ager Group” (Mutti et al., 1973), whereas other investigators have dealt in detail with sequence stratigraphic aspects including also other parts of the Paleogene succession of the central southern Pyrenees (e.g., Canudo et al., 1991; Fonnesu, 1984; Mutti et al., 1988; Puigdefábregas et al., 1986). In addition, I will address some problems related to the correlation of the eustatic cycles as recognized in the southern Pyrenees with those recognized on a global scale (Haq et al., 1987).

STRUCTURAL SETTING

The lower Paleogene strata of the southern Pyrenees have been deposited in a foreland basin which developed during early Paleogene time. To the north, it is limited by the Axial Zone and to the south by the frontal thrust of the “Sierras Marginales” (Fig. 1). During early Paleogene time, the compressive transtensive wrench basin (Puigdefábregas and Souquet, 1986) caused by the rotation and oblique collision of Iberia with the European Plate (e.g., Choukroune, 1976) changes into a foreland basin with westward to southward moving thrusts (Garrido, 1973; Eichenseer, 1988). A major part of the south-Pyrenean Foreland Basin evolved into a thrust-sheet-top basin (Ori and Friend, 1984; Fig. 2).

The evolution of the thrusts during Paleogene time leads to differential tectonic subsidence which strongly controls the distribution, thicknesses, and hydrodynamic conditions of depositional systems. Thrust activity progressed from east to west. The easternmost San Corneli Thrust (Fig. 3) was active already during Maastrichtian time (Simó, 1986; Mutti et al., 1988) or even earlier (Rosell, oral communication). The Col de Vent Thrust influenced the distribution of the depositional systems mainly during Thanetian time, the Turbón Thrust during Ilerdian time (Eichenseer, 1988). The interference of the Turbón Thrust with the Cotiella Thrust causes the high subsidence rates in the area of Campo during Ilerdian and Cuisian times (Garrido, 1973; Van den Hurk, 1990). The southern flank of the foreland basin rests on the backlimb of the frontal thrusts of the Sierras Marginales where incipient detachment started already in the latest Cretaceous. This frontal thrust became increasingly active during the Ilerdian causing a northward shift of the axis of the basin.

The interplay between these incipient thrusts and relative sea-level changes controls the geometry and distribution of the depositional systems. Carbonate systems dominate during relative sea-level highstands, clastic systems during sea level lowstands (Fig. 4). In addition, carbonate systems are widespread during prolonged sea-level rises.

A comparison of the chronostratigraphic section (Fig. 5) with the corresponding cross section (Fig. 6) suggests that thrusting is rather continuous but slow as compared to the changes which may be attributed to fluctuations of the relative sea-level. In addition, the chronostratigraphic section shows some small scale facies distribution patterns which are beyond the graphic resolution of the cross section.

During transgressive systems tracts, retreatning homoclinal carbonate ramps cover most of the basin margins. Their facies succession and geometry vary mainly as a function of the gradient of the depositional relief induced by thrusting. Generally, the ramp systems evolve from transgressive beaches during increased flooding. Beach deposits, rich in representatives of the larger foraminiferal genera Alveolina and Orbitolites, are the shallowest part of the ramps. Highly bioturbated mudstones with Nummulites and Operculina are deposited in deeper water depths. Locally, sediment starved deeper intervals with Assilina and Discocyclina may develop.

In steeper ramps, coralgal reefs may form, in which the encrusting foraminifer Solenomeris plays a major role (see also Plaziat and Perrin, 1992). These coralgal growths thicken at minor morphological breaks into large reef complexes. Reef growth reaches its maximum development at the carbonate bank margin during times of increased flooding. Nummulites bars may develop at the western bank margins, where offbank transport is less important. However, these Nummulites bars are always considerably smaller than those developed during the younger part of the Lower Eocene and the Middle Eocene succession on the platform domains of Libya and Egypt (e.g., Arni, 1965). During the early part of the Eocene Epoch, Nummulites were still of relatively small size and sufficiently large platform domains could not develop in the tectonically active foreland basin of the southern Pyrenees.

Siliciclastic depositional systems dominate during sea-level lowstands. The location of the subaerial parts (mainly incised-
valley fills) of the coarse-grained deltaic systems (braided stream deposits enclosed in red floodplain fines with calcimorphic soils) is controlled by tectonics (Fig. 4). The corresponding lower delta plain deposits are fluvial to tide-dominated distributary channels, marshes and swamps bordered by wave- and/or tide-dominated delta-front sands and silts which grade distally into prodelta shales with small *Nummulites* and *Operculina*. Muddy limestones with *Alveolina* and calcareous shales with *Operculina* are thought to be interdistributary bay and lagoon deposits. The source area of these relatively small coarse-grained deltaic systems is the tectonically uplifted hinterland to the north.

Fine-grained deltaic systems develop mainly along the main axis of the foreland basin. The subaerial portion of the fine-
SEQUENCE STRATIGRAPHY AND THE LIMITATIONS OF BIOSTRATIGRAPHY IN THE MARINE PALEOGENE STRATA

FIG. 3.—Structural map of the Tremp Basin and evolution of thrust units (from Eichenseer and Luterbacher, 1992).

Grained deltas consist of low-sinuosity river-channels filled with well-sorted sands and associated red floodplain fines. Fluvi- to tide-dominated interdistributary channels, clastic tidal flats and marshes represent the lower delta plain facies. The delta front consists mainly of estuarine sands (shoals, channels, tidal bars) which often become wave-reworked seaward. Shales with small Nummulites and Assilina are thought to represent the prodelta facies.

To the west of the Turbón Thrust Anticline, thin-bedded calcarenitic turbidites, sandy turbidites and canyon-fill sands correspond to the offshore part of the lowstand systems tract. For a more detailed discussion, reference is made to Mutti et al. (1988).

STRATIGRAPHIC SEQUENCES OF THE MARINE PALEOGENE SUCCESSION IN THE TREMP BASIN AND THEIR CORRELATION WITH GLOBAL SEA LEVEL CHANGES

Biostratigraphic control in the marine Paleogene deposits of the Tremp Basin is mainly based on larger foraminifera (Alveolina, Orbitolites, Nummulites, Assilina, Operculina, Discocyclina) and to a lesser extent also on planktonic foraminifera, calcareous nannoplankton and dinoflagellates (see e.g., Molina et al., 1992, with references). Magnetostratigraphic investigations in progress are giving encouraging results (Pascual, 1992), but a continuous succession of magnetochrons has not yet been established.

Fig. 7 shows the biostratigraphic correlations of the stratigraphic sequences of the marine Paleogene succession in the Tremp Basin with the chart of global sea-level changes (Haq et al., 1987). At the first glance, there is a surprisingly good fit with the third-order cycles of the “Global Sea-Level Chart”. However, this may be somewhat an artifact.

In the studied sections, boundaries of biozones frequently coincide with sequence boundaries, transgressive or downlap surfaces. These intervals correspond to times of reduced or non-deposition during which the evolutionary lineages of microfossils become telescoped (Fig. 8, see also Fontaine, 1984; Rey et al., 1986). It is in these levels that evolutionary changes become condensed and where first and last occurrences concentrate. The location of biostratigraphic boundaries is strongly conditioned by sequence stratigraphy. Most of the limits between the larger foraminiferal biozones of the Ilerdian strata of the southern Pyrenees coincide with inherently diachronous transgressive or condensed sections (see also Eichenseer and Luterbacher 1992). However, this diachrony will be generally below biostratigraphic resolution.

A detailed correlation of the stratigraphic sequences observed in the southern Pyrenees with the global eustatic cycles meets several difficulties.

FIG. 4.—Generalized paleogeographic map of the Tremp Basin (from the Ilerdian Serraduy to the Alinya sequences) showing major clastic lowstand deltas and carbonate highstand systems (from Eichenseer and Luterbacher, 1992).
**CHRONOSTRATIGRAPHIC SECTION**

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<th>Depositional Sequences</th>
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| Montsana | Oden | trempina | | | LST |
| Figols | Alinya | corbarica | | | HST |
| | Llimiana | | | | |
| Ager | Ager | ellipso- | | | TST |
| | | deltallis | | | |
| | | cucumbiformis | | | |
| | Serraduy | | | | LST |

**Thrusts**

- sb = sequence boundary
- dls = downlap surface
- ts = transgressive surface

- continental clastics
- calcarenitic bars
- condensed bioclastic grainflats
- deltaic sandstone
- shallow-marine bioclastic carbonates
- thin-bedded outer platform carbonates
- marine shales (undifferentiated)
- reefs
- carbonate debris deposits
- shales with solitary corals ("Pattalophyllia")

**Fig. 5.**—Paleogene chronostratigraphic section of the Tremp Basin. The location of the synsedimentary thrusts is added in order to facilitate the distinction between the influences of tectonic and relative sea-level on the sedimentary evolution (from Eichenseer and Luterbacher, 1992).
**FIG. 6.**—Cross section along the northern margin of the Tremp Basin, showing the main facies of the Ilerdian sequences (from Eichenseer and Luterbacher, 1992).

**FIG. 7.**—Comparison of marine Paleogene sequences of the Tremp Basin with the corresponding interval of the "Mesozoic Cenozoic Cycle Chart" (Haq et al., 1987). Numerical time scale according to Odin and Odin, 1990. Stippled boxes = interval of uncertainty (for discussion see Odin and Luterbacher, 1992).
The documentation of the ages attributed to the different cycles in the “Mesozoic-Cenozoic Cycle Chart” (Haq et al. 1987) is still insufficient. Too many correlations are still on a “take it or leave it” basis.

Problems concerning the Paleogene geochronologic time scale are discussed elsewhere (Odin and Luterbacher, 1992). The numerical time scale of the Paleogene and its correlation with magnetostratigraphy, stages and biozonations is still based on too many extrapolations. At present, a useful numerical time scale cannot be constructed without bridging the gaps between the scarce and irregularly distributed tie-points of reliable radiometric ages by more or less sophisticated extrapolations. However, the degree of uncertainty is necessarily larger than that of the direct radiometric measurements. An extrapolation between two radiometrically dated tie points sterilizes the parameter on which the extrapolation has been based and may lead to circular reasoning. The elaboration of a Paleogene numerical time scale based on a few numerical data and on oceanic magnetic sequences or gradual morphologic changes in microfossils prevents the appreciation of the real rates of spreading of the ocean floor or of the evolutionary rates of microfossils.

One of the main problems of a biostratigraphic correlation of the Paleogene sequences in the southern Pyrenees with the global eustatic cycle chart is the correlation of the “standard” zonations based on planktic microfossils with those based on larger foraminifera. The two types of zonations are based on different approaches. The planktic biozones are mainly based on—possibly somewhat heterochronous (e.g., Weaver and Bergsten, 1991)—first and last appearances of index taxa and have therefore an “event-stratigraphic” connotation. Larger foraminiferal zones are ideally based on successions of biometric patterns of series of cycles (see also Haq, 1991), but there is often the pitfall of circular reasoning. To some extent, this has also happened to me (Luterbacher et al., 1991; Eichenseer and Luterbacher, 1992). Biostratigraphic correlations are always somewhat made with a rubber band and we have to be careful not to overstretch them.

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REFERENCES


PALEOCENE STRATA OF THE BASQUE COUNTRY, WESTERN PYRENEES, NORTHERN SPAIN: FACIES AND SEQUENCE DEVELOPMENT IN A DEEP-WATER STARVED BASIN

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ABSTRACT: Paleocene sediments are not thick in the Spanish Basque Country (usually less than 200 m), largely composed of stacks of hemipelagic limestones and marls deposited in a clastic-starved deep basin. In addition to these, resedimented carbonates accumulated on base-of-slope aprons girding the basin, and resedimented carbonates plus lesser amounts of coarse-grained siliciclastics discontinuously plugged a deep-sea channel system incised on the basin floor.

The building blocks of these sequences are high-order stratification cycles, probably tuned to Milankovitch frequencies. Since these hemipelagic sections contain a nearly continuous stratigraphic record, a reliable reconstruction of the Paleocene sea-level changes that affected the Basque basin has been possible.

A good match has been found between the regional sea-level curve derived from the deep-sea record and that of the global chart of Haq et al. (1988), mainly based on coastal onlap. This correlation clearly demonstrates that the signature of sea-level changes can be confidently unravelled from deep-marine successions, though it remains to be seen whether it reflects an eustatic signature or a bias of the data base.

INTRODUCTION

It is now widely accepted that relative sea-level rises and falls have created a recurring pattern of stratigraphic geometries on many continental margins. This pattern, aptly summarized in the systems tract model of Posamentier and Vail (1988), reflects the back and forth displacement of depocenters across the shelf in response to the changes of the marine base level. The familiar “Exxon’s curves” of sea-level variations are derived from these stratigraphic patterns, mainly from the changes of coastal onlap (Vail et al., 1977). More recently, it has been argued that sea-level changes also affect directly and indirectly the sedimentation in the deep sea (cf. Haq, 1991, 1993). If so, it should be possible to establish variations of sea level from the inspection of the deep marine record. Haq (1991, 1993) explored this possibility by comparing the ages of Cenozoic eustatic cycles with those of several types of oceanographic events. In this paper, we will investigate the same subject through use of a specific field example.

The case study is centered on the Paleocene series of the Spanish Basque Country, represented by a hemipelagic carbonate succession deposited in comparatively deep-marine conditions within the Basque basin. This succession is not thick (up to 250 m, usually less than 200 m and locally just a few 10’s of m). One of our objectives is to demonstrate that this comparatively condensed and seemingly homogeneous succession can be subdivided successfully into genetic units following sequence stratigraphic principles: through outcrop studies, 3rd-order depositional sequences have been recognized, and accurately dated with planktic foraminifera.

A second and more important task of this study is the reconstruction of the history of sea-level changes in the western Pyrenees. For this purpose, the studied succession offers potentially excellent conditions, as tectonism and clastic supply were negligible in the Basque basin during most of the Paleocene time. Therefore, the sea-level signature can be analysed there almost in isolation. Timing of the recognized sea-level changes has been constrained mainly with data obtained in the distal, conformable or near conformable parts of the depositional sequences, where an almost continuous stratigraphic record has been preserved.

GENERAL SETTING

The so-called Basque (or Basque-Bearn) basin was an interplate trough that evolved from latest Cretaceous to at least middle Eocene time in the western part of the Pyrenean area (Plaziat, 1981; Pujalte et al., 1992). The trough was created in early Campanian time, at the beginning of the Pyrenean convergence, by the enlargement and joining of precursor sub-basins of the Aptian-Santonian rifting-spreading stages. It was elongated East-West and comparatively deep (see Table 1), opening westwards into the Bay of Biscay and surrounded elsewhere by shallow shelf areas (Figs. 1A and B).

The Campanian to Eocene record of the Basque basin consists of a thick succession (more than 5000 m) of turbidite and hemipelagic deposits, now preserved in two major outcrops, the Biscay synclinorium and the Gipuzkoa monocline (Fig. 1C). This succession can be subdivided into three broad, informal lithostratigraphic units, each of which was accumulated under different tectonic conditions. The lower one is the “Campanian-Middle Maestrichtian flysch” (i.e., the “Flysch gréseux” of Mathey, 1986), which is a thick and rather monotonous unit of thin- and medium-bedded axially-flowing turbidites deposited during an interval of increased differential subsidence. It reached a maximum thickness of 1700 m in the Zumaia section (Fig. 1C), from where it thins progressively to the north and to the south (Mathey, 1986). The middle unit (Upper Maestrichtian-Paleocene) was deposited during an interval of slight subsidence and relative tectonic stability. This was reflected in the basin by a strong reduction of siliciclastic input and sedimentation rates with development of starved conditions, that were particularly pronounced during Danian and earliest Thanetian time. This interval is therefore much thinner and shows less lateral thickness variations than those of underlying and overlying units. The upper unit, the Eocene Flysch, is incompletely preserved in the Spanish Basque Country because of recent erosion. It is formed by a thick (more than 2000 m), coarse-grained turbidite accumulation (Krüt et al., 1975, Orue-Etxebarria, 1983), recording another phase of rapid subsidence rates.

PALEOCENE PALEOGEOGRAPHY AND MAIN SEDIMENTARY SYSTEMS

The single most distinctive feature of the Paleocene Basque basin was its clastic-starved nature, a fact already noted by sev-
eral previous authors (i.e., Van Vliet, 1982; Puigdefabregas and Souquet, 1986; Razin, 1989). Such starvation was brought about by a combination of causes, the two most important of which were:

1. The amount of siliciclastics delivered from source areas in the western Pyrenees was not very high, because the hinterland relief was subdued as a consequence of the contemporaneous tectonic quiescence. Substantial relief did exist, however, in the eastern Pyrenees, where nappe development and uplift had already occurred (Puigdefabregas and Souquet, 1986; Busquet et al., 1992). Yet, paleocaliche development in the thick continental accumulations coming out of these areas indicates arid climatic conditions (Eichenseer, 1988) and, therefore, reduced run-off.

2. The encroachment of widespread but relatively thin carbonate platforms in the shallow shelves that rimmed the south (Iberian) and north (Aquitanian) sides of the basin, as a result of several regional transgressions (cf. Plaziat, 1981). These platforms were usually wider than 50 km, and laterally continuous (Bureau of Recherches Geologiques et Minieres et al., 1974; Pujalte et al., 1993), and must have acted as efficient barriers to siliciclastics since, as pointed out by Hay et al. (1988, p. 29), the development of carbonate platforms “tends to exclude terrigenous detrital sediments by building up the sea floor to very shallow depth”.

In spite of the starvation, a comparatively minor but still substantial amount of coarse-grained resedimented carbonate and siliciclastic sediments reached the Basque basin during the Paleocene time. Depending on their occurrence and nature, facies associations of three main sedimentary systems have been recognized (Pujalte et al., 1993, and Fig. 2). These are:

1. The carbonate slope-apron: It is made up of variable proportions of carbonate breccias, coarse- and fine-grained carbonate turbidites and (hemi)pelagic limestones and marls. These essentially resedimented carbonates fringe the North Iberian carbonate platform, from which they were clearly shed as evidenced by their position and paleocurrent directions. Outcrops stretch laterally for more than 100 km, from the Orhy peak in the Pyrenees (French-Spanish border) to near the village of Eibar in the Gipuzkoa province (Fig. 2A). Basinwards, however, they grade rapidly into the basin-floor hemipelagites.
2. **The basin-floor hemipelagic stacks:** These consist of alternations of pelagic limestones (globigerinid mudstones and wackestones) and hemipelagic marlstones, with minor intercalations of thin-bedded turbidites. It is the most widespread and characteristic of the three Paleocene facies associations of the Basque basin and the main evidence of the starved conditions. Thin accumulation of hemipelagic stacks (50–135 m) mantle most of the basin floor downcurrent of the carbonate slope apron. Their best outcrops are found in coastal sections of the Bay of Biscay (Sopelana, Zumaia, Hendaià, Bidart), but good inland exposures also exist (see below).

3. **The deep-sea channel:** It may include up to 40% massive coarse-grained siliciclastic turbidites, although it is still composed mainly of reworked carbonates. These are some of the more intriguing Paleocene basinal deposits and certainly a key to understanding correctly the contemporaneous paleogeography of the Basque basin. Accumulations of this association always occur within erosional features in the lower slope and/or basin floor, which are interpreted as parts of an axially-flowing deep-sea channel system in the sense of Carter (1988; i.e., “features incised into unconsolidated sediments of ocean margins troughs or abyssal plains”). Pujalette et al. (1993) suggested a genetic link between the development of this system and the starved conditions, reasoning that its creation coincided with the time of maximum starvation, and lasted while starvation persisted (earliest Danian and early Eocene respectively). This process is further explained below.

Individual channels can be 1–5 km wide and up to 150 m in depth, while the length of the whole system must have exceed 200 km (Fig. 2A). Coarse-grained terrigeneous sediments probably entered the deep-sea channel from the easternmost tip of the basin, where a siliciclastic shelf area did exist (cf. Plaziat, 1981). Carbonate sediments must initially have come from the flanking carbonate platforms, but those bypassing the slope-apron and reaching the channel system were also transported by axial flows. The channel probably acted mainly as a conduit towards deeper waters, as evidenced by the winnowed nature of most of the sediments plugged into it. Thus, it bears some resemblance with the Quaternary Valencia sea-valley, an axially trending deep-marine erosional feature that funnels the sediments coming off the flanking Ebro turbidite system into the Algiero-Pro-
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DEPOSITIONAL SEQUENCES AND CYCLES

Terminology

Six 3rd-order, Vail-type depositional sequences (DS) have been recognized in this study, each of which is ascribed to a cycle of relative sea-level change (see below). The oldest of these spans the latest Cretaceous/earliest Paleocene time, and has therefore been coded as DS-M/D (for Maastrichtian/Danian); the next four were entirely developed during Paleocene time and, in the description that follows, will be respectively identified as DS-P1 to DS-P4; finally, the youngest one includes the Paleocene-Eocene boundary, and will be labelled DS-T/Y (for Thanetian/Ypresian).

In addition to these, a number of high-order stratification cycles are present within the hemipelagic stacks, which might represent parasequences or be the result of climatic forcing (see below). It should be noted also that the term “cycle” will be used with a temporal meaning, namely to designate the time intervals during which relative sea-level changes took place.

Facies and Sequences of the Slope-Apron and Deep-Sea Channel Deposits

An account of the sequence stratigraphy of the Paleocene coarse-grained facies of the Basque basin has been published recently (Pujalte et al., 1993) and, therefore, is only summarized here.

Carbonate Slope-Apron.—

The facies architecture of this system was established by direct correlation of several close-spaced sections situated in the southern limb of the Eibar syncline (Fig. 3A). The resulting cross section gives a transverse view of the accumulation (i.e., perpendicular to paleocurrents), which is shown in Figure 3B along with the planktic foraminifera zonation. The reconstructed Paleocene succession is composed of five stacked sequences (Fig. 3B). They range in thickness from 20 to 50 m and always consist of two parts, respectively, interpreted as lowstand slope fan complexes (LSF) and lowstand wedges (LSW). Their bounding unconformities always have an obvious erosional character, particularly evident in the boundaries underlying DS-P3 and DS-P4, both of which have a differential relief of about 25 m, the older of these having locally eroded out the underlying DS-P2 (Fig. 3B). These erosional boundaries are clear proof of erosion and entrenchement on the slope apron surface caused by massive gravity flows.

LSF’s are typified in the five sequences by variable proportions (40–90%) of polygenetic carbonate breccias (debris-flow deposits), that filled and smoothed the irregular topography of their respective erosional lower boundaries. With the exception of DS-P1, these breccias include shallow-water carbonate clasts that attest to the partial collapse of earlier platform margins. The remainder of the LSF’s is mostly made up of carbonate turbidites, with minor intercalations of hemipelagic marls and limestones. The turbidites are intrabiolastic and usually contain fossils or fossil fragments derived from older deposits. This fact is particularly evident in the turbidites of DS-P1, where Cretaceous globotruncanids and inoceramid prisms are ubiquitous. The hemipelagic marls and limestones, however, represent autochthonous deposition, permitting reliable age-dating of these systems tracts.

The LSW’s of DS-P1 to DS-P4 are composed of alternating globeigerinid mudstones/wackestones with thin-bedded and thick-bedded carbonate turbidites. The latter, shelf-derived bioclastic rudstones and grainstones, are considered “shingled turbidites”. The LSW of DS-T/Y differs from that of the other four sequences in having an essentially terrigenous nature, being composed of dark-grey carbonate-poor shales (<5% CaCO₃), with frequent intercalations of thin-bedded carbonate and mixed turbidites. This accumulation has been found, with variable thickness (1–20 m), in every other studied section of the region, including the hemipelagic stacks (see below). Besides an important influx of fine-grained siliciclastic materials, this particular interval is probably related to a large flooding of dissolved CO₂ (caused by the volcanism which accompanied the opening of the North Atlantic) and led to the dissolution of carbonates. Because of its distinctive lithological character, very different from that of the underlying and overlying sediments, this so-called “dissolution interval” (Canudo and Molina, 1992) or “clayey interval” (Orue-Etxebarria et al., 1996), is very useful for field correlation.

Deep-Sea Channel Sequences.—

They also have erosional boundaries and are composed of LSF and LSW deposits. The character of these systems tracts, however, differs in several important points from those of the slope apron. There are, in fact, two types of LSF’s, siliciclastic and carbonate ones.

Siliciclastic LSF’s are best developed in DS-P3 and DS-T/Y (Fig. 4). In the central parts of the channels, siliciclastic LSF’s consist of medium and coarse-grained sandstones and pebbly sandstones, with quartz pebbles up to 5-cm diameter. They are stratified in meter-scale beds, usually bounded by erosional surfaces created by channelling or scour and fill. In most cases, they show a fining-up size grading, but massive and cross-bedded beds have also been observed. Towards the channel margins, the sandstones interfinger with shales and, in some cases, even with resedimented and hemipelagic carbonates. Where the latter sediments are present, they have been used for dating. Otherwise, the ages of the siliciclastic LSF’s have had to be estimated from their stratigraphic position, as the sandstones are unfossiliferous and the shales contain only scattered agglutinated foraminifera.

Carbonate LSF’s are made up of breccias and debris flow deposits that typically include large clasts derived from older sequences. In the Orio section, for instance, this system tract is represented in DS-P1 by a discontinuous accumulation of Maastrichtian cobbles and boulders of up to 6-m diameter and in DS-P2 by a widespread debris in which both large chunks of Maastrichtian purple marlstones and numerous well-rounded carbonate clasts (some of them of shallow-water origin) “float” in a green marly matrix. These disorganized accumulations were interpreted by Pujalte et al. (1993) as the result of collapse, slide and slump processes on channel walls.

The LSW’s of DS-P1 to DS-P4 are both composed of graded rudstones and bioclastic grainstones (turbidites) that are intercalated with variable proportions of hemipelagic marls and
limestones. The turbidites are frequently thickly stratified, some of them reaching 10 m. This, and their coarse-grained nature, are clear proofs of powerful flows, undoubtedly reinforced by confinement. Dating of these sediments is again based on the planktic foraminifera extracted from their autochthonous components.

The stratigraphic record of deep-sea channel deposits is much more discontinuous than that of the slope-apron. Thus, only one particular section (i.e., Orio, Fig. 4) has representation of the five sequences recognized on the apron. Elsewhere, deep-sea channel successions are made up of one to three sequences. This fact reinforces the interpretation that the deep-sea channel must have been essentially a transport (or bypass) sedimentary system.

**Facies, Depositional Sequences and Stratification Cycles of the Basin Floor Hemipelagic Stacks**

Database.—

The hemipelagic stacks, in contrast to both coarse-grained facies associations, seem to contain a nearly continuous succession, as they lack major internal erosional boundaries or biostratigraphic hiatuses. They offer a much better opportunity to document a detailed Paleocene sea-level history in the Basque basin, provided that the signal of these changes could be recognized.

To explore this possibility, detailed lithologs have been constructed for the five best exposed hemipelagic sections of the region: Zumaia, Sopelana, Hendaia, Berano and Trabakua pass...
been matched bed-by-bed between the different sections (see below). Correlation with the slope-apron and deep-sea channel sequences is based also on the planktic foraminifera that occur in these coarse-grained deposits only within scattered marly intercalations.

Other aspects of the evolution of the planktic foraminifera that might be related to sea-level changes and/or climatic variations are currently being investigated. These include, for instance, the position of First and Last Appearance Datums (FAD and LAD), the relative proportions of warm and temperate species, or changes in coiling direction. Some of the results are discussed below.

**Lithology, Macrofossils and Trace Fossils.**

The lithology of the hemipelagic stacks is comparatively simple, being mostly composed of limestones and marls, with lesser amounts of clastic intercalations. The limestones, either grey or pink in color, are formed by tests of carbonate nanofossils (see Seyve, 1990 for excellent SEM views), planktic foraminifera and calcispheres, all set in a micritic carbonate matrix. But for the fraction of matrix, probably transported as mud from the shallow platforms, the limestones can be considered true pelagic sediments and catalogued as highly indurated chalkstones. The marls, on the other hand, are clearly hemipelagic since in addition to planktic microfossils, they contain variable proportions of illite clays, quartz silts and other fine-grained terrigenous sediments (30–70%, according to Mount and Ward, 1986; Ten Kate and Sprenger, 1993). The color of Danian and early Thanetian marls is usually pink or red, while late Thanetian-early Eocene marls are either dark grey or greenish.

There are two types of clastic intercalations, namely a 3.5-mm-thick clayey intercalation that occurs near the Paleocene/Eocene boundary at the Trabakua, Berano and Zumaia sections, and numerous event turbidites. The former is but the expression in the hemipelagic setting of the “dissolution interval” described earlier in the slope apron deposits. Event turbidites occur frequently in the Zumaia and Berano sections, but are conspicuously absent in the others. They are typically thin-bedded (from a few mm up to 15 cm), show incomplete Bouma sequences (Tbc, Tbd, Tc or Tcd) and, at outcrop scale, are laterally continuous. Based on their composition, they can be divided into carbonate (i.e., purely calcarenitic), siliciclastic (i.e., purely quartzitic), and mixed turbidites. Carbonate and mixed types are far more abundant than siliciclastic ones.

Occasional echinoderms are the only large body fossils found in the succession. By contrast, trace fossils are very common, particularly within the limestone beds which are often pervasively bioturbated. Their association (Planolites, Thalassinoideas, Chondrites and Zoophycos), the scarcity of megafossils and the absence of borings or hardgrounds (cf. Ekdale and Bromley, 1984) are all in agreement with the depth estimates of the basin indicated on Table 1.

**High-Order Cyclicity.**

The lithologies described above are stacked with a cyclic (or rhythmic) bedding pattern. The simplest cycles recognizable on outcrop are the bedding couplets, which are either represented by alternating marls and limestones in different relative proportions or by a vertical variation in carbonate content within.
marl or limestone intervals (Fig. 5). Thickness of these couplets range from 15 to 50 cm, with an average of about 35 cm. The limestone/marl (L/M) couplets, in turn, can be grouped into bundles, which may contain from 5 to 7 couplets (bundles made up of marly or limey couplets must also exist, but are more difficult to see in the field). A whole range of bundles has been observed, the two end-members of which have been respectively termed open and crowded. Open bundles designate bundles in which all, or a majority, of the constituent couplets retain their marly and limey portions; in the crowded type, in contrast, the marly portions of the couplets are missing (sometimes giving the wrong impression of being just one single bed), although successive bundles are still separated by comparatively thick marls (Fig. 6). Bundles of intermediate character are collectively referred to as amalgamated. As a rule, the average thickness of the bundles diminishes from open to amalgamated to crowded ones, indicating a declining sedimentation rate in the same direction (Figs. 5, 6).

The more carbonate-rich segments of the succession are obviously composed of many individual couplets and bundles, but none of them are clearly expressed. Such segments are therefore termed multiplets; three intergradational varieties of these can, in turn, be recognized (Fig. 5): (i) stratified, multiplets with well-defined and laterally persistent internal bedding planes, sometimes outlined by millimeter-scale marly intercalations; (ii) amalgamated, when the internal bedding is discontinuous within a few meters and (iii) massive, when the internal bedding is either weakly developed or can not be discerned. For some unknown reason, multiplets were prone to sliding and slump.

Depositional Sequences.—

Six depositional sequences have been recognized in the Zumaia section, which are expressed in the stratigraphic record in three different ways (Figs. 6, 8):

By Variations of Sedimentation Rates.—Relative rates have been estimated from the number of couplets and/or bundles per meter of section, under the assumption that these stratification cycles represent, on average, similar time spans, whether or not they are tuned to Milankovitch frequencies. The pattern of variation is somewhat different in each case, but in general, the lower parts of the sequences are composed of open bundles suggesting comparatively high-sedimentation rates; these are followed by condensed intervals, during which sedimentation rates were noticeably below average. In the field, these intervals appear either as a stack of crowded bundles (e.g., DS-P1, see Fig. 6) or as a zone with an unusually large number of couplets per meter of succession (DS-M/D, P2 and P3). They are also the most bioturbated parts of the sequences and the ones with a higher percentage and/or absolute abundance in planktic foraminifera. These condensed intervals are interpreted as the basal expression of the transgressive systems tracts (TST) and separate, therefore, the lowstand (LS) (below) and highstand (HST) deposits (above; Fig. 8).

By Long-term Vertical Changes of the Bulk Carbonate/Clay Ratios.—As reflected in the relative proportion of marls and limestones, these changes are readily visible in outcrop, and very useful for field correlation. Also, for the reasons explained in the next section, the interval of minimum carbonate content of a sequence has been used in some cases to separate lowstand deposits into the LSF and LSW.

Fig. 5.—Summary diagram illustrating the range of stratification cycles observed within the hemipelagic sections and descriptive names given to them in this paper. See text for explanation.
By Changes in the Frequency and/or the Composition of Turbidite Intercalations.—Turbidites are either absent or infrequent within condensed intervals (Fig. 8). Also, turbidites attributed to the LSF often have a mixed composition, containing variable proportions of quartz grains and planktic microfossils of intrabasinal derivation, whereas turbidites of the remaining parts of the cycles have a bioclastic carbonate composition, usually including a fraction of shallow-water bioclasts (i.e., algae, benthic foraminifera, coral fragments, etc.) These contrasting compositions are thought to reflect, respectively, the absence or existence of a contemporaneous shallow-carbonate platform.

Except for the turbidites, which only occur at the Zumaia and Berano sections, the depositional sequences exhibit very similar features in the other hemipelagic sections examined for this study (Fig. 9). This correlation is clear proof of their basin-wide development.

The characteristics of the six hemipelagic depositional sequences are shown on Figures 6 to 9, and summarized in Table 2. As shown in these figures, each one has unique lithological features. Also, their thicknesses vary from sequence to sequence. And, finally, their time spans, deduced from their biostratigraphic zonation and the time scale of Berggren et al. (1985), vary between 0.5 and 3.5 Ma. Partly for that reason, 3rd-order sea-level changes are thought to have produced the sequences since this particular forcing is known to have acted at such time range (cf. Haq et al. 1988). Another important support to this interpretation is the fact that, with one exception (i.e., DS-M/D), the sequences recognized within the hemipelagic stacks and those found in the deep-sea channel and/or the slope apron, can be linked biostratigraphically to each other (Figs. 3, 4). Finally, a rapid change in the coiling directions of two planktic foraminifera species, from random to preferred sinistral coiling, has been observed at the boundaries between DS-P2/P3 and P3/P4 (Fig. 8). By analogy with more recent species such as Globigerina pachyderma, these changes are
taken as indicative of contemporaneous sea-water cooling (cf. Ericson, 1959), indirectly pointing to sea-level falls.

The hemipelagic depositional sequences can in turn be subdivided into symmetric (i.e., DS-M/D, P1, P2 and P4) and asymmetric (i.e., DS-P3 and T/Y). Such a distinction initially relies on their pattern of vertical variation in carbonate content: symmetric sequences exhibit a decreasing limestone/marl ratio in their lower part and an increasing ratio in their upper one; asymmetric sequences open with a marly or clay rich interval, while the limestone/marl ratio still increases upward (Fig. 8). The two types of sequences differ in two other important aspects: (i) boundaries below symmetric depositional sequences are gradational, while those below asymmetric ones often have a gentle erosional character and (ii) massive siliciclastic turbidites occur in deep-sea channels only in connection with hemipelagic sequences of asymmetric character (Fig. 4). For these reasons, asymmetric sequences are thought to have been initiated by major relative sea-level falls, a possibility which is further explained in the next section.

A GENETIC MODEL FOR DEPOSITIONAL SEQUENCES

As mentioned above, the development of depositional sequences is thought to have been controlled by 3rd-order sea-level changes. In this section, we describe how these changes may have affected basin sedimentary processes. Previous investigations (Pujalte et al., 1993) have demonstrated that, in this basin, the Paleocene platform-basin transition was relatively narrow and steep, a fact incorporated into the genetic model (Fig. 10). In this context, we will discuss an ideal cycle of 3rd-order sea-level change, beginning with a sea-level fall, following with an early rise, then a rapid rise and platform flooding, and finishing with a relative highstand position. During each of these phases, lowstand slope fan complexes, lowstand wedges, transgressive and highstand system tracts were in turn developed.

Sea-Level Fall

In this carbonate-dominated system, sea-level falls generally produced destabilization, collapse and erosion of the platform margin and upper slope, triggering slumps, debris-flows and turbidity currents. These dense flows had initially an erosional character that resulted in the irregular incision of the slope apron surface and in the excavation (and successive reexcavation) of the deep-sea channel system. Later, these erosional features were partly or completely backfilled with carbonate megabreccias and intraclastic graded calcarenites with some hemipelagic limestones and marls deposited at times of reduced resedimentation.

Most of the basin-floor was little affected by the resedimentation of coarse-grained deposits, receiving mainly hemipelagic sediments and, in some sections, thin-bedded turbidites. However, the sea-level drop and the concomitant basinward shift of the shoreline increased the terrigeneous clay input which, homogeneously dispersed through nepheloid layers, diluted the local pelagic rain.

During gentle sea-level lowerings, this process was gradual, leading to the progressive decline in the carbonate/marl ratio typical of the hemipelagic symmetrical sequences (Fig. 10). Rapid sea-level lowerings left a different imprint. Their high rate of fall eventually caused breach of the carbonate platform by fluvial incision, and substantial amounts of terrigeneous sediments reached the Basque basin for a while, directly supplied from continental sources. Coarse-grained siliciclastics then became trapped in the deep-sea channel system (i.e., LSF's of sequences DS-P3 and T/Y, Fig. 4). At the same time and for some time after, large amounts of fine-grained deposits were
dispersed on the basin floor, causing an abrupt and more or less important dilution of the hemipelagic signal on the lower parts of the asymmetric cycles.

**Early Sea-Level Rise**

The sea level began to rise and the platform margin was again submerged and stabilized, slowing and eventually halting slumps and debris flows. The shallow-water production also resumed. Because of its proximity to the platform edge, a fraction of the shallow-water carbonates were easily transferred to the basin, probably through the action of storms. The bulk of these accumulated as bioclastic turbidites in the slope-apron and in the deep-sea channel system, but some also reached the basin floor. The clay input into the basin gradually declined and the vertical trend of carbonate/ Marl ratio of the hemipelagic sequences was reversed.

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**Fig. 8**—A litholog with planktic foraminifera data showing the vertical changes in lithological parameters and 3rd-order depositional sequences recognized in the Zumaia section.

<table>
<thead>
<tr>
<th>BIOZONES</th>
<th>BIOEVENTS</th>
<th>LITHOLOG</th>
<th>SEQUENCE STRATIGRAPHY</th>
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<td>P. angulata</td>
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<td>A. mayarcensis</td>
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<th>MARLY LIMESTONE</th>
<th>MARLY MARL</th>
<th>THIN-BEDDED TURIDITES</th>
<th>MIXED (CARBONATE AND STENOCLASTIC)</th>
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<tr>
<td>Marly marl</td>
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Rapid Sea-Level Rise

An accelerating sea-level rise pushed the shallow-water factory landwards and caused the drowning of the platform margin. The amount of both coarse-grained carbonate and fine-grained terrigenous sediments reaching the basin was reduced. Probably, pelagic productivity was also cut down, due to decreased nutrient supply to the offshore. It was during these phases when the Basque basin became more severely starved, and the hemipelagic condensed intervals were created.

Sea-Level Highstand

Sea level rose to its apex, stabilized for a while, and then began a relative fall. The shallow-water carbonate platform reached peak development, effectively trapping most of the terrigenous sediments, including clays. For a while, the basin received almost exclusively carbonate deposits, mostly as hemipelagic rain, but with a fraction derived from the shallow platform (i.e., the bioclastic turbidites of Zumaia and Berano sections, and possibly some carbonate muds). These and the sediments produced during the previous phase very likely blanketed the entire basin, including the slope-apron and deep-sea channels. In the last two systems, however, they were subsequently removed by the erosional processes of the next sea-level fall.

COMPARING REGIONAL AND GLOBAL SEA-LEVEL CHANGES

A tentative regional curve of 3rd-order sea-level changes has been constructed with the data discussed in the previous sections (left-hand side of Fig. 11). The calibration of this curve is entirely based on the information acquired from the hemipelagic stacks; the ages of the sequence boundaries and of the systems tracts of DS-P1 have been fixed with planktic foraminifera (Fig. 8). The positions of the boundaries of the other five sequences are still based on biostratigraphy, but their system tracts relied on couplet/bundle counts. Magnitudes of sea-level falls were established by taking into account the following parameters: (i) a comparatively high terrigenous content is taken as proof of an important sea-level fall, with breaching of the
Substrate of the studied succession: Alternating Upper Maastrichtian purple hemipelagic marls and grey pelagic limestones that also contain depositional sequences and stratification cycles similar to those described in this paper (e.g., Mount and Ward, 1986; Ward, 1988; Smit et al., 1992; Pujalte et al., 1993). The base of DS-M/D, not shown in Fig. 9, is always gradational.

Lower part (LS and condensed interval) of DS-M/D: This unit consists of purple marls, usually with a color deeper than the substrate. In Zumaia, Smit et al. (1992) recognized 12 marly couplets in this interval, and Ten Kate and Sprenger (1993) established that the bulk carbonate content decreases gradually in its lower part and then increases up to the K/T boundary. This compositional variation is reflected in a well-defined concave outline of the outcrop (zones with less carbonate more eroded easier). The other four sections exhibit the same weathering profile. Ten Kate and Sprenger (1993) also found a drop in sedimentation rates toward its upper part, here considered to mark the condensed interval. Millimeter-to-centimeter-thick turbidite intercalations occur in Zumaia but not in the other sections.

Cretaceous/Tertiary boundary event section: This event, extensively described elsewhere (e.g., Smit and Romein, 1985), abruptly reversed the latest Maastrichtian carbonate increase; however, the trend resumed shortly afterwards, producing a coarse multiplet.

HST of DM-M/D: A multiplet made up of pelagic limestones, usually grey in color, with few and thin hemipelagic marl intercalations. Except in the Trabakua section, where it has a massive character, this multiplet is either amalgamated or stratified. Its thickness varies in each section, probably reflecting a hiatus in the Trabakua and Hendea sections. These two sections also exhibit slumpform and sliding which, at Hendea, have locally evolved into intraformational conglomerates. Three mixed turbidite intercalations have been observed in the Zumaia section.

Lower boundary of DS-P1: It is drawn at the lowest marly intercalation of the well-defined open bundle succession described at point 6; this boundary often coincides with a change from grey to pink colors in the limestones.

LSF of DS-P1: A well defined suite of 23 L/M couplets, grouped in 4 bundles, that can be correlated basinwide bed-by-bed (cf. Fig. 7). Six mixed turbidite intercalations have been observed in the Zumaia section.

“Marly interval” A single bundle (no. 5 in Fig. 6), that is composed mainly of marl-dominated L/M couplets. This is the portion of DS-P1 with a highest proportion of marls, appearing on the weathering profile as the most recessing segment of the outcrop. This marly bundle, probably a minor condensed interval, is used to separate the LSF and LSW within DS-P1 (Fig. 6).

LSW of DS-P1: It is made up of 7 bundles (6 to 12 in Fig. 6), the lowest one still with an open character, the other six becoming progressively more amalgamated upward in the succession. Five carbonate turbidite intercalations have been found in the Zumaia section.

Main condensed interval of DS-P1: It is composed of 7 crowded bundles (13 to 19 in Fig. 6) that average 50 cm in thickness. This is the reddish and most bioturbated part of DS-P1 and the one with the highest abundance of planktic foraminifera. It also marks a radiation event for the fossil group, expressed by the first occurrence of keeled species (i.e., Morozovelloides).

HST of DS-P1: A multiplet formed mainly by grey hemipelagic limestones, usually amalgamated in its lower part and stratified in its upper one. It is only fully preserved in the Zumaia section, where it reaches 12 m in thickness and is composed of 10 bundles. Three carbonate turbidite intercalations occur at Zumaia, one of which is the thickest one of DS-P1 (20 cm).

Lower boundary of DS-P2: It is drawn at the lowest marly intercalation of the well defined open bundles succession described at point 12. This is again a gradual boundary, very similar to the DS-P1 (see point 5).

LS of DS-P2: A well defined suite of 30 L/M couplets, grouped in 5 bundles, which are correlative basinwide. Couplets no. 22 and 23 of this suite are amalgamated and form an outsized double-bed, about 60 cm thick in the Zumaia section, which is also present at Trabakua, Berano and Hendea. But for this double-bed, the thickness of both individual couplets and bundles diminishes gradually upwards in all sections. Three carbonate turbidites have been found in the Berano section.

Condensed interval of DS-P2: A marly interval, always red in color, highly bioturbated and rich in planktic foraminifera. In Zumaia, the condensed interval is 3.2 m thick and is composed of 17 marly couplets (i.e., average couplet thickness is just below 19 cm).

HST of DS-P2: This interval is present only in Zumaia, where it is 3 m thick and consists of 11 marl-dominated L/M couplets (average couplets thickness around 27 cm).

Lower boundary of DS-P3: In Zumaia, it is marked by a sudden jump in the average couplet thickness (from 27 cm below to >40 cm above) and by the occurrence of numerous, very thin mixed turbidite intercalations. Below the boundary, the planktic foraminifera Planorotalites oculus has a random coil orientation, direction, but above the boundary 75% of the foraminifera have sinistral coil. This boundary has a mild erosional character in Trabakua and in several other sections not mentioned in this paper.

LS of DS-P3: It contains grey marls and limestones, with occasional thin-bedded turbidites. The proportion of limestones vs. marls increases vertically in every section, although the exact pattern may vary. The interval begins either with marly couplets or marl-dominated L/M couplets and then grades vertically up into either intermediate- or limestone-dominated L/M couplets. In Zumaia, this interval can be further subdivided based on the type of turbidite intercalations into LSF (dominated by mixed turbidites) and LSW (dominated by carbonate turbidites). The carbonate turbidites, also present in the Berano section, contain the first occurrences of Paleogene larger benthic foraminifera (Discocyclinids and Operculina). In Zumaia, the number of couplets within the interval range between 90–100 (the exact estimate is hampered by the marly nature of the succession, and by the abundance of turbidites). Thickness of these couplets vary between 30–50 cm.

Condensed interval of DS-P3: It varies in thickness between 3 and 4 m, usually composed of intermediate L/M couplets, often slightly reddish in color and rich in planktic foraminifera. At Zumaia, it contains 15 couplets of an average thickness of 20–25 cm, and is the only segment in DS-P3 devoid of turbidite intercalations.

HST of DS-P3: In Zumaia, it is made up of 27 limestone-dominated L/M couplets, grey in color, and ranging between 30–40 cm in thickness. In Trabakua and Berano, it is represented by a stratified multiplet. Carbonate turbidite intercalations that occur at Zumaia and Berano contain the first occurrence of the larger benthic foraminifera Glomalveolina.

Lower boundary of DS-P4: A gradational boundary marks the change from limestone-dominated L/M couplets to either marl-dominated or intermediate L/M couplets. It also coincides with a change in gravel size for the planktic foraminifera Planorotalites pseudomenardii (from fine gravel to predominantly sinistral coal). In Zumaia and Berano, it is also marked by a change in the type of turbidite intercalations (i.e., from carbonate to mixed type, see Fig. 8).

LS of DS-P4: Similar to those of DS-P1 and P2, it is composed of a well-defined suite of L/M couplets (about 30–40). The proportion of limestone vs. marl decreases upward. In Zumaia and Berano, there occur numerous intercalations of thin-bedded mixed turbidites.

Condensed interval of DS-P4: This is a comparatively thin interval (0.5–1 m) made up of poorly-defined marly couplets and very rich in planktic foraminifera.

HST of DS-P4: It is represented by a thin carbonate multiplet (~<1 m), either amalgamated or massive. Typically, most of the planktic foraminifera tests within this interval are infilled with glauconite, a feature best visible on thin sections.

Lower boundary of DS-T/Y: This is almost a sudden transition, often slightly erosive.

LSW of DS-T/Y: In Zumaia, Berano and Trabakua, it is made up of marly shales, dark grey in fresh outcrops but weathers to reddish tones in old ones, characterized by a very low content in carbonates (<5%), organic matter (<0.5 TOC) and by an unusually low content in planktic foraminifera. This is in fact the so-called “dissolution or clacey interval”.

Condensed interval of DS-T/Y: A thin (30–50 cm), strongly bioturbated interval of red marls which are extremely rich in microfossils, mostly planktic foraminifera.

HST of DS-T/Y: Not studied in detail, it is usually represented by limestone-dominated L/M couplets and locally by a carbonate multiplet. A few carbonate turbidites have been observed in Zumaia.
carbonate platform margins; (ii) the presence of shallow-water limestone clasts within LSF’s carbonate breccias is considered an index of contemporaneous margin collapse and therefore of medium or major sea-level falls; and (iii) the relative relief of the lower boundaries of the sequences is finally regarded as indicative of the importance of the erosional capacity of the bypassing gravity flows carving them.

On that basis, major falls are attached to sea-level cycles P3 and T/Y, because their corresponding deep-sea channels LSF’s are made up of coarse-grained siliciclastics (Fig. 4) and their hemipelagic sequences are asymmetric. Also, their slope apron LSF’s contain numerous shallow-water limestone clasts and their lower sequence boundaries are strongly erosional (Fig. 3). Medium falls are assigned to cycles P2 and P4 because although the slope apron sequences still include shallow-water limestone clasts and have high relief erosional lower boundaries (Fig. 3), their deep-sea channel sequences have a low terrigenous content (Fig. 4). Besides, the hemipelagic sequences P2 and P4 have a symmetric character.

A minor fall is finally ascribed to cycle P1, which in the hemipelagic stacks is represented by symmetric sequences and does not contain shallow-water limestones clasts in its LSF’s carbonate breccias. Their lower erosional boundary is only mildly erosional on the slope apron (Fig. 3). It must be mentioned, however, that this particular boundary has a very important erosional character in the deep-sea channel system. This apparent contradiction is explained below.

The Paleocene epoch began amid a transgression and went into a highstand phase that lasted for about 3 Ma (Fig. 11). During this interval, only a comparatively thin carbonate multiplet was deposited on the basin floor (up to 7 m thick in the Zumaia section), indicating extremely low sedimentation rates. This is a worldwide feature linked to the Cretaceous/Tertiary biological crisis, which caused a drastic drop in pelagic sedimentation rates, due to low ocean productivity (cf., Herbert and Hondt, 1990). In the Basque basin, however, it had an important side effect, namely the first excavation of the deep-sea channel system. During this interval, the position of the basin floor became progressively elevated with respect to that of the Bay of Biscay to the west (Fig. 1), as even a small differential subsidence between the two areas could not be compensated with sediments. The next sea-level fall was not very pronounced, but lasted for about 400 ka (i.e., 4 bundles, see Figs. 6, 7), a time during which turbidite activity resumed. Some of these density flows reached the basin axis and were confined along morphological or structural depressions, where their erosional capacity was strongly enhanced. Evidence for this interpretation is the absence of erosional features on the widespread hemipelagic stacks and the very important but local erosion below the deep-sea channel successions. Judging from the thickness of the missing sections, the base of the channel system must have been initially at least 150 m below the average basin floor. Because of the starvation, this important erosional feature was never completely filled with sediments in the course of the Paleocene time.

In Figure 11, the regional sea-level curve is compared with the cycle chart of Haq et al. (1988), the best model so far available of “global” (eustatic) cycles. The former is based on a deep-marine record; the latter mainly on coastal onlap (Vail et al., 1977; Haq et al., 1988). In spite of their different data bases,
the comparison between the two curves reveals more similarities than discrepancies. Thus, both curves have the same number of Paleocene 3rd-order sea-level changes. The ages and duration of corresponding regional and “global” cycles are roughly similar, differences in timing of the cycle boundaries never being greater than 300 ka. The relative magnitudes of sea-level falls coincide for most of the cycles. In fact, only the fall assigned to cycle T/Y is clearly higher than its “global” equivalent and this is likely the result of a regional tectonic enhancement since it was precisely at this moment when the foreland basin stage began in the Pyrenees (cf. Puigdefabregas and Souquet, 1986).

Two main points arise from the correlation discussed above: (i) deep-sea hemipelagic successions are suitable for reconstruction of sea-level changes. Indeed, if such successions can be dated with accuracy, they are ideal for time calibration of the sea-level changes and (ii) the Paleocene sea-level fluctuations of the Basque basin were predicted with reasonably fidelity by the global chart. However, this segment of the chart is based mainly on data of southern England, Belgium and the Paris basin (cf. appendix C, of Haq et al., 1988). These areas are both relatively close to each other and to the Basque basin. It is also possible, therefore, that the good match between both curves simply indicates a common data base. Clearly, information from localities further afield will be necessary to elucidate whether the Exxon chart displays a true eustatic signature for the Paleocene time, or just a western European one.

**CONCLUSIONS**

This paper has demonstrated the validity of sequence stratigraphic concepts even in a deep-sea clastic-starved succession, such as the one deposited within the Paleocene Basque basin. A recurring pattern of sedimentary organization has been recognized, that is attributed to 3rd-order sea-level changes. On that basis, the studied succession, earlier considered just one individual lithostratigraphic unit, has been divided into six depositional sequences.

The field expression of these sequences and of their constituent systems tracts is different in each of the three major sedimentary systems that coexisted in the basin. In the carbonate slope apron girding the basin and in the deep-sea channel system draining its axial zone, the sequences are mainly made up of coarse-grained resedimented deposits. They directly reflect the changes in accommodation in the surrounding shelfal areas which dictated the timing, nature and volume of resedimentation. Thus, the LSF’s and LSW’s of slope apron sequences respectively record the partial collapse of the platform margin following shallowing or even exposure during relative sea-level falls and its later reconstruction and/or stabilization during re-flooding. The most important sea-level drops, on the other hand, are punctuated by the entrance of coarse-grained siliciclastic deposits into the basin, which form the LSF’s of some of the deep-sea channel sequences.

Changing sea-level also influenced hemipelagic sedimentation in the remainder of the basin by controlling the amount and rate of fine-grained suspensioned sediments and the organic productivity of sea water. A large-scale, basinwide vertical cyclicity pattern has been recognized that reflects 3rd-order sea-level signal imprinted upon the hemipelagic stacks. These hemipelagic depositional sequences contain an almost continuous stratigraphic record and can be accurately dated with planktic foraminifera. Thus, they permit a reliable time calibration of the sea-level changes. Such sequences are in turn made up of stratification cycles, possibly related to Milankovitch frequencies. Further studies of these high-order cyclicities may help to improve time calibration.

The curve of Paleocene sea-level changes obtained with data of the Basque basin has many similarities with the one derived from coastal onlap on the “global” chart of Haq et al. (1988). Such a good correlation indicates that reliable sea-level reconstructions can be obtained from the deep-sea record alone; whether it also reflects a common eustatic signature is still open to question.

**ACKNOWLEDGMENTS**

Research for this paper was funded by the Basque Country University projects UPV 121.310-EA 155/92 and UPV 310-EB233/93, and the DGICYT project PB92-0457. We thank both institutions for their support. We are indebted to B. U. Haq, P. Vail and J. Smit, for stimulating discussions in the field. We also express our gratitude to D. J. Horne (University of Greenwich, U.K.) and SEPM editor Dr. D. S. Ulmer-Scholle (Southern Methodist University, Dallas, USA) who made the English text more readable. Needless to say, however, responsibility for possible misconceptions remains entirely with us.

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PART III
MESOZOIC ERA
CRETACEOUS PERIOD
INTRODUCTION TO THE UPPER CRETACEOUS

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Four papers, addressing all or part of the upper Cretaceous, were submitted for this volume. Two papers concern sequence stratigraphic interpretations of carbonate ramp deposits in northern Spain. Gräfe and Wiedmann (this volume) describe the sequence stratigraphy of the Basco-Cantabrian Basin and Floquet (this volume) describes the sequence stratigraphy of shallow water carbonates of the Castillian Ramp. Robaszynski et al. (this volume) describe and calibrate sequence stratigraphic interpretations for the Cenomanian and lowermost Turonian in five areas of the boreal Anglo-Paris Basin including basin margin sections in Devon (UK) and the type area of the Cenomanian Stage, Maine (France), basinal sections in Kent (UK) and the Boulonnais (France) and an intermediate section in Normandy (France). Philip (this volume) describes sequences and systems tracts in mixed carbonate and siliciclastic deposits in the Provence (southeastern France). All papers illustrate the transgressive nature of the Cenomanian and lower Turonian. Floquet (this volume) begins his late Cretaceous/Danian Megacycle with “Cycle boundary” CB 1 at the inflata—dispar ammonite zonal boundary in the upper Albian. Floquet (this volume) divides his Cretaceous/Danian Megacycle in four “Long-Term Depositional Cycles” with boundaries at the base of the late Cenomanian, earliest Coniacian and latest Santonian.

The Cenomanian transgression in northwestern Europe, described by Robaszynski et al. (this volume), appears to coincide with a basin subsidence event resulting in a rapid increase of accommodation space in the upper lower, middle and upper Cenomanian. Rates of sedimentation are moderate in most of the basin except for a Cenomanian transgression in the Anglo-Paris Basin. The basal boundary is interpreted as an unconformity in all 5 areas studied in the lower third of the Mantelliceras saxbii subzone. Transgressive surface (St. Jouin surface in Normandy) and maximum flooding surface are interpreted near the middle of the M. saxbii subzone.

Sequence 2 (Ce1 on Chart 4 of Hardenbol et al. this volume). The basal boundary is interpreted as an unconformity in all 5 areas studied in the lower third of the Mantelliceras saxbii subzone. Transgressive surface (St. Jouin surface in Normandy) and maximum flooding surface are interpreted near the middle of the M. saxbii subzone.

Sequence 3 (Ce2 on Chart 4 of Hardenbol et al. this volume). The basal boundary is interpreted as an unconformity in 3 areas, (Maine, Normandy (Bruneval 1 Surface) and Devon) whereas in Kent and the Boulonnais, a thin shelf margin wedge is identified in the upper part of the M. saxbii subzone suggesting we are at or approaching continuous sedimentation. The transgressive surface is in the uppermost M. saxbii subzone. The maximum flooding surface is interpreted in the lower third of the Mantelliceras dixoni zone.

Sequence 4 (Ce3 on Chart 4 of Hardenbol et al. this volume). The basal boundary is interpreted as an unconformity in 3 areas (Maine, Normandy (Rouen Surface) and Devon), whereas in Kent and the Boulonnais a well-developed shelf-margin wedge is identified suggesting the sequence boundary may be a correlative conformity. The stratigraphic position of the sequence boundary is in the uppermost part of the Mantelliceras dixoni zone. The transgressive surface is near the boundary between the Cunningtoniceras inerme and Acanthoceras rhotomagenae ammonite zones. The maximum flooding surface is interpreted near the middle of the A. rhotomagenae zone.

Sequence 5 (Ce4 on Chart 4 of Hardenbol et al. this volume). The basal boundary is interpreted as an unconformity in 3 areas (Maine, Normandy (Pavilly Surface) and Devon), whereas in Kent and the Boulonnais a well-developed shelf-margin wedge is identified suggesting the sequence boundary may be a correlative conformity. The stratigraphic position of the sequence boundary is in the upper part of the A. rhotomagenae zone. The transgressive surface is in the lower Acanthoceras jukesbrownei zone and the maximum flooding surface is interpreted in the uppermost A. jukesbrownei zone.

Sequence 6 (Ce5 on Chart 4 of Hardenbol et al. this volume). The basal boundary is associated with unconformities or omission surfaces in all 5 areas (Antifer Surface in Normandy). The stratigraphic position of the sequence boundary is at the base of the Metoicoceras gelsiusinum ammonite zone. Shelf-margin wedges are interpreted for the basinal Kent and Boulonnais sections and the intermediate Normandy section. The transgressive surface is near the middle of the M. gelsiusinum zone and the maximum flooding surface is interpreted in the lower Turonian.

Gräfe and Wiedmann (this volume) and Floquet (this volume) identified sequences in the upper Cretaceous of the Basco-
FIG. 1.—Calibration of Cenomanian and Turonian sequences in Robaszynski et al. (this volume), Robaszynski et al. (1990, 1993), Graefe and Wiedmann (this volume), Graefe (1994), Floquet (this volume) and Philip (this volume), with the Chronostratigraphic framework and composite sequences in Hardenbol et al. (this volume) and sequences in Haq et al. (1988), recalibrated to Gradstein et al. (1994, 1995).
Cenomanian/Turonian boundary interval. A precise stratigraphic positioning of sequence boundaries in
ammonite zones between the different areas prevents differences in the stratigraphic position could be caused by (1)
interpretation of the position of the sequence in the studied sections or (2) by calibration uncertainties between the bio-
stratigraphic (ammonite) zonations used in the various basins.

The base of the Cenomanian, in its type area in France, is a
combined sequence boundary and transgressive surface on which the first Cenomanian ammonites are found. Lowstand deposits are not developed in the type area. In Tunisia, the underly-
ing lowstand contains an Albian ammonite assemblage (Robaszynski et al., 1993). Therefore the sequence boundary
below the first Cenomanian deposition is placed in the upper-
most Albian (A111). Gra¨fe (1994) and Floquet (this volume) recognize this sequence as A9 and CB2 respectively. Both au-
thors place the sequence boundary at the Albian/Cenomanian
boundary, which suggests that the sequence boundary and trans-
gressive surface coincide and the lowstand deposits are missing
at their localities. Two minor sequences Ce1 and Ce2 described by Robaszynski et al. (1990, 1993) and Robaszynski et al. (this volume) were not reported from Spain by Floquet (this vol-
ume). Gra¨fe and Wiedmann (this volume) identify sequence
Uc1, probably equivalent to Ce1, but do not report a sequence comparable to Ce2. The most prominent Cenomanian sequence
boundary Ce3, near or at the lower/middle Cenomanian bound-
ary was identified in Spain by Gra¨fe and Wiedmann (this vol-
ume) and Floquet (this volume) as Uc2 and CB3 respectively. Sequence boundary Ce4 close to the FAD of Acanthoceras ju-
kesbrownei was also found in Spain by Gra¨fe and Wiedmann
(this volume) and Floquet (this volume) as Uc3 and CB4, re-
spectively. Gra¨fe and Wiedmann (this volume) and Floquet (this volume) recognize two more sequence boundaries in the up-
permost Cenomanian Uc4, Uc5 and CB5, CB6 respectively, whereas Robaszynski et al. (1990, 1993) and Robaszynski et al. (this volume) recognize only one sequence near the FAD of the Metoicoceras gelsiusianum zone. Ambiguities in the cali-
bration of ammonite zones between the different areas prevent a precise stratigraphic positioning of sequence boundaries in the Cenomanian/Turonian boundary interval.

Three of four Turonian sequence boundaries of Robaszynski et al. (1990), Tu1, Tu2 and Tu3 on Chart 4 compare well with sequence boundaries Uc6, Uc7 and Uc8 described from Spain by Gra¨fe and Wiedmann (this volume) and sequence boundaries CB6’, CB6” and CB7 of Floquet (this volume). Sequence boundary Tu4, of Robaszynski et al. (1990), just below the Coniacian boundary was placed instead in the lowermost Con-
iacian by Gra¨fe and Wiedmann (this volume) as sequence boundary Uc9 and Floquet (this volume) as sequence boundary CB8 based on their records from the Basco-Cantabrian basin in Spain. In platform sections in Europe (Mons Basin) the base of the Coniacian is a transgressive surface and the lowstand deposits are not developed. These different interpretations can be reconciled only if the ammonite content of the lowstand suggests the lowstand straddles the Turonian-Coniacian boundary.

Philip (this volume) identifies 5 sequences (Sb1–Sb5) in the Cenomanian to lowermost Turonian interval in the Provence (southeastern France). The lowest Cenomanian strata (Mantell-
ceras mantelli) onlap Aptian deposits therefore the lower bounding surface combines a number of sequence boundaries. The transgressive surface is at the base of the Cenomanian and compares well with the onlap in northwestern Europe and northern Spain. The stratigraphic position of sequence bound-
ary Sb2 compares with Ce3 of Robaszynski et al. (this volume), at least to the extent that the unconformity separates Mantell-
ceras from Acanthoceras, Uc2 of Gra¨fe and Wiedmann (this volume) and CB3 of Floquet (this volume). Three sequences in the upper Cenomanian-lower Turonian interval recognized by Philip (this volume) (Sb3, Sb4 and Sb5) do not compare well with the interpretations of Robaszynski et al. (this volume), Gra¨fe and Wiedmann (this volume) and Floquet (this volume). Stacking the sequence interpretations for the uppermost Albian, Cenomanian, Turonian and basi Coniacian interval of Roba-
szynski et al. (this volume), Robaszynski et al. (1990, 1993), Gra¨fe (1994), Gra¨fe and Wiedmann (this volume), Floquet (this volume) and Philip (this volume) produces a maximum of 13 sequences for the interval (Table II). Additional biostratigraphic analysis and improved stratigraphic calibration may or may not eliminate the additional sequences. The sequence stratigraphic interpretation for the Cenomanian-Turonian interval is better understood and documented than for most other Mesozoic or Cenozoic stages. The addition of one or even two additional


<table>
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<tr>
<th>Stage/Member</th>
<th>Hag et al (1987, 1988)</th>
<th>Robaszynski et al. (this volume, 1990, 1993)</th>
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INTRODUCTION TO THE UPPER CRETACEOUS 331
sequences, especially in the uppermost Cenomanian and lowermost Turonian interval, through more precise calibration of ammonite chronozones and improved interpretation of the Cenomanian-Turonian sedimentary succession, cannot be ruled out. Sequences Uc10–Uc20 identified in the Coniacian-Maastrichtian Interval by Grae and Wiedmann, (this volume) and CB9–CB14 by Floquet (this volume) can not be reliably positioned stratigraphically because of the poor calibration of biostratigraphic subdivisions with standard stages for this interval. For the same reason Hardenbol et al., (this volume), include a North American sequence record for the Coniacian-lowermost Maastrichtian interval calibrated to north American ammonite zones on the Cretaceous chronostratigraphic chart (Chart 4). For the upper lower and upper Maastrichtian Hardenbol et al., (this volume) include a tentative sequence record based on outcrop data from the type areas of the Maastrichtian Stage in The Netherlands and Belgium calibrated to belemnite zones.

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SEQUENCE STRATIGRAPHY ON A CARBONATE RAMP: THE LATE CRETACEOUS BASCO-CANTABRIAN BASIN (NORTHERN SPAIN)

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ABSTRACT: The investigation of the Late Cretaceous sedimentary series in the Basco-Cantabrian Basin (BCB) using sequence stratigraphic methods allows the establishment of a regional chart of relative sea-level changes. Twenty Late Cretaceous sequence boundaries with 19 depositional sequences can be followed through the whole basin. The sequence stratigraphic model explains lateral sediment thickness and facies variations. High-resolution biostratigraphy was successfully applied to correlate sections and to compare the BCB with other basins. The regional cycle chart of the BCB correlates with some sequence boundaries on published global cycle charts, but there are also important misties. The correlation with more recent regional sequence stratigraphic results from Tunisia and northwestern Europe is much better.

The analysis of sedimentary cycles in the Late Cretaceous carbonate ramp and 2-D reflection seismic profiles. The interpretation of outcrop sections of deepening events, facies stacking patterns and depositional geo cycles tract; however, it contributed with additional information. In sedimentary settings dominated by a strong 2nd-order transgression of calciturbidites or absent. In distal sedimentary environments, the alternations and the highstand systems tracts (HST) mainly of micritic limestone. The basinward part of a TST in such a sequence is formed of neritic marls and marly limestones with detrital calciturbidites or absent. In distal sedimentary environments, the alternations and the highstand systems tracts (HST) mainly of micritic limestone. The basinward part of a TST in such a sequence is formed of neritic marls and marly limestones with detrital calciturbidites or absent. In distal sedimentary environments, the alternations and the highstand systems tracts (HST) mainly of micritic limestone. The basinward part of a TST in such a sequence is formed of neritic marls and marly limestones with detrital calciturbidites or absent. In distal sedimentary environments, the alternations and the highstand systems tracts (HST) mainly of micritic limestone. The basinward part of a TST in such a sequence is formed of neritic marls and marly limestones with detrital calciturbidites or absent. In distal sedimentary environments, the alternations and the highstand systems tracts (HST) mainly of micritic limestone.

INTRODUCTION

Currently, the application of sequence stratigraphic concepts to the investigation of sedimentary cycles and the driving forces behind the observed 3rd-order cyclicity are vehemently debated (e.g., Vail et al., 1991; Schlager, 1993; Miall, 1992; Posamentier and James, 1993; Christie-Blick and Driscoll, 1995). In this paper, sequence stratigraphic methods are used on a Late Cretaceous carbonate ramp system on a passive margin at the southern border of the Bay of Biscay. This area, called the Basco-Cantabrian Basin (BCB), is situated between the Pyrenees in the east and Cantabrian Mountains in the west (Fig. 1). The Late Cretaceous carbonate ramp and the related pelagic sediments are preserved in tectonically undisturbed exposures in contrast to the southern Pyrenees. The model described in Engeser and Schwentke (1986), Malod (1989) and Schwentke and Kuhnt (1992) is used for interpreting the tectonic development of the basin. The BCB lies south of a sinistral strike-slip fault system that marked the plate boundary between Iberian and European plates in the Late Cretaceous period (Fig. 2). In Aptian to Santonian time a counterclockwise rotation of the Iberian plate occurred with respect to the European plate, resulting in an extensional tectonic regime, which is clearly displayed on thermo-tectonic subsidence curves (Gräfe and Wiedmann 1993). The first compressive events, recorded by local unconformities, are found in Campanian deposits in the eastern parts of the basin. Campanian compressional events, recorded for example by uplift on subsidence curves (Gräfe and...
FIG. 3.—Regional cycle chart for the Cenomanian to Santonian stages of the BCB. Thick lines in the onlap columns are sequence boundaries, dashed lines indicate maximum flooding surfaces. The cycle chart is compared with the global cycle chart from Haq et al. (1988) and additional results from Robaszynski et al. (1990, 1993). Sources for the biostratigraphic zonation of the regional chart are Wiedmann (1965), Wiedmann and Kauffman (1978), Lamolda (1982), Küchler and Ernst (1989), Lopez-Sanjaume (1990), Grae (1994) and Gischler et al. (1994). Modified from Grae and Wiedmann (1993) using the absolute time scale from Gradstein et al. (1994).
The very detailed biostratigraphic framework for Cenomanian to Santonian deposits in the BCB is remarkable (Wiedmann, 1965; Wiedmann and Kauffman, 1978; Lamolda, 1982; Lopez-Sanjaume, 1990; Santamaria-Zabala, 1991; Floquet, 1991; Gräfe and Wiedmann, 1993; Gischler et al., 1994) and is based on ammonites, inoceramids, rudists, planktic foraminifers, and benthic foraminifers (Fig. 3). This allows precise correlation between outcrop sections in the basin itself and also the correlation to extrabasinal cycle charts with a resolution much higher than the observed 3rd-order cycles.

METHODS AND SEQUENCE STRATIGRAPHIC CONCEPTS

Several outcrop sections, 700 km of seismic profiles and more than 20 boreholes throughout the basin were used for this study (Gräfe, 1994, 1996). All profiles were interpreted in terms of sequence stratigraphy using methods described in Vail et al. (1987, 1991). The techniques and the sequence stratigraphic and carbonate ramp models used were discussed in detail in Gräfe and Wiedmann (1993) and in Gräfe (1994).

Sequence boundaries are placed according to Gräfe (1994, 1996). On the proximal ramp, sequence boundaries are recognized at major marine-flooding surfaces (Van Wagener et al., 1988) or at unconformities that are associated with a basinward shift in facies (Fig. 4). The candidates for sequence boundaries are correlated by means of biostratigraphy basinward onto the distal ramp. Surfaces with onlap of younger strata above older strata are candidates for sequence boundaries on the distal ramp. Downlap surfaces are followed landward, so that a rock unit is interpreted as LSW if it has a basal downlap surface, pinches out landward and marks a basinward shift in facies (Fig. 4). The basal downlap surface of such a LSW is then established as sequence boundary (SB). In this investigation the basal surfaces of rock units interpreted as LSW on the distal ramp always correlate with major marine-flooding surfaces on the proximal ramp.

Sequence boundaries in this work are named UCn, where n is a number between 1 and 20. Figure 3 shows the correlation of the sequence boundaries found in the BCB with those from Haq et al. (1988) and Robaszynski et al. (1990, 1993). Sequences are named by their lower and upper SB (e.g., sequence UC2/3). The maximum flooding surface (mfs) is characterized in the same way (e.g., mfs 2/3).

LATE CRETACEOUS SEISMIC STRATIGRAPHY OF THE BCB

The seismic stratigraphic interpretation procedure applied here is derived from Vail et al. (1987). First, strong reflectors are picked, correlated through the whole seismic grid and tied to the starting point (Gräfe 1994). Second, points of onlap, downlap, truncation and toplap (Vail et al., 1977, 1987) are identified in the seismic grid. Units that have a basal downlap surface and that pinch-out at the point of offlap break (Vail et al., 1991) are interpreted as LSW in the seismic grid. Downlap surfaces without landward pinch-out of the superimposing systems tract are considered to be maximum flooding surfaces. Third, boreholes as well as ties to outcrops are used for establishing ages at selected reflectors, unconformities and sequence
Fig. 5.—Interpreted version of seismic profile S 82-16 (time section). Abbreviations used in the borehole Boveda 1-Bis: S. = Keuper salt, J. = Jurassic, Apt. = Aptian, Cen. = Cenomanian, Con.-San. = Coniacian-Santonian. Location see Figure 1. The seismic profile is a conventional, 2-dimensional industry standard line processed with a CDP stack and a 50 Hz notch filter. All standard procedures (deconvolution, datum plane correction, NMO correction etc.) are done but the line, shot in 1982, is not migrated. From Graefe (1994).

Fig. 6.—Interpreted version of seismic profile S 82-13 (time section). Location see Figure 1. The technical parameters of this seismic line are the same as those of line S 82-16 (Fig. 5). From Graefe (1994).
SEQUENCE STRATIGRAPHY ON A CARBONATE RAMP

3rd-order cycles have been interpreted in the seismic profile (Fig. 5). Each of these depositional sequences is bounded at the base by an onlap surface (Fig. 5, arrows). The onlap shows a distinctive backstepping towards the south, towards the more landward positions and reflects the strong transgressive trend in the Cenomanian to Middle Turonian deposits. The Turonian sequences do not show significant depositional geometries in this line, due to the overall transgression in this period, which shifts the point of offlap break more landward. Sequence boundaries are formed here by strong reflectors that can be tied easily to other lines where onlap or downlap geometries are visible. The pronounced reflectors forming the Turonian sequence boundaries (SB UC8 for example) are probably caused by a thin calcareous interval, which creates the reflectivity in the argillaceous Turonian series.

The second seismic profile S 82-13 runs subparallel to the basin axis with more basinward positions to the northwest (Fig. 6). The downlap surface at the base of sequence UC4/5 is evident in the northwestern corner of the profile. The other Late Cretaceous sequences show a more or less parallel seismic facies, as is common in seismic profiles striking subparallel to the basin axis. There is also a small salt pillow obvious in the southeastern part of the line. This salt structure reveals a distinct Early to Middle Albian (sequences AL1 to AL6) halokinetic period recorded in the thinning of depositional sequences towards the salt pillow (Fig. 6).

Fig. 7.—Cross section through the Late Turonian to Early Coniacian age depositional sequences UC8/9 and UC9/10 (lower part) based on measured outcrop sections and boreholes. For the distribution of the foraminiferal fauna in the sections Villasana and Gordoa see Figure 8. For biostratigraphic data and description of measured sections and boreholes, see Gräfe (1994).
Seismic lines shot on the area of the proximal ramp exhibit parallel running reflectors and a much thinner Late Cretaceous sedimentary series (Gräfe, 1994). One cannot expect pronounced depositional geometries on such a proximal ramp setting (Burchette and Wright, 1992). However, the pronounced marine-flooding surfaces that separate intertidal to subtidal carbonates from superimposing neritic marlstones (see examples in Floquet, 1991; Gräfe and Wiedmann, 1993; Gräfe, 1994; Segura et al., 1994; Alonso et al., 1994) can be recognized in seismic lines as strong parallel reflectors created by the impedance contrast at the limestone/marl contact. Therefore, such reflectors were chosen as sequence boundaries on lines situated on the proximal ramp, if these reflectors could be correlated with basinward sequence boundaries.

Generally, the resolution of sedimentary cycles on seismic profiles is much lower than the resolution that can be obtained from the study of outcrops and borehole sections (Biddle et al., 1992). However, the pronounced marine-flooding surfaces that separate intertidal to subtidal carbonates from superimposing neritic marlstones (see examples in Floquet, 1991; Gräfe and Wiedmann, 1993; Gräfe, 1994; Segura et al., 1994; Alonso et al., 1994) can be recognized in seismic lines as strong parallel reflectors created by the impedance contrast at the limestone/marl contact. Therefore, such reflectors were chosen as sequence boundaries on lines situated on the proximal ramp, if these reflectors could be correlated with basinward sequence boundaries.

The details of outcrop and well log sequence stratigraphy in the BCB were presented in Gräfe and Wiedmann (1993) and Gräfe (1996). Therefore, only a short summary is given here, and some principles that govern the distribution of lithofacies in a depositional sequence in this carbonate ramp system will be discussed.

For the interpretation of the sedimentary facies, Gräfe and Wiedmann (1993) established 14 facies types that describe the distribution of facies in a depositional sequence. The facies types were also interpreted with respect to their paleobathymetric significance, and a paleobathymetric model obtained from the study of benthic foraminifers allows the accurate position of deepening events in a given section (Gräfe, 1994). The interpretation of depositional geometries, facies analysis and deepening events allowed the recognition of 19 3rd-order depositional sequences in the Upper Cretaceous strata that were bounded by 20 Late Cretaceous sequence boundaries (Fig. 3).

The correlation of several interpreted borehole and outcrop sections as well as the analysis of kilometer-sized outcrops yielded precise information about the distribution of lithofacies in the depositional sequence. Generally, the lithofacies in such a sequence in the BCB is a four component system composed of: (1) autochthonous carbonate grains, (2) detrital carbonate, (3) pelagic and hemipelagic carbonate and (4) terrigenous mud and silt. Two lithofacies distributions in a depositional sequence are presented here.
The first example is from the Middle Cenomanian UC2/3 sequence, a period that is governed by a strong transgressive trend (Wiedmann et al., 1983; Floquet, 1991; Alonso et al., 1994). In this sequence, the parts of the transgressive systems tract (TST) deposited on the distal ramp are composed of marl- and limestone alternations rich in hemipelagic and pelagic components but relatively poor in detrital carbonate (Gräfe and Wiedmann, 1993). In the highstand systems tract (HST), the relative amount of clay- to silt-sized terrigenous input increased leading to the deposition of tens of meters of marl- and claystones without out wackestone and mudstone beds. Autochthonous carbonate grains were not observed in this distal setting; however, they are naturally the major component in the TST and HST on the proximal ramp (Gräfe and Wiedmann, 1993). Lowstand wedges were not recorded in this sequence, probably due to the strong transgressive trend in this period. This transgressive pulse is related to pronounced thermo-tectonic subsidence in the BCB and masks the widespread recorded sea-level lowstand around the Early/Middle Cenomanian boundary (Haq et al., 1988; Robaszynski et al., 1990, 1993). If Cenomanian lowstand wedges occur (for example in the UC3/4 and UC4/5 sequences), they are thin (2 to 20 m thick) and composed of calciturbidites from the nearby proximal ramp. Therefore, the distribution of sedimentary facies in systems tracts resembles in this case the model that was proposed in Robaszynski et al. (1990). Highstand-shedding was not recorded in Cenomanian sequences; whereas, lowstand wedges are only thin and not pronounced in the sedimentary series as expected by models proposed in Haak and Schlager (1989).

A different situation can be observed in the Late Turonian to Early Coniacian sediments. This period is governed by a regressive trend. A very thick hemipelagic series of Late Turonian and earliest Coniacian age is suddenly replaced by 50 to 200 meters of shallow-marine, autochthonous bryozoan-rotalidal packstones and rudist-rudstones (Gräfe and Wiedmann, 1993). Hemipelagic sedimentation resumed above the shallow-marine limestone series in the distal parts of the carbonate ramp. Figure 7 summarizes the distribution of lithofacies in sequences UC8/9 and UC9/10. The pinch-out of the UC9/10 LSW can be established by biostratigraphic data (the missing Petrocierocene Zone on the proximal ramp in Figure 3) and also by field observation (Gräfe and Wiedmann, 1993). Remarkable are the increase in detrital carbonate in the late HST of sequence UC8/9 compared with the early HST and the TST of this sequence. The fine-grained detrital carbonate material was transported basinward from its source on the proximal ramp by wave or storm action and dispersed in the water column. On the distal ramp, it settled out from suspension with the hemipelagic background biogenetics.

The transgressive systems tracts of sequences UC8/9 and UC9/10 are composed mainly of marlstones and pelagic to hemipelagic mudstones (Fig. 7). The LSW of sequence UC9/10 is formed of carbonate grains that show no evidence of large-scale transport. The amount of terrigenous mud is low in the LSW of sequence UC9/10, relatively low in the HST of sequence UC8/9 and medium to low in the TST of this depositional sequence. The input of detrital carbonate is high in the late HST of sequence UC8/9 (Fig. 7). Therefore, the distribution of lithofacies resembles the model proposed by Haak and Schlager (1989, highstand-shedding) with important modifications, however. There is no pronounced series of calciturbidites in the HST of sequence UC8/9, probably due to the low relief (around 1–2 degree) even on the distal parts of the ramp. The occurrence of thick autochthonous lowstand carbonate successions can be easily explained by the basinward migration of the carbonate factory during relative lowstand on the carbonate ramp. The absence of a dominant mud component in the HST, contrary to the observation in the Cenomanian sequences of the BCB and the results from Robaszynski et al. (1990), is explained by the paleogeographic situation. The Upper Turonian HST carbonates on the proximal ramp extended far to the south (Floquet, 1991; Alonso et al., 1994) and prevented the shedding of terrigenous mud to the distal parts of the basin. A similar situation was also recognized in the Santonian period (Gräfe, 1994; Gischler et al., 1994).

The quantitative evaluation of the foraminiferal content in the Upper Turonian to Coniacian deposits shows distinct trends in sections of the outer ramp and deep basin (Fig. 8). Sequence UC 8/9 in the Villasana section on the distal part of the carbonate ramp exhibits a pronounced cyclicity with respect to the percentage of planktic and benthic foraminifers. The sequence boundary UC 8 is marked by a comparatively low planktic foraminiferal content which increases in the TST of this sequence up to the mfs. The amount of planktic foraminifers stays high in the early HST and then dramatically diminishes towards the sequence boundary UC 9 (Fig. 8). This is mirrored by the increase in skeletal and non-skeletal components around this SB, which reflects the basinward shift in shallow-marine lithofacies. Especially the middle part of the UC9/10 LSW is dominated by non-skeletal components (mostly pellets). The top of this LSW is formed by bryozoan-patchreef with a high amount of skeletal components. After the deposition of the lowstand wedge, the planktic foraminiferal fauna recovers partly. Therefore, maximum flooding surfaces in the Coniacian sediments are again characterized by high plankton content, and sequence boundaries are characterized by a relative decrease in planktic foraminifers with respect to the total foraminiferal fauna. Basinward of Villasana, the section Gordoa shows similar trends. A low plankton content is recorded around sequence boundaries and in lowstand wedges, but high contents of planktic foraminifers are recorded in the TST and HST and especially around the mfs. In lowstand wedges, the strong increase in the non-foraminiferal components of the samples is remarkable. In the Gordoa section, this non-foraminiferal content is composed mainly of sponge spicules and also of mollusc debris, which was obviously transported from shallower areas of the BCB into the deeper basin. In each lowstand complex the percentage in non-foraminiferal components increases up to the top of the LSW and then decreases sharply. In the TST and HST non-foraminiferal components are insignificant in the samples.

CONCLUSIONS

The results obtained from this study of Late Cretaceous sedimentary 3rd-order cycles in the BCB are summarized in Figure 3. Twenty Late Cretaceous sequence boundaries can be followed throughout the basin. The sequence stratigraphic model explains lateral variation of thickness and facies of sediments. High-resolution biostratigraphy was successfully applied in correlating sections and for correlation with other basins. Over-
all basin development is tectonically controlled by the opening and closure of the Bay of Biscay, while local tectonic phenomena were sorted out by basinwide correlation of surfaces.

The regional cycle chart of the BCB correlates with some sequence boundaries of the Haq et al. (1988) chart, but there are also important misties (e.g., sequence boundaries UC2, UC4, UC6, UC9). However, the correlation with more recent sequence stratigraphic results from Tunisia (Robaszynski et al., 1990, 1993) and Northwest-Europe (Juignet and Breton, 1992; Owen, 1996) is much better (Fig. 3).

The sequence stratigraphic interpretation of outcrops is dependent on depositional models and the correct interpretation of sedimentary facies. Therefore, generalized models for distribution of carbonate lithofacies in systems tracts (e.g., lowstand vs. highstand shedding) in a given depositional sequence should be treated with caution and must be modified for local depositional processes. The distribution of carbonate lithofacies in a depositional sequence depends on the geometry of the depositional realm (flat-topped platform vs. ramp setting), paleogeographic setting, climate, the amount of thermo-tectonic subsidence and the rate of absolute sea-level change.

The ratio between planktic and benthic foraminifers as well as the amount of skeletal components and non-skeletal components mirrors sequence stratigraphic surfaces (SB and mfs) and systems tracts. The planktic foraminiferal content is low around sequence boundaries, low to moderate in lowstand wedges and high in the TST, the early HST, and highest around the mfs. The amount of skeletal material is highest in the mfs of sections on the distal carbonate ramp and in the deep basin.

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OUTCROP CYCLE STRATIGRAPHY OF SHALLOW RAMP DEPOSITS: THE LATE CRETACEOUS SERIES ON THE CASTILIAN RAMP (NORTHERN SPAIN)

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ABSTRACT: Latest Albian to late Maastrichtian sedimentary deposits on the Castilian ramp (northern Spain) contain a great diversity of facies punctuated by numerous biostratigraphic discontinuities that reveal many paleoenvironmental changes and provide a precise record of stratigraphic cycles. These deposits represent a transgressive-regressive depositional megacycle (i.e., a very long-term cycle, 34 my). This late Cretaceous megacycle is subdivided into four long-term (4–20 my) transgressive-regressive depositional cycles: (1) latest Albian to middle Cenomanian cycle, (2) late Cenomanian to earliest Coniacian cycle, (3) early/middle Coniacian to late Santonian cycle and (4) latest Santonian to late Maaschrichtian cycle. These long-term cycles are made up of 13 well-characterized short-term transgressive-regressive depositional cycles, plus other cycles of shorter duration. Most of the cycles recognized correlate with those defined from the series of the steeper, distal part of the ramp and from the outer series of the deep basque Basin. The megacycle and the long- and short-term cycles were largely the result of the tectonic behavior of the Basque-Cantabrian passive margin and were closely controlled by the evolution of the bay of Biscay in conjunction with the relative movements of the Iberian and European plates.

AIM OF THE STUDY AND METHODOLOGY

The best way to pinpoint the factors that govern the cyclic pattern of sedimentation is to conduct an exhaustive analysis of regional cycle stratigraphy. The late Cretaceous deposits of the Castilian sedimentary basin in northern Spain are highly amenable to this type of analysis because:

1. the series crop out widely, so the three-dimensional arrangement of the lithostratigraphic units can be easily reconstructed;
2. the series can be studied across the full extent of the sedimentary basin, from shoreline to continental slope and from the time the basin formed; and
3. the sedimentary basin was a shallow epicontinental ramp on which almost all the environmental changes, especially those related to relative sea-level fluctuations, were very well recorded by both carbonate and siliciclastic sedimentation;
4. the biostratigraphic data are good enough to ensure precise dating of most of the lithostratigraphic units.

Thus, as inferred from the changes in the depositional environment revealed by the biosedimentary record, the various marine deepening-shallowing and transgressive-regressive depositional cycles that occurred on the Castilian ramp can be recognized with precision. By tracing changes in ramp development, it should be possible to distinguish between the impact of regional and external factors on the cyclic stratigraphic pattern.


The lithostratigraphic units of the northern and central parts of the Castilian ramp, were previously defined by Floquet et al. (1982) and described in detail by Floquet (1991). The lithostratigraphic units of the southern parts of the ramp were defined by Vilas et al. (1982), Canérot et al. (1982) and described by Meléndez et al. (1985), Calonge (1989), Carenas (1987), Gimenez (1987), Segura et al. (1993) and Martin-Chivelet (1992).

The stratigraphic cycles described are transgressive-regressive depositional cycles. They are defined from a succession of different paleogeographies recognized on the Castilian ramp. The detailed paleogeographic maps published by Floquet (1991) and the numerous paleogeographic sketches provided by Alonso et al. (1993) are used to define these cycles.

Stratigraphic cycles are ranked according to their relative duration. Therefore, very long-term, long-term, short-term and very short-term designate different cycles. There is no link between this rank-order and the genesis of the cycles.

The lithostratigraphic units and stratigraphic cycles are dated by a rich variety of fossils including ammonites, rudists and other bivalves, benthic and planktonic foraminifera, ostracods, echinoids, algae and gastropods. Detailed biostratigraphic data can be found in Floquet (1991, Figs. 283–301). Studies of specific fossil groups are given by Rodriguez-Lazaro (1985, on ostracods), Lopez-Sanjaume (1990, on inoceramids) and Santamaria-Zabala (1991, on ammonites).

The depositional systems tracts on the Castilian ramp are characterized here by geometry and facies associations and are defined on the basis of the stratal stacking pattern within a stratigraphic cycle: (1) landward-stepping systems tracts (i.e., retrograding or transgressive) are generally though not necessarily related to deepening of the depositional environment; (2) seaward-stepping systems tracts (i.e., prograding) are generally though not necessarily related to shallowing of the depositional environment and to a regressive trend; and (3) intermediate vertical stepping or aggrading systems tracts. This labelling is based on an analysis similar to the genetic stratigraphy method illustrated by Homewood et al. (1992). However, the boundary selected to separate depositional cycles corresponds to that between a shallowing (and regressive) trend and a deepening (and transgressive) one as in the definition proposed by Hunt and Tucker (1992). This boundary is particularly obvious in the Castilian ramp deposits.

When defining the depositional systems tracts in accordance with the fundamentals of sequence stratigraphy as postulated...
by Posamentier et al. (1988), Van Wagoner et al. (1988), Vail et al. (1991) and Duval et al. (1992) (i.e., referring to the location of the systems tracts relative to the shelf break), only so-called transgressive and highstand systems tracts are found on the Castilian ramp. So-called lowstand systems tracts lie in the Basque basin to the north, beyond the late Cretaceous Basque-Cantabrian shelf break (Fig. 1).

GEOLOGICAL SETTING

Beginning in the latest Albian time, marine gulfs encroached upon the Iberian craton from the north-north-west and from the south-south-east, between the Iberian massif or Meseta in the west and the Ebro massif in the east (Alonso et al., 1989, 1993). These gulfs developed to form the Atlantic ramp on one side of the craton and the Tethyan ramp on the other (Fig. 1). Under major marine flooding conditions, particularly from late Cenomanian to late Santonian times, the Atlantic ramp extended and joined with the Tethyan ramp to form a seaway known as the Iberian strait (Floquet, 1987, 1991; Alonso et al., 1989, 1993) across the Iberian craton (Fig. 1). During minor marine flooding episodes and marine regressions, the Atlantic ramp formed a wide embayment with no direct connection with the Tethyan ramp, which was confined to the southeasternmost areas.

The activity of the Iberian plate controlled the changing geography of the Iberian strait and its ramps (Floquet, 1991, 1992; Alonso et al., 1993). This strait gradually formed and developed through the first half of the late Cretaceous. The sizeable extension of the strait was the result of the southeastward movement of the Iberian plate and of the opening and distension of the bay of Biscay (Fig. 2). The strait and the ramps gradually disappeared in latest Cretaceous and early Tertiary times as compressive forces built up in response to the reversed northward movement of the plate (Fig. 2).

The Atlantic ramp was composed of a distal part, termed the Navarre-Cantabrian part, and a proximal part, termed the Castilian part (Fig. 3).

The Navarre-Cantabrian part of the ramp was relatively narrow and bordered on the north by the continental slope that dipped towards the deep marine Basque basin (Fig. 3) containing flysch deposits (Mathey, 1987; Razin, 1989; Schwentke, 1990, Schwentke and Kuhnt, 1992; Floquet et al., 1996). Thick sediments (up to 4000 m during late Cretaceous times alone) of marls and marly limestones, characteristic of deeper outer shelf environments, were deposited predominantly on this Navarre-Cantabrian part of the ramp (Amiot, 1982; Wiedmann et al., 1983; Schwentke and Wiedmann, 1985; Gräfe and Wiedmann, 1993; Gräfe, 1994).

The Castilian part of the ramp was wider and comprised varied depositional environments characteristic of several domains (the northern, central and southern domains on Fig. 3; Floquet, 1983, 1987, 1991, 1992; Floquet et al., 1996) ranging from outer shelf to coastal and even terrestrial environments. Detailed litho- and biofacies analysis of the series from each domain (Figs. 4–6) and the resultant reconstruction of depositional environments in space and time lead to recognition of transgressive-regressive depositional cycles in both marine and continental settings. Where the marine ramp extended widely so that environments under Atlantic influences covered not only the present-day Castilian regions but the southernmost areas of the present-day Iberian ranges too, the term Castilian ramp sensu lato (Fig. 1) is applied. The term Castilian ramp sensu stricto (Fig. 1 and 2) is used where the marine environments were confined to the present-day Castilian regions (“Castilla la Vieja and Castilla la Nueva”).

THE TRANSGRESSIVE-REGRESSIVE DEPOSITIONAL CYCLES

The biosedimentary record of relative sea-level megachanges, illustrated by a complete series for the northern domain of the Castilian ramp (Fig. 4), reveals a very long-term marine deepening-shallowing and transgressive-regressive depositional cycle lasting some 34 my, from latest Albian to late Maastrichtian times. This cycle is termed the late Cretaceous megacycle (LCMC). In the other domains of the ramp, the marine megacycle lasted from early or middle Cenomanian to latest Santonian or latest Campanian times, the differences depending on regional variations, as shown by the sections from the central domain (Fig. 5) and southern domain (Fig. 6). The latest Cretaceous continental deposits in the central domain (at least) are included in the megacycle because the changes in their depositional environments are analogous to the changes in the marine environments in the northern domain (Floquet, 1991).

The megacycle is subdivided, throughout the ramp, into four long-term marine transgressive-regressive cycles (LTC, Figs. 4–6) ranging respectively from latest Albian to middle Cenomanian times (LTC 1), from late Cenomanian to earliest Coniacian times (LTC 2), from early/middle Coniacian to late Santonian times (LTC 3), and from latest Santonian to late Maastrichtian times (LTC 4). The four long-term marine trans-
gressive-regressive cycles are arranged into two long-term marine transgressive-regressive cycle sets (LTCS, Figs. 4–6) ranging respectively from latest Albian to earliest Coniacian times (LTCS 1–2; a duration of about 10 my) and from early/middle Coniacian to late Maastrichtian times (LTCS 3–4; a duration of about 24 my). Each LTC is subdivided into a variable number of short-term marine transgressive-regressive depositional cycles (DC, Figs. 4–7).

**LTC 1 (latest Albian to middle Cenomanian times, circa 4.5 my)**

The depositional megacycle starts, throughout the study area, with thick siliciclastic deposits that overlie unconformably all the earlier rocks (from early Cretaceous to Paleozoic substratum). Thus, a major unconformity (cycle boundary, CB) marks the base of the megacycle (CB 1). Those terrigenous deposits form the continental Utrillas Formation consisting mainly of fluvial conglomerates, sands and clays. It is a diachronous lithostratigraphic unit, ranging in age from late Albian to Cenomanian times, or to Turonian time or younger towards the continental borders of the depositional area (Floquet, 1991; Alonso et al., 1989, 1993). The entire formation exhibits the same general sedimentological trend in time and space: fluvialite conglomerates and coarse sands prevail at the base and towards the proximal parts of the depositional area; fine-grained, often ligniferous sediments are concentrated at the top and towards the distal parts of the depositional area (Floquet, 1991).

Subsequently, a series of marine incursions from the Basque-Cantabrian margin encroached upon the central high that separated the Atlantic and Tethyan ramps (Alonso et al. 1989, 1993) while continental terrigenous facies continued to be deposited in the hinterland. The thickest marine formations (Valmaseda pro parte and Dosante Formations in Fig. 4, the latter up to 150 m thick) accumulated in the northern outermost domain of the Castilian ramp. Rudists (Ichthyosarcolithes, Polyconites, Caprina, etc.), planktonic and benthic foraminifera (Rotula, orbitolinids, praealveolinids) date the Valmaseda Formation (in this domain) as late Albian and early Cenomanian in age and the Dosante Formation as early and middle Cenomanian in age. The succession of depositional environments displayed by the Dosante Formation provides evidence throughout of a general marine transgressive trend, as illustrated by the typical section at Monte Humión (Fig. 8). However, these depositional environments remained very shallow because the high biogenic/bioclastic carbonate production and the continuing terrigenous supply were sufficient to fill the available space. The formations and corresponding depositional environments can be arranged into four short-term marine transgressive-regressive depositional cycles:
The Valmaseda pro parte, Dosante, Santa Maria de las Hoyas Formations, and the corresponding DC 1–4, are thought to form part of a mega-retrograding/transgressive systems tract at the scale of the megacycle because of their general backstepping arrangement.

The overall sedimentary development of the Utrillas Formation, judging from its mostly fining-upward depositional sequences, appears to have been linked to the general marine transgression (i.e., to have been a result of the beginning of the base level rise). Thus, the Utrillas Formation is included within the above-mentioned mega-retrograding/transgressive systems tract despite its continental facies and depositional environments. Only the unconformity at its base (CB 1), associated with intense erosion, is believed to be the expression of the base level maximum fall preceding the onset of the depositional megacycle.

**LTC 2 (late Cenomanian to earliest Coniacian times, circa 5.5 my)**

**DC 5 (late Cenomanian time),—**

A single extensive formation (Abejar Formation sensu Floquet, 1991; renamed Cabrejas del Pinar Formation by Alonso et al., 1993; and equivalent to the upper part of the Arceniega Formation in the Navarre-Cantabrian part of the Atlantic ramp) overlies the major unconformity CB 5 (Fig. 4). This formation overlaps far southward on to the sediments that accumulated on the earlier Tethyan ramp. It comprises micritic limestones with *Pithonella, Heterohelix, Rotalipora* and other planktonic foraminifers and ammonites (*Calycoceras naviculare* in the lower part, *Metoicoceras geslinianum* in the upper part) which give a late Cenomanian age. Facies and fossils indicate that this unit formed after a distinct deepening together with a broad marine transgression. The maximum flooding event occurred in the *Naviculare Zone*.

This major flooding event caused the Iberian strait to be under a prevailing Atlantic influence. The Castilian ramp extended (Castilian ramp sensu lato) and its environments overlapped the greater part of the Iberian strait.

Later, in the Geslinianum Zone, partial sedimentary filling and consecutive shallowing of the ramp occurred in relation to northward sedimentary progradation (top of the Abejar/Cabrejas de Pinar Formation), including rudist-bearing facies in the southern domain (Ciudad Encantada Formation in the Albacete and Valencia provinces; Gimenez, 1987, 1989; Alonso et al., 1993; Martin-Chivelet and Gimenez, 1993). This sedimentary regression ended with CB 6. Evidence of subaerial exposure can be found in the southernmost domain of the ramp (lower part of the Alarcón Formation in the Albacete and Valencia provinces; Gimenez, 1987; Martin-Chivelet and Gimenez, 1992), and on its western and eastern borders to the north (Floquet, 1991).

DC 5 is assumed to constitute the major retrograding/transgressive systems tract of LTC 2 on the Castilian ramp sensu stricto at least. Moreover, this late Cenomanian DC 5 should also be included in the mega-retrograding/transgressive systems tract at the scale of LCMC, in conjunction with the earlier DC 1–4 of LTC 1, as it is part of the same general backstepping trend.
Fig. 7.—Chart of the late Cretaceous transgressive-regressive depositional cycles on the northern domain of the Castilian ramp (modified from Floquet, 1992). Sequence boundaries (UC 1 to UC 20) of Gräfe and Wiedmann (1993) in the Navarre-Cantabrian part of the Atlantic ramp are also given.
DC 6 (latest Cenomanian to middle Turonian times).—

Micritic and marly/sandy limestones, together with marls rich in ammonites and planktonic microfauna (list in Floquet, 1991), constitute the Puente de Oyarzún, Santa Cruz del Tozo, Picoferentes and Monterde Formations (the latter modified sensu Floquet, 1991), (Figs. 4–7). These formations represent the reappearance of fully open marine environments and a new marine transgression during latest Cenomanian to early Turonian times. Ammonites date the basal flooding event as the Juddii Zone and the maximum flooding event as the Nodosoides Zone.

On the Castilian ramp sensu stricto, the DC 6 maximum flooding event also corresponds to the LTC 2 maximum flooding event (in this specific area, it overlapped the DC 5 transgression to the east and to the west). However, southward, on the Castilian ramp sensu lato, the DC 6 transgression (recorded in the Casa Medina Formation as defined by Vilas et al., 1982, in the southern Iberian ranges) does not appear to have reached the southernmost domain of the Atlantic ramp as did the DC 5 transgression. So, at the scale of the Iberian strait, this flooding event cannot be chosen as the maximum flooding event of LTC 2 and, obviously, of LCMC (see DC 9 and DC 10).

Besides, on the Castilian ramp sensu stricto, the DC 6 marine transgression coincided with the maximum deepening within LTC 2: the circalittoral environments encroached far onto the ramp. Furthermore, in this specific area, this deepening also corresponds to the maximum deepening within LCMC. Thus, it must be noted that, for the megacycle, maximum flooding and maximum deepening do not coincide.

In places, it seems that phases of slight progradation and shallowing on the Castilian ramp sensu stricto (i.e., weak regressive trends due to partial sedimentary filling of the available space) occurred successively at the end of early Turonian time (within the Nodosoides Zone) and in middle Turonian time (within the Ornatissimum Zone) and thus may correspond to very short-term and restricted transgressive-regressive depositional cycles (DC 6a-b-c, see Figs. 4, 7).

Northward biosedimentary progradation occurred on the ramp in early and middle Turonian times, as represented by the Barranco de los Degollados and Jaraba Formations in the southeastern domain, the upper part of Ciudad Encantada Formation in its type locality in the Cuenca province, the upper part of the Picoferentes Formation and the lower part of the Muñecas Formation in the central domain, the lower parts of the Revilla de Pomar y Hornillatorre Formations in the northern domain (see Figs. 4–7). The wide seaward-stepping systems tract progressively filled the available space and led to a sedimentary (normal) marine regression. This was obviously diachronic, and finally the southern and central domains emerged so that more than the half of the ramp was subaerially exposed at the end of middle Turonian time or at the beginning of late Turonian time (CB 7 in Figs. 4–7).

DC 7 (late Turonian to earliest Coniacian times).—

New marly or micritic limestones, locally sandy, bearing fauna from open marine environments are evidence of a further marine transgression in late Turonian time. These deposits form
the upper parts of the Hornillatorre and Revilla de Pomar Formations and the middle part of the Muñecas Formation (Figs. 4–7). The basal flooding event occurred during the Deverianum Zone, and maximum flooding and deepening can be dated from the Neptuni Zone (dating is based on the presence of Romaniceras deverianum and Subprionocyclus neptuni). But this transgression chiefly involved the Castilian ramp sensu stricto.

Mainly bioclastic and biogenic carbonates accumulated on the ramp during latest Turonian and earliest Coniacian times (dating is based on the occurrence of numerous rudists, see Floquet et al., 1982; Floquet, 1991). These deposits constitute, from the south to the north, the lower member of the Pantano de la Tranquera Formation, the middle part of the Muñecas Formation, the Villaescusa de las Torres and Cueva Formations (Figs. 4–7). The formations are arranged in a large northwestward prograding systems tract. The progradational events actually occurred successively in several very short-term depositional cycles (see Fig. 9). All these events involved progressive and consequently diachronous filling of the available space on the ramp and seaward shift of the shoreline. Finally, the whole ramp became subaerially exposed, probably during earliest Coniacian time. Erosion affected deposits at the top of the cycle (CB 8, Fig. 9) and might be related to a base level fall, meaning that a forced regression may have been superimposed on the normal sedimentary regression.

The upper part of DC 6 (6b-c) and the whole of DC 7 are thought to make up the large seaward-stepping systems tract of LTC 2 (Fig. 4).

**LTC 3** *(early/middle Coniacian to late Santonian times; circa 4 my)*

**DC 8** *(late early/middle Coniacian to early Santonian times).*

During the late early/middle Coniacian (Tridorsatum Zone) a major marine transgression flooded the ramp once more from the north. In the northern domain, bioclastic limestones and ammonite-bearing argillaceous limestones, constituting the lower parts of the Valle de Losa Formation and of the Nidaguila Formation, characterize the flooding phase. The fining-upward bioclastic limestones (Riberia Alta Member at the base of the Nidaguila Formation) form the retrograding systems tract (Fig. 4). The argillaceous limestones (Nidaguila Member sensu stricto) correspond to the DC 8 maximum flooding event and then to the base of a prograding systems tract (Fig. 4). The maximum flooding event is dated from the Margae Zone (just above the “Calcareñas intermédiaires” in Floquet, 1991). It also coincided with the maximum deepening of LTC 3 (Fig. 4).

In the central and western domains, fine-grained calcarenites (upper part of the Muñecas Formation = “Upper tidalites”, Fig. 5) form a retrograding systems tract. Pycnodonte-bearing micritic limestones (Hortezuelos Formation) first mark the maximum flooding event and then make up the lower part of a prograding systems tract.

In the eastern and southern domains, coastal sabkha deposits ( laminated dolomitic limestones of the middle member of the Pantano de la Tranquera Formation, Fig. 6; base of the Brechas dolomiticas de Cuenca Formation sensu Vilas et al., 1982) form a retrograding systems tract. Above, micritic lagoonal limestones (upper member of the Pantano de la Tranquera Formation) and dolomites (lower part of the Brechas dolomiticas de Cuenca) constitute the base of a prograding systems tract after maximum flooding (Fig. 6).

In the southernmost domains, very thin supratidal and tidal deposits (top of the Alarçon Formation and base of the Sierra de Utiel Formation) may represent the flooding phase. Above them, restricted lagoonal limestones (in the lower part of the Sierra de Utiel Formation; Gimenez, 1987; Martin-Chivelet, 1992 or in the middle part of the first sequence of this formation sensu Martin-Chivelet and Gimenez, 1992) could correspond to a maximum flooding event and to an early prograding systems tract.

Latest Coniacian and earliest Santonian deposits include, from the south to the north, lagoonal and intertidal-supratidal limestones of the lower part of the Sierra de Utiel Formation (top of the first sequence sensu Martin-Chivelet and Gimenez, 1992) and of the upper member of the Pantano de la Tranquera Formation (calcareous breccia after evaporite dissolution, in Floquet, 1991), calcarenites and rudist-bearing limestones of the lower parts of the Hontoria del Pinar Formation and Nocedo de Burgos Formation (lower part of the Tobalienilla Member), argillaceous limestones and fine calcarenites of the middle part of the Valle de Losas Formation (Figs. 4–7). These deposits correspond to a northward prograding bioseddimentary filling of the available space on the ramp. Such filling gave rise to a typical seaward-stepping systems tract and led to the partial shift of the shoreline towards the north and to the subaerial exposure of part of the ramp, thus partly generating CB 9.

DC 8 is subdivided, in the northern domain, into two shorter cycles: DC 8a from late early/middle Coniacian to middle/late Coniacian times (including the “Marnes inférieures” and the “Calcareñas intermédiaires” in Floquet, 1991); and DC 8b from late Coniacian to early Santonian times (including the “Marnes supérieures” and part of the Nocedo de Burgos Formation in Floquet, 1991). These cycles are separated by a weak
regressive trend and associated final discontinuity (at the top of the “Calcarénites intermédiaires”, CB 8’ on Fig. 10).

**DC 9 (early-middle Santonian to late Santonian times).—**

During early-middle Santonian times, a marine transgression from the Atlantic developed widely on the ramp. It is represented, from north to south, by marly limestones of the middle part of the Valle de Losas Formation, micritic limestones of the lower part of the Nocedo de Burgos Formation (including the upper part of the Tobalinilla Member that constitutes the retrograding systems tract in the northern domain; Figs. 4, 10, 11), fine-grained bioclastic calcarenites of the lower part of the Hontoria del Pinar Formation (Figs. 5, 6), and lagoonal biomicrites of the middle part of the Sierra de Utiel Formation (*Lacazina*-bearing first level, in Floquet, 1991; second sequence in Martin-Chivelet and Gimenez, 1992).

The maximum flooding event, above the Tobalinilla Member in the northern domain (Figs. 4, 10, 11), is dated from the Texanus Zone. At this time, the Iberian strait was wide open. The DC 9 flooding event is also regarded as the maximum flooding event of LTC 3 and perhaps also of LCMC (see DC 10), while deepening was moderate (less than during LTC 2).

In middle Santonian and beginning of late Santonian times, deposits included, from south to north, lagoonal and intertidal-supratidal limestones of the middle part of the Sierra de Utiel Formation, subtidal to supratidal limestones of the Cañadilla Formation, rudist-bearing biostromes and lagoonal micrites of the Burgo de Osma Formation (Floquet, 1982), bioclastic calcarenites of the upper parts of the Hontoria del Pinar and Nocedo de Burgos Formations, and bioclastic micrites and fine-grained calcarenites of the upper part of the Valle de Losas Formation (Figs. 4–7). All these deposits corresponded to a new northward prograding biosedimentary filling of the available space. Progradational events occurred as a chain of very short-term depositional cycles. Finally, the filling was complete and resulted in the subaerial exposure of nearly the whole ramp, partly generating the main final unconformity CB 10.

Taken together, these progradational events form the major seaward-stepping systems tract of both DC 9 and LTC 3 (Figs. 4–6). Erosion at the top of the depositional cycle concurred with the formation of the CB 10 (Figs. 12, 13) and indicates that a base level fall should have occurred and that a forced regression may have been superimposed on the DC 9 sedimentary regression.

**LTC 4 (latest Santonian to late Maastrichtian times, circa 20 my)**

**DC 10 (latest Santonian time).—**

In the northern domain, bioclastic calcarenites and argillaceous micrites (Tubilla del Agua Formation) indicate renewed flooding on the ramp. The fining-upward calcarenites (San Pantaleon de Losa Member at the base of the Tubilla del Agua Formation) constitute a retrograding systems tract above the basal unconformity CB 10 (Figs. 12, 13). The biomicrites and fine-grained calcarenites occur in numerous very short-term depositional cycles (each 1–10 m thick, Tubilla del Agua Member sensu stricto). These deposits corresponded first to the maximum deepening-flooding phase and then to the lower part of
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FIG. 12.—Landscape in the Sobrón gorges (eastern limb of the Sobrón anticline, above the road to the old village, uphill, long.: 3°6′11″ and lat.: 42°46′10″). Top of DC 8b (late Coniacian to early Santonian age), entire DC 9 (early to late Santonian age) and base of DC 10 (late Santonian age). Systems tracts and cycle boundaries. Note, at the base of DC 10, the thick and massive calcarenitic unit considered as an early retrograding/transgressive systems tract. R.: retrograding systems tract, P.: prograding systems tract, e.: early, l.: late.

FIG. 13.—Top of DC 9 (early to late Santonian age), DC 10 (late Santonian age) and DC 11 (Campanian age) in the Tubilla del Agua outcrops (type locality of the Tubilla del Agua Formation, long.: 3°48′20″ and lat.: 42°42′50″). Here, landwards from the Sobrón outcrops (Fig. 12), the retrograding/transgressive systems tract within DC 10 is thin. R.: retrograding systems tract, P.: prograding systems tract, e.: early, l.: late, M.D./F.: maximum deepening/flooding.

A prograding systems tract (Fig. 13). The maximum deepening-flooding event (MD/F on Fig. 13) is dated from the Syrtale Zone by foraminifera and ammonites (Floquet, 1991).

In the central and southern domains, coastal sabkha laminated and brecciated dolomitic limestones, sometimes with marls and evaporites (lower member of the Santo Domingo de Silos Formation, Figs. 5 and 6; lower part of the Fortanete Formation, base of the upper member of the Cuenca Formation, and Bascuñana Formation) are believed to represent this latest Santonian flooding phase. Subsequently, rudist-bearing limestones (middle member of the Santo Domingo de Silos Formation, upper part of the Sierra de Utiel Formation including the Lacazina-bearing second level in Floquet, 1991, or middle of the third sequence in the Sierra de Utiel Formation sensu Martin-Chivelet and Gimenez, 1992) are thought to represent the maximum flooding phase and the base of a prograding systems tract (Floquet, 1991).

The DC 10 maximum flooding event is also the maximum such event of LTC 4. It appears also to have been a major flooding event for LCMC, at least on the Castilian ramp sensu stricto, with nearly the same extent as the DC 9 flooding event. However, the corresponding deepening was slight.

Terrigenous deposits were laid down at the end of Santonian time, especially in the northern domain. They appear in the upper part of the Tubilla del Agua Formation and constitute part of the Moradillo de Sedano Formation, including the lower part of the Ríoseco Member (Fig. 4). They form the upper part of the above-mentioned prograding systems tract and correspond to a sedimentary marine regression. Locally, and particularly on the northwestern border of the ramp, these siliciclastic deposits are arranged in a clearly eastward-prograding deltaic system (Ríoseco delta in Floquet, 1991).

DC 11 (early Campanian to earliest Maastrichtian times).—

A further marine flooding event during early Campanian time was restricted to the northern and central domains of the ramp. In the northern domain, the flooding phase is marked by rudist-rich limestones of the Quintanaloma Formation. Directly overlying deltaic dolomitic red sandstones of the Moradillo de Sedano Formation and above a sedimentary discontinuity (CB 11, Fig. 4), the development of coral reefs (lower member of the Quintanaloma Formation, Fig. 13) is evidence of the basal
floodling event. The rudist fauna indicates early Campanian age (Floquet, 1991 and Floquet, unpubl. data).

In the newly settled embayment, rudist buildups, especially biostromes (upper member of the Quintanaloma Formation), were deposited throughout Campanian time. Numerous relative sea-level fluctuations, obvious in such shallow-water environments and recorded by very short term depositional cycles, occurred over this long period (Floquet, 1991, unpubl. data). Unfortunately, the lack of biostratigraphically significant faunas and floras results in difficulty in dating them more accurately than middle to late Campanian in age. This upper member of the Quintanaloma Formation forms the lower part of a large prograding systems tract and reflects a general regressive trend connected with overall biosedimentary filling of the available space on the ramp. To the west and laterally equivalent to the Quintanaloma limestones, coarse sandstones and conglomerates, constituting the upper part of the Rioseco Member, were deposited in the eastward prograding Rioseco delta.

In the central domain, the Campanian marine deepening-flooding is not so distinct. It is presumably expressed in the reappearance of the last rudist-bearing facies in the upper member of the Santo Domingo de Silos Formation (Fig. 5).

It seems that the fully marine environments failed to reach the southeastern, southwestern and southern domains. Only intertidal to supratidal (including palustrine) environments occurred there at this time, as recorded in the Sierra de la Pica Formation, in the upper part of the Fortanete Formation, in the lower part of the Villalba de la Sierra Formation, and in part of the Perenchiza Formation (Floquet, 1991).

Terrigenous sediments were deposited right across the ramp during earliest Maastrichtian time (probably since Campanian time on the borders of the ramp). They form:

1. the upper part of the Villalba de la Sierra Formation in the southernmost domain, the Embid de Ariza Member (sensu Floquet, 1991) of the Santibañez del Val Formation (Fig. 6), and the base of the Bea-Fonfría-Segura de los Baños Formation (sensu Babinot and Moissenet, 1985) in the southeastern domain, mostly composed of sands and sometimes conglomerates that represent distal to proximal fluvial environments;
2. the base of the Valdelacasa Member and the Castroceniza Member of the Santibañez del Val Formation in the central domain (Fig. 5), mainly composed of red sands that represent distal fluvial flood plains; and
3. the Sedano Formation in the northern domain, mostly composed of oyster-bearing silty clays with occasional conglomeratic sands (Escan˜o outcrops) and of dolomitic limestones, which represent subtidal to supratidal coastal environments (Fig. 4).

These siliciclastic sediments clearly prograded northward and are believed to constitute the upper part of the above-mentioned large prograding systems tract of DC 11. Progradation led to a net marine regression, clearly shown by the gradual and diachronous northward shift of the shoreline and by the spread of continental environments across the main part of the vanishing marine ramp.

DC 12 (early Maastrichtian to middle/late Maastrichtian times).—

In the northern domain, above CB 11 at the top of the Sedano terrigenous coastal facies, rudist buildups and associated carbonitic bars (Valdenoceda Formation) formed in shallow subtidal environments during early-middle Maastrichtian times (dated by mean of rudists). These facies express a marine transgression (Figs. 4, 14). Thin white sandstones or sands, in lenticular bodies, underlie the biogenic limestones locally; they are interpreted as representing a transgressive lag.

In the central domain, lacustrine-lacustrine limestones were deposited during this time (La Yecla Member of the Santibañez del Val Formation in the Santo Domingo de Silos area, Fig. 5; and upper part of the Valdelacasa Member of the same Formation in the Talveiga-Aranga area). The development of lakes and marshes may have been a result of a base level rise and subsequent damming of fluvial outlets, related to the marine flooding phase in the north (Floquet et al., 1993). Terrestrial gastropods and dinosaur eggs confirm a Maastrichtian age.

During middle/late Maastrichtian times, variegated clays deposited in coastal plains (the Sobrepeña Formation, westwards) and palustrine-lacustrine limestones (the Ircio Formation, eastwards, with terrestrial gastropods providing an age) formed in the northern domain (Fig. 4). Fluvial sands or conglomerates (part of the Las Porqueras Member and El Colmenar Member of the Santibañez del Val Formation) continued to be deposited in the proximal domains (Fig. 5). Thus, a further marine regression occurred with northward progradational filling with terrigenous continental and littoral facies. This seaward-stepping systems tract in the northern domain is truncated (erosion of the variegated clays of the Sobrepeña Formation, for example) and emphasizes CB 13 at the top of DC 12.

DC 13 (late Maastrichtian time).—

A slight marine transgression is recorded in late Maastrichtian time, but only in the northernmost domain of the ramp. Deposition of calcareous sandstones and of Orbitoides apiculata, O. cf. gensacicus and Omphalocyclus macroporus-bearing limestones to the north and west (Torme Formation) as well as deposition of evaporite-bearing dolomicrotes to the south and east (Bozóo Formation sensu Floquet, 1991) mark these re-
stricted flooding event (Fig. 4). Conglomerates and sandstones form the base of the Torme Formation; they are regarded as representing a transgressive lag. The Bozó Formation is attributed to late Maastrichtian time because of the occurrence of gastropods such as Lychnus giganteus, Palaeostoa hispanica, Proalbinaria cf. matheroni and Dissostoma sp. (Floquet, 1991).

Later, in this northernmost domain, coastal dolomitic marls, clays and sands (upper part of the Torme Formation) accumulated. These deposits correspond to a prograding and regressive systems tract.

No marine sedimentation occurred at that time in the central domain. Only fluvial sands and conglomerates (part of the Las Porqueras Member and El Colmenar Member of the Santibañez del Val Formation) continued to be deposited there. Part of this terrigenous sedimentation is believed to represent the continental prograding systems tract of DC 13.

The upper part of DC 10 to the end of DC 13 is thought to form the very large seaward-stepping systems tract of both LTC 4 and LCMC (Fig. 4). Strong erosion emphasizes the contact between DC 13 (or even the underlying depositional cycles) and DC 14 or DC 15.

In the northernmost (outermost) domain, dolomitic calcarenites which include marine bioclastic remains (Fresnedo Formation sensu Pluchery, 1995; and sensu Floquet and Pluchery, unpubl. data), forming part of DC 14 (14a on Fig. 4) and attributed to Danian time, overlie an erosional surface at the top of the Torme Formation (Fig. 4).

Further south in the northern domain, the succession varies:

1. the Fresnedo Formation (DC 14a) directly overlies the green clays of the Sobrepeña Formation (DC 12). The Torme Formation is absent (Pluchery, 1995), and there is a stratigraphic gap of about 3 my (Urria section, Fig. 15);
2. sandy and bioclastic marine calcarenites, dated as early Thanetian age and which constitute the Landraves Formation (Pluchery, 1995), and form part of DC 15, directly overlie the dolomiticites of the Bozó Formation (DC 13) so that the Fresnedo and Escaño Formations are absent with a stratigraphic gap of some 6–7 my (Sobrón section, in Floquet, 1991; Fig. 16); and
3. the Landraves Formation (DC 15) directly overlies the Sobrepeña Formation (DC 12) indicating the absence of the Torme, Fresnedo and Escaño Formations and a stratigraphic gap of some 8–9 my (Frias section, in Floquet, 1991 or Valdenoceda section, Pluchery, 1995).

Erosion indicates the influence of a base level fall (probably due to tectonic uplift, see Floquet, 1991; Pluchery, 1995; Floquet and Pluchery, unpubl. data) which induced a forced regression superimposed upon the prior normal regression at the end of DC 13. Above the erosional surface, the Fresnedo Formation represents a marine flooding event that slightly onlapped the previous DC 13 marine transgression, while the Landraves Formation marks a new marine flooding phase that widely overlapped the previous one.

Accordingly, CB 14 at the top of DC 13 is regarded as a major unconformity and is chosen to separate the megacycle LCMC and the next megacycle, the early Paleogene depositional megacycle EPMC (Pluchery, 1995; Floquet and Pluchery, unpubl. data).

**CONTROLS ON THE DEVELOPMENT OF THE TRANSGRESSIVE-REGRESSIVE CYCLES**

*The Tectono-Eustatic Framework*

Since there is no irrefutable proof of glaciation during the middle and late Cretaceous period because of greenhouse effect and globally warm climates (see Barron, 1983; Sloan and Barron, 1990; Spicer and Corfield, 1992), it is difficult to refer to glacio-eustasy as a control for sea-level changes. Further, the worldwide increase in the rate of ocean crust formation between 120 and 80 Ma, and particularly the middle Cretaceous “superplume” (see Larson, 1991; Garzanti, 1993), undoubtedly did cause a rise in sea level. The significant downturn in the rate at
which ocean crust was forming about 80 Ma (Larson, 1991) presumably resulted in a sea-level fall. Note that, from this time, the marine ramps on the Iberian craton were disappearing. Thus, tectono-eustasy related to the evolution of the ocean ridges, and especially of the nearby Atlantic ridge, may have governed the main trends in sea-level fluctuations.

The Tectonic Control

Within the supposed tectono-eustatic framework, referred to above, it is assumed that: (1) the marine flooding events that affected extensive areas of the Iberian craton and the subsequent origination and development of the marine ramps from late Albian time through the first half of late Cretaceous times were mainly due to sizeable downwarps of the Atlantic margins of the Iberian plate in conjunction with a generally rising sea-level; and (2) the progressive disappearance of the marine ramps in latest Cretaceous times resulted chiefly from the restriction of the downwarps of the margins and from the generally falling sea-level.

(1) From late Albian times onwards, the Utrillas Formation plus a part of the Valmaseda Formation in the Navarre-Cantabrian depositional area and the upper part of the Black Flysch Group in the Basque basin formed a thick and extensive siliciclastic cover that buried early Cretaceous sedimentary basins. This terrigenous cover created a new flat topography on which the first late Cretaceous marine ramp could develop. The appearance of this totally new paleogeography was a consequence of general downwarping, corresponding to the individualization of the late Cretaceous Basque-Cantabrian margin. This downwarping is believed to indicate a post-rift thermal subsidence stage and to be linked to pronounced distension between the Iberian and European plates. The spreading of the siliciclastic facies is regarded as the sedimentary response to the post-rift break-up (Rat et al., 1982; Floquet and Pascal, 1996), which sealed the syn-rift “Urgonian” depositional complex (late Aptian to late Albian times) in the Basque-Cantabrian gulf (Rat et al., 1982; Pascal, 1985; Garcia-Mondejar, 1990; Garcia-Mondejar and Fernandez-Mendiola, 1993).

From that time and for a large part of the late Cretaceous times, thermal subsidence markedly controlled changing sedimentary patterns on the Iberian plate. First, the marine ramps could settle while this plate was sinking. Further, changes in the rate of subsidence of the Basque-Cantabrian margin could have partly brought about the four long-term transgressive-regressive depositional cycles on the Castilian ramp.

The progressive and general relative sea-level rise that lay behind LTC 1 seems to have resulted mainly from increasing regional subsidence as the Basque-Cantabrian margin sank as a result of the net distension associated with the southeastward movement of the Iberian plate. The alkaline volcanism (spilites and pillow lavas), characteristic of thinning continental crust (Azambre and Rossy, 1976; Rossy, 1988) that occurred in the Basque basin from late Albian time (Lamolda et al., 1983) is evidence of extensional movements. Changes in subsidence rates also could have partly generated the successive short-term DC 1–4. Thus, the movement of the Iberian plate and the dynamics of the Basque-Cantabrian margin, combined with the tectono-eustatic sea-level rise, are believed to have induced the short-term depositional cycles.

The extensive and homogeneous marine flooding during late Cretaceous time (DC 5), at the base of LTC 2, could be specifically related to a tectono-eustatic sea-level rise. The fact that such flooding was mirrored in other regions (in southwestern Europe, Africa, etc.; see Floquet, 1984) points to a partly eustatic control. A tectono-eustatically-induced sea-level rise is presumably connected with intense volcanic activity along the mid-Atlantic ridge. However, a general sinking and a northward tilting of the Iberian plate were likely, given the large decrease in the areas of Tethyan ramp on the southern side of the plate while the Atlantic ramp was increasing (Garcia et al., 1985; Floquet, 1987, 1991; Alonso et al., 1993). Such tilting certainly favored Atlantic flooding. Latest Cenomanian-earliest Turonian and early Turonian net deepenings (within DC 6) seem also to have been partly induced by tectono-eustatic sea-level rises. Most workers recognize eustatic sea-level rises at a worldwide scale for this period. But in the case of the Castilian ramp, they appear to have been associated with new northward tiltings of the Iberian plate. A reduction or lack of subsidence seems to have favored the sedimentary infilling in early/middle Turonian times and the resultant exposure of the ramp almost everywhere near the end of middle Turonian time (top of DC 6). A further change in subsidence rate (and renewed tilting) could have generated the late Turonian and earliest Coniacian DC 7 (alkaline volcanism activity was still intense at that time in the Basque basin, Amiot et al., 1982; in conjunction with distension between the European and Iberian plates). The LTC 2 boundary (CB 8) resulted in part from a relative sea-level fall, probably related to tectonic uplift. Positive movements due to transpression have been recognized for that time in the eastern Pyrenees (Souquet and Deramond, 1989) and in the Iberian-Betic area (Vilas et al., 1983). The deposition of a megaturbidite in flysch deposits of the Basque basin at the same time (Offroy, 1984; Mathey, 1987, Razin, 1989) could indicate such movements, since megaturbidites are sometimes interpreted as seismiturbidites derived from highly mobile margins (Mutti et al., 1984).

The origin and development of LTC 3, and of its short-term cycles (DC 8a–b, 9), and the origin of LTC 4 are similarly regarded as consequences of changes in subsidence rates, also related to the dynamics of the Basque-Cantabrian margin and to the movement of the Iberian plate. It is believed that the extension between the European and Iberian plates reached its maximum during early and middle Santonian times (Mathey, 1987), during the third long-term cycle. This corresponded to the widespread development of carbonate flysch sedimentation in the Basque basin (Mathey, 1987; Razin, 1989) and thus to the main downwarps of the Basque-Cantabrian margin. During the same period, the Iberian strait was wide open (at its maximum during early Santonian time, at the beginning of DC 9) and dominated by Atlantic influences (Floquet, 1991, Alonso et al., 1993).

The southeastward drift of Iberia, from Albian to Campanian times (black arrows on Fig. 2), resulted in the opening of the bay of Biscay and in a gap of at least 100 km between Iberia and Europe (Olivet et al., 1982). It is thought that the bay of Biscay stopped widening during Campanian time, between 80 Ma and 75 Ma (this event is dated by the occurrence of magnetic anomaly 34 and not of magnetic anomaly 33 in the bay of Biscay, Boillot, 1984). The reversal in the direction of drift, to the north-northwest, of Iberia (as a consequence of the north-
eastward drift of Africa and of the northward drift of Apulia; white arrows on Fig. 2) must have been pronounced from this time onward.

Cumulative thickness curves for the different series in the various domains of the Castilian ramp show a strong inflexion at about 84 or 83 Ma (during latest Santonian time or at the Santonian-Campanian boundary). In the northern domain, the apparent deposition rate decreased by about 500% after this time (Fig. 17). This could be roughly regarded as the result of a decrease in the subsidence rate possibly connected with the compressive movements that affected the hinterlands of the Castilian ramp on the Iberian craton. Such movements might have occurred around 84 or 83 Ma, before the bay of Biscay was completely open: compressive movements are recognized at that time in the eastern Pyrenees (Bilotte, 1985). On this assumption, from this time onwards, the northwestern convergence of Iberia towards Europe should have greatly affected the development of LTC 4. From latest Santonian time, the marine ramps became progressively more restricted until they almost completely vanished at the end of LCMC (the marine ramps finally disappeared in the early Paleogene depositional megacycle; Pluchery, 1995; Floquet and Pluchery, unpubl. data). The large terrigenous deposits that characterized LTC 4 certainly derived from these movements. The first significant siliciclastic input occurred in latest Santonian time (at the end of DC 10, lower part of Moradillo de Sedano Formation) and the next inputs became larger and coarser during Campanian time (upper part of the Moradillo de Sedano Formation = Riotoso Member and deltaic system pro parte within DC 11). The erosional events that underline the cycle boundaries seem to have been successively more intense from CB 12 to CB 14 and are associated with more plentiful and coarser siliciclastic deposits. In the southern domain and especially on its eastern border, tectonic activity was probably strong: conglomerates, presumably Maastrichtian or even late Campanian in age, were laid down unconformably after erosion, above the sabkha and supratidal deposits from the last marine transgression, which is thought to have occurred during latest Santonian time.

Further, the widespread appearance throughout the Basque basin of siliciclastic flysch sedimentation, including large debris flows, in early Campanian time (Mathey, 1987) were surely consequences of the initiation of the compressive movements to the east.

While the general trend of a relative sea-level fall was enhanced by the tectonic behavior of the Iberian plate, further downwarps and changes in the rate of subsidence of the Basque-Cantabrian margin, coupled with Atlantic tectono-eustasy, gave rise to and controlled the short-term depositional cycles of LTC 4. However, these downwarps appear to have been successively weaker and the depositional areas of the marine cycles shifted northwards (i.e., basinwards).

The Role of Sedimentation and Physiography

The physiography of the ramp seems to have been important in the development of depositional cycles. Physiography was linked to the tectonic behavior of the ramp (differential subsidence, increasing northwards) and also to the state of the sedimentary filling. Partial filling produced residual depositional morphologies while total filling produced a flat morphology. This was the main reason why maximum deepening often failed to coincide with maximum flooding (Fig. 17).

Thus, successive onlaps of the DC 1–4 on the gently sloping northern domain, gradually built a virtually flat ramp over which the DC 5 marine transgression (Naviculare Zone) was able to extend far to the south, with no major increase in depth. As a result, this DC 5 flooding event appears to have been the maximum flooding episode of LTC 2 on the Castilian ramp sensu lato. On the Castilian ramp sensu stricto, the flooding events during DC 6 (Judiddi Zone, Nodosoides Zone) overlapped the DC 5 event and were linked to the maximum increases in depth within LTC 2 in this area. However, on the Castilian ramp sensu lato, the DC 6 flooding events were not the maximum flooding events of LTC 2 and a fortiori of LCMC. This was brought about by the northward tilting of the Iberian plate, which made the ramp steeper and so limited marine transgressions southwards. Analogous phenomena occurred in late Turonian and early-middle and late Coniacian times.

The near-complete sedimentary infilling of the available space on the ramp in early Santonian time (end of DC 8b) gave rise to a very flat physiography over which the DC 9 shallow marine transgression extended widely. So, the DC 9 flooding event appears to have been the maximum flooding event of the Iberian strait for LCMC. The total sedimentary filling at the beginning of late Santonian time (end of DC 9) led to a new flat topography over which the DC 10 marine transgression spread in much the same way. Consequently, the DC 10 flooding event was the maximum such event on the Castilian ramp sensu stricto.

Climatic Control and Biologic Productivity

Climate was important in the development of the depositional cycles on the Castilian ramp. The subtropical position of the Iberian plate provided generally warm and humid conditions suitable for high biogenic production and, consequently, a high growth potential of the carbonate ramp. However, local climatic gradations led to major differences in sedimentation. For example, wetter conditions on the western border favored terrigenous supplies and deltaic systems. By contrast, more arid conditions towards the eastern border of the ramp favored pure carbonate sedimentation, including sabkha deposits, further encouraged by the flat topography.

Rudists and other bivalves, foraminifera, algae, gastropods and corals made up a large part of the carbonate factory and were particularly effective in building the prograding systems tracts on the ramp (the best examples are from the upper parts of DC 7, 9 and 11). This factory was also largely responsible for the thick series of peloidal (fine-grained bioclastic) mudstones/wackestones to micropackstones deposited on the distal or Navarre-Cantabrian part of the ramp.

Because a large proportion of the sediments are bioclastic material and because available space for sedimentation was mainly created by tectonics, the name of tecto-biogenic systems has also been applied (Floquet, 1987, 1991) to the long-term depositional cycles.

The cooling that Francis and Frakes (1993) argued occurred during latest Maastrichtian time and could have led to the formation of the polar ice caps, may have enhanced the sea-level fall.
Fig. 17.—Main changes of sea depth (right curve), main coastal onlaps and offlaps (central curve) and cumulative deposit thickness versus time (left curve) during the late Cretaceous depositional megacycle in the northern domain of the Castilian ramp. Note: (1) imprecise dating of some cycle boundaries (CB 2–4, 9–11 and particularly 12–15 during Maastrichtian and Danian times with a possible 2–3 my margin of error as indicated by arrows); (2) occurrence of erosion “Er”, especially at the main cycle boundaries, inducing major uncertainties in plotting a thickness curve and, a fortiori, a subsidence curve; and (3) a strong inflexion of the thickness curve at about 83 Ma which probably expresses a tectonic event. “Eu” means eustatic sea-level rise. 1-2-3-4, close to “Coniacian”, indicate the successive Petrocoriense, Tridorsatum, Margae and Serratormarginatus ammonite zones.
CONCLUSIONS

Outcrop cycle stratigraphy of the sedimentary series deposited on the Castilian ramp in late Cretaceous times based on the identification and ordering of marine deepening and shallowing trends and of transgressions and regressions provides the following main results:

1. A transgressive-regressive depositional megacycle extended from latest Albian to late Maastrichtian times and can be subdivided into 4 long-term and 13 short-term transgressive-regressive depositional cycles (not including shorter term cycles).

2. Maximum deepening and maximum flooding do not always coincide. It would therefore be wrong to argue for a so-called global eustatic sea-level curve on the basis of deepening or shallowing in outer basinal settings and of movements in coastlines.

3. Regional tectonics exerted a major control over the origins and development of the transgressive-regressive cycles and compounded or reduced the effects of Atlantic tectono-eustatic sea-level changes.

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SEQUENCE STRATIGRAPHY IN THE UPPER CRETACEOUS SERIES OF THE ANGLO-PARIS BASIN: EXEMPLIFIED BY THE CENOMANIAN STAGE

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ABSTRACT: Five Cenomanian successions, situated in contrasting positions within the Anglo-Paris Basin and on its margins, are described in detail, and their sequence stratigraphies analyzed and compared. The Cenomanian Stage was chosen specifically for this study because of the high biostratigraphical resolution (about 0.5 myr/zone/subzone) achievable for this stage by the use of ammonites. The successions in Kent (south east England) and the Boulonnais (north east France) are the most basinward in position and comprise rhythmically bedded argillaceous micrites (ch racks and marls) with few major hiatuses. The thinner succession of coarser carbonates in Normandy (north west France) contains larger gaps. The succession in Maine (north west France) was deposited in relatively shallow water near the basin margin and includes several sand bodies and well-developed hardgrounds. The highly condensed succession of sandy limestones, containing major hardground-bounded hiatuses in Devon (south west England) represents deposition closest to the basin margin.

Systems tracts are characterized and defined by diverse criteria, including geometry of sediment bodies, basin margin onlap, discontinuity surfaces (omission surfaces and hardgrounds) and lithological characters, including variations in quartz sand content, clay:carbonate ratios and the presence of authigenic glauconite and phosphate. Detailed comparison of the successions enables us to demonstrate the existence of five basinwide sequences and the lower part of a sixth.

INTRODUCTION AND OBJECTIVES

The concept of sequence stratigraphy has been developed over the last 10 years, and the essential principles summarized in several papers (Vail et al., 1987; Van Wagoner et al., 1988). The pattern of eustatic sea-level change responsible for the control of post-Paleozoic sedimentary sequences has been shown in cycle charts developed mostly from the study of stage stratotype areas (Haq et al., 1987, 1988).

One of the problems encountered in the study of Upper Cretaceous stage stratotype regions is that they are mostly marginal in position, and markedly incomplete (Birkelund et al., 1984); thus, one cannot assume that all the constituent sequences are represented in these regions. The Turonian Stage serves well as an example of these difficulties. In the type region (Touraine, western France), a detailed study of the exposures, for which the litho and biostratigraphy has been thoroughly documented (Robaszynski et al., 1982) lead to the recognition of three sequences (Haq et al., 1987). However, the Upper Turonian is here developed in a sandy facies unsuitable for the recognition of sequences. Later work on an expanded basinal Turonian succession in central Tunisia demonstrated the presence of a fourth sequence in the upper part of the Turonian Stage (Robaszynski et al., 1990).

A central aim of the present work is to test the synchronicity and lateral extent of the sequences present in the Cenomanian Stage of the Anglo-Paris Basin. The Cenomanian Stage lends itself particularly well to this type of study, as the biostratigraphical resolution for this stage is very high. We have compared successions on different massifs bordering the basin, in the basin itself, and intermediate situations, in a variety of facies.

Previous Work

Cenomanian sea-level changes in the Anglo-Paris Basin have long attracted the attention of geologists. For example, Jukes-Browne (1902) noted the widespread “overlap” (onlap) of transgressive Cenomanian deposits. Important recent discussions of Cenomanian sea-level changes are found in Hancock (1989) for southern England, and Juignet (1980) for Normandy and Maine. Sequence stratigraphic treatment of these deposits has been undertaken recently by Juignet and Breton (1992), Robaszynski and Amédro (1993) and Hart et al. (1992).

Depositional Sequences

For the purposes of this study, we have chosen well-studied fossiliferous successions that fall in three contrasting regional positions (Fig. 1):

1. The central basinal successions of Kent (south east England) and Boulonnais (north east France), characterised by relatively continuous deposition of rhythmically bedded hemipelagic chalks and marls; major hiatuses are present only in the Lower Cenomanian Stage.
2. The basin margin setting of Normandy and Maine (north west France), which comprise fairly thick successions of mixed carbonate-clastic sediments, containing many hardground-bounded hiatuses.
3. The highly condensed carbonate-clastic succession in Devon (south west England) with few very major breaks marked by strongly developed hardgrounds.

Biostratigraphical Correlation

The Cenomanian Stage was selected for this study partly because an exceptionally high level of biostratigraphical reso-
The zonation of the Cenomanian Stage has been developed over the past three decades, largely through the publications of Hancock (1960), Kennedy (1969), Kennedy (1971) and Wright et al. (1984).

Stratigraphical Distribution of Ammonites in Different Regions

For Normandy and Maine, the records of ammonite species shown in Figure 2 are based on Juignet (1974), Juignet and Kennedy (1976), Juignet et al. (1978) and Kennedy and Juignet (1981, 1983, 1984, 1991, in press); in the Boulonnais (Figure 3) from Amédro et al. (1978), Robaszynski et al. (1980), Amédro (1986), Robaszynski and Amédro (1986), and Amédro (1994); for Kent (Figure 4), the work of Kennedy (1969) and Gale (1989, 1995).

Integration of the data from the three regions allows us to produce a composite table giving the comparative vertical ranges of the main ammonite species (Figure 5). We subdivide the Cenomanian into 8 well characterized zones of variable duration (cf. Gale, 1990), which are enumerated below in ascending order.

Lower Cenomanian.—

Mantelliceras mantelli Interval-Zone (base coincident with the base of the Cenomanian): from the appearance of M. mantelli to that of M. dixoni. This zone can be subdivided into a lower Neostlingoceras carcintanense Subzone and an upper Mantelliceras saxbii Subzone.

Mantelliceras dixoni Interval-Zone: from the appearance of M. dixoni to the appearance of Cunningtoniceras inerme.

Middle Cenomanian.—

Cunningtoniceras inerme Interval-Zone: from the appearance of C. inerme to that of Acanthoceras rhomagense.

Acanthoceras rhomagense Interval-Zone: from the appearance of A. rhomagense to that of A. jukesbrownei. The zone is divided into two subzones, successively those of Turrilites costatus and T. acutus.

Acanthoceras jukesbrownei Total Range Zone.

Upper Cenomanian.—

C. naviculare Interval-Zone: from the disappearance of A. jukesbrownei to the appearance of Metoicoceras geslinianum. In France, C. naviculare is a appropriate index, but in England this species is only certainly recorded from the overlying zone and C. guerangeri is preferred as an index.

Metoicoceras geslinianum Total Range Zone.

Neocardioceras juddii Interval-Zone: from the disappearance of M. geslinianum to the appearance of Watinoceras devonense.

Base of the Turonian.—

This is now generally taken at the entry of Watinoceras devonense (see Kennedy and Cobb, 1991), which is often associated with the appearance of other species of the same genus and coincides approximately with the appearance of the inoceramid genus Mytiloides.
FIG. 2.—Ranges of ammonites in the Cenomanian succession of Normandy plotted against marker surfaces and ammonite zones (after Juignet and Kennedy, 1976, modified from subsequent data).
FIG. 3.—Ranges of ammonites in the Cenomanian succession of the Boulonnais, and ammonite zonation (numbers and letters after Amédro, 1986, 1993, with modifications).
Fig. 4.—Ranges of stratigraphically important species of ammonites in the Cenomanian of Kent and Sussex. Open circles represent phosphatised remanié occurrences (after Gale, 1989 and unpub. data). See Figure 6 for explanation of abbreviations.
3. Horizons of calcisiltite and calcarenite (inoceramid prisms, calcispheres, echinoderm debris): Represent current winnowing and contain intraclasts (variably mineralized), granular glauconite and sometimes quartz sand. Such lithologies are commonly and characteristically associated with transgressive systems tracts.

4. Geometry and relationships of individual sedimentary units: Erosional surfaces cut progressively deeper, and often channel into underlying sediments towards basin margins, and onlap is observable in the overlying sediments.

In addition to these major sedimentary criteria, a diversity of accessory data help to define systems tracts:

1. Geochemical and mineralogical features: Include variation in illite/smectite ratios, which may be related to proximity to provenance and variance in δ13C which can be related directly to the eustatic curve (Jenkyns et al. 1994).

2. Palaeoecological information: Includes the abundance of larger planktic foraminiferans (cf. Hart et al., 1992) and the occurrence of discrete trace fossil assemblages. Care must be taken here because other factors such as temperature, oxygenation and water mass distribution may be equally important controlling factors with depth.

The practical approach adopted here for identifying sequences is by initial recognition of the relatively conspicuous
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Introduction

In the north of the Anglo-Paris Basin, Cenomanian deposits comprise 60–80 m of rhythmically bedded hemipelagic marls and marly chalks, which are well exposed in sealifts facing each other (40 km apart) across the Straits of Dover/Pas de Calais, between Cap Grand Blanc-Nez and Petit Blanc-Nez in the Boulonnais and between Folkstone and Dover in east Kent. The successions are extremely similar on both sides of the Channel and can be correlated on a decimeter bed scale (Robaszynski and Amédro, 1986; Gale, 1989, 1990, 1995; Amédro, 1993); for this reason, they are described concurrently here.

The Cenomanian succession is relatively expanded and complete in this area and includes the Lower Chalk, 65–75 m thick (see below for corresponding French formation names), Plenus Marl or F. des Crupes; (1.5–2.5 m) and the lowest 1.5 m of the overlying white nodular chalk (basal Melbourn Rock beds or F. du Grand Blanc-Nez).

In east Kent, the succession is well exposed in the cliffs and foreshore between Folkstone and Dover and was first described in detail by Price (1877). The description of Jukes-Browne and Hill (1903) was based on this work. Jefferies (1962, 1963) described the detailed succession in the Plenus Marls, and Kennedy (1969) published a lithological log of the Lower Chalk succession and established the boundaries of the ammonite zones. Robinson (1986) erected a formal lithostratigraphy for the Chalk of the region, and Jarvis et al. (1988a) described the micropaleontology of the Cenomanian-Turonian boundary. Gale (1989) and Jenkins et al. (1994) provided a lithological log of the Chalk Marl and modified the ammonite zonation. The lithological succession of the Folkstone-Dover Cenomanian is given in Fig. 6. The faunal records in Figure 4 represent a combination of data from Southerham in Sussex and from Folkstone itself. This is necessary because parts of the Folkstone succession do not yield well preserved diagnostic ammonites.

In the Boulonnais, the cliff and foreshore sections in the region of the Cap Blanc-Nez were first studied and subdivided biostratigraphically by Chollonneix (1872) and Barrois (1875, 1879). Lately, Jefferies described the stratigraphy of the Plenus Marls (1962, 1963), Amédro et al. (1976, 1978), Robaszynski et al. (1980), Robaszynski and Amédro (1993) and Amédro (1994) described the lithostratigraphy and biostratigraphy of the sections, establishing formal formation names and zonal boundaries. Amédro (1986) provided detailed records of the ammonites. Robaszynski and Amédro (1986), Gale (1990, 1995) and Amédro (1993) have discussed correlation between Kent and the Boulonnais. The succession in the Boulonnais is given in Figure 7, and important fossil records in Figure 3.

The Cenomanian succession in the Straits of Dover/Pas de Calais area comprises a basal unit, laterally variable in thickness, of glauconitic calcareous sand, overlain by decimeter-scale alternations of more and less marly chalk, which in the lower part contain concretions full of sponges. Above, the clay content decreases progressively through the overlying marls and chalks, and the rhythmicity becomes less conspicuous. Throughout the Cenomanian succession individual rhythmical couplets can be traced across the northern Anglo-Paris Basin and are inferred to be Milankovitch band precession cycles (Hart, 1987; Gale, 1990, 1995). Near the summit of the Cenomanian, the Plenus Marl/F. des Crupes displays a temporary increase in the proportion of clay, and are overlain by white, nodular chalks of the Melbourn Rock beds/F. du Grand Blanc-Nez at the summit of the Cenomanian.

Evidence of a general increase in sea-level throughout the Cenomanian is provided by an overall progressive decrease in the detrital clastic component (Destombes and Shephard-Thorn, 1971; Marie, 1960).

Sequences may be defined in relation to conspicuous sedimentary breaks, overlain by coarser beds of calcarenite or calcisiltite, which usually contain phosphatised intraclasts, some quartz sand and glauconite and represent deposits formed immediately overlying the transgressive surface. These surfaces are widely traceable to the west of Folkstone (Sussex, Isle of Wight, Dorset) and show a progressively onlapping relationship onto palaeotopographic structures through the Cenomanian succession (cf. Kennedy, 1970).

Description of the Succession

Sequence 1.—

In cliff and foreshore exposures to the east of Folkstone, the base of the Glauconitic Marl rests abruptly and disconformably on pale grey marls of Bed XIII of the Gault Clay, which have been dated as late (but not latest) Stoliczkaia dispar Zone age (Owen, 1976) and yielded rare specimens of the highest Albian planktic index Planomalina buxtorfi. In offshore cores a thin intervening unit of micaceous glauconitic sand referred to as “Zone 6a” by Carter and Hart (1977) is locally present (Figure 8); it is not known whether this is of Albian or Cenomanian age. In the Boulonnais, the earliest Cenomanian glauconitic sands rest on a thin bed (Bed 1a; 0.15 m) of glauconitic clay, overlying Mortoniceras infiltratum Zone clays of the St. P6 Formation, equivalent to Bed XI of the Gault Clay at Folkstone.

At Folkstone, the sandy Glauconitic Marl (5–7 m) is intensely bioturbated, weakly rhythmic in the upper part, and poorly fossiliferous; it yields rare phosphatised and indigenous ammonites of the lower N. carcitaneus Subzone (Gale, 1989).

In the Boulonnais, the equivalent F. de Strouanne comprises 2 m of glauconitic sandy chalk (Beds 1, 2 of Amédro, 1986)
**Fig. 6.**—The Cenomanian lithological and faunal succession of Kent and sequence stratigraphic interpretation. Abbreviations: qz.: quartz grains; gl.: glauconite grains; ph.: phosphatised intraclasts; l.c.: *Inoceramus crippsi*; l.v.: *Inoceramus virgatus*; O.m. 1-2-3: beds containing brachiopod *Orbirhynchia mantelliana*; E.o.: bivalve, *Entolium orbiculare*; P.v.: oyster, *Pycnodonte vesiculare*; A.: oyster *Amphidonte*; M1–M6: marker beds from Gale (1989); sequence stratigraphical nomenclature as in text.
which contain abundant phosphatised intraclasts in the lowest meter; these include numerous remanié ammonites of the earliest Cenomanian *N. carvitanense* Subzone (Amédéro, 1986).

The Gault/F. de St. Pô—Glaucolithic Marl/F. de Strouanne contact marks a hiatus of considerable duration, equivalent to the Octeville erosional surface of Haute Normandie (see below); in the Boulonnais, the latest Albian *S. dispar* Zone is represented only by 0.15 m of sediment. The burrowed erosional surface surmounting the Late Albian marly clays is a composite sequence boundary overlain by a transgressive surface at the base of the Glaucolithic Marl/F. de Strouanne. The erosively based lenses of Zone 6a may be relics of lowstand channel-fills or may represent earlier parts of the transgressive systems tract of the first Cenomanian sequence (Figure 8). The slowly deposited, winnowed Glaucolithic Marl/F. de Strouanne formed during the transgressive and early highstand intervals of Sequence 1.

The basal 3–4 m of the Chalk Marl and equivalent basal 2–3 m of the F. du Petit Blanc-Nez (Beds 3, 4 of Amédéro, 1986) are rhythmically bedded marly clays containing numerous sponigenous concretions, horizons full of inoceramid debris (*Inoceramus criptsi criptsi* Mantell) and small phosphatic intraclasts, surmounted by a thick limestone bed. The top surface of this bed is an erosion surface, and the overlying chalk (Bed 5 in the Boulonnais) contains glauconite, a little quartz sand and (at Folkstone) the equivalent M3 contains rare phosphatised pebbles. This erosional surface corresponds to the St. Jouin Hardground in Haute Normandie, and is the limit between Sequences 1 and 2; it represents a sequence boundary overlain directly by a transgressive surface.

**Sequence 2.—**

The overlying marly chalks (Beds 6–7 of Amédéro, 1986) are transgressive deposits of Sequence 2, and Beds 8 and lower 9 represent a highstand. The upper part of 9 is a thin shelf margin wedge.

**Sequence 3.—**

In both Kent and the Boulonnais, the base of Sequence 3 (Bed 10 i in the Boulonnais, M4 at Folkstone) is a coarse, arenitic sediment containing sparse quartz and glauconite and rare phosphatised intraclasts. It rests non-sequentially on pale marly chalk, into which it is piped deeply (1 m) by *Thalassinoides* systems. The contact is interpreted as a transgressive surface resting within the *M. saxbii* Subzone. This erosional surface corresponds to the Bruneval Hardground in Haute Normandie. In the southern Isle of Wight, this transgressive surface onlaps sequences 1 and 2 to rest directly on the Albian Upper Greensand.

The overlying 15 m of Chalk Marl (Dover/Folkstone) and F. du Petit Blanc-Nez (Boulonnais) represent the transgressive systems tract and highstand of Sequence 3; the downlap surface is marked by a decrease in the proportion of clay in the sediment.

**Sequence 4.—**

Above, a decrease in bed thickness and an increase in clay content mark the sequence boundary at the base of Sequence 4 shortly beneath the summit of the *M. dixoni* Zone. The shelf margin wedge is terminated by two prominent limestones, the higher of which is overlain by a 1 m bed of marly calcisiltite containing abundant calcitic bivalves (notably *Entolium orbiculare*), brachiopods, serpulids and small corals; the stratigraphy of this interval was described in detail by Paul et al (1994). The lower boundary of this bed is erosive and strongly burrowed, with dark *Chondrites* passing down into the subjacent light limestone (Gale 1995), and is interpreted as the transgressive surface of Sequence 4.

The overlying 10 m of marly chalk (F. du Cran, Beds 17–19) are rhythmically bedded and represent the transgressive systems tract of Sequence 4. Above, a sharp decrease in the clay content at the base of the Grey Chalk marks the downlap surface. The overlying highstand chalks are conspicuously rhythmic and thinly bedded and are terminated by a horizon which displays shallow channelling and is taken as the sequence boundary at the base of Sequence 5.

**Sequence 5.—**

The transgressive surface of Sequence 5 is marked by a thin marl, above which the sediment becomes silty and arenitic, and the chalk contains numerous laminated structures variously described as burrow fills or scours (Jukes-Browne Bed VII at Dover/Folkestone, Bed 23 in the Boulonnais) and rare phosphatised intraclasts in the Boulonnais. The coarseness is attributed to winnowing during the early stages of transgression. This level is overlain by 5 conspicuous rhythms, above which the carbonate content increases (downlap surface); the 13 m of nearly pure chalk up to the Plenus Marls (F. des Crupes) are inconspicuously bedded highstand deposits.

**Sequence 6.—**

The base of the Plenus Marls/F. des Crupes is represented everywhere in the Anglo-Paris Basin by a marked omission surface—the sub-plenus erosion surface of Jefferies (1962, 1963), which is the sequence boundary at the summit of Sequence 5. Immediately above this surface is a striking increase in clay content and coarseness of sediment. The upper limit of carbonate-rich Bed 3 is also an erosional surface (locally a hardground at Dover), and the base of Bed 4 of Jefferies is the transgressive surface of Sequence 6.

The upper parts of the Plenus Marls/F. des Crupes and the overlying Melbourn Rock beds/F. du Grand Blanc-Nez are silt-grade sediments, rich in calcispheres and inoceramid debris; the lowest meter of Melbourn Rock contains numerous chalk intraclasts and several discrete hardground surfaces. These sediments represent the transgressive systems tract of Sequence 6.
cally in the cores of anticlines or where it is brought up by faults.

The first (palaeontological) descriptions of the Cenomanian of Normandy were provided by Brongniart (1822) and Passy (1832); later, d’Orbigny (1852) cited the cliffs of La Hève (Le Havre) in support of his definition of the Cenomanian Stage. Hébert (1864, 1875) established several divisions of the Cre- naceous of Nornandy, and Lennier (1867) produced detailed sections of the cliffs. Detailed litho- and biostratigraphical descriptions of the region were provided by Juignet (1974) and Juignet and Kennedy (1976). A sequence stratigraphical interpretation of the succession was proposed by Juignet and Breton (1992).

The lithological succession of the Cenomanian at St. Jouin—Bruneval, near Antifer, is shown in Figure 9. In this several major discontinuities are present, represented by hardground surfaces which are regionally extensive and separate the succes- sive formations of St. Jouin, Rouen, Cap d’Antifer and Cap Fagnet. These deposits are made up dominantly of biomicrite and fall into 12 lithological units (VI–XV) separated by dis- continuity surfaces. Within each unit a more or less well- marked rhythmicity is picked out by variations in carbonate percentage and bioturbation—often replaced by beds of flint nodules, of which about 60 are present in the Craie de St. Jouin and Rouen. In the Formations d’Antifer and Cap Fagnet, the rhythmity is shown by beds of nodular chalk.

Where they are close to tectonic structures, the formations are reduced in thickness and sometimes dramatically condensed as a result of synsedimentary movements. This takes place in the Craie de St. Jouin near the Lillebonne fault, in the Bray anticline (Villequier) and in the Craie de Rouen bordering the Fécamp fault and the Rouen anticline (Pavilly; Juignet 1974).

The major sedimentary discontinuity surfaces form region- ally extensive datum levels, and the vertical ranges of fossil species (Figure 2) were plotted in relation to these (Juignet and Kennedy, 1976, Kennedy and Juignet, 1991). It is clear that the ranges are themselves controlled by discontinuous sedimentation.

**Description and Sequence Stratigraphy**

**Sequence 1.**

In the cliffs of St. Jouin-Antifer, the base of the Craie de St. Jouin is a conglomeratic level, containing intraclasts, bivalves, sponges and bryozoans set in an intensely bioturbated, very glauconitic matrix. This rests on dark grey bioturbated mica- ceous silty marls of the Gaize d’Octeville, which are of Late Albain (Stoliczkaia dispers Zone) age. The contact, called the Octeville ravinement surface, is irregular and modified by a network of centimeter-scale burrows with a glauconitic fill, which penetrate several decimeters down into the marls. The top of the Gaize d’Octeville has undergone significant erosion. Sandy nodular beds, about a decimeter in thickness, originally interbedded with the marls, were reworked and incorporated as large clasts in the basal glauconitic bed of the Craie de St. Jouin, together with bored green pebbles and phosphatised fossils of the earliest Cenomanian subzone of Neostlingoceras carcitansense.

The Octeville ravinement surface is a sequence boundary (SB), overlain directly by a transgressive surface (TS) on which rests 2 m of glauconitic sediment and 3 m of chalk in which the glauconite decreases upwards. This higher unit includes seven rhythms picked out by beds of flint nodules. These sediments represent the transgressive systems tract (TST) of the basal Cenomanian sequence. Above, a 1-m-thick flint-free nod- ular chalk is surmounted by a second, discontinuity, the St. Jouin Hardground. These sediments are associated with a highstand, a further sequence boundary lies at their summit.

**Sequence 2.**

The second sequence, which belongs to the Subzone of Man- telliceras saxbi, repeats the same theme. The basal contact shows evidence of erosion, with reworked sediment and fossils set in a glauconitic matrix infilling a network of burrows. The sequence boundary is here overlain by a transgressive surface, above which are 6 m of the transgressive system tract which contain cherty horizons and bands of grey flint, with a flinty nodular bed at the top (DLS). This transgressive facies is over- lain by a 5–6 m thick beige chalk containing little glauconite, numerous spicules of opaline silica and bioclasts. It shows rhythmity in flint development. The top of this chalk, which is identified as a highstand deposit, is sharply truncated by the Bruneval Hardground 1.

**Sequence 3.**

The first deposits of the third sequence (M. dixoni Zone) are coarse glauconitic calcarenites containing green intraclasts and lumachelles of an inoceramid (I. virgatus) and sponges. This shelf margin wedge is 3 m thick and includes a further break
Fig. 8.—Lithological succession in the Channel Tunnel boreholes T2 and PM (Point Métrique) 16754, after Amédro, (1994), to show the position and possible sequence stratigraphic interpretation of the “Zone 6a” level of Carter and Hart (1977). The succession is comparable to that seen in the cliff sections of Cap Blanc-Nez and Dover-Folkestone, but here “Zone 6a” sediments are intercalated between the Albian Gault Clay and the Cenomanian Chalk. These comprise several m. of sandy glauconitic sediment which contain reworked phosphatized pebbles at the base and the upper marly beds contain abundant large mica flakes. The development of “6a” is discontinuous and variable in thickness across the channel; it may be interpreted as a shelf magin wedge formed in erosional channels on the surface of the Gault Clay. O.m.: Orbirhynchia mantelliana, I.v.: Inoceramus virgatus, I.c.: Inoceramus crippsi, Sp: sponges. Horizontal squiggles represent strongly bioturbated levels.
**Sequence 4.**

The highstand deposits of the 3rd sequence are truncated by the major discontinuity of Rouen Hardground 1 (Juignet, 1980; Juignet and Louail, 1986), which marks a sequence boundary. This is also the transgressive surface at the base of the fourth sequence, on which lies the Rouen fossil bed, with its phosphatized, condensed fauna of *C. inerme* Zone and lower *T. costatus* Subzone age. The transgressive systems tract includes a basal glauconitic unit with quartz sand and phosphatized pebbles, overlain by a glauconitic chalk containing Rouen Hardground 2. This latter level can be interpreted as a downlap surface (DLS), preceding the highstand system tract of chalk containing calcareous spicules, bioclasts and flints that persists through the remainder of the *A. rhotomagense* Zone. The Pavilly Hardground or an equivalent bioturbated horizon with calcareous nodules at the base of the *A. jukesbrownei* Zone marks a further sequence boundary.

**Sequence 5.**

The shelf margin wedge above comprises a more or less nodular chalk containing bands of flint with opaline silica spicules and abundant pithonellids, alternating with more argillaceous, bioturbated beds (9 rhythms). Above, a marly chalk with laminated structures, glauconite, gravel-grade phosphate clasts and large ammonites (4 m) marks the transgressive systems tract of the fifth sequence and contains about 12 rhythms picked out by subtle differences between marly chalks and chalks with burrows and flint bands.
Towards the top of the Craie de Rouen (C. naviculare Zone), the sediment becomes richer in pithonellids, calcite spicules, and the rhythmicity is less distinct (4 flint beds). This highstand systems tract (3 m thick) is truncated by the Antifer Hardground 1, which is the local expression of the sub-plenus erosion surface of the basin and is the sequence boundary at the base of the sixth sequence.

Sequence 6.—

The lower part of the Zone of Metoicoceras geslinianum (1–2 m) shows 3 rhythms. At the base of each is a glauconitic sediment containing green gravel-grade material, overlain by nodular chalks containing shell debris and flints, surmounted by bioturbated nodular chalk with a green, phosphatized upper surface. This appears to be a lowstand wedge (SMW) with Antifer Hardground 2 as a transgressive surface at the top (Figure 10). The overlying transgressive systems tract (upper part of the M. geslinianum, N. juddii and W. devonense Zones) contains several successive discontinuities (Antifer Hardgrounds 3a, 3b, 3c) and continues up into the Lower Turonian Craie du Cap Fagnet. In Maine, the Cenomanian stratotype of Sarthe is developed dominantly in a clastic facies strongly affected by the proximity of the emergent Armorican Massif. The succession is divided into the following ascending formations: the Marnes de Ballon (Ballon Formation), the Sables et gres de la Trugalle et Lamnay (Trugalle and Lamnay Formations), the Sables et Gres du Mans (Mans Formation), the Marnes de Nogent Bernard and the Craie de Théligny (Théligny and Nogent Bernard Formations), the Sables du Perche (Perche Formation), the Marnes à P. biauriculata, the Sables à Catopygus obtusus and the Craie à Terebratella carantonensis. The average thickness is about 90 m. The incomplete exposures are limited to several quarries, but cores penetrate the entire succession.

Biostratigraphical study of this region was initiated by d’Orbigny (1847, 1850, 1852), and completed by the listing of local sections (Gueranger, 1867; Triger, 1869; Guillier, 1886). Revision of the Sarthe Cenomanian succession was undertaken most recently by Hancock (1960), Juignet et al. (1973), Juignet
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FIG. 11.—Summary of the lithological and faunal succession in the Cenomanian of Maine and its sequence-stratigraphic interpretation.


The lithological succession in the region of Le Mans is given in Figure 11. This provides the section of a cored borehole drilled at Le Mans and is completed by several exposed sections nearby in the region to the north and east of Sarthe.

As in the chalk successions of the Normandy Basin, several important discontinuity surfaces have been identified within the clastic succession of Maine. These are found in the form of erosion surfaces overlain by shelly glauconitic sands, ravine-mant surfaces underlying remanied sediments or more commonly, hardgrounds formed by early diagenetic hardening within quartz sands in shallow water. These surfaces are situated on the proximal platform in a series of lithological sequences that have been used to define the different formations. They correlate well with equivalent deposits of the contiguous distal chalk succession of Normandy.

Description and Sequence Stratigraphy

Sequence 1.—

The base of the Marnes de Ballon (N. carcitanense Subzone) is underlain by a burrowed erosion surface equivalent to the Octeville surface in Normandy. This is cut in the Glauconie à P. asper of Late Albian (S. dispar Zone) age and is the sequence boundary of the first Cenomanian sequence, overlain directly by the transgressive surface. The transgressive system tract (15 m thick) includes beds of glauconite and glauconitic silts with bivalve debris, which are overlain by more argillaceous levels that are rich in glauconite. At the top is a facies of the highstand with glauconite-free silty clays (3 m).

Sequence 2.—

Above, a second sequence is discernable, with the Ballon surface perforated by a network of burrows and overlain by shelly glauconitic sands (SB/TS); the transgressive deposits display a series of meter-scale rhythms including glauconitic silts and interbedded clays (6 m). The overlying black clays (3 m) mark a highstand.

Sequence 3.—

The third sequence commences with the Sables Hardground, which is a discontinuity at the base of the Sables de la Trugalle Formation (summit of the M. saxbii Subzone and base of the M. dixoni Zone). About 12 m of coarse, glauconitic sands with lignite form a shelf margin wedge overlain by a transgressive alternation of glauconitic sands and calcareous sandy limestones or sandy calcarenites 10 m thick. Minor breaks in the succession and early diagenetic cementation are common at this level, especially in the upper part where the Lamnay Hardground is developed (DLS). Above this hardground a 4 m thick prograding subtidal sand body representing the highstand sys-
tem tract is capped by the Théligny Hardground or its equivalent.

**Sequence 4.—**

The Théligny Hardground forms a sequence boundary equivalent to Hardground Rouen 1 in Normandy and is overlain by the transgressive surface of the fourth sequence. The transgressive systems tract of the Craie de Théligny, of which the basal meter is glauconitic and contains a phosphatized fauna of the *T. costatus* Subzone, passes laterally into obliquely stratified glauconitic sands of the lower part of the Sables du Mans (12 m thick). Above these sands the top of the Craie de Théligny becomes sandy and several hardgrounds are present (notably the Jalais Hardground at the summit) marking the highstand systems tract.

**Sequence 5.—**

The boundary between the *A. rhotomagense* and *A. jukes-brownei* Zones corresponds to a sequence boundary. Thus, above the Jalais Hardground, the lower part of the Sables du Perche (total thickness 25 m) is a fine sand with clay layers, comparable to a prograding shelf margin wedge with a terminal discontinuity, the Fayau Hardground 1 (TS). In the upper part of the Sables du Perche coarse, often glauconitic sands, containing large ammonites afford evidence of a transgressive systems tract. At the top, a hardground is often developed (DLS; Hardground Fayau 2 and equivalents). This unit underlies the Marnes à *P. biauriculata* (average thickness of 4 m) which consists of alternating calcarenites and glauconitic calcirudites with lenses of oysters and silty marls. This highstand facies belongs largely to the Zone of *C. naviculare* and is truncated by an erosion surface that marks the base of the sixth sequence.

**Sequence 6.—**

This boundary is the base of the Sables à *Catopygus obtusus* (*M. geslinianum* Zone), of which the lower part comprises bioturbated calcareous sandstones containing quartz gravel and many bivalves, including *Exogyra columba gigas*. These deposits are less than a meter thick and constitute a shelf margin wedge. The Bousse Hardground forms a transgressive surface; the transgressive systems tract commences with silty chalks (upper part of the Sables à *Catopygus obtusus*) and continues with the Craie à *T. carantonensis* (*N. juddii* Zone), which is a glauconitic horizon containing phosphatized fossils. The two are separated by the Mezières erosion surface (a flooding surface), which is strongly bioturbated with evidence of reworking. Above this surface the transgressive systems tract continues into the Craie à *Inoceramus labiatus* which is of early Turonian age. The biomicrites of this formation contain abundant calcispheres and onlap onto the residue of the Armorican Massif.

**HIGHLY CONDENSED BASIN MARGIN: SOUTH EAST DEVON**

In this region, Cenomanian deposits occur in a series of Cretaceous age outliers between Lyme Regis and Sidmouth, and at Wilmington, 10 km inland. Extensive exposures on the coastal cliffs and landslipped blocks allow the stratigraphy to be traced laterally in great detail. The succession was described by Meyer (1874, 1878), and subsequently by Jukes-Browne and Hill (1896, 1903). Smith (1957, 1961, 1965) described the lateral correlation of beds in the Cenomanian succession and the conditions of deposition. Kennedy (1970) recognised ammonite zones within the succession. Most recently, Jarvis and Woodroof (1984) and Jarvis and Tocher (1987) have introduced a new and very detailed lithostratigraphy for the region, and for the first time demonstrated the precise correlation between Wilmington and the coast. Simmons et al. (1991) discussed sea-level changes around the Albian-Cenomanian boundary in this area.

On the coast between Lyme Regis and Sidmouth and in the Wilmington outlier, the summit of the Late Albian Upper Greensand is marked by the strongly lithified Small Cove Hardground (Fig. 12; Jarvis and Woodroof, 1984). The higher part of the Upper Greensand, some distance beneath this hardground has yielded a small ammonite fauna of *S. dispar* Zone age at Shapwick (Hamblin and Wood, 1976). The hardground is a local representative of the widespread hiatus in north western Europe between latest Albian and Early Cenomanian sediments, and may represent several superimposed sequence boundaries.

The overlying Cenomanian Limestone (Meyer, 1874, 1878; Smith, 1957, 1961, 1965; Kennedy, 1970) called the Beer Head...
Limestone by Jarvis and Woodroof (1984) is a condensed de-posit, containing several major hiatuses, which attains a maximum thickness of about 13 m in the Hooken Cliffs west of Beer. It comprises nodular sandy bioclastic limestones, including four regionally extensive hardgrounds, which represent single and composite sequence boundaries. Sequences onlap onto contemporaneous structures, and in an extreme example the Cenomanian is onlapped by early Turonian chucks which rest directly on the Albian (Charton Cliff; Smith, 1965, pl. 5).

**Description and Sequence Stratigraphy**

**Sequence 1.—**

The lowest sequence (Bed A1 of Jukes-Browne and Hill 1903; Pound Pool Sandy Limestone Member of Jarvis and Woodroof, 1984; probably equivalent to the sediment immediately underlying the Wilmington Hardground of Jarvis and Tocher, 1987) contains a significant (<50%) upwardly decreasing proportion of quartz sand, numerous glauconitized intraclasts and abundant bivalve, brachiopod and bryozoan debris. This lowest unit is of *N. carcitanense* Subzone age (at Wilmington yielding abundant *Mantelliceras couloni*) and is interpreted as winnowed coarser material deposited in topographic lows on the inner shelf during maximum flooding and highstand intervals of the earliest Cenomanian sequence (Sequence 1). The summit is marked by the Weston Hardground, on which Sequence 2 of the Cenomanian succession in the Anglo-Paris Basin is missing.

**Sequence 3.—**

Cenomanian Sequence 3 is represented by Bed A2 (= the Hooken Nodular Limestone Member of Jarvis and Woodroof, 1984) which comprises nodular intraclastic shelly limestones, locally sandy, that yield ammonites and inorganic debris of the lower *M. dixoni* Zone. To the north of Beer Head, Bed A2 onlaps the underlying A1 sequence to rest directly on the Albian Upper Greensand. The extensively reworked intraclastic debris is interpreted as representing slow deposition during highstand.

The top is marked by the mineralized King’s Hole Hardground, on which there is a hiatus of considerable magnitude.

**Sequence 4.—**

Middle Cenomanian Sequence 4 is represented by Bed B (= Little Beach Bioclastic Member of Jarvis and Woodroof, 1984). In expanded successions (about 1–2 m thick) the sandy glauconitic lower part contains one or two hardgrounds and probably represents winnowing during the transgressive part of Sequence 4. Rare remanié ammonites from the lower part are indicative of the *Turrilites costatus* Subzone, although most of the sediment is probably of *T. acutus* Subzone age as inferred from the abundant *Holaster subglobosus*. The sequence is terminated by the massively lithified, extensively mineralized Humble Point Hardground, on which sequence 5 is lost.

**Sequence 6.—**

The highest Cenomanian sequence (Sequence 6) is represented by a sandy glauconitic chalk unit commonly known as Bed C (= Pinnacles Glauconitic Limestone Member of Jarvis and Woodroof, 1984). It attains a maximum development of over 2 m in the west part of Hooken Cliffs (Jarvis et al., 1988b). The lower part of Bed C immediately above the Humble Point Hardground contains phosphatized intraclasts, many of which are remanié fossils of the *Calycoceras guerangeri* Zone, and (less commonly) of the *Acanthoceras jukesbrownei* Zone. Bed C also contains lightly glauconitized ammonites of the *Metiococeras gesticinum* Zone and, in the region of Lyme Regis, indigenous ammonites of an un-named higher zone (Kennedy, 1970; Wright and Kennedy, 1981). In the most expanded Hooken Cliff sections, several nodular hardgrounds are present in the lower part of the unit, of which the Limonitic nodule hardground (Jarvis and Woodroof, 1984; Jarvis et al., 1988b) is probably equivalent to Antifer Hardground 2 in Normandy. Evidence from micropaleontology and carbon isotope stratigraphy led Jarvis et al. (1988b) to the contrary interpretation that the beds between the Humble Point Hardground and the Limonitic nodule Hardground are of *C. guerangeri* Zone age.

The latest Cenomanian Haven Cliff Hardground, which surmounts Bed C, is ubiquitously present in south-east Devon, and has yielded an abundant and diverse fauna of the *Neocardioceras juddii* Zone (Wright and Kennedy, 1981). The overlying nodular chalk contains ammonites of the earliest Turonian *Watinoceras coloradoense* Zone. In thinner successions, the Haven Cliff Hardground rests directly above the intensely mineralized (phosphate, glauconite) Humble Point Hardground.

The Humble Point Hardground is interpreted as representing two coalesced sequence boundaries: (1) at the base of sequence 5 and (2) at the base of sequence 6. Sequence 5 is thus not represented in Devon, other than as rare remanié phosphatized ammonites in the overlying sediment. The lower part of Bed C in expanded sections is a thin lowstand deposit, overlain by a transgressive surface at the top of the Limonitic nodule hardground. The Haven Cliff Hardground is a starved, current-scoured surface developed within the transgressive systems tract of sequence 6, overlain by a flooding surface of earliest Turonian age.

**CORRELATION OF THE SEQUENCES**

**Basin-Wide Persistence of Discontinuity Surfaces**

One distinctive feature that emerges from our comparison of different successions is the lateral persistence of major discontinuity surfaces over much of the basin. These pass from burrowed omission surfaces in the marls and marly chalks of the central basin, into glauconitized and phosphatized hardgrounds representing significant hiatuses on the basin margins. The contemporaneity of these surfaces can be demonstrated from the ammonite faunas, although the age of both under- and overlying sediment will vary depending on the extent of, respectively, erosion and onlap (Figure 13). The interpretation of these surfaces in terms of sequences leads us to suggest that the majority are sequence boundaries and transgressive surfaces, or an amalgamation of the two. These are described in ascending sequence, and numbered I–VIII on Figure 14.

I. Unamed surface (Maine), Octeille surface (Normandy)—Small Cove Hardground (Devon)—Gault-Glaucnonitic Marl contact (south-east England), St. Pô—Fm. de Strouanne contact (Boulonnais). This is coincident with the sequence boundary/transgressive surface at the base of Sequence 1.
FIG. 13.—Correlation of Cenomanian sequences and systems tracts in the Anglo-Paris Basin.
Fig. 14.—The development of sequences and systems tracts in the various successions studied, plotted against ammonite zones. Note the widespread development of the 8 discontinuity surfaces across the region.

II. St. Jouin Hardground (Normandy)—Ballon surface (Maine)—marker M4 (Kent)—Bed 4–5 boundary (Boulonnais)—Wilmington Hardground (Devon). This is coincident with the sequence boundary/transgressive surface at the base of sequence 2.

III. Sables Hardground (Maine)—Bruneval Hardground 1 (Normandy)—Weston Hardground (Devon)—Bed 9–10 boundary (Boulonnais)—marker M4 (Kent). This is the transgressive surface at the base of sequence 3.

IV. Rouen-Théligny Hardground (Maine)—Rouen 1 Hardground (Normandy)—King’s Hole Hardground (Devon)—sub-Totternhoe erosion surface (London Platform). This is the sequence boundary at the base of sequence 4.

V. Jalais Hardground (Maine)—Pavilly Hardground (Normandy)—Rag (Hertfordshire)—sub-Nettleton Stone erosion surface (eastern England). This is the sequence boundary at base of sequence 5.

VI. Bousse Hardground (Maine)—Antifer Hardground 1 (Normandy)—sub-plenus erosion surface (Boulonnais-S. England). This is the sequence boundary at the base of sequence 6.

VII. Bousse Hardground (Maine)—Antifer 2 Hardground (Normandy)—Limonitic nodular hardground (Devon)—Bed 3–4 boundary Plenus Marls (Boulonnais/S England). This is the transgressive surface of sequence 6.

VIII. Mezières surface (Maine)—Hardground Antifer 3a (Normandy)—Haven Cliff Hardground (Devon)—nodular chalks at base of Melbourn Rock (south east England). This is the flooding surface within sequence 6.

**Development of Sequences and Systems Tracts Across the Basin**

Figures 13 and 14 show the correlation of the systems tracts and sequences in the Anglo-Paris Basin based respectively on actual thicknesses and relative completeness against a hypothetical complete succession. Boundaries of the ammonite zones and subzones are taken to approximate to time lines.

In Figure 15, each of the 6 sequence boundaries is a horizontal line of correlation; this expresses visually the lateral variations in thickness of individual systems tract development for each sequence and facilitates discussion of the eustatic control on sedimentation.
Sequences 1,2.—

These are only thinly represented, and very little if any shelf margin wedge is developed, except locally in cuvettes in the Pas de Calais (Zone 6a). The preserved sediments fall essentially within the transgressive and highstand systems tracts. Although the earliest two Cenomanian transgressions extended sedimentation far onto the edges of the basin, accommodation space was restricted everywhere, probably because the sea remained relatively shallow. This resulted in a marked non-sequence at the Albian-Cenomanian boundary and condensation of the earliest Cenomanian sediments, so that the earliest Cenomanian fauna is invariably preserved as phosphatized intraclasts.

Sequence 3.—

A slight sea-level fall resulted in a break in sedimentation in the M. saxbii Subzone and the deposition of a thin shelf margin wedge in all 4 regions. The ensuing transgressive surface is widely recognized, and in the basin often contains reworked fossils from the underlying high M. mantelli Zone in a winnowed glauconitic, slightly sandy matrix of earliest M. dixoni age in which Inoceramus virgatus is abundant.

Sequence 4.—

Within the late M. dixoni Zone, a significant sea-level fall took place, which resulted in the development of an erosional or ravinement surface along the basin margins (sequence boundary), often cutting down into the M. dixoni Zone. The resulting sedimentary break is a characteristic feature of the Lower-Middle Cenomanian boundary in marginal settings and can be traced to Mangyslak in the trans-Caspian region (Marcinowski, 1980). In Normandy and Maine, this event is expressed in the development of a major hardground (Rouen—Théligny); in Devon, the King’s Hole Hardground; and on the London-Brabant Massif the erosional surface underlying the Totternhoe Stone. In the chalk basin, sedimentation was uninterrupted, but lowered sea levels resulted in the development of a more clay-rich shelf margin wedge of latest M. dixoni and C. inermis Zone age. The ensuing transgressive surface, of earliest A. rhotomagensis Zone age, onlaps onto the basin margins and commonly contains remanié phosphatized fossils in a glauconitic, sandy chalk matrix representing the transgressive systems tract. The superposition of chalks on shallow-water sands of Early Cenomanian age in Maine is evidence of the considerable magnitude of the Middle Cenomanian transgression. In the basin, the downlap surface is marked by an increase in the percentage of carbonate and lies some distance above a greatly increased abundance of larger keeled planktic foraminifers, the “mid-Cenomanian non-sequence” of Carter and Hart (1977).

Sequence 5.—

The sea-level fall at the base of this sequence is only locally marked on the basin margins, with the development of the Pavilly and Jalais Hardgrounds in Normandy and Maine, respectively, and an un-named hardground on the south side of the London Platform. In Maine, the progradational Sables du Perche forms a thick shelf margin wedge. The transgressive systems tract is widely recognizable in the basin, characterized by the development of coarse, winnowed chalks with scour structures (Bed 23,24 in the Boulonnais; Jukes-Browne Bed VII in east Kent), which extend into north east England as the Nettleton Stone. The return to finer, coccolith bearing chalks above represents the highstand systems tract.

Sequence 6.—

The sea-level fall at the base of the M. geslinianum Zone was rapid and of short duration; it is represented everywhere in the region by an omission surface or hardground (e.g. sub-plenus erosion surface in the basin, Antifer Hardground 1 in Normandy). The overlying shelf margin wedge is thin but laterally persistent and underlies the widely recognizable and strongly marked transgressive surface (base Bed 4 of the Plenus Marl; Antifer 2 Hardground). This vast and rapid transgression initiated in late M. geslinianum and N. juddii Zone time resulted in the complete cutoff of terrigenous clastic sediment on the basin margins and in the blanket-like spread of chalk sedimentation across the entire region, locally extending onto the surrounding Palaeozoic massifs.

REMARKS

Although the sequences described above can be traced throughout the region, there is a tectonic modification of the succession in some areas at certain times. For example, in the Middle Cenomanian succession of Normandy, uplift resulted in reduced sedimentary thickness and accentuated erosion on submarine hardground surfaces. The magnitude of the Albian-Cenomanian break is probably augmented by tectonic uplift (cf. Drummond, 1969).

Hart et al. (1992) present a brief and very different account of the Cenomanian sequence stratigraphy of southern England, based in part on evidence from the relative abundance of planktonic to benthonic foraminifers (P/B ratios) and on inferred breaks in sedimentation. They identify 3 sequences in the Cenomanian (their Fig. 3) and place the lowest sequence boundary at the base of the stage (Gault Clay:Glauconitic Marl boundary). Their first sequence extends up to the “mid-Cenomanian non-sequence” (cf. Carter and Hart, 1977), taken as a sequence boundary and marked by a major increase in the abundance of planktic foraminifers and an apparent local sedimentary
break in the English Channel (see Carter in discussion at end of Kennedy, 1969). Evidence of a sedimentary break at this level was not found in any of the sections examined in this study and instead the data indicates that this horizon falls within the transgressive systems tract of our 4th sequence. The horizon of the second sequence boundary of Hart et al. 1992 cannot be identified with certainty from their figure, but presumably falls in the lower part of the Upper Cenomanian succession. The base of the Plenus Marls and the base of the Melburn Rock are marked as maximum flooding surfaces (Hart et al., 1992, Fig. 3), but because no supporting evidence is provided, discussion is impossible.

Owen (1995) included a discussion of the Cenomanian of the Anglo-Paris Basin within a wider study of the sequence stratigraphy of this stage across western Europe. His results compare favourably with our own in placement of stratal surfaces and sequence interpretation of the regions described herein, with one major difference. Owen identified only 5 sequences in the Cenomanian of the Anglo-Paris Basin and interpreted our second sequence as a maximum flooding surface of sequence 1.

CONCLUSIONS

The synthesis presented allows the comparison of the effects of the same eustatic event on different regionally developed sedimentary facies. Maine is dominated by siliciclastic sediments, in contrast to the dominantly hemipelagic and pelagic carbonate facies of Normandy and Kent/Boulognais. Comparison of the regions demonstrates many similarities between the diverse successions of which the most important are enumerated below.

1. The ubiquitous presence of five sequences and the lower part of a sixth, in all the studied regions;
2. The existence of regionally persistent, biostratigraphically synchronous erosional surfaces which underpin the sequence correlation;
3. The presence of an important sea-level fall at the base of the fourth sequence, represented on the basin margins by a marked break at the Lower-Middle Cenomanian boundary, and by a clay-rich shelf margin wedge in the basin itself. This level is also marked by important faunal changes which are probably related in some way to this regression;
4. The subsequent sea-level rise early in the *A. rhotomagense* Zone, which resulted in the spread of chalk facies onto the basin margins; and
5. The great transgression of latest Cenomanian time saw pelagic chalk sedimentation extend onto and beyond the basin margins.

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REFERENCES


SEQUENCES AND SYSTEMS TRACTS OF MIXED CARBONATE-SILICICLASTIC PLATFORM-BASIN SETTINGS: THE CENOMANIAN-TURONIAN STAGES OF PROVENCE (SOUTHEASTERN FRANCE)

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ABSTRACT: The application of the sequence stratigraphy concepts to mixed carbonate-siliciclastic platforms and adjacent basins is of great interest in reconstructing the paleogeographic evolution of pericratonic domains. The Cenomanian-Turonian stages of Provence are well adapted to this application because the connection between the platform and the basin is preserved, thus leading to accurate facies and lithostratigraphic correlations. Moreover, both the paleogeographic setting of carbonate platforms and the source of siliciclastics can be determined. Biostratigraphic data based on ammonites, rudists, planktonic and benthic foraminifers, allows high resolution chronostratigraphic dating and correlation from the platform to the basin. Two detailed sections have been described as typical of the regional stratigraphical series: one for the basinal area, another for the carbonate platform. Sedimentology, biostratigraphy and sequence stratigraphy, lead to distinguish five third order sequences and their correlative systems tracts encompassing the Early Cenomanian-Middle Turonian time.

Distribution of facies and evolution of paleoenvironments are analysed with respect to relative sea-level changes. The amplitude of sea-level changes at the 3rd-order scale has been estimated by the method by Vail et al. (1977).

Sequence boundaries are of several types: tectonically enhanced, or controlled by eustasy. Two tectonically induced 2nd order regressive-transgressive cycles have been identified. The second one, from Middle Cenomanian to Middle Turonian time displays a succession of phases (forestepping, aggrading, backstepping) that compare well to those described by Jacquin et al. (1992) for the Middle Jurassic Paris Basin.

INTRODUCTION

Applications of sequence stratigraphy concepts to mixed carbonate-siliciclastic platforms and adjacent basins have been, until now, poorly documented in their general aspects (Dolan, 1989) as well as in their regional ones (Chafetz et al., 1986; Garcia Mondejar and Fernandez-Mendiola, 1993). Thus, a detailed stratigraphic study of such sedimentary systems is of great interest in reconstructing the paleogeographic evolution of pericratonic domains, because the latter are controlled by the combined effects of eustasy, tectonics and climate. Owing to the preservation of the connection between the platform and the basin, the Cenomanian-Turonian stages of Provence provide a significant example for this type of study.

PALEOGEOGRAPHIC SETTING

The study area (Fig. 1) is located south of the Durance swell, which formed as a consequence of a Late Aptian-Albian extensional phase (Philip et al., 1987; Chorowicz and Mekrania, 1992). Albian bauxitic layers were deposited on the Durance swell. Cenomanian and Turonian transgressions flooded the southern border of the swell, leading to the growth of several superimposed rudist carbonate platforms. These platforms graded transitionally towards a basinal area located to the south (Provence basin). Coarse-grained siliciclastic deposits are interbedded with the platform carbonate layers. They are derived from the erosion of the Corsica-Sardinia massif, which was located to the south east. Carbonates and siliciclastics therefore prograded in perpendicular directions (Philip, 1974).

STATIGRAPHY

An integrated biostratigraphic scale, based on ammonites (Fabre, 1940), foraminifers (Tronchetti, 1981), rudists (Philip, 1974) and ostracods (Babinot, 1980) allows both high-resolution dating and correlation from the basin to the platform facies. Special emphasis has been given to the Cenomanian-Turonian boundary in previous papers by Crumière-Airaud (1991) and Philip and Airaud-Crumière (1991).

Two sections have been chosen as typical of the regional stratigraphical series: Cassis, for the basinal area and Font-Blanche, for the carbonate platform.

The Cassis section (Fig. 2) crops out along the coast, to the south of the town, between the “Pointe des Lombards” and the “Anse de l’Arène”. Five sequences, bounded by conspicuous surfaces or abrupt facies changes have been distinguished. The first one (the “Banc des Lombards”) is represented by intensively bioturbated calcareous bioclastic sandstones. It rests unconformably upon Upper Aptian marls with a basal conglomerate and reworked ammonites of Clansayesian age (Fabre-Taxy and Thomel, 1964). Numerous ammonites and belonites of Early and Middle Cenomanian age have been recorded from the first sequence (revisions made by Fabre, 1940, Fabre-Taxy and Thomel, 1964 and Kennedy, 1994). It is topped by a ferruginous hardground.

The second sequence comprises three parts. The lowest is formed by ammonite bearing sandy-marls that contain interbedded carbonate debris-flows, made up of rudists (Caprina adversa, Ichthyosarcolithes triangularis) and orbitolinid limestones. The second part is formed by marls with ammonites and planktonic foraminifers. The third part of the sequence is represented by an alternation of sandstones and sandy-marl containing a poor foraminiferal fauna.

The third sequence begins with a graded sandstone with water escape structures and followed by slump sands. The second part of the sequence is formed by an alternation of bioturbated calcareous sandstones and sandy marls. The upper part displays an alternation of marls and calcareous marls with ammonites and planktonic foraminifers topped by a conspicuous bioturbated surface.

The fourth sequence is very thin (4 m thick); it has a lower calcisphaerite-rich slumped limestone with rare reworked rudists (Durania), then planktonic foraminifer-bearing (White innella ar chaeocretacea) marls followed by calcisphaerite-rich limestones with reworked benthic foraminifers and rudists (Durania). A bioturbated Planolites surface caps the fourth sequence.

The fifth sequence is formed, in its lower part, by strongly slumped calcisphaerite-rich limestones. The main direction of slump is from north-west to the south east. The second part contains a thick series (up to 200 m) of marls (marnes de l’Anse de l’Arène”) with planktonic foraminifers.

Because of the presence of Mantellliceras mantelli and Mant elliceras martimpreyi, Fabre-Taxy and Thomel (1964) have as-
signed the “Banc des Lombards” to the Early Cenomanian interval. Middle Cenomanian ammonites (e.g. *Acanthoceras rhoto-magense*) have been recorded from the hardground capping the first sequence and from the sandy marls forming the lower part of the second sequence (*Acanthoceras jukesbrownet*). The Lower Cenomanian-Middle Cenomanian boundary probably occurs at the hardground on top of the “Banc des Lombards”. The Upper Cenomanian succession is very well developed compared to the Lower and Middle Cenomanian. Characteristic Late Cenomanian ammonites (e.g., *Eucalyco-ceras pentagonum*) are present in the middle part of the second sequence. Thus, the Middle Cenomanian-Upper Cenomanian boundary probably fits at the top of the debris-flow unit. Ammonites are very scarce in sequences 3 and 4, but planktonic foraminifers are abundant and diverse. A typical Cenomanian planktonic foraminiferal association occurs in the middle part of the third sequence. The first appearance of *Pseudaspidoceras footeanum* (F. Amedro, pers. commun.) in the slumped limestones of the fifth sequence. The Cenomanian-Turonian boundary probably lies in the fourth sequence, which corresponds to the *Archaeocretacea* Zone.

The Font-Blanche section (Fig. 3) crops out along Route D.3a, located between La Bédoule and Le Camp. In contrast to the Cassis section, only four sequences (1 to 4) are represented here.

The first sequence rests unconformably on Upper Aptian marls and is formed, from base to top, of coarse bioclastic sandstones with orbitolinids and praevalveolinids followed by an alternation of praevalveolinid mud-wackestones and oyster-bearing clays. This first sequence is capped by a hardground with ferroan encrustation and erosional features.

The second sequence begins with an alternation of lenticular sandy marls and praevalveolinid-miliolid mud-wackestones; ostracods and charophytes are frequent. These facies are followed by bioclastic limestones with echinoderm fragments, bryozoans, sponge spicules and calcispheres. The upper part of the sequence is formed by rudistid (caprinids, radiolitids) floatstones. A leached surface ends this second sequence.

The third sequence is formed by bioclastic-rich bryozoan pack-grainstones with scarce planktonic foraminifers, followed by rudistid pack-grainstones; in these, radiolitids are abundant in the lower part, while hippuritids occur in the uppermost 2 m (Philip, 1978). The third sequence is capped by a conspicuous hardground with iron encrustation and erosive features.

The fourth sequence begins with a thin (0.5 m) bioclastic and glauconitic bed with bryozoans, calcispheres, planktonic foraminifers and ammonites. Above, there is an alternation of bioclastic sandstones and rudistid wackestones.

The first occurrence of praevalveolinids and orbitolinids in the lowermost part of the first sequence suggests its age as Middle Cenomanian; the upper part (without orbitolinids) is referred with some uncertainty to the Late Cenomanian interval. The benthic association: rudists (e.g. *Caprinula-Sauvagesia-Dura-nia*), foraminifers (e.g., praevalveolinids, *Chrysalidina gradata*) and bivalves (*Chondrodonta joannae*) provide a Late Cenomanian age to the second sequence (Philip, 1978). This age is confirmed by *Eucalyco-ceras pentagonum*, found near the La Bégude section (Babinot, 1980, Tronchetti, 1981).

Bioturstratigraphic data are less precise in the third sequence. Its base belongs probably to the Late Cenomanian subdivision because of the presence of *Chrysalidina gradata*. The onset of the first hippuritids (*Vaccinites fontalbensis*) in the upper part of the sequence argues for an earliest Turonian age (Philip, 1978). Some chronostratigraphic uncertainties remain for the age of the middle part of the third sequence. Philip (1978) interpreted this part as a transitional undifferentiated interval between the Cenomanian and the Turonian. The Cenomanian-Turonian boundary is probably in this interval, but its exact position cannot be specified.

Ammonites indicate that the lowermost part of the fifth sequence is of Early Turonian age, but the Lower Turonian-Middle Turonian boundary cannot be placed with precision.
Fig. 2.—The Cenomanian-Lower Turonian basinal type section at Cassis.
Fig. 3.—The Cenomanian-Middle Turonian platform type section at Font-Blanche.
Correlations with the Cassis section, based on sequence analysis, biostratigraphic data and mapping, indicate the absence of Lower Cenomanian strata in the Font Blanche area. The sequences 1 to 4 of Font Blanche can easily be correlated with the sequences 2 to 5 of Cassis.

In the Provence series, the Lower Cenomanian (in part), the Middle Cenomanian and the Lower Turonian (in part) successions are so characterized by the standard ammonites zones. Upper Cenomanian strata are characterized by *Eucalycoceras pentagonum*, which is considered partly coeval with the *Calycoceras naviculare* zone (Robaszynski et al., 1993). Fabre (1940), Babinot (1980) and Tronchetti (1981) cited the presence of *Neolobites vibrayeanus* (d’Orbigny) in levels coeval with the ammonitic bioclastic biozone of the second sequence of Font Blanche. In the Paris Basin, this ammonite occurs just below the *Metoicoceras geslinianum*/*Sciponoceras gracile* standard ammonites have not been found until now in Provence.

**SEQUENCES AND SYSTEMS TRACTS**

Detailed stratigraphic sections, lithological correlation and mapping between Cassis and Camps (Fig. 4B), have allowed five 3rd-order sequences and their constitutive systems tracts *sensu* Van Wagoner et al. (1988) to be distinguished (Fig. 4A).

The first sequence is dated from the Early Cenomanian *Mantelloceras mantelli* Zone to the Middle Cenomanian *Acanthoceras rhoto magense* Zone. The transgressive systems tract (TST) includes skeletal bioclastic limestones with echinoderm clasts, ammonites and belemnites in the distal (basinal) part (Cassis) but sands and sandstones in the proximal part (La Bédoule). Hardgrounds or erosive surfaces are found. They can be documented by the condensation of the ammonite zones (at the Cassis section, Fabre, 1940, Fabre-Taxy and Thomel, 1964). The highstand systems tract 1 (HST) comprises caprinid and orbitolinid limestones developed in the La Bédoule area. Sequence 1 is bounded at its base by a regional angular unconformity (SbH), which is typically a tectonically enhanced sequence boundary. It probably results from a relative sea-level fall, coeval with the uplift and erosion of the Duranian swell during Albian time.

The second sequence starts with the sequence boundary Sb2. It is marked by a low angle unconformity in the proximal area. It is identified by a well-marked hardground (the so-called “Banc des Lombards”) in the basin (Cassis). The overlying lowstand deposits (LST, “relative lowstand”, according to Dolan, 1989), is assigned to the Middle Cenomanian *Acanthoceras jukesbrownei* Zone. At La Bédoule, lithologies include coral-stromatolite bioherms (“mud-mounds” in Maurin et al., 1981) associated with orbitolinid-bearing limestones. North-west-south east oriented channels with calcareous debris-flows give the

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**Fig. 4.—**(A). Cenomanian and Turonian sequence stratigraphic interpretation of Provence. (B). Location of the sections and paleogeographic setting.
overall trend of the slope. Several prograding parasequences that pinch out towards the basin have been identified within this lowstand carbonate complex. The transgressive deposits (TST) of the second sequence are represented (i) in the proximal part by backstepping rudist banks, (ii) in the border of the platform by aggrading caprinid-coral buildups and (iii) in the basin by ammonite-bearing marls dated as the Late Cenomanian Calycoceras naviculare Zone. The highstand deposits are characterized by prograding praealveolinid wackestones deposited in inner platform environments and perideltaic sandy marls in the basin.

The base of the overlying sequence (SB) is marked by a surface of emergence in the inner platform. The third sequence itself belongs mainly to the Late Cenomanian Calycoceras naviculare and Metoicoceras geslinianum Zones. Lowstand deposits (LST) are formed by slumped sands and sandstones extending from the margin of the platform into the basin. The corresponding transgressive deposits (TST) extend far onto the Durance swell. They comprise two backstepping parasequences. The lower one contains: (i) lagoonal paralic carbonateous shales and oyster-bearing clays, interbedded with paleosilts and sandy tidal channels (Babinot et al., 1988), (ii) praealveolinid limestones characteristic of inner carbonate platform, (iii) rudist banks and (iv) sands and sandy marls in the basin. The second parasequence is more transgressive. It comprises open platform limestones with ammonites (Eucaleyoceras pentagonum, Neolobites vibrayeanaus) that grade basinward to pelagic marls with planktonic foraminifers (Rotalipora cushmani). The highstand deposits (HST) are represented by limestone beds with rudists (radiolitids) and bentic foraminifers (Chrysalidina gradata). These grade basinward to pelagic marls. A conspicuous erosional surface interpreted as an unconformity (SB) ends this third sequence on the platform.

The fourth sequence is assigned to the latest Cenomanian-earliest Turonian. The lowstand systems tract (LST) is formed by slumped pelagic limestones in the basin. The transgressive systems tract (TST) is represented on the shelf by limestones with sponge spicules and calcispheres. These grade basinward into marls rich in planktonic foraminifers (heterohelicids). The highstand deposits (HST) are formed on the shelf by rudistid buildups (radiolitids, then hippuritids, Philip, 1978) coeval with sponge-rich and planktonic foraminiferal (Whiteinella archaeocretacea) limestones in the outer platform and the basin. A well-marked unconformity (SB), corresponding to a hiatus, tops this sequence on the platform. On the distal platform, the hiatus is probably in the Coloradoense Zone (Philip and Airaud-Cruzière, 1991), while in the proximal areas, it is longer and encompasses the entire Lower Turonian strata down the Mammites nodosoides Zone. This unconformity is characterized by erosion and leaching in the proximal area. In the distal platform, early diagenetic features such as micritic, fibrous or syntaxial cements are typical of submarine hardgrounds (Garrison et al., 1987).

The fifth sequence (Early Turonian-Middle Turonian age) begins with lowstand deposits (LST), first represented by pelagic slumped mudstones with Helvetoglobotruncanca praevelvetica and H. helvetica. They are interbedded with debris-flows that include elements reworked from the underlying sequence (Philip, 1974). Other interbeds are organic-rich marls that onlap the border of the platform. The transgressive systems tract (TST) of this fifth sequence displays a well-marked change in thickness and lithology from the basin to the platform. These are: (i) up to 100-m thick hemipelagic marls in the basin, (ii) less than 1 m of condensed glauconitic limestones with ammonites (Mammites nodosoides) on the outer platform, and (iii) echinoderm-bearing sandy grainstones deposited in high-energy environments on the inner platform (Camps area). The highstand deposits (HST) are dated as Middle Turonian age. They are characterized by a new carbonate platform with rudists (radiolitids and hippuritids), passing basinward to limestones with sponge spicules and echinoderm clasts, then pelagic marls. This carbonate platform is composed of three parasequences that clearly prograde basinward. A general progradation of carbonates is still recorded in the Upper Turonian strata (Philip, 1974).

**DISTRIBUTION OF FACIES AND EVOLUTION OF PALEOENVIRONMENTS**

The role of relative sea-level changes on the genesis and the distribution of carbonate facies (Sarg, 1988) and adjacent siliciclastics can be documented in the study area.

Carbonate geometries and facies are driven by the gradual sea-level rise on the Durance swell. During the relative lowstand of sequence 2, carbonate facies develop on the border of the platform. They are formed by slope deposits such as mudmounds and debris-flows. During the transgressive interval of the second sequence, the biologic diversity of the benthic association increased. The relative rise of the sea level allowed aggrading rudist buildups, but the areal extent of carbonate construction was limited by the influx of clay and sands. The maximum extent of the carbonate facies on the platform occurred during the peak transgression of sequence 4. The large inundated surface on the shelf allowed major development of rudists. These extended carbonate ramp systems were exposed and eroded during subsequent lowering of sea level that induced the SB unconformity. The next episode of transgression (in Early Turonian time) produced open-marine condensed sequences. These overlie and drown the former carbonate ramp deposits (Floquet et al., 1987).

During the deposition of the third sequence, lagoonal paralic facies are well developed in transgressive settings, especially when barriers built by rudists or by sand bars have formed lagoons. No coals were deposited in the lagoons, but carbonateous shales, formed in peat swamp environments are known. The lagoonal parasequence of the Late Cenomanian age third sequence contains a meter-scale asymmetrical alternations of shales and carbonates. The thickness of the shale units clearly decreases upward as a consequence of the overall backstepping on the platform.

Siliciclastics are mixed with carbonate facies only in the platform border mainly at the base of the third sequence. The sands are derived from the Corsica-Sardinia block and prograded into the Provence basin during lowstand phases. Accordingly, the siliciclastics overlie the border of the platform and constitute a shelf wedge.

The basinal deposits are characterized by regularly cyclic alternation of pelagic limestones and marls. The total organic carbon content is low (less than 0.7%). Nevertheless, hypoxic facies characterized by a decrease in bioturbation occurred during the Cushmani and Helvetica Zones (i.e. during the peak
transgression on the platform). It records one of the well-known global anoxic events.

THIRD-ORDER CYCLES

The amplitude of sea-level changes corresponding to the observed 3rd-order cycles can be estimated by the method by Vail et al., (1977) (Fig. 5). A water depth of approximately 100 m has been estimated on the distal part of the platform after the first sequence, compatible with the facies at Cassis. Our first sequence is relatively thin with respect to the overlying ones. Several 3rd-order sequences have been described for this period in the London Paris Basin (Robaszynski et al., this volume), but they are not clearly expressed in Provence as a consequence of the condensation of strata related to the low rate of subsidence in the proximal setting. Our sequences 2 to 5 are better developed due to the increased subsidence rate of the Provence platform with the maximum rate being reached by Early Turonian time.

The sequence boundaries (Sb) are of several types. Sb₁ and Sb₂ are typically tectonically enhanced unconformities (sensu Vail et al., 1991). The Cenomanian beds sit on an erosional surface (Sb₁) that intersects previous layers dated Aptian in the south west down to the Jurassic age in the north east area of our transect (Fig. 4A). Sedimentary breccias and debris flows belonging to the second sequence rest upon the first one by a low-angle tectonically-enhanced unconformity (Sb₂) resulting from the tilting of the outer platform.

Sb₁ is marked by exposure of the inner platform and is followed by siliciclastic influx. It is interpreted as a consequence of a sea-level fall but with no evidence of tectonic deformation. Sb₂ displays exposure features in the inner platform only; it can be interpreted as a result of a decrease in the rate of subsidence in the eustatic rise that allowed depositional filling on the platform. Sb₃ resulted from a sea-level fall subdued by reactivation of subsidence on the distal platform.

The sequence stratigraphic analysis at the 3rd-order scale (Vail et al., 1991, Jacquin et al., 1992) documents the step-by-step rise of relative sea level on the Durance swell (Fig. 4A and 5). The rate of the rise was not constant. It was moderate during Early and Middle Cenomanian time and increased during Late Cenomanian and Early Turonian time. This could result both from an increase in subsidence or a variation in the rate of the sea-level rise (Rowley and Markwick, 1992). The maximum flooding began during the Late Cenomanian Metoicoceras geslinianum Zone (sequence 3). The latest Cenomanian and earliest Turonian transgressions (sequences 4 and 5) did not exceed the limits of the Late Cenomanian. The amplitude of the Early Turonian Mammites nodosoides Zone (sequence 5) coastal onlap is equivalent to that of the Late Cenomanian Metoicoceras geslinianum Zone (sequence 3). In fact, the Early Turonian ammonite-bearing levels are only represented on the outer Provence platform, and they have not been identified on the inner part. They probably grade laterally to platform carbonates that have not been differentiated yet from the Middle Turonian ones by biostratigraphical data. Work in progress focuses on Turonian carbonates in order to define the extent of the Early Turonian transgression. It could show the age of the 2nd-order peak transgression, leading to the drowning of the carbonate platform.

SECOND-ORDER CYCLES

Two 2nd order cycles can be identified (Fig. 5). The lower one is represented pro parte by our first sequence, which resulted from the condensation of several 3rd-order sequences. This cycle is interpreted as a consequence of the rise of the Durance swell dated from Late Albian to Middle Cenomanian (Rhotomagense Zone) age. Because the study area is located in too proximal a setting, the first 3rd-order sequence that belongs to the regressive part of this cycle cannot be observed. This sequence can be documented in the Toulon area (Southern Provence), which subsided more and was deeper than the northern part from Late Albian through Late Cenomanian time (Philip et al., 1987) and consequently allowed the first sequence of this cycle to be formed.

The second regressive-transgressive cycle comprises the 3rd-order sequences 2 through 5. It displays a succession of phases that compare well to those described by Jacquin et al. (1992) for the Middle Jurassic Paris Basin. Thus, from base to top, the following evolution can be distinguished: forestepping (mudmounds and debris flows of the 3rd-order sequence 2) aggrading (sequence 2 pro parte) backstepping (sequences 3 and 4) and restriction (lower part of the sequence 5). Moreover, the upper part of our 3rd-order sequence 5 can also be compared to the infilling phase of the model by Jacquin et al. (1992), following the 2nd-order peak transgression. The infilling part, which is part of the regressive phase at the beginning of the 2nd-order cycle, is missing here. This can be explained by the existence of erosional features on the shallowest part of the Durance swell. The corresponding sediments have been redeposited as blocks of early lithified carbonate grainstones within breccias of the forestepping deposits.

The transgressive-regressive 2nd-order cycle is bound by two major surfaces. The lower one is the tectonically enhanced surface boundary of the 3rd-order sequence 2 (Sb₃). The upper one is the drowning unconformity, which is particularly well marked on the outer platform. The great thickness of the restricted part of the 2nd-order cycle is explained by outer platform flexure that occurred mainly by earliest Turonian time.
CONCLUSIONS

The Cenomanian-Turonian stages of Provence have provided interesting applications of sequence stratigraphic concepts to a mixed carbonate-siliciclastic platform and adjacent basins. The inherited tectonic setting is a fundamental factor that controlled the relative dominance of carbonate-siliciclastic sedimentation. The Late Aptian-Albian tectonic extensional phase created a paleo-high (the Durance swell) on which carbonate platform sedimentation developed during Cenomanian and Turonian time and siliciclastics were shed off of the Corsica-Sardinia massif.

An integrated biostratigraphic scale has given accurate chronostratigraphic data and led to correlations between platform carbonates and basinal siliciclastic facies. In Provence, ammonites, planktonic and benthic foraminifers and rudists have allowed zonal subdivisions of the Cenomanian and the Turonian strata. Moreover, correlations have been extended by mapping the lithostratigraphic units between the innermost platform and the basin.

Five 3rd-order depositional sequences, their constituent systems tracts and bounding surfaces have been distinguished in the Lower Cenomanian-Middle Turonian interval. Siliciclastics dominated basinal sequences, while mixed carbonate-siliciclastics deposits occurred in the platform-basin transition. From the Early Cenomanian to the Early Turonian, a general coastal encroachment and aggradation of the sequences were inferred from successive positions of coastal onlap onto the Durance swell. Following this, backstepping of the carbonate platform occurred, while a carbonate progradational trend developed during the Middle Turonian time.

Relative changes of sea level seem to be a major factor driving the genesis and distribution of facies and the evolution of paleoenvironments. Carbonate geometries and facies were driven by the 3rd-order cyclicity of sea-level rise on the Durance swell. Lagoonal paralic facies are well developed in the platform basin during lowstand phases. Hypoxic facies in the basin occurred during the 2nd-order peak transgression on the platform.

At the scale of the 3rd-order cycles, the relative sea-level changes on the Durance swell, occurred gradually with varying rates. The peak transgression occurred during the Lower Turonian time leading to the drowning of the carbonate platform. At the scale of the 2nd-order cycles, relative sea-level changes have been tectonically induced. In the Middle Cenomanian, platform tilting induced the formation of a carbonate forestepping sequence, while at the Cenomanian-Turonian boundary a platform flexure was responsible for the genesis of a thick restricted sequence in the adjacent basin.

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THE NORTH ATLANTIC CYCLE: AN OVERVIEW OF 2ND-ORDER TRANSGRESSIVE/REGRESSIVE FACIES CYCLES IN THE LOWER CRETACEOUS OF WESTERN EUROPE

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ABSTRACT: Up to six 2nd-order transgressive/regressive facies cycles have been identified within the overall Lower Cretaceous transgressive phase. They are defined on the basis of a widespread set of data ranging from the northern North Sea to southern Italy. They are long-term duration (3 to 14 my) cycles, made of higher order depositional sequences which have been described in both carbonate and siliciclastic settings and mainly from the London-Paris basin to southern Italy. The large-scale stratal pattern of these cycles document the relationship of the type and amount of sediment supply to changes in the rate of accommodation. The Pre-Aptian cycles which are mostly aggradational, relate to high-rates of sediment supply together with changing rates of accommodation. They are limited by major erosional unconformities to which forstepping sequences that include major lowstand deposits, are merging. The Aptian and post-Aptian cycles are mostly retrogradational; these lead to the widespread development of organic-rich condensed horizons at peak transgressions in basin settings. The unconformities bounding these cycles are also highly erosional and relate to major intraplate reorganization of the strain field in connection with the North Atlantic rifting.

INTRODUCTION

The North Atlantic cycle refers to the major, 1st-order, Cretaceous transgressive/regressive facies cycle during which the North Atlantic rift developed (Jacquin et al., this volume). It is made up of thirteen 2nd-order, shorter duration, transgressive/regressive cycles (Fig. 1). This paper concerns the Lower Cretaceous transgressive/regressive facies cycle during which the North Atlantic rift developed (Jacquin et al., this volume). It is made up of thirteen 2nd-order, shorter duration, transgressive/regressive cycles 11 to 15 of our chart). The main reason for dividing this long-term interval into different parts is linked with the large-scale development of the chalk facies over northwestern Europe during Upper Cretaceous times.

2ND ORDER CYCLE 11: UPPERMOST RYAZANIAN (NORDIC) OR UPPERMOST BERRIASIAN (TETHYAN) TO LOWER VALANGINIAN.

Main Characteristics

The boundaries are dated as Upper Ryazanian (pre-Surites (Bojarkia) stenomphalus Zone) to Lower/Upper Valanginian boundary in the Nordic areas. They are dated as upper Berriasian (Upper Berriasella picteti Zone) to lowermost upper Valanginian (Saynoceras verrucosum ammonite Subzone) in the Tethyan areas. The duration is 3 my (depositional sequences Be7 to Va3). The peak transgression is dated as Uppermost Ryazanian (Peregrinoceras albidum ammonite Zone) in the Nordic areas and lowermost Valanginian (Thurmaniceras otopeta ammonite Zone) in the Tethyan areas. The initial definition of 3rd-order depositional sequences is due to Arnaud-Vannee and Arnaud (1991; 1992); Detraz and Stanhauser (1988); Detraz (1989); Detraz and Mojon (1989); Hesselbo and Allen (1991); Jan Du Chêne et al. (1993), Steffen and Gorin (1993); Strauss et al. (1993); Rusciadelli and Jacquin (in prep.).

Lowermost Unconformity

The lowermost unconformity is not a major angular unconformity but a disconformity surface at the turning point between regressive facies below and transgressive deposits above.

Northern Subalpine Chains and Jura.—

The disconformity surface DIII of Detraz (1989) is marked by the emergence of the reeval rim Formation d’Allèves au Mollard de Vions (Detraz and Stanhauser, 1988) and by the development of terrigenous siliciclastics into the laterally subsiding lagoons (Calcaires Gréseux et Marnes à Characées; Detraz, 1989). This discontinuity has been well dated in the lagoonal setting of the platform by ostracods (Cypridea valdensis obliqua Zone) and by charophytes (M5a/M5b zonal boundary) by Mojon (in Detraz and Mojon, 1989). Down-dip the discontinuity surface is tied to the basal D3 Calpionellid Zone (Detraz, 1989; Detraz and Mojon, 1989) and to the M16/M15 magneto-zone boundary (Jan Du Chêne et al., 1993).

London–Paris Basin.—

In the London-Paris Basin (Fig. 2), a similar erosional surface without any angular relationship is observed at the top of the Purbeckian succession. In the Paris Basin, a widespread hiatus covering most of the Berriasian can be documented between the transgressive Marnes Noires dated by dinocysts of the Surites (Bojarkia) stenomphalus to Platylenticeras sp. ammonite Zones and the regressive Purbeckian dolomite below. Within the Wealden facies belt in Dorset and Yorkshire (U.K.), the basal transgressive deposits (Lower Speeton Clay) yield charophyte and dinocyst assemblages similar to those of the Jura mountains and ammonites of the Surites (Bojarkia) stenomphalus ammonite Zone (Rawson and Riley, 1982; Hesselbo and Allen, 1991; Allen and Wimbledon, 1991). These transgressive deposits have been also tied to the magneto-zone M16/ M15 boundary (Ogg et al., 1991), which corresponds to the onset of the starved transgressive deposits in the Subalpine (Vocontian) Tethyan marginal area.

Germany.—

The onset of the Surites (Bojarkia) stenomphalus transgression within the Bückeburg Formation is reflected by a general increasing dinocyst diversity within the Wealden facies from at Mesozoic and Cenozoic Sequence Stratigraphy of European Basins, SEPM Special Publication No. 60
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Fig. 1.—Compared transgressive/regressive facies cycles and depositional sequences for the Early Cretaceous western Europe.
Fig. 1.—Continued.
Transgressive Phase T11 and Peak Transgression

The transgressive phase T11 comprises the two depositional sequences Be7 and Be8 of the chart. They are dated *Surites (Bojarkia) stenomphalus* and *Peregrinoceras albidum* Zones in the Boreal realm, *Berriasella callisto* and *Thurmaniceras otopeta (pars)* in the Tethyan areas. The main characteristics of the transgressive phase T11 are: (1) the persisting anoxia within the northern North Sea basins (Viking Graben), (2) the development of anoxic lakes within intracratonic European enclosed basins such as the Paris Basin and the German Wealden Basin, and (3) the progressive stratification of the water column in the marginal Tethyan subsiding areas such as the Vocontian basin (SE France). Such a trend is found also or even enhanced during the subsequent Lower Cretaceous transgressive episode.

**North Sea.**—

The lowermost surface of cycle 11, where preserved, is not a major erosional unconformity, but a disconformity surface associated with the last rapid backstepping phase of the Draupne interval. On footwalls of major northern North Sea tilted blocks, conglomeratic lag deposits are observed along that surface.

Transgressive Phase T11 and Peak Transgression

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**Northern Subalpine Ranges (Savoy).**—

Bioclastic shelfal limestones named as Chambotte Inférieure Formation fine upwards to an open marine marly level (Marnes d’Arzier) of the *Thurmaniceras otopeta* ammonite Zone (Detraz, 1989). This bioclastic facies fine also seaward by sediment starvation into ammonite-bearing condensed levels and hardgrounds on the outershelf (Detraz and Stanhauser, 1988). More distally on the outerslope (Chartreuse mountain; Fontanil section), the Be8 sequence that underlies the peak transgression shows a well developed lowstand prograding complex named as Valetière member of the Fontail Formation (Arnaud-Vanneau and Arnaud, 1991; Ferry and Rubino, 1988b; 1989).
Subalpine Vocontian Basin (SE France).—

The transgression is marked (1) by an overall increasing dinocyst diversity (Steffen and Gorin, 1991), (2) by an increasing upwards Mn content (Jan Du Chêne et al., 1993) and (3) by the better preservation of the organic matter towards the maximum flooding surfaces of sequences Be7 and Be8 (Steffen, 1993).

Western European Shelves.—

They are dominated by Wealden-type sediments (Germany) or by Purbeck-type deposits (Dorset basin), or by shallow-marine sediments (Yorkshire and southern North Sea). The transgressive phase and the peak transgression are characterized by an overall upward increase of the organic matter content. In the Yorkshire (Steffen, 1993), both offshore and onshore within the Speeton clay, TOC values reach 6% at the peak transgression. In the Paris basin the lagoonal Marnes Noires, which develop from the basal unconformity to the peak transgression, contain type III OM with TOC up to 4 to 5% (Rusciadelli and Jacquin, in prep.). In Germany the transgressive Wealden sandstone interfinger with organic-rich lacustrine shales of the Bückerberg Formation (Strauss et al., 1993) as well (Ruffel, 1991).

North Sea.—

The last occurrence of organic matter-rich shales within the Draupne Formation corresponds to that transgressive phase T11. The shales better develop within the hanging walls of the tilted blocks, and grade updip into backstepping shoreface sandstones, still belonging to the Munin Fm. Crest lines of the structures could have been drowned at the peak transgression.

Regressive Phase R11

The regressive phase R11 comprises the highstand systems tract of 3rd-order sequences Be8, Va1 and Va2 and the lowstand systems tract of sequence Va3. It is the first strong terrigenous influx of the Lower Cretaceous on European shelves (continental Wealden deposits). In the Tethyan-regarding shelves, it is a pronounced prograding interval such as the Chambotte Supérieure and the Rivoire Formations of the French northern Subalpine chains in Savoy (Detraz, 1989; Arnaud-Vanneau and Arnaud, 1991). In the Vocontian basin of southeastern France, the Lower Valanginian lowstand deposits of sequences Va1 and Va2 comprise mainly bioclastic and quartzose limestone named as Calcaires Roux (Ferry, 1991; Ferry and Rubino, 1989). They yield consistent and frequent representatives of the palynological Muderongia group, which reflects the reworking of shallow marine to emersive deposits from the surounding shelves (Wilpshaar and Leereveld, 1994).

In the northern North Sea basins (Viking Graben), it corresponds to the break down of the Upper Ryazanian stratified water column which resulted in well oxygenated bottom conditions. The onset of the regressive phase is a major isochronous environmental change from dyseraerobic conditions below to oxygenated outershelf setting above (Rawson and Riley, 1982). It is also the lowestmost part of the shallow-water Mime limestone which represents a turnover event associated with the Lower Valanginian regression.

2nd ORDER CYCLE SET 12: UPPER VALANGINIAN TO LOWER APTIAN.

Main Characteristics

The boundaries of 2nd order cycle set 12 are dated as lowestmost Upper Valanginian (Sayneceras verrucosum ammonite Subzone) to Lower Aptian in the southeastern France (intra-Deshayesites deshayesi ammonite Zone) or to the Barremian/Aptian boundary (Boreal). The duration is estimated to 15 my (Va3 top lowstand to Ap3 top lowstand). The peak transgression is dated as lower Hauterivian (base Lycoceras nodosoplicatum ammonite Zone). The initial definition of 3rd-order depositional sequences is due to Arnaud and Arnaud-Vanneau (1989); Arnaud-Vanneau and Arnaud (1992; 1990); Brink et al. (1993); Bulot (1995); Charollais et al. (1994); Clavel et al. (1986a; 1986b; 1989; 1994a; 1994b); D’Argenio et al. (1993); Emmanuel (1993); Ferry and Rubino (1988–89); Gjelberd and Steel (1995), Hunt (1992); Hunt and Tucker (1992); Jacquin et al. (1991); Jacquin and Schlager (1993); Jan Du Chêne et al. (1991); Kühnau and Michelsen (1992); Hunt and Tucker (1993); Magniez-Jannin (1991), Magniez-Jannin and Jacquin (1990); Ruffel (1991); Rusciadelli and Jacquin (in prep.).

The cycle T/R 12 can be subdivided into four depositional sequences sets numbered as 12a (depositional sequences Va3 pars to Ha4 pars), 12b (depositional sequences Ha4 pars to Ba2), 12e (depositional sequences Ba2 to Ba3 pars on our chart) and 12d (depositional sequences Ba3 pars to Ap1 or Ap3). The stratigraphic record of these sub-cycles seems to be dependent of local sediment supplies, therefore they are not recorded systematically within all basins.

Tethyan Areas

Lowermost Unconformity.—

The lowermost unconformity is a major erosional unconformity which can be even angular at some places. In the Swiss Jura mountains, the unconformity surface is induced by the southwestward tilting of the platform with a level of erosion reaching the underlying micritic limestone of the Vions Formation belonging to regressive phase (Detraz, 1989).

Transgressive Phase T12 and Peak Transgression.—

The transgressive phase is well documented on the northern margin of the Vocontian Subalpine basin, from the Jura mountain to the French Subalpine chains in Savoy (Clavel et al., 1986a; 1986b; 1989; Busnardo and Thieluloy, 1989; Detraz, 1989; Reboulet et al., 1992; Ferry, 1991). It comprises the five backstepping Late Valanginian sequences numbered Va3 to Va7 on our chart. As a consequence of the overall backstepping, the depositional profile of the platform margins evolved progressively from a rimmed shelf during the Lower Valanginian to a distally steepened ramp during late Valanginian and lowermost Hauterivian times. Along that ramp, lithofacies typically grade from high-energy bioclastic limestone shelfward (Formation du Bourget) to ammonite-bearing marls and shales seaward. Third-order bioclastic lowstand deposits (locally named as Calcaires Roux) commonly develop along the slope during that transgressive phase. At the edge of the Vocontian basin this interval develops “healing type” sequences with thick shaly lowstand and highstand members (Jacquin and Graciansky, this volume). These backstepping sequences thin upward to a surface of drowning and a starved interval dated as Acanthodiscus radia tus Zone from the Jura mountains (Marnes à Astieria; Detraz, 1989; Remane, 1991) to the northern edge of the Vocontian basin. On the southern margin of the Vocontian Basin (Provence platform), the overall backstepping continues into the Lower
Hauterivian reaching the lowermost Lytoceras nodosoplicatum ammonite Zone (MFS sequence Ha2). This is consistent with Boreal basins.

The transgressive phase T12 is marked by major biological events: (1) renewal of the ammonites with great abundance and diversity (Reboulet et al., 1992), (2) significant decrease of the palynological Manderorgia and Systematophora groups which indicates the lowering of continental influx, (3) an increase of the Dinophyceae Cribroperidinium group pointing out the overall deepening (Wilshpaar and Leereveld, 1994), and (4) decrease of the benthic foraminifer content associated with dysaerobic bottom waters (Magniez-Jannin and Dommergues, 1994).

**Regressive Phase R12.**

The regressive phase R12 is a major progradational interval of the so-called Urgonian carbonate platforms surrounding the Tethyan basins. Their detailed sequence stratigraphic framework has been intensively studied in the French Subalpine chains during the last five years due to the high quality of the seismic-scale exposures. Dysaerobic to anoxic bottom-water conditions characterize several western European basinal settings at that time (Vocontian basin, southeastern France: Magniez-Jannin, 1991; southern Germany basin: Kemper, 1979; Kemper and Zimmerle, 1982; U.K. area: Rawson and Matterlos, 1983; North Atlantic: Graciansky et al., 1987; North Sea: work in progress). Several organic-rich layers develop from the Late Hauterivian to the lowermost Aptian 3rd-order condensed section with TOC maximum values up to 4.5%. Note these anoxic events correlate at distance with the time equivalent 2nd-order transgressive phases known from the North Sea to the Arctic areas.

The exact timing of the onset of the rudist lagoonal mudstone, typical of the Urgonian facies, is still a matter of debate. The stratigraphic interpretation rests upon two distinct conceptions of the distribution of large calcareous benthic foraminifers, which are the only biostratigraphic markers in the absence of ammonites within the Urgonian carbonate platforms. The Grenoble University conception (Arnaud-Vanneau and Arnaud, 1990) supports the idea that the Urgonian -rudist bearing- lagoonal mudstones only develop during the deposition of aggrading sequences dated as Upper Barremian through Lower Aptian. They unconformably overlie the Hauterivian infilling sequences landward, such as the Pierre Jaune de Neufchâtel Formation of the Jura Mountains. But they conformably overlie the Lower Barremian forestepping sequences seaward, such as the Glandasse bioclastic Formation of the southern Vercors. The two main points of that model are the following: (1) the forestepping sequences are sequences Ha7 to Barr5. They merge landward into a Lower Barremian unconformity, between the Pierre Jaune de Neufchâtel below and the Urgonian limestone above, and, (2) the Urgonian lagoonal mudstones only develop during the aggrading phase following the forestepping phase, which is dated of the late Barremian and upwards. The Geneva University conception (Clavel et al., 1995) supports the idea of an overall and continuous progradation of carbonate depositional environments, starting in the Lower Hauterivian (sequence Ha2) above the Marnes d’Hauterive Formation in the Jura, and ending in the Upper Barremian (sequence Barr5) in the southern Vercors. The carbonate depositional environments are diachronous following the overall progradation, the Urgonian lagoonal mudstone starting landward in the Jura in the middle Hauterivian (HST of sequence Ha4) and seaward in the northern Vercors in the lower Barremian (Gorges du Nant; sequence Barr2 probably). According to that model, the forestepping sequences are sequences Ha4 to Barr5 and merge landward into an upper Barremian unconformity. The Urgonian lagoonal facies is not limited to the aggradational phase, but starts earlier within the forestepping phase. Our previous definition of infilling, forestepping, aggrading and backstepping sequences sets (Jacquin et al., 1991) was resting originally upon the Grenoble University biostratigraphic conception. But the recent discovery in the southern Vercors of Lower Barremian ammonites laterally to the Urgonian lagoonal facies (forestepping sequences) would confirm the Genève University views (Jacquin and Schlager, 1993).

However, whatever the interpretation of the biostratigraphic data, four Hauterivian (Ha2 to Ha6), seven Barremian (Ha7 to Barr6) and two Aptian (Ap1, Ap2) 3rd-order depositional sequences are the building blocks of that regressive phase. They can be grouped on the basis of their facies and stratal stacking pattern into four progradational tongues: R12a (Pierre Jaune de Neuchâtel), R12b (Archiane lower tongue), R 12c (Archiane upper tongue), and R12d (uppermost Urgonian cliff). These are separated by backstepping phases: T12a (Angulicostata Marls), T12c (Fontaine Colombette Marls), and T12d (Heteroceras Marls).

**Boreal Areas**

**Lowermost Unconformity.**

As in the Tethyan areas, the lower boundary is a major erosional unconformity (Fyfe et al.). In the Paris Basin (Fig. 2), the Wealden Sandstone (Sables de Griselles) of the underlying regressive phase are totally truncated on the basin margin. In addition, this unconformity is marked by a hiatus covering the entire late Valanginian time over the whole area (Rusciadelli and Jacquin, in prep.). In the North Sea basins, the Lower/Upper Valanginian boundary corresponds to the most pronounced erosional unconformity of the Lower Cretaceous, with almost no known record of Late Valanginian strata except in subsiding areas (Rawson and Riley, 1982). This unconformity is the youngest and the last of the so-called Late Cimmerian unconformities that characterize the overall lowering of the sea level during Late Jurassic and Lower Cretaceous times.

**Transgressive/Regressive Subcycles 12a, 12b, 12c, 12d**

Within the northern European basins several Lower Cretaceous continental to deltaic sandstone tongues (regressive phases) interfinger with shallow-marine shale and mudstones (transgressive phases).

**Paris Basin.**

Only three sub-cycles are recorded (Fig. 2 and Rusciadelli and Jacquin in prep.). The first two marine incursions (T12a and T12b) are dated by ammonites (outcrops) and palynomorphs (outcrops and subsurface) as Lower and uppermost Hauterivian respectively. The lower Hauterivian incursion (T12a) corresponds to the Marnes Bleues that grade seaward to
the Calcaires à Spatangues. The uppermost Hauterivian incursion (T 12b) corresponds to the Marnes Jaunes. The regressive phases R12a (Sables de Chateau-Landon) and R12b (Sables de Chateau-Renard) comprise fluviatile to deltaic sandstone which prograde far from the basin margins. The maximum regressions coincide with widespread erosional unconformities along which hiatuses can be documented at the scale of the entire basin. The last sub-cycle (T/R 12c) do not correlate with its equivalent of the Tethyan areas. A main marine incursion (T12c) in the lower part of the Argiles Ostréennes, named as the Zone Lumachellique, occurred during lowermost upper Barremian time. It is followed by a widespread aggradational terrigenous interval (Argiles et Sables Panachés). Then the Paris Basin was totally emerged at the maximum regression (R12) at the end of Barremian times as a consequence of the rapid progradation of fluviatile environments.

**Western Netherlands and Lower Saxony Basins.**

A similar framework (but with four sub-cycles) is observed on the margins of these basins, around the Ardennes-Brabant highs and intrabasinal high (Nederlandse Aardolie, 1980). The peak transgression 12a coincides with the Westerbork Shales in the Lower Saxony Basin but a hiatus still develop during these times in the Western Netherlands. The peak transgression 12b coincides with the Berkel Shale and the 12c with the Jisselmonde Shales in both the West Netherlands and Lower Saxony Basins. The peak transgression 12d corresponds to the Eemhaven Shales in the West Netherlands basin, the time equivalent formation being truncated by the uppermost Barremian unconformity (maximum regression R12) in the Lower Saxony Basin.

**North Sea.**

Only the two lowermost sub-cycles T/R 12a and T/R 12b are strictly recorded. They form stacked retrogradational/progradational intervals within the Mime Limestones. The Lower Barremian transgressive phase T12b (Paracricoceras rarocinctum ammonite Zone) forms a widespread throughgoing fining- and thinning-upward succession with development of organic matter-rich black shales at the peak transgression (work in progress). The regressive phases (R12a and b) coincide with the return to aerobic environmental conditions as indicated by the overall reddish colors of the sediments. An erosional surface frequently marks the surfaces of maximum regression. A hiatus probably develops between cycles 12b and 12d. The Upper Barremian transgressive phase (T12d) marks the onset of an overall deepening phase with a thinning-upward pattern, which culminates with a widespread black shales at the peak transgression of the Aptian cycle T/R 13.

**Spitsbergen.**

The Helvetiafjellet Formation which forms a prograding/re-retreating deltaic system is dated as Barremian and overlies unconformably Upper Jurassic deep-marine shales (Gjelberg and Steel, 1995). The turning point between overall progradation and retrogradation is dated as Barremian with not more precise information. The transgressive phase culminates in the Lower Aptian such as in the North Sea.

**Cycle 12 Summary**

The two following points can be highlighted from the comparison between the Boreal and the Tethyan realms: (1) the four sub-cycles 12a, 12b, 12c and 12d of the Wealden domain (London-Paris Basin, Saxony and Netherlands) can be correlated with time-equivalent progradational depositional sequence sets of the Tethyan areas and (2) the four peak transgressions 12a, 12b, 12c and 12d could correspond in age to major 3rd-order flooding events in the Tethyan basins (base Lytiococeras nodosospliacatum, top Pseudothurmania angulicostata, Moutoniceras moutoni and Imoretites giraudi ammonite Zones respectively). However, the last 12d cycle is frankly transgressive in Spitzbergen, starved in the North Sea Basins, aggradational in the Paris Basin but still aggradational to progradational in the sub-Alpine (Tethyan) domain. Such discrepancies result from differences of sediment supply and tectonic development between individual basins.

**2ND ORDER CYCLE 13: APTIAN**

**Main Characteristics**

The boundaries are dated as Barremian/Aptian boundary (Boreal) or Lower Aptian in the southeastern France (intra-Des Hayesites deshayesi ammonite Zone) to uppermost Aptian (intra-Clansaysesian, boundary of Nolaniceras nolani/Hypacanthoptilites jaobi ammonite Zones, sensu Breistroffer, 1947). The duration is estimated to 8 my. The peak transgression is dated as Lower Aptian (Dehayesites deshayesi ammonite Zone). The initial definition of 3rd-order depositional sequences is same as for transgressive/regressive cycle 12, and d’Argenio et al., 1993; Hesselbo et al.; 1990; Ruffel, 1991; Ruffel and Wach, this volume.

**Lowermost Unconformity**

In the Nordic areas, the lowermost unconformity is the Ap1 sequence boundary which marks the onset of the Aptian transgression. It is not a major erosional unconformity. Within several basins, the Aptian transgression seems to be in continuity with the Late Barremian cycle T/R 12d (see above). In the Tethyan areas, the onset of the Aptian transgression is at the Ap3 sequence boundary, which is one of the major Mesozoic sea-level downward shift.

**Transgressive Phase and Peak Transgression**

The Early Aptian drowning (Des Hayesites deshayesi Zone) is one of the most widespread drowning events of the geological record (Jenkyns, 1980; Schlager, 1981). It is documented in different tectonic settings: on passive margins (Australia, Austria, southern France, northern Spain, Mexico), on isolated oceanic platforms (Apulia, Pacific gyauts), within intracratonic basins (Arabia), and also within active margins (Venezuela: Maracaibo Lake). This drowning is not limited to tropical carbonate platforms, but also developed on temperate siliciclastic shelves (Paris Basin and southern Germany) and within cold siliciclastic settings (Spitsberg).

The final and definitive drowning of the Urgonian platforms (MFS Ap3) is the peak transgression of transgressive/regressive cycle 13. It is preceded by a major third-order eustatic fall (Ap3b, 120 my coeval with the 112 my sequence boundary of the Haq et al. chart, 1988) that subaerially exposed the Urgonian platform (Arnaud-Vanneau and Arnaud, 1990; Jacquin and Vail, 1993).
A similar story can be documented from various Peritethyan carbonate platforms such as in Provence (South-eastern France: Masse, 1993), in Apulia (southern Italy), in Texas and New Mexico (Sligo and Cupido platforms: Scott, 1993; Goldhammer et al., 1991) and in the Arabian peninsula (Shuaiba platform: Alsharhan and Nairn, 1993; Pratt and Smewing, 1993). However the age of the peak transgression as well as the age of the 3rd-order sequences which forms the transgressive remains a problem of biostratigraphic correlation. Another point which would need further investigation concerns the growth potential of carbonate platforms with respect to the space being created at long term.

Within the northern Tethyan basins, the peak transgression (MFS Ap3) is a widespread anoxic event (Goguel Level: Breheret, 1994) with TOC values up to 4%.

Northwestern European Siliciclastic Shelves: London-Paris Basin, Western Netherland and Lower Saxony Basins.—

The transgressive phase is uniformly recorded over this wide area (Fig. 1) by marly open-marine deposits, often bituminous at the vicinity of the peak transgression (Western Netherland: Lower Holland Marls; Paris Basin: Argiles à Plicatules; Germany: Fish Shales (Kemper, 1979); U.K.: Hatherfield Clay in the Wessex basin (Hesselbo et al., 1990); Danish North Sea: Sola Formation (Jensen and Buchardt, 1987). On the northern border of the Paris Basin (Boulonnais), the Lower Aptian phosphatic nodule bed (Cat-Cornu horizon) are time equivalent of the peak transgression in the Paris Basin directly overlie Lower Cretaceous Wealden sandstones (Robaszynski and Amédro, 1986).

Northern North Sea.—

The transgressive phase T13 consists of fining-upward alternating black fissile shales and bioturbated mudstones known as Sola Formation. These lithologies look very similar to others deposited around the anoxic event OAE1 which has been described in the North Atlantic DSDP cores (Jenkyns, 1980, Graciansky et al., 1987). Landward on the Horda platform the transgressive phase T14 form a major backstepping interval (Skibeli et al., 1995) belonging to the Sola Formation.

Regressive Phase 13

The time interval is characterized by the Bay of Biscay opening (Montadert et al., 1989) and by the creation of the Norwegian-Greenland rift system (Ziegler, 1988). An echo of such events was found as far as the Subalpine zone (Hirsch et al., 1992; Joseph et al., 1989). The results were the renewal of erosion on shallow and emerged areas and the production of huge amounts of siliciclastics over the Western European Shelves. As a consequence of the frequent erosional surfaces and of the abundance of terrigenous deposits, the biostratigraphic calibration of 3rd-order sequences and their correlation from shelf to basinal areas is not yet accurate as for other stages. Additional work on the sequence stratigraphic point of view is still needed there. Nevertheless only three 3rd-order depositional sequences seem to form that regressive phase. The best documentation was found in the Vohcontian basin of the French Subalpine Zone (Ferry and Rubino, 1988a; b; Bréheret and Delamette, 1988) and in its transition to the Plate-forme Rhodanien (Rubino, 1988) and to the Helvetic platform (Delamette, 1989; 1994). This regressive phase ends at the top of the No- lanticeras nolani ammonite Zone as shown by the return to Wealden-type lithologies in the Anglo-Paris Basin, noticeably in the Boulonnais (Robaszynski et al., 1980).

2nd ORDER CYCLE 14, TETHYAN AREAS: UPPERMOST APTIAN TO MIDDLE/ UPPER ALBIAN

Main Characteristics

The boundaries are dated as uppermost Aptian (base of Hypacanthoplites jacobi ammonite Zone sensu Breistroffer, 1947) to Middle/Upper Albian (Dimorphoplites bicipicatus/Dipoloceras cristatum ammonite zonal boundary). The duration is 10 my. The peak transgression is dated as lower Albian (Leymeriella tardifurcata ammonite Zone, Leymeriella regularis Subzone). The initial definition of 3rd-order depositional sequences is due to Breheret and Delamette (1988); Ferry and Rubino (1988a, 1988b, 1989); Delamette (1989, 1994); Föllmi (1989); Gräfe (1994).

Lowermost Unconformity

The lowermost unconformity is a major basin-scale erosional surface which is over lain by a strongly backstepping interval. That surface is frequently associated with the renewal of terrigenous supply. It correlates also with the onset of extensional activity in relation with the North Atlantic margin development (onset of rift climax).

Transgressive Phase and Peak Transgression

The transgressive phase is one of the major flooding and drowning events of the Mesozoic record. In basinal areas, it corresponds to a regionally extensive black shale in the Tethys (Breheret, 1985) named as Paquier level and dated as Leymeriella tardifurcata ammonite Zone, Leymeriella regularis Subzone. This organic rich interval is coeval with the OA1 event (Jenkyns, 1980) of the Atlantic ocean (Robaszynski, 1989). Strong condensation does not permit good sequence stratigraphic resolution. Northern Peritethyan shelfal equivalent deposits are also highly starved. They form a widespread condensed phosphate conglomerate, known across the Alpine belt, and overlying unconformably older rocks (Delamette, 1989; Föllmi, 1989).

Regressive Phase

The sediment starvation associated with the onset of the Lower Albian transgressive phase lasted until the lower part of the Middle Albian in basinal settings. This means that the first well-recorded depositional sequence of the overall regressive phase is not older than A15. (Hoplitites dentatus ammonite Zone; Ferry and Rubino, 1989). During this regressive phase the rate of sediment accumulation in the basins increased sharply. Correlated with the overall preservation of the organic matter decreases. Surrounding shelfal members are strongly prograding. On the borders of the Vohcontian basin, shoreface siliciclastic glauconitic sandstones are reported by Ferry and Rubino (1989).
2nd ORDER CYCLE SET 14: NORTH SEA, EASTERN ATLANTIC MARGIN FROM IBERIA TO SCANDINAVIA AND LONDON-PARIS BASIN

Main Characteristics

The boundaries are dated as uppermost Aptian (boundary of the Nolaniceras nolani/Hypacanthoplites jacobii ammonite Zones) to upper/lower Albian boundary. The duration approximately is 10 my. The peak transgression is dated as lowermost Middle Albian (Hoplitites dentatus ammonite Zone, Hoplitites benettianus = Lyelliceras lyelli Subzone). The initial definition of 3rd-order depositional sequences is due to Juignet, 1980; Hesselbo et al., 1990; Ruffel, 1991; Ruffel and Wach, 1991; Amédro, 1992; García-Mondejar and Fernandez-Mendiola, 1993; Skibel et al., 1995; Ruffel and Wach, 1991. The work is still in progress in the Boreal realm.

The time equivalent of the Tethyan 2nd order cycle 14 can be subdivided into two 2nd order cycles numbered as 14a (depositional sequences Ap5 pars to A13) and 14b (depositional sequences A14 to A16).

North Sea.—

Two widespread and correlatable starved intervals can be documented and dated from micropaleontological and palynological data. One is Leymeriella tardefurcata Zone equivalent and the other one is Hoplitites dentatus Zone equivalent. These two condensed sections are characterized by high abundance and diversity of organisms. Frequently a hiatus associated with an erosional surface separates these two flooding events. It has been tentatively calibrated to the A14 sequence boundary and interpreted as the 2nd-order maximum regression R14a, but the work is still in progress.

London-Paris Basin.—

The lowermost unconformity is a strong erosional surface particularly on the basin margins (Fig. 2 and Rusciadelli and Jacquin, in prep.). Upper Aptian transgressive deposits may rest unconformably on various levels of the underlying Cretaceous and Jurassic strata (Juignet, 1980; Amédro, 1985; Robaszynski and Amédro, 1986; Rat et al., 1987, Amédro et al., 1995). As in the Tethyan areas, the Leymeriella tardefurcata through Hoplitites dentatus interval (Lower Albian) can be highly starred and combine into a single condensed section. But when moving westwards, (i.e., towards more subsiding areas with high rate of sediment supply), this condensed section separates into two subcycles, one with the peak starvation in Leymeriella tardefurcata Zone, Leymeriella regularis Subzone, the other one in the Lyelliceras lyelli = Hoplitites benettianus Subzone (Amédro, 1992; Amédro et al., 1995). The peak transgression dated as Lyelliceras lyelli Subzone is marked by a temporary maximum abundance of planktonic foraminifers (Hedbergelliids and Faivuselliids: Magniez-Jannin, 1979; 1983), the appearance of foraminifers and nanofossils on the northern margin of the Paris Basin (Ardennes border of the Paris Basin: Amédro, 1985) and the association of ammonite faunas of both Boreal and Tethyan origins which indicate a full connection between the North Sea and the Tethyan through the Paris Basin (Amédro, 1992).

The Lower Albian R14a maximum regression and unconformity is indicated by a widespread hiatus which has been followed on both side of the English Channel (Amédro, 1992; Hesselbo et al., 1990). It separates regressive sand-rich layers (Folkestone Beds in U.K. and Sables Verts in the Paris Basin) below from clay-rich layers (typical Gault facies) dated as Middle Albian above.

The Middle Albian regressive phase following the Hoplitites dentatus flooding is characterized by continuous and widespread clay-rich deposits over the entire Paris basin, indicative of renewal and homogenization of the subsidence pattern. This trend is coeval with the resuming of the Middle Albian sedimentation on the northern Tethyan margins.

2ND ORDER CYCLE SET 15: MIDDLE/UPPER ALBIAN TO ALBIAN/ CENOMANIAN BOUNDARIES

Main Characteristics

The boundaries are dated as lowermost upper Albian (Diploceras cristatum/Mortoniceras pricei Subzones boundary) to Albian/Cenomanian boundary. The uppermost boundary is still questionable. It could coincide with the major Middle Cenomanian relative sea-level downward shift in the Acanthoceras jukesbrownei ammonite Zone as shown by Juignet and Breton (1991a; 1991b) in the Paris Basin. The duration is 3 my (or 6 my ?); it coincides with depositional sequences A1 7 to A1 10). The initial definition of depositional sequences is same as for cycle 14, and Robaszynski and Amédro (1986); Juignet and Breton (1991a, 1991b). The peak transgression is dated as Mortoniceras inflatum Zone, Upper Albian.

Lowermost Boundary

There is a well-marked widespread (worldwide) erosional surface between cycles 14 and 15, as documented noticeably by Owen (1971). A hiatus develops in western Normandy (northwestern Paris Basin) between the Middle Albian Gault Formation and the Upper Albian Gaize Formation (Juignet, 1980).

Transgressive and Regressive Phases

The overall transgressive phase in the Tethyan basins is relatively starved. It corresponds to the Breistroffer Level (TOC: up to 2%) dated as of the Stoliczkaia dispar ammonite Zone in southeastern France (Bréheret, 1994). It is not as much condensed as the underlying Albian transgressive half-cycles. Following the Late Albian disconformity the T15 transgressive phase in the London Paris Basin is the last major bioevent of the Albian, where large-scale communication between Tethyan and Boreal realms can be documented. Sediments are frequently glauconitic, ferruginous and phosphatic such as in southeastern England (bed XII of Owen, 1971), in northern France (Formation de Lottinghen; Robaszynski and Amédro, 1989; Gaize de Youziers, Amédro et al., 1995). Backstepping depositional sequences of the long term transgressive interval thin upwards and step-back further landward onto older rocks including Late Jurassic layers (Juignet, 1980). The peak transgression is the last widespread clay-rich layer (Gault facies) of the Albian time. The uppermost Albian regressive phase is indicated by the renewal of clastic sedimentary supply, both siliciclastic and carbonate, in all basins. The Albian/Cenomanian (Ce1) sequence boundary is a pronounced relative sea-level downward shift with a pronounced hiatus in shelfal settings.
CONCLUSION

The lowermost boundary of the North Atlantic cycle is defined by the onset of the *Surites* (Bojarkia) *stemonophalus* transgression (Upper Ryazanian) onto the western European craton. It is the major turning point between the overall progradation of the underlying Late Jurassic/Lowermost Cretaceous regressive phase and the overall Cretaceous transgression (Fig. 2). This surface is not necessarily the most erosional of the interval. The Lower Valanginian unconformity between cycles 12 and 11 frequently truncates much more deeply the underlying sediments. This particular unconformity is the last of the four Late Cimmerian unconformities which are dated respectively as Lower Volgian, Upper Volgian, Late Ryazanian and Lower Valanginian.

The Lower Cretaceous succession has a characteristic long-term, 2nd-order, stratigraphic signature, which may provide the means to predict the age of sedimentary deposits ahead of the drill or biostratigraphic evaluation (Vail et al., 1991). It can be summarized as follows:

—The lowermost Cretaceous cycles 11, 12a, 12b and 12c lap out at or near the shelf margins. They are strongly prograding during the regressive phases, commonly ending with major lowstand deposits. As an example, the top of the Barremian progradational carbonate platform in southeastern France (regressive phase 12c) coincides with the maximum seaward extent of shallow water complex towards the basin for the entire Mesozoic section. Similar observations can be done in the North Sea for the same interval.

—The Late Barremian cycle 12d is highly aggradational, indicating a lot of accommodation space being created at that time together with high rates of sediment supply. This is true in both the Tethyan carbonate and Nordic siliciclastic settings.

—The APTian through Turonian cycles (numbered as 13 through 17) step back quickly landward, with major flooding and drowning at or near peak transgressions and major progradation at or near maximum regressions. However the overall trend is clearly and rapidly backstepping instead of being mostly aggradational as previously. The unconformities which limit these cycles are linked with major intraplate tectonic events, in connection with the rifting episodes in the Bay of Biscay. The peak transgression of these cycles lead to the development of starved conditions in basinal settings, with widespread occurrence of organic-rich layers, coevally with the oceanic anoxic event 1 in the Atlantic Ocean. Both the duration and amplitude of the 3rd-order relative sea-level changes differ from the pre-Aptian times (ice house versus greenhouse?).

High-amplitude high-frequency cycles characterized pre-Aptian deposits whereas low amplitude low frequency fluctuations dominate during Aptian and Albian times.

—The Aptian drowning T13 is one of the most widespread and correlatable of the Mesozoic record with even much better organic-matter preservation in the European basinal settings than the well-known Cenomanian-Turonian event. The Lower Albian peak transgression (14a) merges frequently by sediment starvation with the Middle Albian one (14b). It is why they have been frequently interpreted as a single event.

—The Late Albian regressive phase R15 is strongly aggradational with multiple high frequency lowstand deposits and ends with a major sea-level downward shift at the Albian/Cenomanian boundary.

The Lower Cretaceous long term stratal pattern seems to be closely dependent of the western European Craton tectonic evolution. The pre-Aptian mostly aggradational cycles relate to relatively stable conditions between the Late Jurassic North Sea and Middle Cretaceous North Atlantic rifting phases. The Lower Valanginian and Barremian unconformities are associated with intraplate transpressional events recording the changes between Late Jurassic and Cretaceous rift systems. The post-Aptian mostly retrogradational cycles with a rapid encroachment towards the continent relate to the successive rift climax in the neighboring Atlantic realm. The overall 1st-order Late Cretaceous regressive phase developed during the North Atlantic post-rift phase.

REFERENCES


2ND-ORDER TRANSGRESSIVE/REGRESSIVE FACIES CYCLES IN THE LOWER CRETACEOUS OF WESTERN EUROPE


GRAFI, K.-U., 1994, Sequence Stratigraphy in the Cretaceous and Paleogene (Aptian to Eocene) of the Basco-Cantabrian Basin (N. Spain), Tübinger Geowissenschaftliche Arbeiten (TGA), Band 18, 417 p.


JACQU, T., ARNAUD-VANNEAU, A., ARNAUD, H., RAVENNE, C., and VAIL, P. R., 1991, Systems tracts and depositional sequences in a carbonate setting:
ABSTRACT: In this work models are developed for the sequence stratigraphic interpretation of estuarine-shoreface-offshore reservoir sands found throughout Europe at the Aptian-Albian boundary. The models so developed are critically appraised through vertical facies analysis, facies distribution and paleogeographic reconstruction combined with correlation of depositional sequences. This process leads to the recognition of depositional sequences for which definitive criteria may vary, depending on paleogeographic and tectonic location. It is only after the effects of paleogeographic position are considered that an accurate account of sea-level history can be deduced in different environments. Arenaceous strata are present throughout southern England at the Aptian-Albian stage boundary (Lower Cretaceous); for example, the Sandrock/Folkestone Formation/Woburn Sands Formations occur both as part of a conformable succession in the Weald and Channel Basins and as an unconformable succession onlapping Jurassic and older strata of the basin margins. The facies and stratigraphy of three type successions are highly variable, suggesting environmental contrasts in Aptian-Albian times across southern England. Comparison of sequence stratigraphic analyses from each succession indicates that the depositional facies developed and preserved at sequence boundaries, as well as the number of sequence boundaries resolved, varies in direct response to the paleoenvironment.

RECOGNITION OF ESTUARINE-SHELF DEPOSITIONAL SEQUENCES

General

In this study, facies analysis and the application of the sequence stratigraphic methods described in Van Wagoner et al. (1990) were used on marine-estuarine successions of eastern and southern England across the Aptian-Albian boundary. At an early stage in this work, different stratigraphic levels in different paleogeographic locations and with highly variable facies were used in defining depositional sequences. We were able to develop a set of criteria in three key areas, providing the basis for depositional sequence and paleogeographic facies models to describe and correlate Aptian-Albian estuarine—offshore strata.

Cross-stratified and clay-draped quartzose sandstones are found in the Upper Aptian-Lower Albian (mid-Cretaceous) strata in southern and eastern England and form a distinctive lithofacies on outcrop (Fig. 1). These contrast with the Lower Aptian strata below that are composed of grey-green bioturbated sandstone and the overlying Albian Gault claystones (a massive, bioturbated, dark grey to black silty clay). Marine fossils are generally rare in these Aptian-Albian strata, yet they are common in the strata above and below, indicating differential preservation and, most importantly, the existence of upper and lower (bounding) marine flooding surfaces onlapping onto regional unconformities. Facies documentation reveals higher-frequency sequences within the Aptian-Albian strata, some of which correspond to parasequence sets.

The Aptian-Albian strata of southern and eastern England are bounded at their upper and lower surface by regional unconformities associated with phases of active extensional tectonics that appear to be synchronous with eustatic lowstand-highstand cycles (Rawson and Riley, 1982; Ruffell, 1992). The tectonically-influenced unconformities are overlain by a flooding surface (Ruffell and Wach, 1991). Within the Aptian-Albian strata, the geometries of estuarine and offshore deposits vary widely. Two common parasequence stacking geometries are presented in Fig. 2. These may be characterized by the predominance of retrogradational or progradational parasequence set geometries.

Retrogradational Parasequence Sets

Retrogradational sets are commonly preserved in incised valleys and represent fluvial/estuarine environments. The upper parts to these sets show strong reworking within the transition from the lowstand through the transgressive and highstand systems tract (Fig. 2). The erosional topography provides early accommodation space during this stage of sea-level rise. When this topography is filled a low-gradient shoreface develops, over which the effects of transgression will be widespread. Evidence for previous incision is rarely preserved except in channels and offshore environments. In shoreface sediments, the late transgressive-highstand systems tract forms a relatively conformable retrogradational succession, with each parasequence and encompassing parasequence set capped by a minor transgressive flooding surface.

Progradational Parasequence Sets

Progradational parasequence sets develop where sea-level fall is slow or where tectonic subsidence equals or outstrips eustatic fall, resulting in decreased accommodation space (Van Wagoner et al., 1990). Thus the offshore is not subjected to erosion, and the decrease in bathymetry is marked by a “turn around” from offshore to nearshore estuarine sediments. Any sequence boundary is marked by a correlative conformity found down dip from the unconformable surface. Recognition of the surface may depend on outcrop analysis of mud to silt ratios and laboratory analysis of increasing kaolinite and brackish microfauna and may otherwise be confused with a maximum flooding surface (see Fig. 2 and discussion of Isle of Wight). Conversely a gradual basinward shift of facies may be present, with no single surface dominating the process of incision. Retrogradational and progradational parasequences in the estuarine environment may not stack in any consistent manner and definition of parasequence sets may prove difficult.
TECTONIC SETTING

Cretaceous System: Southern and Eastern England

Early Cretaceous basin development in southern England was characterized by the formation of two major depocenters: the Weald and Channel sub-basins of the Wessex Basin “complex” of Stoneley (1982). Following marine withdrawal from these basins in latest Jurassic times, deposition of the non-marine limestones and evaporites of the Purbeck Group was followed by non-marine Wealden Group sands and clays. The thickest Lower Cretaceous deposits are preserved in what were probably strongly subsident areas, controlled by basin-bounding syn-sedimentary faults (e.g., Hog’s Back Fault, Weald Basin; Isle of Wight Monocline, Channel Basin; see Fig. 1). For this reason, the Wealden succession is thought to represent a rift/synrift stage in the evolution of the Weald and Channel Basins (Karner, Lake and Dewey, 1987).

An overall rise in sea level occurred from Aptian to Albian times, punctuated by brief but significant sea-level falls. Mid-Aptian- to Albian-age strata are thus found onlapping and overstepping the margins of the Wessex Basin and are observed presently resting with angular unconformity on faulted and folded Mesozoic-Paleozoic strata. The successions under consideration are interpreted to be part of an early post-rift succession (Karner, et al., 1987).

Aptian-Albian Stages: Southern and Eastern England

Aptian—Albian strata in southern England are preserved in two tectono-stratigraphic positions, consideration of which is essential to the understanding of depositional sequence stratigraphy. The first is a relatively conformable basin-fill succession. The second is found resting directly on faulted-folded Jurassic, Triassic or Paleozoic sediments and metasediments of the basin-margin, occasionally with a thin veneer of earlier Aptian or non-marine Lower Cretaceous (Wealden) strata preserved below. Basin-fill successions include those found in the Weald and Channel Basins, which were both inverted in Upper Cretaceous and Miocene times. Basin-margin successions are widespread in the subsurface of southern England below Upper Cretaceous and Cenozoic cover. Exposures of this marginal succession occur along a northern outcrop of the Cretaceous strata in Eastern England and the southern Midlands, one of the best-known examples being Leighton Buzzard (Casey, 1961).

Subsurface borehole records of the Aptian-Albian strata are extensive and generally of good quality in southern England. However, the majority of hydrocarbon tests have been drilled on structural highs, where the Lower Greensand strata are found to be thin, condensed correlatives of strata found at outcrop (Ruffell, 1991). Nonetheless, these subsurface records demonstrate the areal extent of Aptian-Albian strata preserved across former Lower Cretaceous basin margins and intra-basinal highs as a result of the transgressive phases during the \textit{nutfieldiensis} and \textit{jacobi} zones (Bridges, 1982).

FACIES

General

The lithofacies of the Aptian-Albian strata in southern England comprise predominantly white or yellow, orthoquartzose, medium-grained, moderately-rounded and often un lithified sandstone. The beds are cross-stratified, ripple cross-laminated with abundant clay drapes. This broad lithofacies was common to all Aptian-Albian strata in England, and represents depositional conditions that ranged from brackish restricted estuarine facies in the area of the Isle of Wight to more open estuarine and marine tidal conditions throughout the Weald and Leighton Buzzard outcrops. Sediment sources lay to the north and west, although transport/rewriting mechanisms were variable (Garden, 1991).

These uppermost beds of the Lower Greensand Group are known variously as the Sandrock Formation (Isle of Wight); Folkestone Formation (Weald); Calne Sands (Wiltshire), and Woburn Sands (eastern England). They are here collectively referred to as “Aptian-Albian strata”. Similar Aptian-Albian strata are known from subsurface records in southern England, the English Channel and specific areas of the North Sea and Western Approaches Trough. In the North Celtic Sea Basin (included on Fig. 3), arenaceous Aptian-Albian strata form the...
uppermost beds of the Kinsale Head Gasfield Reservoir (Colley et al., 1981). Fossiliferous and calcareous-cemented cross-stratified sands are found in the Tethyan basins of southern France (‘Lumachelle’: Middlemess and Moullade, 1970). Conversely, condensed, glauconitic, time-equivalent strata are common on highs in the North Sea (the “Greensand Streak” at Speeton, Rawson et al., 1978; Texel Greensand, N.A.M. and R.G.D. 1980). There is a considerable volume of published and unpublished data on each of the successions listed above. Two recent comparative studies concern the stratigraphy and sea-level changes of the whole Lower Greensand Group (Bridges, 1982; Ruffell and Wach, 1991).

The facies of the beds marking the Aptian-Albian boundary found in the Isle of Wight, Weald and Leighton Buzzard Lower Greensand Group successions are variable: estuarine facies with 5 to 20-m thick mudstone intercalations pass north into an offshore, tide-dominated sand wave facies relatively poor in mudstone (Fig. 3). The aim of this study is to compare these variations (with reference to pre-existing work on facies, bedform size/geometry and sediment provenance) within the depositional sequence stratigraphic framework set out in Fig. 2. Facies strongly controlled the development and preservation of depositional sequences. The criteria used in defining and correlating the depositional sequences included facies analysis, parasequence stacking patterns, and geochemical characterisation of clay mineralogy.

The well-preserved sedimentary structures of the Folkestone Formation and Woburn Sands have been the focus of much research on the sedimentological processes responsible for their formation (Allen and Narayan, 1964; Narayan, 1971; Allen, 1982). These works have demonstrated a tidal control on the deposition of these offshore sand wave, shoreface and estuarine sediments.

The Isle of Wight (Channel Basin) succession has been documented in Dike (1972a,b); Nio (1976); Wach and Ruffell (1991) and Wach (1991). These also suggest that tidal processes dominated sedimentation. Shallow-water facies are preserved at various horizons in the Sandrock succession of the Isle of Wight, and are also recorded in the topmost beds of the Woburn Sands and Folkestone Formation (this study).

**Facies and Correlation**

Outcrop and subsurface examination of the Aptian-Albian Lower Greensand Group (Ruffell and Wach, 1991), combined with a review of historical published literature, has revealed the stratigraphic variations found between key sections in southern England (Fig. 4). There is a lack of biostratigraphic control on correlations between the study areas. A consequence of this is that basic variations in stratigraphy may be due to diachronity. However, each succession is biostratigraphically constrained at the top and base, and the sequence stratigraphic interpretation
Fig. 3.—Palaeogeography of southern England (and surrounding areas) in Upper Aptian—Lower Albian times, with locations mentioned in text introduction. F = Faringdon; C = Calne. Box insert = limits of location map in Fig. 5. Similar palaeogeography to that depicted would have developed under transgressive conditions immediately after incision and the formation of a sequence boundary.

illustrates that although the facies relationships are variable, correlation is possible. There are many smaller and less well-exposed sections throughout southern England than those described here. For full details, the reader is advised to consult Casey (1961). The Weald and Isle of Wight are the thickest and best-exposed sections available and they are considered representative of the majority of exposures.

DOCUMENTATION OF DEPOSITIONAL SEQUENCES

Weald (Folkestone Formation)

The Folkestone Formation represents Aptian-Albian strata exposed in the Weald Basin. “Folkestone Beds” (as previously used by Allen, 1982; Casey, 1961; Ruffell, 1990) is a misnomer; the type section at Folkestone (TR240363) is developed as a calcareous/siliceous cemented succession, the upper parts of which are younger than much of the exposed Folkestone Formation of the rest of the Weald. This stratigraphic variation was demonstrated by Casey (1961), and may be due to the thicker *Douvilleiceras mammillatum* zone deposits of the type locality being deposited in an area where continuous fault-controlled subsidence added accommodation space and preserved sediments eroded from adjacent structural highs (e.g., fault-blocks).

The Folkestone Formation sediments of the type locality are poorly- to moderately-sorted glauconitic quartzose sands with calcareous- or silica-cemented “stone bands”. These cemented beds preserve a firmground fauna of large bivalves, echinoids and bryozoa. Thin sections reveal the presence of sponge spicules, which led Padgham (1970) to invoke an early diagenetic origin to the silica cement. Phosphate nodule beds with concentrations of reworked ammonite steinkerns and increased radioactive content (Ruffell, 1990) suggest that even in this subsident area, the succession is incomplete, and has a number of hiatuses. The upper parts of the Folkestone Formation developed at the type locality downlap onto the basal beds of the Gault Clay throughout much of the rest of the Weald Basin (Fig. 5; Table 1).

Throughout the remainder of the Weald outcrop, the Folkestone Formation comprises unconsolidated well-sorted sands that are usually cross-stratified. Pebby or argillaceous beds are also common (Fig. 5). In areas of good outcrop, large subtidal channels or scours (up to 200 m across) are discernible. It is within these channels that the sand wave bedforms described by Allen (1982) occur; they are typical of the lowstand and early transgressive deposits found in dominantly retrogradational depositional parasequence sets. The Folkestone Formation outcrop is large (Fig. 5) and thus the bedforms vary in structure and size, ranging from 1 to 3 m high. These are often the first preserved deposits of a transgressive systems tract fol-
FIG. 4.—Three different stratigraphic interpretations of key successions discussed in text, showing stratigraphic variation between the variable shelf—estuary succession of the Isle of Wight, and shelf succession of the Weald and Leighton Buzzard. The formations are biostratigraphically-constrained at top and base, and by occasional ammonites in the centre: otherwise correlation is poorly constrained. Correlations A and B show how the Isle of Wight succession could be younger or older than that at Leighton Buzzard. The sequence stratigraphic correlation developed here (C) provides a realistic and conservative correlation, and is thus the best.

lowing incision. Nio (1976) recognized sand waves as forming (and being preserved) under transgressive conditions. In the eastern part of the North Downs, well-sorted fine to medium sands infill channels with a north-south orientation. At the western end of the North Downs, the exposed Lower Greensand represents strata preserved in the depocenter of the Weald Basin (Fig. 5, Locations 8–11). Consequently, a thicker Folkestone Formation succession (up to 130 m), comprising poorly-sorted, pebbly and argillaceous sand, is found. The underlying formations of the Lower Greensand Group are also thickest and of a medium to coarse-grained nature in this area: the Bargate Member, found a few meters below the Folkestone Formation contains abundant pebble-size clasts of derived Upper Jurassic lithologies. This greater thickness and lithological variation may be attributed to the proximity of the basin-bounding faults (Hog’s Back) along the London-Brabant Massif sediment source terrain, which was to the north (Fig. 1).

Biostratigraphic control on the age of the Folkestone Formation is poorly constrained because fossils are generally rare. The ammonites used by Spath (1923), Casey (1961) and Owen (1988) to construct biostratigraphic schemes, have been obtained from phosphate nodule beds common at the Folkestone type section. These are interpreted to be condensed sections formed during an hiatus in sedimentation (Ruffell, 1990). Similar, early diagenetic nodule beds are found associated with clay beds in the centre of the Folkestone Formation of the Farnham area (Fig. 5), where they preserve a variety of fossils in a mixed calcareous/phosphatic matrix (Shepherd, 1934; Wright and Wright, 1942; Casey, 1961). The ammonites from this smectite-rich clay band formed the basis for Casey’s (1961) *Farnhamia farnhamensis* subzone (*Leymeriella tardefurcata* zone). The same bed forms a fuller’s earth bentonite bed (Morgan, Highley and Bland, 1978) near Godstone 30 km to the east of Farnham and may be equivalent to one of the clay beds within the Sandrock Formation of the Isle of Wight (Fig. 4). The presence of marine fossiliferous nodules above brackish estuarine sediments is indicative of a flooding surface. Here the maximum flooding surface is found capping retrogradational parasequence sets.

The basal beds of the Folkestone Formation are transgressive and onlap the London Platform over a greater area than the underlying Sandgate Formation below (Middlemiss, 1962). It must be noted however that the exact position of the base of the formation is unclear and not well-dated.

Discussion: Folkestone Formation

From regional sequence stratigraphic correlation of the Folkestone Formation, it is evident that the type section at Folkestone displays a unique facies younger than other Aptian-Albian outcrops of the Weald. Here, the succession of phosphate nodule beds represent a number of sequences condensed into the basal Gault deposits of the rest of the Weald Basin (Hesselbo et al., 1990). Away from Folkestone, definition of depositional sequences may overcome the imprecise biostratigraphic control available and enable us to resolve an *Hypacantholithes anglicus* subzone sequence boundary. Above this sequence boundary, some locations display coarse-grained deposits representative of the lowstand and early transgressive systems tract (e.g., Guildford, Location 8, Fig. 5). Claystones of the *Farnhamia farnhamensis* subzone clays cap the infill of these coarse sediments, or rest directly on the sequence boundary (e.g., Kingsley, Location 4 and Borough Green, Location 13, Fig. 5). The claystone of the *Farnhamia farnhamensis* subzone represents basinwide flooding during the latter stages of the transgressive systems tract. These sections display retrogradational parasequence sets with common amalgamation of flooding surfaces upon the sequence boundary.

Leighton Buzzard (Woburn Sands)

A large volume of published work exists on the Lower Greensand Group of the Leighton Buzzard area, including the
comprehensive accounts of sedimentology and sequence stratigraphy of Johnson and Levell (1995) and Wonham and Elliot (1996). These works provide an excellent lithostratigraphic framework, which the summary of Fig. 6 is based on.

The Woburn Sands of the Leighton Buzzard area average 70 m thick, although boreholes to the northeast and southeast indicate Lower Greensand in excess of 100 m thick (Shephard-Thorn et al., 1986). The beds rest unconformably on Upper Jurassic Oxford, Ampthill or Kimmeridge Clay, and are well-exposed in numerous sand-pits north of Leighton Buzzard (Table 2). The sands can be subdivided into upper and lower intervals. The Lower Woburn Sands are cross-stratified, clay-draped, well-sorted, fine- to medium-grained sandstones, usually unlithified and totalling around 30 m in thickness. Bedforms average 1 m in height or more. Cross-sets can occur with no clay drapes, and an erosive, sometimes pebbly base. These probably represent storm-enhanced tidal deposits.

The Upper Woburn Sands comprise coarse to medium-grained, unconsolidated white sandstones that are mined for quality quartz sand. The beds are up to 20 m in thickness, and contain 0.5- to 2- m-high bedforms interpreted as sand waves formed during the latter stages of the transgressive and early highstand systems tract. Preservation of the sand waves occurred during the remainder of the highstand systems tract when intense bioturbation indicates reworking of the sand wave upper surfaces. The base of the Upper Woburn Sands comprises a pebble bed containing phosphate and quartz pebbles, clay rip-up clasts and abraded, derived fossils (Fig. 6). This horizon is undated and may represent the Hypacanthoplites jacobi zone unconformity (Hesselbo et al. 1990; Ruffell and Wignall, 1990), or the Farnhamensis farnhamensis hiatus described above. No thick clay beds (over 1 m) such as those found in the Sandrock Formation of the Isle of Wight, or of the Folkestone Formation of the Weald, are preserved.

The Silty Beds (Fig. 6) comprise fine glauconitic beds showing slight discordance with respect to the underlying, retrogradational Upper Sands, the result of being deposited in a quiet marine environment. These fine-grained sands and silts are preserved in broad, low-relief channels. The sandstone bodies of the Coarse Red Sands are interpreted as fluvial channels which cut down through the Silty Beds and Upper Sands to rest on Lower Sands at some localities (Fig. 6). The Coarse Red Sands are cemented by iron oxides and infill channels up to 6 m deep. Sedimentary structures in the coarse-grained fill of these chan-
nals may be difficult to observe: some lateral accretion surfaces and trough cross-stratification occur. The channels are interpreted here to represent lowstand deposits.

**Discussion: Woburn Sands Formation**

The lower surface of the Woburn Sands appears to represent a transgressive surface above a sequence boundary, as there is clear truncation of Jurassic strata below. This is another example of an amalgamated transgressive flooding surface and sequence boundary which has been called a “flooding surface/sequence boundary (FS/SB)” by Van Wagoner et al. (1990).

A second transgressive surface lies above at the Lower—Upper Woburn Sands (Formation) sequence boundary, although the degree of truncation is not always obvious at outcrop, due to a lack of marker beds in the Lower Sands. The sand waves of the Upper Woburn Sands indicate the development of strong, offshore tidal currents following a change in relative sea level, accompanied by modification of the basin geometry after deposition of the Lower Sands in a shoreface environment. A similar ravinement occurs at the base of the Silty Beds above (Fig. 6). A relative sea level fall, associated with incision, formed the channels later filled by Coarse Red Sands. The base of the Coarse Red Sands is interpreted by Wonham and Elliot (1996) as a sequence boundary (Fig. 6).

**Isle of Wight**

The Sandrock Formation of the Lower Greensand Group represents the late Aptian—early Albian strata exposed on the Isle of Wight (Table 3). These strata outcrop along the northern margin of the Channel Basin. The Sandrock Formation is thus part of a conformable basin-fill succession, the stratigraphy of

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**TABLE 1.—GAULT-LOWER GREENSAND STRATIGRAPHIC SUCCESSION THROUGHOUT THE WEALD BASIN**

<table>
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<tr>
<th>STRATIGRAPHIC INTERVAL</th>
<th>ENVIRONMENT and LITHOLOGY</th>
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<tr>
<td><strong>GAULT CLAY</strong></td>
<td>Open marine: dark grey, silty clay, calcareous and phosphate nodules</td>
</tr>
<tr>
<td><strong>LOWER GREENSAND GROUP</strong></td>
<td></td>
</tr>
<tr>
<td>Folkestone Formation</td>
<td>Offshore tidal and waves: cross-stratified sandstones</td>
</tr>
<tr>
<td>Sandgate, Hythe and Atherfield Fms.</td>
<td>Offshore marine: glauconitic, fossiliferous and bioturbated sands, clays (often bentonites)</td>
</tr>
</tbody>
</table>

*Stratigraphic terminology from Rawson (1992).
which is shown in Fig. 7. Much of the Aptian Lower Greensand on the Isle of Wight comprises bioturbated, mixed argillaceous/glauconitic-limonitic, quartzose and fossiliferous sands. In the upper beds of the group, white, cross-stratified quartzose sands with clay drapes form the Sandrock Formation. The type stratigraphic section for the Sandrock Formation is at Blackgang, in Chale Bay (SZ 485 767): the succession is summarized in Fig. 7. The known biostratigraphy (Casey, 1961) suggests that the Sandrock is broadly equivalent to the Folkestone Formation and Woburn Sands, yet the lithofacies are somewhat different. Beds may be bioturbated with glauconite representing offshore deposits (similar to the ammonite-bearing *Farnhamia farnhamensis* clays of the Weald Basin), or they may be laminated black clays with fine white sand laminae that are interpreted as estuary-fill deposits (Wach, 1991). Recognized facies transitions between these estuarine-fill deposits and shoreface sands are shown on Fig. 7 and Table 3. The sediments of the Sandrock Formation form laterally extensive sand bodies separated by 10 to 20 m-thick intervals of bioturbated muddy sandstone and laminated mudstone. The exposed sections show rapid lateral variations in thickness. Sands generally coarsen-upwards and contain evidence of erosional surfaces, flaser bedding, wavy lenticular bedding, unidirectional and bi-directional bedding, trough cross-bedding, large-scale ripple forms, intraformational clasts and lag deposits. Bedforms average 1 m thick or less. The sedimentary structures are interpreted to be tidal shoals and channel-fill. The mud facies are either finely-laminated or extensively bioturbated: the presence of a brackish microfauna in the laminated beds (D. J. Batten pers. commun., 1989) suggests marginal-marine conditions such as the mud-fill in a subtidal estuary or intertidal flats. Detailed lithofacies and ichnofacies descriptions of the Sandrock Formation are given in Wach and Ruffell (1990) and Wach (1991).

**PALAEOGEOGRAPHIC INFLUENCE ON DEPOSITIONAL SEQUENCES: CONCLUSIONS**

Two essential differences between the depositional sequences developed in each of the three study areas are the variation in bedform size and the argillaceous content of each stratigraphic succession. The basic lithofacies type, comprising orthoquartzose, clay-draped sands remains the same in all the study areas. The striking feature of the Upper Woburn Sands is the height of preserved sand-wave bedforms, some reaching 3 to 4 m in thickness. Similar, although generally thinner, sand-wave bedforms are known from the Folkestone Formation, interbedded with clay and silt beds. The lower Woburn Sands is characterized by widespread bioturbation and laminae, indicating a brackish to marginal-marine environment. The Lower Greensand Group is characterized by a sequence of estuarine deposits, including sandy estuarine sands and bentonites.

**TABLE 3.—GAULT-LOWER GREENSAND STRATIGRAPHIC SUCCESSION, ISLE OF WIGHT**

<table>
<thead>
<tr>
<th>STRATIGRAPHIC INTERVAL</th>
<th>ENVIRONMENT and LITHOLOGY</th>
</tr>
</thead>
<tbody>
<tr>
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</tr>
<tr>
<td><strong>LOWER GREENSAND GROUP</strong></td>
<td>Mixed shallow and offshore marine facies: chronostratigraphically condensed conglomerates and sandstones</td>
</tr>
</tbody>
</table>

*stratigraphic terminology from Shephard-Thorn et al. (1986).
with pebbly, bioturbated or argillaceous sands. In the Sandrock Formation on the Isle of Wight, cross-bedded sandstones average 1m or less in thickness. Occasional sand-wave deposits preserved in the open, outer estuary setting may be up to 2 m thick (Dike 1972b; Nio 1976; Wach, 1991). In all areas, the variation in bedform size may be reflected in the predominance of retrogradational vs. progradational parasequence sets, the latter commonly containing large bedforms.

Lithostratigraphic variation between the Woburn Sands, Folkestone, and Sandrock Formations is also evident. In the Woburn succession clay beds comprising amalgamated drapes thicker than a few decimeters are rare. The identification of an intra-Woburn Sands sequence boundary is dependent on observation of parasequence stacking patterns and an incised surface. In the Folkestone Formation, a dark, clay-silt bed up to 10 m thick is documented from the center of the succession along the North Downs. This bed contains marine fossils, including ammonites (Casey, 1961). The bed commonly overlies an erosion surface and may be correlated across the western Weald Basin; it may represent a transgressive surface above a sequence boundary. On the Isle of Wight, the Sandrock Formation has been subdivided into members on the basis of 1- to 10-m—thick dark silt/clay units between white sands. These mudstone intercalations are the key to a sequence stratigraphic interpretation of the succession. They are either bioturbated marine mudstones incised by sand-filled channels, representing base-level fall; or they represent laminated subtidal to intertidal muds, preserved within estuaries, incised during relative sea-level fall, and filled during subsequent rise.

The variable stratigraphic successions outlined above are related to the different environments described from each location. Absence of fine-grained sediment at Leighton Buzzard and their large bedform size (3–4 m: Upper Sands), reflects the strength of tidal currents in this area. Conversely, the abundant clay beds of the Isle of Wight succession, preserved in an estuarine environment (Fig. 3), suggests shoreline embayments...
subject to tidal influence. The variable Folkestone Formation represents deposition in a mixed offshore-estuarine environment somewhere between energetic offshore and relatively quiet estuarine conditions.

The succession preserved in each of these areas relates closely to paleo-environment. Published discussions of the controls on depositional sequences have been concerned with the relative importance of tectonism and eustasy: differential tectonic movements in each of the study areas are quite probable (Eyers, 1991; Ruffell, 1992), yet no over riding tectonic control on sequence development can be demonstrated for the variation in each of these successions. Both the Leighton Buzzard and Isle of Wight successions show a paradox in interpretation: at Leighton Buzzard, water depths were greater than on the Isle of Wight, as demonstrated by the preservation of large bedforms and the predominance of retrogradational sequences. Yet at Leighton Buzzard we see the strongly erosive sequence boundary of the Coarse Red Sands (107.5 Ma of Fig. 6) indicative of exposure and few thick clay beds. On the Isle of Wight, shallow-water depths were maintained on what was probably the "correlative conformity" to the exposure surface (sequence boundary), observed in the Leighton Buzzard succession. This is no doubt a product of accommodation space.

Within the pre-existing biostratigraphic constraints, a preliminary depositional sequence stratigraphic interpretation can be applied to each section. The Sandrock succession contains three sequences, yet the Folkestone Formation/Woburn Sands succession has only two sequences within correlatable sections. The conclusion is that understanding paleo-environmental changes may explain why different sedimentary sequences are preserved in similar lithofacies. Even if the same number of sequences could be picked in this succession, using a range of facies indicators, it still appears that environment played a dominant role in how sedimentary sequences formed and were preserved.

The mature, texturally similar sands in the Aptian-Albian strata of southern England are indicative of this sediment having been through previous cycles of erosion. Such deposits were possibly provenanced from the arenaceous strata of the Triassic and Carboniferous hinterlands to the Wessex Basin. Seasonal braided-fluvial systems probably transported sediment from the hinterlands: the climate is thought to have been arid in the Lower Cretaceous, changing to more humid by the mid-Cretaceous (Weissert, 1989). Aptian-Albian transgressions created an extensive seaway across southeastern England, subject to strong tidal currents associated with the opening of the Atlantic and continued rifting in the North Sea. Such shallow seas would be sites of continual sorting for the mature sands being swept into the Wessex Basin. While the whole North Sea was an area of marine conditions (Rawson and Riley, 1982), the Channel Basin (tectonically related to the North Atlantic) shows more evidence of enclosed marine deposition, away from strong tidal influence. The dominance of tidal reworking may also have led to some sediment starvation in the Leighton Buzzard area, where tidal currents were stronger (Fig. 3) Sedimentation in the Channel Basin may have kept pace with subsidence, precluding the deposition of open marine sediments until the mid-late Albian.

The association of progradational parasequence sets and large bedforms may be a function of increased accommodation space. Where incision is clear at a sequence boundary (picked from definition of a basinward shift of facies), accommodation space is limited and erosion dominates over accumulation. Where initial accommodation space is high, incision may occur sequentially on each 4th-order fall of an already 3rd-order falling curve. This is demonstrated in our interpretation of Aptian-Albian correlative conformity development illustrated in Fig. 8.

ACKNOWLEDGMENTS

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REFERENCES


ABSTRACT: The sequences and the depositional systems tracts of the Berriasian (nine), Valanginian (nine), Hauterivian (seven), Barremian (six, perhaps seven) and lowest Aptian (two) in the Río Argos succession near Caravaca (Southeast Spain) are delimited and their chronostratigraphic position established in terms of ammonite biochronozones. Also their position in relation to zones of planktonic fossil groups (calpionellids, dinoflagellate cysts, nannofossils) is given. Thirty-five sequences and 6 supersequences were delimited. The Río Argos succession is an ammonite-bearing cyclic alternation of coccolith limestone beds and marlstone interbeds. The sequence boundaries are correlated in detail with well-described sections in southeast France, such as the Berriasian type section near Berrias, the Broyon section (Lower Berriasian), the La Charce section (Upper Valanginian, Hauterivian) and the Angles section (Berriasian, Valanginian, type section of the Barremian). This correlation shows that in France the same number of sequence boundaries occur at the same chronostratigraphic levels. This adds faith in the reliability of the number and chronostratigraphic position of the Lower Cretaceous sequences. Special attention is given to the sedimentological interpretation of the position assigned to the sequence boundaries in such cyclic limestone marlstone successions, which is different from the one currently used in France.

INTRODUCTION

The ongoing Río Argos Project is an interdisciplinary research project of the National Museum of Natural History of the Netherlands aiming at the integration of microfossil zonation against the standard ammonite zonation and sequence stratigraphy of the Lower Cretaceous succession along the Río Argos. This succession is situated in the frontal parts of the Subbetic Zone of the Betic Cordilleras and outcrops over a length of 8 km along the meandering Río Argos and its tributaries west of Caravaca, Province of Murcia, southeast Spain (Fig. 1). Twenty-five parallel and partly overlapping sections (indicated with the letters A to Z) were measured whereby each bed was numbered (A-5-225, Z1-278, etc.; Fig. 2). In four instances, the sections do not overlap. However, as no ammonite zones are missing, these observational gaps are small and do not seem to hide vital stratigraphic information.

When the sequence stratigraphy of the Río Argos succession was compared with some sections in southeast France, it appeared, as could be expected, that the same number of sequences were found in similar stratigraphic positions. This convinced us of the reliability of the results. As the sequence-stratigraphic signal in the Río Argos succession is very clear (in general clearer than in southeast France), it has been used for field instruction during courses in sequence stratigraphy.

LITHOLOGY AND FACIES

The entire Río Argos succession consists of a rather monotonous cyclic alternation of olive-grey marly micritic limestone beds and dark-grey, shaly marlstone interbeds. The thickness of the limestone beds varies between 0.5 and 75 cm, those of the marlstone interbeds between 0.5 and 1000 cm. A limestone marlstone couplet has an average thickness of 58 cm. In the limestone beds and marlstone interbeds, the clay fraction is the main detrital fraction. From the middle of the Barremian stratigraphic framework, 2 formations have been distinguished (van Veen, 1969): the Miravetes Formation without turbidites (from bed Z1 to V63) and the Argos Formation with turbidites (from bed V63 upward).

Diagenetic overprinting has been indicated: besides mechanical compaction (±80–90% for the marlstone interbeds and ±40–60% for the limestone beds) differential dissolution and cementation occurred, which gave rise to contrasting bedding enhancement and nodular bedding.

Ten Kate and Sprenger (1989) and Sprenger and Ten Kate (1992) demonstrated that periodicities in the succession of layers in the outcrops along the Río Argos closely match the Milankovitch frequency bands of the precession, obliquity and eccentricity. A limestone marlstone couplet has generally been deposited during 20,000 years.

The entire Lower Cretaceous Río Argos succession has been deposited within a basal pelagic environment. Several paleoecological arguments suggest deposition in an epibathyal pa-
Fig. 2.—Correlation of the sequence boundaries of the Río Argos succession with some important sections in southeast France: the lower Berriasian strata of the Broyon Quarry, the Berriasian stratotype near Berrias, the Berriasian and Valanginian strata and the Barremian stratotype near Angles, and the upper Valanginian and Hauterivian strata near La Charce.
Fig. 2.—Continued.
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leoenvironment varying between 200 and 400 m: nektonic ammonites constitute 99% of the megafossil content. The ammonite fauna is dominated by phylloceratids, lytoceratids, haploceratids and desmoceratids. Such a domination characterizes depths below 200 m (Ziegler, 1967). Besides frequent bioturbation, large Zoophycos feeding burrows are present throughout the succession. This so-called ‘Zoophycos ichnofacies’ is commonly interpreted to indicate depths below the neritic environment (Pemberton et al., 1992). These depths are in accordance with the general scarcity of benthic megafossils and foraminifera.

CHRONOSTRATIGRAPHIC FRAMEWORK

In the IGCP-Project 262, it had been agreed upon that all Cretaceous biozones should be calibrated against ammonite chronozones, because all Mesozoic stages are defined by ammonite chronozones and because ammonites, if frequent, produce the highest stratigraphic resolution (down to 100 ky). Also, depositional sequences, which are lithostratigraphic intervals, can be treated as chronostratigraphic intervals because a sequence represents all rocks deposited within a particular interval of time. Sequence boundaries are not time lines, but time-bounding surfaces that separate older rocks from younger ones. The unconformable sequence boundaries on the shelf are represented by their correlative conformities in pelagic basins. If, as can be expected, 3rd-order depositional sequences prove to be of eustatic origin, then the correlative conformities would be virtually isochronous and would allow precise global correlation. Sequence-stratigraphic boundaries subdivide ammonite chronozones into still finer chronostratigraphic intervals. Magnetostratigraphy could not be applied because of strong magnetic overprinting, possibly caused by the deformation of the Betic Cordilleras.

In the Río Argos succession, the combination of ammonite chronozones and sequence stratigraphy forms a firm and unsurpassedly fine chronostratigraphic framework against which the biozones of calpionellids, dinoflagellate cysts, nannofossils, and foraminifera can be calibrated precisely (Fig. 2). The ammonite zonation used here is the one adopted during the international workshop on the Lower Cretaceous standard ammonite zonation of the Mediterranean faunal province held in Mula (Spain) in July 1992 (Hoedemaeker et al., 1993; Fig. 3). The integrated biozation of the Río Argos succession has been published by Hoedemaeker and Leereveld (1995). It will only be shortly referred to briefly here, because it is important for well-founded correlation to know the exact chronostratigraphic position of boundaries of sequences and systems tracts.

MAIN SEQUENCE-STRATIGRAPHIC SIGNAL

The monotonity of the cyclic alternation is broken by marl-rich bed sets, which alternate with marl-poor bed sets. In the marl-rich bed sets (containing more terrigenous siliciclasts), limestone beds are relatively far apart and, consequently, marlstone interbeds are thick. In the marl-poor (less siliciclastic, but more pelagic) bed sets, limestone beds are relatively closely spaced and the marlstone interbeds are thin. The variation in the thicknesses of the limestone beds is smaller than the variation in thicknesses of the marlstone interbeds. This suggests that changes in the supply of siliciclastic detritus are greater than changes in the rain of Nannococcos tests.

The sequence-stratigraphic signature is mainly manifested by these alternating marl-rich and marl-poor bed sets and by the pattern in the succession of beds. Since the clay fraction is the main terrigenous siliciclastic component in the rocks, only the marlstone interbeds should be considered during sequence-stratigraphic analysis. The pelagic limestone beds should be left out of consideration, because the source of the limestone components is totally unrelated to that of the terrigenous siliciclastic input; they are not detrital.

Roughly speaking, the thickness of the marlstone interbeds gradually diminishes upward from the marl-rich to the marl-poor bed sets, whereas the lithologic change from marl-poor to the marl-rich bed sets is commonly abrupt and attended by a notable jump in the thickness of the marlstone interbeds. The thicknesses of the Milankovitch cycles (i.e., of the limestone-marlstone couplets) are a measure of the degree of starvation in the basin. When the marlstone interbeds are thin, the deposits are more starved, which is also shown by an increased ratio of dinoflagellate cysts to sporomorphs and by the greater frequency of ammonites. Because of the closer spacing of the limestone beds, the frequency of nannofossils also is greater. The less siliciclastic, more calcareous bed sets are interpreted as transgressive to highstand systems tracts in which the most starved part represents the maximum flooding surface.

When the marlstone interbeds are thick, the input of siliciclasts is great and the limestone beds are relatively far apart. The frequency of ammonites and nannofossils is less and also the ratio of dinoflagellate cysts to sporomorphs is notably less than in the more calcareous bed sets. The more siliciclastic bed sets were interpreted as lowstand systems tracts in which the input of terrigenous siliciclasts into the basin is greatest.

The abrupt transition from thin to thick marlstone interbeds at the base of the marl-rich bed sets is interpreted as the sequence boundary. This sequence-stratigraphic interpretation is in contrast to the opinion of French sedimentologists (Jaquin et al., 1991; Ferry, 1989, 1991; Reboulet et al., 1992), who place the sequence boundaries in the pelagic successions of the southeastern French Fosse Vocontienne at the tops of the marl-rich bed sets instead of at their bases. They interpreted these bases as tops of lowstand surfaces or as maximum flooding surfaces. This would be right if the limestones were composed of platform-derived calciclastic detritus; this is commonly not the case in truly pelagic basins. When examining the Río Argos succession in 1991, a group of Dutch stratigraphers/sedimentologists discussed the interpretation used here at length with P. R. Vail and we were convinced that the new interpretation far better accords with the depositional model of sequence stratigraphy: the siliciclastic clay fraction is mainly deposited in the lowstand systems tract, whereas the transgressive and highstand systems tracts are starved and less clayey. The contrasting French interpretation causes a consistent shift in the ages of the sequence boundaries; they are invariably too young. Moreover the correlations of shelf successions with basin successions (from which the ages are derived) are demonstratably out of phase.

Since the flooding of the shelf on top of the lowstand systems tract comes about rather quickly, the transition from relatively thick to relatively thin marlstone interbeds at the top of the marl-rich bed sets also may be rather quick, though certainly
**Fig. 3.**—Calibration of short-term sea-level fluctuations (3rd-order) and long-term sea-level fluctuations against the standard ammonite zonation. This zonation is the one adopted at the International Meeting of the Cephalopod Working Group in Mula in 1992, held within the framework of the IGCP-project 262: Tethyan Cretaceous Correlation.
not as abrupt as at the base of the marl-rich bed sets. The high-
stand systems tracts show a gradual thickening of the marlstone
interbeds just below the jump in thickness at the sequence
boundary.

The smaller input of nutrients during sea-level highstands is
often shown by the limestone beds being thinner, implying
smaller nannoplankton production during each Milankovitch
cycle. During sea-level lowstands, greater input of nutrients of-
ten translates into a greater nannoplankton production and
thicker limestone beds.

The gradual changes in thickness of the marlstone interbeds
were interpreted as prograding, aggrading, or retrograding,
which helps to identify the various systems tracts.

An important consequence of this sequence-stratigraphic in-
terpretation of pelagic successions with respect to chronostra-
igraphic correlation is that marl-rich bed sets on the carbonate
platform generally correspond to limy bed sets (pelagic nann-
ofossil limestones) in the basin. The normal correlation of
marl-rich bed sets on the platform with marl-rich bed sets in
the pelagic basin can be erroneous, at least in settings compa-
rable to the Vocontian Trough and the Río Argos basin. A sa-
lient example is the shallow-marine marls of the basal Hauter-
ivian radiatus zone near Neuchâtel (Switzerland) and the
conspicuous bedded limestone interval of the same zone in the
Vocontian basin (France). However, the distinction between
marl-rich and limy bed sets may not be apparent in some pelagic
basins because of little or no clay influx (e.g., Maiolica For-
mation, Italy). In these circumstances depositional systems
tracts are difficult to discern.

The previously described criteria are, except for a few ad-
ditional features, the only ones by which depositional sequences
in pelagic successions (such as along the Río Argos) can be
delimited. The stratigraphic positions of the sequence-strati-
graphic boundaries are very clear. I consider this as an advan-
tage of pelagic successions over shelf successions, because one
does not have to make difficult sedimentological and facioli-
gorical interpretations to solve the position of the different sys-
tems tracts; the interpretation is straightforward. Moreover,
bathyal pelagic successions are generally more complete.

SEQUENCES IN THE RIO ARGOS SUCCESSION

For the various sequence boundaries, the French code (Ar-
naud-Vanneau and Arnaud, 1990, 1991, Jacquin et al., 1991,
Hunt and Tucker, 1993) is used instead of the absolute age code
used in the cycle chart of Haq et al. (1988). The Río Argos
sequences correlate marvelously with the French ones, whereas
the correlation with the cycle chart of Haq et al. (1988) is not
always clear. The codes BE1–7 are used for the Berriasian,
VA1–6 for the Valanginian, HA1–7 for the Hauterivian and
BA1–5 for the Barremian Stage. However, several new se-
quence boundaries have been indicated that still lack a code.
The presence of the same number of marl-rich bed sets at the
same chronostratigraphic levels in so many different sections
in France and Spain is really striking. It is very convincing that
the same supraregional signals can only be explained by sea-
level fluctuations. Correlations of sequence boundaries between
the Río Argos succession and several important French sections
are given (Fig. 2).

Berriasian Sequences

Within the Berriasian strata of the Río Argos succession, nine
sequence boundaries have been determined. Eight of them also
have been established in southeast France (Jan du Chêne et al.,
1993) at similar stratigraphic levels in the stratotype section
near Berrias, the reference section near Angles, and in the sec-
tion of the Broyon quarry. Confusingly, the biostratigraphic
subdivision used in the columns by Jan du Chêne et al. (1993)
is not in accordance with the published source data. Perhaps
more confusing is that each investigator uses his own bed num-
bering. We follow the bed numbers used by Jan du Chêne et
al. (1993). Correlations of the Río Argos sequences with those
of France are given (Fig. 2).

Sequence Boundary BE1.—

BE1 is at the base of a grainflow bed (Z1) near the lower
boundary of the jacobi subzone (approximate base of calpi-
onellid zone B). The grainflow bed is followed by a marl-rich
bed set and a slumped interval. This sequence boundary cor-
relates with BE1 in the Berrias (base of bed 139/1) and Angles
(base of bed 68) sections and with “TI6?” in the Broyon section
(base of bed 15). According to Le Hégarat (1971), Le Hégarat
and Remane (1973) and Cecca et al. (1989), the latter level
represents the base of calpionellid zone B. The biostratigraph-
y of this level has been misinterpreted by Jan du Chêne et al.,
1993. According to calpionellid zonation (Le Hégarat and
Ferry, 1990) sequence boundary TI5 of the Broyon section cor-
relates with “TI6” of the Angles section.

New Sequence Boundary.—

A new sequence boundary on top of bed Z139 is needed to
explain the sudden thick marlstone interbed at this level and to
explain the widespread extinction wave among the ammonites
at the top of the jacobi subzone (Hoedemaeker, 1981). It can
be correlated with “BE1” at the base of bed 18A in the Broyon
section. This level is, in contrast to the interpretations of Jan
du Chêne et al. (1993), at the top of the jacobi subzone (Cecca
et al., 1989; Le Hégarat and Remane, 1973). This new sequence
boundary has not been found in the Berrias section. It may
correspond to “BE2” at the base of bed 78 in the Angles section.
In the Río Argos succession, the transgressive and highstand
systems tracts of this sequence are strongly enriched with
ammonites.

Sequence Boundary BE2.—

BE2, at the base of the marl-rich bed set on top of bed Z173
in the lower part of the grandis subzone. It correlates with BE2
at the top of bed 23 in the Broyon section and at the base of
bed 143/2 in the Berrias section (both near the base of the gran-
dis subzone; Le Hégarat, 1971; Cecca et al., 1989). It also cor-
relates with “BE3” at the base of bed 88 in the Angles section.

Sequence Boundary BE3.—

BE3, at the base of the marl-rich bed set, on top of bed Z203
at the top of the grandis subzone. It correlates with BE3 at the
base of bed 147/14B of the Berrias section. This implies that
the lower part of bed 147 (numbering of Le Hégarat, 1965a),
from which no ammonites have been reported, belongs to the
grandis subzone instead of the subalpina subzone. It also cor-
BERRIASIAN-BARREMIAN SEQUENCES IN THE RIO ARGOS SUCCESSION NEAR CARAVACA (SOUTHEAST SPAIN)

Sequence Boundary BE4.—

BE4 explains the thick marlstone interbed on top of bed Z241 at the base of the calpionellid zone C in the upper part of the subalpina subzone. It correlates with BE4 at the base of bed 150/22 in the Berrias section. It is interpreted to be at the base of bed 30 in the Broyon section and it is incorporated within the hiatus at the base of bed 88 in the Angles section, where the totality of the occitanica zone is missing.

New Sequence Boundary.—

A new sequence boundary may be drawn at the thick marlstone bed on top of bed Y49 in the lower part of dalmasi subzone. In the Río Argos succession, the frequency of ammonites peaks in the interval from bed Y34 to bed Y49, which may be interpreted as transgressive and highstand systems tracts.

In the Berrias section, bed 155, also in the lower part of the dalmasi subzone, contains many ammonites. Le Hégarat (1971) even speaks of a “lumachelle” of Dalmsiceras dalmasi. The manganese content also peaks in bed 156 (bed 155 not sampled; Emmanuel and Renard, 1993). Bed 156 (bed 155 not sampled) contains a peak in the frequency of dinoflagellate cysts (Monteil, 1993). Bed 155 may therefore be interpreted to represent the maximum flooding surface. In the Berrias section, the new sequence boundary may be situated at the base of bed 161/31, a thick detrital limestone bed exhibiting a minimum in the frequency of dinoflagellate cysts (Monteil, 1993) and in manganese content (Emmanuel and Renard, 1993). In the Río Argos succession, the paramimouna subzone begins at the top lowstand surface at the base of bed Y122 above the first occurrence of Calpionellopsis simplex in bed Y105. The transgressive and highstand systems tracts correspond to the beds with the foraminifer Pavlovecina allobrogensis in southeast France.

Sequence Boundary BE5.—

BE5 is drawn at the base of the prominent marl-rich bed set on top of bed Y136, at about the middle of the paramimouna subzone. It correlates approximately with the base of bed 49 in the Berrias section, which corresponds to a low in the manganese content (Emmanuel and Renard, 1993) and low dinoflagellate cysts frequency (Steffen & Gorin, 1993). This level is just above a peak in the frequency of opaque, highly oxidized, terrestrial higher plant debris (Steffen & Gorin, 1993), which indicates the highest part of the highstand systems tract. This level corresponds with the position advocated by Emmanuel and Renard (1993) for their “BE6” and is just below the base of bed 180, which has been interpreted as the position of BE5 by Jan du Chêne et al. (1993). In the Angles section, BE5, at the base of bed 120, represents a hiatus comprising the upper part of the paramimouna subzone; only the lower part of this subzone is present (beds 92–119). The maximum flooding surface in bed Y180 of the Río Argos succession corresponds to depositional discontinuity Di1 established in southeast France (Darsac, 1983) and to bed 188 in the Berrias section. This level presumably corresponds to the highest point the sea level ever reached during the Berriasian time and contains the maximum diversity in Berriasian ammonite species.

Sequence Boundary BE6.—

BE6 is drawn at the base of the marly bed set on top of bed Y189, approximately in the middle of the picteti subzone. It represents a notable jump in the thickness of the marlstone interbeds. It correlates with BE6 at the base of bed 190 in the Berrias section and at the base of bed 132 in the Angles section.

Sequence Boundary BE7.—

BE7 is drawn on top of bed Y234 at the base of the marl-rich bed set near the base of the alpilensis subzone. It represents a notable jump in the thickness of the marlstone interbeds. It correlates with BE7 on top of bed 162 in the Angles section and with the top of bed 197 (probably 196 of Le Hegarat, 1965a, 1971) of the Berrias section, which represents an important hiatus. The greater part of the deposits corresponding to the alpilensis subzone are missing (Hoedemaeker, 1982, 1983, 1984) as is the entire magnetostratigraphic chron M15n. Bed 198 of the Berrias section yielded the Valanginian ammonites Thurmanciceras thurmanni and T. gratianopolitense (Le Hégaret, 1971; Galbrun et al., 1986) and also contains the boundary between calpionellid subzone D3 with C. oblonga (bed 198a–c) and D3 without C. oblonga (bed 198d–f) (Le Hégaret and Remane, 1968). This calpionellid boundary is located within the highstand systems tract below sequence boundary VA1. This implies that the lowstand systems tract and the greatest part of the transgressive systems tract are missing. Sequence boundary BE7 correlates with the so-called “Late Cimmerian Unconformity,” well-known from subsurface stratigraphy in the North Sea. This is a type 1 sequence boundary (Håg et al., 1988), which in the Mediterranean faunal province is attended by an important turnover in the ammonite fauna (Hoedemaeker, 1982, 1983, 1984, 1995a). The transgressive and highstand systems tracts yielded the first Thurmanciceras otopeta and Tintinnella pertransiens. The maximum flooding surface at the base of bed Y266 corresponds to depositional discontinuity Di2 established in southeast France (Darsac, 1983).

Valanginian Sequences

In the Valanginian of the Río Argos succession, ten sequence boundaries have been indicated. Six of them have a French code. Correlation with the Angles (numbering of Busnardo et al., 1979) and the La Charce sections (numbering of Bulot et al., 1992) in southeast France are given (Fig. 3). In the La Charce section the bed numbers of Reboulet et al. (1992) also are given.
**Sequence Boundary VA1.**

VA1, situated at the base of the marl-rich bed set on top of bed Y267a, is a type 1 sequence boundary (Haq et al., 1988). It represents a notable jump in the thickness of the marlstone interbeds. It corresponds to the top of bed 194 in the Angles section and with the top of bed 200 in the Berrias section. It is just above the boundary between the beds of calpionellid subzone D3 with *C. oblonga* and the beds without *C. oblonga* and it nearly coincides with the appearance of *Remaniella margoeauui*. *Calpionellites darderi* appears in the transgressive systems tract and marks the base of calpionellid zone E (Busnardo et al., 1979).

**New Sequence Boundary.**

A new sequence boundary was interpreted at the top of bed Y267c, which is the base of a marl-rich bed set. It was correlated with the top of limestone bed 223 of the Angles section and is situated in the lower part of the *pertransiens* zone.

**Sequence Boundary VA2.**

VA2, situated at the base of the marl-rich bed set on top of bed Y319 (= M248), represents a notable jump in the thickness of the marlstone interbeds and corresponds to the top of bed 254 in the Angles section. Bed M259 is the maximum flooding surface that corresponds to depositional discontinuity Di3 in southeast France (Darsac, 1983).

**Sequence Boundary VA3.**

VA3, situated at the base of the marl-rich bed set on top of bed M261, corresponds to the top of bed 290 in the Angles section. In the Rio Argos as well as the Angles sections, the shales of the lowstand systems tract fall apart into thin plates. The cause of this similarity over such a great distance is unknown. In the Rio Argos succession very thin (a few millimeters) silty laminae are intercalated and are interpreted as the distinct-most parts of turbidites, whose proximal parts form part of the Los Villares Formation.

**New Sequence Boundary.**

A new sequence boundary was interpreted at the base of the slump interval on top of bed M320 in order to explain the peak in relatively shallow water dinoflagellate cysts which separates two distinct peaks of relatively deep water dinocysts. This sequence boundary can also be recognized at the base of the marl-rich bed set on top of bed 303a in the Angles section. It correlates with the hiatus at the base of the Calcaire Roux turbidites (*campionotoxus* zone; the index species was sampled up to the top) in the La Charce section. The micritic limestone bed 204 (= 100 of Rebloulet and others) on top of these turbidites, already represents the base of the *pronecostatum* horizon (i.e. the middle part of the *verrucosum* zone). The entire *verrucosum* horizon (i.e. lower part of *verrucosum* zone) is missing in La Charce. The *verrucosum* horizon is equivalent to the transgressive and lower highstand systems tracts of this sequence. The top lowstand surface corresponds to the so-called “Discontinuité du Valanginien moyen” of Autran (1989).

**Sequence Boundary VA4.**

VA4 is situated at the base of the marl-rich bed set on top of bed M344. This sequence boundary is situated in the lower part of the *pronecostatum* horizon (middle part of the *verrucosum* zone). It correlates with the top of limestone bed 321 in the Angles section, which marks a notable jump in the thickness of the marlstone interbeds at the base of a marl-rich bed set. It also correlates with bed 206 (103 of Rebloulet et al., 1992) of the La Charce section, which is also at the base of an appreciable jump in the thickness of the marlstone interbeds at the base of a marl-rich bed set. The lowstand systems tract is virtually equivalent to the *pronecostatum* horizon (except for its basalmost part). This sequence boundary is attended by an important turnover in the ammonite fauna (Hoedemaeker, 1995a) not only in the Mediterranean region, but also in the boreal realm (Kemper, 1978; Bartenstein and Bettenstaedt, 1962; Kemper and Wiedenroth, 1987). Therefore it probably represents a type 1 sequence boundary (Hoedemaeker, 1995a). Sequence VA4 and the lowstand systems tract of the next higher sequence constitute the so-called “Grande Lumachelle” of the Carejuan and Allaves sections in southeast France (Thieuloy et al., 1990).

**New Sequence Boundary.**

Rebloulet et al. (1992) described the sequences in the La Charce section. However, they interpreted the bases of the marl-rich bed sets as transgressive surfaces instead of as sequence boundaries as interpreted here. The same sequences were indicated in the Rı´o Argos succession and in the Angles section. Of these four sequence boundaries only the last two bear a French code; the first two do not. The bed-by-bed correlation of Atrops and Rebloulet (1993) between the La Charce and Angles sections only differs in minor details from the correlation preferred here.

In the Rı´o Argos succession, the first new sequence boundary was drawn at the base of the marl-rich bed set in the middle of the unnumbered interval between M344 and N1. In the La Charce section, this new sequence boundary is situated at the base of the marl-rich bed set on top of bed 213c (124 of Rebloulet et al., 1992) in the middle of the *periginus* horizon (upper part of the *verrucosum* zone). In the Angles section it is situated on top of limestone bed 338a. The “Discontinuité du Valanginien supérieur” (DVS) of Autran (1989) corresponds to the top lowstand surface close to the base of the *nicklesi* horizon (base of *trinodosum* subzone of the *pachydicranus* zone) in the La Charce section (bed 217).

**New Sequence Boundary.**

The second new sequence boundary is situated at the base of the marl-rich bed set on top of bed N1. It correlates with the top of bed 221 (= 141 of Rebloulet et al. 1992) in the La Charce section in the lower part of the *nicklesi* horizon (= lower part of the *trinodosum* subzone), and with the top of limestone bed 348 in the Angles section.

**Sequence Boundary VA5.**

VA5 is the code of the sequence boundary in the basal part of the *furcillata* horizon (= upper part of the *trinodosum* subzone). In the Rı´o Argos succession, it is situated on top of bed N13. It correlates with the top of bed 232 (166 of Rebloulet et al., 1992) in the La Charce section and with the top of limestone bed 361 in the Angles section.
Sequence Boundary VA6.—

VA6 is situated in the observational gap (fault) between beds N40 and N41. In the La Charce section, it is situated on top of bed 238 (176 of Reboulet et al., 1992) in the middle of the *furcillata* horizon (upper part of the *trinodosum* subzone). In the Angles section it is situated on top of limestone bed 372.

Hauterivian Sequences

Seven sequence boundaries have been indicated in the Hauterivian Stage. The Hauterivian sequence in the sections near Chamaloc and Vergons in France shows a remarkable resemblance with the Hauterivian sequence in the La Charce and Río Argos sections (Fig. 3); both contain the same number of sequence boundaries at the same chronostratigraphic levels.

Sequence Boundary HA1.—

HA1 was established on top of bed P12 at the top of the *radiatus* zone and at the base of a marl-rich bed set. It correlates with the top of bed 270b (214 of Reboulet et al., 1992) at the base of a slumped interval in the La Charce section. The lowstand surface practically coincides with the base of the *jeannoti* subzone (the *loryi* subzone is confined to the lowstand systems tract), and the maximum flooding surface corresponds to the “Discontinuité de la Zone à Loryi” (DZL) of Autran (1989).

Sequence Boundary HA2.—

HA2 is situated at the base of the marl-rich bed set on top of bed P31 near the middle of the *loryi* zone, i.e. in the lower part of the *jeannoti* subzone. It correlates with bed 291 in the La Charce section.

New Sequence Boundary

For sake of completeness we have to mention that in the section of Vergons in southeast France, a sequence boundary is present at the base of slumped bed 53, close to the base of the *nodosoplicatum* zone. In the Río Argos succession, this sequence boundary should be situated in the observational gap between the *loryi* and *nodosoplicatum* zones. In a section parallel to section A of the Río Argos succession, there is a marl-rich bed set at the base of the *nodosoplicatum* zone, but the sequence boundary is not exposed. Its position in the La Charce section is not clear.

Sequence Boundary HA3.—

HA3 is situated at the base of the marl-rich bed set and on top of bed A25, near the middle of the *nodosoplicatum* zone. It correlates with the top of bed 307b in the La Charce section. This sequence boundary might be of type I on account of the profound, though relatively gradual, change in the ammonite fauna (Hoedemaeker, 1995a).

Sequence Boundary HA4.—

HA4 is situated at the base of the marl-rich bed set on top of bed A57, near the middle of the *crusensis* horizon (basal part of *sayni* zone). It correlates with the top of bed 327 in the La Charce section.

Sequence Boundary HA5.—

HA5 is situated at the base of the marl-rich bed set on top of bed A78 approximately at the top of the *sayni* zone. It correlates with the top of bed 358 in the La Charce section.

Sequence Boundary HA6.—

HA6 is situated at the base of the marl-rich bed set on top of bed A135 in the middle of the *balearis* zone. It correlates with the top of bed 26 (numbering of Busnardo, 1965) in the Angles section. The base of the *angulicostata auctorum* zone coincides with the top lowstand surface. The valid name for *Pseudothurmannia angulicostata auctorum* is *P. ohmi* (Winkler); the zone should be rebaptized accordingly.

Sequence Boundary HA7.—

HA7 (or BA0) was drawn on top of bed A154 at the base of the marl-rich bed set near the base of the *catulloi* subzone approximately in the middle of the former *angulicostata auctorum* zone. It correlates with the top of limestone bed 64 (numbering of Busnardo and Vermeulen, 1986) in the Angles section. This sequence boundary is attended by an important turnover in the ammonite faunas in the Mediterranean faunal province (Hoedemaeker, 1995a,b). It is a type I sequence boundary on account of the development of large basin floor fans in the departments of Drôme and Hautes-Alpes (France) (Arnault-Vanneau and Arnaud, 1990). The base of the *hugii* zone nearly coincides with the top lowstand surface of this sequence.

Barremian Sequences

Seven sequence boundaries have been recognized in the Barremian Stage of which five have a French code. Correlations with the type section near Angles are given (Fig. 3). In southeast France the sequence boundaries within this stage have been well established in platform facies (Jacquin et al., 1991; Hunt and Tucker, 1993). In the Barremian type section in the adjacent basin, however, they have been indicated according to the French interpretation (i.e., at the tops of marl-rich bed sets instead of at their bases which is our interpretation; condensed marl-rich bed sets on the platform correspond to condensed limy bed sets (pelagic nannofossil limestones) in the basin. This causes differences in age assignment of the sequence boundaries, because their ages are established in the basin. Moreover, the exact chronostratigraphic position of several sequence boundaries in the Angles section is uncertain because of the absence of a good sea-level fluctuation signal. The sequences in the Río Argos succession may help to solve the uncertainties. Correlation with the Angles section in southeast France is given; the bed numbering of Busnardo (1965) is used.

Sequence Boundary BA1.—

BA1 is situated on top of bed A195 at the base of the marl-rich bed set near the base of the *nicklesi* zone. This level correlates with the top of limestone bed 101 in the Angles section. This bed corresponds to a peak in the ratio between sporo-morphs and dinoflagellate cysts (Wilpshaar, pers. commun., 1990).

New Sequence Boundary.—

A possibly new sequence boundary may be interpreted at the top of bed V28 (of section V1), which is the base of a slumped
interval on top of condensed deposits (thin limestone beds and thin marlstone interbeds) approximately in the middle of the *caillaudianus* zone. It may be a pronounced parasequence. It may correlate with the top of limestone bed 111c of the Angles section (bed number of Vermeulen, pers. commun., 1990).

**Sequence Boundary BA2.**—

BA2 is situated either at the top of bed Q69 or in the observational gap directly below that bed, which forms the base of a marl-rich bed set in the upper part of the *caillaudianus* zone. The lowstand systems tract contains olistostromes with olitholiths as large as 50 m and with allochthonous olitholiths (conorted blocks of Barremian sandstones from the shelf/slope environment of the internal Prebetic zone of the Betic Cordilleras). This means severe destruction of the shelf/slope region and the protrusion of the slope fan far into the basin. It may be interpreted as a type 1 sequence boundary. Also, Hunt and Tucker (1993) reported that this sequence boundary has type 1 characteristics in the Vercors Mountains (southeast France). A possible basin-floor fan is the Caprés Sandstone in the Sierra del Corque (southeast Spain), whose base is dated Caillaudianus chron (Company et al., 1992). The correlation with the Barremian type section near Angles is not clear, because of the absence of a clear unambiguous signal for sea-level fluctuations in the middle part of the type section and because of the scarcity of diagnostic ammonites in that part. The sequence boundary may be situated at the base of bed 137; the marlstone interbed below this bed corresponds to a peak in the ratio of sporomorphs to dinoflagellate cysts (Wilpshaar, pers. commun., 1990). The entire sequence BA2 (between the approximate top of the *moutoni* horizon and the approximate appearance of the dinocyst *Odontochitina operculata* in bed 142) seems to be missing. This absence may be explained by assuming a hiatus in the Angles section at the base of bed 137. Sequence boundaries BA2 and BA3 may be incorporated in this hiatus.

**Sequence Boundary BA3.**—

BA3 is situated on top of bed Q104 (in section Q2) at the base of a marl-rich bed set with slumped intervals, situated in the lower part of the *vandenheckii* zone. It is just below the first appearance of the zonal index *Odontochitina operculata* in bed Q108 in the lower part of the *vandenheckii* zone. This species also appears in bed 142 (numbering of Busnardo, 1965) just above bed 137 of the Angles section (Wilpshaar, pers. commun., 1990). Also BA3 may therefore be incorporated in the hiatus at the base of bed 137 as well as the base of the *vandenheckii* zone.

**New Sequence Boundary.**—

A new sequence boundary is interpreted at the base of the marl-rich bed set on top of bed V40 (of section V2) in the upper part of the *vandenheckii* zone. This conspicuous marl-rich bed set in the Río Argos succession cannot be unambiguously correlated with the Angles section. Candidates could be the beds 144 or 151. Of these two, only the marlstone of bed 144 shows a small peak in the ratio between sporomorphs and dinoflagellate cysts (Wilpshaar, pers. commun., 1990). The transgressive systems tract of this sequence is equivalent to the upper part of the *sartousiana* zone, the beds around the maximum flooding surface correspond to the *Camereiceras limentinum* level in southeast France, the highstand systems tract (beds V53–56 of the Río Argos succession = 162–169 of the Angles section) is equivalent to the *feraudianus* zone.

**Sequence Boundary BA4.**—

BA4 is situated at the base of the marl-rich bed set on top of bed V56 (in section V2) at the base of the *giraudi* zone. It correlates with the top of limestone bed 169 in the Angles section. In the Angles section, there is a peak in the ratio between sporomorphs and dinoflagellate cysts in these levels (Wilpshaar, pers. commun., 1990). The lowstand systems tract of this sequence represents the so-called “Marnes à Heteroceras” in the basin, which is not equivalent to the *Camereiceras limentinum* beds on the platform. In the Río Argos succession, the presence of many distal sandy turbidites begins in the upper part of the lowstand systems tract. The turbidites characterize the Argos Formation, which begins with the first turbidite in bed V63. The turbidites form part of an extensive slope fan.

**Sequence Boundary BA5.**—

BA5 is situated on top of bed V107 at the base of a marl-rich bed set in the upper part of the *sarasinii* zone. It correlates with the top of limestone bed 194 in the Angles section. The top lowstand surface (Angles base of bed 197) is currently considered to be the base of the Aptian Stage. In the Río Argos section, the transgressive and highstand systems tracts yielded Aptian ammonites of the *weissi* zone. Bed 200 of the Angles section yielded the first deshayesitid ammonite specimen (M. Kakabadze, pers. commun., 1995).

**Lower Aptian Sequences**

Only the two lowest Aptian sequence boundaries are discussed here, because the younger beds are tectonically disturbed. The correlations with the Angles section are given (bed numbering of Busnardo, 1965; Fig. 3).

**Sequence Boundary AP1.**—

AP1 is situated at the base of a marl-rich bed set and on top of bed U23, whose top is iron encrusted. It correlates with the top of limestone bed 209 (numbering of Busnardo, 1965) in the Angles section. This level is close to the boundary between the *weissi* and *dehayesi* zones. The transgressive systems tract of this sequence yielded ammonites characteristic of the *dehayesi* zone and correlates with the “Couches supérieur à Orbitolines” in southeast France and with the Selli level in Italy.

**Sequence Boundary AP2.**—

AP2 is situated on top of bed U32, at the base of a marl-rich bed set. It correlates with the top of bed 219 of the Angles section within the *dehayesi* zone and is a type 1 sequence boundary because of the deeply incised valleys on the time-equivalent top of the Urgonian Limestone Formation in southeast France. It coincides practically with the entry of the planktonic foraminifer *Shackoina cabri*.

**SUPERSEQUENCES**

In the Río Argos succession, supersequences were identified with help of ammonite diversity (Hoedemaeker, 1995a; Fig. 3).
It appears that at certain levels, which can be discerned in the Berriasian to Aptian stages all over the Mediterranean region, the ammonite diversity dropped conspicuously; for a short time, extinction is predominated (crisis). After 60–80% of the biotopes had emptied (diversity minimum), about 50–60% of the empty biotopes became quickly filled again with new taxa (faunal renewal). This rapid origination rate marks the beginning of a new biozone, the faunal content of which differs conspicuously from that of the previous zone. The minimum level of the diversity at the end of the crisis is the level of the faunal turnover.

These faunal turnovers are explained as the result of severe eustatic sea-level falls, because they occur at levels that are elsewhere commonly marked by well-known hiatuses and because they virtually coincide with sequence boundaries. The conspicuous disappearance of ammonite taxa are explained by assuming that, as a consequence of such high-amplitude sea-level fall, many biotopes of ammonite species were pushed over the shelf edge and severely telescoped. This reduction of biotopes would have enhanced selection pressure and ultimately led to extinction.

Only 7 of the thirty-six sequence boundaries are attended by such faunal renewals; they are separated by an average of 4 to 5 sequence boundaries without faunal extinction or renewal greater than normal background extinction or renewal. These severe sea-level falls generate type 1 or possible type 1 sequence boundaries and are interpreted to result from long-term eustatic sea-level fluctuations, which are of larger amplitude and longer duration than those of 3rd order. The average lapse of time between these 7 high-amplitude sea-level falls is in the order of 3 my. They are quite similar to the sea-level fluctuations recognized by Hallam (1988) in the Jurassic System.

Sequence boundaries BE3, BE7, VA4, HA3, HA7, BA2 and AP2 are related to the high-amplitude sea-level falls attended by important faunal renewals (Fig. 3). They occur near the base of the subalpina subzone, near the base of the alpilensis subzone, near the base of the pronecostatum horizon (in the middle of the verrucosum zone), in the middle of the nodosoplicatum zone, near the base of the alpilensis subzone, in the middle of the catullosi subzone, in the middle of the caillaudianus zone and near the top of the deshayesi zone respectively (Hoedemaeker, 1995a). Halfway between these falls relatively strong sea-level rises occur, which mark important transgressions in Europe. Transgressive periods were around the base of the lower Berriasian jacobii zone, around the base of the upper Berriasian paramimouna subzone, in the lower part of the lower Valanginian campylotexus zone, around the base of the lower Hauterivian radiatus zone, in the middle of the Hauterivian sayini zone, around the base of the lower Barremian caillaudianus zone, and around the base of the upper Barremian feraudianus zone respectively (Hoedemaeker, 1995a). The highest sea-level stands are reached in the upper part of the jacobii subzone, in the highest paramimouna subzone and lower picteti subzone, in the verrucosum horizon (= basal part of verrucosum zone), in the lower nodosoplicatum zone, in the highest part of the balearis zone and lower ohmi subzone, in the lower part of the caillaudianus zone, and in the deshayesi zone (Hoedemaeker, 1995a).

These supersequences are the result of the same high-amplitude sea-level fluctuations as those that have been called “variations relatives du niveau de la mer” by Arnaud-Vanneaux and Arnaud (1991).

CONCLUSIONS

Thirty-five sequences and 6 supersequences from the Berriasian to lower Aptian strata of the Río Argos succession have been delimited by thirty-six sequence boundaries and 7 supersequence boundaries. It appears that the same number of sequences and supersequences have been or (as to the newly interpreted ones) could be indicated in southeast France at the same chronostratigraphic levels, which is a measure of the reliability of the number and ages of the sequences and their depositional systems tracts. The positions of the sequence boundaries in the French sections in Figure 2 are in accordance with the new sequence-stratigraphic interpretation of pelagic cyclic marlstone-limestone alternations with minor siliciclastic input that has been advocated by Vail and several Dutch stratigraphers/sedimentologists during the field excursion in Spain in May 1991. This interpretation is published here for the first time.

ERRATA

In the long lapse of time between the submission of the manuscript and the proofreading, some new data were published, which should be incorporated in this manuscript. Firstly, on the basis of the investigations of Delany (1995) I was compelled to change the biostratigraphy of sequence AP1, which does not belong to the deshayesi zone but still to the weissi zone which extends up to sequence boundary AP2. This should be corrected in Figs. 2 (8) and 3. Secondly I propose to draw a new sequence boundary on top of limestone bed V82 (section V2, upper Barremian of the Río Argos succession). This level correlates with the top of limestone bed 177 in the lowest part of the sarasini zone in the type section of the Barremian near Angles (S.E. France) (Fig. 2 (8)). Finally, I changed my interpretation of the positions of the sequence boundaries Ba2 and Ba3 in the Barremian stratotype section near Angles. On the basis of peaks in the sporomorph/dinoflagellate cyst ratio Ba2 should be drawn at the top of limestone bed 129, and Ba3 on top of limestone bed 135 (Fig. 2 (7)). There is no hiatus, only condensation of the sequences Ba2 and Ba3.


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PART IV
MESOZOIC ERA
JURASSIC PERIOD
THE NORTH SEA CYCLE: AN OVERVIEW OF 2ND-ORDER TRANSGRESSIVE/REGRESSIVE FACIES CYCLES IN WESTERN EUROPE

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ABSTRACT: Four 2nd-order transgressive/regressive (T/R) facies cycles, including about forty 3rd-order depositional sequences, have been identified within western European Basins. They are: T/R 7 (Bajocian-middle Bathonian), T/R 8 (Upper Bathonian-middle Oxfordian), T/R 9 (Upper Oxfordian-Lower Volgian) and T/R 10 (Ryazanian). These long-term cycles are surprisingly synchronous over widespread areas, even though extensional tectonics were particularly active at that time. Only two regressive pulses (lower-middle Callovian and lower Kimmeridgian) are locally reported as a direct consequence of synrift block tilting. The 3rd-order depositional sequences, which are the building blocks of these 2nd-order T/R cycles, have been calibrated and defined from the North Sea to the Tethyan margin. This allows precise correlation between different tectonic settings and depositional environments. Major erosional unconformities relate to critical events of North Sea rifting. However, they are recorded as major 3rd-order sequence boundaries on surrounding shelves, even far away from the North Sea, such as on the Tethyan margin in the French sub-Alps. Most of the forty defined 3rd-order sequences can be objectively correlated within the limits of the available biostratigraphy. They represent a very efficient tool to constrain the timing of the tectonic control, to quantify the tectonic subsidence and the sediment supply and to predict the development of stratigraphic features.

INTRODUCTION

A comprehensive sequence stratigraphic study of the Jurassic layers in the northern North Sea in the Paris Basin and in the Dauphiné Basin (southeastern France) has been performed to build the hierarchy of stratigraphic cycles and to better understand its relationship with the tectonic evolution of these Basins. The data base consists of more that 450 interpreted wire-line logs in the Paris Basin, 250 wire-line logs in the northern North Sea and numerous well-dated core sections from the North Sea and outcrop sections from the Paris Basin and southeastern France Dauphiné Basin. Additional documentation (Fig. 1) comes from the Mid-Norway Basins (Halten Banken), Moray Firth Basin (southwestern North Sea: Stephen and Davis, this volume), Yorkshire and Dorset (Coe, 1992; 1995), Swiss Jura Mountains (Gygi et al., this volume), Aquitaine Basin (Hantzpergue, 1985, 1991, 1993) and central Lusitanian Basin (Leinfelder and Wilson, this volume). The present paper compares the stratigraphic signature of these different Basins (Fig. 2). We have emphasized the 2nd-order cycles of the Paris Basin because of its location between the Tethyan Margin to the south and the Atlantic Margin to the north. In addition, the Paris Basin was, at the time of the North Sea cycle, the boundary between the carbonate and siliciclastic facies belts. A forthcoming paper (Jacquin et al., in prep.) will document the northern North Sea sequence stratigraphic framework, thus we do not present here any precise documentation for that area.

The time interval corresponding to the North Sea major cycle covers the Middle Jurassic (Dogger), the Late Jurassic (Malm) and the earliest part of the Early Cretaceous epochs (Ryazanian or Berriasian). This long span of time is particularly well controlled by biostratigraphic tools in both the Tethyan realm from ammonites and in the Boreal realm from dinoflagellates.

The long-term North Sea cycle coincided with the entire evolution of the Late Jurassic North Sea rift. It started at the mid-

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Fig. 2.—Compared stratigraphic signature of western European basins for Middle and Upper Jurassic times.
AN OVERVIEW OF 2ND-ORDER TRANSGRESSIVE/REGRESSIVE FACIES CYCLES IN WESTERN EUROPE

RANCE IN PROVENCE PLATFORM

ZONES STAGES

TETHYAN AMMONITES CHRONOZONES

SECOND ORDER

THIRD ORDER

144.2 ± 2.6
159.4 ± 3.6

BATHONIAN
CallistoPicteti
Paramimounum
Dalmasi
Privasensis
Subalpina
Jacobi
Alpillensis

AALENIAN
Murchisonae
Opalinum
Concavum
Bradfordensis

BAJOCIAN
Upper
Parkinsoni
Garantiana
Niortense
= Subfurcatum
Humphrie
sianum
Sauzei
Laeviuscula
Discites

Upper
Middle
Lower
Discus
Zigzag
Progracilis
Subcontractus
Bremeri
Morrisi

Upper
Middle
Lower
Retrocostatum
Bullatus/Macrocephal.

Durangites
Microcanthum

TITHONIAN
Lower
Upper
Middle
KIMMERIDGIAN
Lower
Upper
Middle
OXFORDIAN
Upper
Middle
Lower
BERRIASIAN
Lower
Upper
Middle

137.0 ± 2.2
154.1 ± 3.2
150.7 ± 3.0

Be8 : 137.4
Be6 : 138.6
Be7 : 138.0
Be4 : 141.0
Be3 : 141.8
Be2 ? : 143.0
Be1 : 144.2
Ti4 : 145.8
Ti3 : 146.5
Ti2 ? : 147.7
Ti1 : 150.0
Ki4 : 152.0
Ox8 : 154.6
Ki3 : 152.7
Ox7 : 155.2
Ox6 : 155.8
Ox5 : 157.0
Ox4 : 157.4
Ox2 : 158.5
Ox1 : 159.0
Ca5 : 159.9
Ca4 : 161.0
Ca3 : 161.5
Ca2 : 162.8
Ca1 : 163.7
Ca0 : 164.4
Bt4 : 165.7
Bj5 : 169.7
Bj4 : 171.7
Bj3 : 173.2
Bj2 : 174.6

R6
Aa1 : 179.5
Aa2 : 178.1
Bj1 : 176.5

Fig. 2.—Continued.
overall regressive phase, most of the western European Basins experienced a transpressional stress regime associated with the activation of the North Atlantic Cretaceous rifting. It ended with the so-called “Base Cretaceous unconformity”, which belongs to the Late Cimmerian phase of deformation (Stille, 1924) that recorded the onset of the North Atlantic rifting.

SEQUENCE STRATIGRAPHIC FRAMEWORK

Four 2nd-order transgressive/regressive facies cycles, including about forty 3rd-order depositional sequences, have been identified. These 2nd-order cycles are labelled 7 to 10, following the 6 already defined 2nd-order T/R cycles of the eastern Tethys and Ligurian major 1st-order T/R cycles (Gianolla et al., this volume; Graciansky et al., this volume).

2nd-Order Cycle 7: Upper Aalenian, Bajocian, Lower and Middle Bathonian Substages

Age of Key Surfaces.—

The basal boundary is dated as Early/Late Aalenian boundary (Ludwigia murchisonae-Graphoceras concavum ammonite zonal boundary; 177 Ma). It coincides with the mid-Cimmerian unconformity (Stille, 1924; see discussion in Jacquin and Graciansky, this volume). The upper boundary is dated as Middle/Late Bathonian boundary (Morrisiceras morrisii/Procerites hodsoni ammonite zonal boundary; 167 Ma. The duration is 10 MY. The peak transgression is dated as Parkinsonia acris Subzone of Parkinsonia parkinsoni Zone (Late Bajocian) or Morphoceras macrescens/Parkinsonia convergens subzonal boundary of the Zigzagiceras zigzag Zone (Early Bathonian) depending on the area.

The initial definition of 3rd-order sequences has been given by Floquet et al., 1992; Graciansky et al., 1994; Partington et al., 1993 a, b; Ponsot, 1994; Rioult et al., 1991.

Transgressive Phase.—

The transgressive phase generally comprises two parts: Lower Bajocian layers are mostly aggradational and the Upper Bajocian layers are retrogradational. These features are independent of the nature of sediments, whether they are carbonates or siliciclastics. On structural highs inherited from the previous mid-Cimmerian (Stille, 1924) structures, the aggradational lower part may be missing. This partitioning was controlled by a major tectonic pulse (Tarbert unconformity, Fig. 2), that corresponds to the onset of the block tilting and rifting in the North Sea. Outside the North Sea area, this event is recorded as a major 3rd-order sequence boundary (Bj3 sequence boundary, Stephanoceras humphriesianum Zone), with widespread exposure of shelfal areas and the development of large lowstand deposits in the more subsiding parts (Teloceras bladgeni-T. banksi ammonite Subzones).

Northern North Sea.—In the Brent delta province, the aggradational part of the transgressive phase corresponds to stacked alluvial plain deposits (Ness Formation) that grade seaward into estuarine sand bodies. The thickness variations within that interval may be rapid following lateral changes in subsidence (Fig. 3; North Viking graben versus Huldra Terrace for example). The sediments of the main part of the Ness Formation, were not deposited during the advance of the Brent delta system, but during its retreat, which means they are not time-equivalent to the sand-prone prograding Ranoch-Etive delta front (Fig. 4). These aggradational alluvial plain deposits are bracketed between two mappable and easily recognizable markers on well logs and seismic lines: the one at the base is the maximum flooding surface of the Aa2 sequence (Mid-Ness shale of Mitchell et al. (1992) and the one at the top is the Tarbert unconformity that is overlain by marine Tarbert sandstones.

The backstepping part of the transgressive phase corresponds to the rapid retreat and drowning of the Brent delta system (Fig. 3). It is characterized by nearshore to shelfal sand bodies onlapping the aggradational alluvial plain deposits of the underlying Ness Formation. The increasing upward marine influence is recorded by increasing abundance of marine palynomorphs (Williams, 1992), in particular the base acme biozone of Distilosidinium willei. This acme is calibrated to outcrop sections in East Greenland and to cores from the United Kingdom North Sea, both of which dated by macrofossil of the Parkinsonia parkinsoni ammonite Zone (Upper Bajocian; Callomon, 1975; Morton, 1993). Similarly the J32 maximum flooding surface of Partington et al. (1993 a, b), which is characterized by a dino-flagellate association of the Parkinsonia parkinsoni ammonite Zone, coincides with the drowning surface at the top of the Tarbert Formation. This points to a Late Bajocian age for the Tarbert backstepping phase and not Bathonian or even Callovian as usually proposed (Mitchener, 1992; Helland-Hansen et al., 1992; Whitaker et al., 1992).

Mid-Norway.—The aggradational part of the transgressive phase is represented by the Fangts group (Fig. 2). It is exactly time-equivalent to the aggradational phase of the Ness Formation of the Brent delta system. Shallow-marine to coastal/deltaic facies dominate in the lower part (Ile Formation), whereas open-marine shaly environments develop at the top (Not Formation; Volslet and Doré, 1984; Dalland et al., 1988). The Not Formation is interpreted as an abandonment phase of the deltaic progradation as a consequence of a major Bj2 3rd-order maximum flooding of the Stephanoceras humphriesianum Zone (Fyfe and Cecchi, 1994). A similar event, but within more restricted environments, is also recorded within the Ness Formation of the Brent delta province.
The coarse-grained shoreface sandstones of the Garn Formation, which follows the Not shales, are time-equivalent to the Tarbert sandstones (Fig. 2). They represent the backstepping phase of the overall transgression. The unconformity at the base of the Garn reflects extensional block-faulting similarly to the Tarbert unconformity in the northern North Sea (Fyfe and Cecchi, 1994). As in the northern North Sea, the drowning is of Lowermost Bathonian age. Within both the Tarbert and the Garn, two or three depositional 3rd-order sequences may be recorded (Bj3 to Bj5), each of them backstepping and displaying thick lowstand prograding complexes.

Western Paris Basin.—The Normandy Coast section (Fig. 5, 6) in France (Rioult et al., 1991) and the Dorset Coast section (Fig. 7) in southern United Kingdom (Callomon and Chandler, 1990) are similar and record submarine swell environmental conditions. The environment was dominated by carbonates. The sedimentation rate was very low. The transgressive part may be very thin (even partly missing) unconformably overlying by places the thin prograding sequences of the preceding regressive phase. The upper part is starved (Oolite Ferrugineuse de Bayeux in Normandy; Astarte obliqua bed in Dorset) and unconformably overlies the lower deposits. This important hiatus between the two sequence sets relates to the Tarbet unconformity of the North Sea (which in the Dorset section has been called Vesulian unconformity between the Astarte obliqua beds of the Parkinsonia acris Zone and the “Red Conglomerate” of the Stephanoceras humphriesianum Zone).

Center and Eastern Paris Basin.—The Paris Basin shows a similar evolution, with aggradational shelfal limestones (“Calcaire à Entroques” Formation in Burgundy, Fig. 2) sitting unconformably on Aalenian to Toarcian shales (Fig. 5, 8). Near the top of the aggradational Calcaire à Entroques, the maximum flooding surface of the sequence Bj2 is well recorded throughout the Paris Basin. It brings open-marine shaly facies and associated ammonites (Stephanoceras humphriesianum Zone) onto the carbonate platform (Fig. 5, 9). These shales are coeval to the Not shales of the mid-Norway Basins. The Upper Bajocian retrogradational interval is modulated by successive 3rd-order exposures, which led to the development of thick lowstand deposits (named respectively Teloceras blagdeni, Garantiana garantiana and Parkinsonia bomfordi from the ammonites Subzones) in more subsiding areas (Graciansky et al.,...
This is indicative of high-amplitude 3rd-order sea-level oscillations during the backstepping phase. The tectonic event equivalent to the Tarbert unconformity relates here to major exposure surfaces (Sb Bj3, Bj4), which may cause the Strenoceras subfuscum ammonite Zone to be missing on structural highs (Durlet and Loreau, 1995).

The southern Jura Mountains on the eastern side of the Paris Basin yield spectacular seismic-scale outcrops showing the Bajocian transgressive phase (Ferry and Mangold., 1994). The setting corresponds to the outer shelf part of the carbonate platform that covered the Paris Basin and to the northern margin of the subalpine Dauphiné Basin (southern end member on Fig. 5, 8). As a consequence of the distality of the section, the five Bajocian sequences are well expressed (Fig. 5, 8). In spite of the complexity of the extensional structural setting, the 3rd-order depositional sequences is still preserved, with (1) Lower Bajocian transgressive deposits following Aalenian regressive bioclastic limestones, (2) a major flooding dated as Stephanoceras humphriesianum zone and (3) three major Upper Bajocian 3rd-order lowstand deposits superimposed on second-order overall backstepping.

Subalpine Dauphiné Basin.—In the most subsiding parts of the Subalpine Dauphiné domain, such as the Digne Basin in southeastern France, the subsidence and sediment supply allowed the deposition of several hundred meters of hemipelagic sediments comprising alternating calcareous mudstones and marls. These have been dated with abundant ammonites at the subzone scale, or even at the horizon level (Pavia, 1971). The lower aggrading part is dated from the Hyperlioceras discites through Stephanoceras humphriesianum ammonite Zones of the Lower Bajocian stage and shows a major development of lowstand prograding complexes with respect to highstand deposits. This aggrading part does not extend over the adjacent Provence carbonate platform. The upper part typically is retrograding, as shown by the incursion of middle to late Bajocian ammonite deposits resting in places on lowermost Jurassic strata of the Provence platform. The transition from the aggrading to the backstepping sequences is dated as Early/Late Bajocian boundary. It corresponds to a major phase of block tilting and extension in the subalpine domain (passive margin of the Ligurian Tethys), including the eastern border of the Massif Central (Coadou and Beaudoin, 1975; Elmi, 1984).
at the lowestmost surface of the Fullers Earth (Fig. 11). It is the maximum flooding surface of sequence Bj5. In Normandy, on the other side of the Channel, the peak transgression is within the Couches de passage at the base of the Marnes de Port (Fig. 5, 6; Rioul et al., 1991). It coincides with the same maximum flooding surface (Bj5, Zigzagiceras zigzag Zone, Morphoceras macrescens Subzone).

Center and Eastern Paris Basin.—Within most of the Paris Basin, the peak transgression is within the Ostrea acuminata marls (Figs. 2, 5, 9) and dated in the Parkinsonia acris Subzone (mfs Bj4), where the Bajocian carbonate platform developed (Floquet et al., 1989; Jacquin et al. 1992). On the eastern margin of that carbonate platform, such as in southern Germany (Dietl, 1977), the peak transgression follows the Parkinsonia oolith at the base of the Dentalien Top and is dated of the Morphotrenceras macrescens Subzone (mfs Bj5).

Subalpine Dauphiné Basin.—In the subalpine domain, as in the adjacent surrounding platforms such as in the Jura to the north (Ferry and Mangold, 1994) and the Provence to the south (Dardeau, 1983), the peak transgression is dated in the Zigzagiceras zigzag Zone, Morphoceras macrescens Subzone (MFS Bj5; Graciansky et al., 1994).

In summary, the age of the peak transgression is dated as being close to the Bajocian-Bathonian boundary. It generally coincides with the maximum flooding surface of the 3rd-order depositional sequence Bj5 (Morphoceras macrescens Subzone of the Zigzagiceras zigzag Zone). But local features, such as the combined effects of low sediment supply and increasing tectonic subsidence, may force that peak transgression to be younger.

Regressive Phase.—

Northern North Sea.—The regressive phase is defined by the lower part of the widespread tongue of Heather Shales that covers the Brent delta deposits (Fig. 2). Landward, such as on the Horda platform, a series of basinward stepping shelfal sandstones (Krossfjord Formation, Vollset and Doré, 1984), Early to Middle Bathonian in age, prograde westward out into the Heather Shales (Steel, 1993). These sandstones have a limited extension due to synrift block faulting and are frequently truncated by a mid-Bathonian unconformity. Commonly the Bathonian Heather Shales, time-equivalent seaward to the regressive Krossfjord Sandstones, are missing from the Tampen Spur wells due to deep truncation by a mid-Bathonian unconformity and subsequent Upper Jurassic and Cretaceous erosional surfaces (Fig. 10). They are preserved, however, within some of the deepest parts of the half grabens, as indicated in well 34/71-15S for example.

Paris Basin.—The Bathonian regressive phase shows the restoration of carbonate platforms over the Ostrea acuminata marls (Fig. 5, 8). Infilling sequences developed during the Early Bathonian times with low-energy mud-rich facies; whereas high-energy, oolite-rich forereefing sequences dominated during the Middle Bathonian (Fig. 11). Linked with the forereefing sequences, two major features develop: (a) large prograding lowstands on the margins of the Paris Basin carbonate platform (Calcitres de Saint Pierre du Mont Formation and Calcaires de Blanville Formation in Normandy; Rioul et al., 1991; Oolite Blanche Formation in Burgundy; Floquet et al., 1989; Manalt, 1993; Duch, 1994), (b) a widespread hiatus of the mid-

Peak Transgression.—

Northern North Sea.—The peak transgression is a major drowning event (base Heather Shales) that is characterized by a major increase in the abundance and diversity of dinocysts, reflecting a fully open-marine environment (Williams, 1992). Where sections are continuous, this drowning coincides with the maximum flooding surface of sequence Bj5 (Zigzagiceras zigzag ammonite Zone). Frequently, however, the lowermost part of the Heather shale is so condensed that a hiatus may develop such that younger sections of the Bathonian Heather Shales may directly overlie the Tarbert sandstones (Fig. 10).

Mid-Norway.—The situation and dating are similar (Fig. 2). The drowning coincides with the lowestmost surface of the Melke Formation in the Viking Group (Vollset and Doré, 1984; Daland et al., 1988).

Western Paris Basin.—In the Dorset section, the peak transgression corresponds to the top of the Zigzagiceras zigzag beds,
TOARCIAN TO OXFORDIAN
N-S SECTION OF THE EASTERN PARIS BASIN

Fig. 8.—North-south cross section on the eastern margin of the Paris Basin.
AN OVERVIEW OF 2ND-ORDER TRANSGRESSIVE/REGRESSIVE FACIES CYCLES IN WESTERN EUROPE

FIG. 8.—Continued.
Bathonian on structural highs and on raised parts of major tilted blocks, such as in the southeastern Paris Basin, Lorraine, Luxembourg, southeastern Germany (Braun Jura, Dietl, 1977) and southern Jura mountains (Mangold, 1970).

Subalpine Dauphine Basin.—Several hundred meters of silty shales, known as the Terres Noires accumulated in the highly subsiding Dauphine Basin, from latest Bajocian to Early Oxfordian times (Fig. 2). The Bathonian part is generally undated due to the rarity of macrofossils (Darreau et al., 1994). However, where ammonites have been found such as in the Digne area, main part of the Upper Bathonian deposition is missing (Sturani, 1967) as a result of a major phase of extensional tectonics and block tilting.

Second-Order Cycle 8: Upper Bathonian through Lowermost Upper Oxfordian Substages

Age of Key Surfaces.—

The basal boundary is dated at Middle/Upper Bathonian boundary (Morrisiceras morrisi/Procerites hodsoni ammonite zonal boundary; 167 MY). The upper sequence boundary is dated as lowermost Upper Oxfordian (base of the Epipeltoceras bimammatum Zone in the Tethys and base of the Ringsteadia pseudocordata in the Boreal; 155.5 MY). The average duration is 12 MY. The peak transgression is dated as Lower Oxfordian, Cardioceras scarburgense Subzone.

The initial definition of third order sequences has been given by Dardeau et al., (1994); Floquet et al., (1989, 1992); Garcia, (1993); Gygi et al., (this volume); Jacquin et al., (1992); Javaux, (1992); Partington et al., (1993 a,b); Ponsot, (1994); Stephen et al., (this volume), Rioult et al., (1991); Vidier et al., (1995).

Lowermost Unconformity.—

The basal unconformity represents a major extensional event in relation to the North Sea and Tethyan rift systems. It is characterized by frequent intrabasinal hiatuses covering all Middle Bathonian strata at least. On the Ligurian Tethys passive margin, this event was the last extensional episode prior to the post-rift thermal subsidence (Lemoine et al., 1986, Lemoine and Graciansky, 1988; Rudkiewicz, 1988). Within the northern North Sea basins, it was the second extensional pulse (Fig. 2). It can also be characterized as a turning point between overall progradation and overall retrogradation. This major change does not occur simultaneously from the north to the south. Around the Ligurian Tethyan margin, the Upper Bathonian strata (Cadomites bremeri Zone, Bt 3 flooding) can be transgressive onto older rocks, including Triassic or even basement. This is known on structural intrabasinal highs of the subalpine domain (Coadou and Beaudoin, 1975), on major tilted blocks of the Tethyan margin such as the Briançonnais (Lemoine and Graciansky, 1988) and on the eastern (Massif Central) and southern (Provence) margins of the Dauphine Basin (Bodeur, 1994; Elmi, 1984, Dardeau et al., 1988). Within the Paris Basin, the onset of the transgression can be dated of the Upper Bathonian (Prohecticoceras retrocostatum Zone, Bt4 flooding). It is documented in Normandy to the west (Rioult et al., 1991), in the Boulonnais to the north (Vidier et al., 1995) and in the Braun Jura in Southern Germany (Dietl, 1977). In the United Kingdom area, the turning point between the two cycles can be dated as latest Bathonian (Clidoniceras discus Zone, Bt5 flooding; Stephen et al., this volume; Underhill and Partington, 1993 a, b). In the northern North Sea, the top of the prograding Krossfjord sandstones corresponds to the Bt 3 sequence boundary (Jacquin et al., in prep). This time delay reflects the diachronity, at the scale of the European craton, of the onset of the thermal subsidence following the mid-Bathonian rifting phase.
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**Fig. 10.—**Comparison between Horda Platform and Tampen Spur sequence stratigraphic frameworks. Note: we do not extend the Tarbert Formation up to the mid-Bathonian unconformity as proposed classically (Mitchener et al., 1992). The Tarbert sandstones are strictly limited to the backstepping phase at the top of the Brent delta and prior to its drowning. The Krossfjord sandstones are limited to the infilling sequences that prograde over the drowned Brent delta.

Transgressive Phase.—

The stratal and facies stacking patterns of this transgressive phase vary according to location, independently of the nature of the facies concerned. This shows that the tectonic development, in response to the North Sea rifting, was the most prominent factor, at least during the early stage of the overall transgression.

Northern North Sea.—The overall transgressive phase from the Middle Bathonian (Bt3 Sb) to the lowermost Oxfordian strata (MFS Ox1, top *Cardioceras mariae* Zone), was modulated by a renewal of extensional activity, with block faulting, syn-rift progradation and the development of a pronounced basin-scale unconformity. On the Horda platform, a first peak transgression is recorded in the lowermost Callovian deposits (MFS Ca1, *Proplanulites koenigi* Zone). It corresponds landward to a tongue of Heather-type shales, between the Krossfjord sandstones below and the regressive Fensfjord sandstones above (Fig. 2). This flooding is followed by the progradation of shoreface sandstones of the Fensfjord Formation, which form large syntectonic wedges. Locally a mid-Callovian erosional unconformity may develop as a consequence of block tilting. In areas where the series were well developed, the top of the regressive Fensfjord sandstones corresponds to the Callovian/Oxfordian boundary. This temporary regressive phase is also known in Jameson Land (East Greenland; Surlyk, 1991; Partington et al., 1993 a, p. 366). Southern North Sea.—In the Moray-Firth Basin of the southern United Kingdom North Sea, Stephen and Davies (this volume) document a succession of marginal marine backstepping sequences, onlapping Upper Bathonian regressive coal-rich deposits, as well as older rocks landward. These culminate at a peak transgression dated at the middle/upper Callovian boundary or Callovian/Oxfordian boundary depending on the location.

Paris Basin.—The transgressive phase shows the classical succession of lagoonal, mud-rich, aggrading sequences and high-energy, oolitic-rich, backstepping sequences (Fig. 13; Jacquin et al., 1992). The aggrading sequences, dated as Late Bathonian, correspond to the main development of the Comblanchien platform (Fig. 5, 8). This aggradational phase ended with a major 3rd-order (eustatic?) oscillation (Bt5 Sb = 164.7 Ma; previous 158.5 Ma of the Haq et al. 1988 chart), which shows...
records the same intraplate tectonic event as in the North Sea (Fig. 12). The location of oil and gas discoveries in the Paris Basin is closely controlled by the distribution of the remnant Ca2 highstand carbonate wedges below the Ca3 sequence boundary and by the associated perched bioclastic lowstands (Javaux, 1992).

**Peak Transgression.**—

The peak transgression is one of the most widespread and correlatable drowning events on the European craton. It is dated as close to the Callovian/Oxfordian boundary. In the Northern North Sea (on Tampen Spur) the only parts of the Heather Shale Formation that cover the structural highs are those dated in the Lower Oxfordian, time-equivalent to the known peak transgression landward. On the Horda platform, the Callovian Fensfjord sandstones are drowned by recurrent lowermost Oxfordian Heather Shales (Van der Zwan, 1989). A similar situation is observed and described within different fields of southern North Sea (Partington et al., 1993 a, b; Stephen and Davies, this volume). In the Paris Basin, the peak transgression is characterized by a marker bed named as the Oolite Ferrugineuse, where a large span of time can be highly condensed on structural highs, from middle Callovian to middle Oxfordian times (Fig. 2). The ferruginous oolite is frequently a stack of several condensed intervals (Marchand et al., 1982), separated by erosional surfaces induced by the late Callovian deformation phase (Javaux, 1992). In the subalpine Dauphiné Basin, the Cardioceras scarburgense Subzone (earliest Oxfordian) is well known as a period of relative starvation and fossils accumulation. The platforms surrounding the Dauphiné Basin (Provence, Eastern Massif Central and Swiss Jura) were drowned at this time (Dardeau et al., 1994; Gygi et al., this volume).

**Regressive Phase.**—

This phase dated Early to Middle Oxfordian is particularly well calibrated with ammonites from the Tethyan margins, Paris Basin and United Kingdom areas and with dinoflagellates in more northern areas (Stephen et al., this volume; Partington et al., 1993 a, b; Marchand et al., 1990; Fortwengler, 1989; Dardeau et al., 1994; Coe, 1992, 1995; Cope et al., 1980; Stephen et al., 1993; Van du Zwan, 1989; Milton, 1993; Cox, 1990).

The regressive phase comprises two successive parts that are characteristic of basins undergoing strong extensional tectonic activity: infilling-type sequences with a moderate amount of progradation during Early Oxfordian times and then forestepping type sequences with tectonic enhancement during the Middle Oxfordian. These large-scale stratal patterns can be observed in both siliciclastic (northern and southern North Sea) and carbonate settings (Paris Basin and Tethyan margins).

**Northern North Sea.**—On the Horda Platform, the sandstones of the Sognefjord Formation form a series of shoreface Basinward stepping wedges that change facies westward into the Heather shales (Fig. 2). They mainly developed during the forestepping stages of the overall regression (from OX4 to OX6 top LST). These sandstones form the reservoirs of the giant Troll field on the Horda Platform (Osborne and Evans, 1987). Most of these sandstones merge landward (i.e., the Norwegian hinterland) towards a major tectonically enhanced unconformity. The mid-Oxfordian unconformity is also very active on widespread exposure and incision at the sequence boundary (Floquet et al., 1989; Rioult et al., 1991; Javaux, 1992; Ponsot, 1994). The subsequent retrogradational phase is modulated by high-amplitude 3rd-order oscillations, noticeably the MFS Bt5, MFS Ca1, MFS Ca3 and MFS Ca5 and Ca1/Ca2 sequence boundary. These are well recorded throughout the Paris Basin (Garcia, 1993). The carbonate depositional environments were definitively drowned, after the Ca 3 sequence boundary, which was responsible for the almost complete exposure of the carbonate platform. The north-south transect (Fig. 5, 8) on the eastern side of the Paris Basin shows how the last remnant carbonate wedges, belonging to the Ca2 highstand, are overlain to the north and west by relatively thick upper Callovian shales and to the south and east by very thin,starved, upper Callovian and even Lower to Middle Oxfordian ferruginous oolites (Marchand et al., 1982). It is possible that this upper Callovian westward tilting of the southeastern border of the Paris Basin
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PARIS BASIN BURGUNDY DAUPHINE

**A. -Mid-Cimmerian unconformity : 180.5 Ma**
- Uplift and associated erosional truncation in the north-south cross section displayed on Fig. 7.

**B. -Earliest Bajocian max. progradation R6 : 176.0 Ma**
- Forestepping sequences

**C. -Latest Bajocian peak transgression T7 : 170.0 Ma**
- Backstepping sequences

**D. -Mid-Bathonian maximum regression R7 : 166.8 Ma**
- Infilling Sequences

**E. -Drowning of the Callovian platform (Ca3 Sb) : 161.9 Ma**
- Aggrading sequences

**F. -Earliest Oxfordian peak transgression : 159.0 Ma**
- Backstepping sequences

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**FIG. 12.**—Major steps in the evolution of the eastern margin of the Paris Basin (from north-south cross section displayed on Fig. 7). (A) Mid-Cimmerian Unconformity (180.5 Ma): Uplift and associated erosional truncation in the north. (B) Earliest Bajocian Maximum Regression (176.0 Ma): Development of a subaerial exposure surface over most of the Paris Basin. Rapid progradation of carbonate platforms towards the Dauphine Basin margins, during Aalenian and early Bajocian times. (C) Latest Bajocian Peak-Transgression (170.0 Ma): Widespread development of carbonate platforms over the Paris Basin and surrounding areas (aggrading first and then backstepping during the Late Bajocian). (D) Mid-Bathonian Maximum Regression (166.8 Ma): Reestablishment of carbonate platforms and rapid progradation with high-energy facies. Locally (eastern margin, Jura) a mid-Bathonian unconformity may develop. (E) Drowning of the Callovian Platform (161.9 Ma): Widespread development of carbonate platforms over the Paris Basin (Comblanchien limestone) and surrounding areas (aggrading during the Bathonian and backstepping during the Lower Callovian). A westward large scale tilting may be deduced by the wedge shape of the Lower Callovian backstepping sequences. (F) Earliest Oxfordian Peak-Transgression (159.0 Ma): Continued westward tilting of the eastern margin of the Paris Basin (Ca3 to Ox0). Widespread peak-transgression reached during the earliest Oxfordian (mfs Ox0) along a ferruginous oolite layer.

Paris Basin.—The Paris Basin presents another good example of the regressive phase but in carbonate setting. The infilling sequences form widespread blankets of hemipelagic marls and mud-dominated shelfal sediments downlapping onto the highly starved ferruginous peak transgression surface (Fig. 5, 13). These marls have various names according to the location: Creniceras rengeri Marls Formation in the Eastern Paris Basin (Gygi et al., this volume), Ancy le Franc Marls Formation in Burgundy, (Marchand and Menot, 1980), Marines de Villers Formation in Normandy (Rioult et al., 1991). During that infilling phase, carbonate factories reestablished, starting to build outward and shedding carbonate materials downdip. Middle Oxfordian forestepping sequences show rapid progradation of reefal and associated high-energy platform rim environments towards the most subsiding areas (Fig. 5, 13). During that time, thick prograding bioclastic and reefal lowstand deposits develop forming potential reservoirs for gas storage (Blondin, 1994). On structural highs, these forestepping sequences may merge towards a major mid-Oxfordian unconformity such as their time-equivalent on the Horda platform in the northern North Sea. At the northwest corner of the Paris Basin in Normandy, these bioclastic lowstand deposits show a progressive westward enrichment in detrital quartz deriving from the Armorican Massif (Rioult et al., 1991; Dugué and Rioul, 1989). This trend was more pronounced during the subsequent 2nd-order facies cycle. From wireline-logs correlation across the Paris Basin, it is possible to show how purely massive carbonate lowstand deposits in the east become purely siliciclastic in the west.

Subalpine Dauphine Basin.—The infilling sequences form huge shale-prone accumulations of hemipelagic silty shales, up to 600 m thick (Terres Noires Formation, Dardeau et al., 1994). Middle Oxfordian equivalent forestepping sequences become more calcareous (Argovian facies) and thinner, as a consequence of the overall Late Jurassic decreasing subsidence (Dardeau et al., this volume).

Within the carbonate facies belt, a pronounced marker for large-scale correlations characterizes the first middle Oxfordian
Fig. 13.—Upper Jurassic strata of the Paris Basin center. R8: Early to Middle Oxfordian regressive phase. Infilling sequences are mud-dominated and cover the entire center of the Paris Basin. They are coeval with the Oxford Clays or Heather shales in the North Sea. Middle Oxfordian forestipping sequences (Ox4 to Ox6) form large reefal prograding wedges, whose facies change into open-marine marls. Observe the migration towards the north-east (seaward) of the successive pinchout points of LST Ox5 and Ox6, with the overall 2nd-order progradation. T9: Late Oxfordian to Late Kimmeridgian (sensu gallico) transgressive phase. Aggrading sequences correspond to the development of a widespread carbonate platform (Calcaire de Tonnerre) with a reefal rim. The platform temporarily drowned at the Ki1 MFS (Pictonia baylei Zone) and definitively at the Ki3 MFS (Aulacostephanoides mutabilis Zone). The Calcaires à Astartes are bioclastic facies dominated by foramol faunal associations, which form the transition between the “healthy” reefal Late Oxfordian carbonate platform (prior to Ki1) and its definitive drowning. The peak transgression is recorded at the point of maximum encroachment of open-marine facies landward (Nanogyra virgula marls) at the MFS of Ki6 sequence. R9: Tithonian regressive phase. Carbonate depositional environments reestablished quickly after the peak-transgression and formed the last extended carbonate platform of the Jurassic period in the Paris Basin (Calcaire du Barrois). At the top, emersive gypsiferous layers (Purbeckian facies) are the last deposits prior to the late Cimmerian (Stille, 1924) unconformities in the lowermost Cretaceous.
foresteepping sequence. It is a through going reefal episode, corresponding to the transgressive systems tract of the 3rd-order sequence Ox4, which is dated as late Periphiocystes (Dichotomocystes) antecedens to P. (D.) antecedens/Periphiocystes (P.) parandieri Subzones boundary of the Gregoryceras transversarium Zone (Enay, 1966). Within the Anglo-Paris Basin (Fig. 2), this interval is marked by the Coral Rag in Yorkshire and Dorset (Cope et al., 1980; Wright, 1986; Coe, 1992), by the Calcaires à Polypiers in Maine and Normandy (Rioul, 1971) and/or by the reefal Calcaires de Sainte Ursanne in the Swiss Jura (Gygi et al., this volume). In main parts of the White Jura, the interval coincides with the onset of extensive reefal building, that lasted throughout the entire Late Oxfordian stage (Enay, 1966).

Second Order Cycle 9: Upper Oxfordian Lower Volgian (= Tithonian) Substages

Age of Key Surfaces.—

The basal boundary is dated at the Lower/Upper Oxfordian boundary (Dichotomocystes bifurcatus/Epiphiocystes bimammatum ammonite Zone; 155,5 Ma). The upper boundary is dated at the Lower/Upper Volgian boundary (Boreal) or Tithonian/Berriasian (Tethyan) (144 Ma). The duration is 11,5 Ma. The peak transgression is dated as Late Kimmeridgian, upper Aulacostephanus eudoxus or Aulacostephanus autissiodorensis ammonite Zones depending on the area.

The initial definition of third order sequences has been given by Coe (1992; 1995); Hantzpergue, (1985, 1991, 1993); Herbin et al., (1995) and previous works; Leinfelder and Wilson, (this volume); Leinfelder, (1993); Partington et al., (1993a, b); Prousot, (1994); Proust et al., (1993); Stephen et al., (this volume); Stephen et al., (1993); Vidier et al., (1995); Wignal, (1991).

Basal Unconformity.—

The basal unconformity corresponds to the mid-Oxfordian unconformity. It is well dated at the turning point between the overall progradation of the foresteepping sequences below and the overall aggradation of the transgressive sequences above. This turning point is the top lowstand of the last foresteepping sequence of the preceeding regressive phase. It appears to be relatively synchronous over the whole European craton, whatever the setting, siliciclastic or carbonate. This particular point can be precisely dated in areas where the series is continuous. Generally, it corresponds to the flooding surface of the Ox6 sequence, but locally it may be slightly younger, at flooding surfaces of Ox7 or Ox8 sequences which are dated as Late Oxfordian. In the northern North Sea on the Horda Platform, the fault activity has been extremely reduced during the Kimmeridgian epoch was also a time of rapid subsidence whatever the structural framework: postrift (Tethyan margin) or synrift (southern North Sea).Northern North Sea.—On the Horda Platform, the uppermost sandstones of the Sognefjord Formation quickly backstep during latest Oxfordian and earliest Kimmeridgian times, without any development of an aggradational phase (Fig. 2). They interfinger with high gamma-ray shales of the Draupne Formation (previously called Kimmeridge Clays: Wollset and Doré, 1981). These first Draupne Shales are truncated by a regional Early Kimmeridgian unconformity (sensu anglico, SB Ki6), which developed in both the greater Tampen Spur area and the Horda Platform (Fig. 10). This unconformity relates to the end of the synrift period, as most of extensional faults are sealed by further deposits. On the greater Tampen Spur, the Lower Kimmeridgian Draupne Shales are missing from intrabasinal highs, whereas Upper Kimmeridgian hot shales (Aulacostephanus eudoxus zone for the oldest) are sitting unconformably over intrabasinal structural highs and over graben thick successions. On the Horda Platform, the Aulacostephanus eudoxus and Aulacostephanus autissiodorensis Draupne Shales also overlie unconformably older deposits.

Southern North Sea.—Whereas along the axis of the Viking Graben fault activity has been extremely reduced during the Kimmeridgian backstepping phase (Brown, 1987), the fault activity in the South Viking Graben and Moray Firth reached a maximum at that time (Rattey and Hayward, 1993; Boldy and Breailey, 1990). This produced fault-controlled detritics interlayered with organic shales, ranging from alluvial fans to submarine fans, that have been interpreted as regressive deposits. This includes the main part of the Brae and Miller reservoirs, the Fulmar Formation and the upper Ula Sandstones in the Central Graben and the Piper Sandstones in the Ground Graben.

Paris Basin.—Two types of sequences occur, aggrading during the upper Oxfordian-lowermost Kimmeridgian and backstepping during most of Kimmeridgian times (Fig. 2). On its western side (Normandy and Boulonnais), widespread thick piles of Upper Oxfordian shales, similar to those found in the Dorset
(Sandsfoot Clay Formation), characterize the aggrading part of the transgressive phase (Rioul et al., 1991). They interfinger towards the center of the Paris Basin with an aggradational carbonate platform that developed further eastward covering widespread areas on the northern Tethyan margins. An east-west transect across the Paris Basin (Fig. 14) shows that the Upper Oxfordian aggrading sequences were continuous and isopach, whether their substratum is a thick Lower Jurassic and Triassic series or directly basement. This is indicative of high rates of tectonic subsidence reached at that time (up to 150 m/my).

During the Kimmeridgian backstepping phase, the overall stratigraphic pattern is similar (Fig.), even though high-amplitude third-order oscillations modulated the record (Fig. 15). These are the Ki2Sb (153.5 Ma equivalent to the D2 discontinuity of Hantzpergue, 1985), the Ki4Sb (152.2 Ma, equivalent to his D8) and the Ki6 Sb (151.5 Ma, equivalent to his D10). These sequence boundaries led to the widespread development of aggradational lowstands, which grade from purely lime-mudstone wackestones in the center of the Paris Basin to purely sandstones on its north-western margin (Boulonnais, Proust et al., 1993). The temporary drowning of the aggradational carbonate platform in the earliest Kimmeridgian (Fig. 14), as well as its definitive drowning during the Aulacostephanos autissiodorensis Zone, correspond with major 3rd-order flooding events (mfs Ki1 for the first drowning, mfs Ki3 for the second). They are also indicated by the turnover of the ammonite-fauna within the whole European craton (Hantzpergue, 1985, 1990).

**Peak Transgression.**

The peak transgression must be defined on shelfal areas by the age of the maximum extension of open-marine facies towards the land masses (Nanogyrus virgula Marl's Formation in the Paris Basin for instance, Figs. 11, 12, 14). Within the western European carbonate platforms, this point was reached during the Aulacostephanus eudoxus (mfs Ki5) or Aulacostephanus autissiodorensis (MFS Ki6) ammonite Zones. The onset of the following regressive Portlandian limestones (“Calcaires du Barrois” Formation in the Paris Basin: Menot, 1981) and the onset of the Tithonian platform-derived gravity deposits in the Tethyan marginal basins, are dated as Early Volgian (Ki7 sequence boundary) following the Aulacostephanos autissiodorensis (Ki7) MFS.

Organic-rich, condensed shales have been deposited within the North Sea Volgian layers; they overlie Kimmeridgian organic-rich shales. As source-rocks typically develop during transgressive phases (Jenkyns, 1980, 1985), the age of the peak transgression has been debated recently (Wignal, 1991; Clark et al., 1993; Herbin et al., 1995; Jacquin et al., this volume). We consider that the deposition of Volgian organic-rich shales resulted from local physiographic conditions in relation with the development of the Late Jurassic North Sea rift during the post-Kimmeridgian 2nd-order regressive phase.
Regressive Phase.—

One of the main characteristics of the regressive phase is the widespread development of mud-dominated Portlandian carbonate platforms, succeeded by emmersive, lacustrine to evaporitic layers (so-called Purbeck facies).

Northern North Sea.—Sand-prone gravity deposits followed by shoreface progradations (Magnus Formation in United Kingdom, Munnin Formation in Norway; Vollset and Doré, 1984) are interfingered with the Draupne Shales, around different raised edges of tilted blocks, such as the Snorre high. Their occurrence covers most of the lower Volgian, from *Pectinatites* (*Virgatosphinctoides*) wheatleyensis to *Titanites* anguiformis Zone. They are bounded at the base by a major erosional surface, which corresponds to a pronounced downward shift of the coastal onlap (Ti2 Sb?). These features are probably controlled by a significant phase of structural inversion that started exactly at that time and was recurrent until the Early Cretaceous times (Frost, 1987; Booth et al., 1992).

Paris Basin.—The regression is marked by two major steps, one being dated as uppermost *Aulacostephanus autissiodorensis* Zone (Ki7 Sb) at the base of the Portlandian regressive limestones and the other one of the lowermost *Pectinatites* (*Virgatosphinctoides*) wheatleyensis Zone (Ti2 Sb) at the base of the Purbeckian facies (Fig. 5, 9).

The top *Aulacostephanus autissiodorensis* event (Ki7 Sb of our chart or D13 discontinuity by Hantzpergue, 1985; Ki6Sb of Wignal, 1991) is a major sea-level drop that dramatically initiated the overall regression. It has a widespread record from the Dinarides (Karstic features and bauxite development in Slovenia; Strohmenger et al., 1991), to the Swiss and French subalps (Detraz et al., 1987; Detraz and Steinhauser, 1988; Mojon and Stasser, 1987; Enay, 1984), to the Northern Aquitaine platform (Gability et al., 1985) and to the Lusitanian Basin (Leinfelder, 1993). On the western margin of the Paris Basin (Boulonnais), a major siliciclastic lowstand progrades over the organic-rich *Aulacostephanus autissiodorensis* shales (Herbin et al., 1995). In Dorset and Yorkshire, this Ki7 (eustatic?) sea level fall temporarily interrupts the organic-rich shale deposition (Herbin et al., 1995).

The base *Pectinatites* (*Virgatosphinctoides*) wheatleyensis major exposure event (Ti2 Sb or D14 of Hantzpergue, 1993)
caused the death, by exposure, of the Portlandian carbonate platform. It is followed in relative continuity (in the center of the Paris Basin at least) by the deposition of gypsiferous Purbeckian facies (Fig. 5, 13). The same major exposure surface is recorded in the Aquitaine Basin (D14 discontinuity of Hantzpergue, 1993 and K17 Sb of Wignal, 1991) and in the Jura on the northern margin of the Dauphiné Basin (Detraz and Steinhauer, 1988).

Subalpine Dauphiné Basin.—The Tethyan Tithonian (= Lower Volgian of the Boreal realm) was a time of overall regression. This was marked by continuous subaerial exposure conditions on the peri-Tethys margins, and by the associated long-term forced regression and gravitational deposits in the Basinal areas (Beaudoin, 1977; Beaudoin et al., 1975, Joseph et al., 1988; Detraz et al., 1987; Detraz and Steinhauer, 1988). Within the northern North Sea Basins, a second and last occurrence of organic-rich Draupne shales (dated upper Volgian) overlies either unconformably older strata, or conformably non-organic Draupne shales from the Titans anguiformis Zone. These various observations constrain the timing of the unconformity as being post-Titans anguiformis and pre-Berriasella (Hegaratella) jacobis, that is around Ti8 Sb or Be1 Sb, whatever the length of the gap in various places.

2nd-Order Cycle 10: Ryazanian (Nordic) or Berriasian (Tethyan) Stages

Age of Key Surfaces and Forewords.—

The basal boundary is dated at the Lower/Upper Volgian boundary (intra-Titans oppressus; 144 Ma). The upper boundary is dated as upper Ryazanian (pre-Surites (Bojarkia) stenomphalus (Surites iceni ?) = Upper Berriasella picteti of the Tethyan realm; 138 Ma). The duration is 6 my. The peak transgression is dated as Middle Ryazanian (Hectoroceras kochi ammonite Zone) in the Nordic or Middle Berriasian (uppermost Subthurmanna (Strambergella) subalpina ammonite Zone) in the Tethyan areas. Initial definition of third order sequences was given by Jan Du Chêne et al., (1993) and key paper for correlation at the European scale by Allen and Wimbledon (1991).

The cycle 10 corresponds to the Purbeckian interval pro parte, known on the European shelves in Poland, Lower Saxony, West Netherland, Anglo-Paris Basin, Aquitaine, Basco-Cantabrides and Betics in Spain. It is time-equivalent to the Berriasian calcareous mudstones and gravitational deposits layered down conformably with the Tithonian limestones in the Tethyan Basinal areas. The “Purbeckian” and Berriasian successions developed at the end of the 1st-order major regression that began in the latest Jurassic, before the long term, 1st-order Early Cretaceous transgression. It is why, in spite of its earliest Cretaceous age, the Purbeckian cycle is still grouped within the North Sea, mainly Jurassic in age, 1-st order cycle.

Lowermost Unconformity.—

A major tectonically enhanced unconformity separates the lower from the upper Volgian strata. This unconformity has been well documented in the Weald subbasin (United Kingdom) where gypsiferous beds (English Lower Purbeck) containing Berriasian charophytes overlie disconformably “Portlandian” limestones dated of the Galbanites kerberus Zone (Wimbledon and Hunt, 1993, in Allen and Wimbledon, 1991). In the Paris Basin, a similar erosional unconformity can be shown (Fig. 5, 13), but within the Purbeckian Formation, between the lower gypsiferous regressive part and the upper calcareous transgressive part. Similar observations have been made in Belgium, west-Netherlands and Denmark subbasins (Allen and Wimbledon, 1991). In the Wessex Basin, in the type area of the Purbeck Formation, the basal gypsiferous “Purbeck beds” (Upper Volgian) conformably overlie the Titans anguiformis Zone of the type “Portlandian” section (Wimbledon and Cope, 1978). On the Tethyan Basins margins, such as in the Jura, the uppermost Jurassic platform carbonates end with a series of paleosoils and karstic surfaces (the so-called Tidales de Vouglans; Bernier, 1984). They are overlain by marine “Purbeckian” limestones, yielding Berriasella (Hegaratella) jacobis and Pseudosubplanulites grandis ammonites (Be, MFS: Detraz et al., 1987; Detraz and Steinhauer, 1988). Within the northern North Sea Basins, a second and last occurrence of organic-rich Draupne shales (dated upper Volgian) overlies either unconformably older strata, or conformably non-organic Draupne shales from the Titans anguiformis Zone. These various observations constrain the timing of the unconformity as being post-Titans anguiformis and pre-Berriasella (Hegaratella) jacobis, that is around Ti8 Sb or Be1 Sb, whatever the length of the gap in various places.

Transgressive phase and maximum flooding surface.—

United Kingdom.—In both the Weald and Wessex subbasins, the succession shows above the basal unconformity (1) gypsiferous beds, (2) Classopolis rich Freestone layers, (3) the Cypriade (ostracod rich) Cockle beds and finally (4) the marine Cinder beds that record the peak transgression (Allen and Wimbledon, 1991).

Paris Basin.—The upper calcareous part of the Purbeckian succession, from wire-line logs and (sparse) cores from the Basin center where the section is the most complete, shows three marine (oolitic, ostracod, bryozoan-rich) backstepping depositional sequences. They onlap at the basin margins onto the lowermost unconformity. They thin upwards and end with more open-marine bioclastic facies with palynological association characteristic of the middle Berriasian (Jan du Chêne, pers. commun., 1994). They are overlain unconformably by widespread black shales, dated of the uppermost Berriasian. These black shales belong to the next transgressive cycle (T11a).

A good stratigraphic correlation, using palynofacies and dinoflagellates, can be made between the marine “Cinder beds” and the calcareous “Purbeck” in the Paris Basin. This peak-transgression is also described in the boreal realm, including Greenland (Surlyk, 1973), where it corresponds to the last “Hot Shales” of the Draupne Formation (dated by dinocysts as intra-Hectoroceras kochi ammonite Zone).

Lower Saxony Basin.—The Purbeck time-equivalent is the Weiteven Formation (Fig. 2). It overlies unconformably the Lower and Middle Jurassic strata of the Altena Group, with conglomeratic time transgressive deposits at the base, rapidly followed by evaporitic and shaly playa sediments. The peak-transgression is reached with a marginal marine limestone-rich interval (Serpulite member or Munder Formation), that has been proved to be coeval with the Cinder beds (Strauss et al., 1993).

Subalpine Dauphiné Basin.—The peak transgression is defined by the most starved interval and by geochemical criteria (Jan du Chêne et al., 1993). This event is dated as intra-Subthurmanna (Strambergella) subalpina Zone (mfs Be3) at the Berriasian stratotype at Berrias (France).

Regressive Phase and “Base Cretaceous Unconformity”.—

The regressive phase is widespread all over Europe (Fig. 2). It shows the return, from open-marine environments at the peak
transgression, to brackish and then to continental environments at the end of the regression (middle to upper Purbeck in United Kingdom: Allen and Wimbledon, 1991; Buckeberg Formation, Oenkirchen member in the Lower Saxony Basin: Strauss et al., 1993; Coevorden Clay in West Netherlands: Nederlandse A.M., 1980).

The maximum regression can be precisely calibrated when combining information from the German Wealden, the English Purbeck and the Tethyan Berriasian. Strauss et al. (1993) documented within the Wealden Group in southern Germany that the turning point between the regressive phase and the subsequent Cretaceous transgression is between the Wealden 3 and Wealden 4 members defined by Wolburg (1949). Palynological evidence shows that this change occurs between the Surites (Bojarkeia) stenophalus and Peregrinoceras albidum ammonite Zones, near the base of the Pseudoceratium pelleferum dinocyst Zone. This precise point exactly coincides with the Be7 sequence boundary, which has been calibrated with the same dinocyst association to the Berriasella callisto ammonite Zone at various type sections in southeastern France (Jan Du Chêne et al., 1993). In the French subalpine chains, rimmed carbonate platforms extended seaward by overall progradation during the middle and late Berriasian times. During the late Berriasian times, the carbonate platforms were rapidly covered by backstepping bioclastic limestones (Vions Formation: Detraz et al., 1987; Detraz and Steinhauser, 1988). This change occurred with the Be7 sequence boundary, which is here dated at the lower part of the D3 Calpionellid zone (i.e., of the Berriasella callisto ammonite Zone).

Paris Basin.—The “Purbeckian” succession is truncated by a major Basin-scale unconformity. Widespread black shales of the next transgressive cycle (Ti1a) overlie the Purbeckian limestones. The minimum hiatus between the black shales and the limestones ranges from Hectoroceras kochi to Surites (Bojarkeia) stenophalus ammonite Zones (from dinocysts).

North Sea.—A major unconformity named Base Cretaceous Unconformity (“BCU”: Johnson, 1975), frequently with an important time gap, characterizes the lowermost surface of the Lower Cretaceous hemipelagic series. This surface may erode deeply the underlying Jurassic strata (Fyfe et al., 1981). It is a major basin-scale onlap surface. In areas where the succession is more conformable, such as in the Viking graben or within the subsiding parts of its margin, a late Ryazanian, non-organic, hemipelagic wedge onlaps the unconformity. It can be dated as younger than the Hectoroceras kochi Zone and older than the Surites (Bojarkeia) stenophalus Zone, which indicates that the lowermost onlap surface fits with the Be4 Sb.

CONCLUSION

Since the North Sea Cycle was a time of active crustal separation in the central Atlantic and in the North Sea domain (Emery and Uchupi, 1984), we might expect that the stratigraphic signature of the different basins’ development would vary with local tectonic features. In fact, only two periods are significantly controlled by local tectonic factors: the middle Callovian synrift regressive phase on the Horda Platform (R8a) and the Lower Kimmeridgian fault-controlled regressive deposits (R9a) in the southern North Sea. The four other major 2nd-order regressive phases (R7: Lower to Middle Bathonian, R8: Lower to Middle Oxfordian, R9: Lower Volgian and R10: Ryazanian) are surprisingly synchronous throughout western European basins, within the limits of the available biostratigraphic tools. This of course would suggest that local tectonics have little effect upon the development of long-term stratigraphic cycles. This statement should be tempered by the following remarks:

1. When comparing accommodation curves from different basins there is no doubt that 1st-order long-term tectonic subsidence controls most of the space being created and thus controls the overall stratigraphic pattern. This of course would suggest that local tectonics have little effect upon the development of long-term stratigraphic cycles. This statement should be tempered by the following remarks:

2. The amplitude, the timing and the wavelength of the 2nd-order changes of the accommodation are quite similar from Basin to Basin, if the 1st-order tectonic trend is subtracted. These measurements could accurately estimate a possible 2nd-order eustatic signal, whatever its origin.

3. Concerning the four synchronous T/R cycles, there is no physical relationship between their development (i.e., the overall progradation) and block faulting and tilting. Generally, the tectonic features superimpose their effects on the overall progradation during the regressive phase or on the overall retrogradation during the transgressive phase, but do not create the progradation or the retrogradation. These latter are definitely controlled by long-term changes of the base-level with respect to sediment supply. Tectonic features will create erosional surfaces, whether they are basinward or landward. They will enhance an overall progradation, locally creating forstepping surfaces, or enhance an overall retrogradation, locally creating starved intervals or backstepping sequences instead of aggrading sequences.

4. Some major 3rd-order exposure surface, that occur at critical points in the 2nd-order evolution, are closely associated with the tectonic development of the Basins, mainly during the rifting phases. We can emphasize the Bj3 and Bj4 sequence boundaries, which are recorded as major downward facies shifts on European platforms and which are linked with the onset of extensional activity in the northern North Sea at least. The overall situation of Bt5, Ca1 and Ca3 sequence boundaries during the cycle 8, and K2, K4 and K6 sequence boundaries during cycle 9 is similar; they are all associated with major extensional activity somewhere in the North Sea and recorded as major exposure surfaces on surrounding platforms and shelves.

5. Major third-order floodings seem to be more independent of the tectonic evolution. The maximum flooding surfaces of the following sequences: Bj2, Bj4, Bj5, Bt5, Ca1, Ca3, Ca5, Ox4, K11, Ki2, Ki4, Ki5 and Be3 are recorded as major drowning events within carbonate settings as well as within silicilastic settings and during overall regressive phases as well as during overall transgressive phases. On the contrary, peculiar long-term sediment starvation in the Basinal settings (Volgian and Ryazanian in the North Sea), do not relate to any major flooding on surrounding shelves, but do relate to regional increasing subsidence and decreasing sediment supply.

6. The uppermost Jurassic-lowermost Cretaceous overall regressive phase, at the end of the North Sea Cycle, relates to a transpressional stress field (Booth et al., 1992) associated...
with the reactivation of old northwest-southeast lineaments. This was coupled with the onset of the Norwegian Green-
land sea rift to the north (Ziegler, 1988) and to the Atlantic opening in the Galice area (offshore Iberia) to the south (Le-
moine et al., 1984). However, the relative synchronicity of the influx of clastics in the North Sea Basins, of carbonate
debris in the Tethyan Basins and of the widespread expansion of surrounding shelves suggests a lowering of the sea-level
of that time. The resulting late Cimmerian unconformity ends the North Sea Cycle.

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THE LIGURIAN CYCLE: AN OVERVIEW OF LOWER JURASSIC 2ND-ORDER TRANSGRESSIONAL/ REGRESSIVE FACIES CYCLES IN WESTERN EUROPE

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ABSTRACT: The Lower Jurassic-Aalenian transgressive/regressive major cycle transects the western European craton from the nascent Tethyan margin, now involved in the Subalpine folding in the south, to the mid-Norway basin in the north. It includes the eastern Aquitaine, the Anglo-Paris and the North Sea basins. This cycle has been named Ligurian from the main phase of the rifting that affected the southern part of the craton and led to the opening of the Ligurian part of the Tethys. The lower boundary of this major cycle is the early Cimmerian unconformity dated as latest Norian age. The upper boundary is dated as late Aalenian (mid-Graphoceras concavum ammonite Zone) and is related to the mid-Cimmerian unconformity. Peak transgression coincides with deposition of the organic-rich lower Toarcian black shales (Harpoceras falciferum ammonite Zone). Three 2nd-order transgressive/regressive facies cycles have been identified from the Tethyan margin; they record the successive steps of Ligurian rift development and subsequent phases of subsidence. However, four 2nd-order cycles are recognized in the Paris Basin and five onshore United Kingdom basins. The number of cycles, ages of peak transgressions and main cycle boundaries vary regionally according to the local extensional tectonic development. For example, a major regressive episode is apparent at the end of the Sinemurian stage in northern areas but is not prominent in the southern areas. Instead, a regression dated as early Pliensbachian (Tragophylloceras ibex ammonite Zone) is dominant in the southern areas. Most of the 27 3rd-order depositional sequences comprising the Ligurian major cycle can be documented from the North Sea to southern Europe. Precise age dating was done using ammonites (down to the “horizon” level, where possible), from the Provence shelf in the south to the Hebrides basin in the north, allow correlation. In more northern areas where ammonites are missing, palynological determinations provided the biostratigraphical control.

INTRODUCTION

Second-order cycles in several Early Jurassic basins and sub-basins, located on the western European craton from the Provence Platform on the nascent Tethyan Ligurian margin (Fig. 1) to the mid-Norway basins (Halten Terrace area) on the future Norwegian Sea margin, are reviewed along a transect that trends roughly north-south across the western European craton. We do not attempt to incorporate Lower Jurassic data from the Greenland, Denmark, Germany and Iberia. The 1st-order Ligurian Cycle is bounded by the early Cimmerian (latest Norian) and mid-Cimmerian (late Aalenian) unconformities (Jacquin and Graciansky, this volume). The peak transgression is recorded everywhere by lower Toarcian organic-rich shales.

Stratigraphic units that underwent the Ligurian rifting processes (i.e., the Subalpine area sensu lato, including the Provence platform, Dauphiné basin, Jura platform and eastern Aquitaine areas) are compared to other intracratonic basins where crustal stretching and subsequent subsidence were less pronounced (the Paris Basin; onshore United Kingdom basins: including the Wessex/Dorset, Bristol Channel, Cleveland/Yorkshire and Hebrides basins; offshore North Sea; and mid-Norway basins). Three to five 2nd-order transgressive/regressive (T/R) facies cycles have been identified within these basins recording slightly different timing according to the local tectonic and sedimentary settings (Fig. 2). One cycle is dated as Hettangian to mid-Pliensbachian in age (T/R cycle 4) in the Alpine area and two cycles (T/R cycle 4a and T/R cycle 4b) are related to the same period in the other basins studied; one is dated as Pliensbachian pars (T/R cycles n°5) and one is dated as Toarcian-Aalenian (T/R cycle 6). The numbers (4), (5) and (6) refer to the T/R facies cycles on our chart. The T/R facies cycles are described and compared below in the following order: (a) the southeastern France basin coinciding with the present day folded, Tethyan margin of southwestern Europe, (b) Paris Basin.

FIG. 1.—Outcrops of Early Jurassic sediments in western Europe and study areas (from Dean et al., 1961).
and borders, (c) onshore United Kingdom and (d) the North Sea basins (Fig. 2).

Even though 3rd-order sequences can be identified and mapped over the mid-North Sea and mid-Norway areas, the lower liassic successions of these northern areas cannot be dated accurately enough to be compared with confidence to any relative sea-level chart based on other successions. The only horizons that are well calibrated are the peak transgressions. Nevertheless, using physical criteria such as the stratal patterns and their rank within 2nd-order cycles, we have defined the most probable framework for 3rd-order depositional sequences that are the building blocks of 2nd-order cycles.

In other areas such as the onshore United Kingdom and southward to the Tethyan margin, abundant and well-preserved ammonites provide precise ages at subzone level for most of the stratigraphic cycles. The average duration of a Liassic ammonite zone is 1.14 Ma and a subzone is 0.40 Ma. Palynological determinations allow biostratigraphic correlation between southern and northern areas, because the northern successions rarely contain ammonites.

SECOND-ORDER CYCLE 4: MAIN FEATURES

In Tethyan (Subalpine) areas, there is only one 2nd-order transgressive/regressive cycle which ranges from Rhaetian to early Pliensbachian in age. Cycle 4 was probably a consequence of crustal extension and subsidence events that occurred episodically during Rhaetian to Early Sinemurian times (Dumont, this volume). The transgressive and regressive parts are equally well developed. The regressive part is particularly thick in the half-grabens of the Tethyan marginal areas.

In the Paris Basin and more northern regions that experienced Atlantic and North Sea rift system development, an additional
2nd-order regression is evident close to the Lower/Upper Sinemurian boundary (upper part of the Caenisites turneri ammonite Zone). Thus two 2nd-order cycles, numbered 4a and 4b are observed in the northern part of Europe and are collectively equivalent in age to the whole Tethyan cycle 4. The age of the lower boundary of 2nd-order T/R facies cycle 4a is approximately the same in the Anglo-Paris Basin, but the ages of the lower and upper boundaries of T/R cycle 4b differ according to location.

The earliest deposits (Hettangian) of cycle 4 are predominantly carbonates up to the latitude of the Hebrides and follow the progressive encroachment of the marine domain onto the previously emergent areas. In contrast, the Sinemurian is marked by a southward migration of the northern boundary of the carbonate facies. Well-developed uppermost Sinemurian and lower Pliensbachian limestones are limited to the Aquitaine and Tethyan areas.

**Second-Order Cycle 4, Tethyan Margin: Rhaetian to Early Pliensbachian Times**

**Main Characteristics.**

The lower boundary is dated as Late Norian (211 Ma; see discussion in Jacquin and Graciansky, this volume). The upper boundary is dated as early Pliensbachian (mid-T. ibex Zone: 193 Ma; top LST P12). The duration is 18 my.

**Transgressive Phase.**

In the internal Alpine zones (Brianconnais and Piemontese), the onset of transgression is marked by the rapid gradation from...
the intertidal/supratidal Norian Hautpodolit to the backstepping intertidal to shallow subtidal Rhaetavcula beds (Dumont, this volume). In the Subalpine (external) zones, Rhaetavcula beds with similar lithologies overlie Late Triassic subkba type evaporites and sandstones (Mégard-Galli and Faure, 1988). The next phase of the transgression is marked by the invasion of the platforms by Psiloceras planorbis faunas. This is indicative of both deepening and opening of the marine environment during the earliest part of the Hettangian stage. In proximal areas of the Tethyan realm, the Hettangian succession comprises aggrading dolomitc platforms or tidal sands which are equivalent to the ammonite-bearing P. planorbis, Alsatites liasicus and Schlotheimia angulata beds of more basinal areas (Mouterde et al., 1980). The P. planorbis beds are not known further west than the Ardèche (= eastern) border of the French Massif Central in the Rhône valley areas as it is in the Anglo-Paris Basin. The following episode of the transgression is marked by the deposition of the Gryphaeabearing limestones which are widespread in the Subalpine domain.

**Peak Transgression.**

This is dated as Caenisites turneri ammonite Zone, C. turneri/bordoti subzonal boundary, corresponding to the maximum flooding surface of sequence S1. In areas marginal to the Tethys, the extensional phase occurred around the Early/Late Sinemurian boundary (Coadou and Beaudoin, 1974; Lemoine et al., 1986), and the subsequent phase of erosion on elevated parts of tilted blocks eliminated a part of the stratigraphic record. Hence, it is difficult to assign a precise age to the peak transgression. Nevertheless, an age of Caenisites turneri or Asteroceras obtusum Zones can be proposed (MFS S1 or S2; Graciansky et al., 1990). This took place during a period of overall subsidence linked with the Tethyan rifting.

**Maximum Regression.**

The regressive part of cycle 4 in the Subalpine zones (Tethyan margin area) is recorded by widespread layers of well-cemented cherty crinoidal bioclastic limestones. They are progradational units with internal cross-bedding and erosional surfaces, and they grade to fine-grained lithologies in basinal areas. The entire succession is dated Late Sinemurian to Early Pliensbachian (late Asteroceras obtusum through early Tragophylloceras ibex ammonite Zones; Corna et al., 1990; Graciansky et al., 1993).

**Second-Order Cycle 4a, Paris Basin and Onshore United Kingdom: Rhaetian-Hettangian-Early Sinemurian Times**

**Main Characteristics.**

The lower boundary is dated as Upper Norian (211 Ma; see discussion in Jacquin and Graciansky, this volume). The upper boundary is dated as Lower/Upper Sinemurian boundary (199 Ma) in the United Kingdom and middle Lower Sinemurian (Agassiceras scipionianum/Enagassiceras saaezianum subzonal boundary of the Arnioceras semicostatum ammonite Zone; 200 Ma) in the Paris Basin. The duration is 11 to 12 my.

**Transgressive Phase.**

In the Wessex and Paris Basins, the turning point from continental to marine deposits in the uppermost Norian strata is marked by the gradation of coastal plain dolomitic shales to shallow marine backstepping sandstones and rare bioclastic limestones with Rhaetavcula, deposited in intertidal to shallow subtidal marine conditions (Goggan et al., this volume). In the Paris Basin small hydrocarbon accumulations have been found locally in the Rhaetian backstepping sandstones (Morlot, 1986). Their fluvial lateral equivalents have been used for gas storage (Leblanc et al., 1991).

In the Netherlands, the fossiliferous marine Sleen Shale of the Altena Group is dated as Rhaetian from palynological data. It separates the red anhydritic beds of the Keuper stage from overlying Early Jurassic argillaceous limestones (Nederlandse, 1980).

The next step in the transgression is marked by ammonite-bearing P. planorbis beds indicating earliest Hettangian age and the first arrival of open-marine environments onto the post-Hercynian platform. In the United Kingdom area, these are not known from further north than the Cleveland Basin and locally in the Hebrides (Oates, 1978; Bradshaw et al., 1992; Fyfe et al., 1993), nor in the Paris Basin west of its center part (Mouterde et al., 1980; Poujol, 1960).

**Peak Transgression.**

The peak transgression is dated as mid-Alsatites liasicus or Arnioceras semicostatum ammonite Zones, depending on the location. In the eastern Paris Basin a peak transgression strongly expressed on well logs is dated A. liasicus Subzone of the A. liasicus zone belonging to depositional sequence He2. It is dated by ammonites recovered in several wells within calcareous shales towards the base of the Calcaires à Gryphées (Graciansky et al., this volume) and corresponds to the drowning of a carbonate platform dated as Hettangian age in the eastern Paris Basin (Champagne area; Poujol, 1960). In the west-central Paris Basin, the stratigraphic pattern as recorded in the well logs, together with the areal extension of the Pararoniceras charlesi (Coroniceras lyra) Subzone (Arnioceras semicostatum ammonite Zone) recovered in boreholes (Graciansky et al., this volume), suggest that the uppermost Pararoniceras charlesi Subzone is a good candidate for the 2nd-order peak transgression that is identical in age to the peak transgression in United Kingdom. The peak transgression in onshore United Kingdom occurred within the Coroniceras lyra (= Agassiceras scipionianum) Subzone of the Arnioceras semicostatum Zone (middle Early Sinemurian; MFS of sequence S1). It is recorded in the Dorset section at the boundary between limestone-marl rhythms of the Blue Lias and the “Shales with Beef” (Hesselbo and Jenkyns, this volume).

**Regressive Phase.**

In the Paris Basin, the regressive phase is marked by the widespread development of carbonate facies (Calcaires à Gryphées). In the United Kingdom the return to siliciclastic shales and silts (e.g., at Robin Hoods Bay in the Yorkshire section; Hesselbo and Jenkyns, 1995) follows the peak transgression.

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Second-Order Cycle 4a, North Sea Basins: Rhaetian-Hettangian-Early Sinemurian (Pars) Times

Lower Unconformity.—

The lower Rhaetian beds, dated by palynology (Jan du Chêne, unpublished data, 1995), are separated from the underlying Triassic siliciclastics by a major regional unconformity. See the discussion in Jacquin and Graciansky (this volume).

Transgressive Phase.—

In the North Sea and mid-Norway areas, mixed flood plain and coastal lagoonal deposits succeed the Norian sand-prone fluvialite deposits (Steel and Rysseth, 1990). This change corresponds to the lower boundary of the Statfjord Formation that is widespread and uniform over the North Sea. In the mid-Norway area the environments were still restricted during the transgressive phase. There, the Aare Formation is characterized by braided alluvial plain deposits evolving upwards, at the peak transgression, into swamp and delta plain deposits with interbedded coal seams and marine influenced layers. Unfortunately dating in these boreal areas is no more precise than stage level.

Peak Transgression.—

In the North Sea, the peak transgression is located in the most shale-prone and coal-bearing level of the Statfjord group (Raude Formation, Steel, 1993) and can be palynologically dated as Hettangian age but with no more precision than it is related to the Coromoceras lyra Subzone (Jan du Chêne, unpublished data, 1995)

Regressive Phase.—

In the North Sea, one the main characteristics of this regressive phase is the widespread extent of continental siliciclastic terrigenous facies (lower part of the Eriksson Member of the Statfjord Formation in the North Sea and upper part of the Are Formation in the Norwegian basins). The fluvialite sand bodies become progressively more amalgamated towards the maximum regression which is marked by the most basinward shift of fluvial deposition (Steel, 1993; Parkinson and Hines, in press). The most significant oil discovery in this regressive phase is the Snorre field in the Tampen Spur area (northern North Sea) on the crest of the Snorre tilted-block escarpment (Karlsson, 1986).

Second-Order Cycle 4b, Paris Basin: Late Sinemurian-Early Pliensbachian Times

Main Features.—

The 2nd-order cycle 4b of the Paris Basin is highly asymmetrical in many places, with a short, thin backstepping phase and a thick regressive phase comprising infilling and then restreepting depositional sequences developed in response to lower Pliensbachian block faulting and tilting. The asymmetry is observed also in the North Sea area, even though the North Sea had not yet reached a true rifting stage at this time. The lower boundary is dated as middle-early Sinemurian (Acanthopleuroceras obtusum Zone; intra-Acanthopleuroceras valdani Subzone: 193 Ma; top LST P12). The duration is 7 my.

Lower Boundary.—

An unconformity in the Paris Basin separates the underlying regressive Calcaires à Gryphées (Lower Sinemurian) from the overlying Argiles à Promicroceras (mainly Upper Sinemurian). This sharp contact is erosional on uplifted fault blocks, such as at the Vert le Grand well (14 km south of Paris). In Burgundy, in the Seigm area (locus typicus for the Sinemurian Stage) the unconformity is marked by an extreme attenuation of the Upper Sinemurian with 1-m-thick nodular limestones associated with phosphate and glauconite (Dommergues, 1993). This marked condensation can be related to the “wave-base razor” effects of tempests on submarine shallow highs where bioclastic sediment was generated by organic production (Mettraux and Home-wood, 1994).

Transgressive Phase.—

This transgressive phase had a short duration that marks the asymmetry of this cycle. It gave birth to attenuated sequences dated as Asteroceras obtusum Zone (early Late Sinemurian). These correspond to the transition between the regressive Calcaires à Gryphées to the transgressive marlstones named Argiles à Promicroceras in Lorraine (Hanzo and Espitalié, 1994).

Peak Transgression.—

The peak transgression is dated as late Sinemurian Asteroceras obtusum Zone, Asteroceras stellare or Eoarietites denotatus Subzones (MFS S13). The peak transgression is marked by a prominent gamma ray peak dated as A. stellare Subzone (Asteroceras obtusum Zone; MFS S13) not far above the top of the Calcaires à Gryphées. It is a good shaly marker in the basin and its extension onto the Burgundy swell consists of crinoidal wackestone to marlstone with a high contents of phosphate, pyrite and glauconite in some places. Locally, it coalesces with the overlying and underlying 3rd-order depositional sequence boundaries (Dommergues, 1993).

Regressive Phase.—

Deposition of the Late Sinemurian (Lotharingian) marlstone named as Marnes à Promicroceras in Lorraine (eastern Paris Basin: Hanzo and Espitalié, 1994), Marnes inférieures in Normandy (northwestern Paris Basin: Berthe et al., 1961) and of the Early Pliensbachian marlstones named as Marnes à Numismalis in Lorraine (Hanzo and Espitalié, 1994) comprise the regressive phase. In Lorraine, the interbedded Calcaires Ocreux dated Late Sinemurian (mainly of the Echioceras raricostatum Zone; Guérin et al., 1961) show attenuated sequences and internal hiatal surfaces at the approach of the maximum regression. Such features can be reconstructed also by correlation of well logs in Lorraine wells (e.g., Courdemange, Heitz le Hutier, Trois Fontaines; Graciansky et al., this volume) and also shown in outcrops in Burgundy (Dommergues, 1993).

Second-Order Cycle 4b, United Kingdom: Late Sinemurian Main Characteristics.—

The lower boundary is dated as Early/Late Sinemurian boundary (i.e., Caenisites turneri/Asteroceras obtusum ammo-
nate zonal boundary, 199 Ma, Sb Si3). It corresponds to a maximum regression at the top of a sandy interval interpreted as a 3rd-order lowstand systems tract in the Hebrides, Morvern, Yorkshire and Dorset sections; there is no unconformity along this surface in the more basinal settings (Hesselbo and Jenkyns, 1995 and this volume). The upper boundary is dated as latest Sinemurian (Echioceras raricostatum Zone, 194 Ma; Sb Si4). The duration is 5 my.

Transgressive Phase.—

In the Dorset coast section, the transgressive phase is thin and is intersected by an erosional surface located at the Coinstone level (Hesselbo and Palmer, 1992). The transgressive phase corresponds to the aggradational part of the Sinemurian Shales in the Yorkshire section; the aggradational character being suggested by the stratatal patterns as visible on the section drawn by Hesselbo and Jenkyns (1995).

In the North Sea, the transgressive deposits correspond to the Nansen marine backstepping sandstones and to their thin marine shale equivalents in the uppermost part of the Statfjord Formation (Steel, 1993). Microfaunas from mudstones within the Nansen sandstones suggest an age within the range of Arnioceras semicostatum to early Echioceras raricostatum ammonite Zones (Partington et al., 1993).

Peak Transgression.—

This peak transgression is dated as Late Sinemurian Asteroceras obtusum/Oxynoticeras oxynotum Zones boundary (MFS of sequence Si3). In the Yorkshire and Western Scotland (Hebrides) sections the peak transgression is located within Siliceous Shales and Pabay Bed respectively, dated as Asteroceras obtusum/Oxynoticeras oxynotum zonal boundary (MFS Si3; Hesselbo and Jenkyns, Fig. 7, this volume). In Dorset, the peak transgression coincides with a hialtal surface (Coinstone); the late Asteroceras obtusum Zone and the entire Oxynoticeras oxynotum Zone are missing (Hesselbo and Jenkyns, 1995, this volume).

Regressive Phase.—

In the Hebrides Basin, the regressive phase is represented by the thickening-upward part of sandstones related to the Pabay Beds. This corresponds to the upper part of the Siliceous Shales in Yorkshire and to the upper part of the Black Ven Marls in Dorset. All are dated as Echioceras raricostatum Zone of the Late Sinemurian (Hesselbo and Jenkyns, Fig. 10, this volume). Thus the regressive phase was much shorter in duration in the United Kingdom than in more southern European areas.

Second-Order Cycle 4b, North Sea Basins: Early Sinemurian (Pars) to Early Pliensbachian (Pars) Times

Main Features.—

The lower boundary is dated as Early Sinemurian, Arnioceras semicostatum Zone (200 Ma, Sb Si2). The upper boundary is dated as uppermost Lower Pliensbachian (= Carixian; Productiloceras davoei Zone, 192 Ma; Sb P14). The duration is 8 MA.

Lower Boundary.—

This lower boundary coincides with the lower boundary of depositional sequence Si2. It is marked by a change within the upper part of the Statfjord Formation of the North Sea and within the Are Group of the Norwegian Sea. It is expressed by a sharp gradation from progradational alluvial systems to marine backstepping sandstones. Several ravinement surfaces with linked hiatuses occur. It is a major boundary but less so than the overlying and underlying ones; there is neither a visible angular relationship between the strata nor tectonically enhanced erosion as seen in the areas affected by the Tethyan rifting.

Peak Transgression.—

This transgression coincides with the MFS of the depositional sequence Si3 and is a major drowning event in the area. It is dated as Late Sinemurian, around the Asteroceras obtusum Zone/Oxynoticeras oxynotum Zone boundary, as in the United Kingdom.

In the northern North Sea, the transgression culminated in the lowermost part of the Amundsens Formation. Offshore marls and siltstones are dated by palynological associations within the Echioceras raricostatum ammonite Zone. The lateral equivalents may extend as far as the Norwegian hinterland (Steel, 1993). In the mid-Norway Basin, the peak transgression is taken to be a through-going mudstone interval from the lowermost part of the Tilje Formation. It has been recognized in most wells on Halten Terrace but is unfortunately poorly dated (Dalland et al., 1988).

Regressive Phase.—

On the Horda Platform of the northern North Sea, the regressive phase corresponds to progradational shoreface sandstones of the Johansen Formation that interfinger with the offshore marine shales of the Amundsens Formation (Steel, 1993). The sandstones produced by that regressive phase do not extend a great distance basinward from the margin. Frequent erosional surfaces are linked with the regression on both landward and intrabasinal structural highs. Analogous deposits in the Norwegian Sea belong to the Tilje Formation and comprise recurrent alluvial plain deposits with some coal seams (Dalland et al., 1988). There is no known oil discovery linked with these regressive deposits.

SECOND-ORDER CYCLE 5

Main Features

The Late Pliensbachian 2nd-order cycle 5 is asymmetrical everywhere. For example, in the Yorkshire section within the Cleveland Basin or at various places within the central Paris Basin, the regressive phase is up to an order of magnitude thicker than the underlying transgressive one. The transgressive phase was short and thin with widespread backstepping deposits. It does not contain 3rd-order aggrading sequences. This is the consequence of an episode of extensional tectonics and of subsequent rapid subsidence that affected most of the southern part of the European craton and the northern Tethyan margin. The carbonate belt continued to move southward with respect to the previous cycles, and at the end of the cycle its northern boundary reached the latitude of the northern Aquitaine Basin.

Intracratonic basin sedimentation is everywhere characterized by fine-grained siliciclastics, shales and siltstones during the infilling phase, including the northern Tethyan margin. Dur-
ing the forestepping phase, relatively coarse-grained, prograding sediments extended further seaward than during the regressive phase of the previous 2nd-order cycle. This occurred from the Boreal to the Tethyan realms (in the siliciclastic belt from the North Sea, onshore United Kingdom, and Paris Basins, as well as in the calcareous belt in Aquitaine and the Tethyan margins).

The ages of the cycle boundaries and peak transgression differ according to the location, which shows again that the tectonic development was different from the north to the south of the European continent. Cycle 5 began earlier and had a longer duration in the British Isles (Hesselbo and Jenkyns, this volume) than in the Tethyan domain (Graciansky et al., 1993). This records the development of the North Atlantic rift system in northern Europe.

Second-Order Cycle 5, Subalpine Zone, Eastern Aquitaine Basin and Paris Basin: Pliensbachian

Main Features.—

The lower boundary of the 2nd-order cycle 5 is dated as Pliensbachian, Tragophylloceras ibex Zone, Acanthopleuroceras valdani Subzone (194 Ma; top LST of depositional sequence P12). The lower boundary is dated as uppermost Pliensbachian, Amaltheus spinatus Zone, Pleuroceras hawskerense Subzone (190 Ma; SbP18). The duration is 4 my.

Lower Boundary.—

An erosion surface dated as the Acanthopleuroceras valdani Subzone intersects the cherty crinoidal limestone belonging of the previous regressive phase on major tilted blocks in the Digne area of the Subalpine-Tethyan margin-area (Dommergues, unpublished data, 1994). In a similar manner in the Paris Basin, a major erosional surface dated as the Tragophylloceras ibex Zone (Acanthopleuroceras valdani Subzone) is recorded on both its eastern and western borders from outcrop and subsurface data. In Lorraine (eastern Paris Basin), the Calcaires à Davoei closely overlies the erosional surface that intersects the Marnes à Numismalis and the Calcaire Ocreux (Courdemange, Heitz le Hutier and Trois Fontaines wells; Graciansky et al., this volume; Guérin et al., 1961). In Burgundy (southeastern Paris Basin), a hiatal (erosional?) surface is dated as early T. ibex Zone.

Transgressive Phase.—

In the Digne area of the Subalpine zone (Tethyan margin), a set of condensed sequences in the uppermost lower Pliensbachian (Prodactylioceras davoii and most part of Tragophylloceras ibex ammonite Zones) represents the transgressive phase (Dommergues, unpublished data, 1994). Relatively attenuated sequences of interbedded marlstone, mudstone and bioclastic limestone of the same age play the same role in the Grands Causses (Meister, 1989) of the southern French Massif Central.

From our well-log correlations in the Paris Basin, an unnamed marly interval related to the Beauinoceras luridum Subzone of the T. ibex Zone (overlain by the Calcaire à Davoei in Lorraine; Hanzo et al., 1994) sits on the regional unconformity and is a marker in both the basinal and shelfal areas. It is less than 10 m thick. Crinoidal wackestones to packstones are the time equivalent of this marker in the outcrops of the Burgundy swell (Dommergues, 1987).

Peak Transgression.—

In the Subalpine (Tethyan margin) area, the peak transgression is located towards the top of attenuated depositional sequences of bioturbated cherty and crinoidal limestones. These yield ammonites of the Beaniceras luridum to early Amaltheus stokesi Subzones (Turin et al., 1984). The peak transgression probably corresponds to the uppermost part of this condensed interval (i.e., to the Stokesi Subzone that is recorded in the Monestier horizons in the Grands Causses; Meister, 1989).

In the Paris Basin, the peak transgression is located at the transition between alternating marlstones and wackestone (named as Calcaires à Davoei in Lorraine; Hanzo and Espitalié, 1994) and the infilling marly sequences dated as the uppermost Prodactylioceras davoii Zone (upper Aegoceras capricornus and Oistoceras figitum Subzones) and the Amaltheus margaritatus Zone. It can be recognized from a characteristic well-log signature on both the basin and swell areas. It has been dated by ammonites in several wells (Poujol, 1960; Graciansky et al., this volume). It corresponds to a crinoidal limestone sitting on the Burgundy High and dated as the Aegoceras capricornus Subzone (MFS P13; Dommergues, 1987).

Regressive Phase.—

In many areas of the Subalpine zones (Tethyan marginal domain), there are crinoidal cherty limestones including characteristic large pectinids. These constitute the Barre à Pecten in western Aquitaine (Quercy: Rey and Cubaynes, 1991) and the Calcaires du Domérien in the Subalpine area (Graciansky et al., 1993). They all prograde towards basinal areas where they grade into silty shales.

In the Paris Basin, eastern Aquitaine (Quercy), Causses, and Subalpine (Tethyan margin) areas, the deposition of black siltstones, commonly micaceous shales is a typical feature of infilling 3rd-order sequences, which belong mainly to the Amaltheus margaritatus ammonite Zone. The forestepping sequences are dated principally as Amaltheus spinatum Zone. They range from progradational siltstones to fine grained sandstones in the shallower parts of the Paris Basin and are named as Grès médioliasiques in Lorraine (Hanzo et al., 1993) and Banc de Roc in Normandy (Berthe et al., 1961).

Second-Order Cycle 5, United Kingdom; Uppermost Sinemurian-Pliensbachian (Pars)

Main Features.—

The lower boundary is dated as uppermost Sinemurian (Echioceras raricostatum ammonite Zone; 195 Ma; top LST Si4). The lower boundary is dated as Pliensbachian (Amaltheus margaritatus Zone, Amaltheus stokesi Subzone, 191 Ma; Sb P1 5). The duration is 4 my.

Lower Boundary.—

In Yorkshire, the sands of Landing Scar that separate the Siliceous Shales from the Pyritous Shales rests upon the sequence-boundary surface and are dated as the Paltechioceras aplanatum Subzone of the Echioceras raricostatum Zone. On Pabay and Skye (Hebrides Basin), sands belonging to the up-
permmost *E. aricostatum* Zone record the maximum regression at the base. It is well marked by a sharp surface on the Dorset coast (Wessex Basin) along the “Hummocky bed” that separates the Black Ven Marls from the Belemnite marls, with *Leptechioceras macdonelli* and *Paltechioceras aplanatum* Subzones missing. This hiatus at the Hummocky Bed may be caused by shallow submarine erosion or, alternatively, by sediment starvation related to the subsequent deepening.

**Transgressive Phase.**—

The Pyritous Shales in Yorkshire (North Cheek, Robin Hood’s Bay), and time equivalents in the Hebrides Basin (Pabay and Raasay sections) and the Belemnite marls in the Dorset section record a very rapid, short-term transgression. This is accompanied by the development of high-organic-matter content (Van Buchem and McCave, 1989; Van Buchem et al., 1992).

**Peak Transgression.**—

In Hebrides, Cleveland and Wessex basins, the age of the peak transgression is earliest Pliensbachian, *Phricodoceras taylori* ammonite Subzone (MFS St5). The organic-rich Pyritous shales dated as *P. taylori* Subzone of the *Uptonia jamesoni* Zone in the relatively proximal Yorkshire section. In the basinal Dorset section, an erosional hiatus caused by shallowing at the top of the underlying depositional sequence was followed by sediment starvation, which is represented by the Hummocky Limestone.

**Regressive Phase.**—

Thickening and coarsening-upward sandy layers in the Hebrides Basin (Pabay), Yorkshire (Ironstone Shales) and Dorset coast (Belemnite Marls overlain by the Green Ammonite Beds) record this regressive phase (Hesselbo and Jenkyns, 1995, in prep.). In both the Yorkshire and Dorset coastal sections, a surface dated as the *Acanthopleuroceras validani* Subzone (i.e., the 2nd-order peak regression in the Paris Basin and Subalpine areas), corresponds to a well-marked change in lithology and stratigraphic pattern. A hiatus, at which the *Tropidoceras masseanum* and part of the *A. validani* Subzones (lower *Tragophylloceras ibex* Zone) are missing, is shown from subsurface data in the Cleveland Basin (Felixkirk borehole: Ivimey-Cook and Powell, 1991) and on the Market Weighton Block (Gaunt et al., 1980). The hiatal surface may be a good 3rd-order sequence boundary but it is not considered by Hesselbo and Jenkyns (this volume) to be a 2nd-order maximum regression, as it is in more southern areas.

Regression culminated with the prograding Staithes Sandstone in Yorkshire and with the prograding Third Tiers, Eype Clay and Downcliff Sands in Dorset, dated as *Amaltheus stokesi* Subzone. In the Humber boreholes of the Market Weighton Block, an erosional hiatus is dated by Gaunt et al. (1980) as *Oistoceras figulinum* through *Amaltheus gibbosus* subs of the *Prodactylioceras davoei* and *Amaltheus margaritatus* Zones. The maximum regression in the Dorset is probably slightly later and coincide with the Thorncombe Sands (i.e., *Amaltheus subnodosus* Subzone). The maximum regression in the Pabay and Raasay section of the Hebrides Basin is marked by a hiatal surface that separates sandstones dated as *Prodactylioceras davoei Zone* (*Aegoceras capicornus* Subzone) from hummocky calcareous levels dated as *Amaltheus stokesi* Subzone.

**Remark.**—

The upper Pliensbachian part of the Scalpa Sandstone of the Hebrides section may constitute a good T/R facies cycle, as drawn by Hesselbo and Jenkyns (this volume, Fig 7). The condensed sandstone at the Davoei/Margaritatus boundary and the uppermost Scalpa Sandstone (close to the Pliensbachian/Toarcian boundary) may represent the lower and the upper boundaries of this T/R cycle. With such an interpretation, the upper boundary of the Scalpa Sandstone T/R cycle of the Hebrides marks the end of a forced regression that is known at this age in most European basins. It may be coeval with the upper boundary of the Late Pliensbachian T/R cycle 5 of the Paris, Eastern Aquitaine and Subalpine basins. The point of uncertainty raised by Hesselbo and Jenkyns (this volume) is whether the condensed sandstone at the Davoei/Margaritatus boundary records maximum or minimum accommodation space.

**Second-Order Cycle 5, North Sea: Mainly Late Pliensbachian Main Features.**—

The duration of cycle 5 was 2 my only, that is small as compared with other cycles. It was bounded by two well-marked surfaces, the lower one being strongly erosional. The cycle itself was asymmetrical, with a short transgressive phase relative to the regressive. The lower boundary is dated as late Early Pliensbachian (= Carixian; *Prodactylioceras davoei* ammonite Zone; 192 Ma; lower boundary of sequence P14). The lower boundary is dated as latest Pliensbachian (= Domerian; *Amaltheus margaritatus* Zone, *A. stokesi* Subzone, 190 Ma; lower boundary of sequence P18).

**Lower Boundary.**—

This boundary is a major erosional surface in the North Sea cutting across the Amundsen open marine shales (Tampen Spur area) and the prograding Johansen sandstone on the Horda Platform (T. Jacquin, unpublished data, 1995). It is observed even on top of the intrabasinal structural highs.

**Transgressive Phase.**—

The transgressive phase of the Johansen Formation corresponds to near-shore shelf sandstones that progressively backstep landward and interfinger with marine shales of the lowermost Burton Formation. In the mid-Norway basin, the transgressive deposits comprise part of the Tilje Formation and form a thin, tidal to shoreface interval intercalated with coastal plain deposits. Age control is poor.

**Peak Transgression.**—

This transgression-is dated as lower Uplkr Pliensbachian (= Lower Domerian; base *Amaltheus margaritatus* Zone). The peak transgression is located at a gamma ray peak at the base of Burton Formation (northern North Sea) and Aalburg Shales (Central Graben and Danish and Dutch basins) and can be dated by palynological associations related to the base of the *Amaltheus margaritatus* Zone (P15 MFS; Partington et al., 1993). In the Mid-Norway basin, the base *A. margaritatus* bio-event is also recorded.
Regressive Phase.—

The regressive phase corresponds to the progradational part of the Cook Sandstones, which are mainly near-shore to shoreface deposits derived from erosion of the Norwegian hinterland (Steel, 1993). During the forstepping phase, they covered wide areas North Sea tens of kilometers away from the basin margins (Livbjerg and Mjøs, 1989). The sandstones are related to an erosional surface recorded in a proximal landward location but never on top of the intrabasinal structural highs, such as during the preceding 2nd-order regressive phase. Biostratigraphic data indicate late Pliensbachian ages with no more precision. These sandstones can locally be good reservoirs with significant oil discoveries such as the Gullfaks field and the Statfjord field on the Tampen Spur (Spencer et al., 1986).

SECOND-ORDER CYCLE 6: LATE PLIENSBACHIAN-TOARCIAN-AALENIAN:

Main Features

Cycle 6 is highly asymmetrical in many places, especially in southern Europe. The transgressive phase is generally of 2nd-order peak transgression, within a starved interval, characterized by abundant authigenic minerals (such as glauconite, phosphate and/or ferruginous oolites) on intrabasinal structural highs and in distal shelf areas. In basin areas, the time-equivalent deposits are organic-rich shales known as Posidonien Schifer, Paper Shales or Schistes Carton (Jenkyns, 1984, 1988). The cycle corresponds to both a 1st- and 2nd-order peak transgression.

The regressive phase is a thick interval with shale-prone infilling depositional sequences overlain by thick forstepping sequences in the basin areas. Forstepping sequences are either sand-prone (Brent province) or calcareous (Jura platform and Subalpine basin margin) or shale and mudstone prograding deposits (in the Subalpine basin itself).

During the latest stage of the regression (i.e., during the forstepping phase), foramol-type carbonate factories were established as far north as the London-Brabant massif. Extensional tectonic activity led to block faulting, tilting and uplift, major characteristics of the regressive phases in all European basins. In the North Sea this has been interpreted as the time of the prerift thermal doming (Ziegler, 1988, 1989; Underhill and Partington, 1993), whereas in the Alpine area it is one of the major episodes of the Tethyan rifting (Lemoine et al., 1980). In areas with the most subsidence such as the La Javie-Digne area, the earliest Toarcian faunas are Dactylioceratidae belonging to the Dactylioceras semicelatum Zone of the upper Dactylioceras tenuicostatum Zone. This would imply that the lower part of the Tenuicostatum Zone (i.e., sequence P18 of our chart) is missing or that it cannot be identified for paleoecological reasons. The earliest transgressive layers are alternating marlstones and calcareous mudstones dated as Dactylioceras semicelatum, Harpoceras strangewayi and H. pseudoserpentinus Subzones (Mouterde et al., 1980). The shales related to the H. falciferum and H. sublevisoni Subzones are not clearly identified in the subsiding Digne area since the organic-rich Schistes Carton has undergone alpine recrystallisation (Gracianky et al., 1994). The organic-rich Schistes Carton are well developed in more external areas (Mouterde et al., 1980).

In eastern Aquitaine (Quercy: Cubaynes, 1986), the transgressive phase is recorded only by encrusted surfaces dated as Dactylioceras tenuicostatum Zone overlying the Amaltheus spinatun (P. hawskerense) strongly regressive limestones. These are overlain by organic-rich Schistes Carton dated as Harpoceras serpentinum Zone in the Aquitaine subsiding areas. The black shales grade laterally onto the submarine highs to calcareous and ferruginous oolites dated as H. serpentinum to lower Hildoceratids bifrons Zones (Rey and Cubaynes, 1991).

In the Paris Basin, the earliest stage of the transgression, dated as Tenuicostatum, is marked either by ferruginous crusts or by ferruginous oolites deposited under starved conditions from Burgundy to Poitou (southeastern to southwestern boundary of the basin) or by marlstones with variable sandy intercalations (Normandy and Lorraine: northwestern and eastern boundaries of the Paris Basin). These are overlay by a continuous blanket of organic-rich shales (Schistes Carton) dated as Serpentinum (including the Harpoceras falciferum Subzone). These shales range in thickness from 0.3 m (Argiles à Poissons in Normandy) to 50 m in Lorraine (for more details, see Mouterde et al., 1980). The early Toarcian was a time of maximum landward extent of open-marine lithologies onto the continental areas during Jurassic times. It is one of the main source rocks for the southern North Sea, German and Paris Basins (Jenkyns, 1984, 1988).

Peak Transgression.—

This peak transgression is dated as Harpoceras falciferum or H. sublevisoni Subzones, around the Lower/Middle Toarcian

Second-Order Cycle 6, Subalpine Zones, Eastern Aquitaine and Paris Basin: Late Pliensbachian-Toarcian-Aalenian

Main Features.—

The lower boundary is dated as Amaltheus spinatus Zone, Pleuroceras hawskerense ammonite Subzone of the late Pliensbachian (190 Ma, Sb P18). In the Paris Basin, eastern Aquitaine (Quercy: Rey and Cubaynes, 1991) and Subalpine (Tethyan margin) areas, a tectonically enhanced unconformity dated of the Late Pliensbachian (A. spinatus Zone, Pleuroceras hawskerense Subzone) is well expressed on uplifted or tilted blocks (Coadou and Beaujouin, 1974; Graciansky et al., 1993). The duration is 13 Ma.

Transgressive Phase.—

In the Subalpine zones, earliest Toarcian sediments are generally missing following the principal regression dated as late Amaltheus spinatun Zone (Mouterde et al., 1980). In areas with the most subsidence such as the La Javie-Digne area, the earliest Toarcian faunas are Dactylioceratidae belonging to the Dactylioceras semicelatum Subzone of the upper Dactylioceras tenuicostatum Zone. This would imply either that the lower part of the Tenuicostatum Zone (i.e., sequence P18 of our chart) is missing or that it cannot be identified for paleoecological reasons. The earliest transgressive layers are alternating marlstones and calcareous mudstones dated as Dactylioceras semicelatum, Harpoceras strangewayi and P. pseudoserpentinus Subzones (Mouterde et al., 1980). The shales related to the H. falciferum and H. sublevisoni Subzones are not clearly identified in the subsiding Digne area since the organic-rich Schistes Carton has undergone alpine recrystallisation (Gracianky et al., 1994). The organic-rich Schistes Carton are well developed in more external areas (Mouterde et al., 1980).

In eastern Aquitaine (Quercy: Cubaynes, 1986), the transgressive phase is recorded only by encrusted surfaces dated as Dactylioceras tenuicostatum Zone overlying the Amaltheus spinatun (P. hawskerense) strongly regressive limestones. These are overlain by organic-rich Schistes Carton dated as Harpoceras serpentinum Zone in the Aquitaine subsiding areas. The black shales grade laterally onto the submarine highs to calcareous and ferruginous oolites dated as H. serpentinum to lower Hildoceratids bifrons Zones (Rey and Cubaynes, 1991).
boundary, depending on the location (MFS To2 or To3). In the Digne thrust sheet of the subalpine domain the peak transgression is represented by a thin layer (20 cm) of nodular limestones dated as *Lusitanicum* horizon (upper *H. sublevisoni* Subzone of lower *Hildoceras bifrons* Zone; i.e., MFS To3). This layer overlies a discontinuous veneer of black shales that rests on the adjacent Provence carbonate platform at the south. (Gracianky et al., 1994). In Quercy (eastern Aquitaine), the peak transgression is related by Rey and Cubaynes (1991) to the Lusitanicum horizon as well but the black shale layer (Schistes Carton) is limited to the *H. strangewaysi* Subzone (lower *H. serpentinum* Zone; MFS of sequence To1 on our chart). In the Paris Basin for the peak transgression, a consensus has been reached for a *Harpoceras falciferum* Subzone age (MFS of sequence To2) corresponding to the maximum total organic carbon values in the area (Jenkyns, 1984, 1988).

**Regressive Phase.**—

In the Subalpine (Tethyan margin) area, the transect from the Provence Platform to the Digne basin clearly illustrates the three successive steps of the regressive phase that gave birth to sediments locally ten times thicker than the transgressive one. These steps were: (1) a time of starvation during the middle and early late Toarcian, which followed the peak transgression, (2) a time of infilling where several hundred meters of shales accumulated in the more basinal parts during *Dumortieria pseudoendoradiosa* Zone of Late Toarcian times (at the south of La Javie pars; in the Digne area; or near the lake of Serre Ponçon, on the high Durance valley) and (3) a time of foresteeding, as indicated by the pinching out on the basin margins of the depositional sequence Aa2 dated as late *Ludwigia murchisonae* and early *Graphoceras concavum* Zones (Graciansky et al., 1994).

In Jura and Bresse, recent grabens located at the transition between the structural high of the southern Paris Basin and the subsiding, nascent Tethyan margin. The foresteeding phase forms a large prograding carbonate platform, more than 100 m thick, resting unconformably on the Toarcian shales. Here, the truncation at the late-Cimmerian unconformity (Stille, 1924) affected all the infilling sequences and truncated part of the peak transgression beds of the lower Toarcian (Jacquin et al., this volume).

In the Paris Basin, the infilling sequences comprise mainly offshore relatively organic-poor shales with few calcareous mudstone intercalations. These are dated as middle and upper Toarcian by their ammonites and are named *Marnes Supérieures* in Normandy or partly *Marnes à Bifrons* at other places. Their thickness reaches one-third of the total thickness of the whole Lias interval, representing a probable duration of 6 my compared to the 26 my for the total cycle (Graciansky et al., this volume). This shows that the 1st- and 2nd-order peak transgression dated Early Toarcian corresponded to maximum accommodation space in the area. The foresteeding sequences are dated mainly as *Pleydellia aalensis* Zone (uppermost Toarcian) and Aalenian. On the structural highs such as the central part of the Paris Basin or on the Burgundy high (southeastern border of the Paris Basin), they are represented only by condensed ferruginous deposits representing the maximum flooding surfaces of 3rd-order depositional sequences dated *Leioceras opalinum*, *Ludwigia murchisonae* and *Graphoceras concavum* Subzones respectively (Mouterde, 1942). The same depositional sequences comprise shaley masses in more subsident areas. The entire succession is intersected by a through-going erosional surface dated uppermost Aalenian so that the Bajocian layers may rest in places on Mid-Toarcian strata (e.g., Vert le Grand well in Normandy). This surface records here the mid-Cimmerian unconformity.

**Second-Order Cycle 6, United Kingdom: Late Pliensbachian-Late Aalenian**

**Introduction.**—

British authors consider that cycle 6 can be divided into two subcycles, the lower one dated as late Pliensbachian-late Toarcian and the upper one dated as late Toarcian-late Aalenian. This idea comes mainly from the following two points: (1) the pronounced regression dated as *Pleydellia aalensis* Zone, which may define the boundary between the two subcycles and (2) the strong transgressive character of thin layers dated as *L. opalinum* and *L. murchisonae* Zones in the United Kingdom (such as the Dogger ferruginous oolites of the Cleveland Basin) corresponding to the To7, Aa1 and Aa2 transgressive systems tracts (Knox et al., 1991). The MFS of the lower subcycle is dated as *Harpoceras falciferum* Subzone as in other European areas. The MFS of the upper sub-cycle is tentatively dated as mid-*L. murchisonae* Subzone based on an interpretation of the Yellow Conglomerate in Dorset as being starvational in origin (Hessbo and Jenkyns, 1995), the occurrence of thick shallow water carbonates of this age in the Worcester graben, the presence of marine *Murchisona* Shales above the Dogger locally in Yorkshire, and a condensed *L. murchisonae-Zone* ammonitic limestone at the Duncaan Shales/Barrereraig Sandstone boundary in the Hebrides (the *Black Bed* of Barrereraig).

The interpretation by French authors is somewhat different in the sense that they consider the whole late Toarcian-Aalenian interval as a long term regression dated as middle and late Toarcian-late Aalenian (*H. falciferum* Subzone to *G. concavum* Zone), which is a common characteristic of all the European basins. This regression is induced by the late Cimmerian unconformity dated as upper *P. aalensis* Zone (top To7 LST) corresponding to the uplift of the North Sea Rift dome (Ziegler, 1988; Underhill and Partington, 1993). It is recorded by strongly prograding sediments, typical forced of regression in ramp and basinal environments. On the submarine highs, the accommodation space was nearly zero, as shown by the thinness of preserved sediments. Basin borders were characterized by erosional features. All this is well documented in the Subalpine, Paris and North Sea Basins (Graciansky et al., 1993; Mouterde et al., 1980; Steel, 1993). During this time interval, short-term but high-amplitude transgressions corresponded to the maximum flooding surfaces, To7, Aa1 and Aa2, recorded by the deposition of thin ferruginous oolite layers on the submarine highs of the Anglo-Paris Basins. These two alternative interpretations are proposed on Figure 2. The characteristics of the main cycle 6 in the United Kingdom are summarized below.

**Main Features.**—

The duration is 34 MA. The lower boundary is dated as in the lower part of the *Amaltheus stokesi* Subzone of the late Pliensbachian (191 Ma, SbP15). It is marked by the base of the Staithes Sandstone in Yorkshire and of the Three Tiers in Dorset.
(Wessex basin) and the Scalpa Sandstone in the Hebrides basin. The upper boundary cannot be dated with certainty in the Dorset Coast profile within the extremely attenuated Aalenian-Bajocian Inferior Oolite. In comparison to what observed in other areas, it may be proposed that the upper boundary coincides with the *Brasilia bradfordensis/Graphoceras concavum* ammonite zonal boundary, i.e. the Mid-dle/Late Aalenian boundary (177 Ma, top LST of sequence Aa2).

**Transgressive Phase.**—

A major flooding event dated as upper Pliensbachian and lower Toarcian covered nearly all Europe without any major aggradational shelfal interval. Carbonate shedding and terrigenous supply ceased at that time all over Europe. It is recorded for example on the Dorset Coast by the *Junction Bed*, where the *Amaltheus spinatus* Zone and the lower and middle Toarcian substages are highly condensed and incomplete (Hesselbo and Jenkyns, 1995). In the United Kingdom (with the possible exception of the Hebrides basin), there is no clearly identified 2nd-order cycle of Late Pliensbachian age as in other areas. Therefore the interval of shales and sandstones dated *Margaritatus* and *Spinatum* must be considered as occupying part of the transgressive phase of 2nd-order cycle 6a.

**Peak Transgression.**—

Highly organic shales in the Yorkshire coast mark the peak transgression. In the Hebrides basin it corresponds also to highly organic-rich shales and extremely condensed layer (Raa-say Ironstone). In the Wessex basin (Dorset coast), only an extremely condensed interval occurs (Junction bed).

**Regressive Phase.**—

In the Cleveland Basin at the Ravenscar section, the upper Toarcian comprises a prograding set of depositional sequences that include the upper part of the Whitby Mudstone Formation (Alum Shale, Peak Mudstone and Fox Cliff Siltstone Members) overlain by the Blea Wyke Sandstone Formation (Grey Sandstone and Yellow Sandstone Members; Hemingway et al., 1968; Knox et al., 1991). The lower part of the prograding sandstones and siltstones of the basal Saltwick Formation constitute the regressive phase on the Yorkshire coast section (Knox et al., 1991). Good sections in the Inferior Oolite of the Worcester graben, immediately north of the Wessex basin, indicate increased siliciclastic supply and formation of a widespread unconformity in the *Graphoceras concavum* Zone (e.g., Cope et al., 1980). In the Hebrides, the *G. concavum* Zone represents a strongly prograding phase (Morton and Hudson, 1995). Relatively small thicknesses are in contrast to the several hundred meters of prograding coeval strata in many subsiding parts of the southern subapline areas.

**Second Order Cycle 6, North Sea and Mid-Norway Basins: Latermost Pliensbachian-Toarcian-Aalenian**

**Main Features.**—

The lower boundary is dated as late *A. margaritatus* Zone of the Upper Pliensbachian (SpP18 of our chart), within the Cook Formation of the Dunlin Group (distal) or Orrin Sands (proximal); Partington et al., (1993). The upper boundary is dated as Late Aalenian (probably top LST of sequence Aa2; base *G. concavum* Zone: 177 Ma). This upper boundary corresponds to the maximum progradation of the Brent Group. The duration is around 13 my.

**Transgressive Phase.**—

Lower shoreface to shoreline backstepping sandstones and siltstones comprise the transgressive phase that belongs to the Cook Formation (Partington et al., 1993). This transgressive phase is marked in many places by lower Toarcian strata onlapping onto structural highs. For example it directly overlies the Triassic deposits of the United Kingdom side of the Viking Graben.

**Peak Transgression.**—

In the mid-Norway basins, the peak transgression is identified as a strong gamma-ray peak that is well dated by palynomorphs as *H. falciferum* Zone. It corresponds to a burrowed, dark-colored open marine shale developed at the base of the *Ror* Formation and unconformably overlying the Pliensbachian Tilhe sandstone. In the North Sea basins, the peak transgression is identified also along a major gamma ray peak within the lowermost part of the Drake Formation. It is not as well dated as in the mid-Norway basins, but it is close to the Falciferum MFS (Partington et al., 1993).

**Regressive Phase.**—

In both the mid-Norway and North Sea basins, the infilling and forestepping depositional sequences can be easily separated. The infilling sequences form large progradational intervals evolving from nearshore and shoreface environments landward (Oseberg Formation on the Horda platform) to offshore marls (Drake Formation on Tampen Spur or Ilje Formation in the mid-Norway area). Such regressive sequences do not merge towards a major erosional surface. Detrital deposits were derived from the Norwegian hinterland.

Forestepping sequences developed in response to thermal doming in the southern North Sea whose onset is dated in the mid-*P. aalensis* Zone (i.e., at the boundary of sequence To7). This coincides with the mid-Cimmerian unconformity, which is marked by widespread erosional hiatuses in northwestern Europe (Ziegler, 1988 and 1989). This induced a complete reorganization of the depocenters and sediment transport directions in the northern North Sea Brent province. They include the Broom, Rannoch, Etive and partly Ness facies belonging to the well-known deltaic Brent Group (Morton et al., 1992). In the mid-Norway basins, forestepping sequences, equivalent to the *Ilje* Formation, form shoreface progradational units still derived from the Norwegian hinterland but structurally controlled by tilted faulted blocks.

The boundary between the infilling and forestepping phases is well marked by prograding sandstones analogous and time equivalent to the Bridport, Blea Wyke and Bearreraig Sandstones of the onshore United Kingdom (Partington et al., 1993). The sandstones are considered to be the regressive part of cycle 6.

In all these areas, the maximum progradation of shoreface deltaic sandstones towards the basin is reached during latest Aalenian times at the top lowstand of sequence Aa2. This is overlain by a major (3rd-order) flooding surface dated as *G. concavum* Zone.
The Ligurian cycle has been named from the main episode of the rifting that affected the southern part of the western European craton and led to the opening of the Ligurian (= western) Tethys. This phase of rifting lasted approximately 40 my and was discontinuous. It comprised discrete phases of extensional tectonics, shown by block faulting and tilting, and subsequent phase of rapid subsidence, recorded by regionally extensive erosional surfaces in the Western Alps (Lemoine and Graciansky, 1988).

The Ligurian major transgressive/regressive cycle comprises three to five 2nd-order transgressive/regressive facies cycles. Twenty seven 3rd-order depositional sequences make the building blocks of the 2nd-order cycles. Most of the 3rd-order depositional sequences can be documented from the North Sea to southern Europe. Moreover, when precise age dating can be provided by ammonites synchronity can be established even at distances as far as from Yorkshire to the southern Subalpine zones (Graciansky et al., this volume).

The organization of the 3rd-order sequences within each of the four 2nd-order cycles follows similar patterns from the North Sea to the Tethyan margin, whether in siliciclastic or carbonate settings. The lower and middle Lias (4 and 5) 2nd-order T/R facies cycles record the long term (1st-order) progress of Early Jurassic continental encroachment. As an example, this trend is shown by the progressively recent ages of the peak transgression of cycle 4 from the north to the south. It is dated as Caenisites turneri in the Subalps.

The upper Lias regressive phase of the major Ligurian cycle is coeval from northern to southern Europe. It is marked by the late Toarcian (P. aalenensis Zone) discontinuities (SbTo7) that date the mid-Cimmerian unconformities in the Paris Basin and more northern areas and that induced the emplacement of time equivalent forestreeing sediments in the most subsiding areas. The number and the ages of the 2nd-order cycles are not the same from the north to the south. Examples are, first, the structural evolution of the Paris Basin, which is relatively time-independent from those of both the Subalpine and United Kingdom basins during the Hettangian-Sinemurian but which followed closely the Tethyan margin development during the Pliensbachian, Toarcian and Aalenian stages. Second, the upper Sinemurian T/R facies cycle boundaries are approximately coeval in the United Kingdom and North Sea areas, but the peak transgressions differ by one ammonite zone between these two areas.

The age discrepancies can be explained as recording the local tectonic development of the sedimentary basins located on the western European craton. In consequence, the subsidence events may appear to be the cause of the transgressive phase of the 2nd-order transgressive/regressive cycles. Second-order peak transgressions that have been characterized and dated in the Western Alps are strictly coeval and can be correlated with 3rd-order maximum flooding surfaces in the Paris Basin and more northern areas. This observation suggests that the phases of Tethyan rifting have been recorded at long distances from the nascent margin. A possible tectono-eustatic effect may have been created by a change in size of the ocean basin during rifting.

One of the merits of sequence stratigraphy is to provide methods for long distance correlation if done with appropriate chronostratigraphic control. As our biostratigraphic assignments are relatively accurate, another explanation for age discrepancies of characteristic surfaces between areas may be due to different interpretations of the successions rather than different histories of the accommodation space evolution with time. We hope that such comparisons between remote areas will induce new studies for solving the remaining uncertainties.

NOTE ADDED IN PROOF
Since this was prepared, lithostratigraphic revisions have been proposed for the Hettangian Lower Pliensbachian succession of the Hebrides Basin (Hesselbo et al. 1998). An important change is that strata formerly included in the Upper Broadford Beds of former usage are now incorporated within an extended Pabey Shale Formation. Strata belonging to the former Lower Broad Beds are now referred to the Broadford Formation. (Hesselbo, S. P., Oates, M. J. and Jenkyns, H. C. 1998, The Lower Lias of the Hebrides Basin: Scottish Journal of Geology, v. 34, p. 23–60.)

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DOCUmENTATION OF JURASSIC SEDIMENTARY CYCLES FROM THE MORAY FIRTH BASIN, UNITED KINGDOM NORTH SEA

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ABSTRACT: The sequence stratigraphic analysis of the Lower to Upper Jurassic (Hettangian to Kimmeridgian) succession of the Moray Firth basin synthesizes sedimentological and paleontological data from approximately 400 wells in the offshore domain with scattered outcrops along the eastern coastline of Scotland. The stratigraphic framework produced provides a consistent framework for understanding and predicting basin-wide sedimentary facies distribution.

Three hierarchies of sedimentary cycles are distinguished: (i) major transgressive/regressive (T/R) cycles (with durations between 10 and 50 my), (ii) transgressive/regressive (T/R) facies cycles (with durations between 3 and 10 my), and (iii) regressive/transgressive (R/T) cycles (with durations between 0.5 and 3 my). Two major T/R cycles are identified, which are bounded by three important surfaces of maximum regression. These boundaries, which generally correspond to regionally extensive subaerial unconformities, occur close to the Triassic/Jurassic boundary, the Lower/Middle Jurassic boundary (often termed the “Mid-Cimmerian” unconformity) and the Ryazanian/Valangian boundary (often termed the “Base Cretaceous” unconformity). Each major T/R cycle can be subdivided into component T/R facies cycles, which are also bounded by surfaces of maximum regression. Both types of cycle are synchronous across the basin, although high sediment flux locally affects the timing of peak transgression.

Difficulties exist in objectively defining and correlating most higher-frequency (so-called 3rd-order) sequence boundaries in distal depositional settings. Instead maximum flooding surfaces have proven to be the most reliable, distinctive and easy-to-date surfaces with which to define a rigorous sequence stratigraphic framework. Consequently, higher-frequency cycles are bounded by maximum flooding surfaces, are centered upon transgressive surfaces and hence are termed regressive/transgressive (R/T) cycles. Fourteen R/T cycles are defined, each of which developed simultaneously across the Moray Firth basin. The fact that these R/T cycles can be correlated from both subbasin to subbasin, and from the Inner Moray Firth to Outer Moray Firth basin, indicates that local tectonics had little effect upon controlling the timing of 3rd-order cycle development, even during the Middle and Late Jurassic Epochs when extensional rifting is known to have exerted a pronounced control upon basin development.

INTRODUCTION

The Moray Firth (MF) basin was one of the key areas for the early application of seismic and sequence stratigraphy to North Sea Jurassic geology (Vail and Todd, 1981; Vail et al., 1984). Since these pioneering studies, significant new geological information has become available which has enabled a more rigorous sequence stratigraphic subdivision to be made. It is the aim of this paper to describe in detail this newly established stratigraphic framework. By undertaking this type of study, it should become possible to assess the geographical extent and synchronicity of such stratigraphic sequences. Only then will it be possible to attempt to ascertain the primary controls upon sequence development on a regional and perhaps global scale.

Several recent studies have demonstrated the existence of correlatable sequences in the Jurassic succession of the North Sea area (e.g. Rattey and Hayward, 1993; Partington et al., 1993a, b; Stephen et al., 1993; Underhill and Partington, 1993, 1994; Sneider et al., 1995). These papers document sedimentary cycles at a variety of scales in terms of both duration and spatial distribution. This paper builds upon the previous work of Stephen et al. (1993) by classifying the various stratigraphic cycles into the hierarchical divisions as proposed by Vail et al. (1991) and by extending the study area from the Inner Moray Firth (IMF) to Outer Moray Firth (OMF) region. The emphasis of this study is to document the sequence stratigraphy of the evolving shallow marine deposystems of the MF basin. As no Volgan or younger shallow marine sediments are yet known, the study interval is restricted to the Lower Jurassic, Middle Jurassic and lower Upper Jurassic Series (Kimmeridgian sensu gallico).

Rationale

The ability to identify, correlate and accurately date key stratal surfaces that have chronostratigraphic significance is essential in any sequence stratigraphic analysis. Three such surfaces have been used to define chronostratigraphically significant cycles of sedimentation: (i) a sequence boundary to define depositional sequences (Vail et al., 1977a), (ii) a transgressive surface to define transgressive/regressive cycles (Embry, 1993), and (iii) a maximum flooding surface to define genetic stratigraphic sequences (Galloway, 1989). In this study, the type of bounding surface used depends upon whether or not it can be objectively identified, correlated and dated which depends upon the hierarchy of the sedimentary cycle being described.

Hierarchies of Sedimentary Cycles.—

The idea of an ordered hierarchy of sequences within sedimentary successions was introduced by Vail et al. (1977b). This concept has since been refined and at present six orders of sequence have been defined based upon estimates of their perceived duration (Vail et al., 1991; Jacquin and de Graciansky, this volume). Three hierarchies of sedimentary cycle are of importance in this study: (i) durations between 10 and 40 my (1st-order sub-cycles), (ii) durations of between 3 and 10 my (2nd-order cycles), and (iii) durations between 0.5 and 3 my (3rd-order cycles).

Many of the higher-frequency cycles described in this paper are not directly comparable to the 3rd-order sequences documented by other authors in this volume. This is because of the difficulty in defining and correlating all such sequences in the condensed Jurassic sedimentary record preserved in the MF basin. This problem of cycle identification is compounded in subsurface datasets because of the less precise sedimentologic and biostratigraphic information they afford. As such, the higher-frequency cycles defined herein may be composite and comprise a number of the 3rd-order cycles described from other well-exposed and well-dated Lower to Upper Jurassic sections.
Much debate has ensued regarding the best sequence stratigraphic approach to use in basin analysis (e.g. Posamentier and James, 1993). In order to provide consistency and to enable comparisons to other papers in this volume, the approach advocated by Vail et al. (1991) has been adopted in this study wherever feasible. Consequently, the boundaries of 1st-order sub-cycles, also termed major transgressive/regressive (T/R) cycles, are defined at ubiquitous erosional unconformities that coincide with the position of maximum regression of the last 2nd-order cycle. In the MF, these unconformities are generally subaerial and may be tektontically enhanced. Major T/R cycles are centered upon the position of peak transgression. Second-order cycles, also known as transgressive/regressive (T/R) facies cycles, are identified in a similar manner with the cycle boundaries being defined at the position of maximum regression of the last 3rd-order cycle. Second-order peak transgression equates to the 3rd-order maximum flooding surface of the last 3rd-order sequence.

In contrast to the two types of T/R cycle described above, the higher-frequency cycles identified in this paper are not bounded by sequence boundaries. It is recognized that the sequence boundary should theoretically be the best candidate isochronous surface with which to bound this type of cycle, primarily because it is the only key sequence stratigraphic surface thought to form independently of sediment supply (Van Wagoner et al., 1990). However, in a well-based study such as this, sequence boundaries that develop due to minor “3rd-order” relative falls in base-level are difficult to objectively recognize and correlate into the basin interior where they have little or no lithological expression (Stephen et al., 1993; Embry, 1993). As such this surface can only be reliably used for local correlation. The problem in recognizing the correlative conformity makes the accurate dating of sequence boundaries difficult. Where the sequence boundary is objectively identified, it commonly has time duration, and its frequent location at the base or within sediments that do not lend themselves to good biostratigraphic recoveries also precludes precise dating. Instead, in this study, higher-frequency cycles are bounded by regional maximum flooding surfaces. These surfaces are the most distinctive and reliable horizons on which to base a stratigraphic framework at this hierarchical scale in the North Sea (Underhill and Partington, 1993, 1994). Associated with these maximum flooding surfaces are diagnostic faunal influxes that include ammonites, dinocysts, foraminifera, radiolarians and ostracods, which enable these sedimentary cycles to be biostratigraphically well-constrained (Partington et al., 1993b). Maximum flooding surfaces form the most reliable chronostratigraphic horizons for correlating from the basin margin into the basin interior, although difficulties can exist with their recognition in upper delta-plain environments.

The highest-frequency cycles documented in this study define progradational-retrogradational couplets that are bounded by maximum flooding surfaces and their correlative hiatal surfaces. The approach followed is very similar to that of genetic stratigraphy sensu Galloway (1989). However, we consider it only possible to distinguish two geometric tracts (regressive and transgressive) in comparison with the later modifications of the Galloway (1989) model in which three geometric systems tracts are defined (Xue and Galloway, 1993). Our cycles also differ from genetic stratigraphic sequences in that they are centered upon transgressive surfaces and have therefore been termed regressive/transgressive (R/T) cycles. Sequence boundaries may coincide with the transgressive surface, but where they do not every effort has been made to locate and accurately date this horizon. The transgressive surface has not been used to bound the higher-frequency cycles documented in this paper primarily because of the potential for miscorrelation.

A variety of types of transgressive surfaces have been recognized that may be easily confused (Dermarest and Kraft, 1987), especially in the absence of outcrop or cored material.

**Geographical and Structural Setting**

The MF basin of the North Sea United Kingdom is a Mesozoic rift basin which trends east-west and stretches from the intersection of the Central and Viking Grabens in the east to the Jurassic exposures of the Scottish coastline in the west (Fig. 1A). This graben has historically been subdivided into two basins, the IMF and OMF, along a 0-mGal Bouguer anomaly contour (McQuillin et al., 1982; Fig. 1B). Both basins are delineated by a series of major normal fault systems, and as a consequence of Late Jurassic extensional tectonism each basin comprises a number of well-defined subbasins (Andrews and Brown, 1987). Thick Jurassic successions (up to 4 km) are commonly preserved in many of these half-graben, whereas on the crest of horst-blocks the Jurassic Series may be much thinner or even absent. In the western IMF Jurassic strata subcrop at sea floor as a result of Tertiary uplift and erosion (Thomson and Underhill, 1993).

The tectono-sedimentary history of the Jurassic in the MF basin is complex but can be readily subdivided into the following: (i) an Early Jurassic prerift phase, following an earlier Permian rifting event, (ii) a Middle Jurassic thermal doming phase, where the whole of the MF area was uplifted and significant volumes of sediment eroded, (iii) a Middle to Late Jurassic prerift phase, when initial deflation of this topographic high resulted in sediments passively onlapping the “Mid-Cimmerian” unconformity, and (iv) a Late Jurassic synrift phase when differential subsidence across major synsedimentary faults led to the establishment of numerous half-graben (Thomson and Underhill, 1993). The onset of accelerated rifting in early Kimmeridgian times resulted in tectonic subsidence outpacing sedimentation rates across much of the MF basin. As a consequence, the basin margin profiles changed from gently dipping ramps to shelf-break margins with the well-defined breaks in slope occurring across synsedimentary fault planes. Consequently, the sedimentary architecture of lower Kimmeridgian and younger stratigraphic cycles are generally significantly different from those developed prior to the climax of rifting (Davies et al., 1996).

**Database and Methodology**

Initial studies undertaken by Underhill (1991a, b) used a dense regional 2-D seismic grid to erect a seismic stratigraphy for the IMF basin. However, many of the stratigraphic cycles identified in this study are too thin to be resolved on seismic data. Sedimentary cycle correlation is therefore primarily based upon well-log analysis (approximately 400 wells were used in...
FIG. 1.—Location maps. (A) The North Sea rift system and the relative position of the Moray Firth graben. (B) Major tectonic elements of the Moray Firth graben and the position of the main oil-fields. (C) The four Jurassic outliers exposed on the eastern coastline of Scotland.
this study), incorporated with biostratigraphic data and sedimentary logging of cored intervals. It has been possible to correlate this subsurface dataset to the four Jurassic outcrop sections along the north-eastern Scottish coastline (Fig. 1C).

The systematic documentation of the various hierarchies of sedimentary cycle is biased towards the outcrop sections, to which the subsurface database is compared, because of the greater biostratigraphic control such sections generally provide. However, stratigraphic gaps occur within the Jurassic System onshore, most notably from the Pliensbachian to Bathonian and from the upper Oxfordian to lowermost Kimmeridgian Stages (Fig. 2). Over these intervals recourse has been made to the subsurface database, where more stratigraphically complete successions enable these intervals to be more fully investigated. Under such circumstances key sequence stratigraphic surfaces are invariably dated by microfossil recoveries (e.g., dinocysts, foraminifera, ostracods, etc.) from sidewall cores, cored sections or cuttings. In order to achieve consistency these bioevents are calibrated to standard Jurassic ammonite zonations (Cope et al., 1980a, b) using the biostratigraphic framework of Partington et al. (1993b) unless definitive evidence suggests this scheme may be locally redundant (Brealey, 1990). The biostratigraphic scheme of Partington et al. (1993b) is also utilised when microfossils have been used to date nonmarine or marginal-marine successions. Although some notable modifications to the zonation schemes of Cope et al. (1980a, b) have recently been made (e.g., Page, 1989 for the lower Callovian), the relatively poor ammonite recoveries documented from the onshore sections by Sykes (1975) have prevented these higher-resolution biostratigraphic revisions from being incorporated in this study.

No Lower Jurassic sediments are preserved in the OMF basin (Andrews and Brown, 1987). Therefore, where exposure is incomplete, offshore wells from the IMF have been used to investigate this interval. Documentation of Bathonian to middle Oxfordian sedimentary cycles has also been preferentially undertaken in the IMF basin. In this area, more detailed biostratigraphic information is available, resulting from the economic importance of this interval (n.b., the Beatrice Field, Block 11/30, is reservoired primarily in Callovian sandstones; Linsley et al., 1980). In comparison, the documentation of upper Oxfordian to Kimmeridgian sedimentary cycles is biased towards the OMF basin. This area is a prolific hydrocarbon-bearing province (Harker et al., 1993). Subsequently biostratigraphic information is good enabling detailed correlations.

RESULTS

A systematic documentation of the three hierarchies of sedimentary cycle recognized in Lower Jurassic (Hettangian) to lower Upper Jurassic (Kimmeridgian) strata of the Moray Firth region is presented below and summarized on Figure 3. Sedimentary logs are provided for the outcrop sections to illustrate the sedimentary manifestation of key sequence stratigraphic surfaces (see Fig. 4 for the legend) and to give pertinent information to allow their inspection by interested parties.

Major Transgressive/Regressive Cycle Ja

The base of this cycle is placed at an unconformity which occurs between the Dunrobin Bay and Lossiemouth Sandstone

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**Fig. 2.—Summary lithostratigraphic breakdown for the Jurassic succession encountered in the Moray Firth.**
Formations (Batten et al., 1986; Fig. 5A). Assigning a precise age to this unconformity is extremely difficult. However, the recovery of Rhaetian microfossils in the Dunrobin Bay Formation in offshore wells indicates a Rhaetian or older Triassic age. The Stotfield Cherty Rock, which occurs at the top of the Lossiemouth Sandstone Formation, is a silcrete paleosol which can be correlated into offshore wells (Naylor et al., 1989; Fig. 6). The Stotfield Cherty Rock records a significant regional phase of subaerial exposure and soil formation and the top of this unit is a major sequence boundary.

During the Hettangian and Sinemurian Ages sedimentation resumed in the IMF with marine influence generally becoming progressively more important further up-section (Batten et al., 1986; Stephen et al., 1993). Peak transgression, which corresponds to the most distal marine facies, occurs within lowermost Pliensbachian strata (Fig. 5A). Ammonite recoveries from the Lady's Walk Shale Member enable this surface to be accurately dated as intra-Phricodoceras (P.) taylori Subzone of the Uptonia (U.) jame soni Zone. The remainder of the Lower Jurassic Series can only be investigated from offshore wells (Fig. 6). This interval consists of a coarsening-upwards sand body (the Orrin Formation) formed by the progradation of a significant deltaic system. A major unconformity, often termed the "Mid-Cimmerian" unconformity (MCU; Ziegler, 1975), is pres-
ent at the top of the Orrin Formation, which progressively truncates older Jurassic strata in an eastwards direction (Stephen et al., 1993). The youngest rocks beneath this unconformity can only be assigned a broad late Pliensbachian to early Toarcian age, and the oldest rocks above this unconformity are most likely to be Bathonian in age. Regional correlations suggest the correlative conformity to this 1st-order sequence boundary is of Aalenian Age (Underhill and Partington, 1993, 1994).

Transgressive/Regressive Facies Cycle Ja.2.—

During the Hettangian and Sinemurian a ubiquitous fining-upwards sedimentary section was deposited by a retreating alluvial fan (Figs. 5A, 6). The basal Lower Jurassic succession at Golspie consists of a (30-m-thick) interval of braided stream deposits (Batten et al., 1986). No sediments of analogous origin have been identified from time-equivalent strata across the remainder of the IMF. At present, it is unclear whether these braided stream sediments represent incised valley deposits, because current exposures do not allow for an accurate understanding of their three-dimensional distribution.

This interval of backstepping facies belts culminates in a thin zone showing temporary marine influence in the Golspie (Lam and Porter, 1977; Fig. 5A) and Lossiemouth areas (Berridge and Ivimey-Cook, 1967; Fig. 5B). This horizon corresponds to
the position of peak transgression that can only be assigned a broad lower Sinemurian age. The correlative interval in the Beatrice Field and adjacent areas is a thin interval of lacustrine mudrocks (Fig. 6). Above this zone of temporary marine influence, a return to freshwater environments is apparent as is evidenced by the deposition of an aggrading to weakly prograding distal alluvial fan succession. This section is severely truncated by the overlying White Sandstone Unit with this erosive junction marking the position of maximum regression that defines the top of this cycle (Fig. 6).

No definitive higher-frequency stratigraphic cycles can be identified within this 2nd-order cycle. Such a situation arises from the difficulty in distinguishing and correlating key sequence stratigraphic surfaces within predominantly nonmarine intervals in subsurface datasets when core coverage is limited.

Transgressive/Regressive Facies Cycle Ja.4.—

The basal part of this cycle is comprised of a series of stacked fluvial/estuarine distributary channel systems (White Sandstone Unit), which contrast strongly with the underlying silt-dominated units. Detailed well correlations demonstrate that the underlying alluvial plain sediments are markedly truncated (Fig. 6). The amount of erosion (estimated as 35 m in well 12/22-3 and possibly more in the Lossiemouth borehole; Fig 5B) exceeds the thickness of individual distributary channels (up to
10-m-thick) in the overlying arenaceous section and as such cannot be explained by distributary channel switching alone. Hence, this period of erosion records an interval of widespread subaerial exposure, truncation and valley cutting. Poor palynological recoveries only allow a broad lower to upper Sinemurian age estimate to be made for this cycle-bounding unconformity. The sedimentary fill of the incised valleys reveals an upwards-increase in marine influence (Stephen et al., 1993). The dimensions of the incised valley complex are difficult to ascertain, but this depositional system appears to have been very broad (Fig. 6). Such a feature may be explained by the coalescence of originally disparate channels into one large valley system. In the Lossiemouth area, a very distinctive (15-cm-thick) erosively based conglomerate documented by Berridge and Ivimey-Cook (1967; Fig. 5B) is interpreted as the correlative interfluve.

Close to the top of the White Sandstone Unit occurs an interval of reworked rounded siderite clasts (Fig. 7A). The siderite clasts are most likely to have been reworked during transgression from an easily eroded soil horizon or hardpan formed during a phase of earlier subaerial exposure. This association suggests the presence of an additional sequence boundary (in this case coincident with a transgressive surface). This sequence boundary, which can only be assigned a broad upper Sinemurian age and for which a hierarchical status is not yet known, can only be identified and correlated in cored wells within the Beatrice Field area (Block 11/30) and has not been used to define a new stratigraphic cycle in this study. However, it is possible that this unconformity may correlate to the 2nd-order surface of maximum regression of upper Sinemurian age documented by Hesselbo and Jenkyns (this volume).

The juxtaposition of offshore marine mudstones of the Ladys Walk Shale Member on top of this incised valley/interfluve complex represents a distinct break in deposition. This erosive ravinement surface, above which a ubiquitous fining-upwards trend is developed, is no younger than uppermost Sinemurian, *Paltechioceras (P.) aplanatum* Subzone (Stephen et al., 1993). At Golspie, peak transgression is defined at the position of finest-grained mudrock deposition which occurs within the *P. taylori* Subzone and coincides with a band of prominent carbonate concretions (Fig. 5C). In offshore wells, this surface is easily identified at the position of gamma-ray maxima and sonic-velocity minima (Fig. 6). The conformable progression from the upper part of the Ladys Walk Shale Member to the Orrin Formation records a shallowing of depositional environment. The top of this cycle is bounded by the MCU.

**Regressive/Transgressive Cycle Ja.4.2.—Offshore-marine mudstones form the lowermost part of the Ladys Walk Shale Member and the base of this cycle is picked at the position of finest-grained sedimentation within this succession (Figs. 5C, 5D).** This maximum flooding surface is dated as uppermost Sinemurian (*P. aplanatum* Subzone) on the basis of offshore ammonite recoveries (Stephen et al., 1993). A similar succession of aggrading mudstones is developed above this surface, which in turn are sharply overlain by offshore-transition zone silty-sandstones (Fig. 7B). Minor truncation can be demonstrated below the junction between these two lithologies, which may be lined by a bioclastic lag deposit. A distinct basinward shift in facies occurs across this junction, that is a sequence boundary of latest Sinemurian age. Above this sequence boundary occurs a package of transgressive sediments of variable thickness. Therefore, a transgressive surface is believed to coincide with this sequence boundary across much of the IMF (Fig. 6). The top of this cycle coincides with the surface of peak transgression previously identified within the lower Pliensbachian (*P. taylori* Subzone; Figs. 5A, 5D).

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**Fig. 7.—Key Lower Jurassic sequence stratigraphic surfaces.**  
(A) Well 11/30-2, depth 7125–7129 feet. *Diplocraterion parallelum* burrows overprint an earlier freshwater ichnofabric (rootcasts) and are associated with reworked siderite clasts suggesting the presence of a sequence boundary coincident with a transgressive surface (SB/TS).  
(B) Well 11/30-a6, depth 6978–6982 feet (see Figure 5D for stratigraphic summary). Transgressive surface coincident with a sequence boundary (SB/TS) within the basal part of the Ladys Walk Shale Member.  
(C) Well 12/27-1, depth 3694–3698 feet. Lower Callovian transgressive marine sandstones unconformably overlie uppermost Sinemurian siltstones. This major break in deposition has been termed the “Mid-Cimmerian” unconformity.
Regressive/Transgressive Cycle Ja.4.4.—Our detailed knowledge of this cycle is limited to offshore boreholes. The sedimentology of this succession is very similar to the previous cycle, comprising a lower regressive unit and upper transgressive unit separated by a basinward shift in facies (Fig. 5D). A similar sequence stratigraphic interpretation is therefore proposed, with a sequence boundary coincident with a transgressive surface marking the junction between these two lithotypes. This composite surface can only be assigned a broad lower Pliensbachian age. The upper boundary to this cycle is defined at the position of finest-grained deposition within the upper part of the Lady's Walk Shale Member. This corresponds to a thin zone of organic-rich mudstones with exceptionally low sonic velocity values (Fig. 5D). It is at this distinctive petrophysical log marker where Linsley et al. (1980) consider that an unconformity exists (Fig. 6). If this is the case, then this break in deposition, which probably formed as a result of sediment starvation associated with peak relative sea-level rise, is most likely to have been of relatively minor duration, confined to the lower Pliensbachian interval (Stephen et al., 1993). Vail and Todd (1981) suggest that this “unconformity” corresponds to their lowermost Pliensbachian (179 my) sequence boundary, although as demonstrated here this surface is in fact a maximum flooding surface.

Regressive/Transgressive Cycle Ja.4.6.—This cycle comprises part of a major coarsening-upwards unit (the approximately 45-m-thick Orrin Formation; Fig. 6). The lower part of the Orrin Formation (the I Sand) can be split into two genetic units. The lower regressive unit was deposited in a shallow-marine environment by a prograding wave-dominated delta. These marine, fine- to medium-grained sandstones are sharply overlain by coarse-grained, stacked fluvial sandstones which comprise part of the upper genetic unit of the I Sand. This juxtaposition of facies marks a basinward shift in facies, and significant incision and valley formation can be documented into the underlying marine sediments (e.g., well 12/26-1). The absence of lower Pliensbachian marine mudstones in the Lossiemouth borehole is probably a result of this erosive event (Fig. 5B). These multi-storied fluvial channel sandstones are gradually succeeded by more isolated and more sinusous fluvial channel deposits towards the top of the I Sand. Shanley and McCabe (1993) suggest that such stacking patterns reflect increasing rates of base-level rise. The position of the maximum flooding surface that defines the top of this cycle can be determined by trace fossil analysis with zones of Diplocraterion indicating the positions of significant marine influence. Marine dinocysts recovered from the horizon of maximum marine influence only allow for a broad late Pliensbachian to early Toarcian age assignment to be made for this event (Stephen et al., 1993).

Regressive/Transgressive Cycle Ja.4.8.—Above this zone of temporary marine influence, a rapid return to freshwater environments is evidenced by the predominance of the alga Botryococcus sp. (Stephen et al., 1993). The H Sand, which occurs above this maximum flooding surface and comprises the topmost part of the Orrin Formation, was deposited in a variety of nonmarine environments. However, isolated high-sinuosity fluvial channels and (6-m-thick) mouthbars predominate volumetrically and interdigitate with discontinuous coals and carbonaceous shales. This aggradational alluvial architecture may reflect accommodation space broadly remaining constant or decreasing slightly.

A major unconformity (MCU) occurs at the top of the Orrin Formation and represents an important phase of sedimentary bypass and erosion across the whole of the IMF basin (Figs. 6, 7C). The youngest preserved strata below this unconformity can only be assigned a broad upper Pliensbachian to lower Toarcian age (Stephen et al., 1993). However, with the knowledge that lower Toarcian strata in many of the onshore and offshore basins of the United Kingdom are of a deep-marine mudrock facies deposited during a major transgressive episode (Partington et al., 1993b; Hesselbo and Jenkyns, this volume), it might be expected that a similar stratigraphic development be encountered in the IMF if Toarcian strata were present. Since such sediments are not present, it is probable that lower Toarcian sediments are not preserved in the IMF area.

Major Transgressive/Regressive Cycle Jb

The MCU delineates the base of this cycle and separates the Dunrobin Bay and Humber Groups. At least the Toarcian, Aalenian and probably the Bajocian Stages are absent (Fig. 2). Sedimentation recommenced in the IMF basin with the deposition of the non/marginal-marine Brora Coal and Pentland Formations. Exact dating of these units is precluded by the paucity of marine palynomorphs, although a Bathonian age is preferred for the Brora Coal Formation (IMF; Stephen et al., 1993), and a broader Bajocian/Bathonian age is postulated for the Pentland Formation (OMF; Harker et al., 1987). This alluvial/coastal plain succession is succeeded by Callovian shallow-marine sediments in the IMF (Figs. 8, 9, 10). In the OMF, similarly-aged marine successions have been considered absent (e.g. Andrews and Brown, 1987), although recent work indicates their local occurrence (Davies et al., 1996). This backstepping and easterly migration of depositional systems, which in other parts of the North Sea began in the upper Aalenian Age (Underhill and Partington, 1993, 1994), continues through the Oxfordian and into the Kimmeridgian but poor paleobathymetric control makes it difficult to ascertain the position and timing of peak transgression. Partington et al. (1993a) suggested that Jurassic relative sea level reached a maximum in the North Sea area in the late Kimmeridgian Age (Aulacostephanus (A.) eudoxus Zone), although recent work undertaken in the OMF (L. Riley, pers. commun., 1995) may indicate that deepening continued into the latest Kimmeridgian or earliest Volganian time.

During the peak part of this major transgressive episode arenaceous turbidite deposits are volumetrically restricted. The first significant influx of arenaceous sediments into the MF basin began in early Volganian time and continued into the late Ryazanian. This dramatic increase in coarse clastic sedimentation indicates that little or no accommodation space existed in shallow-shelf settings at this time. Consequently, Volganian to Ryazanian deep-marine sediments were probably deposited at a time of decreasing shelfal accommodation space. Maximum regression is positioned at the boundary between the J2.5/K.1 seismic sequences of Underhill (1991a, b), also termed the “Base Cretaceous” unconformity that is dated as uppermost Ryazanian in age (close to the Surites (S.) stenomphalus-Peregri noceras (P.) albidum Zonal boundary). This important seismic marker is not an unconformity across most of the North Sea.
area but instead is the result of stratigraphic thinning below the resolution of the seismic method. This is a consequence of the facies change from hemipelagic sediments of the Kimmeridge Clay Formation to the pelagic deposits of the Cromer Knoll Group (Rattey and Hayward, 1993).

Transgressive/Regressive Facies Cycle Jb.2.—

This cycle comprises most of the Brora Coal Formation, was deposited in a variety of alluvial and coastal plain environments and is characterized by an overall fining-upwards trend accompanied by a decrease in marine influence (Figs. 8, 10). Severe difficulties exist in understanding the sequence stratigraphic subdivision of the Brora Coal Formation because of a lack of prominent stratigraphic markers and poor biostratigraphic recoveries. However, comparisons to the outcrop sequence stratigraphic studies of alluvial systems by Shanley and McCabe (1993) suggests that the Brora Coal Formation was probably deposited during a period of overall base-level rise. Also the observed trend of fluvial channels becoming increasingly more isolated within finer-grained alluvial plain deposits is most likely to indicate deposition occurred during a period of decreasing rates of base-level rise. Therefore, the bulk of this sedimentary cycle was probably deposited during a 2nd-order regressive phase. Maximum regression occurs near the top of the Brora Coal Formation, and at both onshore localities corresponds to the junction between the Doll and Inverbrora Members of Hurst (1981; Figs. 8, 9). Poor paleontological recovery precludes the exact dating of this surface and a broad age assignment can only be made by considering the microfossil recovery from the basal part of the overlying 2nd-order cycle. This lagoonal succession is characterized by the persistent occurrence of dinocysts typical of the uppermost Bathonian Cly- doriceras (C.) discus and lowermost Callovian Macrocephali- tes (M.) macrocephalus ammonite Zones (MacLennan and Trewin, 1989). Thus, the top of this cycle is no younger than latest Bathonian in age (C. discus Zone). Sedimentary logging...
of the Brora Coal Formation has not yet enabled the definition of the intermediate surface of peak transgression that theoretically should occur between the two unconformities that bound this cycle.

It is likely that higher-frequency sedimentary cycles are present within the Brora Coal Formation. Indeed abrupt changes in fluvial stacking patterns, as indicated by amalgamated fluvial channel systems, can be recognized (e.g., wells 12/21-3 and 12/16-1, where 30-m-thick stacked fluvial sandbodies are locally developed). According to the models of Shanley and McCabe (1993), this style of fluvial architecture most likely indicates a basinward shift in facies and the presence of a sequence boundary at the base of the amalgamated channel system. However, regional and even local correlations of such higher-frequency sequence boundaries are fraught with difficulty. Intermediate maximum flooding surfaces also cannot be identified. Consequently, this 2nd-order cycle has not been subdivided into constituent higher-frequency sedimentary cycles.

Transgressive/Regressive Facies Cycle Jb.4.—

The base of this cycle is uppermost Bathonian, *C. discus* Zone in age. As previously discussed, the Doll Member of the Brora Coal Formation is considered as having been deposited during a period of decreasing rates of base-level rise. Maximum regression occurs at the top of this lithostratigraphic unit with well log correlations and seismic studies suggesting that subaerial erosion truncation has occurred (Vail and Todd, 1981; Underhill 1991a, b; Fig. 10). At Brora (Fig. 11A) this unconformity corresponds to a brecciated siderite cementstone horizon, interpreted as a paleosol by Hurst (1985).

A significant regional change in sedimentary architecture occurred in latest Bathonian times with marginal and shallow marine sediments transgressing the older Bathonian alluvial plain, as well as onlapping Lower Jurassic and older strata in an eastwards direction (Stephen et al., 1993; Fig. 10). At the onshore localities, finest-grained deposition and therefore the position
of peak transgression occurs close to the middle/upper Callovian boundary (Figs. 8, 9). Across much of the remainder of the basin this surface is younger in age (close to the Callovian/Oxfordian Stage boundary; Fig. 10). This discrepancy in timing may be explained by regional variations in the rates of sediment supply, as sediments accumulating in the western extremities of the basin are most likely to have been derived from the Scottish Highlands landmass, whereas most of the central and eastern parts of the basin were probably sourced from intra-basinal highs to the east (and possibly the subsiding remnants of the Middle Jurassic thermal dome). The anomalous stacking of Callovian higher-frequency cycles in the west may therefore have resulted from localized high sediment flux exceeding regional increases in accommodation space.

The upwards progression from the Broyra Argillaceous to Brora Arenaceous Formations onshore (Figs. 8, 9), and similar trends recognized within the Uppat Formation in the subsurface (Fig. 10) records a regression and shallowing in depositional environment. During middle Oxfordian time a significant unconformity developed in more proximal settings, such as in the Ross Field area (Blocks 13/28 and 13/29; Fig. 10) and in the sites of embryonic, intra-basinal, structural highs due to a lack of sufficient shelfal accommodation space. The timing of maximum regression can only be assigned a broad late middle Ox-
fordian age (Cardioceras (C.) tenuiserratum or C. blakei ammonite Subzone) on the basis of ammonite recoveries at Balintore (Fig. 9) and microfossil recoveries from offshore wells.  

Regressive/Transgressive Cycle Jb.4.2.—The maximum flooding surface that defines the base of this cycle is positioned within the lagoonal sediments of the Inverbrora Member (Fig. 11A). The best candidates for this surface are either of the two Neomiodon sp. and Isoynoman sp. shell beds that have associated dinocyst assemblages which are found close to the Bathonian/Callovian Stage boundary (MacLennan and Trewin, 1989). Above this maximum flooding surface, there is a relatively thin succession of aggradational lagoonal sediments with a variable marine influence that are sharply overlain by the Brora Coal that was deposited in a freshwater-floating swamp environment (Figs. 11A, 12A, 13A). This basinward shift in facies records an early Callovian fall in relative sea level which can only be dated as occurring between the Bathonian/Callovian Stage boundary and either the Macrocephalites (M.) macrocephalus ammonite Zone or the Proplanulites (P.) koenigi ammonite Subzone.  

At both onshore localities, the Brora Coal is unconformably overlain by a thin transgressive marine sandstone (the Roof Bed; Figs. 11B, 12A, 13B). This unconformity developed as a result of wave and tidal erosion as the shoreface profile translated eastwards through a process of shoreline retreat. In offshore wells, the Brora Coal is overlain by estuarine and fluvial sediments with marine transgression and ravinement erosion occurring somewhat later (Fig. 13A). Overlying this shallow/marginal-marine succession is a widely developed retrogradational offshore marine mudstone package where the maximum flooding surface, which defines the top of this cycle, provides an easily recognized datum (Stephen et al., 1993; Fig. 10). At Brora, this surface is positioned at a major horizon of carbonate concretions within the Brora Shale Member (Fig. 11B), which formed during a break in deposition resulting from very low sedimentation rates associated with the peak rate of relative sea level rise. The maximum flooding surface is coincident with the boundary between the Sigaloceras (S.) calloviense and S. endodatum Subzones. At Balintore this surface is more difficult to position and date, but probably occurs towards the base of the Cadh’ an Righ Shale Member or at the top of the Brora Roof Bed (Fig. 12A).  

Regressive/Transgressive Cycle Jb.4.4.—A well-defined regressive package of sediment occurs directly above the lower Callovian maximum flooding surface which bounds the base of this cycle (Fig. 10). At Brora, this prograding interval is abruptly terminated by a sharp basinward shift in facies and an influx of sand-grade material (Figs. 11C, 13C). This junction between the Brora Shale Member and Glauconica Sandstone Members is a sequence boundary that is Kosmoceras (K.) jason Subzone in age. Across most of the eastern part of the IMF basin, this unconformity is of greater duration and corresponds to a discrete zone of sedimentary bypass (Figs. 10, 13D). Significant truncation can be demonstrated beneath this sequence boundary, especially in the Beatrice Field area, although no conclusive evidence for incised valley fill development has been identified.  

The Glauconica Sandstone Member is a condensed, aggradational to slightly progradational unit deposited when sedimentation rates were broadly equivalent to the rate of creation of accommodation space. The junction between this unit and the overlying strongly retrogradational Brora Brick Clay Member is a transgressive surface that is dated as K. grossouvrei Subzone (Fig. 11C). Across the remainder of the IMF basin a similar backstepping belt of facies is developed above this sequence boundary and the transgressive sands deposited at this time (the upper part of the A Sand) form part of the main hydrocarbon reservoir in the Beatrice Field (Fig. 10). At Brora, the maximum flooding surface that culminates this transgressive trend is coincident with a major horizon of carbonate concretions and occurs approximately (1.5 m) above the base of the K. phaenium Subzone (i.e approximately coincident with the middle-upper Callovian boundary; Fig. 11D). At Balintore, a non-sequence spanning most of middle Callovian time has been documented by Sykes (1975; Fig. 12A). It is probable that this unconformity represents a significant period of composite erosion/nondeposition related to either insufficient accommodation space at the time of deposition of the Glauconica Sandstone Member, or sediment starvation and winnowing during the later transgressive period, or both.  

Regressive/Transgressive Cycle Jb.4.6.—Across much of the basin, the basal part of this cycle is characterized by offshore-marine mudstones that coarsen upwards into offshore-transition zone siltstones. At Brora, a distinct change from a dominantly progradational to an aggradational/weakly progradational stacking pattern can be discerned within these sediments (Fig. 11D). This subtle surface is dated as K. phaenium Subzone in age and records the change from a time when sedimentation rates exceeded the rate of creation of accommodation space, to a time when sedimentation rates broadly equalled accommodation rates. Such a change in stacking patterns may indicate that this surface is a Type 2 sequence boundary (sensu Van Wagoner et al., 1990). However, severe difficulties exist in identifying and correlating this surface in the shale-prone environments encountered across much of the remainder of the basin. This weak progradational trend is terminated by a regionally correlatable transgressive surface (Fig. 10). At Brora, this surface is dated as Quenstedoceras (Q.) lamberti Subzone (Figs. 11E, 13E). The remainder of this cycle exhibits a retrogradational stacking pattern culminating at a well-defined maximum flooding surface. This surface has been dated as no younger than upper Callovian in the offshore wells (Fig. 10; Stephen et al., 1993) an age that can be confirmed and refined as Q. lamberti Subzone at Brora.  

Regressive/Transgressive Cycle Jb.4.8.—Disparities exist between the uppermost Callovian to middle Oxfordian strata exposed at Brora, and those encountered across the remainder of the basin in that an additional, higher-frequency sedimentary cycle can be defined in the lowermost part of this cycle at this onshore locality (termed sedimentary cycle Jb.4.8a; Figs. 11E, 14A). Above the basal cycle bounding maximum flooding surface occurs a thin (0.5-m-thick) interval of prograding siltstones that is sharply attenuated by the Clynelish Quarry Sandstone Member. The junction between these two units is marked by a sudden influx of well-sorted fine-grained sandstones (Figs. 13E, 13F), and a significant change in depositional environment is envisaged, from offshore-transition zone to subtidal nearshore. This junction marks a distinctive basinward shift in facies and this sequence boundary is dated as Q. lamberti Subzone in age.
FIG. 11.—Stratigraphic breakdown of key Middle Jurassic successions in the Brora area, Sutherland. (A) Brora Coal Formation exposed on the foreshore (NC 905032). (B) Brora Shale Member exposed on the foreshore (NC 905032). (C) Glauconitic Sandstone Member exposed on the northern bank of the Brora River cliff section (NC 888040). (D) Upper Callovian succession exposed on the foreshore south of the Brora River estuary (NC 909031). (E) Upper Callovian succession exposed on the southern bank of the Brora River (NC 898038).
The Clynelish Quarry Sandstone Member has an overall slightly progradational stacking pattern, exhibits ample evidence for deposition from tidal currents and exhibits many features characteristic of an incised valley fill succession. However, the exposures at Brora are essentially two-dimensional and patchy, therefore a valley morphology cannot be verified. An abrupt change to retrogradational stacking patterns occurs near the top of this Member, and the ravinement surface that defines this position is also of Q. lamberti Subzone age (Fig. 14A). The maximum flooding surface which would be used to
define the top of this higher-frequency sedimentary cycle has been truncated by a sequence boundary at the base of the Brora Sandstone Member. A potentially correlative maximum flooding surface which is not widely developed and is poorly dated (appears to occur close to the *Q. mariae/C. cordatum* zonal boundary), can be identified in some wells with relatively thick lower Oxfordian sections (e.g. well 12/21-4).

Across the remainder of the basin, a well-developed lower and middle Oxfordian regressive section occurs above the uppermost Callovian maximum flooding surface that delineates the base of this cycle (Figs. 9, 10). At Balintore, the top of this regressive section occurs at the junction between the Shandwick Siltstone and Port an Righ Ironstone Members (Fig. 12B). This transgressive surface is dated as either *C. cordatum* or *C. vertebrale* ammonite Subzone. No sequence boundary is identified in this section. However, in several offshore wells an early Oxfordian (*C. cordatum* Zone) basinward shift in facies has been recognised (e.g., well 12/30-1, Fig. 10). The inability to recognise the correlative conformity in the section at Balintore precludes a more definitive age determination.

It is not apparent whether the basinward shift in facies recorded at the base of the Brora Sandstone Formation (Fig. 14A) corresponds to: (i) the lower Oxfordian sequence boundary recognized in wells such as 12/30-1, (ii) a sequence boundary as-
FIG. 14.—Stratigraphic breakdown of key Upper Jurassic successions in the Brora area, Sutherland. (A) Brora Arenaceous Formation exposed on the southern bank of the Brora River (NC 900040). No ammonites have been recovered from the Brora Sandstone Member. Consequently, the exact position of the Callovian/Oxfordian Stage boundary in this area is not known, and the duration of the hiatus between the Brora Sandstone and Clynelish Quarry Sandstone Members is not yet apparent. (B) Brora Sandstone Member exposed along the southern side of the Brora River (NC 900040-906039). The transgressive surface is placed near the top of the logged section although the lack of continuous exposure makes the position of this surface extremely tentative. (C) Middle Oxfordian strata exposed at Ardassie Point (NC 913041).

Regressive/Transgressive Cycle Jb.4.10.—At Balintore, above the middle Oxfordian maximum flooding surface an increase in clastic as opposed to chemical sedimentation occurs when progressing from the Port an Righ Ironstone into the Port an Righ Siltstone Member (Fig. 12B). Maximum water depths, as interpreted from the position of finest-grained sedimentation, developed in the C. tenuiserratum Subzone, some time after peak transgression. Such a situation is plausible where the decreasing rate of accommodation space creation locally exceeded sediment supply. The remainder of the middle Oxfordian section at Balintore and in many of the offshore wells is weakly progradational (Fig. 10). The top of this regressive succession is marked by a fundamental change in facies, and rapid deepening is indicated, for example, by changes in microfaunal assemblages (Andrews and Brown, 1987). At Balintore, the junction between these progradational and the overlying strongly retrogradational sediments is interpreted as a transgressive surface that can only be broadly constrained as latest middle Oxfordian in age (C. tenuiserratum or C. blakei Subzone; Fig. 12B). No sequence boundary can be recognized. The transgressive surface marks the position of maximum middle Oxfordian regression and corresponds to the junction between the Uppat and Heather Formations. Due to the creation of accommodation space across the MF basin in latest middle Oxfordian times, sedimentation resumed in many areas that were previously bypassed. For example, the transgressive marine uppermost middle Oxfordian Ross Sandstone Unit unconformably overlies Callovian fluvial sediments in Blocks 13/28 and 13/29, where it forms an important hydrocarbon bearing interval (Figs. 10, 15C). The culmination of this retrogradational stacking pat-
Fig. 15.—Key Upper Jurassic sequence stratigraphic surfaces. (A) Brora Sandstone Member exposed at Strathsteven Cliffs, south of Brora (NC 886021). Large-scale tabular cross-bedding characterizes the unit, which was deposited by a series of south eastward-migrating sand dunes in a strongly tidally influenced sea. The strong tidal influence, sharp and erosional nature of the lower boundary (Fig. 14A) and aggradational to weakly progradational stacking pattern of this member may indicate deposition occurred during the early stages of base level rise. (B) A transgressive surface (TS) marks the junction between the Shandwick Siltstone and Port an Righ Ironstone Members at Balintore (NH 853733, see Figure 12C for stratigraphic detail). The maximum flooding surface (MFS) is positioned at the most intense zone of sediment starvation and siderite cementation (Bed 3 of Sykes, 1975). (C) Well 13/28-2, depth 9641–9643 feet. Boundary between the fluvialite Parry and transgressive marine Ross sandstones, interpreted as a transgressive surface coincident with a sequence boundary (SB/TS; the junction is a preserved sample). (D) Well 15/21a-A1, depth 14884–14886 feet. The basinward shift in facies at the base of the upper Oxfordian Scott Member is interpreted as a sequence boundary (SB). (E) Allt na Cuille Sand Member exposed at Lothbeg Point, south of Helmsdale (ND 961093; see Figure 17A for stratigraphic detail). (F) Loth Burn Siltstone Member exposed in the banks of Loth Burn adjacent to the railway bridge (ND 954099; see Fig. 17B for stratigraphic detail). This member consists of a series of lenticular sandstones deposited within isolated submarine channels by immature turbidity currents, interbedded with siltstones deposited from dilute turbidity currents (Wignall and Pickering, 1993).
tern and the top of this cycle occur at a well-defined maximum flooding surface. As detailed by Stephen et al. (1993), this surface is dated as lowermost upper Oxfordian, Amoeboceras (A.) glosense Zone.

Transgressive/Regressive Facies Cycle Jb.6.—

The base of this cycle is positioned at the surface of middle Oxfordian maximum regression, which can be mapped across the IMF and into the OMF using lithological and seismic criteria. Above this important regional event, the rapid creation of additional accommodation space, inherently linked to the onset of accelerated rifting (Davies et al., 1996), resulted in a major shift in the locus of shoreline positions from the IMF to the eastern part of the OMF basin. Individual higher-frequency cycles, which are detailed below, exhibit an overall retrogradational stacking pattern (Fig. 16), as is clearly demonstrated by a progressive decline in the proportion of alluvial/coastal plain to marine sediments (Davies et al., 1996). Hence, the upper Oxfordian and Kimmeridgian sediments of the MF basin were deposited during a 2nd-order transgressive phase. The timing of peak transgression is difficult to ascertain because poor palaeobathymetric control precludes such determinations. It is most likely, however, that peak transgression occurred in latest Kimmeridgian or earliest Volgian times. Additional work needs to be undertaken to ascertain when maximum regression occurred.

During late Oxfordian and Kimmeridgian times much of the north-eastern part of the OMF (the Witch Ground Graben; Fig. 1B) was characterized by shallow-marine to non-marine environments, whereas the remainder of the basin was characterized by deeper-water (shelf and slope) mud-prone sedimentation. As such, the Witch Ground Graben area shows greater lithological variation where the identification, correlation, and dating of higher-frequency cycles is best achieved (Fig. 16). The lithostratigraphic nomenclature for this interval is extremely complex, but in the following discussion the scheme advocated by Harker et al. (1993) is employed unless otherwise stated.

Regressive/Transgressive Cycle Jb.6.2.—Sedimentation in the OMF recommenced with the deposition of the non- to fully-marine Sgiath Formation. Middle Oxfordian, or possibly older Oxfordian (Harker et al., 1993), paralic sediments of the Skene Member onlap the Bajocian/Bathonian Fladen Group (Pentland and Rattray Volcanics Formations) and Triassic red beds. These are in turn sharply overlain by open-marine siltstones and mudstones of the Saltire Member deposited as a result of a ubiquitous lower Kimmeridgian marine transgression (Fig. 16). The base of this cycle is picked at a widespread gamma-ray maximum/sonic minimum identified within the Saltire Member, which correlates to the lowermost upper Oxfordian (A. glosense Zone) maximum flooding surface recognized in the IMF.

Above this zone of maximum flooding, an upward-decreasing serrate gamma-log profile is a common feature, consistent with deposition by a prograding shoreline system (Fig. 16). In many Witch Ground Graben wells (e.g., 15/21a-A1, Fig. 15D), this regressive unit is often highly attenuated and in one presently unreleased well, alluvial fan conglomerates of the Scott Member abruptly overlie offshore marine mudstones of the Saltire Member. This surface is the only basinward shift in facies that can be confidently identified and correlated within the OMF. Accurate dating of this sequence boundary is problematic due to the very poor microfossil recoveries from the Scott Member, as well as the difficulty in identifying and therefore dating the correlative conformity in more distal wells. Consequently, the age of this surface can only be constrained by the ages of the maximum flooding surfaces that define the base and top of this cycle (i.e., lowermost upper Oxfordian to intra-upper Oxfordian).

In localized areas of the Witch Ground Graben (for example block 15/21), stacked distributary channel facies deposited above this sequence boundary mark the renewed onset of accommodation space creation on the alluvial plain. The subsequent abandonment of these incised valleys and the deposition of retrogradational marine successions indicates that local sediment supply was outpaced by relative base level rise. Dating this transgressive event is difficult due to unsuitable lithologies. This composite phase of shoreline progradation, subsequent fluvial incision and later transgression gives rise to a discrete unit forming the lower part of the Scott Member. The maximum flooding surface that terminates this retrogradational trend can only be assigned a broad intra-upper Oxfordian age. The correlative event in the IMF has been dated as A. serratum Zone by Copestake (1993). However, this age assignment is based upon widely spaced cuttings samples. Therefore, this maximum flooding surface is assigned a more pragmatic intra-upper Oxfordian age.

Regressive/Transgressive Cycle Jb. 6.4.—It is possible to recognize this cycle only in specific parts of the Witch Ground Graben (Block 15/21, the Scott Field area, Fig. 16; Davies et al., 1996). Above the intra-upper Oxfordian maximum flooding surface, there is commonly an abrupt shift from offshore-transgression zone siltstones to shoreface sandstones, consistent with the progradation of a wave-dominated shoreline system. It is uncertain whether this lithological junction corresponds to the position of a sequence boundary. A second sharp and erosive junction exists between these sandstones and an overlying thin succession of transgressive marine siltstones. Once again this transgressive surface cannot be accurately dated. The maximum flooding surface that caps this retrogradational interval and bounds the top of this cycle occurs within the I Shale (as dated and defined by Maher, 1981). Contradictory ages have been assigned to this pulse of shale deposition (cf. Maher, 1981; O’Driscoll et al., 1990). It has however been assigned to the A. rosenkrantzi Zone by Harker et al. (1987, 1993). Importantly, the A. rosenkrantzi maximum flooding surface cannot be easily correlated out of the Piper Field (contra Harker et al., 1993). Instead, these authors have confused this event with the distinct and ubiquitous lower Kimmeridgian maximum flooding surface dated as Pictonia (P.) baylei Zone by ammonite recoveries from the Mid-Shale of the Scott, Ivanhoe, and Rob-Roy Fields (Brealey, 1990).

Regressive/Transgressive Cycle Jb. 6.6.—This sedimentary cycle is particularly well developed in Blocks 15/17 and 15/21 (Fig. 16). Above the basal bounding maximum flooding surface, there exist sedimentary and biogenic trends similar to those identified in the previously described higher-frequency cycles. In more proximal areas, aggrading alluvial/coastal plain sediments were deposited at the same time as the regressive shoreline succession. This progradational stacking trend, which attains maximum thicknesses in the Piper Field area and thins drastically to the south and south-west, is abruptly terminated by a pronounced transgressive surface that upon structural highs
is characterized by erosion and/or nondeposition (this may account for the absence of the *A. rosenkrantzi* Zone maximum flooding surface in many such areas). It is possible that evidence for a basinward shift in facies may have been removed by erosion across this surface. This transgressive pulse marks the depositional initiation of the Kimmeridge Clay Formation (*sensu* Deegan and Scull, 1977) in both the IMF and western areas of the OMF basin. On the basis of microfossil recoveries, the age of this distinct lithostratigraphic boundary falls close to the Oxfordian/Kimmeridgian Stage boundary.

In the Witch Ground Graben area, the earliest Kimmeridgian transgressive episode resulted in a distinct landward shift in facies belts and a sharp lithological change (e.g., from upper-shoreface sands to open-marine mudstones and siltstones in the Scott Field). This retrogradational facies stacking is capped by a ubiquitous condensed horizon (Fig. 16). The age of the maximum flooding surface that defines the top of this cycle is well constrained as a result of the extraction of ammonites from cores that place this surface close to the *P. baylei/Rasenia (R.) cymodoce* Zonal boundary (Brealey, 1990).
Regressive/Transgressive Cycle Jb. 6.8.—A significant change in the sedimentary architecture of Jurassic higher-frequency cycles in the IMF and western parts of the OMF follows the lowermost Kimmeridgian transgression. In these areas, the most common Kimmeridgian and younger depositional processes are those of mass flow and turbidity currents. In contrast, shallow-marine deposystems prevailed in the Witch Ground Graben area and above the lowermost Kimmeridgian maximum flooding surface occurs a similar regressive shoreline succession to those previously described (Fig. 16). Using the ammonite recoveries as determined by Brealey (1990), this progradational trend continues from the *R. cymodoce* into the *A. mutabilis* Zone. Delta front turbidites, not to be confused with “lowstand” submarine fan deposits, are developed locally (e.g., western part of the Scott Field, Block 15/21).

At the onshore localities, a significantly different succession is encountered (Fig. 17). In the Helmsdale area, pronounced lateral variation occurs with sediments deposited by debris flows and slides interdigitating with submarine canyon/channel sandstones (Pickering, 1984; Figs. 15E, 17A). All these deposits developed on the downthrown side of the active Helmsdale Fault. Because a gap in exposure exists between the outcrops around Kintradwell and the middle Oxfordian strata at Brora, a sequence stratigraphic interpretation for these lower Kimmeridgian deposits is problematic as it is not clear what sediments underlie these units (Fig. 2). However, the fact that significant quantities of sand were transported from a narrow shelf on the footwall to the Helmsdale Fault to accumulate in a channel/canyon system on the hangingwall side, indicates that sufficient accommodation space did not exist on the upthrown side of this active fault during the deposition of this member. This sudden influx of mature sand into the basin may therefore record a basinward shift in facies (“lowstand” deposition) and a sequence boundary is envisaged to occur at the base of this channel/canyon system. This candidate sequence boundary is dated as *R. involuta* Subzone or older. Correlative turbidite sandstones (*R. cymodoce* Zone) are recognised in Blocks 12/21, 13/30 and 14/26, indicating that this fall in relative sea level was not confined to the area around the Helmsdale Fault.

An important change in sedimentation occurs near the top of the *R. uralensis* Subzone in the Kintradwell area, with the sudden abandonment of the amalgamated submarine channel complex and the commencement of finer-grained deposition by dilute turbidity currents. This sharp boundary, corresponding to the junction between the Allt na Cuille Sand and Lothbeg Siltstone Members (Fig. 17A), is equivalent to the “top basin floor fan surface” of Vail et al. (1991). The Loth Burn Siltstone Member, the up-dip time-correlative to the Lothbeg Siltstone Member, contains (1.5- to 2-m-thick), isolated submarine channel
sandstones interbedded with fissile siltstones (Figs. 15F, 17B). Both members are analogous to the “slope fan” complex of Posamentier and Vail (1988). A rapid deepening is envisaged from the A. mutabilis into the A. eudoxus ammonite Zone (Wignall and Pickering, 1993) and no “prograding wedge” can be identified. At Eathie (Fig. 17C), paleoecological work undertaken by these authors indicates a deepening in environment close to the R. cymodoce/A. mutabilis Zonal boundary, which may be equivalent to the “top basin floor fan surface” recognized at Kintradwell.

Distinct disparities therefore exist regarding the stratigraphic architecture of the middle Kimmeridgian successions of the IMF and eastern parts of the OMF basin. In the IMF, a type 1 sequence boundary is identified within the R. cymodoce Zone, with retrogradational stacking patterns beginning near the top of the R. uralensis Subzone. In the Witch Ground Graben, the additional age constraints afforded by Kimmeridgian ammonite recoveries (Brealey, 1990) indicates that shallow marine progradation continued from the R. cymodoce into the A. mutabilis Zone. Consequently, the middle Kimmeridgian basinward shift in facies identified in the IMF cannot be identified in this area of the OMF. The problem of sequence stratigraphic correlation is compounded by the fact that identification of a sequence boundary in the Witch Ground Graben within this cycle is equivocal. Any evidence for subtle changes in stacking patterns may have been stripped away by ravinement erosion as a result of the later subsequent rise in relative sea level. The maximum flooding surface that defines the top of this cycle is marked by a distinctive log motif (Fig. 16) but can only be convincingly identified in the OMF. The ammonites recovered from Block 15/21 by Brealey (1990) suggests that this surface is either A. mutabilis or A. eudoxus Zone, although the palynological recovery of the OMF is more consistent with an intra-A. mutabilis Zone age. In the IMF, this intra-A. mutabilis Zone maximum flooding surface cannot be distinguished and a composite Jb.6.8./6.10. cycle must be defined (Fig. 18).

Regressive/Transgressive Cycle Jb. 6.10.—This cycle is commonly missing in many of the footwall locations in the Witch Ground Graben because of erosion and/or nondeposition associated with Late Jurassic fault block rotation. In favourable structural settings, however, such as Block 15/21 in the Theta Graben, this sedimentary cycle is identifiable (Figs. 16, 19). A well-defined progradational to retrogradational trend characterizes this stratigraphic interval and is similar to those described previously. It is possible that in places the sequence boundary and transgressive surface are coincident, although no basinward shift in facies has been identified. Although an exact age for the turnaround surface cannot be obtained, an intra-A. eudoxus Zone age is preferred.

The late Kimmeridgian transgression had a profound effect on the paleogeographies of the eastern parts of the OMF. The remaining shallow-marine shorelines (e.g., Piper Formation) were drowned and the deposition of anoxic mudstones ensued (Kimmeridge Clay Formation, sensu Deegan and Scull, 1977). Only in the Claymore Field area (Block 14/19) and adjacent to the Renee Ridge (Block 15/30) are post-A. eudoxus Zone shallow-marine shoreline sediments encountered in the OMF (Riley, pers. commun., 1995). Within the lowermost part of the Kimmeridge Clay Formation, in well 15/21a-25, ammonites have been identified that correspond to the A. eudoxus ammonite Zone (Brealey, 1990), and these can be used to assign a late Kimmeridgian age to the maximum flooding surface that bounds the top of this cycle. Biostratigraphic information in conjunction with petrophysical log trends from wells in the IMF confirms the presence of a basinwide A. eudoxus Zone maximum flooding surface (Fig. 18). This concurs with the independent work of Wignall and Pickering (1993), who have shown that the lower to middle Kimmeridgian transgressive trend adjacent to the Helmsdale Fault culminated in maximum water depths in the A. eudoxus Zone.

**DISCUSSION AND CONCLUSIONS**

This paper improves upon previous Jurassic sequence stratigraphic studies undertaken in the MF basin by providing a rigorous documentation of the various hierarchies of sedimentary cycle present within this area (Figs. 18, 19). The Jurassic System comprises two major T/R cycles bounded by three basinwide surfaces of maximum regression which occur close to: (i) the Triassic/Jurassic boundary, (ii) the Lower/Middle Jurassic boundary (also termed the “Mid-Cimmerian” unconformity), and (iii) the S. stenomphalus/P. albidum Zonal boundary (uppermost Ryazanian; also termed the “Base Cretaceous” unconformity). The Lower Jurassic major T/R cycle comprises two 2nd-order T/R facies cycles bounded by three ubiquitous surfaces of maximum regression that occur: (i) close to the Triassic/Jurassic boundary, (ii) within the Sinemurian, and (iii) at the “Mid-Cimmerian” unconformity. The Middle Jurassic to Lower Cretaceous (Bathonian to Ryazanian) major T/R cycle comprises at least three 2nd-order T/R facies cycles. These cycles are bounded by four ubiquitous surfaces of maximum regression that occur: (i) at the “Mid-Cimmerian” unconformity, (ii) in upper Bathonian time (C. discus Zone or older), (iii) in upper middle Oxfordian time (either the C. tenuiserratum or C. blakei Subzone), and (iv) in middle Volgian time. Additional work is required to define the exact position and age of the fourth cycle boundary. All these surfaces of maximum regression are defined on the basis of maximum progradation of facies belts, indicative of minimal accommodation space in the basin margin settings.

Each T/R facies cycle has been divided into constituent higher-frequency cycles whenever possible. Because of difficulties in objectively defining and correlating higher-frequency sequence boundaries out of local proximal depositional settings, these cycles are bounded by maximum flooding surfaces that are used to define R/T cycles. Maximum flooding surfaces and their correlative hiatal surfaces have proven to be the most distinctive, reliable, and easy-to-date horizons on which to base a regional sequence stratigraphic framework at this hierarchical scale. Fourteen R/T cycles have been defined (Figs. 3, 18, 19). Lower Jurassic maximum flooding surfaces can be identified within the P. aplanatum Subzone (uppermost Sinemurian) and P. taylori Subzone (lowermost Pliensbachian). Two additional maximum flooding surfaces are also recognized within the lower and upper Pliensbachian Stage, although their zonal significance cannot yet be ascertained. Four Middle Jurassic maximum flooding surfaces are recognized: (i) close to the Bathonian/Callovian boundary, (ii) at the boundary between the S. calloviense and S. endodatum Subzones (lower Callovian), (iii) close to the middle/upper Callovian boundary within the K.
JURASSIC SEDIMENTARY CYCLES FROM THE MORAY FIRTH BASIN, UNITED KINGDOM NORTH SEA

Fig. 18.—Summary stratigraphic breakdown of the Jurassic succession (Hettangian to Kimmeridgian) encountered in the Inner Moray Firth well 12/21-4. In the IMF the intra-A. mutabilis Zone maximum flooding surface cannot be distinguished and a composite Jb.6.8./6.10. cycle has to be defined. An additional lower Oxfordian maximum flooding surface may occur at a depth of 6600ft.

<table>
<thead>
<tr>
<th>AGE</th>
<th>GAMMA RAY</th>
<th>DEPTH (FT)</th>
<th>SONIC</th>
</tr>
</thead>
<tbody>
<tr>
<td>KIMMERTIDIAN</td>
<td></td>
<td>0</td>
<td>150</td>
</tr>
<tr>
<td>L.M.</td>
<td></td>
<td>140</td>
<td>40</td>
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<tr>
<td>OXFORDIAN</td>
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<td>L.M.U.</td>
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<td>CALCVIAN</td>
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<td>BATHONIAN</td>
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<tr>
<td>PRIEBENTTO-SINEMIRIAN</td>
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<tr>
<td>TRIASSIC</td>
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</tbody>
</table>

**LITHO.**
- Unconsolidated sandstone
- Kimberidge Clay
- Heather
- Bovaird Coal
- Barretts
- Ficksburg
- Rovena Bay
- Van Coloured Sand
- Seiche Bay

**SED. CYCLE**
- G. jenkinsi 5900'N
- E. lindgren
- C. pumilum and P. pansoni and G. pumilum increase to calcareous-bioclastic transgressive
- C. pumilum
- S. crystallinum
- G. jenkinsi pumilum
- S. crystallinum
- R. claviformis
- R.-sulcata R. claviformis
- E. claviformis
- S. l. excubitor
- S. l. excubitor
- C. calatum
- S. l. calatum
- G. s. acidum
- G. s. acidum
- A. murchisoni
- A. murchisoni
- E. index
- E. index
- M. gracilis
- M. gracilis
- C. pensylvanicus
- C. pensylvanicus
- A. eudoxus
- A. eudoxus
- A. murchisoni
- A. murchisoni
- C. spinulosus
- C. spinulosus
- V. baylei/R. cymodoce Zonal boundary (lower Kimmeridgian), (vi) intra-A. mutabilis Zone (middle Kimmeridgian), and (vii) intra-A. eudoxus Zone (upper Kimmeridgian). Whenever possible, all other higher-frequency sequence stratigraphic surfaces have been dated and correlated.

It is beyond the scope of this paper to comprehensively compare and contrast the stratigraphic cycles documented in the MF basin with those of other areas. Interested readers should consult the references provided.
Fig. 19.—Summary stratigraphic breakdown of the Jurassic succession (Bajocian to Kimmeridgian) encountered in the Outer Moray Firth well 15/21a-15 (Scott Field).

The fact that key sequence stratigraphic surfaces can be correlated across the various North Sea basins that are known to possess differing tectonic histories suggests that local tectonics (e.g., individual fault block rotation) had little effect upon controlling the timing of cycle development at this hierarchical scale. This even appears to be the case during the Middle and Late Jurassic Period when extensional rifting exerted a profound control upon gross basin development. As such, the statement of Rawson and Riley (1982) that, “even in the tectonically active areas of the North Sea, the effects of eustatic...
sea level changes were never completely masked by local tectonics" appears to hold partly true. However, it is unsound to invoke a global mechanism from such a local dataset. Further detailed studies from beyond the North Sea and European domain are necessary in order to gain further insights into the mechanisms driving the formation of these stratigraphic cycles.

ACKNOWLEDGMENTS

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THIRD-ORDER SEQUENCES IN AN UPPER JURASSIC RIFT-RELATED SECOND-ORDER SEQUENCE, CENTRAL LUSITANIAN BASIN, PORTUGAL

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ABSTRACT: A Middle Oxfordian to Tithonian transgressive/regressive 2nd-order sequence is recognized over most of the Lusitanian Basin of Portugal. This paper describes the nature of this sequence and its constituent 3rd-order sequences in the Arruda subbasin situated 30 km north of Lisbon. Lack of good outcrops precludes the identification of 3rd-order sequences in the transgressive part of the 2nd-order sequence, but they are easily identified in the regressive part in four different tectonic settings.

The transgressive part of the 2nd-order sequence is related to rift movements that created the subbasin. During rift initiation, carbonate depositional systems dominated. These were drowned during the rift climax phase when footwall uplift caused local erosion and karstification, and the influx of coarse siliciclastic sediments near active faults. 2nd-order maximum flooding occurred during the Late Oxfordian bimammatum zone at which time the subbasin was relatively starved of sediment and was a deep depression.

Third-order sequences deposited during the immediate postrift phase (i.e., at the beginning of the 2nd-order regression) are aggradational lowstand arkosic submarine fan deposits. As accommodation was reduced by sedimentation, localized transgressive to highstand reefal carbonates formed on the shallow proximal part of the fan. The late postrift phase was heralded by progradational sequences consisting of lowstand fine-grained slope deposits capped by transgressive/highstand coral boundstones and oolites. These filled the basin virtually to sea level, so that succeeding 3rd-order sequences lacking lowstands developed in shallow, low-energy carbonate and siliciclastic fluvial facies.

Limited biostratigraphic control suggests that the 11 3rd-order sequences may be co-eval with those recognized elsewhere in Europe. If this is correct, the ages of these European sequences suggest that the rift climax event in the Arruda subbasin lasted only 1–2 my, and that subsidence rates approached 2 m/ky in the center of the subbasin.

INTRODUCTION

The Middle Oxfordian to Tithonian succession of the central part of the Lusitanian Basin contains several mixed siliciclastic-carbonate associations. These accumulated during extremely rapid rift-related basement subsidence and subsequent slow regional subsidence. This paper shows how the sequence stratigraphic approach provided new perspectives concerning the subsidence history and origin of the complex facies mosaic of one subbasin.

The Arruda subbasin, situated 35 km north of Lisbon, is a half-graben about 20 km wide that developed during the Late Oxfordian/earliest early Kimmeridgian Age (Figs. 1, 2). It was filled by four major depositional systems (Fig. 3): (1) carbonate buildups and associated deep-water sediments (Ellis et al., 1990; Leinfelder, 1994), (2) coarse-grained siliciclastic submarine fan (Leinfelder and Wilson, 1989), (3) southward prograding fine-grained siliciclastic slope capped by carbonates (Ellwood, 1987; Wilson, 1989; Leinfelder and Wilson, 1989; Nose, 1995), and (4) coastal plain and shelf (Leinfelder, 1986, 1987a). The distribution of these systems in space and time was largely controlled by changes in accommodation space caused by tectonism and sediment infilling, resulting in a 2nd-order transgressive/regressive sequence. Third-order sequences are recognized within the 2nd-order sequence in different tectonic settings in the subbasin.

The first section of the paper summarizes the tectonic setting and stratigraphic framework of the Arruda subbasin, after which (in the second section) the nature of the Late Jurassic transgressive/regressive 2nd-order sequence that fills it is discussed. The seismic and sedimentologic features of the rift initiation, rift climax, immediate postrift and late postrift phases of the subbasin’s development are documented. The third section of the paper describes the nature of 3rd-order sequences...
that occur in the regressive part of the 2nd-order sequence in four distinct tectonic settings within the subbasin. In light of the probable correlation between the 3rd-order sequences in the four tectonic settings, the final section of the paper discusses the timing of rifting and possible correlation with sequences recognized elsewhere in Europe.

GEOPHYSICAL FRAMEWORK

Tectonic Setting

Three subbasins occur in the central part of the Lusitanian Basin to the north of Lisbon (Fig. 1). They began to develop during the mid-Oxfordian and probably ceased to exist as separate subbasins by the end of the Late Jurassic Epoch. Wilson et al. (1989) suggested that the transtensional rifting episode that created them was the precursor to Late Jurassic ocean spreading in the Tagus Abyssal Plain to the west.

The half-graben structure of the Arruda subbasin is known largely from seismic data (Fig. 2A). It is separated from the Bombarral subbasin by a saddle formed by the Torres Vedras-Montejunto anticline. This anticline (Fig. 2B) was initiated during the Late Jurassic as a salt pillow structure that was further deformed by Miocene transpressional movements (Wilson et al., 1989). The west side of the subbasin is bounded by the Runa fault zone, the northern sector of which consists of a graben in which Upper Cretaceous sediments and volcanics are preserved (Fig. 4). On seismic lines, this zone is seen to be underlain by a major westward-dipping normal fault that forms the eastern margin of the Turcifal subbasin. A piercement diapir occurs at the north end of the fault zone. The eastern margin of the subbasin is a complex zone consisting of two horsts separated by a probable transfer zone. Miocene inversion produced a broad low-amplitude domal structure (the Arruda anticline, the crest of which is situated in the area around Arruda #1) above the thick fill of Upper Jurassic sediments. The location of the southern limit of the subbasin is unknown, because no seismic data have been shot in this area.
**Stratigraphy**

Wilson et al. (1989) recognized four megasequences in the Mesozoic succession of the Lusitanian Basin and linked them as follows to events in the opening of the Atlantic:

4. **Late Aptian—Campanian**: Ocean spreading around west and north Iberian margins.
3. **Valanginian—Early Aptian**: Rifting around north and northwest Iberia.
2. **Middle Oxfordian—Early Berriasian**: Rifting, and ocean spreading beneath the Tagus abyssal plain.
1. **Triassic—Callovian**: Triassic rifting and later thermal subsidence, but no ocean opening.

Figure 5 shows the lithostratigraphic units within the second and part of the third megasequence that are discussed in this paper. During the Triassic and earliest Jurassic periods, movements along Hercynian basement faults produced basins that were filled with red siliciclastics (Silves formation*) and evaporites (Dagorda formation). The latter are relatively thin in the study area but, to the north beneath the Bombarral subbasin, were thick enough to be mobilized to produce salt structures (Fig. 1). The Triassic and Hettangian sediments accumulated in grabens and half-grabens, but the younger Lower and Middle Jurassic sediments (Brenha and Candeeiros formations) blanketed the entire Lusitanian Basin and exhibit simple facies geometries indicative of a westerly inclined carbonate ramp system.

The early part of the second megasequence is characterized by extremely high apparent basement subsidence rates (Wilson et al., 1989) and (apart from the basal Cabacós formation) major lateral changes in facies (Figs. 3A, B). Its base is marked by a basin-wide hiatus spanning latest Callovian to Early Oxfordian time. The rest of the second megasequence is described in the next section of the paper. Megasequence 3 shows a relatively simple facies distribution, with dominantly fluvial siliciclastics of the Torres Vedras formation interfingering southwards and westwards with the marine carbonates of the Cascais formation (Fig. 5).

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*The Term ‘formation’ is not shown with a capital ‘F’ in this text, as the stratigraphic scheme shown in Figure 5 is informal.*
history is described in detail, and interpreted in terms of tectonic systems tracts defined by Prosser (1993). The former part of this section of the paper provides a brief summary of the late Jurassic tectono-sedimentary history of the subbasin. Figure 6 shows the relationships between the lithostratigraphic units that comprise the transgressive/regressive upper Jurassic sedimentary fill of the Arruda subbasin.

In many places in the Lusitanian Basin, the top of the Middle Jurassic succession is marked by karstification, rubification and caliche formation (Felber et al., 1982; Leinfelder, 1983; Ruget Perrot, 1961; Wright and Wilson, 1987). The period of exposure indicated by these features is linked to the presumed hiatus extending from the topmost part of the Callovian to the top of the Lower Oxfordian. This hiatus, the increased subsidence rates during the Oxfordian and early Kimmeridgian Stages compared with those during the Early and Middle Jurassic Epochs, and the change from shelf carbonates at the top of the Middle Jurassic formation to lacustrine/marginal marine carbonates of the Cabaços formation, all indicate a significant change in the tectonic framework of the Lusitanian Basin (Wilson et al., 1989).

Lacustrine and marginal marine carbonates of the Cabaços formation were deposited over the entire Lusitanian Basin during early Middle Oxfordian time. Thickness variations of this formation and the overlying marine carbonates of the Montejunto formation show that differential subsidence resulted in the formation of separate subbasins. Rapid deepening occurred over much of the Arruda subbasin during the deposition of the Montejunto formation, except along its eastern margin where shallow water carbonate buildups formed (Fig. 3A).

There was a sudden influx of siliciclastic material all over the Lusitanian Basin during the Late Oxfordian bimammatum zone. In the Montejunto area on the northwest flank of the Arruda subbasin, this change occurs at the boundary between the Montejunto formation and the Tojeira member of the Abadia formation. Foraminiferal evidence suggests that the greatest water depths occurred during deposition of the Tojeira member (Stam, 1985), suggesting that the 2nd-order maximum flooding event occurred during the bimammatum zone.

The bulk of the sedimentary fill of the Arruda subbasin consists of the Custanheira member, which reaches a thickness of over 2 km beneath Arruda dos Vinhos. The lateral equivalent of this unit in the Montejunto area is the Mirante member, which is only about 50 m thick. Both members consist of arkosic sandstones and gravels.

The uppermost unnamed member of the Abadia formation consists of siliciclastic mudstones, shales and siltstones with subsidiary sandstones. On seismic sections, it corresponds to a cliniform reflection package (Wilson, 1989: Leinfelder and Wilson et al., 1989) indicating southward progradation of a slope
THIRD-ORDER SEQUENCES IN AN UPPER JURASSIC RIFT-RELATED SECOND-ORDER SEQUENCE

Fig. 6.—Lithostratigraphic cross section showing the nature of the 2nd-order transgressive-regressive sedimentary fill of the Arruda subbasin. The nomenclature is informal: formations are shown in capitals, and members in lower case. Text in italics indicates depositional environment. (1): marls with siliciclastic and carbonate turbidites and debris flows with allochthonous shallow water karstified limestone blocks (80 m); (2): sandstones and conglomerates with limestone clasts (10–30 m) in north-south channel 3.5 km wide; (3): euryhaline limestones and marls; CR: relic of Castanheira reef.

Our work indicates that the sedimentary fill of the Arruda subbasin exhibits many of the characteristic features described for rift basins by Prosser (1993), allowing for the following characteristics that are not encompassed by her model. (1) carbonates formed the first phase of the basin fill, (2) the growth of the salt pillow beneath the present-day Torres Vedras-Montejunto anticline caused reflector convergence onto this structure, and (3) there is no evidence for the derivation of sediments from the hanging wall of the Arruda half-graben.

As will be shown in the final section of the paper, integrating the 3rd-order sequence stratigraphy with Prosser’s (1993) tectonic systems tract model suggests that the period of rifting was relatively short (1–2 my) in the Arruda subbasin.
such movements resulted in a radical reorganization of the distribution of carbonate depositional systems, from a prerift westward-dipping ramp to buildups forming over fault or diapirically controlled highs during rift initiation.

On seismic sections across the Arruda subbasin, the Oxfordian carbonates thicken towards the southeast and east due to movement along the eastern boundary fault complex, and the growth of a salt pillow beneath the present day Montejunto-Torres Vedras anticline. Such thickening is characteristic of the rift initiation and rift climax phases of Prosser (1993). However, as the sedimentary record shows no evidence of either significant subaerial exposure or the erosion of the thick argillaceous and carbonate-rich Lower and Middle Jurassic formations, the carbonate deposition is interpreted to represent rift initiation rather than rift climax.

Ellis et al. (1990) and Leinfelder (1994) recognized several types of buildups in the Lusitanian Basin, based on their facies characteristics and tectonic setting. Fault-controlled buildups occur on the east side of the basin. They exhibit shelf profiles, are relatively thin (200–500 m), show well-developed lateral facies zonation and are dominated by lime mudstones and wackestones, with lesser amounts of packstones, grainstones and boundstones. In the Arruda subbasin, buildups of this type occur along its eastern margin from Montejunto in the north (late Oxfordian, possibly extending into the early Kimmeridgian), to the south at Ota (Kimmeridgian) and in the subsurface in Montalegre #1 (late Oxfordian). In the western Montejunto area, basinal deep-water lateral equivalents are exposed and were encountered in Benfeito #1 and Sobral #1 (for locations of these boreholes, see Figs. 2B, 4), and they are presumed to extend beneath the entire subbasin. There are no occurrences of the salt-controlled buildups described by Ellis et al. (1990) in the study area.

The lacustrine marginal marine Cabacós formation and deep-water part of the Montejunto formation show semi-continuous low-amplitude reflections that diverge towards the center of the Arruda subbasin. The strong double reflector at the base of this reflection package (Fig. 7) is caused by a mixed anhydrite/carbonate unit in the lower part of the predominantly lacustrine Cabaços formation. At the eastern margin of the subbasin, there is an area of chaotic reflections (Fig. 7A) that Leinfelder and Wilson (1989) interpreted as being caused by the massive coarse-grained facies in the proximal part of the Castanheira fan (Fig. 2B). On most seismic lines, the anhydrite and top Montejunto reflectors fade out into the chaotic zone (Fig. 7B), which suggests that seismic energy is dissipated by the overlying proximal fan facies of the Castanheira member of the Abadia formation.

After the episode of emergence that resulted in the late Callovian—early Oxfordian age hiatus, relative sea-level then rose sufficiently to produce the lacustrine and marginal marine conditions in which the Cabaços formation was deposited. Accommodation then increased rapidly so that hemipelagic carbonates of the Montejunto formation were deposited over most of the subbasin. However, in the northern part of the eastern boundary fault complex in the east of the Montejunto area, a carbonate buildup developed over the footwall (Ellis et al., 1990). It is probable that faults did not break through to the surface at this time (as no coarse carbonate slope deposits indicative of exposed scarps are present), but that fault-tip folds produced depositional profiles upon which shallow-water carbonate buildups could develop over the higher fold limb.

**Rift Climax**

During the rift climax phase, the maximum rate of fault displacement occurs, and “so sedimentation is likely to be outpaced by subsidence and differential relief will be created across the fault scarp” (Prosser, 1993). During this phase, basins are likely to be sediment-starved, as rates of subsidence are very high compared to erosion and new drainage systems have not become established.

In the Arruda subbasin, the sedimentological features of the Tojeira member indicate that it was deposited during the rift
climax phase. It is only exposed in the Montejunto area, where it spans the upper part of the *bimammatum* zone and the *planula* zone of the late Oxfordian (Atrops and Marques, 1986, 1988). The member consists of an alternation of marls and thin (up to 50 cm) turbidites composed of arkosic and carbonate sand. Bedded micritic ammonite-bearing limestones similar to those of the Montejunto formation are present in places. Occasional debris flow units also occur, not only containing siliciclastic pebbles and sand, but also allochthonous shallow-water carbonate blocks up to house size, some of which exhibit karstification. The shedding of large blocks of karstified shallow-water carbonates suggests significant footwall uplift causing subaqueous diageneisis and erosion. For the first time since Late Triassic time, coarse-grained Hercynian basement material appears in sediments, suggesting that it was exposed in nearby footwall blocks.

On seismic sections, the interval interpreted to be the lateral equivalent of the Tojeira member thickens significantly towards the eastern boundary fault zone (Fig. 7), which together with the observed sedimentological features described above, is consistent with the interpretation that it represents the rift climax systems tract.

**Immediate Postrift**

At the immediate postrift stage in rift basin development, differential subsidence across boundary faults ceases, but the basin continues to subside due to lithospheric cooling. Prosser (1993) suggested that this phase of basin development is characterized by a change from divergent to parallel reflections as fault-block tilting ceases and that this is accompanied by strong onlap updip on the hanging wall and possible downlap in the center of the basin.

Relatively continuous high-amplitude reflections showing convergence towards the western and northwestern margins of the Arruda subbasin characterize the arkosic sands and gravels of the Castanheira member. Some onlap is seen at the base of this reflector package on the southeastern flank of the Torres Vedras-Montejunto anticline. In the Arruda subbasin, seismic sections do not show a marked change from a divergent to nondivergent reflection package characteristic of the rift-climax to immediate postrift transition. However, there is a significant change in the degree of divergence (Fig. 7) that probably occurs (it has not been drilled) at or near the base of the arkosic sands of the Castanheira member. The divergent reflection pattern in the immediate postrift systems tract shown on Fig. 7 is interpreted as being due in part to the growth of a salt pillow beneath the present day Montejunto-Torres Vedras anticline and to thinning of sediments away from the source of the submarine fan across the probable transfer zone within the eastern boundary fault zone. Therefore, there is both a halokinetic and sedimentary overprint on the seismic character of the immediate postrift systems tract.

On the eastern margin of the Arruda subbasin, about 350 m of coarse arkosic sandstones and conglomerates of the top part of the Castanheira member are exposed. A total minimum thickness of 2200 m was proved by Arruda #1, which did not penetrate the base of the member. The sandstones contain pebbles and cobbles of granite, gneiss, slate, quartzite and vein quartz, and in places carbonate clasts occur. The massive structural features sometimes contain large reefal blocks. Claystone boulders up to 8 m across also occur in which the original bedding is sometimes orientated vertically. These features indicate deposition by debris flows. In the finer facies, amalgamated channels occur, showing fining-upward trends; these suggest deposition from turbiditic flows. Fine-grained shaly intervals in the cores from Arruda #1 contain marine palynomorphs. Evidence from cores suggests an overall coarsening-upward trend. The ammonite *Ardescia pseudolictor*, found at outcrop near the top of the member, indicates a middle early Kimmeridgian age (middle part of *hypselocyclum* zone, Leinfelder, 1994).

On the basis of outcrop, borehole, and seismic data, Leinfelder and Wilson (1989) interpreted the Castanheira member as a submarine fan supplied with sediment through a gap in the eastern boundary fault complex (Fig. 2B). The zone of chaotic reflections on the eastern margin of the subbasin was interpreted by them to be caused by the proximal massive and coarse-grained part of the fan system.

**Late Postrift**

Prosser (1993) stated that the late postrift systems tract is the result of the continued peneplanation of topography caused by faulting and further filling of accommodation space formed by earlier tectonism. Parallel reflections are likely, though they may show some divergence towards the basin center due to compaction of previously deposited sediments. Erosion of fault block crests may result in the development of fining-up sequences, and eustatic sea-level changes are now more likely to control sediment input rates and change the amount of accommodation available.

The top part of the Castanheira member shows a fining-up trend, and contains two reefal intervals related to 3rd-order sequences (see below) that is identified as the basal late postrift systems tract. The uppermost unnamed part of the Abadia formation, the Ota limestone and Amaral, Lourinhã and Farta Pão formations, represent the remainder of this tract.

The Ota limestone cannot be distinguished on seismic sections, because a zone of chaotic reflections (typical of seismic data in many places in the Lusitanian Basin where thick carbonates occur at or near the surface) characterizes the horst zone. It is contemporaneous with the upper part of the Castanheira member and the prograding top part of the Abadia formation (Ellis et al., 1990; Leinfelder et al., 1988; Leinfelder and Wilson, 1989). It developed as a narrow reef-rimmed platform on a horst on the eastern margin of the Arruda subbasin (Leinfelder and Wilson, 1989; Leinfelder, 1992, 1994). The aggradational geometry of the platform and its facies zonation are the result of its growth on top of the a horst formed during earlier rifting. No reef talus or deeper-water sediments are exposed; they are presumed to have developed at the foot of the fault scarp and, if present, are now buried by, or interfering with, the Castanheira and uppermost members of the Abadia formation. The Ota limestone represents the immediate postrift systems tract developed in a siliciclastic-starved setting above the footwall on the northern segment of the Arruda subbasin boundary fault system.

The southward-prograding clinoforms seen on seismic sections (Wilson, 1989; Leinfelder and Wilson, 1989) show clearly
that the siltstones and marls comprising the uppermost unnamed member of the Abadia formation were deposited in a prograding slope setting. The presence of turbiditic sandstones, slumped horizons, mud pebble breccia beds and reseidmented ooid grainstones seen at outcrop are consistent with this interpretation. Occasional higher-amplitude reflections can be traced downdip into the continuous high-amplitude reflections of the underlying submarine fan deposits of the Castanheira member. The transition between the two seismic facies units occurs at successively higher reflections in the older package, so that as the clinoform unit thins southwards, it exhibits a kind of “climbing downlap” relationship with the underlying unit. This indicates that the submarine fan system (sourced from the east) continued to be deposited as the slope system prograded southwards.

Outcrop and well data show that the slope and shelf sediments extend over much of the central part of the Lusitanian Basin. In the Bombarral and Turcifal subbasins and along the northern and eastern margins of the Arruda subbasin, these sediments are about 550 m thick but thin to 60 m in the vicinity of Castanheira, due presumably to lateral replacement by the Castanheira member. Unlike the deposits of the first two depositional systems, they do not show significant thickness changes over major tectonic structures, except in the northern part of the eastern boundary fault complex, where they are replaced by the Ota limestone. Throughout the southern part of the Lusitanian Basin the Amaral formation is recognizable on seismic sections as a strong reflection capping the clinoform reflection package at the top of the Abadia formation. It was deposited in a shallow high-energy shelf setting on top of the southward-prograding marls and siltstones. In the Arruda subbasin, the Amaral formation contains lenses of coral boundstones (some of which are thrombolitic) overlain by ooid grainstones (Leinfelder et al., 1993b; Nose, 1995). Patches of grainstones that occur above the karstified top of the Ota limestone are relics of the Amaral formation (Leinfelder, 1994). This is the earliest occurrence of a lithostatigraphic unit extending from the Arruda subbasin across onto the hangingwall of the boundary fault system. This indicates that the basin was “full” and beginning to overflow across its eastern margin and that the depositional profile was no longer related to topography produced during the rift climax phase. However, the effects of differential compaction across the fault resulted in the Ota block remaining slightly higher during the deposition of the Lourinhã and Farta Pão formations (Leinfelder, 1985).

The prograding slope and shelf system is overlain by mostly red siliciclastic fluvial deposits of the Lourinhã formation (Hill, 1989). Only at the base are deltaic sediments present (Sobral member) that are probably latest Kimmeridgian in age (Leinfelder, 1986), possibly extending into the earliest Tithonian (Manuppella, pers. comm. 1994). Immediately to the east of the Ota platform, freshwater oncocid horizons occur that formed in spring-fed streams and lakes (Leinfelder, 1985). The Arranhó member of the Farta Pão formation consists of limestones and marls with a rich fauna of euryhaline to partially brackish bivalves in the lower part and coral biostromes in its upper part. The Arranhó member occurs only in the southern part of the Arruda subbasin. The overlying Freixial member also consists of limestones, marls and sandstones with a euryhaline to brackish fauna (Leinfelder, 1986, 1987a).

The Lourinhã and Farta Pão formations are situated near the surface over most of the Arruda subbasin, so that their true seismic characters usually are not well recorded. Where clear reflection characteristics of the Lourinhã formation can be discerned, moderate to strong parallel discontinuous reflectors occur, consistent with the presence of fluvial or deltaic sand bodies of limited lateral extent.

The relatively simple facies relationships in the Lourinhã and Farta Pão formations, with fluvial sediments being replaced southwards by marine limestones and marls, contrasts with the complexity of earlier facies distributions in the subbasin. This simple pattern, which extends across the Lusitanian Basin to the present-day coastline, indicates that by Tithonian time, depositional profiles were no longer linked to rift-related subsidence or topography.

Summary of Tectono-Sedimentary History

The transgressive/regressive 2nd-order sequence of the Arruda subbasin was produced by a combination of rapid rates of subsidence during the late Middle and Late Oxfordian rift initiation and rift climax episodes and the subsequent reduction of accommodation space by sedimentation.

The nature and location of the depositional systems filled by the subbasin were controlled by the topography produced during rifting and later erosion and burial as new drainage systems were established. During initial rifting, differential subsidence across faults was not sufficient to cause exposure above sea level, but it did result in the deposition of aggradational carbonate buildups along the eastern margin of the subbasin. Strong relief was produced during the rift climax, but only the crests of hanging walls were emergent or covered by shallow-marine water, after which the subbasin was a large depression waiting to be filled once new drainage systems had developed. During the immediate postrift phase, coarse-grained sediment was transported via a transfer zone in its eastern boundary fault zone by high-density turbidity flows. The final filling of the subbasin virtually to base-level was accomplished by a southward-prograding fine-grained slope-shelf system that in terms of Eliet and Gawthorpe’s (1995) drainage domains, was axial. By middle late Kimmeridgian times, the accommodation space created during the rifting episode and later subsidence caused by sediment loading and/or lithospheric cooling had virtually been eliminated, and so the remainder of the subbasin fill was deposited in fluvial (axial) and shelf settings. The footwall and hangingwall drainage domains of Eliet and Gawthorpe (1995) were not developed, though their fifth, karst, domain probably developed over a small area associated with the Ota platform.

The rift initiation and climax phases of development of the Arruda subbasin represent the transgressive phase. Foraminiferal studies (Stam, 1985) suggest that deepest water conditions are represented by the Tojeira formation, in which case the 2nd-order maximum flooding surface occurs within the bimammatum zone.

The following episodes in the tectono-sedimentary history of the Arruda subbasin can be distinguished (see Table 1 for a summary of these).
Rift Initiation (Cabaços and Montejunto formations).—

Regional uplift resulted in the late Callovian–Early Oxfordian hiatus. Later fault movements in the basement produced flexures in the overlying Triassic–Middle Jurassic cover, which in places triggered salt migration. This led to the development of subbasins and shallow-water carbonate buildups on the elevated parts of the flexures or salt highs. The change from the lacustrine carbonates and evaporites of the Cabaços formation to the shallow- and deep-water limestones of the Montejunto formation indicates a relative rise in sea level during Middle Oxfordian time.

Rift Climax (Tojeira member).—

During Late Oxfordian time, differential subsidence was accentuated, and fault scarps were formed. Hangingwall subsidence drowned older carbonate systems, and footwall uplift resulted in meteoric diagenesis and shedding of blocks of older carbonates into the basin. The sudden influx of basement-derived clastics into the area indicates significant fault-related topography to the east and northeast.

Immediate Postrift (Castanheira and Mirante members).—

Reflective divergence towards the footwall of the eastern boundary fault complex is probably not indicative of continued displacement but is due to the influx of sediment through a transfer zone between two fault segments along the eastern margin of the Arruda subbasin. Thus the initial “hole” produced during the rift-climax was only filled when a new drainage system was established that could transport basement-derived debris across presumed fault-related topography situated to the east and northeast. Large thickness variations (~50–2000 m) resulted from the infilling of fault-related topography.

Late Postrift (Top Castanheira and Uppermost Unnamed Member of Abadia formation, Amaral, Lourinhã and Farto Pão formations).—

There is little variation in the thickness of these units, as by now only the Ota platform remained as a significant high along the eastern boundary fault zone. This area was starved of siliciclastic sediments, and a reef-fmbed carbonate system developed over it at the same time as the fine-grained siliciclastic slope system prograded southwards. Rift-related topography was finally eliminated (apart from the effects of differential compaction) by the time the Amaral formation was deposited.

**Montejunto—Torres Vedras Anticline**

Superficially, the Oxfordian carbonate succession in this area appears to be one sequence (Fig. 8). A subaerial unconformity or hiatus occurs at the base and is overlain by lowstand lacus-

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**UPPER OXFORDIAN TO TITHONIAN 3RD-ORDER SEQUENCES**

In this section, the 3rd-order sequence stratigraphic interpretation of successions developed in the regressive part of the 2nd-order sequence is discussed. The successions occur in four contrasting tectonic settings within the Arruda subbasin:

- **Montejunto—Torres Vedras Anticline**: salt pillow on north-west margin of the subbasin
- **Arruda area**: depocenter of the Arruda half-graben
- **Castanheira—Vila Franca de Xira**: transfer zone on eastern margin of the subbasin
- **Ota carbonate platform**: crestal position immediately to the east of the footwall of the northern part of the eastern boundary fault system

Biostratigraphic correlation of sequences in the four areas is combined to provide a 3rd-order sequence stratigraphic framework for the entire subbasin. This is used to add precision to the probable ages of the tectonic phases already described. We also describe the way in which the nature of 3rd-order sequences changed as the subbasin was progressively filled. The descriptions and interpretations are divided into sections dealing with the nature of sequence boundaries (SB) and lowstand (LST), transgressive (TST) and highstand (HST) depositional systems tracts. The constituent parts of the sequences are named according to their location (e.g. SB M1, LST M1, etc; M: Montejunto).
trine and marginal marine deposits with evaporites of the Cabaços formation. Above this shelf and deeper water carbonates of the Montejunto formation suggest transgressive and highstand depositional systems. However, we believe that it is likely that several 3rd-order sequences are present in this succession, but poor outcrops and tectonic complications make it difficult to piece together an accurate lithological succession. Therefore the 3rd-order sequence interpretation presented here begins at the base of the Tojeira member.

**Sequence M1.**

**SB M1.**—As described earlier, the Tojeira member consists of shales and marls with siliciclastic and carbonate turbidites and debris flows containing allochthonous shallow-water limestones. The contact with the underlying deep-water carbonate facies of the Montejunto formation is erosional in places (Ellwood, 1987), which, together with the abrupt facies change across the boundary, indicates a sequence boundary.

**LST M1.**—On the northern side of the Montejunto massif, to the southeast of Pragaça, the Tojeira member contains a basal carbonate rudstone package at least 20 m thick, which is composed of clasts and boulders derived from the Montejunto carbonate platform to the east. Some of the boulders were karstified prior to transport. The sediments grade upward into a marl succession with intercalated beds of allochthonous grainstones and packstones. To the south of Montejunto, these allochthonous intercalations are largely absent and are replaced by ammonite-bearing lime mudstones that we interpret as being a more distal facies deposited at a greater distance from fault scarps. In the Torres Vedras and Abadia areas, the lateral equivalent of the Tojeira member consists of marls similar to those that occur in the upper clinoform part of the Abadia formation.

**Age of Sequence M1.**—Ammonites from the Tojeira member indicate a Late Oxfordian age (late *binammatum* to *planula* zone, Atrops and Marques, 1986, 1988).

**Sequence M2.**

**SB M2.**—In the Montejunto area, Ellwood (1987) described sandstones and conglomerates (which he named the Mirante conglomerate member) abruptly overlying marls. This facies shift, and the local erosional base to the member, indicates the presence of a sequence boundary.

**LST M2.**—This system tract comprises the Mirante conglomerate member and the overlying succession of Abadia formation marls, which are over 500 m thick.

According to Ellwood (1987), the Mirante member ranges in thickness from about 20 m to 60 m. It contains clasts of quartzites, schists and granite up to 30 cm across, and limestone clasts over 1 m in diameter. The sand fraction contains abundant fresh feldspar crystals up to 0.5 cm long. The clast composition is very similar to that of the Castanheira member that reaches a thickness of over 2200 m in the deepest part of the Arruda subbasin, 25 km to the south. Ellwood (1987) suggested that the Mirante member was deposited in a north-south-oriented channel at least 3.5 km wide, through which sediment was transported southwards towards the depocenter around Arruda dos Vinhos, where the material met with coarse Castanheira fan siliciclastics derived from the east (as indicated by poorly developed clast imbrication and cross stratification).

The Mirante member is overlain by Abadia formation marls that display south-dipping clinoform reflectors on seismic sections. The marls contain occasional sharp-based sandstones less than 1 m thick. These show flute and groove casts and contain Bouma sequences $t_{dc}$ or $t_{e}$ or only parallel lamination or current ripple cross stratification. Rare paraconglomerate intervals occur, containing large limestone boulders showing boundstone lithologies similar to those present in the overlying Amaral formation, 20 km to the south (see below), but which do not occur in this formation in the Montejunto area.

As discussed below, we believe that no transgressive or highstand deposits are preserved in Sequence M2.

**Age of Sequence M2.**—Atrops and Marques (1986) reported *platynota* zone and lowermost *hypselocyclus* zone ammonites from the marls above the Tojeira member, indicating that Sequence M2 commences in the latter zone. Mouterde et al. (1972) assigned the marls above the Mirante member to the *divisum* zone, but in the Montejunto area it is not possible to constrain the age of the top of the Abadia marls.

**Sequence M3.**

**SB M3.**—The sequence boundary is identified at the base of the Amaral formation. In the Montejunto area, this formation is only about 10 m thick, compared with thicknesses approaching 100 m around Torres Vedras and in the Arruda area. It consists of marls, sandstones and ooid grainstones, and they could be interpreted as a transgressive systems tract at the top of Sequence M2. However, viewed in a regional context, the absence of the coral-rich boundstone facies characteristic of the topmost HST and earliest TST above the LST Abadia marls in the Arruda area (see below) and the sharp contact between slope marls and ooid grainstone described by Ellwood (1987) suggests the presence of a sequence boundary. An alternative interpretation is that the Amaral represents a carbonate-dominated TST/HST of the underlying sequence (Leinfelder, 1993) overlying siliciclastic lowstand sediments (Mirante member and overlying LST M2 marls).

**TST M3.**—Based on Ellwood’s (1987) description of the Amaral formation in the Montejunto area, we interpret it as a TST deposit. A 1-m-thick sharp-based ooid grainstone with low-angle truncation surfaces marks the base of the formation where it is best exposed on the south side of the Montejunto anticline at Portela de Sol, some 3 km east of Vila Verde dos Francos. The remainder of the formation consists of five coarsening-up 1.0 to 3.5-m-thick parasequences. These contain marls interbedded with fine-grained sandstones that show an upward increase in thickness. The intervals are capped by a sandstone and/or a fine to medium ooid grainstone. Ellwood suggested that the sharp base of the Amaral formation indicates that the wave-base was shallow (<5 m) and/or that there was a rapid change of slope at the shelf edge. This would explain the absence of hummocky, swaley and wave-ripple cross stratification. The sequence stratigraphic approach leads to an alternative interpretation: the expected wave-dominated shoreline transitional deposits between the top of the Abadia and the Amaral formations were eroded prior to deposition of the TST of Sequence M3 (i.e., during lowstand or as the transgression began). This interpretation also explains the absence of boundstone facies at the top of the Abadia formation that occur further south in the Arruda area and their occurrence as resedimented blocks.
in the LST (clinoform Abadia marls) of sequence M2 in the Montejunto area.

The nature of the contact between the Amaral and Lourinhã formations is not exposed in the Montejunto area. Therefore, it is not possible to identify the next sequence boundary, or to decide whether the fluvial sediments of the Lourinhã represent HST deposits of Sequence M3, or a younger sequence.

**Age of Sequence M3.**—The Amaral formation has so far yielded no biostratigraphically significant fossils, so its age can only be inferred from regional stratigraphic considerations. The youngest known ammonites found in the Abadia formation are indicative of the *divisum* and *acanthicum* zones (Ruget Perrot, 1961; Mouterde et al., 1972; Atrops and Marques, 1988). Younger ammonites giving a basal Tithonian age occur in the top part of the Sobral and the base of the Arranhol formation (cf. Ruget-Perrot 1961; Atrops and Marques 1986; Leinfelder et al., 1993a). Thus, it is probable that the Amaral formation in the Montejunto area represents the Late Kimmeridgian *eudoxus* zone.

**The Arruda Area**

In this area, the boundary between the Oxfordian and Kimmeridgian strata is not exposed. Subsurface data indicate that this area was the deepest part of the Arruda subbasin and is now filled with over 2200 m of arkosic sands and gravels of the Castanheira member (Leinfelder and Wilson, 1989). Above the Castanheira member there occur lithostratigraphic units broadly comparable to the clinoform Abadia marls (but only ~200 m thick in this area) and the Amaral and Lourinhã formations in the Montejunto area (Fig. 9). In addition, marine carbonates of the Farta Paão formation interfinger northwards with the Lourinhã formation.

**Sequence A1.**—

**LST A1.**—The base of Sequence A1 cannot be seen at outcrop. Arruda #1 showed that over 2200 m of arkosic sandstones and gravels of the Castanheira member lie beneath the top few tens of meters of it that are exposed at the surface. This member is interpreted as a LST fan. The overlying slope marls and siltstones of the top part of the Abadia formation floor much of the wide Arruda valley but are not well exposed. Occasional slump horizons, mud pebble layers and beds rich in lignite debris occur. The Abadia marls are interpreted as a prograding LST wedge.

**TST and HST A1.**—Within the Abadia marls, about 30–40 m below the Amaral formation, a distinctive condensed section is relatively well exposed at Serra Isabel (Fig. 10) to the north of Arruda dos Francos. It can be followed in an east-west direction along the strike of the prograding top Abadia slope system. This condensed interval is named the Serra Isabel unit and is up to 10 m thick. It consists of marly limestones with a rich benthic fauna often stained by iron hydroxides. Ammonite-rich beds are present, as are coral and crinoid meadows, and microbial thrombolite reefs up to 7 m thick that in places contain corals and siliceous sponges (Leinfelder et al., 1993a; Werner et al., 1994). Extremely low rates of background sedimentation are indicated by these features, particularly the microbial crusts. These low rates suggest rising relative sea level and the deposition of a TST. The occurrence of thrombolites lacking corals in a fairly shallow-water setting, the abundance of authigenic glauconite, and the presence of clusters of the dysaerobic bivalve *Aulacomyella* indicate that sea-level rise was accompanied by oxygen depletion in shallow water (Leinfelder et al., 1993b). The Abadia marls above the Serra Isabel condensed unit contain some low-diversity coral meadows (Nose, 1995) and are interpreted as the highstand.

**Age of Sequence A1.**—Ammonites from the Serra Isabel unit indicate that the top of the sequence may span the top *hypsioyclum* to *divisum* zones (Leinfelder et al., 1993a, b).

**Sequence A2.**—

**SB A2 and LST A2.**—Near the top of the Abadia marls, the occurrence of a series of oolitic sandstone channels marks a facies shift indicative of a sequence boundary (Nose, 1995).
TST A2.—Coral thrombolite bioherms and pure thrombolitic bioherms occur within a marl succession immediately above the sandstone channels and grade into a discontinuous carbonate coral bioherm/biostrome level of the lower Amaral formation that is interpreted as a flooding surface. This lower biohermal level is overlain by bioclastic limestones and oolitic carbonates. This parasequence is in turn overlain by a second, more prominent and continuous parasequence occurring over much of the Arruda area. It is composed of mostly crust-rich, coral bioherms and biostromes, representing the maximum flooding event. Frequently, the lower coral level is not developed and the upper coral limestones of the Amaral formation are in sharp contact with the underlying Abadia marls. This probably represents a ravinement surface within the TST A2. Alternatively, Nose (1995) discussed the existence of an additional depositional sequence, based on the fact that erosional channels may also occur locally at the base of the upper coral limestones. However, the laterally discontinuous character of the lower coral limestones and the probable short duration of its deposition are arguments in favor of a parasequence interpretation. The upper coral limestones tend to become more bioclastic towards the top (Nose, 1995), which is interpreted as early highstand deposits of HST A2 (Fig. 10).

Sequence A3.—

SB A3.—Within the upper coral limestones, a discontinuous level exhibiting karst features is present. Sometimes this shows greenish clays containing exclusively calcitic fossils. The enrichment of calcitic fossils (brachiopods, crinoids, pectinid bivalves) and the disappearance of the normally dominant aragonitic elements (particularly corals) is interpreted to be due to early meteoric dissolution of aragonite along an intraforma-}

![Diagram](image_url)
upper Arranhó and the tongue of the fluvial Lourinhã formation are interpreted as the TST and HST. There are insufficient field data to separate these two tracts.

**Sequences A6–A8.—**

The Freixial member of the upper Farta Pão formation (upper part of lower Tithonian to base of Cretaceous rocks, according to microfossil zonation) consists of carbonate-dominated marine to marginal marine intervals alternating with fluvial red beds of the Lourinhã formation (Leinfelder, 1986). The marine to brackish carbonates deposited in a shallow-ramp setting represent transgressive and, possibly, early highstand phases. As the exposures are not good enough to permit examination of the boundaries between the marine and terrestrial facies, it is not clear whether fluvial progradation occurred during lowstands or highstands. Since the transgressive marine sediments mostly exhibit very shallow-water characteristics, the latter interpretation is more likely. This conclusion is consistent with the fact that by this time the Arruda subbasin was virtually full of sediment so that a relative fall in sea level would result in no accommodation space being available and therefore no lowstand deposits would be preserved.

**Castanheira—Vila Franca de Xira Area**

In this area, arkosic sands and gravels of the Castanheira member are developed in a more proximal setting, close to the probable transfer zone in the eastern boundary fault complex through which the sediments were supplied (Fig. 2B). In addition, two reefal limestone intervals developed on siliciclastic-starved areas of the fan and shed large allochthonous blocks into adjacent areas (Fig. 11).

**Sequence C1.—**

The base of this sequence must occur in the subsurface, but it was not penetrated by Arruda #1. It is probably identifiable on seismic sections at the top of the uppermost interval that shows significant divergence towards the eastern boundary fault zone (shown as equivalent to the Tojeira member on Fig. 7). The sedimentary characteristics of the Castanheira member described earlier indicate that only a LST occurs in Sequence C1.

**Sequence C2.—**

**SB C2 and LST C2.—**At Monte Gordo (a prominent hill above Vila Franca de Xira), the sequence boundary is interpreted to be the contact between the Castanheira arkoses of Sequence A1 and a rudstone to grainstone interval several meters thick at the base of the Monte Gordo limestone that contains many black pebbles. Interparticle crystal silt is very frequent in places and indicates repeated subaerial exposure during a lowstand. **TST and HST C2.—**The grainstone to rudstone interval is overlain by the Monte Gordo limestone, which is about 60 m thick. At the base, there are several meters of grain-rich coral limestones dominated by microsolenids. The lime mud content then increases upwards, and siliceous sponges, indicative of deeper settings, become fairly abundant (Leinfelder et al., 1993a). This deepening trend is reversed towards the top of the Monte Gordo limestone, where coral bafflestones dominate and siliceous sponges disappear (Leinfelder, 1994). The deepening-up part of the succession is interpreted as the TST, and the shallowing-up part as the HST.

**Sequence C3.—**

**SB C3.—**This sequence is characterized by a paleokarst surface at the top of the Monte Gordo limestone. **LST C3.—**Parautochthonous to allochthonous limestone boulders embedded in arkosic sands and conglomerates occur above and on the flanks of the Monte Gordo reef relic. Block formation was caused by karstification during SB C3 together with tectonically or gravitationally induced collapse. Both the allochthonous blocks and the reef relic are overlain by coarse-grained siliciclastic sediments containing reworked reef pebbles and poorly preserved ammonites that represent the remaining part of LST C3. Further north, about 1 km west of Castanheira, an interval of allochthonous reef boulders can be correlated with those associated with the Monte Gordo reef. This indicates a time of more extensive reef growth within the Castanheira fan (Leinfelder, 1994) during the TST and HST of the preceding sequence. In the Castanheira area, the boulder level is overlain by about 150 m of Castanheira fan conglomerates. **TST and HST C3.—**In the Castanheira area the siliciclastic sediments above the lower boulder level are capped by a 3 m thick reefal limestone. This is probably a relic of a once-thicker reefal interval. The limestone consists of coral bafflestones, most of which are in life position, and frameneites with a fairly high diversity of corals (Leinfelder, 1994). A similar autochthonous reef relic occurs about 1 km to the south at the same level. The boundary between the limestone and the underlying conglomerates is transitional: the grain size and amount of terrigenous material decreases upwards as the carbonate content increases. This trend is consistent with the interpretation that the carbonate interval represents a TST/HST, with relative sea-level rise progressively reducing the influx of siliciclastic material.
Sequence C4.—

**SB & LST C4.**—The Castanheira reef limestone contains karstic cavities filled with siliciclastic material, indicating that a subaerial exposure surface developed on top of the former reef surface; this is SB C4. A significant period of karstification is indicated by the existence of an adjacent field of huge parautochthonous to allochthonous reefal boulder blocks that were karstified prior to transport. Most of the blocks occur close to the two occurrences of the Castanheira reef limestone, and their lithology is identical. Block formation probably was caused by the collapse of tower karsts at the beginning of LST C4 times.

To the east of Castanheira, the Abadia marls onlap the Castanheira formation. The base of the marls contains carbonate debris flows with large allochthonous blocks of karstified reefal limestone. The marls exhibit a more proximal character compared with the Arruda or Montejunto areas, for they are rich in plant debris and contain occasional allochthonous charophyte gyrogonites (Manuppella, pers. commun., 1994). These deposits and the boulder fields associated with the Castanheira reef relic and are interpreted as a lowstand fan, with the overlying marl succession representing a prograding lowstand wedge.

**Sequence C5.**—

**SB C5.**—As in other locations in the Arruda subbasin, the abrupt change from slope deposits of the Abadia marls to shelf carbonates of the Amaral formation is a significant shift from deeper to shallower conditions and may be interpreted as a sequence boundary, above which lowstand deposits are missing (cf. the Montejunto area). The Amaral carbonates clearly represent TST deposits, since carbonate formation in a setting with a high siliciclastic influx requires this input to be shut off by a transgression.

**TST and HST C5.**—Unlike the Amaral formation in the Arruda area, exposures of this formation north of Castanheira show no reefal facies. There it consists entirely of ooid grainstones that contain a significant amount of siliciclastic detritus, both as nuclei to the ooids and as small pebbles of vein quartz and basement material.

**Sequence C6.**—

As in the Montejunto area, the red beds of the Lourinhã formation lie above the Amaral formation; a sequence boundary is interpreted to occur between them. However, without reference to the interpretations made in the Montejunto, Arruda and Ota areas, local interpretation in the Castanheira challenges this interpretation. Lack of exposure of SB C5 and CB 6 in the area leaves open the possibility that the Amaral formation represents TST C4, overlying LST C4, with the Lourinha red beds interpreted as HST C4 (Leinfelder, 1993). However, the regional context together with the presence of reworked pebbles of Amaral limestones at the base of the Lourinhã formation supports the position of sequences C5 and C6 shown on Fig. 11.

**Age of Sequences C1 to C6.**—

Biostratigraphic indicators are extremely rare in the Castanheira-Vila Franca area. The ammonite *Ardescia pseudolictor* was found at the top of the Castanheira member, indicating a mid-early Kimmeridgian age (middle part of the *hypselocyclum* zone) for Sequence C3.

### Ota Carbonate Platform

The succession shown in Figure 12 was deposited on a narrow reef-rimmed aggradational platform (Leinfelder, 1992, 1994). This developed above the footwall of the eastern boundary fault complex of the Arruda subbasin at approximately the same time as the slope/shelf system of the top Abadia formation prograded across the subbasin (Fig. 12) (Leinfelder et al., 1988; Leinfelder and Wilson, 1989).

**Sequence O1.**—

The base of the Ota limestone is not exposed. Therefore, that part of the buildup occurring below a wedge-shaped unit of black pebble conglomerate (see Sequence O2) cannot be interpreted in sequence stratigraphic terms.
**Sequence O2.**

**SB and LST O2.**—A wedge shaped unit of black pebble conglomerate with an erosional base marks the SB and a locally developed LST. Over a distance of 1 km it thins from 4–5 m at the eastern margin of the Ota platform to only 20 cm towards the platform interior. Many of the clasts are reworked dasyclad and charophyte limestones of the Middle Oxfordian Cabaços formation. The conglomerates consist of poorly rounded to angular clasts set in a micritic groundmass. They were deposited as debris flows forming a small lowstand alluvial fan that developed away from a narrow, subaerially exposed area at the eastern margin of the platform (Leinfelder, 1987b, 1994). The wedge shape of the deposit and lack of oncoidal encrustation of pebbles clearly precludes interpretation of these deposits as a transgressive lag.

**TST and HST O2.**—Platform carbonates sandwiched between the wedge of black pebble conglomerate and a higher black pebble horizon represent TST and HST O2. The strongly aggradational nature of the Ota buildup shows that it was able to keep up with a relative sealevel rise during TST O2. Aggradation continued during the early part of HST O2, because progradation was prevented by the existence of a steep, tectonically induced bypass margin. This interpretation is consistent with the crust-rich, high-energy character of the Ota reef at the western platform margin (Leinfelder, 1992).

**Sequence O3.**

**SB O3 and TST O3.**—The base of this sequence is marked by a 10 to 50-cm-thick lithoclastic and oncolitic black pebble horizon. This horizon occurs right across the platform and rests unconformably with an angular discordance (<5°) on older strata, indicating tectonic tilting. Occasionally, tilting is overprinted by subaerial erosion, resulting in an irregular erosional surface coated by caliche crusts (Leinfelder, 1987b). Elsewhere, the horizon is a marine hardground bored by lithophage bivalves (Leinfelder, 1994), representing an amalgamation of SB O3 and a transgressive surface at the base of TST O3. The oncolitic character of most of the horizon, with black pebbles serving as nuclei for oncoids, is typical of a basal TST lag deposit.

**TST O3 and HST O3.**—This sequence consists of aggradational platform carbonates similar to those of sequence O2.

**Sequences O4–O6.**

The karstified top of Sequence O3 and the presence of relics of the Amaral formation and the Sobral and Arranhó members (Leinfelder et al., 1988; Leinfelder, 1994) suggests the presence of at least two sequence boundaries beneath the fluvial Lourinhã formation (the base of which is a third sequence boundary). It is probable that the younger lithostratigraphic units onlapped the Ota platform during transgressive phases, only to be almost totally removed during succeeding lowstands.

**Discussion**

**Composite sequence stratigraphy of the Arruda subbasin.**

Although good biostratigraphic data are lacking in places, there is enough information available, in combination with using the Amaral formation as a marker horizon, to constrain the ages of the sequences identified at the four separate localities as shown in Fig. 13. Fig. 14 shows the composite sequence stratigraphy for the Upper Oxfordian and Kimmeridgian strata that was established by comparing the interpretations of the successions studied at four separate localities in the subbasin.

**3rd-Order Sequence Stacking Patterns.**

Sequences 2–5 in the Arruda subbasin are dominated by lowstand deposits, consisting either of arkosic submarine fan sands and gravels derived from the east or southward-prograding fine-grained slope siliciclastics (Fig. 14). This dominance of lowstand systems tracts in the early sequences is a result of the deep basin filling that occurred after the rift climax at the end of Oxfordian time. These sequences change from aggradational (Sequences 2 and 3) to progradational (4 and 5). Within them, transgressive/highstand deposits formed only in two settings:

1. on the proximal part of the submarine fan system in the east of the subbasin, where the subbasin had filled sufficiently to make conditions shallow enough for carbonate deposition to have occurred (Monte Gordo and Castanheira reefs) when the siliciclastic supply was shut down by a relative sea-level rise; and

**Fig. 13.**—Summary of the probable ages of local depositional sequences and the proposed composite sequence stratigraphic scheme for the Arruda subbasin. Asterisks indicate presence of biostratigraphically significant ammonites.
2. on the uplifted footwall of the eastern boundary fault system (the Ota platform).

At the end of Sequence 6 time, the Arruda subbasin was virtually full of sediment so that only during periods of relative sea-level rise and highstand was space available for sediments to accumulate. It is also notable that from Sequence 6 onwards, relics of formations present in the basin are preserved above the Ota platform on the footwall of the eastern boundary fault zone, indicating that by this time the basin was beginning to overflow. The footwall crest would have remained a relatively positive feature due to the compaction of the thick pile of sediments in the basin to the west.

Our work shows that from Middle Oxfordian to middle late Kimmeridgian time, the nature and location of depositional systems were not controlled primarily by 3rd-order relative sea level changes but were only modified by them. Relative sea level rises and highstands appear to have shut down the influx of siliciclastic sediments into the subbasin during the time of Sequences 2 to 5. From Sequence 6 time onwards, siliciclastic systems were still effectively shut down during relative rises, but were able to prograde during highstands. Lack of accommodation space prevented deposition during lowstands.

**Comparison with Proposed European 3rd-Order Sequence Chart.**

Up to this point, applying the sequence stratigraphic approach has not relied on the use of so-called global sequence correlation charts. For Upper Jurassic successions, such charts (e.g., Haq, 1987; Ponsot and Vail, 1991a, b; Jacquin et al., this volume) are based largely on studies of successions that accommodate on relatively tectonically undifferentiated ramp settings in contrast to the rift origin of the Arruda subbasin. Nevertheless, from the planula zone (uppermost Oxfordian) to the top of Tithonian stage, 3rd-order sequences of the Arruda subbasin can be tied to the scheme for European basins proposed by Jacquin et al. (this volume) as summarized in Fig. 15. Sequence boundary 2 of the Arruda subbasin can be tied with SB Ox8 of Jaquin et al. (this volume) but Sequence 2 of the Arruda subbasin corresponds to two sequences (Ox 8, Ox 9). It is possible that Sequence Ox 8 has not been identified in the Arruda subbasin because of the poor exposure of the top part of the Tojeira member, where it would be expected to occur. Recognition of the earlier Oxfordian sequences (Ox 0–7) in the Arruda subbasin is not possible due to poor exposure and tectonic deformation of rocks of this age in the Montejunto area.

European-wide biostratigraphic and hence sequence stratigraphic correlation of the Upper Oxfordian/lower Kimmeridgian strata is far from being resolved. Problems exist particularly between the correlation of zones from the boreal to sub-Mediterranean realm. Taxonomic revisions of Upper Jurassic ammonites from southern Germany resulted in the recent recognition of Amoeboceras bauhini in the uppermost bimammatum zone, a classical “Upper Oxfordian” biozone of the sub-Mediterranean realm. However, this ammonite is the index fossil for the base of the Kimmeridgian (base of baylei zone) in Great Britain (i.e., in the boreal realm, G. Schweigert, 1995). This discovery shows that the definition of base of the Oxfordian in the sub-Mediterranean zonation (base of platynota zone) does not coincide with the definition of the base of the Oxfordian in the boreal classification (base of baylei zone). The Lusitanian Basin stratigraphy is based on the sub-Mediterranean classifi-
THIRD-ORDER SEQUENCES IN AN UPPER JURASSIC RIFT-RELATED SECOND-ORDER SEQUENCE

Fig. 15.—Composite sequence stratigraphic interpretation of the Upper Jurassic strata of the Arruda subbasin compared with the 3rd-order sequence chart Jacquin et al. (this volume). Key to letters and numbers: LCF: lower part of Castanheira submarine fan; MGR: Monte Gordo reef; UCF: upper part of Castanheira submarine fan; CR: Castanheira reef; Fx: Freixial member; 1: change from dominantly carbonate to siliciclastic deposition; 2: Mirante sand interval in Montejunto area; 3: condensed interval exposed beneath Serra Isabel (Fig. 10); 4: reworked caliche nodules in marine conglomerate within Sobral member.

cation. In the light of these new results, the top of the Montejunto formation and the entire Tojeira member may be early Kimmeridgian age in the boreal sense. Since 3rd-order relative sea-level curves were largely established in the boreal realm, a complete revision of the curve for the Oxfordian/Kimmeridgian transition interval may be necessary.

Constraining the Timing of Rifting and Major Sedimentary Infilling.—

The sequence stratigraphic interpretation and correlation presented in Figures 14 and 15 enables the timing of the rift climax and subsequent basin filling to be constrained (Table 1). Most of the subsidence probably occurred during the rift climax phase that coincides with Sequence 2 (Ox 8 and 9) times, and nearly 2 km of arkosic sands and gravels were deposited during Sequence 3 (K1) time. According to Jacquin et al. (this volume), these two intervals span about 1.3 my, which gives subsidence/sedimentation rates in the order of 2 m/ky, more than double the tectonic subsidence rates obtained by Wilson et al. (1989) using the total thickness of the Castanheira member and the total time duration estimated for its deposition. However, exact dates will only be obtained once the problems in biostratigraphic correlation of the Oxfordian/Kimmeridgian boundary, as outlined above, are solved.

CONCLUSIONS

The overall distribution in space and time of major depositional systems within the 2nd-order sequence of the Arruda subbasin was linked to tectonic subsidence, but shorter and smaller-scale variations in facies distributions were controlled by 3rd-order changes in relative sea level. Table 1 provides a summary of the relationships between lithostratigraphic units, rifting phases and third order sequences discussed in this paper.

Four major depositional systems fed sediments into the Arruda subbasin (Fig. 3): (1) carbonate buildups and associated deep-water sediments; (2) coarse-grained siliciclastic submarine fan, sourced from the east through a gap in the eastern boundary fault zone; (3) prograding fine-grained siliciclastic slope showing clinoform reflectors on seismic sections capped by shallow-water carbonates; and (4) coastal plain and shelf. The distribution of these systems in space and time was controlled largely by changes in accommodation space caused by tectonism and sediment infilling, resulting in a 2nd-order transgressive-regressive sequence (Fig. 6).

Increased subsidence rates during middle Oxfordian time compared with those earlier in Jurassic time marked the onset of rifting. Differential subsidence during rift initiation resulted in aggregational carbonate buildups forming along the eastern margin of the subbasin, with hemipelagic carbonates being deposited to the west. Rifting climaxed around the Oxfordian-Kimmeridgian boundary, resulting in footwall uplift of the eastern margin of the subbasin that caused meteoric diagenesis and shedding of blocks of earlier Oxfordian carbonates. At the beginning of Kimmeridgian time, it is probable that the Arruda subbasin was a deep “hole” in which over 2500 m of largely siliciclastic postrift sediments subsequently accumulated. The 2nd-order maximum flooding occurred during Late Oxfordian bimammatum zone time.

The coarse arkosic sediments of the Castanheira member were deposited by new drainage systems that were established in the recently rifted source area to the east and northeast. They formed the immediate postrift fill of the subbasin, which eliminated much of the accommodation space created during the rift climax. The late postrift phase is characterized by progradation of the slope-shelf system over the arkosic submarine fan sediments. This filled the basin virtually to sea level, after which coastal plain and shallow shelf sediments accumulated during a period in which subsidence largely resulted from sediment loading and differential compaction.

Four key findings result from our sequence stratigraphic interpretation at the 3rd-order scale of the sedimentary fill of the
### Table 1. Tectonic Phases, 3rd Order Sequences and Lithostratigraphic Units of the Upper Jurassic Arruda Subbasin

<table>
<thead>
<tr>
<th>Rifting phase/tectonic system tract and age</th>
<th>3rd order sequences A in Arruda subbasin (Ox, K) and (T): correlation with European sequences (Jacquin et al., this volume)</th>
<th>Lithostratigraphic units</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Late Post rift</td>
<td>A6–11 (K4, T1–76)</td>
<td>Lourinhã fm. (up to 400 m): fluvialite red beds, with deltic Sobral member at base; interbedding southwards with: Faria Plo fm. (up to 100 m): Annuná and Freixaí mbrs.): limestones and marls with euryhaline to brackish bivalves. Top Abadia fm. (550 m): southward-prograding fine-grained siliciclastic slope capped by coral boundstones and ooid grainstones of the Amapal fm (~80 m). Top part of Castanheira mbr. (~350 m): arkosic submarine fan with carbonate buildups. Ota limestone (~160 m): aggradational reef-rimmed carbonate platform situated on horst in boundary fault complex.</td>
<td>Very low basement subsidence rates partly due to compaction of older sediments, so that eustatic control of accommodation space may have become very significant. Formations can be identified across several subbasins, in contrast to the greater complexity of facies distributions in earlier systems. Third order sequences changed from progradational lowstand dominated (top Abadia) to transgressive and hightstand dominated from the Amapal fm. onwards.</td>
</tr>
<tr>
<td>Immediate Post rift</td>
<td>A3–5 (K1–3)</td>
<td>Lower part of Castanheira mbr (~2000 m): arkosic sandstones and gravels built westwards from gap in eastern boundary fault complex.</td>
<td>The influx of coarse clastics was probably caused by the establishment of a drainage system that crossed inactive fault systems and so could transport basement debris from the east and northeast. Carbonate clasts within the Castanheira member may be derived from footwalls of boundary fault complex and/or further fault systems to the east. Aggradational lowstand systems only were deposited.</td>
</tr>
<tr>
<td>Rift Climax</td>
<td>A2 (OX 8, 9)</td>
<td>Tojeira mbr. (&lt;100 m): carbonate and siliciclastic turbidites and debris flows with allochthonous karstified shallow-water limestone blocks.</td>
<td>Rifting and regional subsistence resulted in hanging wall drowning of earlier carbonate depositional systems, and footwall uplift resulted in meteoric diagenesis and shedding of allochthonous blocks of earlier carbonates into the basins. Basement rocks became exposed in postulated footwall blocks to the east, which resulted in significant siliciclastics being deposited for the first time since the Late Triassic Epoch.</td>
</tr>
<tr>
<td>Rift Initiation</td>
<td>A1 (OX 73–7)</td>
<td>Montejunto fm.: deep-water carbonates in basin (2–400 m) carbonate buildups above footwall of eastern boundary fault complex (500 m). Cabaço fm. (2–400 m): lacustrine and shallow-marine limestones, with anhydrite.</td>
<td>Initial uplift produced hiatus followed by lacustrine and shallow-water carbonates. Faulting produced topographic differentiation into subbasins though flexuring (i.e., fault-tip folds) leading to carbonate buildups forming on eastern margin. Salt movement was triggered by faulting.</td>
</tr>
</tbody>
</table>

Arruda subbasin. (1) 3rd-order sequences developed during all stages of the infilling of the subbasin, suggesting that 3rd-order changes in relative sea level controlled smaller scale facies distributions. (2) Aggradational followed by progradational lowstand depositional systems tracts dominated the 3rd-order sequences in the early regressive part of the subbasin fill. Slightly progradational to aggradational transgressive/highstand-dominated 3rd-order sequences formed once the subbasin was filled near to sea level, from *eudoxus* time onwards. (3) Biostratigraphic calibration, though varying in precision in different parts of the successions, is sufficient for 11 sequence boundaries to be identified within the rift-climax and postrift fill of the subbasin (mid-bihimmatum zone to top Tithonian).

Despite the limited biostratigraphic control, the ages of these boundaries appear to correlate well with the European sequence boundaries proposed by Jacquin et al. (this volume), though their Ox 9 boundary has not been identified. (4) The identification of 3rd-order sequences enables the timing and rates of tectonic subsidence and subsequent deposition to be estimated with greater precision, suggesting that during the rift climax phase they approached 2 m/ky over a relatively short period of time (1–2 my).

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THIRD-ORDER SEQUENCES IN AN UPPER JURASSIC RIFT-RELATED SECOND-ORDER SEQUENCE


SEQUENCE STRATIGRAPHY OF THE OXFORDIAN AND KIMMERIDGIAN STAGES (LATE JURASSIC) IN NORTHERN SWITZERLAND

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ABSTRACT: The Oxfordian and Kimmeridgian exposures of the Jura of northern Switzerland show a transition from a shallow-water carbonate platform to a deeper marine basin. The carbonate platform was well established in the northern part of the Jura by Middle Oxfordian time and continued to develop through Late Oxfordian and Kimmeridgian time. The platform is characterized by shallow-water carbonates, sometimes rimmed by coral reefs. The basal sediments grade distally from marls or turbiditic carbonates into condensed, highly fossiliferous limestones or ironstones. The study is based upon 220 sedimentary logs and utilizes a very detailed chronostratigraphic framework established from a combination of ammonite biostratigraphy and clay mineralogy.

Sequence stratigraphic analysis of the succession reveals that the Oxfordian sediments as a whole represent a 2nd-order regressive-transgressive cycle, punctuated by eight 3rd-order cycles. The onset of the 2nd-order transgression is coincident with a change in the morphology of the carbonate platform from a rimmed-shelf to an epicontinental carbonate platform. The lower Kimmeridgian sediments represent a minor 2nd-order regressive cycle, punctuated by three 3rd-order scale cycles. This was followed by a 2nd-order transgression, the base of which is dated as *eudoxus* Zone. The Oxfordian sequence boundaries are denoted O1 to O8, and the Kimmeridgian sequence boundaries are denoted K1 to K3. Relative sea-level changes are supported by the progradational, aggradational and retrogradational relationships of the facies within the sequences and by detailed study of the genesis and correlation of the key stratigraphic surfaces.

The sequence stratigraphic method allows further refinement of correlations between the basin and carbonate platform, and provides an explanation and framework for the complexity of carbonate facies.

INTRODUCTION

Aims and Methodology

The Oxfordian and Kimmeridgian exposures in the Jura of Northern Switzerland (Fig. 1) provide an ideal opportunity to test the concepts of sequence stratigraphy and to study the internal architecture of sequences because they:

—can be placed in a precise biostratigraphic framework based on ammonites
—show sediments deposited in a spectrum of settings from a carbonate platform to deep neritic basin, and
—have been studied in great detail by the first author and others for many years.

The aim of this paper is to present a sequence stratigraphic summary of these exposures and to examine how the key stratigraphic surfaces and facies packages change both laterally and vertically. This paper utilizes the sequence stratigraphy model and definitions proposed by Vail et al. (1991), Van Wagoner et al. (1988) and Duval et al. (1992). The sequence stratigraphic model has been adapted to take into account the different sediments and facies found in environments ranging from the carbonate platform to the deep basin, the intermixing of terrigenous siliciclastics with carbonates and the epicontinental setting of the Swiss Jura.

This paper is based upon a large geological dataset. The sequence stratigraphic interpretation was completed from field studies and 220 detailed sedimentary logs completed by R. Gygi but as yet mostly unpublished. The time framework (Fig. 2) is based upon all of the published ammonite biostratigraphy for the area and a clay mineralogy study by Gygi and Persoz, 1986. The sedimentary logs published in this paper are summary logs from selected representative points along the platform to basin profile (Figs. 1, 3), and therefore the exact bed numbers mentioned in the text are not shown on the summary logs (Figs. 5A, 5B, 6A, 6B, 7—see Fig. 4 for key).

Geological Setting

The successions studied were deposited in an epicontinental sea situated between the emergent land of the Ardenne–Rhenan massif to the north and the Tethys ocean basin to the south (Gygi and Persoz, 1987, Fig. 7; Gygi, 1992, Figs. 1 and 2). The outcrops occur partly in the folded and partly in the tabular Jura range (Fig. 1). Synsedimentary tectonics were mostly restricted to uniform regional subsidence (Gygi, 1986) with localized synsedimentary faults, including the Late Oxfordian Effingen Member of north-central canton Aargau may also be the effect of synsedimentary faulting (Gygi, 1990c, Figs. 5, 6; Fig. 3).

Previous Work

The Oxfordian and Kimmeridgian exposures in northern Switzerland display a full transition from a carbonate platform to a deeper marine basin. This transition of sediments was first recognized by Gressly (1838–1841), and it inspired him to develop the concept of facies. These exposures were also used by Oppel (1863) in his pioneering work on modern zonal stratig-
Fig. 1.—Non-palinspastic map of northern Switzerland showing the localities and towns cited. The measured sections (black dots with numbers) are projected onto two transects that are perpendicular to the depositional strike. The two transects are linked by a line of depositional strike.

A short review of diagnostic facies and of the factors controlling sedimentation is in Gygi and Persoz (1987), and the paleogeography is summarized in Gygi (1990c). Methods to reconstruct the paleobathymetry down to the deep subtidal zone were suggested by Gygi (1981, 1986, Fig. 6).

SEQUENCE STRATIGRAPHIC ANALYSIS OF THE THIRD-ORDER SEQUENCE CYCLES

Late Callovian and Early Oxfordian Time

A depositional sequence spans across the Callovian–Oxfordian boundary. At the base of this sequence is the thin but widespread Lamberti Bed, which in northwest Switzerland comprises 10 cm of marl-claystone with abundant iron ooids. The Lamberti Bed contains a rich macrofauna that consists mainly of ammonites indicative of the lamberti Subzone, the last Middle Jurassic subzone. Below this bed is a hiatus that encompasses the henrici Subzone of the lamberti Zone. This hiatus is present throughout northwest Switzerland and is interpreted as a sequence boundary. The top of the Lamberti Bed (Callovian–Oxfordian boundary) is interpreted as a maximum flooding surface because of the rich macrofauna and condensed nature of the bed.

In northwestern Switzerland, the base of the Oxfordian succession coincides with the base of the Renggeri Member (Figs. 2, 5A, 8). The lowermost 20 cm of the Renggeri Member is a widespread calcareous, slightly silty claystone with scattered iron ooids (Fig. 5A; Gygi, 1990a, Fig. 2, section RG 280 near Liesberg, upper part of bed 6). The abundance of ooids decreases to zero at the top of the bed. This bed contains a ma-
Fig. 2.—Chronostratigraphical table showing the correlation of the latest Callovian, Oxfordian and Kimmeridgian units across northern Switzerland, together with the sequence stratigraphic interpretation. The colors usually adopted for sequence stratigraphy (e.g., Rioult et al., 1991), have been used to illustrate the full sequence stratigraphic interpretation of the section. A different color is used for each systems tract and its corresponding upper boundary. The color code is as follows: Lavender: shelf-margin systems tract and transgressive surface, green: transgressive systems tract and maximum flooding surface, and orange: highstand systems tract and sequence boundary.
crofauna of mainly ammonites including Scarburgiceras cf. scarburgense near Péry (Gygi, 1990a, Fig. 4). Gygi (1986, Fig. 3A) estimated that the water depth at the time of deposition of this bed was between 80 and 100 m. The iron ooids are interpreted to have formed in situ during a time of very low mud

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**Fig. 3.**—Synthetic cross section of the platform-to-basin transition of the Oxfordian and Kimmeridgian Stages in northern Switzerland. The cross section is palinspastic, is assembled from the two transects shown in Figure 1 and is restored to Eudoxus Chron times. Note that there is a strong vertical exaggeration. The water depth is inferred from the calculations by Gygi (1981, 1986). The sediment thicknesses shown are averaged and compacted. The base of the section is the effect of differential subsidence and compaction under the varying sediment load that is superimposed on a supposedly uniform regional tectonic subsidence. The bathymetric profile at the beginning of the Oxfordian is shown in Gygi (1986, Fig. 3A). The figure presented is modified from Plate 1 in Gygi and Persoz (1986). Sequence boundaries—Oxfordian: 01 to 08; Kimmeridgian: K1 to K3. The sequence stratigraphic interpretation of the Renggeri Member (REN) and the existence of sequence boundary 01 are uncertain. Correlation of the Gerstenhübel Limestone (GER) and Effingen Member (EFF) to the southeast of Mellikon is uncertain. The Oxfordian Stage is interpreted as a 2nd-order regression followed by transgression. The lower Kimmeridgian Stage is interpreted as a very minor 2nd-order regressive phase. Second-order transgression resumed in mid-Eudoxus Chron times (in the sequence above K3). For key to colors see caption for Figure 2. Blue: coral bioherms. Abbreviations for lithostratigraphic units: BAD—Baden Member, BAN—Banné Member, BIR—Birkenstock Member, BUR—Bure Member, CRE—Crenularis Member, EFF—Effingen Member, GEI—Geissberg Member, Gh.B—Gerstenhübel Beds, GUN—Günsberg Member, HMB—Hauptmunienbank, HOL—Holzflue Member, KUS—Küssaburg Member, Lam—La May Member, LAU—Laufen Member, LET—Letzi Member, Lie—Liesberg Member, ROS—Röschzen Member, PIC—Pichoux Formation, POR—Porrentruy Member, REU—Reuchenette Formation, STE—Steinebach Member, SUF—St-Ursanne Formation, TAC—Terrain à Chaillies Member, VER—Verena Member, VOR—Vorbourg Member, WAN—Wangen Member, WET—Wettingen Member. For key to symbols see Figure 4.

**Fig. 4.**—Key to the sedimentary logs and cross section (Figs. 3, 5, 6, 7).
Fig. 5.—(A) Composite section of Lower Callovian to upper Kimmeridgian strata in the region of Liesberg (the composite is made up from the following individual sections: RG 280, clay pit of Amphil, near Liesberg; RG 306 clay pit and quarry of Chestel, near Liesberg; RG 402, Röschenz, Cantonal road to Müli; and, RG 398 limestone quarry, near Liesberg). (B) Composite section from the Lower Oxfordian to the Lower Kimmeridgian near Sornetan (the composite is made up from the following individual sections: RG 314 and 315, in the Gorges du Pichoux). The ammonite zones and subzones are biostratigraphic. For key to symbols see Figure 4.
Fig. 6.—(A) Composite section of Lower Callovian to lower Kimmeridgian strata from near Péry and Oberdorf (the composite is made up from the following individual sections: RG 307, quarry of La Charuque, near Péry; and RG 433, Wäberhüsli, near Oberdorf). (B) Composite section of Lower Bathonian to lower Kimmeridgian in canton Aargau (the composite is made up from the following individual sections: RG 37, quarry of Jakobsberg, near Auenstein; RG 226, road cut, near Auenstein; RG 62 Schrannechopf, near Villigen; and RG 70 large quarry near Mellikon). The ammonite zones and subzones are biostratigraphic. For key to symbols see Figure 4.
OXFORDIAN–KIMMERIDGIAN SEQUENCE STRATIGRAPHY NORTHERN SWITZERLAND

Fig. 7.—Section of the Middle Oxfordian to the Lower Kimmeridgian strata near Moutier, RG 381, Gorges de Court. The ammonite zones and subzones are biostratigraphic. For key to symbols see Figure 4.

The sedimentation rate at, or near, storm wave base. This basal bed with iron ooids and the lower part of the overlying slightly silty calcareous claystones of the Renggeri Member comprise the highstand systems tract (Figs. 2, 3, 5A).

This Callovian–Oxfordian sequence thins out altogether in the distal direction. It reappears in the deep part of the basin in the form of thin lenses of iron oolite and marl that are traceable laterally over several kilometers (Fig. 2).

Sequence Boundary O1 (Praecordatum Subzone?)

Ammonites indicate that the scarburgense, praecordatum, bukovskii and costicardia Subzones are all represented by the Renggeri Member (Gygi, 1990a), but the only continuous exposure of this member is at Ampthil Clay Pit, Liesberg (Figs. 5A, 8). In general, the member comprises up to 70 m of very poorly bedded, variably silty claystone with scattered ammonites, belemnites, brachiopods and small pyrite filled burrows. The clay pit exposure is difficult to examine in detail sedimentologically due to the formation of a thick superficial crust that forms as the exposure continually weathers and slips. However, examination of the clay-pit exposure, and samples taken from every 2 m, shows that there is a sharp color change in the middle of the member from a light bluey-grey below to a more uniform light grey above (Fig. 8). This color change is coincident with an increase in silt content and is interpreted as sequence boundary O1. The ammonites show that this change is near to the top of the praecordatum Subzone. Near to the base of the costicardia Subzone, there is a band of medium- to dark-grey mudstone that is interpreted as the transgressive surface. Small exposures of the costicardia Subzone mudstones near Liesberg have recently yielded several pyritized ammonites and probably indicate the proximity of the maximum flooding surface. Rioult et al. (1991, Fig. 17) and Coe (1992a, 1992b, 1995) document the first Oxfordian sequence boundary in France and England to be at, and near, the top of the praecordatum Subzone, respectively.

Sequence Boundary O2 (Costicardia–Cordatum Subzone Boundary)

The Renggeri Member is overlain by the much more calcareous Terrain à Chailles Member. The boundary between the two members is well-defined by the lowest band of approximately spheroidal carbonate nodules. The Terrain à Chailles comprises interbedded marl-claystones, bands of carbonate nodules and continuous beds of marly limestone (Figs. 5A, 8). The spacing and relative abundance of these different beds varies systematically through the member and is clearly indicative of two 3rd-order cycles. The macrofauna of the member includes ammonites, brachiopods and infaunal bivalves.

The sharp change to more carbonate-rich facies at the base of the Terrain à Chailles represents the incoming of shallower water facies and is interpreted as sequence boundary O2 at its correlative conformity. Ammonites from either side of the boundary indicate that sequence boundary O2 corresponds to the base of the cordatum Subzone.

The Terrain à Chailles Member pinches out completely in the distal (southeast) direction (Figs. 2, 3). Further into the basin, it reappears as a thin discontinuous bed, the Schellenbrücke Bed, which contains ammonites that indicate that it belongs to the cordatum Subzone (Fig. 2; note that the bed is too thin to be represented in Fig. 3). The Schellenbrücke Bed is composed of scattered ferruginous ooids (10–20%) supported by a matrix of ferruginous lime mudstone. The macrofauna of the Schellenbrücke Bed is dominated by ammonites (Gygi, 1981, Table 1). The bed is interpreted as a shelf-margin systems tract, and the proposed mode of formation of the iron ooids in relation to
the marls of the Terrain à Chailles Member is shown in Figure 9. The formation of thin, widespread iron oolites in relatively deep water with a muddy matrix and a macrofauna of mainly cephalopods was first discussed by Gygi (1981). Basinward, the Schellenbrücke Bed grades into the Glaukonitsandmergel Bed, which is a 10-cm-thick marl clay with up to 30% glauconite.

**Transgressive Surface.**

The lower portion of the Terrain à Chailles Member, interpreted as the shelf-margin systems tract, contains evenly and well-spaced interbedded marl-claystones, limestone nodules and limestones; this is overlain by an approximately 10 m package that lacks continuous limestone beds. This decrease in carbonate, and thus relative increase in argillaceous material, is interpreted as representing the transgressive systems tract because the carbonate is a shallower water facies. In the marginal part of the Terrain à Chailles, the transgressive surface at the base of this argillaceous package is marked by a limestone nodule bed that contains a small concentration of macrofossils, mainly ammonites (Gygi and Marchand, 1993; Fig. 2). This is the “Fossil bed” of Gygi and Persoz (1986).

**Maximum Flooding Surface.**

Near the top of the transgressive systems tract, continuous limestone beds reappear; above this, the bed spacing decreases until the beds are very closely spaced just above the middle of the member. These closely spaced beds also contain ammonites and brachiopods. The maximum rate of decrease in bed spacing is explained as being due to decreasing sediment supply and is thus interpreted as the maximum flooding surface. The highstand systems tract is comparatively thin (2–3 m) and comprises interbedded marl-claystones, limestones nodules and limestones. In the Jura collection of the Natural History Museum, Basel, there are several ammonites and belemnites encrusted with serpulids and oysters from the Terrain à Chailles, but the exact horizon from which they came is unknown. However, it seems likely that they were found near this maximum flooding surface. In the basin, the transgressive surface and maximum flooding surface have coalesced and are represented by a limonitic crust on top of the Schellenbrücke Bed (Fig. 2).

**Sequence Boundary O3 (Mid-Densiplicatum Subzone)**

**Sequence Boundary and Transgressive Surface.**

A further cycle is recognized within the Terrain à Chailles Member. The upper part of the carbonate-rich package above the middle of the member contains several thick, well-defined, tabular limestones interbedded with marl-clays that contrast with the closely bedded, irregular, fossiliferous limestones and limestone nodules immediately below. The first thick tabular limestone is interpreted as sequence boundary O3. As in sequence O2, the transgressive surface is marked by a change to a package of interbedded marl-claystones and limestone nodules; continuous limestone beds do not reappear again until near the top of the transgressive systems tract (Fig. 5A).

In the basin the equivalent stratum is a thin marly limestone with iron ooids and ammonites. This is the lowest bed of the Birmenstor Member and contains ammonites indicative of the densiplicatum Zone.

**Maximum Flooding Surface.**

The boundary between the Terrain à Chailles Member and the overlying Liesberg Member is well-defined. It is characterized by the sudden appearance of a profusion of hermatypic corals and is interpreted as a maximum flooding surface. Litho-
ologically, the Liesberg Member contains an abundant fauna that is dominated by corals but also includes well-preserved serpulids, crinoids, brachiopods, echinoids and bivalves. The Liesberg Member is a marl-claystone with carbonate nodules similar to the Terrain à Chaîlles Member, but the nodules of the Liesberg Member have an irregular shape, are partly silicified and contain an abundant fauna. The hermatypic corals of the Liesberg Member form a densely colonized biostrome; no bioherms have been observed. The mass occurrence of hermatypic corals with mostly dish-shaped colonies in the Liesberg Member is noteworthy because the sedimentation rate of siliciclastic mud was high (up to 25 m were deposited near Liesberg during only part of the Antecedens Subchron, Fig. 2). This member is interpreted to comprise all of the highstand systems tract. The time-equivalent sediments of the Liesberg Member in the basin are thin and condensed. They are part of the condensed basal bed of the Birmenstorf Member and the laterally equivalent upper part of the Mumienmergel Bed (Fig. 2).

The widespread Mumienmergel Bed is a glauconitic marl with oncoids. The bed is only 10 cm thick and is interpreted to have been laid down at a depth of more than 120 m (Gygi, 1992). In spite of this relatively great depositional depth, the bed contains some discoid oncoids that are up to 30 cm in diameter and formed around parts or entire casts of ammonites. These ammonite casts were overturned during the formation of the oncoidal crust (Gygi, 1992), but the cause of this overturning is uncertain. Water currents strong enough to overturn discoidal oncoids with a diameter of up to 30 cm are thought unlikely to have occurred in this environment, so the ammonites could have been overturned by fishes (Gygi et al., 1979, p. 946; Gygi, 1981, p. 249).

**Initiation of the Carbonate Platform**

The sediments of the mid-antecedens Subzone indicate that a carbonate platform started to develop in the area following deposition of the sequence above O3 (the term carbonate platform is used as a general term for a thick sequence of shallow-water carbonates in the sense of Tucker and Wright, 1990). The facies deposited between the mid-antecedens Subzone and mid-hypsulenum Subzones indicate that it was a rimmed-shelf type of carbonate platform. Reefs and oolitic sandbodies were developed along the shelf margin, proximally from this there was a lagoonal area in which stromatolitic limestones, micrites, and peloidal and oncoidal wackestones and packstones were deposited and in the lower part of the succession lagoonal reefs developed. In the basin, micrites and rhythmically bedded arcellar micrites and micritic limestones, probably deposited by turbidity currents, accumulated. The sequence stratigraphic analysis is based upon the development and superimposition of these facies.

**Sequence Boundary O4 (Mid-Antecedens Subzone)**

**Sequence Boundary.**

The Liesberg Member is overlain by the St-Ursanne Formation; the base of the latter is marked by the sudden appearance of oncoidal and oolitic grainstones with oncoids (Fig. 5A). This indicates a rapid basinward shift of carbonate facies and the establishment of a carbonate platform in this area. The base of the St-Ursanne Formation is interpreted as a sequence boundary (O4). The lower half of this formation is very heterogeneous with no well-defined platform rim of oolite shoals. The time-equivalent sediments deposited at the platform edge and in the basin also show a marked change in character from those beneath, indicating a sequence boundary (Figs. 5B, 6). At the outer fringe of the platform, the first true Oxfordian coral bioherms started to grow. These bioherms form the lower part of the St-Ursanne Formation in this area (Fig. 3; Gygi and Persoz, 1986). Distally from the platform edge, bedded micrites of the Pichoux Formation were deposited, which prograded basinward and thus progressively overlie older strata until at Pery they rest unconformably on a small thickness of the Renggeri Member (Figs. 3, 6A). The lower part of both the St-Ursanne Formation and the Pichoux Formation are interpreted as a shelf-margin systems tract. In the deep part of the basin, sequence boundary O4 is within the very thin (less than 10 cm), condensed basal bed of the Birmenstorf Member (Fig. 2).

**Transgressive Surface.**

The start of transgression is marked by a distinct facies change from oolitic grainstones to chalky micrites with coral bioherms over a wide area of the platform interior. The transgressive surface is coincident with the onset of coral bioherm growth far into the platform interior, for example at St-Ursanne and Liesberg (Fig. 3). These bioherms are composed of up to 40% coral colonies, most of which are massive. At Roche au Vilain near St-Ursanne, a large number of regular echinoids was found adjacent to one of the lagoonal bioherms (V. Pümpin, pers. commun., 1968); this provides evidence that at least some of the chalky micrite was produced through bioerosion. The bioherms and the chalky micritic sediments in between are interpreted as the transgressive systems tract. The time-equivalent sediments on the slope are a marly limestone and the overlying micritic limestones of the middle to upper part of the Pichoux Formation (e.g. near Sornetal; Figs. 2, 5B).

**Maximum Flooding Surface.**

In the upper part of the St-Ursanne Formation, the bioherms of the platform interior all appear to have been simultaneously suffocated by a rapid increase in lime mud supply. This is interpreted as reflecting a rapid relative sea-level rise, and thus the top of the bioherms is interpreted as the maximum flooding surface. The highstand systems tract is characterized by thickly bedded marine lime-mudstones over the platform interior, by oolite shoals and a fringe of coral mudmounds containing up to 10% typically massive corals at the platform margin and by the bedded micrites of the upper Pichoux Formation on the slope. The time-equivalent strata of the upper St-Ursanne Formation in the basin is the noncondensed part of the Birmenstorf Member (Figs. 2, 6B). This is a densely colonized biostrome of siliceous sponges interpreted to have grown at a water depth of more than 100 m (Gygi, 1986, Fig. 3B).

**Sequence Boundary O5 (Parandieri–Schilli Subzone Boundary)**

**Sequence Boundary.**

Sequence boundary O5 and the overlying shelf-margin systems tract are well developed both on the carbonate platform
and in the basin. On the carbonate platform, the contact between the St-Ursanne Formation and the overlying Vorbourg Member (Vellerat Formation; Fig. 2) is interpreted as sequence boundary O5. The top of the St-Ursanne Formation is usually a hardground and is directly overlain by an argillaceous and sometimes sandy marl representing renewed deposition of terrigenous sediment. For example at Röschenz (Fig. 5A) there is a limonitic crust overlain by a thin argillaceous marl, and at Gorges du Pichoux, near Sorpetan (Fig. 5B), the top of the St-Ursanne Formation is a sharp erosion surface overlain by a thin, sandy marl. Pfitzer (1982, p. 22) recorded that the top of the St-Ursanne Formation was a bored, encrusted and mineralized hardground at three localities near Moutier. Similarly, near the base of the carbonate platform slope at Péry, the sequence boundary is a well-developed limonitic and pyritized crust on top of the Pichoux Formation (Fig. 6A). In the basin, sequence boundary O5 is marked by the sharp boundary between the sponge biostratums of the Birmenstorf Member and the redeposited marls and argillaceous limestones of the lower Effingen Member (e.g., Auenstein; Figs. 3, 6B, 10).

The shelf-margin systems tract is represented on the carbonate platform by the Vorbourg Member and by the lower part of the Röschenz Member. The Vorbourg Member is a thickly bedded micritic limestone with clay and detrital quartz, especially near the base. The member also contains fenestral stromatolites with dewatering cracks (Gygi, 1992, Fig. 10), stromatolites with rootlets (Gygi, 1992, Fig. 13) and lagoonal oncoids (section RG 396, Kleinlützel, unpublished, northwest of Liesberg, Fig. 1). Pumplin (1965, Fig. 26) found a tidal channel in this member. The Vorbourg Member grades laterally towards the platform margin into cross-bedded grainstones and stromatolitic limestones of the Günberg Member (e.g., Gorges du Pichoux, near Sorpetan; Fig. 5B). The time-equivalent sediments on the slope and in the basin are a very thick succession of rhythmically bedded argillaceous micrites (40–70% CaCO₃) and micritic limestones (70–95% CaCO₃). These form the lower Effingen Member (Fig. 3; Gygi and Persoz, 1986, Plate 1). The thickness of the lower Effingen Member changes between the slope and the basin, indicating that the member has a sigmoidal progradational clinoform geometry. The change in sediment type and sedimentation rate across sequence boundary O5 is dramatically displayed near Péry (Fig. 6A) where all of the Effingen Member exposed is assigned to this shelf-margin systems tract. Spectacular, large erosional scour features are developed near the base of this member near Auenstein, indicating temporary high-energy conditions in the basin (Fig. 10). The Effingen Member is interpreted to have been deposited by turbidity currents because of the presence of sharp based fining-upward cycles visible in polished slabs (Gygi, 1969, Plate 4, Fig. 12), the decimeter-scale cyclicity, erosional scours, geometry and paleogeographic position of the member. The source of the large amount of micrite in the Effingen Member is problematic because there is no obvious large carbonate platform preserved that could have produced all of this sediment. However, we envisage that in part it was produced by bioerosion from the coral bioherms of the Günberg Member and in part by whittings (Macintyre and Reid, 1992; Milliman et al., 1993) from the central part of the platform. The argillaceous content of the member was sourced from the rivers draining the land and bypassing the carbonate platform.

The lateral facies boundary between the shallow-water limestones of the Vorbourg and Günberg Members and the argillaceous micrites of the Effingen Member is marked by coral bioherms that prograded nearly 20 km basinward during deposition of the Effingen Member. The facies relationship between the Effingen Member and the coral bioherms is well-exposed behind the silos on the west side of the road at La Reuchenette cementworks, Péry (Figs. 1, 3, 11; Gygi, 1992, Fig. 9).

The Vorbourg Member grades vertically into the Röschenz Member (formerly Natica Member; Gygi and Persoz, 1986) which is an intercalation of marls and limestones deposited in very shallow water. The marine strata contain gastropods like Neritoma (formerly Natica) and sometimes very numerous nerineids. Ostreid bivalves may also be common. Wavy laminated stromatolites with fenestrate pores indicate an intertidal environment in the lowermost Röschenz Member near Röschenz (Fig. 5A).

Transgressive Surface.—

At the platform margin, transgression is marked by the change from prograding to aggrading coral bioherms. In the basin, the lateral equivalent contains a relative increase in the amount of carbonate deposited compared to the rest of the Effingen Member (Fig. 6B). In the platform interior, the carbonate facies belts backstep so that peloidal and oolitic mudstones and wackestones overlie stromatolitic limestones (e.g., Röschenz; Fig. 5A).
In the platform interior, the maximum flooding surface is fossiliferous; for example, at Röschenz (Fig. 5A), it is indicated by a bed with nerineid gastropods, ostreid bivalves and *Pleuronoma* within thinly bedded peloidal packstones and wackestones. At Gorges de Court, near Moutier (Fig. 7), the maximum flooding surface is at the top of a bioturbated nodular peloidal packstone bed that contains perisphinctid ammonites, brachiopods and gastropods. Across the carbonate platform, the highstand systems tract is composed of peloidal and oolitic limestones; near the top of the highstand systems tract, stromatolitic limestones again prograde basinward immediately prior to the next sequence boundary O6 (e.g., Sornetan (Gorges du Pichoux), Fig. 5B). In the basin, the maximum flooding surface is at the base of the Gerstenhübel Limestone (Fig. 6B) which is traceable over tens of kilometers and marks the maximum increase of carbonate content within the basin compared to the rest of the Effingen Member. The highstand systems tract comprises the tabular micritic limestones of the Gerstenhübel Beds. Occasional blocks of plastically deformed micritic mudstone have been found within the micritic limestone beds of this unit, suggesting that they were deposited by submarine debris flows.

**Sequence Boundary O6 (Grossouvrei–Hypselum Subzone Boundary)**

The depositional sequence above O6 contains a facies assemblage similar to that overlying sequence boundary O5. Sequence boundary O6 is conspicuous in many places; at Röschenz (RG 402; Fig. 5A) it is marked by a renewed influx of sandy marl (3.2 m); at Gorges du Pichoux, near Sornetan, the boundary is a distinct angular erosion plane within stromatolitic limestones containing freshwater algae (Characean gyrogonites; section RG 315; Fig. 5B). Near Vermes, it is a stromatolite (bed no. 48) below a black pebble conglomerate (Fig. 1; Gygi and Persoz, 1986, Fig. 5, section RG 406). At Gorges de Court, near Moutier, it is marked by stromatolites (Fig. 7). At La Chaques quarry, near Péry, there is a well-developed thin bed of lignite (Fig. 6A). Pieces of wood up to 2 cm in diameter have been found at this locality, and seams of lignite occur at several other localities (Gygi and Persoz, 1986, p. 399).

In the basin, sequence boundary O6 occurs at the top of the Gerstenhübel Limestone and is marked by the renewed input of a thick succession of interbedded argillaceous, locally sandy marls and thin micritic limestones that form the upper Effingen Member. The upper Effingen Member contains erosional truncation surfaces infilled by submarine debris flows of plastically
deformed, laminated, silty pelsparite clasts within a marl matrix (Gygi and Persoz, 1986, Figs. 2, 3). Localized fault topography may have given rise to slopes on which the debris flows formed. Similar to the lower Effingen Member, the upper Effingen Member is interpreted to have been deposited by turbidity currents.

**Transgressive Surface.**

During the Late Oxfordian Age the morphology of the carbonate platform and basin changed from a rimmed-shelf to an epeiric carbonate platform. As a result, the transgressive surface above O6 is distinctive and laterally continuous. The short-term (3rd-order) transgression above O6 caused grainstone shoals to form far into the platform interior such as near Liesberg (Fig. 5A). At Gorges de Court and Vermes, there is a well-developed marine hardground with borings and encrusting bivalves. Coral bioherms resumed growth west of Olten near the edge of the basin and pure micritic limestones developed near Villigen. The transgressive systems tract is thus represented by backstepping of facies belts and re-establishment of carbonate facies.

**Maximum Flooding Surface.**

In the platform interior, the maximum flooding surface above sequence boundary O6 is at the base of the Hauptmumienbank Member (Figs. 2, 3). The onlapping nature of this maximum flooding surface in the proximal (northwest) direction is well-displayed on the Cantonal road 1.7 km east of Liesberg (Fig. 1), where the Hauptmumienbank rests unconformably on planed off sand waves of the peloidal grainstone of the upper Röschenz Member (Gygi and Persoz, 1987, Fig. 2A). The Hauptmumienbank is a lagoonal oncilite with oncocids up to 3 cm in diameter floating in a lime mud matrix (Ziegler, 1956, Plate 1, Fig. 1) and is interpreted as the highstand systems tract. The time-equivalent strata towards the basin are the Steinbach Member, which comprise oolitic grainstones (Figs. 2, 11). This grainstone unit prograded about 20 km basinward over the transgressive systems tracts of the upper Effingen Member (Fig. 3). At the platform margin (e.g., Olten), the fringe of coral bioherms that were initiated at the transgressive surface continued to grow (Figs. 2, 3). The micritic limestones of the Geissberg Member were laid down in the adjacent shallow marginal area immediately distal of the bioherms (Fig. 2). This part of the Geissberg Member contains a macrofauna of mainly infaunal bivalves. The member thins out and becomes marly in the distal direction.

**Modification to the Carbonate Platform Morphology and Facies**

The morphology of the carbonate platform, and hence the type of facies, changed in the middle of the *hysselum* Subzone from the rimmed-shelf type of carbonate platform to an epeiric carbonate platform (at the transgressive surface above O6). The epeiric carbonate platform that developed was characterized by widespread, poorly defined facies belts which included micrites, oncotic and peloidal packstones and wackestones, thickly bedded grainstones and packstones. The more distal sections, southeast of Villigen (Fig. 3), are characterized by variably glauconitic ammonite-rich micrites and sponge bioherms.

**Sequence Boundary O7 (Lower Bimammatum Subzone)**

**Sequence Boundary and Transgressive Surface.**

Across the platform, sequence boundary O7 is a planar erosion surface associated with a renewed input of terrigenous siliciclastic material. At Liesberg and Röschenz (Fig. 5A), the boundary is overlain by a thin bed of sandy claystone and at Gorges de Court, near Moutier, sandy oolitic facies (Oolite Rousse, Figs. 5B, 7). Proximally, i.e. towards the northwest from Liesberg, there is further evidence of terrigenous input from the fairly extensive package of marl deposited that comprises the Bure Member (formerly Humeralis Marl, Gygi and Persoz, 1986; Figs. 2, 3). Near Péry, the boundary O7 is a well-marked limonitic and corroded bedding plane overlain by a peloidal marl containing a variety of macrofossils (Figs. 6A, 11). The superimposition of oolitic grainstones on top of oncotic and peloidal wackestones at this boundary between Liesberg and Moutier, the marine borings at Roches (Fig. 1, RG 372, bed 50, sample Gy 4348) and the fossiliferous marl overlying oolitic packstones near Péry indicate that across the epeiric carbonate platform this sequence boundary and the transgressive surface are superimposed. The transgressive systems tract in the platform interior comprises the oolitic grainstones of the Oolite Rousse Member and the oncotic and oolitic mudstones, wackestones, and packstones of the basal part of the Laufen Member (Figs. 2, 5B, 7). Distally from the main platform, there is an areally restricted shelf-margin systems tract that comprises oolitic grainstones at Linn (Fig. 1) and 40 cm of micritic (poorly glauconitic) limestones at the base of the Crenularis Member (e.g., Villigen, Fig. 6B; section 62 of Gygi, 1969, Plate 17). This is overlain by the glauconitic, fossiliferous micrites of the Crenularis Member, which is interpreted as the transgressive systems tract.

**Maximum Flooding Surface.**

At Gorges du Pichoux, near Sornetan and in a nearby exposure at Souboz (RG 312, bed no 87), the maximum flooding surface is marked by numerous macrofossils (ostreids, brachiopods, serpulids) that form up to 50% of the rock volume. At Wäberhüsli, near Oberdorf (Figs. 1, 6), the maximum flooding surface is marked by a concentration of bivalves within oolitic micrite. In the basin, the maximum flooding surface is indicated by the maximum glauconite content, which is coincident with the top of the Crenularis Member (e.g., Villigen, Fig. 6B; section 62 of Gygi, 1969, Plate 17). This is overlain by the glauconitic, fossiliferous micrites of the Crenularis Member, which is interpreted as the transgressive systems tract.

**Sequence Boundary O8 (Late Hauffianum Subzone)**

**Sequence Boundary and Transgressive Surface.**

Sequence boundary O8 is at the base of the massive, widespread oolitic grainstones of the Verena Member. In fresh outcrops, the boundary O8 can be easily discerned by the color change between the underlying bedded, light-grey to brownish
Laufen Member and the overlying massive yellowish to pure-white Verena Member. The widespread superimposition of oolitic grainstone over different facies across the platform indicates that sequence boundary O8 and the transgressive surface are combined (Fig. 2). We therefore assign the whole of the Verena Oolite to the transgressive systems tract. The great thickness of oolitic limestone preserved is attributed to rapidly increasing sediment accommodation space produced during the combination of a short-term relative sea-level rise and the long-term Late Oxfordian transgression.

The horizon containing large blackened limestone pebbles in the exposure at La Reuchenette quarry, near Péry (Fig. 6A) is only local and probably represents a local flooding surface within the shallow-water Verena Oolite. This horizon may be time equivalent to the micritic intercalation within the upper Holzflue Member, near Balsthal.

Sequence boundary O8 is well-defined in the basin. Near Villigen (Schrannechopf; Figs. 6B, 12), the sequence boundary and transgressive surface have coalesced and form an uneven corroded bedding plane very near the top of the Wangen Member. The corroded bedding plane is overlain by small limestone nodules which are surrounded by a glauconitic marl (Fig. 12; Gygi, 1969, Plate 17, section 62). Below and above the hummocky bedding plane are oxidized nodules of iron sulfide with a diameter of up to 3 cm. The next bed (bed 59, Gygi, 1969, Plate 17, section 62) is slightly condensed and is 0.3 m thick. It contains siliceous sponges, an abundance of other macrofossils and a small amount of glauconite. Bed 59 and the underlying nodules enclosed in glauconitic marl are called the Knollen Bed (Gygi, 1969, p. 72; Fig. 2). The Knollen Bed forms a marker horizon that can be followed over a distance of more than 120 km into southern Germany (Gygi and Persoz, 1986, p. 408) and it is interpreted as the base of the transgressive systems tract. The Knollen Bed was first recorded and described by Mœsch (1867), but surprisingly, it has never been mapped by subsequent geologists working in the Jura.

Maximum Flooding Surface and the Oxfordian–Kimmeridgian Stage Boundary.—

In northern Switzerland, the Oxfordian–Kimmeridgian boundary is conventionally assumed to be between the pure bedded limestone of the Letzi Member and the overlying, somewhat glauconitic Baden Member; this corresponds to the boundary between the galar Subzone of the planulata Zone and the planulata Zone (see Gygi and Persoz, 1986, Table 2). The boundary between the galar Subzone and the planulata Zone was established precisely in an excavation near Schaffhausen (section RG 239; Gygi, 1990b, p. 69). Ammonites are relatively concentrated at the boundary, and it is interpreted as a maximum flooding surface. At Mellikon on the slope (section RG 70; Figs. 3, 6B), the top of the uppermost micritic bed of the Letzi Member is a very distinctive hummocky and mineralized surface.

Proximally, the Oxfordian–Kimmeridgian boundary is difficult to correlate because of the lack of ammonites over this interval. However, it was traditionally placed at the distinct lithological boundary between the very pure, massively bedded limestone of the Verena Member and the somewhat darker, thinly bedded limestones and marls of the lowermost Reuchenette Formation (Figs. 2, 3, 5A, 5B, 7, 6A). This is partly corroborated by mineralostratigraphic correlations using kaolinite (Gygi and Persoz, 1986, Fig. 10) but the correlation is not entirely confirmed because it is at the limit of the resolution as the planulata Zone is only represented by 20 cm of sediment at Mellikon where the mineral stratigraphy was calibrated with the ammonite biostratigraphy. Lithologically, the change from the massive oolitic grainstones of the Verena Member to the thinly bedded lime mudstones of the Reuchenette Formation is interpreted as a maximum flooding surface, which assuming that the maximum flooding surface is isochronous over northern Switzerland, further confirms the correlation of the Oxfordian–Kimmeridgian boundary between the platform and the basin.

Biostratigraphically, correlation of the Oxfordian–Kimmeridgian boundary across Europe is uncertain. In the Boreal and northwest European Province, it is defined as the base of the baylei Zone. Sykes and Callomon (1979, p. 894) thought that “there were indications that the Oxfordian–Kimmeridgian boundary in the Sub-Mediterranean Province must be drawn lower than normally accepted.” Recent biostratigraphic data (Schweigert and Callomon, 1997; Matyja and Wierbowski, 1997) documenting the occurrence of Amoeboceras bauhini (Oppel) and Pictonia densicostata (Salfield ms) in sections in Germany, Poland, England and Scotland indicate that the lowermost faunal horizon of the baylei Zone (northwest European and Boreal provinces) is equivalent to the uppermost faunal
horizon of the *hauffianum* Subzone (*bimammatum* Zone) of the Sub-Mediterranean province. Thus implying that the Oxfordian—Kimmeridgian boundary (*sensu anglico*) correlates with a much older surface in Switzerland than that taken herein. However, other workers (e.g., Mégnien and Mégnien, 1980; Hantzpergue, 1988, p. 497) have placed the stage boundary in the Submediterranean province at the top of the *planulata* Zone, thus indicating that the base of the *platynota* and *baylei* Zones are synchronous.

An excellent example of a maximum flooding surface is recognized in the Wessex Basin, southern England, at the Oxfordian—Kimmeridgian boundary (Coe, 1992a, 1992b). Hence, if the correlation between the English and the Swiss sections is based solely upon the sequence stratigraphic interpretation it indicates that the stage boundary is synchronous in the two sections or that part of one of the sections is absent, thus supporting the biostratigraphic evidence of Mégnien and Mégnien (1980), and Hantzpergue (1988). A third alternative is that the maximum flooding surfaces are of different ages in Switzerland and England.

**Kimmeridgian Sequences**

In Switzerland, the Kimmeridgian sequence boundaries are poorly defined due to the thick shallow-water marine and non-marine limestones on the platform and condensed marine limestones within the basin. However, on the platform, there are three distinct sedimentary cycles. In general, each of these cycles contains the following sediments: at the base occur thinly bedded argillaceous and occasionally sandy limestones associated with hypersaline and/or freshwater faunas, subaerial exposure surfaces and stromatolites. These are overlain by thinly bedded peloidal limestone and capped by a very massive bed of micritic limestone. The influx of terrigenous material and evidence of subaerial exposure is interpreted as the sequence boundary and is usually overlain by a very thin shelf-margin systems tract. The thinly bedded packstones and wackestones are interpreted as the transgressive systems tract deposited under higher energy conditions as relative sea-level rises. The massive micritic beds are interpreted as the highstand systems tract deposited fairly rapidly as relative sea level began to fall again at the end of the relative sea-level cycle.

**Sequence Boundary K1 (Platynota—*Hypselocyclum* Zonal Boundary)**

**Sequence Boundary and Transgressive Surface.**—

The first sequence boundary within the Reuchenette Formation is about 7 m above the base of the formation. At Liesberg (Fig. 5A), it is at the contact between peloidal limestone and the overlying thinly bedded marly limestone. At Courgenay-Chemin Paulin (southeast of Porrentruy; section RG 350; Fig. 1) and Gorges du Pichoux, near Sornetan (Fig. 5B), it is at the base of a distinct package of thinly bedded marly limestone. At Oberdorf (Wäberhüsli; Fig. 6A), it is at the boundary between oncolitic limestone and the overlying thinly bedded sandy micrite. All these sections indicate that K1 is associated with a renewed input of terrigenous material. There is good evidence of subaerial exposure associated with this sequence boundary near Balsthal (Fig. 1; sections RG 438-Steinebach-tobel, RG 439-Innere Klus and RG 450-Chluser Roggen). After removing the shortening effect of Tertiary folding (palinspastic reconstruction), this gives an area with a diameter of at least 2.5 km. This subaerial exposure surface lies several meters above the base of the Reuchenette Formation. At Péry, sequence boundary K1 is marked by a limonitized surface of micritic limestone that is overlain by a stromatolite and peloidal limestone.

At Balsthal (Fig. 1), one ammonite, *Lithacosphinctes evolutus* (Quenstedt) (specimen J 30530 Natural History Museum, Basel) indicative of the *platynota* Zone, has been found 2.5 m below K1 in the lowermost Reuchenette Formation (section RG 439). At the quarry of Born, near Olten, *Ataxioceras* (*Ataxioceras* aff. *catenatum* Schneid (J23097 Natural History Museum, Basel), indicating the middle of the *hypslocyclum* Zone, was found about 15 m above the base of the Reuchenette Formation. Therefore we conclude that sequence boundary K1 is near, or at, the boundary of the *platynota* and *hypslocyclum* Zones. At all localities across the platform, the thin package of sediment containing terrigenous material is interpreted as the shelf-margin systems tract, and the overlying thinly bedded peloidal wackestones, packstones and grainstones are interpreted as the transgressive systems tract.

In the basin, sequence boundary K1 is within the lower part of the condensed, nodular, glauconitic and ammonitiferous limestone of the lower Baden Member (e.g. Mellikon, RG70; Gygi, 1969; Fig. 6B). The sequence boundary and transgressive surface in the basin are difficult to separate in this condensed facies.

**Maximum Flooding Surface.**—

On the platform, the maximum flooding surface is at the boundary between medium-bedded micritic limestones and a thick bed of homogeneous micrite. The homogeneous micrite is typically about 6 m thick and was deposited under quiet water conditions, it forms the highstand systems tract (e.g., Oberdorf; Fig. 6A). At the disused quarry of La Rasse, 1 km south of Porrentruy (Fig. 1), the succession above sequence boundary K1 is well-exposed; here, the transgressive surface is at the base of an oolite containing corals, bivalves and gastropods (bed number 46, section RG 340), and the maximum flooding surface is a slightly glauconitic horizon (bed number 23). This glauconitic horizon is developed as several thick limestone beds with a rich fauna of bivalves and large nautiloids in the old quarry adjacent to the Fontenais cemetery south of Porrentruy.

**Sequence Boundary K2 (Divisum Zone)**

**Sequence Boundary and Transgressive Surface.**—

Sequence boundary K2 is easily recognized both within the basin and on the platform. It is usually just over 20 m above the base of the Reuchenette Formation on the platform. At Oberdorf (Wäberhüsli; Fig. 6A), it is at the corroded upper surface of bed number 34, which is covered by a limonite crust. The sequence boundary can also be recognized in the western part of the section at Steingruben, Lommiswil, west of Oberdorf (RG 434) as a limonitic crust on the planar top surface of micritic limestone. In the Gorges du Pichoux section, near Sornetan, the boundary K2 is marked by a thin marl (Fig. 5B). Near Porrentruy, there is a bored erosion surface that is inter-
interpreted as sequence boundary K2 (section RG 340, bed number 6). Shallow-water facies and subaerial exposure, associated with sequence boundary K2, occur in the section west of Glogelier (RG 352) where about 7 m of sediment below the Banné Member (Figs. 1, 2, 5A) contain stromatolites and bedding planes with prism cracks. Similar to the sequence above K1, the transgressive systems tract again comprises thinly bedded peloidal packstones and wackestones on the platform (e.g., Sorinetan; Fig. 5B).

In the basin, near Mellikon (RG 70), sequence boundary K2 is interpreted as the upper surface of the glauconitic, fossiliferous lower Baden Member (Fig. 6B; Gygi, 1969, Plate 17, bed number 124). There is no evidence of substantial omission on this bedding plane except that it is uneven. The first ammonites of the *divisum* Zone were found just below the bedding plane, indicating that the sequence boundary is within the *divisum* Zone. The sequence boundary is overlain here by 1 m of poorly fossiliferous, massive marl with up to 40% terrigenous clay; and interpreted as a thin, local shelf-margin systems tract. The transgressive surface near Mellikon is at the contact between the marl and the overlying micritic limestone bed that forms the base of the Wettingen Member.

**Maximum Flooding Surface.—**

The Banné Member, which is exposed near Courgenay southeast of Porrentruy (section RG 341), lies 50 m above the base of the Reuchenette Formation (Fig. 2). This marly member has a rich bivalve macrofauna together with some brachiopods, nautiloids and rare ammonites. The maximum flooding surface is interpreted to be within this member. The surface is dated from the ammonite *Aspidoceras cf. acanthicum* (J 30714 Natural History Museum, Basel) that was found by H. and A. Zbinden in section RG 341, probably in bed 37, near Courgenay. This maximum flooding surface is well-marked in southern Germany; it is within the bedded fossiliferous marl (white Jura delta 2) that is also dated as upper *acanthicum* Zone by ammonites (Fig. 2).

The highstand systems tract on the platform is again marked by about 6 m of thickly bedded micrite similar to the highstand systems tract above K1 (e.g., Oberdorf, Fig. 6A).

**Sequence Boundary K3 (Eudoxus Zone)**

**Sequence Boundary and Transgressive Surface.—**

Sequence boundary K3 is again characterized by the influx of argillaceous material and a limonitized surface on the platform. The boundary is well developed near Oberdorf, where it is at the base of the famous Solothurn Turtle Limestone (Fig. 2). Lithologically, near Oberdorf, (section RG 433, Fig. 6A), K3 is the boundary between a micritic limestone and overlying thin layer (3 cm) of soft marl. One kilometer to the west, in the eastern quarry of Steingrueben (section RG 434), the sequence boundary is interpreted as the hummocky upper surface of bed 12, which is covered with a limonitic crust. The exact age of sequence boundary K3 is not clear, but it is interpreted to be at, or near, the base of the *eudoxus* Zone.

The Solothurn Turtle Limestone Member of the lower Reuchenette Formation begins, as defined by Thalmann (1966, Fig. 13), about 35 m above the base of the Reuchenette Formation, a thickness that corresponds remarkably closely to that for the section at Oberdorf where K3 is 37.5 m above the base of the Reuchenette Formation (section RG 433, Fig. 6A). The thickness of the Reuchenette Formation near Solothurn of more than 120 m, as given by Thalmann (1966, p. 89), is erroneous. It is at most 50 m because of thickness reduction by pre-Tertiary erosion. The thickness of the Solothurn Turtle Limestone is, according to Lang and Rütimeyer (1867, p. 11), about 9 m.

Limmic ostracods from the Bargetzi quarry near Solothurn (Thalmann, 1966, p. 105) and footprints of large sauropods (Meyer, 1989, p. 189) in the eastern quarry of Steingrueben west of Oberdorf (section RG 434, upper surface of bed 17) indicate that the Solothurn Turtle Limestone Member and its time equivalents are shallow-water limestones with exposure surfaces associated with sequence boundary K3.

**Maximum Flooding Surface.—**

Near Solothurn and Oberdorf, the freshwater limestones associated with sequence boundary K3 are overlain by a distinct unit of thinly bedded micritic limestones (6–9 m thick) that have been assigned to the Portlandian Stage (*sensu gallico*) by several authors (e.g., Buxtorf, 1907, p. 59) because it contains *Nanogryra cf. virgula* (Meyer, 1989, p. 188). The occurrence of an ammonite together with the oysters suggests that this unit is marine. This thinly bedded limestone begins 46 m above the base of the Reuchenette Formation in section RG 433 near Oberdorf (Fig. 6A). The maximum flooding surface is interpreted to be near the base of these thinly bedded marine limestones.

The ammonite *Aulacostephanus (Aulacostephanoceras) autissiodorensis* (Cotteau) was found in the Bargetzi quarry near Solothurn (Gygi, 1995, Fig. 24; specimen 10842 Naturmuseum Solothurn). The exact horizon of the ammonite in the quarry is not known. Meyer (1989, p. 188) thinks that it was found at the base of the thin-bedded limestone member just above the Solothurn Turtle Limestone. The vertical range of the ammonite species is the lower part of the *autissiodorensis* or equivalent *beckeri* Zone. This would place the maximum flooding surface in the upper part of the *eudoxus* Zone or lower *autissiodorensis* Zone.

Elsewhere in northwestern Switzerland, the maximum flooding surface above K3 is marked by the distinctive and laterally extensive Virgula Member (Fig. 2). The ammonite *Aspidoceras caletanum* (Oppel) (J 27976 Natural History Museum, Basel, Gygi, 1995, Fig. 26), indicative of the *eudoxus* Zone, was found in the lowermost part of the limestones above the Virgula Member at Alle east of Porrentruy by H. Zbinden. Towards the south, the Virgula Member grades into limestone. The Virgula Member contains a glauconitic bed that must be the time equivalent of the glauconitic marl between the White Jurassic delta 3 and 4 in southern Germany. Delta 3 and 4 represent the *eudoxus* Zone (B. Ziegler, 1962, Table 1). The “Portlandian” that Laubscher (1963, p. 12) indicates above the Virgula Member near Porrentruy then has nothing to do with the Portlandian Stage *sensu gallico*. These limestones are the lower part of the upper Reuchenette Formation and are Kimmeridgian in age. This late Eudoxus age maximum flooding surface is well-marked in England and the North Sea, usually by organic-rich marine shales (Coe, 1992b).
SEQUENCE STRATIGRAPHIC ANALYSIS OF THE SECOND-ORDER TRANSgressive-REGRESSive CYCLES

Oxfordian Stage

The Oxfordian Stage of northern Switzerland is interpreted as representing a single 2nd-order regressive-transgressive cycle as defined by Duval et al. (1992) and Vail et al. (1991). The regressive period started at the Oxfordian-Callovian boundary and continued until mid-hypselum Subzone time; this was followed by a transgressive period that lasted until the end of the Oxfordian Stage. This regressive-transgressive cycle is demonstrated by the stacking pattern and relative development of the 3rd-order systems tracts, the vertical changes in facies and by the initiation and subsequent development of the carbonate platform. The progressive development of thick 3rd-order lowermost systems tracts during the 2nd-order regressive period, compared with thick transgressive and highstand systems tracts during the 2nd-order transgressive period, is clearly shown in Figure 3 and is described in the summary below.

The boundary between the regressive and transgressive periods in the Oxfordian Stage is coincident with the morphological change from a rimmed-shelf type of carbonate platform to an epeiric carbonate platform. The epeiric carbonate platform probably extended into northern and central France, and was present until at least the end of the Kimmeridgian Age. Towards the southeast, the epeiric carbonate platform sloped off into the Rhodano-Swabian epicontinental basin (Gygi, 1992) in two stages, with a small flat area in between. The carbonate platform morphology appears to have changed in response to two factors; firstly, because of the great thickness of the underlying Effingen Member which filled a large portion of the sediment accommodation space that was present distally from the edge of the rimmed-shelf. Secondly, the morphology changed because of the overall long-term (2nd-order) late Oxfordian transgression, which was probably caused by a phase of regional, uniform tectonic subsidence. The response to these two factors was that the platform area increased dramatically, and the edge became poorly defined and moved southeastwards to Olten. This 2nd-order transgressive period also appears to have prevented preservation of any lowermost systems tract above sequence boundaries O7 and O8 across the carbonate platform (Fig. 3).

Kimmeridgian Stage

The facies change at the Oxfordian–Kimmeridgian boundary, from thick widespread marine carbonates to thin, laterally variable shallow-marine and nonmarine carbonates, indicates that the lower part of the Kimmeridgian was deposited during a minor regression. This is supported by the development of 3rd-order lowermost systems tracts during this period (i.e., above sequence boundaries K1, K2 and K3; Figs. 2, 3) in a manner similar to the enhanced development of lowermost systems tract during the Oxfordian regressive period. The subsequent Kimmeridgian transgressive phase started near the base of the eu-doaxes Zone and is well marked by the deposition of micritic marine limestones across the carbonate platform and glauconite-rich sediments in the basin. This transgression continued into the Tithonian Age.

SUMMARY OF THE DEPOSITIONAL HISTORY AND THE INTERACTION OF THE SECOND- AND THIRD-ORDER CYCLES

Oxfordian Second Order Regressive Period

Second-order regression throughout the Early and Middle Oxfordian Stage is indicated by the progressive change from marine clays and marls to shallow-marine carbonates, including the establishment of a rimmed-shelf carbonate platform. In addition, terrigenous input into the basin was high during the latter part of the regression and resulted in deposition of the thick Effingen Member. Overall during the regression, the 3rd-order shelf-margin systems tracts show a progressive increase in both absolute thickness and amount of time represented, whereas the transgressive and highstand systems tract decrease (Figs. 2, 3).

The Callovian–Oxfordian boundary is a well-marked 3rd-order maximum flooding surface, because it is coincident with the 2nd-order peak transgression. During the Early Oxfordian Stage, clays were deposited in the northern part of the area (Renggeri Member) and silty clays during relative sea-level fall. In sediment-starved basinal areas, iron oolite-rich sediments were deposited. The first carbonates were deposited during relative sea-level fall during Early Oxfordian time (cordatum Subzone, sequence boundary O2). During deposition of the antecedens Subzone sediments (above sequence boundary O4), a rimmed-shelf carbonate platform developed with fringing coral reefs. Sequence boundary O5 (parandieri–schilli Subzonal boundary) is overlain by a thick, well-developed shelf-margin systems tract in the basin, composed of redeposited carbonates and terrigenous claystones and siltstones (lower Effingen Member). Simultaneously, the coral bioherms on the platform margin prograded, and shallow-water carbonates with subaerial exposure surfaces were deposited across the platform. A 3rd-order transgression is marked by aggradation of the bioherms, retrogradation of the carbonate facies across the platform and decreased terrigenous deposition in the basin. The overlying sequence O6 (hypselum Subzone) is similar to the sequence O5 but lacks the fringing coral reefs. Peak 2nd-order regression is near the top of the Effingen Member.

Oxfordian Second-Order Transgressive Period

The transgressive period is marked by the development of a much more widespread epeiric carbonate platform, with laterally extensive facies belts. The 3rd-order sequences are thick, and lowermost systems tracts are not developed. The onset of 2nd-order transgression in the middle of hypselum Subchron (above sequence boundary O6) is marked by a cutoff in terrigenous input and the deposition of pure carbonate. The 3rd-order maximum flooding surface above O6 is marked by widespread carbonate deposition across the platform and in the basin (Hauptmunienbank Member and equivalent). During the Late Oxfordian Age sequence boundaries O7 and O8 and the transgressive surfaces were superimposed across the carbonate platform. Thus, the overlying transgressive and highstand systems tracts preserved are dominated by thick, relatively homogeneous carbonates due to the rapid increase in sediment accommodation space during this 2nd-order transgression. In the deep basin, fossiliferous glauconitic limestones were deposited during 3rd-order transgressions (e.g., Knollen Bed) and thick carbonate highstands were developed (e.g., Küssaburg Member;
Fig. 3). The Oxfordian–Kimmeridgian boundary is interpreted as a 3rd-order maximum flooding surface and a minor peak 2nd-order transgression.

Kimmeridgian Second Order Minor Regressive Period

The Kimmeridgian sediments display three distinct carbonate cycles. On the platform, they comprise shallow-marine and freshwater carbonates. At the base of each cycle, there is evidence for subaerial exposure; the exposure surface is overlain by wackestones and packstones, and the cycle is capped by thick micrite. Each of these cycles is interpreted as a 3rd-order depositional sequence. In the basin, the three sequences are developed within fairly condensed strata. Above K2 there is a distinct shelf-margin systems tract of terrigenous clay similar to the shelf-margin systems tracts developed during the Oxfordian regressive period. Maximum 2nd-order regression (minor) is marked by the Solothurn Turtle Limestone and equivalent strata.

Kimmeridgian Second Order Transgressive Period

The sediments deposited from the lower *eudoxus* Subzone and into the Tithonian are interpreted as transgressive. On the carbonate platform, micritic limestones with ostreids were deposited, and in the basin glauconitic sediments accumulated. The maximum flooding surface in the upper *eudoxus* Zone is widespread and well-developed, because it is near the start of the 2nd order transgression.

CONCLUSIONS

Sequence stratigraphic analysis of the Oxfordian and Kimmeridgian Stages of northern Switzerland allows further refinement of the correlation between the carbonate platform and the basin, and provides an elegant explanation for the facies complexity observed. The numerous sections along the carbonate platform-to-basin transition enables detailed study of the lateral changes that occur along key sequence stratigraphic surfaces and systems tracts. The Oxfordian represents a long-term regressive-transgressive cycle that is further subdivided into eight sequences. The lower Kimmeridgian strata represents a minor regression that can be subdivided into three sequences and was followed by transgression. The long-term change from regression to transgression in the Oxfordian Stage coincides with the carbonate platform changing from a rimmed-shelf to an epeiric carbonate platform and most likely relates to regional tectonics. The sequence boundaries are the most laterally widespread surfaces and are usually associated with the main facies changes. Within each of the sequences, the transgressive surface (associated with the onset of transgression) and the maximum flooding surface are generally well-developed both on the carbonate platform and in the basin.

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 rewritten in 2018.
ABSTRACT: A virtually complete succession of shallow-marine, epicontinental Lower Jurassic strata crops out in excellent exposures along the North Yorkshire coast (U.K.). Borehole control of the inland continuation of these deposits, together with the availability of a detailed ammonite biostratigraphy, allows for the construction of a cross-section correlating the relatively thick succession of the Cleveland Basin (exposed in the coastal sections) with the more condensed succession on the adjacent Market Weighton High (encountered in boreholes). This study is concerned with only the Hettangian to Pliensbachian part of the Liassic succession, corresponding to the informal ‘lower Lias’ and ‘middle Lias’ divisions of traditional usage.

Detailed examination of sedimentological features within a biostratigraphic framework revealed a systematic set of changes occurring both in the basin and on the high in medium-scale depositional sequences (10 to 60 m thickness; duration of 1 to 3 ammonite biozones). These sequences show an overall shallowing-upward trend from mudstone, deposited below storm wave-base, through quartz siltstone and sandstone or shelly mudstone, deposited above storm wave-base in the shoreface domain, to Fe-enriched deposits (limonite, glauconite, chamosite, ferruginous ooliths) at the top of the sequence. The mudstones were deposited relatively rapidly and extend over both basin and high. The shelly mudstones, siltstones and sandstones represent much lower overall accumulation rates. The Fe-enriched deposits represent periods of condensation and are best developed on and immediately around the topographic Market Weighton High.

Three orders of depositional sequences have been identified: (1) 3 large-scale, 2nd-order sequences, that record major sea-level rises in the planorbis/liasicus zones, semicostatum zone and jamiensis zone and are bounded by major erosional phases in the bucklandi and margaritatus zones and a basin-wide shallowing in the rariocostatum zone; (2) 6 medium-scale, 3rd-order sequences that are bounded by phases of reduced, Fe-rich sedimentation or erosional/nondepositional surfaces and that have a duration in the order of one ammonite subzone; and (3) several types of small-scale, high-frequency, 4th and 5th order cycles.

The transgressions of the 2nd-order depositional sequences (early Hettangian, early Sinemurian, early Pliensbachian times) are recognized worldwide and are thus most probably of eustatic origin. Tectonic uplift of the Market Weighton High influenced the expression of the medium-scale, 3rd-order sequences. Detailed comparison with other British and European sites is required to further clarify this aspect. Long-term climatic influence was also important in setting the scene for the abundant supply of iron, which became concentrated in the marine environment. At a higher frequency, climate influenced sediment distribution within the shallow epicontinental basin by relatively short-term, orbitally induced variations in the weather conditions.

INTRODUCTION

One of the most complete successions of Lower Jurassic strata in northwestern Europe is exposed in the coastal sections of North Yorkshire in the United Kingdom. The excellent quality of the outcrops in the cliff face and on the foreshore has encouraged detailed lithostratigraphic and biostratigraphic studies of these rocks since the early 19th century (e.g., Tate and Blake, 1876; Barrow, 1888; Buckman, 1915; Fox-Strangways and Barrow, 1915; Arkell, 1933; Bairstow, 1969; Hallam, 1975; Cope et al., 1980; Knox et al., 1990; Hesselbo and Jenkyns, 1995).

The rich variety of sedimentological features in the Redcar Mudstone, Staithes Sandstone and Cleveland Ironstone Formations of Powell (1984) (lower and middle Lias of traditional usage) have been the subject of a number of detailed studies (Sellwood 1970a, 1970b, 1972; Shalaby, 1980; Greensmith et al., 1980; Rawson et al., 1982; Howard, 1985; van Buchem and McCave 1989; van Buchem, 1990; van Buchem et al., 1992; van Buchem et al., 1994). These formations display a limited number of distinct lithofacies, consisting of mudstones (with few shells), shell-rich mudstones, siltstones, sandstones and Fe-enriched deposits (e.g., Hallam and Bradshaw, 1979). The occurrence of Fe-enriched deposits (ferruginous oolithes and related chamosite and glauconite) have been the subject of a number of specific studies (Dunham, 1960; Chowns, 1966; Hallam, 1967; Catt et al., 1971; Hemingway, 1974; Greensmith et al., 1980; Rawson et al., 1982; Howard, 1985; Myers, 1989).

Three main forces have been identified as influencing the geographical and temporal distribution of Lower Jurassic sediments in southern Britain: local tectonism, regional tectonism and different orders of eustatic sea-level variation (Sellwood and Jenkyns, 1975; Hallam and Sellwood, 1976). The local tectonism was responsible for the division of the depositional area into a pattern of basins and ‘swells’ (structural and topographic highs). The aim of this paper is to look in detail at the sedimentation pattern in one such basin, the Cleveland Basin, and to compare its sedimentation pattern with that of the adjacent ‘swell’, the Market Weighton High.

To evaluate the expression of the depositional trends in these two settings, we will first describe the distribution of the different lithofacies within the biostratigraphic framework, then propose a subdivision into different orders of depositional sequence, and finally discuss the relative importance of tectonic, eustatic and climatic control on the sedimentation pattern. For this purpose, a cross section has been constructed correlating the basal deposits of the Cleveland Basin (Barnard and Cooper 1983), as exposed in the coastal sections, with the more condensed sections identified in boreholes on the adjacent, tectonically active Market Weighton High (Bott et al., 1978; Fig. 1).

LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY

The study is based on a detailed sedimentological reexamination of the lower Liassic outcrops along the Yorkshire coast (Robin Hood’s Bay, Redcar and Hantscliff; see field logs in Fig. 2) and the reexamination of core material of the Felixkirk, Brown Moor and Howsham boreholes (Fig. 1). The new observations on the core material have been incorporated into the
Cope et al. (1980) and the more recent work by Phelps. The biostratigraphic subdivision is based on the compilation by van Buchem et al. (1992). A revision of the position of the *obtusum* zone by Page (1992) and adopted by Hesselbo and Jenkyns (1995) has been indicated in Figures 2a and 2b. However, this revision has not been applied in the correlation scheme because of the need to maintain compatibility between the coastal section and the borehole sections, all of which have been dated on conventional biozonal criteria.

**FACIES AND SURFACES**

The detailed lithostratigraphic and biostratigraphic data set provides an excellent basis for a sequence stratigraphic analysis (e.g., van Wagoner et al., 1988, 1990; Vail et al., 1991; Homewood et al., 1992). Before proceeding to a subdivision in depositional sequences, a short environmental interpretation of the different lithofacies is given, and the key correlation surfaces are identified.

**Lithofacies Types**

Three main lithofacies associations are distinguished: (a) mudstones (excluding those with abundant shell material), (b) quartz siltstones, sandstones and shelly mudstones, and (c) Fe-enriched deposits. Their environmental interpretation is discussed briefly, and schematically presented in Figure 3. Since this paper focuses on their stratigraphic distribution, the reader is referred to the literature mentioned below for detailed sedimentological information.

**Mudstone Facies.**—

The mudstone units (Figs. 2, 3) consist of dark to light grey clay (illite, kaolinite, smectite; van Buchem et al., 1992), which is commonly very micaceous and is locally silty. Pyrite can be common (especially in the Pyritous Shales) and occurs dispersed throughout the mudstone in the form of small frambooids and spherules. Calcite and siderite nodules are a common feature and normally occur in horizons. The organic carbon content varies around 0.8% for the Redcar Mudstone Formation, with somewhat elevated values at the base of the Pyritous Shales and in the lower part of the Ironstone Shales (‘Banded Shales’ unit, van Buchem et al., 1992, 1994). Generally, the mudstone is homogenized by bioturbation with pyritized burrow systems being the only recognizable trace fossils. As a result, the quartzose silt is mostly dispersed in the mudstone, but patches showing very fine grading and thin streaks occur sporadically. The fauna is generally poor, mainly consisting of protobranchs and thin shelled pectinids with less common small gastropods, bivalves, belemnites and ammonites, locally concentrated in little ‘nests’ (Sellwood, 1972; van Buchem, 1990).

The mudstone facies has been interpreted as the deepest depositional environment in the Redcar Mudstone Formation (van Buchem and McCave, 1989). Very little evidence for storm influence is found in this facies, which places it below the storm wave-base (> 80 m for an epicontinental basin of this size; van Buchem and McCave, 1989). However, the presence of slight bioturbation and the absence of laminated shales excludes the possibility of long-term anoxic conditions and makes the presence of a permanently stratified water column unlikely.

**Quartz Siltstone to Sandstone and Shelly Mudstone Facies.**—

In the Silurian, Pyritous and Ironstone Shales, the siltstones and sandstones occur as discrete layers or localized patches in

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**Fig. 1.**—Generalized geological map of North Yorkshire, indicating outcrop and inland borehole locations (Nettleton: Nettleton Bottom borehole).
LOWER AND MIDDLE LIASSIC DEPOSITIONAL SEQUENCES OF YORKSHIRE

FIG. 2A.—Robin Hood’s Bay—South section; Millers Nab and Wine Haven locations. Coastal sections of the lower Lias stage in Robin Hood’s Bay (N. Yorkshire, United Kingdom). See the lithostratigraphy and biostratigraphy chapter for comments. Location maps can be found in Knox et al. (1990) and in Hesselbo and Jenkyns (1995). Columns: A: Medium-scale depositional sequences, B: Ammonite zones and subzones, C: Bed numbers from Hesselbo and Jenkyns (1995), D: bed numbers from Tate and Blake (1876). Thicknesses are given in centimetres. The alternative position of the Upper Sinemurian zones and subzones following Page (1992) and Hesselbo and Jenkyns (1995) is indicated as follows: 1a,b,c for the base of the obtusum Zone and Subzones; 2a,b for the base of the oxynotum zone and subzones; 3a,b,c,d for the base of the raricostatum zone and subzones; 4 for the base of the jamesoni zone.
Fig. 2B.—Robin Hood’s Bay—North section; Bay Town and Dungeon Hole locations. Coastal sections of the lower Lias stage in Robin Hood’s Bay (N. Yorkshire, United Kingdom). See the lithostratigraphy and biostratigraphy chapter for comments. Location maps can be found in Knox et al. (1990) and in Hesselbo and Jenkyns (1995). Columns: A: Medium-scale depositional sequences, B: Ammonite zones and subzones, C: Bed numbers from Hesselbo and Jenkyns (1995), D: thickness of beds. The alternative position of the Upper Sinemurian zones and subzones following Page (1992) and Hesselbo and Jenkyns (1995) is indicated as follows: 1a,b,c for the base of the obtsuni Zone and SubZones; 2a,b for the base of the oxynotum Zone and subzones; 3a,b,c,d for the base of the raricostatum Zone and subzones; 4 for the base of the jamesoni Zone.
Fig. 2C.—Robin Hood’s Bay—North section; North Cheek and Castle Chamber locations. Coastal sections of the lower Lias stage in Robin Hood’s Bay (N. Yorkshire, United Kingdom). See the lithostratigraphy and biostratigraphy chapter for comments. Location maps can be found in Knox et al. (1990) and in Hesselbo and Jenkyns (1995). Columns: A: Medium-scale depositional sequences, B: Ammonite zones and subzones, C: Bed numbers from Hesselbo and Jenkyns (1995), D: bed numbers from Tate and Blake (1876). Thicknesses are given in centimetres.
a background of argillaceous mudstone facies (Figs. 2, 3; Sellwood, 1970b; van Buchem, 1990). The discrete layers occur in several forms: (i) mm-thick silty streaks; (ii) cm-thick, laterally very continuous silt and sand layers showing parallel lamination, wavy lamination, hummocky cross stratification (HCS) and small wave-ripples; (iii) dm-thick sandy layers, continuous at the scale of the outcrop, generally homogenized by bioturbation (*Cruziana* ichnofacies: *Chondrites*, *Rhizocorallium*, *Thalassinoidea*, *Teichichnus*, horizontal *Ophiomorpha*, and occasionally *Diplocraterion*) and gutter casts (*sensu* Aigner and Futterer, 1978); and (iv) scour-fills, consisting of distinct plano-convex lenticular bodies (diameter up to 3 m) with an erosional base and containing a variety of small-scale cross stratification features (climbing ripples, wave ripples, low-angle cross bedding, etc.). The sandy layers are generally poor in fossil content; bivalves, belemnites and ammonites occur. Siltstone layers

**Fig. 3.**—Idealized medium-scale depositional sequences for the lower Liassic shallow-marine succession in Yorkshire. Typical sequences for a silt/sand-rich environment and for a silt/sand-poor environment are shown. In the latter case, the shell material constitutes the coarse fraction. The top of the sequence in both cases is characterised by a winnowed surface followed by a concentration of Fe-enriched deposits such as limonitised shells, glauconite, chamosite and ferruginous ooliths (see text for discussion). Ch = *Chondrites*, Th = *Thalassinoidea*, Rh = *Rhizocorallium*, Di = *Diplocraterion*; Cl = clay, Si = silt, Sa = sand.
Lateral variation in mineral composition is encountered within individual Fe-enriched units. Our observations show that the chamositic/limonitic oolitic facies are preferentially developed on the high and at the top of coarsening-upward sequences, whereas glauconite is preferentially developed in association with deeper-water facies in a more basinward setting. However, the local co-occurrence of chamosite and glauconite grains indicates that there may be some overlap in their environment of formation or perhaps that conditions have been locally favorable to post-depositional alteration of chamosite ooliths into glauconite peloids (cf. van Houten and Purucker, 1984; Odin, 1988).

There is general agreement that the Jurassic oolitic ironstones in Europe accumulated in shallow-water environments in the vicinity of low-lying, densely vegetated landmasses that underwent lateritic weathering during periods of reduced sediment influx (Bubenicek, 1961; Hallam, 1975; Gygi, 1981; Van Houten and Purucker, 1984; Young and Taylor, 1989). The characteristic occurrence of iron-oolitic deposits towards the top of minor and major regressive sequences has also been widely illustrated in the literature (Hallam and Bradshaw, 1979; van Houten and Purucker, 1984; Bayer et al., 1985; McGhee and Bayer, 1985; Young and Taylor, 1989; Burkalter, 1995). Debate continues, however, as to whether they were formed during maximum sea-level lowstand or during transgression in the early stages of the succeeding sea-level rise. They are perhaps best considered as marking the transition between the culmination of a shallowing phase and the onset of renewed deepening.

Correlation Surfaces

Lithological packets of strata can be defined by the correlation of two main types of horizons:

Erosional Surfaces, Non-Sequences, Condensed Surfaces.—

Two major breaks representing the duration of an ammonite zone have been found in the sequence (Figs. 4, 5). The first is identified by the absence of the bucklandi zone over the Market Weighton High but is represented in the basin by a substantial thickness (approximately 25 m) of mudstone capped by an Fe-oolitic and sandy facies. The second is represented by the absence of the margaritatus zone over the Market Weighton High and is again represented in the basin by a substantial thickness (approximately 30 m) of sandstone and ironstone (Staithes Sandstone and Cleveland Ironstone Formations). The considerable magnitude of these breaks may be assessed by calculating the average duration for a Jurassic ammonite zone, which is in the order of 1.5 my (ammonite zones following Cope et al., 1980; time scale from Harland et al., 1990).

Four minor breaks in the sedimentation are identified that represent hiatuses or periods of strongly reduced sedimentation and that are of the duration of an ammonite subzone (Figs. 4, 5). These are identified in (1) the angulata zone, which is partly missing in the Humber area on the Market Weighton High; (2) the turneri zone, which is poorly developed on the Market Weighton High and mostly represented by the ferruginous facies of the Frodingham Ironstone; (3) the densinodulum subzone (basal raricostatum zone), which over the Market Weighton High is represented by a hiatus or by only a thin Fe-oolitic
level over the Market Weighton High. In the basin, it is more fully represented, but still marked at the top by an Fe-oolitic level (composite coastal section); (4) the *masseanum* subzone (basal *ibex* zone) is missing on the Market Weighton High and strongly reduced in thickness (1 m) in the basin. The succeeding *valdani* subzone is very thin in the basin (2.50 m) and capped by a thin level of glauconite-rich and Fe-oolitic deposits that can be followed onto the high, where they are represented by the Pecten Ironstone.

**Flooding Surfaces.**

An abrupt change to deeper water facies, interpreted as a flooding surface, can be observed at five positions in the succession (Figs. 4, 5):

1. at the base of the Hettangian stage, where the shallow-water siliciclastic deposits of the Cotham Beds are overlain by the open-marine, shell-rich deposits of the *planorbis* and *liasi cus* zones, which are of almost uniform thickness across this area;
2. at the base of the *semicostatum* zone, where deeper water mudstones overlie the ferruginous-oolitic top of the *bucklandii* zone in the basin and the corresponding erosional surface on the Market Weighton High;
3. in the upper *obtusum* zone, where deeper water mudstones overlie silty and sandy deposits in the basin and the Frodingham Ironstone on the Market Weighton High;
4. in the upper *raricostatum* zone and lower *jamesoni* zone, where the upward change to black mudstones marks a major deepening both in the basin and on the Market Weighton High, and
5. in the lower *ibex* zone, where deeper water mudstones overlie the condensed section of the Pecton Ironstone on the high and time-equivalent condensed facies in the basin.

**DEPOSITIONAL SEQUENCES**

Three scales of depositional sequence have been distinguished in other regions (e.g., Einsele et al., 1991). The order of magnitude of the duration and thickness generally attributed to these different scales of sequences are: for large-scale depositional sequences, durations of more than 5 my and sediment thicknesses of hundreds to thousands of meters; for medium-scale sequences, durations of 0.5 to 5 my and sediment thicknesses in the order of hundreds to thousands of meters; and for small-scale sequences, durations of 20 to 400 ky (the Milankovitch Band) and sediment thicknesses in the order of decimeters to tens of meters. This classification is followed hereafter.

Figure 4 presents the sequence stratigraphic correlation scheme. It illustrates in a north-south cross section the thickness

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**Fig. 5.**—Correlation table based on ammonite biostratigraphy for the lower and middle Liassic stages of North Yorkshire. Six medium-scale and three large-scale depositional sequences are indicated. The relative sea-level curve is based on the detailed analysis of the different facies types. For literature references see text. Abbreviations: CALCAREOUS = Calcareous Shales, SILIC = Siliceous Shales, P = Pyritous Shales, L.I. = lower Ironstone Shales, U.I. = upper Ironstone Shales (all units within the Redcar Mudstone Formation); MARLSTONE = Marlstone Rock.
Lower and Middle Liassic Depositional Sequences of Yorkshire

and facies changes that take place in a transect from the axial part of the Cleveland Basin (coastal section), over the Market Weighton High (Acklam area and Humber area boreholes) and into the northern part of the relatively shallow Lincolnshire Basin (Netleton Bottom Borehole). The datum for the cross section is the raricostatum/jamesoni ammonite zonal boundary. The biostratigraphy follows the ammonite zonation of Cope et al. (1980), modified to take account of the updates by Phelps (1985) and Howard (1985). The lithostratigraphic nomenclature follows that proposed by Powell (1984). The Ironstone Shales have been subdivided into a lower and an upper part, whereby the name ‘Banded Shales’ (introduced by van Buchem and McCave, 1989) is an informal unit corresponding to the lower part of the Ironstone Shales.

Large-Scale, Second-Order Depositional Sequences

Three large-scale depositional sequences have been distinguished in the lower and middle Liassic strata of North Yorkshire that roughly correspond to the first three stages of the Lower Jurassic (Figs. 4, 5). The sequence boundaries are formed by two major erosional breaks in the succession at the Market Weighton High (absence of the bucklandi and margaritatus zones) and a phase of widespread sandy sedimentation both in the basin and on the high (raricostatum zone). An additional argument for the large-scale character of these sequences is the worldwide nature of their deepening phases, suggesting an eustatic origin (e.g., Hallam, 1978, 1981, 1988; Haq et al., 1987). The major deepening events occur at the base of the Hettangian, in the planorbis to liasicus zones, at the base of the Sinemurian, in the semicostatum zone, and at the base of the Pliensbachian strata, in the lower part of the jamesoni zone (Figs. 4, 5).

Some evidence for an organization at an even larger scale is suggested by the gamma-ray logs of the Felixkirk Borehole and the outcrop gamma-ray log of the composite coastal section. They show an overall increase in radioactivity through the Hettangian and Sinemurian strata, with a maximum at the base of the Pliensbachian followed by a gradual decrease of the gamma-ray values towards the top of the Pliensbachian section (Fig. 2). The highest gamma-ray values occur at the base of the Pliensbachian, where there is a distinct change in the clay mineral composition marked by an increase of the kaolinite fraction. Kaolinite values subsequently remain high throughout the janesoni, ibex and davoei zones. Based on this observation, which was supported by the AI/K trend, van Buchem et al. (1992) suggested an important climatic change as responsible for the change in mineralogy. However, as was pointed out by Hesselbo (pers. commun, 1995), the outcrop natural gamma-ray values, measured at the same section (van Buchem et al., 1992), show a decrease in the Th/K ratio at the Sinemurian/Pliensbachian boundary, which is contrary to what would be expected from an increase in K-poor kaolinite. A possible diagenetic influence on the clay mineralogy is unlikely (cf. van Buchem et al., 1992), and further study of the mineralogical phases containing K and Th are required to explain this paradox. The observed trend in overall radioactivity most probably relates to the proportion of uranium, which does show a good coherency with the overall increase and decrease of the total gamma-ray curve and is closely reflected in the variations in V and TOC (van Buchem et al., 1992).

Medium-Scale, Third-Order Depositional Sequences

Each large-scale depositional sequence can be further subdivided into medium-scale depositional sequences. This subdivision is based on the identification (in the basin and on the high) of a characteristic, recurring lithofacies rhythm, consisting of argillaceous mudstone overlain by muddy siltstone and sandstone or shelly mudstone, capped by Fe-rich deposits (Fig. 4). Based on the facies analysis, these medium-scale sequences have been interpreted as shallowing-upward cycles (Fig. 3). They were characterized by abrupt flooding events followed by continued deepening (often of the duration of one ammonite subzone), during which packages of mudstone (up to 20 m) were deposited over large areas, both in the basins and on the highs. This was followed by a reduction in sedimentation rate that led either to deposition of silty mudstone to fine sandstone facies, showing abundant evidence for storm-influenced sedimentary structures and winnowing (shell beds, sand layers with HCS, various kinds of scour), or, where there was low or minimal supply of silt and sand, it led to deposition of a very shell-rich mudstone facies (see Figure 3). Generally, at the top of the shallowing-upward trend, ferruginous mudstones are found that contain iron-ooliths, glauconite and phosphate pebbles. The interpretation of this last facies is ambiguous. It probably is a condensed record of several changes in the depositional environment during late highstand, lowstand and beginning of the next deepening and is thus essentially transitional in character. The typical development of thick iron-enriched deposits on the swells suggests that areas of shallow water, close to the low-lying land areas, were important in the generation of the ferruginous facies. However, the thin, time-equivalent iron-enriched deposits in the basin indicate that for limited periods, similar conditions existed in both the basin and on the highs.

Development of the iron-enriched deposits probably occurred at the very early stages of sea-level rise, when the flooding of low-lying land with lateritic soils caused iron-rich material to be carried into the relatively shallow but sediment-starved basins.

Sequence 1.—Sedimentation in the planorbis and liasicus biozones was characterized by the onset of deep-water mudstone sedimentation in all localities, with no significant thinning over the Market Weighton High (Fig. 4). These zones are not exposed in the field, but borehole analysis suggests a shallow marine environment with a shelly mudstone facies deposited between storm and fair weather wave-base (Fig. 3). Shallowing probably took place in pulses over the duration of the angulata zone. At Felixkirk, a shelly layer containing glauconite and phosphate occurs in the middle of the angulata zone. The top of the sequence is situated around the angulata/bucklandi zonal boundary, which is characterised by various types of iron-rich facies: a 5-m-thick interval of glauconitic shell beds, encrusted ammonites and fossil wood in the composite coastal section, limonitic shell beds at Felixkirk, and phosphate and ferruginous ooliths in the Acklam area. The angulata zone thins out over the Market Weighton High, and it is absent in the Cockle Pits Borehole in the Humber area.

The basal Hettangian flooding is a recognized phase of worldwide transgression and interpreted to be of eustatic origin (see references in Hallam, 1978, 1981, 1988). A striking feature of this depositional sequence is the contrast between the lateral
continuity of the facies in the planorbis and liasicus zones and the restricted occurrence of angulata zone deposits around the Market Weighton High. The non-preservation of the angulata zone in the Humber area is possibly related to a local uplift of the Market Weighton High towards the end of the angulata zone or in the beginning of the bucklandi zone. The widespread occurrence of the iron-enriched facies (in the basin and on the high) suggests that a more regional, possibly eustatic, component was also involved.

Sequence 2.—This sequence corresponds to the bucklandi zone. The most complete section is found in the coastal exposures at Redcar (Getty, 1972; Hemingway, 1974), where it reaches the considerable thickness of 30 m (Fig. 4). Here the basal deposits consist of several meters of winnowed glauconitic shell beds. The glauconite pellets have a maximum diameter of 300 μm and are slightly cracked; a feature that according to Odin (1988) indicates a break in sedimentation in the order of 10 to 100 ky. This basal facies is overlain by deeper marine mudstone facies with occasional shell beds. Towards the middle of the bucklandi zone, an interval with limonitic, reworked shell beds occurs (the ‘Cardinia Bed’ of Tate and Blake, 1876). The upper 10 m of the bucklandi zone show a gradual increase in quartz content (see also van Buchem et al., 1992). The boundary with the semicostatum zone is marked by two shell beds with ferruginous ooliths, overlain by 2 m of limonitic shell material and small phosphate pebbles in a mudstone matrix. At Felixkirk, the bucklandi zone is reduced to about 10 m and appears to be represented only by the roiforme subzone (Ivimey-Cook and Powell, 1991). In the Acklam area, the entire bucklandi zone appears to be missing, but the bucklandi zone sediments reappear on the southern flank of the Market Weighton High.

The sedimentological features found in the bucklandi zone at Redcar suggest a clear deepening in the lower part of the zone, and a distinct shallowing at its top. Since the bucklandi zone is entirely absent or strongly reduced over most of the Market Weighton High, a dominantly tectonic control of the shallowing trend is favored. Indeed, an eustatic sea-level fluctuation may have reinforced this effect, acting on the topographic relief inherited from late angulata zone to early bucklandi zone times.

Sequence 3.—The base of this sequence is characterized by condensed deposits with ferruginous ooliths in the composite coastal section and in the Acklam area (Figs. 4, 5). This was followed in mid and late semicostatum zone (scipionianum and sauceanum subzones) times by deeper water mudstone and fine siltstone deposition in all locations, except for the Worlaby Borehole in the Humber area. The thickness reaches up to 40 m in the Cleveland Basin sections and approximately 10 m in the Acklam area on the Market Weighton High.

Around the southern flank of the Market Weighton High, a prolonged period of widespread iron-rich deposition led to the formation of the Frodingham Ironstone, which ranges from latest semicostatum zone to mid obtusum zone (stellare subzones) in age. In the Humber area, the Frodingham Ironstone consists mainly of sideritic limestones containing layers rich in chamosite and limonite ooliths (Gaunt et al., 1980). Much of the shell debris is limonitized. In the Acklam area, the top of the semicostatum zone and all of the proved turneri zone also occur in condensed ferruginous facies not unlike that of the Frodingham Ironstone (Gaunt et al., 1980). Some evidence for the original iron silicate, now in the form of Fe-rich chlorite, has also been found in the Felixkirk ooliths (van Buchem, 1990). More common here, however, are shelly, ferruginous and sandy beds that represent winnowed and reworked horizons. In the Cleveland Basin, the turneri and obtusum zones are less condensed than on the high, and some variations in the quartz silt/sand content occur. The facies becomes more and more silty towards the top of the semicostatum zone (see Hesselbo and Jenkyns, 1995 for the coastal section). The succeeding turneri and obtusum zones also show some variations in the quartz silt content, and in particular the ‘High Scar’ (bed no. 23 of Hesselbo and Jenkyns, 1995 and Fig. 2a) is a prominent feature. We have taken this as the top of our sequence 3, which, following the biostratigraphy in Cope et al. (1980), places it in the upper part of the obtusum zone. In the revised scheme of Page (1992), which, as explained earlier, is not followed here, the top of our sequence 3 would be placed at the top of the turneri zone.

Sequence 3 is thus characterized by an extensive flooding event in mid semicostatum zone times, which is believed to represent a major eustatic sea-level rise and has been recognized over the British Isles, in Europe and around the world (Donovan, 1979; Hallam, 1981, 1988). Through the recognition of clay-rich intervals, deeper water conditions have also been inferred for the top of the turneri zone and the base of the obtusum zone in the Cleveland Basin. Fluctuations in relative sea level were poorly registered on the Market Weighton High, where the time-equivalent deposits are very thin (less than 5 m for the two zones in the Howsley borehole, Fig. 4), and were mostly represented in the condensed iron-rich facies of the Frodingham Ironstone. A probable explanation for this difference in sedimentation pattern is topographic control, which was either actively created through tectonic reactivation of the Market Weighton High at that time or already existed, in which case the amplitude of the relative sea-level variation was probably so low that the high was hardly affected. Evidence for a depositional sequence in the turneri zone is not conclusive in the studied area, and comparison with other areas is needed.

Sequence 4.—In late obtusum zone and early oxynotum zone times, mudstone sedimentation was restored in the basin and on the high, except for the Humber area (Fig. 4). The thicknesses remained limited, and the mudstone was generally more silty than in previous deepening phases.

The upper part of the oxynotum zone and the raricostatum zone show a distinct shallowing trend similarly expressed in the basin and on the high. It is characterized by coarse silt and sand, which form beds of varying thickness, scours and streaks within the mudstone sequence. They were emplaced by storms and show evidence of long periods of winnowing. The base of the raricostatum zone (densinodulum subzone) is absent on the high in the Acklam and Humber areas. In the basin, deposition was continuous, but the top of the densinodulum subzone is marked by a condensed chamositic/oolitic bed. The densinodulum subzone condensation-level represents a somewhat changed situation compared to sequence 3, in that no time-equivalent iron-oolitic deposits of importance (such as the Frodingham Ironstone) have been recorded on the Market Weighton High. In the Acklam area, a thin layer with limonite ooliths has been identified, while at the corresponding level in the Felixkirk Borehole, a high concentration of iron-rich chlorite was
observed, together with some grains in a transitional stage of conversion to glauconite. In the coastal section, chamomite ooliths (concentric lamination, Fe-rich chlorite) have been preserved in a sideritic matrix, found at the top of a thick sand bed. The good preservation of the chamomite and the shape of the grains argue against transportation and indicate a localised mode of formation. This condensed level has been taken as the boundary between sequences 4 and 5. An additional support for placing it at this position is the closer spacing and more amalgamated nature of the sand beds below the oolitic level, reflecting the limited accommodation space. Above this level, sand beds are increasingly widely spaced (see the Siliceous Shales in Fig. 2a).

The late obtusum zone to early oxynotum zone transgression has been recorded in Lincolnshire (Sellwood, 1972; Richard-son, 1979; Bradshaw and Penny, 1982), in Scotland (Hallam, 1978), in northern France (Hallam, 1978) and in the Schwabian Alb (Brandt, 1985) and thus seems to have affected at least all of northwestern Europe. The late oxynotum zone to early raricostatum zone condensed sequence is also a northwest European feature (England: Sellwood, 1972; Donovan et al., 1979; Bradshaw and Penny, 1982; Europe: Hallam, 1978, 1988; Brandt, 1985). The substantial increase in coarse siliciclastics in the central and northern British basins may indicate a regional tectonic uplift of the hinterland at that time. Compared with the shallowing, highstand, part of the previous sequences, the high at the base of the mudstone facies occurred simultaneously in the basin and on the shelf. The abrupt change from sandy shoreface facies to hemipelagic mudstones in a hemipelagic setting below storm wave-base. The middle and upper parts of the jame-soni zone are characterized in all studied locations by a gradual change to a more silty sedimentation. The typical characteristics of this shallowing are the common occurrence of shell beds with a rich macrofauna (notably Pincta and belemnites) and the development of a distinct dm-scale bedding pattern that has been interpreted as Milankovitch cycles (van Buchem and McCave, 1989; van Buchem et al., 1994). This interval corresponds to the lower part of the Ironstone Shales, for which the informal name ‘Banded Shales’ was introduced by van Buchem and McCave (1989). The top of sequence 5 is positioned around the jame-soni/ibex zonal boundary, which is a condensed sequence and/or hiatus in all localities (Fig. 5).

The basal jamesoni zone represents a transgressive deepening phase over most of Great Britain (Sellwood, 1972; Donovan et al., 1979; Bradshaw and Penny, 1982), and has also been recognized in northwest Europe and other parts of the world (von Hillebrandt, 1971; Hallam, 1988), suggesting a eustatic cause. The late jamesoni zone to early ibex zone ‘non-sequence’ has been recognized in central and southern England (Donovan et al., 1979; Horton and Poole, 1983) but so far not in other European localities. It represents a break in sedimentation of considerable importance, in the order of 100 to 1000 ky based on glauconite maturity (cf. ‘highly evolved stage’ of Odin, 1988).

The late valdani zone to early luridum subzone represents a transgressive upward sequence with a pulsed character. This is best expressed in the coastal section, where several siltier and sandier intervals have been found (Fig. 2c) as well as...
a horizon with reworked sideritic nodules (‘snuff boxes’) and chamositic ferruginous ooliths (van Buchem, 1990; bed 21 of Hesselbo and Jenkyns, 1995, Fig. 2c). This bed might be correlatable with a variously glauconitic, phosphatic and ferruginous oolitic bed found in the same stratigraphic position on the Market Weighton High (upper *ibex* zone). Towards the top of the *maculatum* subzone, sedimentation becomes siltier and sandier, finally developing into the bioturbated, cross-bedded and laminated sandstones of the Staithes Sandstone Formation. The sediment source for the Staithes Sandstone is probably in the north or northwest, as suggested by Smithson (1934, 1938) based on the heavy mineral assemblage and, by Shalaby (1980), based on palaeocurrent directions. The influx of sands occurs in the Felixkirk Borehole at the top of the *ibex* zone, and in the coastal section and Acklam area in the middle of the *davoei* zone. In the Humber area, this part of the sequence is strongly reduced in thickness, partly due to erosion and partly to condensation. The upper part of the *davoei* zone and the *margaritatus* zone are characterized by medium- to coarse-grained sandstone (Staithes Sandstone Formation) and overlain by extensive ironstones in the north (Cleveland Ironstone Formation; see Howard, 1985). In the Acklam area, the Staithes Sandstone and Cleveland Ironstone are strongly reduced in thickness, whereas in the Humber area all that remains is in the sideritic, chamositic oolitic mudstones of the Marlstone Rock (Richardson, 1979). In the Nettleton Bottom Borehole, the sandy facies, possibly corresponding to the Staithes Sandstone, is again better developed, but the Fe-rich interval is here also reduced to the Marlstone Rock, which is sandy at the base and consists of sideritic chamositic mudstone at the top (Bradshaw and Penney, 1982).

The late *ibex* zone deepening is recorded in the basin and on the high and seems to represent a relative sea level variation of at least regional importance. The late *davoei* zone to early *margaritatus* zone regression has been recognized in northwest Europe and attributed to regional tectonic uplift of the hinterland (Hemingway, 1974; Hallam, 1988). The different timing of the arrival of the coarse grained siliciclastic sediment at the various locations in North Yorkshire, followed by the extensive, local erosion of the *margaritatus* zone on the Market Weighton High, are clearly in agreement with such an interpretation. The overlying Cleveland Ironstone Formation, together with the Grey Shale Member and Mulgrave Member (Jet Rock) of the Whitby Mudstone Formation, form the deepening-upward trend of the next 2nd-order depositional cycle.

**Small-Scale, Fourth- and Fifth-Order Depositional Sequences**—

High-frequency cyclicity has been observed in several sedimentary facies in the lower and middle Liassic deposits of North Yorkshire (Sellwood, 1972; Howard, 1985; van Buchem and McCave, 1989; van Buchem et al., 1994). In the Calcareous Shales, shell beds locally occur grouped together in packages. A medium scale cyclicity is evident and summarized in Figure 3. A regular pattern at a finer scale has, however, not been observed. The Silicious Shales are characterized by the alternation of dm- to m-scale silty mudstone intervals and fine sand layers. A stacking pattern of more amalgamated and more expanded small-scale cycles can be observed in the coastal section (Fig. 2a), which helped to define the boundary between sequences 4 and 5. A Fourier analysis carried out by Weedon (1987) failed to identify frequencies corresponding to the Milankovitch band in these deposits. A combination of high-energy erosion and deposition probably obscured the original frequencies of the driving mechanism.

The Pyritous Shales do not show good evidence for a regular sedimentation pattern at a high frequency. The only interruptions to the deposition of the black mudstones are represented by horizons of calcitic and sideritic concretions, often displaying *Chondrites* burrows in association with a concentration of shell material. These concretion horizons probably represent short pauses in sedimentation, allowing local carbonate precipitation. Clear evidence for high-frequency cyclicity was again found in the lower part of the Ironstone Shales (Fig. 2b; ‘Banded Shales’ unit of van Buchem and McCave, 1989). Here a decimeter-scale bedding pattern was observed, consisting of an alternation of silt-rich layers with abundant macrofossils and clay-rich layers with few fossils (Sellwood, 1972; van Buchem et al., 1994). An orbital control of the high-frequency cycles has been demonstrated by van Buchem et al. (1994), and comparison to other locations in England shows that very similar types of cyclicity were registered in very different environments over a very similar time interval (Horton and Poole, 1983; Weedon and Jenkyns, 1990; van Buchem et al., 1995). The upper part of the Ironstone Shales displays in the Cleveland Basin a meter-scale to decameter-scale alternation of sandy/silty intervals and mudstone. The Cleveland Ironstone Formation also shows small-scale cycles, which are topped by ferruginous-oolitic deposits (e.g., Howard, 1985).

**DISCUSSION AND CONCLUSIONS**

In the shallow-marine epicontinental setting of the lower Lias and middle Lias succession of North Yorkshire, three large-scale, 2nd-order depositional sequences have been defined, which correspond roughly to the first three stages of the Lower Jurassic Series. The importance of the *semicostatum* zone flooding, which is poorly expressed in part of the Humber area, was pointed out to the authors by Hesselbo (pers. commun., 1995). The first sequence (broadly equivalent to the Hettangian Stage) has an estimated duration of 6 my (four ammonite zones; time estimation based on a Jurassic ammonite zone duration of 1.5 Ma based on Cope et al., 1980 and Harland et al., 1990) with a maximum thickness of 62 m in the basin and a minimum thickness of 21.5 m on the high. The second sequence (equivalent to most of the Sinemurian Stage) has an estimated duration of 6.75 my (four and a half ammonite biozones) with a maximum thickness of 74 m in the basin and a minimum thickness of 28.5 m on the high. The third sequence (broadly equivalent to the Pliensbachian Stage) has an estimated duration of 6 my (four ammonite biozones) with a maximum thickness in the basin of 127 m and a minimum thickness on the high of 27 m. Two important conclusions follow from these general observations. Firstly, it is clear with such low overall sedimentation rates that the sequences must be very incomplete, hence the importance of the Fe-enriched facies in this type of siliciclastic epicontinental basin. Secondly, the Cleveland Basin experienced almost a doubling of the sedimentation rate in Pliensbachian times, as compared with the Hettangian and Sinemurian, whereas the sedimentation rate over the Market Weighton High remained almost the same (see also Table 1).
The cross-section correlating successions on the high to those in the basin (Fig. 4) permits us to evaluate the relative influence of the possible local tectonic control on the sedimentation pattern. In *planorbis* zone and *liasicus* zone times (early Hettangian time), a transgressive package of open-marine mudstones of almost uniform thickness was deposited across the area, suggesting an almost flat topography. In *angulata* zone and *bucklandi* zone times, this situation changed dramatically, with substantial breaks in sedimentation occurring over the Market Weighton High. The *angulata* zone is absent in the Cockle Pits Borehole on the southern flank of the high, while the succeeding *bucklandi* zone is missing in Howsham on the northern flank. The creation of topography and the shift of areas of non-deposition and erosion strongly suggest tectonic activity at that time in the area of the Market Weighton High. The anomalously thick *bucklandi* zone in the coastal section (Table 1) may, seen in this light, partly represent the transport of erosional products from the uplifted Market Weighton High (or other uplifted marginal areas). An interference of the tectonically influenced depositional patterns with medium-scale variations of eustatic origin is likely in this part of the succession. Detailed comparison with time-equivalent strata deposited in areas of minimal tectonic activity is required to evaluate the significance of the eustatic component in this part of the succession.

During the major transgression in *semicostatum* zone times, mudstone deposition was restored both in the basin and on the high, except in the Worlaby area, where a topographic relief was probably maintained. In the late Sinemurian time, the sedimentation pattern was determined by the overall lack of accommodation space, both on the high and in the basin. One medium-scale rise in eustatic sea level (*oxynotum* zone) of at least northwest European extent, created enough accommodation space for sediment to accumulate on the Market Weighton High. In general terms, facies are condensed in the upper Sinemurian strata (Frodsmith Ironstone, Siliceous Shales). There is no specific evidence for local tectonic activity at this time, although some topographic relief appears to have been maintained between the slightly more rapidly subsiding Cleveland Basin and the Market Weighton High. The sedimentation pattern in the basin and on the high becomes increasingly uniform towards the top of the Sinemurian deposits (Siliceous Shales).

The Pliensbachian Stage marks a distinct geographical differentiation in the creation of accommodation space. In the Cleveland Basin, the inferred sedimentation rate doubled, as compared with the Sinemurian and Hettangian time; whereas on the Market Weighton High, the sedimentation rate stayed roughly the same (see Table 1). The overall creation of accommodation space during the *jamesoni* and *ibex* zones was, however, such that a similar sedimentation pattern could be maintained on the high and in the basin (Fig. 4). The *jamesoni* zone flooding was certainly of eustatic origin, and nothing in the studied area argues against a similar origin of the *luridum* subzone deepening. At the end of the Pliensbachian Age, tectonic control in the form of uplift in the hinterland has been invoked as causing the extensive supply of sands (Hemingway, 1974; Hallam, 1988). Erosion and/or non-deposition occurred on the Market Weighton High, where the entire *margaritatus* zone is missing, probably due to active uplift at that time.

Our observations on tectonic control are in accordance with the interpretation of the presence of a granite body underneath the Market Weighton High by Bott et al. (1978). They have suggested that the buoyancy of the relatively low-density granite was responsible for the differential uplift. The uplift was restricted to certain specific periods, occurring through movements on basement faults (Bott et al., 1978).

The typical facies sequences, as defined in Figure 3, may represent a general pattern for the infill of this type of shallow, siliciclastic epicontinental basin. Characteristic are the distinct vertical facies succession and strong lateral continuity of the sedimentary facies. The iron-rich facies may play a key role in these basins, allowing an accurate correlation of basin and high successions. Furthermore, detailed geochemical studies of this type of iron-rich facies may be carried out with more success within a well-defined sequence stratigraphic framework.

To summarize, the sedimentation pattern in the lower Lias and middle Lias stages was influenced at different scales by eustatic, tectonic and climatic factors. The transgressions of the 2nd-order depositional sequences (early Hettangian, early Sinemurian, early Pliensbachian times) are most probably of eustatic origin. The medium-scale, 3rd-order sequences probably reflect both local tectonic control (Market Weighton High) and eustatic variations in sea level. Climatic influence may have played a role at the large scale, by creating the conditions for extensive deposition of iron-enriched sediment in the marine environment and at the small scale by influencing sediment distribution within the shallow epicontinental basin by (orbitally-induced) high-frequency variations in climatic conditions. Our main conclusions are:

1. The lower and middle Liassic strata in North Yorkshire (Redcar Mudstone, Staithes Sandstone and Cleveland Ironstone Formations) are characterized by three large-scale, 2nd-order sequences with durations in the order of 6 my, six medium-scale, 3rd-order depositional sequences with durations varying from 1 to 3 my, and small-scale, 4th- and 5th-order cycles occurring in several facies.
2. The medium-scale depositional sequences show a typical organization of lithofacies in a shallowing-upward trend, from argillaceous mudstone, to quartz siltstone and sandstone or shelly mudstone into Fe-enriched deposits (limonite, glauconite, chamosite, ferruginous ooliths).
3. Both eustatic and tectonic control were found to have influenced the sedimentation pattern. The transgressive phases in the large-scale, 2nd-order depositional sequences are of eustatic, tectonic and climatic control. The transgressive packages of open-marine mudstones of almost uniform thickness were deposited across the area, suggesting an almost flat topography. The shift of areas of non-deposition and erosion strongly suggest tectonic activity at that time in the area of the Market Weighton High.

### Table 1—Duration, thicknesses and sedimentation rate of the medium-scale depositional sequences.

<table>
<thead>
<tr>
<th>Sequence</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ammonite biozones</td>
<td>3</td>
<td>1</td>
<td>2.5</td>
<td>1.75</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>Estimated duration (my)</td>
<td>4.5</td>
<td>1.5</td>
<td>3.75</td>
<td>2.6</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>Thickness in the basin (m)</td>
<td>32</td>
<td>30</td>
<td>50</td>
<td>24</td>
<td>64</td>
<td>63</td>
</tr>
<tr>
<td>Thickness on the high (m)</td>
<td>20</td>
<td>1.5</td>
<td>21.5</td>
<td>7</td>
<td>17</td>
<td>10</td>
</tr>
<tr>
<td>Basin avg. sed. rate (m/m)</td>
<td>7.1</td>
<td>20</td>
<td>133</td>
<td>9.2</td>
<td>21.3</td>
<td>21</td>
</tr>
<tr>
<td>High avg. sed. rate (m/m)</td>
<td>4.4</td>
<td>1</td>
<td>5.7</td>
<td>2.7</td>
<td>5.6</td>
<td>3.3</td>
</tr>
<tr>
<td>Ratio basin/high</td>
<td>1.6</td>
<td>20</td>
<td>2.3</td>
<td>3.4</td>
<td>3.8</td>
<td>6.3</td>
</tr>
</tbody>
</table>

*Thickmesses in the basin are taken from the composite coastal section, thicknesses on the high are taken from the Humber area (see Fig. 4). The estimated duration is based on the average duration of a Jurassic ammonite biozone of 1.5 my (Cope et al., 1980; Harland et al., 1990). Average sedimentation rate is calculated based on the total time represented in the sequence (including hiatuses).*
tatic origin (early Hettangian, early Sinemurian and early Pliensbachian times). Long-term tectonic control was particularly evident during the Pliensbachian Age, when the Cleveland Basin subsided rapidly relative to the then stable Market Weighton High. At the scale of the medium-scale, 3rd-order sequences, eustatic control is inferred in the *oxy-notum* and possibly late *ibex* zones. Tectonic activity at this scale was found to influence sedimentation at various degrees on and around the Market Weighton High through local uplift during *angulata*, *bucklandii* and *margaritatus* zone times. In the long term, climatic conditions set the scene for the deposition of iron-enriched deposits, and, at a higher frequency (4th- and 5th-order), caused variations in sediment flux and transportation within the epicontinental basin.

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ABSTRACT: Biostratigraphically well-calibrated exposures of Lower Jurassic rocks in the Wessex, Bristol Channel, Cleveland and Hebrides basins have been remeasured and interpreted in the context of sequence stratigraphy. The aim has been to see whether the stratigraphy and facies are compatible with the hypothesis that relative sea-level changes were synchronous across all these basins. The Lower Jurassic Series can be subdivided into four large-scale (so-called 2nd-order) lithologic cycles, with durations of approximately 3–10 my, that appear to be synchronously developed in all onshore British basins; the cyclic changes in facies become more extreme as the cycles young. Candidate maximum flooding surfaces in the large-scale cycles, identified on the basis of distal starvation, or facies successions indicative of maximal accommodation space in proximal areas, occur in the lower semicostatum zone (Lower Sinemurian), obtusum–oxynotum zones (upper Sinemurian), lower jamesoni zone (lower Pliensbachian) and falciferum–bifrons zones (lower Toarcian). Candidate sequence boundaries in the large-scale cycles, defined on the basis of major unconformities or facies successions indicative of minimal accommodation space in proximal areas, are recognized in the upper turneri zone (mid-Sinemurian), mid-vanicostatum zone (upper Sinemurian), basal margaritatus zone (mid-Pliensbachian) and levissuei zone (upper Toarcian).

In general, at the large scale, the Lower Jurassic Series of the Dorset area of the Wessex Basin shows the most distal pattern of sediment accumulation, in which condensed sections (limestone or mudrock) correspond to relative sea-level rise or highstand and expanded sections (mudrock or sandstone) correspond to relative sea-level fall or lowstand. In contrast, the Lower Jurassic Series of the Sky, Pabay and Raasay areas of the Hebrides Basin exemplify the proximal pattern of sedimentation in which expanded sections (sandstone and mudstone) correspond to relative sea-level rise or highstand, and condensed sections (sandstone) correspond to relative sea-level fall or lowstand. The Yorkshire coast successions of the Cleveland Basin exemplify an intermediate setting. Significant divergence from this pattern is evident in the Toarcian (and through the Middle Jurassic) deposits over which interval the style of accumulation in the Hebrides is intermediate between that of the Wessex Basin and that of the Cleveland Basin. This indicates a reduction of clastic supply or increase in creation of proximal accommodation space in the Hebrides area relative to Yorkshire that began in the early Toarcian.

Lithologic cyclicity at the scale of ammonite zones and subzones (so-called 3rd-order) is recognized in a variety of facies; durations are inferred to be approximately 0.5 to 3 my. In a manner that contrasts with the large-scale cycles, the medium-scale cycles become more weakly expressed upwards through Lower Jurassic successions. The link between medium-scale sedimentary cycles and relative sea-level change is more interpretative than is the case for the large-scale cycles. There are few surfaces that have a definitive expression in all basins considered here; those that do are: candidate maximum flooding surfaces in the lira and taylori subzones, and at the stokesi-subnodosus subzonal boundary (all major); and candidate sequence boundaries in the mid-jamesoni zone (moderate), and at the base of the stokesi subzone (major). Similarly, there are few surfaces that appear strongly localized, the best examples being candidate sequence boundaries in the subnodosus and gibbosus subzones, which are developed mainly in the south and north respectively. In hemipelagic-dominated, distal facies, there is evidence to suggest that stratigraphic condensation is a consequence of relative sea-level fall rather than rise; relative sea-level rises in these settings appear to have generated erosion surfaces. A new relative sea-level curve is presented with medium- and large-scale cycles shown that are compatible with all the successions considered in this study.

INTRODUCTION

This study utilizes exposures of the marine Lower Jurassic successions of the British Isles to examine the proposition that synchronous depositional sequences, of distinct orders of frequency and magnitude, can be recognized in different basins across the U.K. We intend this work to form a frame of reference against which sequence stratigraphic interpretations from other parts of the world may be compared. Not only can the British successions be correlated with those in other basins in north-west Europe at the level of an ammonite subzone (Dean et al., 1961; Copeland et al., 1989; Cox, 1990), but they may also be compared on a global basis using Sr-isotope stratigraphy with almost equal precision (Jones et al., 1994). It is a central aim of this study, therefore, to root a Lower Jurassic sequence stratigraphic scheme in a careful assessment and documentation of the onshore U.K. exposures.

The development of lithologic cycles within the British Jurassic System, and their relationships to sea-level change has been a subject of much discussion (see Arkell, 1933; Duff et al., 1967; Einsele et al., 1991 for general reviews; see Hallam, 1961, 1964, 1981, 1988; Haq et al., 1988, for the interpretation of lithologic cycles). The successions exposed on the coasts of Dorset and Yorkshire played an important role in the calibration of the cycle chart of Haq et al. (1988). Differences between the Hallam (1988) and Haq et al. (1988) curves stem largely from the fact that the same surfaces in the Dorset and Yorkshire sections were given different interpretations regarding relative sea-level change, and because Hallam (1988) did not consider global sea-level change to have been established unless indicative phenomena could be traced widely across the world. A further difference between these curves arises because Hallam (1988) did not explicitly recognize an intricate hierarchy in the cyclicity.

Different levels in the hierarchy of sedimentary cyclicity (Haq et al., 1988; Vail et al., 1991) have been assigned to orders, defined on the basis of their perceived durations; those of concern to us here are: (1) large-scale cycles, with durations approximating to stages and thought to represent some 10 my (so-called 2nd-order, or transgressive-regressive facies cycles) and (2) medium-scale cycles, approximating to ammonite zones or subzones and representing approximately 0.5 to 3 my (so-called 3rd-order or sequence cycles). In the present study, we discuss separately the large-scale and medium-scale cycles. Lithologic cycles also occur at a smaller scales (so-called 4th order, etc.), but their occurrence is sporadic in the British Lower Jurassic successions, and regional correlation is not possible with the biostratigraphic resolution presently attainable; hence their origin is not discussed. It should not, of course, be assumed that every lithologic cycle at the same scale was produced by the same mechanism.

Recent years have seen the development of sophisticated models describing the architecture of depositional sequences in response to relative sea-level change (Posamentier et al., 1988;
All major British Lower Jurassic sections have been remeasured (Figs. 1–9), including, from south to north, the coastal exposures of Dorset (Wessex Basin), Somerset and Glamorgan (Bristol Channel Basin), Yorkshire (Cleveland Basin) and western Scotland (Hebrides Basin). In areas where lithostratigraphic and biostratigraphic controls were relatively poor, further collections of ammonites were made in the course of the present study, and emphasis was placed on obtaining as detailed and complete a lithostratigraphic succession as possible; this has applied particularly to the Sinemurian strata of Robin Hood’s Bay, Yorkshire (Hesselbo and Jenkyns, 1995; Fig. 3) and most of the Lower Jurassic strata of western Scotland (Hesselbo et al., 1998; Figs. 5–7).

The Lower Jurassic sedimentary rocks considered here were deposited in epicontinental extensional basins with broadly similar tectonic histories and patterns of sediment accumulation (Whittaker, 1985; Chadwick, 1986; Lake and Kamer, 1987; Penn et al., 1987; Milsom and Rawson, 1989; Ziegler, 1990; Jenkyns and Senior, 1991; Bradshaw et al., 1992; Morton, 1992; Hesselbo and Jenkyns, 1995). The strata were probably all laid down in fully marine environments, which are predominantly siliciclastic and are typically, though not exclusively, fine-grained. In all the British onshore basins, the Lower Jurassic rocks are the first major marine accumulations following a pe-
relatively low accumulation rates and pass upwards into lower Pliensbachian mudrocks that had relatively high accumulation rates. The same pattern is evident in upper Pliensbachian through to the lower Toarcian deposits. Therefore, in both cycles, an increase in accommodation space is required to permit shale accumulation. Our measure of accumulation rate is zonal and subzonal thickness: biostratigraphic subdivisions are taken to be of approximately constant time value, an assumption that is, of course, open to question. Any bias related to unequal duration of zones or subzones would probably increase the number of subdivisions recognized in the mudrocks, because preservation is generally better in this lithology, and a finer biostratigraphic precision would be facilitated. Correction for this potential bias would serve to amplify the apparent rate of relative sea-level rise. It is also possible that the mudrocks were deposited in shallow water than the sandstones, but given the very shallow aspect of most of the sandstones, the Staithes Sandstone for example (Rawson et al., 1982; Howard, 1985), such an interpretation has little merit. Furthermore, because of the significantly greater compaction that must have taken place in the mudrocks compared to the sandstones, the different relative thicknesses must originally have been considerably greater. (Differential compaction may also relate to the greater ease with which coarse sediment may build into shallow water compared to fine sediment).

Secondly, unconformities on the basin margins correlate with the late Pliensbachian and late Toarcian phases of sand deposition in the basin. In the case of late Pliensbachian deposition, Howard (1985) and Copestake and Johnson (1989) have illustrated how most of the Cleveland Ironstone and the entire Staithes Sandstone are missing across the Market Weighton High to the south of the Cleveland Basin. The Toarcian/Aalenian unconformity (Black, 1934) is well displayed in the south of Robin Hood's Bay, where its relationship to the Peak Fault, a major synsedimentary structure (Milsom and Rawson, 1989), can be seen clearly. The stratigraphy on the eastern, upthrown, side of the Peak Fault is incomplete with Aalenian strata resting on mid-Toarcian. On the downthrow side, the Blea Wyke Sandstone (upper Toarcian) is preserved, but the stratigraphy is still incomplete as the uppermost *levesquei* Zone is missing (Knox, 1984). Thus, movement on the Peak Fault alone cannot explain the large-scale stratigraphic cyclicity, and sea-level fall of at least regional extent is implied.

**Dorset (Wessex Basin).**

An upward change occurs from relatively slow and continuous accumulation in Hettangian and Early Sinemurian time, through to rapid but highly episodic sedimentation in the mid-Pliensbachian to Toarcian (and Middle Jurassic) time. This trend parallels the change from clays to sands as the main siliciclastic constituent (Fig. 4). Similar variations in lithology and depositional rate are superimposed at the scale of stages, with relatively continuous, gradual deposition characterizing the Hettangian to Lower Sinemurian and the lower Pliensbachian, in contrast to episodic deposition of asymmetric lithologic cycles which characterizes Upper Sinemurian, upper Pliensbachian, and upper Toarcian strata. These stage-scale cycles
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FIG. 3.—Summary sections for the Yorkshire coast Lower Jurassic (Cleveland Basin). See Figure 2 for key. All ammonite zones and subzones of the Lower Jurassic are present in the Yorkshire area except the uppermost, the aalenis subzone of the levelae subzone (Bairstow, 1969; Hemingway, 1974; Getty, 1980; Howarth, 1980a, b). Stratigraphy and sedimentology for the Robin Hood’s Bay area of Yorkshire are discussed in Hesselbo and Jenkyns (1995). The Hettangian to mid-Pliensbachian strata are predominantly mudstone with subordinate sandy and Fe-rich intervals. The Redcar Mudstone (Powell, 1984) is subdivided into a number of informal units correlatable within the basin. Sandstone beds are developed in the Upper Sinemurian Siliceous Shales, whereas mudstone dominates both below, in the Lower Sinemurian Calcareous Shales, and above, in the lower Pliensbachian Pyritous Shales and Ironstone Shales. An organic-rich shale is developed in the Lower Sinemurian (semicostatum zone) Calcareous Shales and the lowermost Pliensbachian (jamesoni zone) Pyritous Shales. The ammonite zones of the Upper Sinemurian are condensed in comparison to those of lower Pliensbachian. The upper Pliensbachian is characterized by the occurrence of sandstone and ironstone: two distinct formations are recognized, the lower being almost entirely sandstone, deposited in a storm-dominated shallow-marine environment (the Staithes Sandstone) and the upper being a unit of mudstone, siltstone, sandstone and oolitic ironstone arranged in a 10-m-scale cycles (the Cleveland Ironstone) (Howard, 1985; Hesselbo and Jenkyns, 1995). Ammonite subzones of the Staithes Sandstone are similar in thickness to those of the underlying rocks, but those in the Cleveland Ironstone are somewhat condensed. The Toarcian comprises predominantly mudrock (the Whitby Mudstone; Powell, 1984; Rawson and Wright, 1995) with the uppermost part coarsening to fine sandstone (Blea Wyke Sandstone). Organic-rich facies occur mainly in the falciferum zone near the base of the succession, the Mulgrave Shale (formerly ‘Jet Rock’ sensu lato; Hallam, 1967, 1978; Morris, 1979; Myers and Wignall, 1987; Jenkyns, 1988; Rawson and Wright, 1995). The ammonite zones of the mudrocks are most expanded in the middle and condensed at the base and top.

broadly comprise the large-scale depositional sequences (Fig. 10), except in Upper Sinemurian deposits where patterns that are apparent in Yorkshire and the Hebrides are obscured in Dorset because of the incompleteness of the record.

The large-scale cycles can be interpreted in the context of relative sea-level change, although for much of the succession interpretation is somewhat difficult without also considering the other basins. The Hettangian and lowermost Sinemurian strata appear to have been deposited in relatively deep water during a relative sea-level rise with sediment starvation occurring in earliest Sinemurian time. Lower Sinemurian strata were probably then deposited during a relative sea-level hightstand and, perhaps, subsequent fall, which allowed mud to prograde into the basin, filling space created during the earlier rise. Consideration of the Yorkshire and Hebrides successions suggests that the gaps in the Upper Sinemurian may have been a consequence of deposition during two episodes of relative sea-level rise, when supply to the basin was greatly reduced.

A return to relatively slow and continuous sedimentation inferred from the lower Pliensbachian strata is compatible with
FIG. 4.—Summary sections for the Lower Jurassic Series of the Dorset coast (Wessex Basin). See Figure 2 for key. The Hettangian to mid-Pliensbachian of the Dorset area is predominantly mudrock, deposited in fully marine environments (Lang, 1924, 1936; Lang and Spath, 1926; Lang et al., 1923, 1928, Hesselbo and Jenkyns, 1995). The mudrock of the Hettangian and Lower Sinemurian section (the Blue Lias) is interbedded with limestone (Hallam, 1960, 1964; Weedon, 1985). Higher parts of the succession, the mid-Sinemurian to lower Pliensbachian (comprising the Shales-with-‘Beef’, Black Ven Marls and Belemnite Marls) also contain discrete and commonly thick intervals with significant volumes of carbonate, albeit more dispersed than in the Hettangian. The Belemnite Marls show a marked interbedding of organic-rich mudstone and carbonate-rich mudstone which is attributed to climatic control (i.e., Milankovitch cycles; Weedon and Jenkyns, 1990). Thick intervals of organic-rich, laminated black shales occur in the succession, particularly in the Lower to lower Upper Sinemurian (senticostatum, turneri and obtusum zones). Also important are biostratigraphic gaps in the Upper Sinemurian, upper obtusum to oxynotum and ramicostatum zones (the Coinstone and the Hummocky respectively; Lang, 1945; Hallam, 1969; Sellwood, 1972; Hesselbo and Palmer, 1992). Lang has documented the ammonite zones and subzones, tied to a detailed lithostratigraphic succession (Palmer, 1972b; Getty, 1980; Howarth, 1980a). Overall, there is an increase in thickness of zones from the Hettangian to mid-Sinemurian (planorbis to turneri zones), above which level the pattern is complicated by missing zones and subzones in the Upper Sinemurian, and condensation in the lower Pliensbachian. (Some authors continue to recognize the bucklandi subzone at the top of the bucklandi zone [e.g., Page, 1992], whereas other authors subsume this within the rotiforme Subzone [Ivimey-Cook and Donovan, 1983]; in this study we follow Ivimey-Cook and Donovan (1983), but indicate in the stratigraphic columns in parentheses the limits of the bucklandi subzone as previously recognized.) In the mid-Pliensbachian (Green Ammonite Beds) the carbonate content decreases and the silt content increases; the zonal thicknesses also increase upwards. The two ammonite zones of the upper Pliensbachian are vastly different in thickness with the lower, margaritatus zone, being very expanded, and the upper, spinatum zone, being very condensed (in the Junction Bed sensu lato). Stratigraphic gaps occur in the succession, but are relatively small. The bulk of the upper Pliensbachian succession undoubtedly represents a shallowing of the depositional environment, with the shallowest water deposition occurring probably in the upper subnodosus subzone of the margaritatus zone. The basal unit of the Junction Bed (sensu lato), the Marlstone, belongs partly to the lowermost Toarcian and is an argillaceous, crinoidal and Fe-oolitic grainstone (Howarth, 1980c). The bulk of the Toarcian is represented by the very condensed limestone forming the Junction Bed sensu stricto which, although containing many gaps locally, is stratigraphically complete at a zonal level (Jackson, 1922, 1926; Howarth, 1980b, c, 1992; Jenkyns and Senior, 1991). Above the Junction Bed, the Toarcian strata comprise a thick coarsening upward succession of silt to fine sand (Down Cliff Clay to Bridport Sands).
**Fig. 5.—** Summary section for the Hettangian–Sinemurian strata of Skye (northern Hebrides Basin). See Figure 2 for key. The succession comprises alternating sandstone and sandy mudstone, with minor pure mudstone, limestone and ironstone (Judd, 1878; Lee, 1920; Lee and Bailey, 1925; Richey and Thomas, 1930; Hudson, 1983; Hallam, 1992; Morton, 1990a, b; Hesselbo et al., 1998) and is relatively complete. Biostratigraphic assignments are based on Hallam (1959) and Oates (1976, 1978). Lithostratigraphy is as revised in Hesselbo et al. (1998). The Hettangian to mid-Pliensbachian is a heterogeneous succession of limestone, sandstone and mudstone with a general pattern of expansion of the ammonite zones from the Hettangian into the lower Pliensbachian and condensation into the top of the lower Pliensbachian, complicated by condensation in the mid-Sinemurian (obtusum and oxynotum zones). The Broadford Fm, of Hettangian and earliest Sinemurian age, comprises interbedded carbonate and siliciclastic rocks deposited in environments as shallow as beach (Searl, 1989, 1992). Ammonites are rare but demonstrate termination of shallow-water deposition at the end of the bucklandi zone. The lower Pabay Shale (Lower to mid-Sinemurian) comprises sandy mudstone and sandstone. The semicostatum and turneri zones are relatively expanded and sandstone of the turneri zone exhibits meter-scale, roughly tabular, cross-bedding. Condensation of the mid-Sinemurian strata coincides with the development of less turbulent-water environments, culminating in the deposition of dark micaceous shales in the oxynotum zone. The upper Pabay Shale (Upper Sinemurian and lower Pliensbachian) is predominantly sandy mudstone. The more sandy units within the Pabay Shale are defined as members (Hesselbo et al., 1998).

deposition in deeper water, which probably began in Sinemurian time because the erosion surface that characterizes the Sinemurian–Pliensbachian boundary (the Hummocky) is less marked than the erosion surface within the Upper Sinemurian (the Coinstone). Milankovitch cycles, present in the Hettangian (Blue Lias) and in the lower Pliensbachian (Belemnite Marls) deposits (Weedon, 1985; Weedon & Jenkyns, 1990), may also indicate deposition in deeper water, more hemipelagic environments, less prone to disruption by near-shore or near-surface processes.

Upper Pliensbachian strata can be interpreted simply as a consequence of progradation of sediments into deeper water during a late Pliensbachian relative sea-level highstand and, possibly, fall; the expansion of the upper Pliensbachian strata with respect to the underlying strata is interpreted here as a consequence of increasing grain-size (silts to fine sands) and supply rate, which permitted the sediment pile to build into shallower, more turbulent water.

Bathymetric interpretation of the Junction Bed (Fig. 4) has always been a matter of some discussion. For this bed, and the lithologically similar Middle Jurassic Inferior Oolite, Sellwood and Jenkyns (1975) favored deposition in water that was too shallow to allow significant accumulation of clay, but they were frank about a lack of diagnostic paleobathymetric criteria. However, limestones such as the Junction Bed clearly characterized areas of reduced subsidence within the Wessex Basin. Other areas within the basin with higher rates of subsidence (e.g., the Winterborne Kingston Trough; Rhys et al., 1982), accumulated a less condensed, muddier lower to mid-Toarcian succession. On the basis of correlation with widespread transgressive facies
FIG. 6.—Summary section for the Pliensbachian–Toarcian of Pabay and Raasay (northern Hebrides Basin). See Figure 2 for key. In the Pabay Shale, a thick homogeneous sandstone member (Suishnish Sandstone) is developed locally at the Sinemurian/Pliensbachian boundary, overlain by dark mudstone with siderite concretions which is an identical facies to the age-equivalent strata in Yorkshire (i.e., the Pyritous Shales). The upper Pliensbachian succession, the Scalpa Sandstone, comprises a very-fine-grained bioturbated sandstone or siltstone. The sandstones locally extend down into the lower Pliensbachian and up into the Toarcian. The Scalpa Sandstone section shown is as seen on Raasay, using the stratigraphy of Howarth (1956), Phelps (1985) and Hesselbo et al. (1998); sedimentology of the upper part of the Scalpa Sandstone has been discussed by Hallam (1967). The Toarcian strata contrast greatly with those underlying, in that they are relatively condensed. Condensation and missing section are particularly marked in the mid-Toarcian Raasay Ironstone which lies between two mudrock units, the organic-rich Portree Shales below, and the Dun Caan Shales above. The Portree Shales are of *falciferum*-zone age, and there is close correlation with the Mulgrave Shale (*H11505* 'Jet Rock') of Yorkshire in terms of age and lithology. These units are representative of widespread carbon-rich *falciferum*-zone-age shales that occur throughout Europe and elsewhere (Jenkyns, 1988). Ammonites within the Raasay Ironstone of Raasay indicate the *falciferum* zone (Howarth, 1992), but on Ardnamurchan (Richey and Thomas, 1930; Dean et al., 1961; Howarth, 1992) indicate additionally the *thouarsense* zone (?*striatulum* subzone): the whole is clearly condensed and incomplete. A further stratigraphic break above the ironstone is indicated by the ammonites in the Dun Caan Shales which are upper *levesquei* zone (*aulensis* subzone), although mostly developed as shale, the Skye sections, particularly those exposed in the Strathaird area, are very sandy.

elsewhere, Hallam (1975, p. 165) considered the Junction Bed to have formed at a time of rising sea-level: its condensed nature being partly a function of sediment starvation (cf. Talbot, 1973), and a conclusion with which we concur. The early Toarcian relative sea-level rise probably began in the late Pliensbachian time, judging from the lack of documented latest Pliensbachian (*spinatum* Zone) sand in the Wessex Basin.

Although the influx of silt and sand from the north in late Toarcian time has previously been interpreted as due to deepening (i.e. to the creation of accommodation space; Sellwood and Jenkyns, 1975), it now seems much more likely that it resulted from a late Toarcian relative sea-level highstand and fall that caused renewed (and latterly rapid) progradation of sands from north to south. It was pointed out by Hallam (1978) that this juxtaposition of fine-grained facies over condensed limestone in the Toarcian (and the Bajocian-Bathonian) strata was local to the Dorset and Normandy areas, and he thus argued for local tectonic override of a eustatic regressive trend. However, if the Junction Bed is viewed as a phenomenon indicative of starvation in distal areas, albeit best developed on intra-basinal highs, the late Toarcian silt to sand (Down Cliff Clay and Bridport Sands) succession is not anomalous within the wider context. Relative sea level may have begun to rise again in the latest Toarcian because a lower Aalenian expanded equivalent
FIG. 7.—Summary section for the Lias of Morvern (southern Hebrides Basin). See Figure 2 for key. The shallow-water Broadford Fm of the northern Hebrides area is not well developed in this part of the basin. Here mudstone/limestone interbeds (Blue Lias) are instead predominant in the Hettangian and Lower Sinemurian (Oates, 1978; Hesselbo et al., 1998). These are overlain, by a coarsening-up succession of sandy mudstone and muddy sandstone of mid-Sinemurian age assigned to the Pabay Shale. The ammonite succession is relatively complete, except at the junction between the Blue Lias and the Pabay Shale where the only subzonal representative of the semicostatum zone is the lyra subzone, occurring locally in a thin pyritized crinoidal limestone at the level marked by an asterisk.

FIG. 8.—Summary section for the Lias of the north Somerset coast (Bristol Channel Basin). Locations of measured sections for individual segments indicated. See Figure 2 for key. Although the Somerset section is one of the most expanded Hettangian onshore successions in the British area, exposure does not permit examination of strata younger than the semicostatum zone. The Lower Jurassic rocks overlie uppermost Triassic strata of the Penarth Group which were deposited in marine, or marginal marine environments subject to subaerial exposure and desiccation. By convention, Cope et al. (1980) take the first appearance of the ammonite Psiloceras planorbis as defining the base of the Jurassic System. The summary section is a composite based on our own work, and integrates data published previously (Palmer, 1972a; Whittaker and Green, 1983; Page, 1992). At the top of the section we observe that the succession comprises interbedded light and dark marls decreasing in thickness upwards. The sedimentary facies are closely comparable with those of the Blue Lias and Shales-with-'Beef' of Dorset with which they correlate. The successions do, however, differ in that the lyra subzone of the semicostatum zone is not condensed in Somerset, where it may be as much as 90 m thick (Page 1992).

Western Scotland (Hebrides Basin).—

Broadly, the large-scale lithologic cycles recognized in the Lias of the Hebrides (Figs. 5, 6) are similar to those of Yorkshire and Dorset (Figs. 3, 4). The overall trend from Hettangian into lowermost Sinemurian strata is one of deepening with shallower, marginal facies (Broadford Fm), passing upwards into quiet-water, open-marine facies (Pabay Shale Fm). It is also clear that the middle Lower Sinemurian rocks are relatively regressive, comprising in part coarse siliciclastic sediment deposited in shallow, highly energetic marine or paralic environments (the Hallaig Sandstone Member of Hesselbo et al., 1998; Fig. 10). Unfortunately, in what are probably the most proximal areas, Raasay and Skye, the mid-Sinemurian stratigraphic relationships are uncertain and, beyond the indication that sandstone of the turneri zone is replaced upwards by sandy mudstone of the obtusum zone and shale of the oxytum zone, we can determine little about the completeness of that transition. Nevertheless, in the Skye area, the evidence does seem to point
to a progressive and sustained deepening from the uppermost *turneri* zone to the *oxynotum* zone. In the more distal regions (e.g., Morvern, Figure 6) the *turneri*-zone to *obtusum*-zone succession coarsens steadily upwards and a subtle fining is evident between the *obtusum* and the *oxynotum* zones, but the existence of an *oxynotum*-zone deepening in the southern area of the Hebrides Basin, particularly on Mull, is far from proven (see e.g., Hesselbo et al., 1998). In contrast to Morton (1989, 1990b), we find no evidence to suggest that the limited *semicostatum*-zone non-sequence between the Blue Liass and Pabba Shale in Morvern is part of a diachronous unconformity, supposedly continuous with an erosion surface at the base of the *obtusum* zone in Skye.

The uppermost Sinemurian succession (*raricostatum* zone) in the proximal areas is, like the mid-Sinemurian, also sandy (the Suishnish Sandstone Member of Hesselbo et al., 1998; Fig. 10), although neither the sediment grade nor the sedimentary structures achieve the dimensions observed in the *turneri* zone (Hallaig Sandstone), suggesting that the former were deposited in a greater depth of water. It is also notable that the *raricostatum* zone is relatively expanded (Fig. 6), which we attribute to the infilling of accommodation space created from the *obtusum* to the *oxynotum* zones and an increasing rate of creation of accommodation space in the time equivalent to the uppermost *raricostatum* zone.

The lower Pliensbachian (*jamesoni* zone) section of Pabay is markedly expanded; indeed, extrapolating from the increased thicknesses of *raricostatum*-zone sandstone on Skye compared to Pabay (Figs. 5, 6), it is likely that in any single vertical succession the strata are more expanded than the underlying *raricostatum* zone (no such section has been discovered at outcrop). The Pliensbachian section of the Pabay Shale is also less sandy than the Sinemurian portion. We take the earliest Pliensbachian to be the time of maximum rate of creation of accommodation space in the large-scale cycle. The mid- to upper Pliensbachian section shows a return to deposition of shallow-marine sandstone. Sedimentary thicknesses are most reduced in the mid-Pliensbachian, around the *davoei-margaritatus* zonal boundary, which we interpret as representing the level of minimum accommodation space in the large-scale cycle. The expanded upper Pliensbachian section is interpreted as reflecting renewed relative sea-level rise.

The *davoei-margaritatus* zonal boundary (as we have argued for the partly age-equivalent Junction Bed of Dorset), but the top surface may also have been winnowed during relative sea-level fall in latest Toarcian time prior to the resumed abundant supply of siliciclastic sediment. The uppermost Toarcian section clearly represents an overall shallowing phase into the Middle Jurassic sandstone above.

**Somerset and South Wales (Bristol Channel Basin).—**

Although the facies in Somerset (Fig. 8) are very similar to those in Dorset, the notable expansion of the *lyra* subzone suggests that sediment supply was sufficiently great to fill all available accommodation space and, hence, that relative sea level showed a progressive and accelerating rise through Hettangian and earliest Sinemurian time.

In the Glamorgan area of South Wales (Fig. 9), the progressive onlap and upward expansion of ammonite-zone thicknesses are compatible with deepening through Hettangian and earliest Sinemurian time. However, an alternative explanation for the onlap, that it is generated by increasing sediment supply driven by relative sea-level fall in the source area, cannot be ruled out on the basis of data presently available. It is relevant to this debate that the ‘marginal’ facies of South Wales have for the most part been interpreted as littoral with the outcrop...
relations thought to indicate progressive inundation of the Carboniferous basement, during Early Jurassic transgression (e.g., Trueman 1922), but interpretation of these deposits is somewhat controversial (Ager, 1986; Hodges, 1986; Fletcher et al., 1986; Fletcher, 1988; Wilson et al., 1990). Fletcher (1988) has detailed the nature of the unconformity surface around Ogmore and the relationships of the basal conglomerates and breccias to that surface. The surface morphology resembles modern tropical shorelines, and the orientations of bored pebbles, cobbles and boulders in the rocks immediately overlying the unconformity, led Fletcher (1988) to suggest a model of tidal-nip formation and subsequent cliff collapse. However, beds higher in the succession at Ogmore, as Ager (1986) pointed out, are matrix-supported conglomerates and may indeed be the products of debris flow. Wobber (1965) suggested a number of depositional processes, ranging from cliff-collapse and wave-action on a beach to density currents, slumps and slides in deeper water. The strong evidence for mass-flow depositional processes lends more weight to the deeper-water hypothesis. Thus the onlap observed may not be limited by shallow-water reworking.

**Comparison of Large-Scale Sedimentary Cycles in the British Area**

**Relative Sea-Level Change and Stratigraphic Response.**—

We have argued for the existence and ages of the large-scale cycles independently for each basin, but we also believe that the large-scale (2nd-order) relative sea-level cycles, which were the driving mechanism behind the formation of the lithologic cycles, were largely synchronous in all British basins (Fig. 10). Four large-scale lithologic cycles are recognized, with durations of approximately 3–10 my. Maximum flooding surfaces, identified on the basis of distal starvation, or facies successions indicative of maximal accommodation space in proximal areas occur in the *lyra* subzone of the *semicostatum* zone (Lower Sinemurian), at the *obtusum–oxynotum* zonal boundary (Upper Sinemurian), in the *taylori* subzone of the *jamesoni* zone (lower Pliensbachian), and within the *falciferum* zone (lower Toarcian). Sequence boundaries, defined on the basis of major unconformities or facies successions indicative of minimal accommodation space in proximal areas, are recognized in the *birchi*
subzone of the *turneri* zone (mid-Sinemurian), the mid-*rari-
costaturn* zone (Upper Sinemurian), the *stokesi* subzone of the *margaritatus* zone (mid-Pliensbachian), and the mid-*levesquei* zone (upper Toarcian).

There is a striking reciprocal relationship in stratigraphic thickness observed in a comparison of the Pliensbachian and Toarcian between Yorkshire and Dorset: thick sections in Yorkshire correspond to condensed sections in Dorset and *vice versa* (Hesselbo and Jenkyns, 1995; cf. Wilson, 1967). This relationship is also evident in a comparison of the Lower Jurassic Series of all the British onshore basins (Fig. 10).

We suggest that those parts of basins that were underfilled with respect to creation of marine accommodation space accumulated sediments preferentially when relative sea-level was falling (i.e., when sediment was forced basinward by lack of proximal accommodation space). Within the Dorset area of the Wessex Basin, thicknesses of strata related to long-term relative sea-level falls are significantly greater than those deposited during long-term relative sea-level rise (Fig. 10).

In contrast, areas of basins receiving large quantities of sediment in comparison with available accommodation space will only be able to accumulate large amounts of sediment during relative sea-level rise, so long as sediment is able to bypass the basin when it cannot accumulate in the local marine environment. This case is exemplified by the Hebrides Basin in the Skye–Raasay area, where thicknesses deposited during 2nd-order sea-level cycles, synchronous with Dorset, show proportionally much greater net accumulation during rises than is the case in Dorset (Fig. 10), particularly for the uppermost Sinemurian–lowermost Pliensbachian and uppermost Pliensbachian sections. In the Hebrides Basin, the Toarcian section is relatively condensed, and the pattern of accumulation is in many respects intermediate between that seen in the Wessex Basin and that seen in the Cleveland Basin, suggesting a relative reduction in the supply of siliciclastic sediment to the Hebrides Basin at this time or an increase in rate of creation of accommodation space closer to sediment source.

The Cleveland Basin (Yorkshire) represents the case in which approximately equal thicknesses of strata accumulated during Early Jurassic long-term rises and the falls (Fig. 10). From early Toarcian and through Middle Jurassic time, the situation changed such that the Cleveland Basin was the most proximal of the three main basins considered here, since most of the Middle Jurassic Series is predominantly non-marine and the succession appears to be punctuated by an unconformity spanning most of the Bathonian stage (Hogg, 1993). Within any one basin, some areas would have been oversupplied with sediment and some undersupplied, and a good example of this is shown by a comparison of northern with southern successions in the Hebrides Basin (e.g., Skye *versus* Morvern, Figs. 5, 7).

Taking this model for development of the British Lower Jurassic stratigraphy further, we can attempt to explain the thickness differences and timing of sandstone formations within each large-scale cycle with the caveat that if we divide the relative sea-level cycles into early and late rise and early and late fall we are approaching a level of analysis close to that of the medium-scale cycles discussed below. In underfilled settings the late stage of relative sea-level fall appears to have resulted in an expanded succession as sediment was forced basinwards into water depths that placed no restrictions on vertical accumulation; a good example is the mid-Pliensbachian of Dorset. In the same setting the early and late rise in sea level is characterized by fining and/or condensation as exemplified also by the Dorset section in the late Pliensbachian/earliest Toarcian (the Junction Bed *sensu lato*). It should be noted, however, that the depth of water which limits accumulation may be greater for clay than for sand. Hence, even though in the Dorset area the Wessex Basin was generally underfilled for sand in Early Jurassic time, it appears not to have been underfilled for clay, which was unable to accumulate substantially through the mid-Sinemurian.

In contrasting the Dorset succession with that of the Hebrides once again, a good case can be made that the mid-Pliensbachian relative sea-level fall created condensed sand and granule facies across much of the Hebrides Basin, as the water became too shallow for significant accumulation to occur and considerable reworking and winnowing took place. During the subsequent early rise, further accommodation space was created which was completely filled with sediment. In the early Toarcian late stage of rise and highstand, only clay was available to accumulate. Similarly, the uppermost Sinemurian sandstones in the Skye area, which occupy the same early rise position in the preceding large-scale cycle as the upper Pliensbachian sandstones, are similar in facies, stratigraphically expanded and overlain rather abruptly by deeper water clays (Fig. 6).

In summary, the Wessex Basin occupied a distal position for much of the Early Jurassic Epoch and large amounts of sediment accumulated at times when accommodation space was not available in more proximal settings such as the Cleveland or the Hebrides basins. Most differences in the sedimentary fills of these basins can be adequately explained by their positions on proximal to distal gradients and, in contrast to Hallam (1984), we see no reason to postulate independent and contrasting patterns of uplift and subsidence.

Where our analysis contrasts with recent studies on neighbouring basins, mainly in the North Sea, discrepancies appear either to be a consequence of different interpretations of the same successions, or else they cannot be be assessed because insufficient evidence has been presented to justify the previous interpretations.

In a study of the Lower Jurassic North Viking Graben, Parkinson and Hines (1995) subdivide the succession into four ‘genetic stratigraphic sequences’ bounded by maximum flooding surfaces of late Sinemurian, late Pliensbachian and early Toarcian age. The Lower Jurassic successions of the North Sea are not well calibrated biostratigraphically, and Parkinson and Hines (1995) placed interpretations on the hiatus-levels in Dorset, different from those made herein, which were used as ‘keys’ to understanding the North Sea stratigraphy. Parkinson and Hines (1995) considered that the Coinstone erosion surface (Fig. 10) formed as a consequence of lack of accommodation space, whereas the Hummocky erosion surface (Fig. 10) formed due to sediment starvation. As discussed in the next section, we interpret both surfaces as having formed by transgression-related sediment starvation, immediately following a prolonged period of regression.

Underhill and Partington (1993) reviewed the Jurassic North Sea and adjacent areas and stage-scale cycles are clearly indicated in the figures of Partington et al. (1993) and Underhill and Partington (1994). These exhibit a striking similarity to the cycles described herein. Within the limitations of the offshore
biostratigraphic resolution, these cycles, which are labelled as ‘3rd-order’, appear to be synchronous from the base of the Sinemurian up to the mid-Pliensbachian with those described in the present study. However, the upper Pliensbachian has been subdivided into several ‘3rd-order’ sequences that bear little relationship to the large-scale cycles described in the present study, and are not easily reconciled with the medium-scale cycles described in the following section. In the absence of a detailed account of their interpretations, it is difficult to identify the reasons for this contrast, which is most marked at the base of the *margaritatus* zone, interpreted as a time of maximum flooding.

The ‘genetic stratigraphic sequences’ of Stephen et al. (1993), described for the Moray Firth Basin, offshore eastern Scotland, do appear to coincide with the large-scale cycles described herein; the setting was considerably more proximal in Early Jurassic time than was the case for any of the basins described in the present study, and only the Upper Sinemurian to upper Pliensbachian lithologic cycle is clearly recognizable. A maximum flooding surface is described from the late *rari-costatum*–lower *jamesoni* zones. The very marked early Toarcian deepening, so well manifested elsewhere, is not reported; this part of the Moray Firth succession is poorly dated, and it is probable that the succession does not extend up to the *fulciferum* zone and is instead truncated beneath the Bathonian strata. The ‘genetic stratigraphic sequences’ of Morton (1989, 1990b) are approximately of the same scale as the cycles discussed in this study, although for the Lower Jurassic strata his interpretation of the Hebridean sections differs considerably from ours, particularly in the Sinemurian interval.

**MEDIUM-SCALE CYCLES**

In the following sections, we identify, on a stage-by-stage basis, candidate sequence boundaries and maximum flooding surfaces within lithologic cycles at a medium scale, and assess to what extent these can be said to be synchronous across the U.K. area. The most parsimonious interpretation is illustrated in Figures 3–9 and Figure 11. We also discuss the occurrence of constituent sedimentary cycles at smaller scales in the relatively few instances in which they have been observed.

**Hettangian Interval**

The Hettangian rocks of southern Britain are characterized by a simple and widespread stratigraphic motif, well exemplified by the section at St Audrie’s Bay on the North Somerset coast (Figs. 1, 8). The successions comprise interbedded limestone and mudrock that is locally organic-rich (the Blue Lias; see e.g., Fig. 4). Limestone is most abundant in the basal (*planorbis* and top (*angulata*) zones, with the thickest and least calcareous mudstones occurring in the middle (*liassicus*) zone. This pattern is well developed over the whole southern British area, and has been noted or described by Donovan et al. (1979), Hallam (1981), Brandon et al. (1990), and many other workers. In Glamorgan, where the typical ‘offshore’ interbedded limestone–marl facies can be traced into ‘marginal’ facies, the evidence has been interpreted as indicating *liassicus*-zone deepening. There, south of the Dunraven Fault, the ‘marginal’ facies which possibly belongs to the *planorbis* zone (Tawney 1866; Hodges 1986) is overlain by shale of *liassicus* zone age (Hodges, 1986; Wilson et al., 1990). *Angulata*-zone ‘offshore’ facies are not observed overlying *liassicus*-zone ‘marginal’ facies, although it might be argued that there is no exposure where this transition would be expected. Hence a candidate maximum flooding surface is recognized in the mid-*liassicus* zone.

An alternative explanation for the stratigraphic relationships in the mid-Hettangian must also be considered; the relatively condensed and carbonate-rich interval of the Pre-*Planorbis* Beds and the *planorbis* zone are sediment starved and correspond approximately to maximum flooding, whereas the *liassicus*-zone shales are the result of an increase in supply of argillaceous sediment to the basin. In this case, onlap across the London Platform (Donovan et al., 1979; Horton et al., 1987) may be the distal expression of progradation, rather than a reflection of increase in relative sea level.

The mudstone of the *angulata* zone is highly calcareous and it has been suggested that it represents a shallower water facies than the underlying *liassicus* zone. Some support for this interpretation comes from the exploratory work of Smith (1989) who suggested that, on the basis of correlation of Milankovitch-scale cycles between Somerset and Dorset, a hiatus is present in the Dorset section in the middle of the *angulata* zone. Further evidence cited in support of a regression at that time, or at least a stillstand, is the cessation of onlap across the London Platform (Donovan et al., 1979). However, a contrasting interpretation, that the *angulata* zone represents a time of maximum flooding, is equally tenable because all of these phenomena can be accounted for by postulating siliciclastic sediment starvation. The evidence in support of an *angulata*-zone relative sea-level fall or stillstand in the U.K. area is weak.

In summary, two competing hypotheses for medium-scale relative sea-level cycles through the Hettangian interval may be entertained, based on the U.K. sections. The evidence for either position is at present inconclusive. We prefer an interpretation of mid-*liassicus* transgression and mid-*angulata* regression.

**Lower Sinemurian Interval**

In Dorset and in Somerset (Figs. 4, 8), the lower *rotiforme* subzone of the *bucklandi* zone is, like the *liassicus* zone, characterized by the occurrence of a relatively argillaceous and, apparently, expanded succession. The mid-*rotiforme* subzone is dominated by calcareous mudstone and limestone, whereas the upper *rotiforme* subzone shows a return to argillaceous deposition. In Glamorgan, a closely similar pattern is observed (Fig. 9) except that the lower *rotiforme* subzone is not markedly argillaceous, but the mid-*rotiforme* subzone is developed as a limestone. Inland in south Wales, in the St Fagan’s borehole (Waters and Lawrence, 1987), a sequence of oolitic and peloidal limestone occurs between the *angulata* and *semicostatum* zones and displays a thickening-up bedding pattern; this is correlated with the limestone on the coast. The sequence in the borehole is capped by a hardground, overlain by limestone–marl facies of the uppermost *bucklandi* or *semicostatum* zones. Although it may be argued that the sequence represents redeposition during relative sea-level rise (i.e., highstand shedding in the sense of Schlager, 1992) an interpretation as shoaling has been preferred by previous authors (Waters and Lawrence, 1987; Wilson et al., 1990). The evidence appears ambiguous, but as in the
FIG. 11.—Summary of sequence stratigraphic interpretations for the Lower Jurassic of the British area and proposed relative sea-level curve compatible with the sections discussed. An alternative curve for the Hettangian-earliest Sinemurian time is shown by the gray stroke.
case of the Hettangian, we prefer to interpret the more argillaceous intervals as representing maximum flooding.

It is now well established that the *semicostatum* zone, in particular the *scipionianum* subzone, was a time of major deepening on the basis of diverse facies changes in many parts of the world including the British area (Hallam 1981). In Dorset, the top few decimeters of the Blue Lias comprise condensed limestone and organic-rich mudstone of the *lyra* subzone. These are overlain by the Shales-with-‘Beef’ which comprises organic-rich mudstone of the *scipionianum* subzone and calcareous mudstone of the *resupinatum* subzone (Fig. 4). The boundary between the *lyra* subzone and the *scipionianum* subzone is a hiatal surface showing evidence of erosion (Hallam, 1960; Hesselbo and Jenkyns, 1995) and is the oldest of several such surfaces in the Hettangian–Pliensbachian mudrocks. Considered in isolation, condensation of the *lyra* subzone may be ascribed to either shallowing or sediment starvation.

The *semicostatum*-zone succession in the northern Hebrides gives an unambiguous indication of deepening followed upwards by shallowing, and may aid interpretation of the southern exposures, if one assumes synchronous relative sea-level changes. In the Skye and Raasay areas, the undoubtedly shallow-marine limestone, mudstone and sandstone of the Broadford Fm pass upwards into the mudstone and sandy mudstone of the lower Pabay Shale (sensu Hesselbo et al., 1998). No where is the transition well exposed, but this deepening occurs at about the boundary between the *rotiforme* subzone of the *bucklandi* zone and the *lyra* subzone of the *semicostatum* zone (Hallam, 1959; Oates, 1978; Searl, 1992; Hesselbo et al., 1998). The *lyra* subzone attains a substantial thickness and is more argillaceous (as opposed to sandy) in the lowermost part and in the uppermost part which extends into the *scipionianum* subzone (Figs. 5, 10). Thus, there are two important phases of deepening in the lower part of the *semicostatum* zone with candidate maximum flooding surfaces near the base of the *lyra* subzone and in the *scipionianum* subzone; candidate sequence boundaries occur in the mid-*lyra* subzone and near the base of the *resupinatum* subzone. The hiatal surface at the top of the Blue Lias in Dorset is thus interpreted as resulting from sediment starvation caused by relative sea-level rise, as is the similar hiatal surface separating the Blue Lias from the Pabay Shale in Morvern (Fig. 7).

Following this interval of starvation in more distal settings and relatively argillaceous deposition in more proximal settings, the sedimentation pattern for the upper *semicostatum* zone is everywhere interpretable as regressive. This is most clearly the case in the more proximal settings (e.g., on Raasay; Fig. 5) where mudstone of *scipionianum*-subzone age is overlain by fine-grained sandstone in the *resupinatum* zone. A similar transition may be seen in Cleveland on the foreshore at Redcar, Yorkshire (Tate and Blake, 1876), although there is no modern published description of either the lithologic succession or the ammonites. Certainly, the mudstone of the uppermost *semicostatum* zone of Robin Hood’s Bay, Yorkshire (Calcereous Shales) is sandy relative to the mudstone yielding *Euaggasiceras scipionianum* at Redcar. This trend cannot be explained by paleogeography; in the Early Jurassic Cleveland Basin, northern localities (e.g., Redcar) were more proximal to sourc elands than were southern localities (Hesselbo and Jenkyns, 1995). In the Wessex Basin, the *resupinatum* zone (Shales-with-‘Beef’) is characterized by relatively carbonate-rich mudstone (Lang et al., 1923; Hesselbo and Jenkyns, 1995); this is the earliest of several intervals in which carbonate-rich facies in Dorset correspond to relatively arenaceous facies in Yorkshire. The *turneri* zone is easiest to interpret in the Cleveland Basin. There, it is characterized by argillaceous deposition (Calcereous Shales), which begins at the base of the zone and terminates near the top through an influx of fine-grained sand (Siliceous Shales); the subzones are not well defined. One bed in the middle of this mudstone interval is unusual in containing ferruginous ooids (van Buchem and McCave 1989; Bed 13 of Hesselbo and Jenkyns, 1995). The mudstone is less silty, has fewer storm-scours than the overlying and underlying strata, and is most simply interpreted as representing deeper water. The equivalent stratum in Dorset is an organic-rich paper shale, in the Shales-with-‘Beef’, at the boundary between the *birchi* and *brooki* subzones, underlain and overlain by more marly mudstone. The cyclic facies arrangement is a repeat of that seen in the *semicostatum* zone and this is particularly strongly evident in the Dorset succession (Fig. 4). In the more distal parts of the Hebrides Basin (Morvern), the basal *turneri* zone is missing. However, sedimentation resumed with silty mudstone deposition (upper *turneri* zone, *birchi* subzone) above which a coarsening is evident.

In summary, the Lower Sinemurian strata of the U.K. area can be resolved into three deepening–shallowing cycles (Fig. 11). Two cycles are present in the *semicostatum* zone with candidate maximum flooding surfaces in the *lyra* and *scipionianum* subzones. One cycle is present in the *turneri* zone with maximum flooding probably in the mid-*brooki* subzone and a candidate sequence boundary in the mid-*birchi* subzone. In the most distal setting (i.e., Dorset), the basal cycle of the *semicostatum* zone falls within a highly condensed interval, but in the most proximal setting (i.e., Skye and Raasay), it is well expressed in open-marine mudstone–sandstone facies. By contrast, in the more proximal settings, the development of the upper, *turneri*-zone cycle is suppressed; this is compatible with its coincidence with shallowest water conditions within larger scale cycles. It should be noted that at no one locality can all four medium-scale lithologic cycles be recognized.

**Upper Sinemurian Interval**

Medium-scale cycles are well expressed in the Upper Sinemurian successions. In Dorset, these comprise alternations of organic-rich and calcareous mudstones (Black Ven Marls) as is the case in the Lower Sinemurian strata, except that erosion surfaces are more clearly developed (i.e., the Cointone and the Hummocky; Fig. 10). In Yorkshire, the cycles comprise alternations of mudstone with sandy mudstone in the Siliceous Shales, and in the Hebrides, mudstone alternates with sandstone in the Pabay Shale (Figs. 5, 10).

The Yorkshire succession is apparently biostratigraphically complete. A mudstone-dominated interval occurs at the *obtusum*–*oxynotum* zonal boundary and in the uppermost *raricosatum* zone, continuing into Pliensbachian strata. The intervening sections contain variable quantities of sand in medium to thick, bioturbated beds. Storm scours are common (Sellwood, 1972; van Buchem and McCave, 1989). Of the sandy intervals, the uppermost *turneri* zone–lowermost *obtusum* zone, and the
upper oxynotum zone–lower raricostatum zone contain the most sand. Thus, candidate sequence boundaries are identified in the upper turneri zone (birchi subzone) and mid-oxynotum zone (simpsoni–oxynotum subzonal boundary) in Yorkshire.

The Dorset Upper Sinemurian strata contain two major hiatal surfaces. The calcareous mudstone that forms the top of the turneri zone (birchi subzone) in Dorset is overlain by a thick organic-rich shale, the Obtusum Shale, belonging to the obtusum zone (obtusum and stellare subzones; Page, 1992). The base of the Obtusum Shale is a minor erosion surface (Hesselbo and Jenkyns, 1995) and the interval overlying the shale, belonging to the upper part of the stellare subzone, is a calcareous mudstone. The general pattern is thus the same as was the case for the underlying turneri-zone lithologic cycle, except for the occurrence of a basal hiatal surface and a reduced total thickness (Fig. 4). Paleobathymetric interpretation is obscure, but on the basis of similarity to the case of the organic-rich facies in the underlying Shales-with-‘Beef’, the Obtusum Shale may best be regarded as having been deposited during relative sea-level rise or a highstand. There is no strong expression of the Obtusum Shale event in the Cleveland Basin. Since this is the probable time of maximum regression in the large-scale lithologic cycle, insufficient accommodation space may have been available in Yorkshire to accumulate a mudstone facies.

In the Hebrides (Fig. 5), the obtusum and oxynotum zones are generally either poorly exposed or poorly dated, and a medium-scale signal cannot be resolved with confidence. Better data exist for the raricostatum zone (Oates, 1978; Getty, 1980) which comprises sandy mudstone and sandstone (Figs. 5, 10). Three sandy intervals are separated by two muddy intervals. The sandstone becomes thicker upwards at the expense of the mudstone. Based on ammonites reported from the northern Hebrides (Oates, 1978; Hesselbo et al., 1998), the mudstone intervals are at the densinodulum–raricostatoides subzonal boundary and in the lower aplanatum subzone.

In summary, although a very clear medium-scale lithologic cyclicity exists for the British Upper Sinemurian units, it is difficult to make comparisons between basins in which good chronostratigraphic control is combined with unambiguous depth-related facies changes. However, based on the Yorkshire succession, candidate sequence boundaries are recognized in the stellare Subzone and at the simpsoni–oxynotum subzonal boundary. Based on Dorset, a candidate maximum-flooding surface is recognized in the obtusum Subzone. In the Hebrides Basin, candidate maximum flooding surfaces are recognized at the densinodulum–raricostatoides subzonal boundary and in the lower aplanatum subzone; candidate sequence boundaries are recognized in the densinodulum subzone, at the raricostatoides–macdonnelli subzonal boundary, and in the aplanatum subzone. It is plausible to relate the hiatuses and erosion surfaces in the Dorset succession to deepenings in the obtusum subzone, the denotatus–simpsoni subzonal boundary (‘the Coinstone’) and at the aplanatum–taylori subzonal boundary (‘the Hummocky’) (Figs. 4, 10).

**Lower Pliensbachian Interval**

Lower Pliensbachian medium-scale lithologic cycles comprise mainly silty mudstone with minor sandstone, and the successions are relatively complete down to a subzonal level. In both Yorkshire and Dorset, decimeter-scale beds are typical of the jamesoni and ibex zones (van Buchem and McCave, 1989; Weeden and Jenkyns, 1990; Hesselbo and Jenkyns, 1995), and stratigraphic variations in thicknesses of these zones give a useful indication of relative sedimentation rates. In Yorkshire, the base of the lower Pliensbachian (taylori subzone; jamesoni zone) is characterized by the occurrence of an organic-rich facies, the Pyritous Shales. Above this level in the, jamesoni zone, the succession gradually becomes coarser, reaching a maximum in the valdani subzone of the ibex zone where sand is fairly abundant. This upward-coarsening corresponds to a reduction in sedimentation rate as inferred from the condensed ammonite zones and subzones, thinning upsection of the decimeter-scale beds, and increasing concentration of belemnites (Hesselbo and Jenkyns, 1995). Making the assumption that coarser sediment in this setting indicates increased proximity to source, the surface of maximum condensation is more likely a surface of sediment by-pass rather than starvation. Biostratigraphic expansion and an abrupt fining characterize the upper ibex zone (luridum subzone) in Yorkshire, which is overlain by a succession that becomes progressively coarser upwards through the davoei zone (Hesselbo and Jenkyns, 1995). The Lower Pliensbachian (Belemnite Marls) section of Dorset shows remarkable parallels with that of Yorkshire (Figs. 3, 4), particularly in the pattern of expansion and condensation through the jamesoni and lower ibex zones. One major difference is that in Dorset, there is no concomitant grain-size variation detectable by field observation. A second significant difference is that in the ibex zone of Dorset, in addition to the condensed horizon in the valdani subzone (the Belemnite Bed), a second condensed horizon occurs in the luridum subzone (the Belemnite Stone). Although the former correlates with an inferred shallowing in the Yorkshire area, the latter corresponds to an inferred deepening. Hence, condensation in the more distal Dorset area of the Wessex Basin may be related in one instance to shallowing and in the other to deepening. Condensation and erosion at the Sinemurian–Pliensbachian boundary in Dorset (the Hummocky) may also be related to deepening. The davoei zone in Dorset (the Green Ammonite Beds) becomes more silty upwards, although the trend is not as marked as it is in Yorkshire, as is consistent with the more distal setting of the Wessex Basin.

A reasonably well-defined succession through the jamesoni zone can be followed on the Isle of Pabay (Figs. 1, 6). The taylori subzone is a black mudstone with sideritic bands very similar to coeval facies in the Cleveland Basin. The mid-jamesoni zone is significantly coarser with shelly fine-grained sands occurring interbedded with mudstone. A minor fining is evident into the upper jamesoni zone but the ibex zone clearly continues the coarsening trend. Whether this reaches a maximum in the valdani subzone is at present unknown because the critical interval is not exposed in the Skye–Raasay–Pabay area. The significance and lateral continuity of the upper jamesoni zone muddy interval is uncertain, but it is not well expressed, if at all, in the Yorkshire or Dorset successions. The uppermost ibex- to davoei-zone succession of the Hebridean area shows progressive coarsening combined with condensation (Phelps, 1985).

In summary, there is a remarkably good correlation in the stratigraphic development of the Cleveland and Wessex basins
through early Pliensbachian time. Candidate maximum flooding surfaces are recognized in the taylori subzone of the jamesoni zone of all British basins, and in the luridum subzone of the ibex zone in the Cleveland and Wessex basins. Candidate sequence boundaries are identified in the valdani subzone of the ibex zone. Some evidence exists locally in the Hebrides Basin for a moderately expressed sequence boundary in the polymorphus subzone and a weakly expressed maximum flooding surface in the jamesoni subzone, both of the jamesoni zone.

Upper Pliensbachian Interval

The Upper Pliensbachian strata are largely arenaceous in the Wessex, Cleveland, and Hebrides basins. Medium-scale lithologic cycles are well expressed in all three basins as mudstone–sandstone alternations.

In Dorset, several cycles are observed in the margaritatus zone (Figs. 4, 11). Sand occurs at four levels within this biozone: (a) the ‘Three Tiers’ unit at the base is the culmination of a coarsening-upwards trend originating in the lower davoei zone; (b) an unnamed, thin, discontinuous bed above the concretionary Eyre Nodule Bed; (c) the Down Cliff Sands, overlaying the concretionary Day’s Shell Bed; and (d) the Thorncombe Sands. All, except the Thorncombe Sands, belong to the stokesi subzone, and the lithologic cycles that they delimit may be regarded as of higher frequency than the cycles described herein as ‘medium scale’. The gibbosus subzone comprises a thin, silty mudstone with no name. It has been argued elsewhere (Ensom, 1984; Hesselbo and Jenkyns, 1995) that the Eyre Nodule Bed and Day’s Shell Bed contain hiatus concretions. Their positions at the bases of sand units (which were probably emplaced by storm processes) suggests an origin due to increased wave action on the seafloor prior to the supply of sand sufficient for deposition to occur (cf. Plint, 1988; Hadley and Elliott, 1993). In contrast, the thin, pebbly sandy limestones that cap the Down Cliff Sands and Thorncombe Sands are more likely to owe their origin to winnowing which occurred as sand supply to the basin was reduced, possibly as a result of deepening in more proximal settings. The Margaritatus Stone, which is the best developed of these possible condensed beds, caps the Down Cliff Sands. It may represent an extended period of time equivalent to the upper stokesi subzone and the lower subnodosus subzone and, as is indicated in Figure 11, is a good candidate as a maximum flooding surface. If the sequence boundary is to be placed at an abrupt juxtaposition of sandstone over mudstone, then the bases of any of the sandstones within the stokesi subzone may be regarded as candidate sequence boundaries, and no one surface has greater merit than another, based on currently available data. A candidate sequence boundary may also be placed at the base of the Thorncombe Sands (in the mid?subnodosus subzone).

In Yorkshire, the Staithes Sandstone is the culmination of the coarsening trend observed from near the top of the ibex zone (Figs. 3, 10). More or less muddy units of sandstone occur within the formation, but the thickest and coarsest development of sand is in the lower part of the stokesi subzone. The succession then finites to a minimum in the Cleveland Ironstone around the stokesi–subnodosus subzonal boundary. As was the case in Dorset, a candidate maximum flooding surface can be placed at this level. The Cleveland Ironstone comprises a series of coarsening-up cycles capped by oolitic ironstones approximately equivalent to the subnodosus subzone, the gibbosus subzone and the spinatum zone (Fig. 3). Not all the oolitic ironstone beds sit directly on the coarsest members of each cycle. This is particularly true of the Pecten Seam at the base of the spinatum zone, which sits unconformably on top of silty clays of the underlying margaritatus zone (Chowns, 1968; Howarth, 1980a; Howard, 1985). Because there is a coarsening up to the top of the spinatum zone, the ironstone seams at the base of the zone may coincide with the maximum flooding surface (Fig. 3; for discussion see Macquaker and Taylor, 1996; Hesselbo, 1997).

The succession in the Hebrides Basin differs from that of Dorset and Yorkshire in that the smaller-scale lithologic cycles are not observed (Figs. 6, 10). The upper Pliensbachian section is represented by the Scalpa Sandstone. The stokesi subzone fines upward into the subnodosus subzone, which is developed entirely as a siltstone/silty limestone facies. An abrupt coarsening occurs into the base of the gibbosus subzone. A weakly argillaceous but very shell-rich interval also occurs in the lower spinatum zone (apyrenum subzone) on Raasay: Howarth (1956) characterized this unit at Rudha Na’Leac as an oolitic sandy limestone and Hallam (1967) has identified the ooids as chamosite (presumably berthierine). The abundant and diverse fauna of ammonites, bivalves and crinoids was taken by Hallam (1967) to indicate condensation. As was the case in both Dorset and Yorkshire, candidate maximum flooding surfaces may be placed at the stokesi–subnodosus subzonal boundary and near the base of the apyrenum subzone (spinatum zone). A notable difference between the Hebrides section and that of either Dorset or Yorkshire, is that the stokesi subzone is less coarse than the underlying upper subzones of the davoei zone. One possible explanation may be an erosional gap at the base of the stokesi subzone, as would befit the more proximal position of the Hebrides area at this time; indeed, at Carsaig Bay on the Island of Mull (Fig. 1), the upper subzones of the davoei zone are missing (Phelps, 1985).

In summary, candidate maximum flooding surfaces are recognized at the stokesi–subnodosus subzonal boundary (margaritatus zone), and near the base of the apyrenum subzone (spinatum zone), at all locations in this study. A candidate sequence boundary is recognized near the base of the stokesi subzone (margaritatus zone), also at all locations studied. On the basis of an abrupt increase in sand deposition, a sequence boundary may be inferred within in the subnodosus subzone in Dorset, although there is no clear evidence of this in the Hebrides. On the same basis, a candidate sequence boundary is recognized near the base of the gibbosus subzone in the Hebrides. In Yorkshire both the subnodosus and gibbosus subzones coarsen up, and it is possible to infer similarly a sequence boundary in both subzones. In Dorset, the silty mudstone comprising the gibbosus subzone is sandwiched between the apparently condensed horizon at the top of the Thorncombe Sands and the condensed Marlstone; thus, it may be interpreted as a the distal expression of a regressive package of sediment and hence compatible with the sequence stratigraphy inferred from other basins. In the case of the gibbosus subzone candidate sequence boundary, its stratigraphic expression becomes stronger towards the north; this is in striking contrast to the subnodosus subzone candidate sequence boundary which finds its clearest expression in the south. This is the best example of asynchro-
nous sequence development at a medium scale; however, the sequences are not diachronous, and the evidence is not sufficient to propose different relative sea-level histories in each basin.

Toarcian Interval

There is little definite that can be determined concerning medium-scale sequences within the lower Toarcian strata. The Dorset succession is strongly condensed and incomplete, and the succession in the Hebrides suffers from poor exposure. In Yorkshire, the lower Toarcian comprises shales, some extremely organic rich, whose sequence stratigraphic significance is somewhat obscure.

A sequence-stratigraphic interpretation of the Yorkshire succession has been made by Wignall (1991) and Wignall and Maynard (1993). It has been argued that a sequence boundary occurs at the top of the spinatum zone on the basis of an interpretation of the Marlsstone as an extensive shallow-water deposit, and because a hiatus at the base of the tenuicostatum zone, occurring over structural highs in southern and central England, is interpreted as being due to lack of accommodation space (Wignall 1991; Wignall and Maynard, 1993). Wignall and Maynard (1993) also argued for a sequence boundary at the base of the falciferum zone on the basis of a minor biostratigraphic gap (see also Wignall and Hallam (1991)). We do not regard the evidence as strong for either of these sequence boundaries in the British area. The facies succession from the spinatum zone to the tenuicostatum zone appears always to be representative of deepening (Hallam, 1967) and, as has been argued in detail for a Sinemurian example (Hesselbo and Palmer, 1992), the occurrence of biostratigraphic gaps in these open-marine, fine-grained facies cannot be ascribed safely to relative sea-level fall. Indeed, both hiatal surfaces in the lower Toarcian strata may be better interpreted as consequential upon rapid relative sea-level rise and sediment starvation. Furthermore, we have found no evidence that the top of the exaratum subzone in the Cleveland basin is abnormally belemniferous, an argument used by Wignall and Maynard (1993) to support interpretation of this level as a maximum flooding surface. Neither is there any support for condensation at this level from consideration of the subzonal thicknesses (Fig. 3).

Most of the British upper Toarcian, like the lower Toarcian, does not evince distinct medium-scale lithologic cycles. By far the most complete and best-exposed section is that in the Cleveland Basin, although the upper Toarcian section is localized to the downthrown side of the major synsedimentary Peak Fault where late Toarcian–early Aalenian erosion was less pronounced. Grain-size variations, repeating at the scale of ammonite subzones, are detectable in the thouarsense and levesquei zones (Whitby Mudstone and Blea Wyke Sandstone; Knox 1984). Each lithologic cycle brings in coarser sand, and an erosion surface occurs within the sandstone at the Toarcian–Aalenian boundary, at which level the uppermost subzone of the levesquei zone (aalensis subzone) appears to be missing (Knox, 1984). It is noteworthy that in both the Hebrides and Wessex basins the uppermost levesquei subzone shows stratigraphic expansion compatible with a candidate sequence boundary at this level, manifested by the forcing of increased volumes of sediment into distal settings.

SUMMARY AND CONCLUSIONS

The Lower Jurassic Series can be subdivided into four large-scale (‘2nd-order’) lithologic cycles, with durations of approximately 3–10 my that appear to be synchronously developed in all onshore U.K. basins; the cyclic changes in facies become more extreme as the cycles young. Maximum flooding surfaces in the large-scale cycles, identified on the basis of distal starvation or facies successions indicative of maximal accommodation space in proximal areas, occur in the lyra subzone of the semicostatum zone (Lower Sinemurian), at the obutsum–oxynotum zonal boundary (Upper Sinemurian), in the taylori subzone of the jamesoni zone (lower Pliensbachian), and within the falciferum zone (lower Toarcian). Sequence boundaries in the large-scale cycles, defined on the basis of major unconformities or facies successions indicative of minimal accommodation space in proximal areas, are recognized in the birchi subzone of the turneri zone (mid-Sinemurian), the mid-rari-costatum zone (Upper Sinemurian), the stokesi subzone of the margaritatus zone (mid-Pliensbachian), and the mid-levesquei zone (upper Toarcian).

In general the Lower Jurassic strata of the Wessex Basin show the most distal pattern of sediment accumulation for large-scale sequences, in which condensed sections (limestone or mudrock) correspond to relative sea-level rise or highstand and expanded sections (mudrock or sandstone) correspond to relative sea-level fall or lowstand. In contrast, the Lower Jurassic strata of the Skye, Pabay and Raasay areas of the Hebrides Basin exemplify the proximal pattern of sedimentation in which expanded sections (sandstone and mudstone) correspond to relative sea-level rise or highstand, and condensed sections (sandstone) correspond to relative sea-level fall or lowstand. The Yorkshire coast successions of the Cleveland Basin occur in an intermediate setting. Significant divergence from this pattern is evident in Toarcian deposits (and through the Middle Jurassic) over which interval the style of accumulation in the Hebrides is intermediate between that of the Wessex Basin and that of the Cleveland Basin. This indicates a reduction of clastic supply, or an increase in creation of proximal accommodation space, in the Hebrides area relative to Yorkshire that began in early Toarcian time.

Candidate sequence boundaries and maximum flooding surfaces defining medium-scale (‘3rd-order’) sequences fall into three distinct categories: (1) those surfaces that occur unambiguously in all basins analysed in this study; (2) those surfaces that occur in more than one basin analysed in this study, but whose existence in all basins cannot be demonstrated unambiguously; (3) those surfaces that show distinct and unambiguous geographic localization. In several cases, the interpretation of the surface in terms of either rising or falling relative sea level may be in question, despite its long-range correlatability. Additionally, it should be borne in mind that some undoubted global sea-level-related events, such as that which produced the Toarcian exaratum-subzone black shale (Jenkyns, 1988), are not expressed unambiguously in all basins in this study.

There are few surfaces that have a definite expression in all basins considered here. Those that do are as follows: candidate maximum-flooding surfaces in the lyra and taylori subzones, and at the stokesi–submodosus subzonal boundary (all major); and candidate sequence boundaries in the mid-jamesoni zone.
(moderate), and at the base of the stokesi subzone (major). Similarly, there are only a few surfaces that appear strongly localized, the best examples being candidate sequence boundaries in the subnodosus and gibbosus subzones, which are developed mainly in the south and north respectively.

Unlike the large-scale lithologic cycles, the medium-scale lithologic cycles cannot be linked definitively to relative sea-level change in preference to changes in sediment supply. However, all the cycles are compatible with relative sea-level change as a driving mechanism. Where good data exist, there is clear evidence in the most distal segments of large-scale cycles for a high degree of synchronicity of medium-scale cycles, particularly in Early Sinemurian and early Pliensbachian times. Asynchronous medium-scale cycles, as recognized on the basis of sharp-based sandstone units, appear to have developed in late Pliensbachian and, possibly, Late Sinemurian times, during the longer-term relative sea-level lows inferred from the large-scale cycles.

If linked to sea level, then condensation within medium-scale cycles in distal settings may be a consequence of relative sea-level fall, as well as rise. This is most persuasively the case for the ibex zone, where in the Dorset succession we interpret condensation of the Belemnite Bed to be related to lack of accommodation space (i.e., winnowing), whereas the similar Belemnite Stone is more likely related to relative sea-level rise (i.e., sediment starvation). Opposing interpretations to ours were placed on these horizons by Haq et al. (1988). Erosion surfaces in distal settings correlate commonly to thinning or deepening in the early Pliensbachian, and, possibly, Late Sinemurian times, during the longer-term relative sea-level lows inferred from the large-scale cycles.

A relative sea-level curve can be constructed, representing large- and medium-scale cycles, that is compatible with all the successions described in this study (Fig. 11). The new curve shows broad agreement with those previously published, which were based wholly or in part on the Dorset and Yorkshire sections (Hallam, 1988; Haq et al., 1988) but differs significantly in detail. The sea-level curve in this study is least certain for the Hettangian and earliest Sinemurian interval, and least detailed for the Toarcian. Large-scale relative sea-level cycles appear to have had the strongest influence on the stratigraphic architecture of the British area in the later Early Jurassic times. Medium-scale relative sea-level cycles appear to have the strongest influence on the stratigraphic architecture during the times of large-scale relative sea-level lows, probably because the apparent effects are most marked in the shallowest water facies. Hence, the structure of the relative sea-level curve is strongly influenced by the superposition of medium-scale cycles upon large-scale cycles; it is simply less elaborate when inferred from deeper water facies.

The new relative sea-level curve differs from proposed eustatic sea-level curves in a number of respects. Our curve is considerably more detailed: most of the large-scale ('2nd-order') cycles recognized in this study correspond broadly in scale and timing to '3rd-order' cycles of Haq et al. (1988) and the smallest scale of fluctuation shown by Hallam (1988). In the present study we propose about 20 medium-scale candidate sequence boundaries, in contrast to the 10 suggested by Haq et al. (1988). Principal differences in timing, apart from in the uncertain Hettangian interval, are in the Late Sinemurian where, in contrast to Hallam (1988), we recognize an important oxynotum-zone deepening, and in the early Pliensbachian, where in contrast to the ibex-zone deepening of Haq et al. (1988) we recognize deepening in the jamesoni zone. These differences stem largely from contrasting interpretations of the same stratigraphic horizons in distal facies.

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AMMONITE BIOSTRATIGRAPHIC CORRELATION AND EARLY JURASSIC SEQUENCE STRATIGRAPHY IN FRANCE: COMPARISONS WITH SOME U.K. SECTIONS

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ABSTRACT: The Ligurian cycle has a duration approximately equivalent to Early Jurassic times. This major transgressive/regressive facies cycle is characterized by several rifting events that are related to the development of the Ligurian Tethys. Its onset is marked by the Early Cimmerian unconformity dated as latest Norian. Its end corresponds to the Mid-Cimmerian unconformity, dated as Late Toarcian (Stille, 1924), which preceded the late Aalenian major regression linked with tectonic uplift known in many areas of northern Europe. One result of this uplift phase was the coeval rapid and major subsidence phase that allowed the accumulation of thick successions in the major half-grabens of the nascent Tethyan margin. The Ligurian cycle can be subdivided into three or four 2nd-order facies cycles, depending on the local tectonic development in the area. The whole cycle comprises 27 3rd-order depositional sequence cycles. These can be dated locally to the precision of an ammonite “horizon”. Our sections are located along a north-south transect, from the Hebrides basin in the northern U.K. to the Alpes Maritimes in southern France. The study areas are located in the southeastern France Basin (now involved in Subalpine folding), the Causses (southern Massif Central of France), the Quercy (eastern Aquitaine) and the Paris Basin. These are compared with the Wessex, Cleveland and Hebrides basin in the U.K. Documentation is provided by ammonite biostratigraphic studies and outcrop and subsurface data. Such data have been reinterpreted using sequence-stratigraphic methodologies. One of the key results is that the 3rd-order depositional sequences that are the building blocks of the longer term cycles are correlatable units the entire western European craton.

INTRODUCTION

The objective of this paper is to document the biostratigraphic data for the Early Jurassic (Liassic) depositional sequences belonging to the Ligurian cycle. This paper concerns mainly the French Liassic basins which are compared with those of the U.K. (Fig. 1). The Ligurian cycle is mostly of Liassic age. The major regression at the lower boundary of this cycle is dated as late Norian (Late Triassic) and has been named the early Cimmerian unconformity (Stille, 1924). It marks the onset rifting of the Ligurian Tethys (Lemoine and Graciansky, 1988). The major regression at the end of the cycle is dated as late Aalenian (Graphoceras concavum Zone). It is related to the mid-Cimmerian unconformity and to a major doming and erosion phase in the North Sea (Underhill and Partington, 1993) and uplift in many areas of northern Europe.

The Lower and Middle Jurassic stratigraphic pile is characterized by a particularly well-refined ammonite biostratigraphy in both the U.K. and France, in which countries most of the stratotypes have been defined. Therefore it seemed critical to us to calibrate our sequence stratigraphic chart in these areas where stratigraphic science was born in the last century. The idea was to compare stratigraphic successions located in various structural and paleoclimatic settings along a transect across the western European craton, from the Provence platform on the Mediterranean Sea to the Hebrides Basin in the northern U.K. (Fig. 1). Ammonite identification and age assignment in this paper come from: (1) previously published studies, (2) unpublished data accumulated during geologic mapping of the study areas and (3) recent field work carried out specifically for this program. The main study areas (Fig. 1) are sedimentary basins located either on nascent Tethyan margin or controlled by the Tethyan structural development. These show a gradation from carbonate platform to mixed siliciclastic-carbonate basins. The data come mainly from outcrop studies. One area, the Provence platform to the southern Dauphinois basin is documented along the transect between the cities of Nice and Digne (southeastern France basin; Graciansky et al., 1993). Another area is the Grands Causses basin, a major half-graben located on the southeastern border of the Massif Central at midway between the southeastern France basin at the east and the Aquitaine basin at the west (Meister, 1986 and 1989). Quercy, a medium-scale graben at the eastern border of the Aquitaine basin was studied in great detail by the Toulouse University (Rey et al., 1991). Another key study area is the Paris Basin, a complex intracratonic basin.
that was subjected to both Tethyan and Boreal in fluences, de-
pending on the time. Information comes from both subsurface
and outcrop. The subsurface data combine well log interpreta-
tion (T. Jacquin) and calibration by ammonites recovered in the
boreholes (mainly unpubl. data by R. Mouterde). The Paris Ba-
sin outcrops are located in Burgundy (Dommergues, 1979,
1987, 1993), Lorraine (Hanzo, 1980; Hanzo and Espitalié,
1994) and Normandy (Rioult, 1988). As a result the duration and characteristics of indi-
vidual Liassic 2nd-order transgressive/regressive cycles differ
according to location, in contrast with the relative uniformity
in Europe all were active. However, the rift development,
including the rate of crustal extension and the rate of subsi-
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ler, 1988). As a result the duration and characteristics of indi-
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the Rhaetian strata over long distances.

On the nascent Tethyan margin (Subalpine zones), one single
transgressive/regressive facies cycles constituting the major Ligurian cycle have been numbered TR 4, 5
and 6 in the Tethyan marginal areas and TR 4a, 4b, 5 and 6 in
more northern European areas (Graciansky et al., this volume).
Each 2nd-order cycle varies relatively in time and space as a
consequence of the localized structural evolution. Their time
duration and time equivalence to third order depositional cycles
have been summarized on Figure 2.

Late Norian to Rhaetian

The Onset of the Jurassic Transgression

During Liassic times, rift systems located in the Arctic, North
Atlantic, central and western Mediterranean and western and
central Europe all were active. However, the rift development,
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Fig. 2.—Time-scale correlation of third-order depositional sequences and second-order transgressive-regressive facies cycles in several European-type areas.
with the overlying marine deposits dated as Hettangian by their ammonites (*Psiloceras planorbis* beds), because they are generally characterized by shallow-marine faunas with a great abundance and diversity of species, including the classical bone beds and *Avicula*-bearing coquinas. Rhaetian beds are widespread at the base of the Liassic basins. In the Netherlands (Sleen Shale; Nederlandse Aardolie, 1980) and in the Germany Basin, the three Rhaetian depositional sequences can be documented with similar features to those of the Paris Basin.

**T5-3 Sequence**

The **T5-3** sequence is physically defined as being the oldest backstepping sequence following the Norian major regressive phase and associated unconformities. In the western Southern Alps (Gaetani et al., this volume) the **T5-3** sequence corresponds to both the Riva di Solto Shales and the lower Zu Limestone, which may reach 300 to 1500 m in thickness and which represent the earliest open marine deposits of various lithologies that overlie the Norian Hauptdolomit as a consequence of its tectonic collapse and drowning. These transgressive shales have been dated late Norian (Sevatian) which shows that the Jurassic transgression actually started during Late Norian time and followed a major distensional event.

Similar results are found from field work in the Briançonnais units of the Internal Alpine Zones. In various places located at several hundred kilometers along strike, the sharp contact between the Norian Hauptdolomit and the Rhaetian beds is considered as intra-Sevatian (Late Norian). The Hauptdolomit have been dated by molluscs (*Worthenia contubulata*) and dasycladales (*Griphoporella vesiculifera*) and foraminifers (*Involutina sp.*). The Rhaetian beds comprise typical pelecypods (*Rhaetavicularia contorta, Dimyopsis emerichi*; Baud and Mégard-Galli, 1975, 1982–1983).

In the Paris Basin, the **T5-3** sequence has been defined by a widespread blanket of fluvial and lagoonal deposits in the west (Argiles de Chalain and Grès de Chalain) to playa type sediments in the east (argiles barioles dolomitiques), both unconformably overlying Norian regressive continental formations (Goggin and Jacquin, this volume; Figs. 3, 4, 5, 6).

In the Germany Basin, the **T5-3** sequence corresponds to the Unterrhät Formation or Ko1 unit of Wolburg (1969) which is characterized by its brackish fauna indicative of playa-type environments in the Postera Schichten and similar to those of the
Paris Basin (Fig. 5). The Unternhät formation can be divided into two parts based on their stacking patterns analysis. The lower one, the Postera Schichten, is transgressive onto strata ranging from Ladinian to Norian in age. It contains brackish faunas related to the uppermost Norian strata. It is indicative of restricted playa-type environments similar to those of the Paris Basin at that time (Fig. 5). The maximum flooding surface is a well marked gamma-ray peak. The upper part, the Grenz Sandstein, is a sand-prone deltaic prograding unit interpreted corresponding to the highstand systems tract. This is contemporaneous with similar sands of the Paris Basin. The entire T5-3 sequence of the Germany basin onlaps landward over the early Cimmerian unconformity.

In the Western Netherlands basin, the T5-3 sequence corresponds exactly to the Upper Keuper Claystone member of the Keuper formation (Nederlandse, 1980; Fig. 5). The sedimentological and stratigraphic characteristics of the sequence are similar to those of the Germany and Paris Basins. In the northern North Sea, the T5-3 sequence corresponds again to the oldest backstepping sequence that forms the base of the Statfjord formation (Steel, 1993). Here, it cannot be age dated but is characterized by its clear transgressive pattern over the early Cimmerian unconformity.

It is important to note the relatively speculative character of any precise definition of depositional sequences and long-distance correlation for this time span as a consequence of: (1) the long duration (10 my) of the Norian, (2) the poor biostratigraphical resolution, (3) the scarcity of biostratigraphical data and (4) the frequent occurrence of hiatuses related to the Early Cimmerian tectonic phase in the Germany and Paris Basins and more northern areas. Nevertheless, the available dates in the internal Alpine Zones (Briàncônnais), in the Southern Alps (Lombardy) and in Germany suggest that the T5-3 sequence boundary is pre-Rhaetian and intra-Sevatian (Late Norian). The T5-3 sequence appears to be restricted to subsiding areas, onlapping onto the early Cimmerian unconformity (Gaetani et al., this volume; Goggin and Jacquin, this volume; Fig. 5).

Rh1 Sequence

Everywhere, from the Tethyan domain to the Germanic domain sensu lato and the North Sea Central Graben, the Rh1 sequence yields the first true Rhaetian marine faunas. In the Alpine area (Dumont, this volume; Figs. 7, 8), the sequence boundary Rh1 (Rh1 Sb) shows neither major erosional features nor evidence for long periods of subaerial exposure,
The presence of characteristic of the Alpine areas. In Lombardy, the sequence corresponds to the middle Zu Limestone Formation (Gaetani et al., this volume). Shallow-marine to fluvial sabkha type to mixed terrigenous/calcareous environments are typical of the Rh 1 transgressive systems tract (Rh 1 TST) consists of thin backstepping sandstones (Flaser Sandstein member) overlain by shallow-marine to fluvial prograding sandstones (Flaser Sandstein member) and fluvial to uviatile prograding sandstones (Flaser Sandstein member) and fluviatile to uviatile prograding sandstones (Flaser Sandstein member). In the Paris Basin, a prograding parasequence set comprises coastal uviatile prograding sandstones (Flaser Sandstein member) and fluviatile to uviatile prograding sandstones (Flaser Sandstein member) and fluviatile to uviatile prograding sandstones (Flaser Sandstein member) and fluviatile to uviatile prograding sandstones (Flaser Sandstein member). In the West Netherlands, Broad Fourteens and southern North Sea Central Graben, the Rh 1 and Rh 2 sequences form the Sleen Shale Formation. Compared to deposits related to T 5-3 sequence, the Sleen Shale Formation covers wide areas including older rocks above the early Cimmerian unconformity. In the Germany Basin, the Rh 1 sequence corresponds to the Mittelrha Formation (Wolburg, 1969; Fig. 5). The transgressive systems tract (Rh 1 TST) consists of thin backstepping sandstones (Ubergany) overstepping out the preceding sequence and older rocks, landward. The maximum flooding surface covers wide areas and is indicative of open marine conditions (Contorta Schichten). The Rh 1 highstand systems tract comprises coastal to fluviatile prograding sandstones (Flaser Sandstein member) whose areal extent is more limited than their equivalent in the Paris Basin.

In the Paris Basin, a prograding parasequence set comprises the Rh 1 highstand systems tract (Rh 1 HST), and the main depocenter lies between the Vittel and Metz faults, as a probable consequence of the reactivation of these faults and a shift of subsidence eastward (Figs. 3, 4, 5, 6; Goggin and Jacquin, this volume).

Fig. 4.—Chronostratigraphic diagram for the Lower Jurassic part of a transect across the Paris Basin. Same profile and same horizontal scale as on Figure 3.
In the Paris Basin, the sequence boundary Rh2 is an erosional unconformity, with erosion being pronounced in Alsace and in the western part of the basin (Fig. 3, 4, 5, 6). In the basin center, the sequence boundary (Sb) is defined as lying between prograding Rh1 HST and retrograding Rh2 TST. In the southeast (Forcelles area; Fig. 5), the relative sea-level fall produced an isolated sediment wedge characterized by aggradational sands. This may represent a forced regressive wedge (Goggin and Jacquin, this volume). The Rhaetian transgression continued northwards and westwards. It has reworked previously deposited shallow-marine sands belonging to the underlying highstand systems tract of sequence Rh1 and redeposited these in the more subsident areas of the basin as barrier sand bars or tidal channel sands (Goggin and Jacquin, this volume). The maximum flooding of sequence Rh2 is the most pronounced of the Rhaetic sequences. It is characterized by diversified accumulations of open-marine benthic faunas including mollusks (commonly Avicula) and brachiopods. The highstand systems tract shows a return to lagoonal conditions over the entire basin with a shaly deposit known as the Argiles de Levallois (Jacquot, 1855).

In the Germany Basin, the Rh1 sequence corresponds to the Oberrhaet Formation (Wolburg, 1969). Its maximum flooding surface is also the most marine of the Rhaetian strata. As in the Paris Basin Argiles de Levallois, the highstand systems tract is highly regressive and yields continental palynomorphs in the Triletes Schichten.

Rh2 stratigraphic features include: (1) a moderate downward shift of coastal onlap at the sequence boundary, (2) rapid and widespread flooding at the maximum flooding surface and, (3) pronounced regression of the highstand systems tract illustrated by distinct progradation. These features are common characteristics of sequence Rh2 from the Paris Basin to the eastern French Massif Central and Alpine Zones.

The two Rhaetian sequences, Rh1 and Rh2, are characterized by an increasing areal extent and consequent backstepping of marine deposits. This trend continued until the Mid-Hettangian peak transgression.

**HETTANGIAN IN THE SUBALPINE ZONES AND EASTERN BORDER OF THE MASSIF CENTRAL**

The earliest pelagic cephalopod bearing sediments deposited on the western European post-Hercynian platform are dated as
**FIG. 5.**—The Early Cimmerian unconformity in both the Paris and Germany basins. The two transects across the Paris and Germany basins illustrate the late Norian-Early Jurassic age deposits. The late Norian unconformity is seen in both cases to erode the underlying sequence K5 (shaded). In the Paris Basin, this truncation is observed on the eastern and western margins, whereas in Germany it is observed on the western margin only. Gr: Gamma Ray; Dt: Sonic; R-ILD: Resistivity.

**FIG. 6.**—Organization of 3rd-order depositional sequences within the framework of the earliest 2nd-order transgressive/regressive facies cycles of the major Ligurian cycle across the Paris Basin from latest Norian to Early Sinemurian times. The datum is the peak transgression dated as *Asteroceras stellare* Subzone. Such reconstructions as this one show the Chaunoy area as a positive zone during Liassic times, which explains the presence of oil discoveries in the area. The high separates the Paris Basin into two parts with relatively different deposits on each side. Detrital sediments are more abundant in the western part.
The Hettangian transgression began with widespread *Psiloceras planorbis* Beds. It advanced progressively westward from the Tethys and covered most part of the future Alpine domain and Rhone valley as far as the eastern border of the Massif Central, but reach 100 m in the internal Alpine Zones (Dumont, this volume) and 500 m in the Rhone valley graben (Baudrimont and Dubois, 1977) and probably several hundred meters in the poorly-dated siliciclastic successions of the North Sea grabens (Steel and Ryseth, 1990). The best dating and definition of depositional sequences and their boundaries is provided by the Subalpine sections (Dumont, this volume).

**He, Sequence**

This sequence corresponds to the *Psiloceras planorbis* Beds. Ages can be assigned from the Digne thrust sheet of the Subalpine Zone (Tanaron section; Mouterde and Coadou, 1971; Dumont, this volume) and in the Ardèche (= eastern) border of the Massif Central (Elmi and Mouterde, 1965).

The He subsequence boundary (He, Sb) coincides everywhere with the first flooding surface since no platform margin existed at that time. No erosional surface is observed here. The He, transgressive systems tract (He, TST) lowest parasequences comprise thick-bedded oolitic and pelloidal packstone to grainstones with a wide distribution from the internal Alpine units through the Ardèche (eastern) border of the Massif Central. The overlying parasequences include crinoidal limestones indicative of increased water depth.

The He, maximum flooding surface (He, MFS) has argillaceous layers that yield *Psiloceras planorbis* and *P. psilonotum* in the Subalpine Zones and in the Helvetic nappe du Mont Joly (Barféty and Mouterde, 1978; Triboulet, 1980) that is the earliest ammonite-bearing Hettangian “horizon” (Mouterde and Tintant, 1980). In the eastern Massif Central border, the maximum flooding surface is dated *P. psilonotum* and *P. plicatulum*, which is the next “horizon” above the *Psiloceras planorbis*
Zone. This corresponds to the most widespread distribution of open marine facies of sequence He1 (Elmi and Mouterde, 1965; Martin, 1985). This illustrates the progressive landward and westward shift of parasequences during the development of the transgression. The MFS belonging to two successive parasequences may be expressed as a diachronous transgression and regression.

The He1 highstand systems tract (He1 HST) occurs as an overall regressive trend that is recorded above the Psiloceras planorbis layers and is marked by increased carbonate content and development of lagoonal environments with patch reefs in the Ardèche (Massif Central side) section (Martin, 1985) and the Massif Central cover (Dumont, this volume).

He2 Sequence

A sharp transgressive surface is recorded in the Digne thrust sheet and foreland (Subalpine domain). It is dated approximately in the Psiloceras planorbis/Johnstoni subzonal boundary within the Psiloceras planorbis Zone. Its marks the sequence boundary He2. The corresponding maximum flooding surface is well-dated in the Caloceras johnstoni Subzone by Caloceras torus in the Digne thrust sheet and foreland (Mouterde and Coadou, 1971) and by Caloceras johnstoni in both the Ardèche (eastern Massif Central; Elmi and Mouterde, 1965) and in the northern border of the Provence platform (Alpes Maritimes; Dardeau, 1983). The Caloceras johnstoni flooding is well recorded also in the central Paris Basin (as shown below) but it is not clearly expressed in the Dorset coast section (Hesselbo and Jenkyns, this volume).

He3 Sequence

This sequence is dated as Alsatites liasicus and lower Schlotheimia angulata ammonite Zones. A tectonic extensional event dated of the early Alsatites liasicus Zone is linked with the development of the Tethyan rifting. It gave birth to several individual subbasins and swells each one having its own tectonic history.

The He3 sequence boundary corresponds to the development of karstic surfaces on emerged areas. Coeval surfaces directly intersect Permian strata in the southern Massif Central (Causses). In the continuously subsiding Digne area, the sequence boundary is located at the reversal between regressive and transgressive trends.

The He3 transgressive systems tract is best expressed in hemipelagic basinal environments such as in the Digne thrust sheet (Mouterde and Coadou, 1971) of the Helvetic nappe du Mont Joly (Barféty and Mouterde, 1978; Triboulet, 1980) and nappe de Roselette (Barféty, 1985) in the external Alps. Thinning-upwards marlstone/limestone parasequences are dated there in the Waehneroceras portlocki and then of the Alsatites laqueus Subzones. On the Ardèche border of the Massif Central, the same fauna have been described within nodular limestones overlying transgressive crinoidal limestones (Lumachelle à Ostreidés; Elmi and Mouterde, 1965; Martin, 1985). The transgressive systems tract is dated in Waehneroceras portlocki Subzone of the Alsatites liasicus Zone. The He3 maximum flooding surface has been assigned to the lower part of the Alsatites laqueus Subzone (Laqueus “horizon”) in the Subalps (Dumont, this volume).

The He2 highstand systems tract corresponds to aggradational dolomitic platforms in the eastern Massif Central and Provence bordering areas, overlain by greenish shaly continental layers associated with evaporites (Baudrimont et Dubois, 1977; Courèl et al., 1984; Melas, 1982). Basinal equivalents are dated of the lower Schlotheimia angulata Zone, (Schlotheimia extrarodosa Subzone) and are mainly nodular limestones with shell debris and crinoid ossicles.

The He3 sequence boundary is recorded by subaerial surfaces with lacustrine deposits and/or karstic features associated with influx of detrital quartz in the platform areas. In several areas the Hettangian/Sinemurian boundary is marked by hiatus or by condensed sections that associate Schlotheimia to Arietites bucklandi or even Arnioceras semicosatum ammonite Zones as a probable consequence of intrabasinal block-tilting (Barféty, 1985; Dardeau, 1983). In subsiding areas where hemipelagic deposits were continuous, the environmental conditions do not show significant variation with time. Nevertheless, on the eastern side of the Massif Central (Elmi and Mouterde, 1965), a flooding surface marked by crinoidal limestones overlies a slowing-upward succession. This turning point is dated intra-Arietites bucklandi or even Arietites bucklandi Zone only (He2 sequence). Deposition of crinoidal limestones with shell debris occur. In other places of
the Anglo-Paris Basin, Hettangian section thickness do not exceed 20 m on outcrop.

Hettangian-Lower Sinemurian times are characterized by a peak transgression, the age of which remains questionable between Alsatites liasicus or Ariettes bucklandi Zones in the Paris Basin. This uncertainty results from the aggradational pattern of the strata belonging to sequences He₂ and He₃.

**He₁ Sequence**

Open marine-ammonite-bearing sediments were limited to the central and eastern parts of the Paris Basin (Fig. 9, 10). They are assigned to the *Psiloceras planorbis* beds. In the central Paris Basin (Brie and Champagne), the He₁ sequence boundary is defined by a disconformable surface at the base of a typical limestone unit with echinoid spines (*Miocidaris*; Poujol, 1961). These overlie the fluvialite, varicolored, Argiles de Levallois of latest Rhaetian age. Thus the sequence boundary is both a ravinement surface and a transgressive surface. The He₁ sequence is apparently more erosional than it is in Alpine areas, according to observations from well log inspection (Fig. 11). The Rhaetian/Hettangian boundary is extremely rare at outcrop.

The He₁ transgressive systems tract comprises alternating crinoidal limestones and dark brown, laminated, black shales, with fish scales and bones, plant fragments and strings of pyrite and of fine-grained quartz. This was observed from the wells of the central Paris Basin (Figs. 3, 4, 11, 12, 13, 14; Poujol, 1961, p. 589–592), and at outcrop near Nancy (Guerin Franiatte et al., 1983).

In the southeastern Paris Basin, the He₁ maximum flooding is shown by *Psiloceras planorbis* beds that has been recorded...
The He1 highstand systems tract is documented by a regressive trend that is well marked on the well logs (Figs. 11, 12, 13). It corresponds to crinoidal marly limestones belonging to the upper Psiloceras planorbis Zone. During this period, the continental encroachment of open-marine ammonite-bearing deposits increased towards the west and easy communication with the Tethyan opened across the Burgundy threshold (Figs. 3, 9).

The He2 sequence boundary is marked by the turning point between an earlier regressive trend and later regressive trend dated as *Waehneroceras portlocki* (Grand-Pré well; Figs. 10, 12; outcrop section in Lorraine; Guérin-Franjatte et al., 1983; Fig. 13). Therefore, the comparison with the Subalpine and eastern border of the Massif Central (Dumont, this volume) suggests that the He2 Sb could be dated as close to the *Psiloceras planorbis*/*Liasicus* zonal boundary.

On the Lorraine outcrop section at Mazerulles (Guérin-Franjatte et al., 1983), the He2 transgressive systems tract comprises a laminated marlstone including phosphate nodules and sparse crinoid ossicles interbedded with pelecypod bearing calcareous mudstones. In the deeper part of the Paris Basin (Poujol, 1961), shales with significant organic carbon contents are interbedded with bioclastic wackestones. They are dated *Waehneroceras portlocki* and *Alsattites laqueus* Subzones in the Grand-Pré well.
AMMONITE BIOSTRATIGRAPHIC CORRELATION AND EARLY JURASSIC SEQUENCE STRATIGRAPHY IN FRANCE

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Fig. 11.—Depositional sequences dated as Rhaetian to Pliensbachian age in three wells located in the Central Paris Basin. The cored parts of the Courgivaux and Bouchy-le-Repos wells have provided ammonites that allowed the calibration of the well logs. Correlations between the three wells show the lower Pliensbachian unconformity. See location on Figs. 10A and 10B (from Guérin-Franiatte et al., 1983).

Fig. 12.—Sequence stratigraphic interpretation of the Rhaetian to Hettangian cored part of the Grand Pré well (central Paris Basin). See location on Figure 10.

The He max flooding surface is recorded by a reversal between a transgressive trend dated as Waehneroceras portlocki and Alsattites laqueus Subzones and a regressive trend dated Schlotheimia angulata Zone (Fig. 12, 13, 14; Grand-Pré, Raulecourt, Dontrien and Rethel wells). A noticeable renewal of foraminifer species is observed in the Mazерulles outcrop section in Lorraine (Guérin-Franiatte et al., 1983) and is dated as intra-Alsatites laqueus Subzone. Therefore, the maximum flooding surface is proposed to be dated as intra-Alsatites laiqueus Subzone. In the eastern part of the Paris Basin (Lorraine, Champagne and Brie; Figs. 3, 4, 6), the Alsattites laiqueus maximum flooding is considered as the peak transgression of the 2nd-order cycle that ends within Early Sinemurian time. It compares well with the flooding event dated as Alsatites laiqueus in the Germany basin (Merzereaud et al., 1995).

The Dorset coast section shows two or three parasequences with increasing clay content, dated as Waehneroceras portlocki and lower half of Alsatites laqueus Subzones, and then the return to calcareous sediments within the upper half of the Alsatites laqueus Subzone. Therefore a good candidate maximum flooding surface could correspond to this turning point dated as intra-Alsatites laqueus Subzone (H68 bed of Hesselbo and Jenkyns, 1995). This has not been taken into account by Hesselbo and Jenkyns (this volume), but is consistent with the Paris Basin and Subalpine areas record.

The He max highstand systems tract is related to the upper Alsatites laqueus and lower Schlotheimia angulata deposits which constitute a well-defined prorading succession recorded in well logs in the eastern and east-central Paris Basin (Figs. 10, 14, 13). These correspond to the “Calcaires et Marnes Foie de Veau” known from outcrops in Burgundy (Mouterde and Tinchant, 1980) and a set of four short parasequences in Lorraine (Fig. 12; Guérin-Franiatte et al., 1983).

He3 Sequence

Sequence He3 is dated as part of the Schlotheimia angulata Zone and most of the Ariettes bucklandi Zone and is therefore mainly of Early Sinemurian age. But we consider that its beginning is still within the Schlotheimia angulata Zone, by com-
The lithologies of the He, transgressive systems tract are similar throughout the Paris Basin (Fig. 9), comprising oyster-bearing nodular limestones (Calcaires à Gryphées; Figs. 3, 4) with the exception of sandstone units on the Ardennes side (Grès de Luxembourg; Hanzo, 1980). A well-marked flooding event dated as Schlotheimia angulata Zone (Schlotheimia complanata Subzone ?) covers much wider areas than the Alsatites laqueus flooding on the western side of the Paris Basin (Figs. 3, 4, 6, 9). It is documented by conglomerates and/or dolomites associated with bioclastic limestones dated as Schlotheimia angulata and resting directly on the Hercynian basement. This was observed in the Ardennes (Horon, 1961, p. 19), in Normandy (Rioult, 1980) and on the Northern Massif Central (Mouterde et al., 1980). At the Puiselet well (50 km south of the city of Paris) black calcareous shales yielded Schlotheimia angulata (Durand, 1961, p. 557). In the Dorset section, an argillaceous level located between the Upper White and Speckelly beds (Hesselbo and Jenkyns, 1995) can be tentatively compared to this flooding surface.

The Sinemurian Paris Basin embayment was open towards the Germany platform. It was bounded by the London-Brabant massif at the northeast, the Massif Central at the south, the Armorican massif at the west and by a threshold located approximately at the Normandy/Channel coast (Fig. 4). There the offshore Calcaires à Gryphées (Figs. 3, 4, 6) grade to subemersive dolomites and to sandstones (Dolomies d’Hugleville and Grès de Berneval; Berthe et al., 1961). This threshold coincided with a loose northern boundary of the calcareous belt which that was north of the U.K. during Hettangian times. In fact, the Dorset and Yorkshire coast sections comprise mainly Sinemurian shales and silts with minor calcareous intercalations (Hesselbo and Jenkyns, 1995, this volume). In the northern North Sea, the 250-m-thick Stanford and lower Amundsen Formations dated Hettangian to Sinemurian were dark colored, with significant TOC contents (Bessereau and

parison with the Ardèche (eastern Massif Central section; Du- mont, this volume; Figs. 7, 8) and the Mont Blanc unit (Eltchaninoff, 1980). Open-marine ammonite bearing sediments covered a wider area in the west than previous sequences. The communication with the Tethys Sea became permanent after that time for the rest Early Jurassic times (Fig. 9).

The He, sequence boundary is poorly dated but within the Schlotheimia angulata Zone on the basis of specific peaks on the Paris Basin well log records (Poujol, 1961; Durand, 1961). It is shown by a sharp contact in the Hettange section (Hanzo et al., 1987). In Burgundy (Mouterde and Tintant 1980, p. 93–97), an erosional surface separates the “Calcaires Foie de Veau”, dated as Schlotheimia angulata from the “Calcaires à Gryphées” dated as Ariettes bucklandi Zone, which records He, Sb. This tentatively can be correlated with the Dorset coast section where a hialtal surface and a marked lithological change record a short term regression, according to Donovan et al. (1979). However, evidence in support of an intra-S. angulata Zone sea-level fall or stillstand in the U.K. area is considered as poor by Hesselbo and Jenkyns (this volume). Nevertheless, a well marked regressive phase is dated as intra-S. angulata Zone in the Germany basin (Merzereaud et al., 1995).

The Sinemurian Paris Basin embayment was open towards the Germany platform. It was bounded by the London-Brabant massif at the northeast, the Massif Central at the south, the Armorican massif at the west and by a threshold located approximately at the Normandy/Channel coast (Fig. 4). There the offshore Calcaires à Gryphées (Figs. 3, 4, 6) grade to subemersive dolomites and to sandstones (Dolomies d’Hugleville and Grès de Berneval; Berthe et al., 1961). This threshold coincided with a loose northern boundary of the calcareous belt which that was north of the U.K. during Hettangian times. In fact, the Dorset and Yorkshire coast sections comprise mainly Sinemurian shales and silts with minor calcareous intercalations (Hesselbo and Jenkyns, 1995, this volume). In the northern North Sea, the 250-m-thick Stanford and lower Amundsen Formations dated Hettangian to Sinemurian are mainly siliclastic alluvial plain deposits grading to offshore shales (Steel, 1993).

In the central Paris Basin (Poujol, 1961), the “Calcaires à Gryphées” are 50 to 100 m thick (Figs. 3, 4, 6). These show relatively regular alternations of calcareous mudstone to wackestone layers and laminated marlstone. The wackestones typically contain crinoid ossicles and shell debris. The marlstones are dark colored, with significant TOC contents (Bessereau and

![Fig. 13.—Sequence stratigraphic interpretation and dating of an attenuated sequence in the Hettange area section (Lorraine, eastern Paris Basin). See location on Figure 10.](image-url)
they grade basinward to fine-grained wackestones and mud-sands mainly in subsiding areas at the platform borders, and thickness is less than a few meters. Thicker sedimentary piles are areas such as the Semur-type Section (Corna et al., 1990), thick-homogenous lithologies and absence of turbidites. On platform crinoids and brachiopods and interbedded with marlstones. The dominant lithologies are relatively uniform, highly bioturbated packstones to wackestones containing debris of bivalves, which are associated with sporadic fish and plant remains. The depositional environment was characterized by low-water flat areas and have been redistributed towards the deeper parts by storm processes (Mettraux and Homewood, 1994). This has led to aggradational patterns, the condensation of deposits on platform areas and relative homogeneity of sediments have probably been produced at a steady rate on shallow-water flat areas where the fauna have been collected (Coma and Mouterde, 1987) shows a condensed Vermiceras conybeari bearing surface is dated as close to the Vermiceras conybeari: Coroniceras rotiforme subzonal boundary of the Arietites bucklandi Zone. It is particularly well shown on the Bouchy le Repos and Rethel well log (Fig. 11, 14; Poujol, 1961) at the trend reversal between a strongly retrograding unit dated as Vermiceras conybeari Subzone and a prograding one dated as Coroniceras rotiforme Subzone (Figs. 10, 13). The outcrop stratotype section at Hettange in Lorraine (Fig. 15; Hanzo et al., 1987) shows a condensed Vermiceras conybeari-bearing bed overlain by massive wackestone beds dated as Coroniceras rotiforme Subzone. In the Sinemurian stratotype section at Semur (Burgundy), condensed marly limestone beds dated as Vermiceras conybeari Subzone are characterized by a maximum abundance and diversity of fossils with respect to overlying massive wackestone assigned to the Coroniceras rotiforme Subzone (Coma and Mouterde, 1988).

The interpretation of the Dorset coast section given by Hesselbo and Jenkyns (this volume) is somewhat different. The Vermiceras conybeari/rotiforme subzonal boundary is considered as a sequence boundary, and the whole Coroniceras rotiforme interval is strongly transgressive. This could be explained by differences in the tectonic development between different parts of the Anglo-Paris Basin, or different interpretation of the accommodation space variation.

He Sequences

This sequence is assigned to part of the Lower Sinemurian Arietites bucklandi and Arnioceras semicostatum Zones. It is considered the last depositional sequence that predates the Sinemurian 2nd-order maximum regression in the Paris Basin. The overall regressive character of Si1 is not obvious from well log inspection alone (Fig. 3), the upper Hettangian and Lower Sinemurian being mainly aggradational. However, the following sequences, the overlying “Marnes à Promicroceras” and coarse deposits are clearly retrogradational with respect to the Lower Sinemurian Calcaires à Gryphées (Fig. 11).

In the Paris Basin, the datation of the Si sequence boundary is proposed as intra-Arietites bucklandi Subzone. This results from comparisons between subsurface (Poujol, 1961) and outcrop data in Lorraine and Burgundy stratotype sections (Figs. 15). The intra-A. bucklandi Subzone sequence boundary is very well expressed in almost all the U.K. sections, noticeably in the Hebrides, Glamorgan and Sommerset (Fig. 11; Hesselbo and Jenkyns, this volume). Nevertheless, uncertainties remain in the dating in the Hebrides and the sedimentological interpretation in Glamorgan and Somerset.

The Si transgressive systems tract is dated as Arietites bucklandi (pars), Paracoroniceras charlesi and Agassiceras scipionianum (pars) Subzones. Ammonites of the Arnioceras semicostatum Zone found in the wells of Ouzoir sur Trezée, Bouchy le Repos, Montmirail and Dontrien (Fig. 11, 15; Poujol, 1961).

Guillocheau, 1994); they include some thin silty to sandy layers with sporadic fish and plant remains.

On the southern side of the Paris Basin and in Tethyan areas, the dominant lithologies are relatively uniform, highly bioturbated packstones to wackestones containing debris of bivalves, crinoids and brachiopods and interbedded with marlstones. These sediments were deposited on extended flat ramps at relatively shallow-water depths (perhaps 100 m) as suggested by homogenous lithologies and absence of turbidites. On platform areas such as the Semur-type Section (Corna et al., 1990), thickness is less than a few meters. Thicker sedimentary piles are found as aggrading accumulations of bioclastic calcareous sands mainly in subsiding areas at the platform borders, and they grade basinward to fine-grained wackestones and mudstones such as in the Subalpine domains (Barfété, 1985). Sediments have probably been produced at a steady rate on shallow-water flat areas and have been redistributed towards the deeper parts by storm processes (Mettraux and Homewood, 1994). This has led to aggradational patterns, the condensation of deposits on platform areas and relative homogeneity of sediments in basinal areas. Such features make the placement of sequence boundaries and maximum flooding surfaces difficult.

He Sequences

The He sequence is dated as part of the Schlotheimia angulata and Arietites bucklandi Zones. The He, maximum flood-
Fig. 15.— Compared sequence stratigraphic interpretation of three lower Sinemurian outcrop sections in the eastern and southeastern Paris Basin border. See location on Figure 1. Ammonites zones and subzones are: ang: Schlotheimia angulata; buck: Arietites bucklandi; cha: Paracoroniceras charlesi; ja: Uptonia jamesoni; obt: Asteroceras obtusum; ra: Echioceras raricostatum; sauz: sauzeanum; sc: Agassiceras scipionianum; tur: Caesinites turneri.

The Si₁ TST is divided into two parts by a high-frequency (possibly 4th-order) regressive phase. It has been found in all the Paris Basin well logs (Figs. 11, 13). In the Hebrides there is a very well developed top Paracoroniceras charlesi Subzone regression. The section in Somerset is consistent with this interpretation. The Dorset, Lorraine and Burgundy sections (Fig. 15) comprise bioturbated, nodular, packstone to wackestone beds, less than 4-m-thick. Precise datation by ammonites (Corna and Mouterde, 1988; Corna and Dommergues, 1994) demonstrate the relative condensation of these levels.

The Si₁ maximum flooding surface is dated as infra-Agassiceras scipionianum Subzone (middle Arnioceras semicostatum Zone). It is marked by a surface of condensation and heavy bioturbation in the Hettange section (Lorraine) and by a level of condensation and phosphate enrichment in the Burgundy stratotype sections (Fig. 15). Well-defined surfaces of maxi-
mum flooding dated as *Coroniceras lyra* Subzone (= Paracoroniceras charlesi Subzone) and *Agassiceras scipionianum* Subzone has been described by Hesselbo and Jenkyns (this volume) in the Hebrides, Somerset and Dorset sections (condensed level at the top of the Blue Lias dated as Lowermost Zone) in the Hebrides, Somerset and Dorset sections (condensed level at the top of the Blue Lias dated as Lowermost Zone). The Si 1 MFS is not yet dated formally in the southern subalpine units.

The main part of sequence Si 1 comprises the transgressive phase. In consequence, its prograding highstand phase is limited to a few meters of *Gryphaea*-bearing limestones in the center of the basin, and is reduced to less than 10 cm at the Mandelot (south Burgundy) outcrop (Fig. 15). The Si 1 HST is dated as late *Agassiceras scipionianum* Subzone.

**Si 1 Sequence**

Sequence Si 1 is dated as upper *Arnioceras semicostatum* Zone (*Euagassiceras resupinatum* Subzone) and part of the *Caesinites turneri* Zone. It corresponds to the maximum regression (Fig. 2) in the Paris Basin and central North Sea and is a little later than the peak transgression in the U.K. according to Hesselbo and Jenkyns (this volume).

For Si 2 sequence boundary, consistent dating is provided by the Raasay (Hebrides) and Yorkshire coast (Cleveland Basin) sections. There, mudstones dated as *Agassiceras scipionianum* Subzone are overlain by fine-grained sandstones of *Euagassiceras resupinatum* (= Sauzeanum) age (Hesselbo and Jenkyns, this volume). This can be correlated with a sandstone layer dated as Sauzeanum overlying marlstones of *Agassiceras scipionianum* Subzone at the Lorraine section at Hettange (eastern Paris Basin, Fig. 15). The corresponding unconformity tentatively can be correlated also with a surface showing significant bioturbation and fossil accumulation, dated as intra-*Euagassiceras sauzeanum* (= resupinatum) Subzone at the Semur stratotype section in northern Burgundy (Fig. 15). On the Paris Basin well logs, a clear and consistent sequence boundary has been dated as *Arnioceras semicostatum* Zone in the Bouchy le Repos well (Fig. 11; Poujol 1961).

The Si 2 lowstand systems tract is fine-grained sandstone regressive layers dated as *Euagassiceras resupinatum* in both the Raasay (Hebrides) and Robin’s Hood Bay (Cleveland basin) sections (Hesselbo and Jenkyns, this volume). They compare well with the coeval Sauzeanum sandstone layer of the Hettange section in Lorraine (Fig. 15; Hanzo et al., 1987) but are not present in the attenuated sections of Burgundy. In the more subsiding parts of Lorraine, a 10-m-thick alternation of *Gryphaea*-bearing wackstone and marlstone shows an aggradational to progradational pattern that can be related to the Si 2 LST.

The Si 2 transgressive systems tract is characterized everywhere by a well-marked retrogradational pattern, noticeably on the Paris Basin well-log records (Fig. 3). Its lower part still belongs to the upper part of the *Arnioceras semicostatum* Zone as deduced from ammonites recovered in the Rethel 1 and Bouchy le Repos wells (Fig. 11). At the Semur stratotype section in Northern Burgundy, an aggrading set of four parasequences, 1.5 m thick all together, is dated as top *Spiroceras sauzeanum* Subzone and (?) lower *Caesinites turneri* Subzone. One bed is characterized by the presence of accumulated large *Caesinites* of the lower *Caesinites turneri* Zone. This suggests that the Si 2 transgressive surface is intra-uppermost *Arnioceras semicostatum* Zone and that the Si 2 TST of is dated as Sauzeanum (pars) and *Caesinites turneri* Subzones in this area.

In the U.K., Hesselbo and Jenkyns (this volume) have shown that the coeval Shales-with-Beef of the Wessex basin and Calcareous Shales of the Cleveland basin are strongly expanded compared to the underlying and overlying beds. In contrast, the Digne section (Corna et al., 1990) in the Subalpine area comprises a 8-m-thick section with thinning upward beds of nodular limestones containing accumulations of *Gryphaea arcuata* and dated as *Caesinites turneri* Zone.

For the Si 2 maximum flooding surface, the most precise dating is provided in the Cleveland basin (Yorkshire coast) section by a ferruginous oolite level dated as mid-*Caesinites brooki* (= C. *turneri*) Subzone (van Buchem et al., 1992 and bed 13 by Hesselbo and Jenkyns, 1995, this volume). In the Paris Basin well logs (Rethel, Bouchy le Repos and Courgivaux), the Si 2 MFS postdates the Late *Arnioceras semicostatum* Zone and precedes the *Microderoceras birchi* Subzone (= Late *Caesinites turneri* Zone; Fig. 11; Poujol, 1961). In the Digne area (Subalpine Zones) and in the Mont Joly section (Helvetic units; Barféty, 1985), the Si 3 MFS is recorded by maximum argillaceous content and abundance and diversity of fossils belonging to the *Caesinites turneri* Zone (bed 2078/2079 of Corna et al., 1990). In the Subalpine areas, the Si 3 MFS coincides with the long-term peak transgression T4 (Fig. 2).
The Si₃ highstand systems tract is marked by a regressive phase dated as Microderoceras birchi Subzone in the Paris Basin wells (Fig. 11; Poujol, 1961) as well as in the Dorset coast and Morvern sections (Hesselbo and Jenkyns, this volume). This phase defines the maximum regression R₄a.

THE LATE SINEMURIAN (LOTHARINGIAN) CYCLE 4b IN THE U.K. AND IN THE PARIS BASINS

REGRESSIVE PHASE OF CYCLE 4 IN THE SUBALPINE AREA

In the Tethyan realm, the Upper Sinemurian layers were deposited during the early part of the long-term regression (R₄) that ended during the Tragophylloceras ibex Zone of the early Pliensbachian. In the Paris Basin (Fig. 1), the Late Sinemurian is marked by the transgressive phase of the 2nd-order cycle 4b, by its peak transgression (dated as Asteroceras obtusum/Oxynoticeras oxynotum zonal boundary) and by the lower part of its regressive phase. In the U.K., the Upper Sinemurian comprises the entire T/R cycle 4b and the lower part of the transgressive phase of the subsequent cycle (Fig. 2).

Si₃ Sequence

The Si₃ sequence is dated as uppermost Caesinites turneri, Asteroceras obtusum and part of Oxynoticeras oxynotum Zones (Upper Sinemurian). Its sequence boundary marks a phase of block tilting and subsidence in the nascent Tethyan marginal area (Fig. 17; Coadou and Beaudoin, 1975). In the central and eastern subsiding parts of the Paris Basin, the Gryphaea-bearing limestones grade to strongly argillaceous deposits named Argiles à Promicroceras in Lorraine (Hanzo, 1980) or Argiles du Lotharingien in other areas.

However on parts of the Paris Basin borders (Fig. 18; Dommegues, 1992), the Calcaires à Gryphées lithological type persisted. At other places, the whole Upper Sinemurian section can be attenuated. It comprises crinoidal packstones, 0.5 m-thick at most (Fig. 18). The increased accommodation space above the rugged morphology of the sea bottoms allows the presence of lowstand deposits for the Si₃ sequence which follows the 2nd-order peak transgression belonging to second order cycle 4b. This is the consequence of the influence of Tethyan extensional events that probably induced a 2nd-order transgression. An echo of such events can be evoked from the erosional surface at the base of the Asteroceras obtusum Shale on the Dorset section but not in more northern U.K. areas (Hesselbo and Jenkyns, this volume). The Si₃ sequence is marked by the Asteroceras obtusum transgressive phase, one of the most significant of the Liassic Anglo-Paris Basin and is considered a 2nd-order transgression.

For Si₃ sequence boundary, the most precise dating comes from the Dorset coast section, where a sharp surface separates calcareous mudstone above from organic-rich shale below and is dated as intra-Microderoceras birchi Subzone (upper Caesinites turneri Zone). In a similar manner, Microderoceras birchi sandstone intersects the top of upper Caesinites turneri mudstone at the Yorkshire section (Hesselbo and Jenkyns, this volume). In the Paris Basin well logs (Figs. 3 and 4), the Si₃ Sb just postdates the Microderoceras birchi Subzone (Bouchy le Repos and Courgivaux wells; Fig. 11) and predates the Asteroceras obtusum Zone (Grand Pré and Puiselet wells; Poujol, 1961). This defines the maximum regression which ends second order cycle 4a. In Burgundy (Posange; Fig. 18), the Si₃ Sb is recorded by a hiatal surface that separates the Microderoceras birchi from the Asteroceras stellare Subzones, with Asteroceras obtusum Subzone missing. In many other places, the whole upper Sinemurian is condensed and the Si₃ Sb is amalgamated with other key surfaces. In the Subalpine areas, the Si₃ Sb is tectonically enhanced (Fig. 17). It is marked by condensed sections or by hiatuses of various duration such as at the Caesinites turneri/Asteroceras obtusum zonal boundary at the Castillon lake section (Corna et al., 1990).

Most of the Argiles à Promicroceras in Lorraine (Hanzo, 1980) or Argiles du Lotharingien in the wells of the central Paris Basin correspond to the Si₃ transgressive systems tract. Towards the borders of the basin, they grade to crinoidal packstones/wackestones with mollusk debris including age-diagnostic ammonites (Fig. 18). In the areas interpreted as Sinemurian submarine highs or plateaus, such as Burgundy, the Si₃ sequence time interval is generally represented by crinoidal wackestones to packstones with phosphate concentration and fossil accumulation. These belong commonly to the Asteroceras stellare and Eoarctites denotatus Subzones, the other subzones related to Si₃ sequence being represented only locally. A well-marked gamma-ray peak is dated as Asteroceras stellare Subzone at several wells (Figs. 7, 11). It marks both the Si₃ MFS and the peak transgression of cycle 4b in the Paris Basin (Figs. 3, 4).
In the Swiss Prealps (Jansegg section), Dommergues and Meister (1987) described a 20-cm-thick bed of crinoidal packstone including phosphate grains and fossil debris related to the upper part of the Asteroceras stellare Subzone and to the Eoarietites denotatus Subzone. Hence, a Asteroceras stellare or Eoarietites denotatus age is proposed for the of Si, MFS in this area.

In the distal parts of the basinal areas, shaly sediments such as the Asteroceras obtusum Shales (Wessex basin) or the Argoles à Promicroceras (Lorraine, eastern Paris Basin) overlie coarser and/or more calcareous sediments. In both the Wessex and Cleveland basins the transgression occurred in two phases, with an earlier MFS dated as Asteroceras obtusum Subzone and a later MFS dated as Eoarietites denotatus Subzone. These are separated by a short-term regression related to the Asteroceras stellare Subzone (Hesselbo and Jenkyns, this volume). In the Paris Basin, there is only one transgressive phase within the Asteroceras obtusum Zone which culminates in the Eoarietites denotatus Subzone. Asteroceras stellare sediments are strongly transgressive. In the North Sea, a maximum flooding surface dated as Oxynoticeras oxynotum Zone (Steel and Ryseth, 1990) is the closest in age with respect to the MFS dated as Eoarietites denotatus Subzone.

The Si4 highstand systems tract is dated as Oxynoticeras simpsoni Subzone. In the Burgundy outcrop section, it is represented only by a discontinuous veneer of crinoidal packstone or wackestone (Fig. 18). In the central and western areas of the Paris Basin (Figs. 3, 4, 11), its corresponds to calcareous, prograding beds, following the Asteroceras stellare Argoles à Promicroceras. In Lorraine, the well logs show that marlstones were deposited continuously through the upper Asteroceras obtusum and lower Oxynoticeras oxynotum Zones.

Si4 Sequence

Uppermost Sinemurian deposition is marked by an overall regression in both the Tethyan and Paris Basin areas. The Si4 sequence is dated as upper Oxynoticeras oxynotum Zone and most of the Echioceras raricostatum Zone. It is the earliest Liassic depositional sequence that shows significant lowstand
deposits. This is a consequence of the maximum accommodation space, due to the 2nd-order peak transgression dated as *Asteroceras obtusum* and to the tectonic instability and differential subsidence during this period.

The *Echioceras raricostatum* Subzone dates a significant 3rd-order maximum flooding surface, which is recorded from Southeastern France through the Anglo-Paris Basin. It marks a transgressive episode within the overall regression phase that culminates in the lower Pliensbachian strata in these areas (Graciansky and Jacquin, this volume). The *Echioceras raricostatum* transgression, also known in the North Sea is not dated at the subzone level (Steel and Ryseth, 1990). It is considered both a 3rd-order MFS and 2nd-order peak transgression in these northern areas. This is a consequence of the different timing of the structural development of the North Sea from Tethyan localities.

The Si₄ sequence boundary is dated as the *Oxynoticeras simpsoni/Oxynoticeras oxytorn* subzonal boundary on the Yorkshire coast section (Hesselbo and Jenkyns, this volume) and in the Paris Basin. On the central and western parts of the Paris Basin well logs (Fig. 3, 4, 11), the stratal pattern trend reversal falls within the *Oxynoticeras oxytorn* ammonite Zone, with no more precision. Good dates are provided by Burgundy outcrop sections interpreted as small-scale graben fillings (e.g., Epiry and Posange; Fig. 18) which show a sharp lithological change at the *Oxynoticeras simpsoni/Oxynoticeras oxytorn* subzonal boundary. On the highest submarine swells such as Remilly and Menetoy, *Oxynoticeras simpsoni* and/or *Crucilobiceras densinodulum* Subzones are missing (Dommergues, 1993). Similarly, the whole *Oxynoticeras oxytorn* Zone is consistently missing in the ammonite harvests from the raised edges of tilted blocks involved in the Swiss Prealp folding (Dommergues and Meister, 1987).

The Si₄ lowstand systems tract can be best defined along an east-west transect drawn from discrete outcrops across the Burgundy swell. The *Oxynoticeras oxytorn* and *Crucilobiceras densinodulum* Subzones are present only in the more subsiding areas and show significant thickness variations (Fig. 18). But the overlying *Echioceras raricostatum* and *Leptechioceras macdonnelli* Subzones continuously blanket both minor horsts and grabens with similar thickness (Mornay-sur-Allier; Fig. 18). The *Oxynoticeras oxytorn* and *Crucilobiceras densinodulum* (pars) Subzones are considered to be the ages of the lowstand systems tracts which follows the *Asteroceras obtusum* 2nd-order peak transgression. In the Subalpine Zones, a massive layer, 10 to 50 m thick, of crinoidal cherty packstone is dated as post-*Asteroceras obtusum* Zone at some places (Corna et al., 1990) and is interpreted from its stratal pattern as Si₄ LST.

The Si₄ transgressive systems tract and maximum flooding surface are dated as *Echioceras raricostatum* Subzone. This is known on the Provence platform (Arnaud and Monleau, 1979), in the Subalpine Zone near Grenoble (Bas, 1988), on the Rosellette Sinemurian platform (Helvetic nappes; Barféty, 1985), on the raised edge of several tilted blocks involved in the Swiss Prealps thrust sheets (Dommergues and Meister, 1987), in the Jura (Salins les Bains section; Dommergues 1993), in Lorraine (Calcaire ocreux; Hanzo, 1980, p. 93–94; Guérin et al., 1961), in the Quercy (Cubaynes, 1986) and in Burgundy (Dommergues, 1993; Fig. 18). The more frequent lithologies around the Si₄ MFS are phosphatic crinoidal packstones and/or ferruginous oolites with accumulated mollusks including ammonites.

From the vertical evolution of the stratatal patterns, the Si₄ maximum flooding surface in Burgundy can be placed at the boundary between the *Echioceras raricostatum* and Boehmi “horizons” of the *Echioceras raricostatum* Subzone (Fig. 18, Epiry section; Dommergues, 1993). Similar conclusions can be drawn from outcrop sections in the Swiss Préalpes médiannes (Dommergues and Meister, 1987; Dommergues et al., 1990). In the Hebrides basin, a candidate Si₄ MFS is recognized at the *Crucilobiceras densinodulum/Echioceras raricostatum* subzonal boundary as well as in the Dorset coast section (Hesselbo and Jenkyns, this volume).

In the Hebrides basin, regressive deposits dated as upper *Echioceras raricostatoides* and *Leptechioceras macdonnelli* follow the maximum flooding surface and are considered as Si₄ highstand systems tract. Correlative wackestones to packstones related to the Boehmi “horizon” of the *Echioceras raricostatum* Subzone and to *Leptechioceras macdonnelli* Subzone date the Si₄ HST in Burgundy (Menetoy section, Fig. 18) and in the Swiss Préalpes médiannes (Dommergues and Meister, 1987). The Si₄ HST is well expressed on Paris Basin well logs by a prograding stratal pattern (Fig. 11).

**THE LOWER PLIENSBACHIAN (CARIXIAN) REGRESSIVE PHASE OF CYCLE 4b IN THE PARIS BASIN.**

**PART OF CYCLE 4 IN THE SUBALPINE AREA AND CYCLE 5 IN THE U.K.**

The Paris Basin and Subalpine Lower Pliensbachian strata are divided into two parts by a 2nd-order maximum regression (top T/R cycle 4) as a consequence of a phase of block faulting and subsequent subsidence dated as mid-*Tragophylloceras ibex* ammonite Zone (Figs. 2, 11, 19). In the U.K., the Lower Pliensbachian units comprise the asymmetrical 2nd-order cycle 5. The lower maximum regression was close to the Sinemurian/Pliensbachian boundary. The peak transgression is reached in the *Phricodoceras taylori* Subzone of the Lowermost Pliensbachian time (Hesselbo and Jenkyns, this volume). This is followed by a long-term regression culminating in the latest Pliensbachian (Fig. 2). The *Phricodoceras taylori* peak transgression is not so well expressed in more southern areas. Nevertheless, the *Phricodoceras taylori* Zone is marked in the Quercy area (Fig. 1) by flooding of the former Sinemurian platform (Meister and Sciau, 1988). Such differences result from different timing in the development of the extensional tectonics between our study areas (Graciansky et al., this volume).

**Si₅ Sequence**

The earliest depositional sequence of the Pliensbachian begins within the uppermost Sinemurian and is named Si₅ according to our rules (Fig. 20). Reference sections are found in the U.K. The Burgundy outcrop sections are frequently condensed or hiatus for such ages (Fig. 19). In the Paris Basin (Fig. 4), the Si₅ sequence is missing in many places as a consequence of a mid-*Tragophylloceras ibex* erosional phase (Lower Pliensbachian; Fig. 11, 19).

The Si₅ sequence boundary is dated as early *Paltechioceras aplanatum* Subzone of the *Echioceras raricostatum* Zone in the Hebrides section (Hesselbo and Jenkyns, this volume). In the
Yorkshire section, the sharp contact at the base Landing Scar sandstone, dated as lowermost *P. aplanatum* Subzone can be correlated with the sequence boundary of \( \text{Si}_5 \), but Hesselbo and Jenkyns (1995; volume) stop short of this interpretation. At the Dorset coast section, \( \text{Si}_5 \) is marked by a gap covering the *Leptechioceras macdonnelli* and *P. aplanatum* Subzones that makes the TST of \( \text{Si}_5 \) (Echioceras raricostatum Subzone) directly overlap by the of \( \text{Si}_4 \) (Phricodoceras taylori Subzone) along the Hummocky Bed (Hesselbo and Jenkyns, this volume).

In the Paris Basin well logs, peaks characteristic of the \( \text{Si}_5 \), have been identified in the range of supra-*Echioceras raricostatum* Subzone and sub-*Uptonia jamesoni* Zone (Fig. 11; Poujol, 1961). It can be considered as the \( \text{Si}_5 \) sequence boundary by comparison with the U.K. section. In Burgundy outcrop sections (Me- netoy, Fig. 18; Dommergues, 1993) the post-*Leptechioceras macdonnelli* and pre-*Uptonia jamesoni* Subzones surface is characterized by a sharp change in lithology and stratal pattern and is interpreted as the \( \text{Si}_5 \) sequence boundary. This sequence boundary also can be found in the Causses area (Southern Massif Central: Meister, 1986; Meister and Sciau, 1988) as a ravinement surface separating Sinemurian platform carbonates from offshore alternating black shales and mudstones dated as *Phricodoceras taylori* Subzone (Fig. 20).

In the Causses section (Meister, 1986; Meister and Sciau, 1988), the \( \text{Si}_5 \) transgressive systems tract shows the lowermost Pliensbachian (*Phricodoceras taylori* Subzone) as transgressive onto Sinemurian carbonate platforms (Fig. 20). This is consistent with the organic-rich Pyritous Shales dated as *Phricodoceras taylori* Subzone in the Yorkshire section and coeval deposits that are considered as clearly transgressive in every studied Liassic U.K. sections by Hesselbo and Jenkyns (this volume). A small point of discussion concerns the shales dated as *Paltechioceras aplanatum* Subzone which overlie the Landing Scar sands at the Yorkshire coast section and show regressive features just below the offshore *Phricodoceras taylori* Pyritous Shales. The Aplanatum shales could be considered as a short-term lowstand deposit as tentatively shown on our chart, but with no better argument than a lithological change around the Sinemurian/Pliensbachian boundary. Another interpretation is to consider the Aplanatum shales as an early parasequence of the \( \text{Si}_5 \) TST, as proposed by Hesselbo and Jenkyns (this volume). In that case, the bed with accumulated fossils of the Donovani “horizon” (Dommergues and Meister, 1992) corresponds to one of the flooding surfaces belonging to the lower \( \text{Si}_5 \) TST.

On the Paris Basin well logs, the interval assigned to the *Uptonia jamesoni* Zone in two wells and overlaying the conventional sequence boundary \( \text{Si}_5 \) is clearly retrogradational (Figs. 10; 18), but no dating at the subzone level is available. In the outcrop sections in Burgundy, the lower part of the *Uptonia jamesoni* Zone (*Phricodoceras taylori* and *Platypleuroceras brevispina* Subzones) is most frequently not represented, and black shales of the *Uptonia jamesoni* Subzone directly overly Upper Sinemurian beds. Exceptions are local small scale graben such as Corbigny or Maconge (Fig. 21). Therefore the reference sections for this interval are chosen in the U.K. and particularly in the Yorkshire coast outcrop.

The \( \text{Si}_5 \) maximum flooding surface coincides with the organic-rich Pyritous Shales in the Yorkshire coast section and similar deposits in other parts of the U.K. (Hesselbo and Jenkyns, this volume), which provides the most precise dates from our available sections. A surface with local fossil accumulation in Burgundy (Fig. 21; Corbigny section: Dommergues, 1979; Hesselbo and Jenkyns, this volume), because of the regressive character of the Ironstone Shales in the Yorkshire section and the Pabay Beds on Pabay in the Hebrides (Fig. 1). A similar trend is observed in the Causses section (Fig. 20), but there is nothing clear on the outcrop sections in Burgundy (Fig. 21).

**\( \text{Pl}_1 \) Sequence**

The \( \text{Pl}_1 \) sequence is dated as Lower Pliensbachian (Upper *Uptonia jamesoni* and Lower *Tragophylloceras ibex* Zones). The stratigraphy of this part of the Paris Basin section (Guérin et al., 1961) was not clear in the past because of erosional unconformities and condensation that were not yet documented from the outcrop studies (Fig. 4). In many parts of the Paris Basin, the well log records show that both sequences \( \text{Si}_5 \) and \( \text{Pl}_1 \) were eroded prior to the deposition of the transgressive \( \text{Pl}_2 \) sequence (Figs. 11, 16). In Lorraine, where the \( \text{Pl}_1 \) sequence was preserved, it is part of the Marnes à Numismalis sedimentation. Its clearer record is given by well logs. The \( \text{Pl}_1 \) maximum
flooding surface is assigned to the *Uptonia jamesoni* Subzone, as suggested from the Burgundy outcrops. There, condensed sections dated as *Uptonia jamesoni* Subzone are present on the structural highs. At some places there is only a veneer of *Uptonia jamesoni* marlstone or crinoidal limestone (Fig. 21, Marseille and Mandelot sections, Domergues, 1979). Similar observations can be made on the Causses sections in the Southern Massif Central, where *Uptonia jamesoni* Subzone has been found within marly layers present in both structural high and basinal environments (Fig. 20; Meister, 1986).

In the Quercy area, Rey and Cubaynes (1991) place a sequence boundary at the base of a regressive trend starting in the *Caesinites turneri* Subzone (Fig. 22). In consequence, they do not document the existence of two individual sequences within the *Uptonia jamesoni* Zone in the outcrops. Therefore the Pl1 sequence of Rey and Cubaynes (1991) in the Quercy is time-equivalent of our Si5 plus Pl1 sequences on our chart. Another interpretation consistent with Paris Basin data would be to have a *Polymorphites polymorphus* Zone sequence boundary.

On subsiding parts of the Tethyan marginal areas, the lowermost Pliensbachian units are represented at many places of the Subalpine Zones by massive, 20- to 100-m-thick cherty cross-bedded crinoidal grainstones. They can be interpreted from their stratal pattern as lowstand deposits. Dating is poor. They are post-*Asteroceras obtusum* and pre-*Tragophylloceras ibex* ammonite Zone. They represent the lowstand systems tract of Pl1 and/or Si5 sequences. But on the submarine swells such as in the Préalpes Médianes, the *Polymorphites polymorphus* to
**Uptonia jamesoni** Subzones time interval is represented by a condensed level (Dommergues, 1987).

In the Yorkshire section, Hesselbo and Jenkyns (1995) have shown a relatively sharp gradation between Pyritous Shales dated as *Phricodoceras taylori* Subzone and Ironstone shales dated as *Polymorphites polymorphus* Subzone. This gradational boundary can be correlated with a moderately expressed sequence boundary in the Hebrides basin and with the Pl 1 sequence boundary in the Paris Basin. Hesselbo and Jenkyns (this volume) do not indicate a MFS in the *Uptonia jamesoni* Subzone on the Yorkshire coast section; nevertheless van Buchem et al. (1992; this volume) mention the existence of a condensed horizon which a shelly, sandy, bioturbated, glauconitic, oolitic level dated as upper *Uptonia jamesoni* Zone in the Nettleton area. An *Uptonia jamesoni* Subzone transgressive phase is evident in the Hebrides (Hesselbo and Jenkyns, this volume). This horizon could represent the equivalent of the Pl 1 MFS in the Paris Basin.

**Pl 1 Sequence**

The Pl 1 sequence is dated as lower to middle Carixian (*Trophiphyloceras ibex* and lower *Productyloceras davoei* Zones, Lower Pliensbachian). It is found in the Paris, eastern Aquitaine, Causses (Fig. 22), and Subalpine Basins by the development of relatively thick lowstand deposits in the more subsiding areas. In the Paris and Subalpine areas, an extensional phase linked with the Tethyan rift-system caused the development of an erosional truncation surface on the raised edges of faulted and/or tilted blocks. This surface is a good marker horizon that can be traced using both subsurface and outcrop data in these areas (Coadou and Beaudoin, 1975; Figs. 11, 17, 19) and is considered as the maximum regression that ends the 2nd-order cycle 4. Similar events are not clearly present in the U.K. stratigraphical record.

On the submarine highs in the Paris Basin, the Pl 1 sequence...
boundary is an erosional surface that has silty shales dated as Tragophylloceras ibex Zone directly overlying the Upper Si- nemurian deposits related to the Si₄ sequence (Figs. 11, 19). The Pl₁ Sb is well shown in the western Paris Basin (Figs. 4, 22) by 2-m-thick calcareous, feldspathic sandstone that overlies crinoidal packstones (Durand, 1961). In the Burgundy outcrop sections, a sharp erosional surface runs near the Tragophylloceras ibex/Uptonia jamesoni zonal boundary (Fig. 21, Malain and Mandelot; Dommergues, 1979). In both the Quercy (Le- favrais and Lafaurie, 1980) and the Causses (Fig. 20; Meister, 1986), a well marked change in stratal pattern dated as close to the Tragophylloceras ibex/Uptonia jamesoni zonal boundary records the Pl₁ Sb. In the Digne area (Subalpine Zones), the Pl₁ Sb separates thick-bedded crinoidal packstones tentatively related to the lowermost Tragophylloceras ibex and Uptonia ja- mesoni Zones, from highly bioturbated, condensed wacke- stones that yielded ammonites belonging to Beaniceras luridum through Aegoceras capricornus Subzones (i.e., to Pl₁ and Pl₂ sequences; Dommergues, 1994, pers. commun.). Detailed mapping in the area shows this surface as clearly erosional (Coadou and Beaudoin, 1975) and as intersecting the raised edges of major tilted blocks (Fig. 17).
From available dates, we consider that sequence boundary Pl2 is lowermost *Tropidoceras masseanum* Subzone in both the Paris and Causses Basins (Fig. 20). In the U.K., Hesselbo and Jenkyns (this volume) have placed the corresponding sequence boundary within the *Acanthopleuroceras valdani* Subzone from their data on the Ironstone Shales in the Yorkshire coast section and on the Belemnite Marls in Dorset. This is consistent with the intra-*Acanthopleuroceras valdani* Pl2 sequence boundary of the Query area (Fig. 22). Nevertheless, both layers dated as *Tropidoceras masseanum* and *Acanthopleuroceras valdani* Subzones are markedly thin in the U.K. sections with respect to those overlying and underlying, and the stratigraphical record is probably incomplete in this part. In consequence we cannot exclude a datation of the Sb Pl2 date as lower *Tropidoceras masseanum* Subzone in the Dorset and Yorkshire sections. The areal extension of *Tropidoceras masseanum*-condensed levels documents its relatively transgressive character over western Europe (Dommergues and Meister, 1992), including several areas of the outer Alps, such as Oisans, Serre-Ponçon and Chablis. Therefore, the sequence stratigraphic interpretation of the lowest Pliensbachian strata could be reappraised later.

The Pl1 lowstand systems tract is not well developed in the Paris Basin. Exceptions are the more subsiding areas such as in Champagne or the Loire graben (Fig. 21; Dommergues, 1979). Within a major half-graben located in the Causses (Southern Massif Central; Meister, 1986), the Pl1 LST corresponds to thickening upward alternations of packstones and marlstones dated as *Acanthopleuroceras valdani* Subzone half of the total. The Pliensbachian succession was deposited within a locally subsiding area (Fig. 20). In the subsiding areas of the Digne unit (Subalpine Zones), alternating mudstones to marlstones must correspond to Pl1 LST, but the ammonites are there extremely rare in relation to the high sedimentation rate, and dates are non-existent. The top LST surface of sequence Pl1 defines the maximum regression of the second order cycle 4 (Fig. 2).

The Pl1 transgressive systems tract extended over wide areas of the European craton. Deposits dated as *Beaniceras luridum* (*Tragophylloceras ibex* Zone) have been recorded in most of the study area (Hesselbo and Jenkyns, this volume; Rey and Cubaynes, 1991; Meister, 1986; Dommergues, 1979). This marks the onset of the long-term transgression T5.

For example in the mostly hiatal sections in Burgundy, crinoid packstones containing *Beaniceras luridum* ammonites are present on the less subsiding and shallower submarine highs (Fig. 21, Belle-Idée; Dommergues, 1979). In the Digne area (Subalpine Zones), highly bioturbated packstones dated as *Beaniceras luridum* Subzone rest on an erosional surface which truncates the raised edge of a major tilted block (Dommergues, pers. comm., 1994). In most studied sections, a change in the stratigraphic pattern is well marked in the uppermost *Acanthopleuroceras valdani* Subzone (Figs. 19, 20) and record the transgressive surface. In the Dorset coast section, which is a distal area of the Wessex basin, the transgressive systems tract comprises two superimposed condensed beds, 0.5 m total thickness only. The lower one is the Belemnite bed dated as upper *Acanthopleuroceras valdani* Subzone, and it may represent the transgressive surface. The upper one is the Belemnite bed, dated as *Beaniceras luridum* Subzone and it is correlatable with the Pl1 maximum flood surface. The total thickness of the two beds is less than a half-meter. The thinness of the TST in this area is evidence of the areal extent of the *Beaniceras luridum* transgression (Hesselbo and Jenkyns, 1995; this volume). The Pl1 MFS in the Yorkshire section belongs to a lower parasequence of the *Beaniceras luridum* Subzone (Hesselbo and Jenkyns, this volume). In the Paris Basin, the maximum flooding surface is dated as *Beaniceras luridum* “horizon” (top of the *Beaniceras luridum* Subzone), and it is marked by the ingress of numerous *Lytoceratidaceae* of Mediterranean type.

The Pl1 highstand systems tract is dated as upper *Beaniceras luridum* Subzone and *Aegoceras maculatum* Subzone. Lithologies in the basinal areas are shales, such as the Ironstone Shales in Yorkshire and the lowermost Green Ammonite Beds in Dorset. Alternating marlstones and mudstones are present in the lower part of the Calcaire à *Prodactylioceras davoei* in Lorraine (Fig. 11), and similar alternations are dated as *Aegoceras maculatum* Subzone in the Causses (Fig. 20; Meister, 1986). On the submarine highs such as in Burgundy (Fig. 21, Belle-Idée) or in the Causses, crinoidal limestones prevail where present.

**Pl2 Sequence**

Sequence Pl2 is characterized by a widespread transgression dated as *Aegoceras capricornus* Subzone. This is interpreted as a 2nd-order peak transgression in the Paris Basin (Figs. 2, 3, 11) and it is recorded in Quercy and Causses areas. In the Subalpine domain, it marks a step of the overall transgression which peak is reached within the *Amaltheus stokesi* Subzone. In the U.K. sections, on the contrary, the *Aegoceras capricornus* transgression is not clearly recorded (except in the Yorkshire coast section) within the overall regression, which culminates at the lower to upper Pliensbachian boundary. In the shallow parts and on the submarine highs of the Subalpine areas, sequences Pl2, Pl1, and Pl0 are generally amalgamated within condensed sections that underlie the Upper Pliensbachian (Domerian) marlstone (Dommergues and Meister, 1990).

The Pl2 sequence boundary is well recorded in the Paris Basin well logs (Figs. 3, 4). It is located within the Calcaires à *Prodactylioceras davoei* in Lorraine (Hanzo and Espitalié, 1994) and corresponds to a sharp lithological change from calcareous shales dated as *Aegoceras maculatum* Subzone to packstones dated as *A. capricornus* Subzone in the more basinal parts of the Burgundy high (Fig. 21, Maconge section; Dommergues, 1979). In Quercy (Cubaynes, 1986, p. 172–206), a sharp contact dated as intra-upper *A. maculatum* Subzone separates bioclastic wackestones from mudstones and represents Pl1 Sb (Fig. 22), although it has not been considered as a sequence boundary by Rey and Cubaynes (1991). In the Yorkshire coast section an erosional surface below the Oyster Bed separates *A. maculatum* from *A. capricornus* deposits and is correlatable with Pl1 Sb in France. But there is no clear equivalent within the Green Ammonite Beds of the Dorset Coast section.

In Lorraine, the upper Calcaires à *Prodactylioceras davoei* and its subsurface equivalent are clearly retrogradational (Fig. 23) and represent the Pl2 transgressive systems tract. These limestones grade to marlstones named Marnes Moyennes in Normandy (Fig. 4; Durand, 1961), the lower part of which is dated as *A. capricornus* Subzone from ammonite recovery in
various wells (Fig. 11). On the structural highs in Burgundy (Fig. 21; Dommergues, 1979), in the Causses (Fig. 20; Meister, 1986) and in the Subalps (Graciesky et al., 1993), crinoidal limestones dated as A. capricornus Subzone correspond probably to the Pl₃, MFS. In the Subalps, the Paris Basin and the Dorset, the A. capricornus/Oistoceras figulinum subzonal boundary is marked by ferruginous encrustments and by accumulations of Prodactylioceras davoei and of other ammonites including species of the southern (Mediterranean) faunal province. In the Yorkshire section the MFS, transgressive surface and Pl₁, Sb lie close together around the Oyster bed (Hesselo and Jenkyns, this volume).

In the Paris Basin (Fig. 19), Causses (Fig. 20; Meister, 1986) and Quercy (Lefavrais and Lafaurie, 1980, Cubaynes, 1986), the Pl₁ highstand systems tract is dated as upper Aegoceras capricornus and lower Oistoceras figulinum Subzones. It is represented by thin layers of crinoidal packstones to wackestones that prograde from the edges of the basins. Coeval lithologies in basinal areas are marlstones that grade upward to the Upper Pliensbachian marlstones named Marnes à Amalthées in Lorraine (eastern Paris Basin) or Marnes moyennes in Normandy (northwestern Paris Basin). These infilled the accommodation space created during the A. capricornus peak transgression (Fig. 11). The situation is different in the U.K. sections, especially in the Yorkshire coast section where the Aegoceras capricornus and Oistoceras figulinum Subzones are coarsening-upward shales and silts that are part of the overall regression (R5) culminating with the Staithes Sandstones dated as Amaltheus stokesi Subzone.

THE UPPER PLIENSCHABIAN (DOMERIAN) CYCLE 5 IN FRANCE AND IN THE NORTH SEA-ONSET OF TRANSGRESSIVE PHASE OF CYCLE 6 IN THE U.K.

In all of western Europe, relative sea-level fluctuations were relatively rapid through late Pliensbachian times when compared with underlying and overlying units. This resulted in short-duration 3rd-order sequences. Sequences Pl₁, Pl₂, and Pl₃ are found from the U.K. to southern France (Fig. 24). The overall stratigraphical trend differs between France and the U.K. (Fig. 2). In France, the upper Pliensbachian substage (Domerian) is characterized by the overall regression R5. This begins in the late Carixian (Aegoceras capricornus Subzone) in the Paris Basin or around the lower/upper Pliensbachian boundary (Oistoceras figulinum Subzone) in Quercy (Fig. 22) and in the Subalps (Graciesky et al., 1993).

In the most subsident areas such as the basinal parts of the Paris Basin (Figs. 3 and 4), the Causses (Fig. 20; Dommergues and Meister, 1986) and the Subalps (Graciesky et al., 1993), 3rd-order sequences Pl₁, Pl₂, and Pl₃ comprise relatively homogenous micaceous shales dated by Amaltheids and are not easily discernible. They constitute the infilling phase of a 2nd-order regressive half cycle (R5). This is followed by the prograding phase of half-cycle R5 comprising coarse deposits related to sequences Pl₁ and Pl₂ (Figs. 21, 25, 26).

The maximum regression is marked by coarse-grained bioclastic limestones and/or by sandy sandstones (Grès médio-liaisiques in the Paris Basin) which reach the Pleuroceras hawskerense Subzone (uppermost Pliensbachian). In contrast, transgression in the U.K. area appears to begin earlier and follows the deposition of the Staithes Sandstones (Yorkshire coast section), dated as Amaltheus stokesi and its time-equivalents in other basins.

Pl₁ Sequence

The Pl₁ sequence straddles the lower and upper Pliensbachian deposits. Its main part is within the Amaltheid-bearing marls in French basinal areas (Figs. 3, 4). In the Subalpine basins, the whole sequence is amalgamated with the underlying sequences Pl₀ and Pl₁ on the positive areas, but in basinal areas dating is poor, and placement of these sequences remains uncertain.

Pl₁ sequence boundary clearer expression can be found in the Quercy (Cubaynes, 1986, p. 208 and 234). There, a regional condensed surface dated intra-Oistoceras figulinum Subzone separates sponge-spicule-bearing wackestones from silty marlstones. It is numbered Pl₁ by Rey and Cubaynes (1991) in the local nomenclature but it correlates well with sequence boundary Pl₁ of our chart. In the Yorkshire coast section, Hesselo and Jenkyns (this volume) have shown the existence of a sequence boundary dated as intra-Oistoceras figulinum Subzone within the Staithes sandstones (Fig. 24); but there is no clear equivalent in the Dorset coast section.

In the most subsident parts of the Paris Basin, the Pl₁, lowstand systems tract can be related to a modestly progradational unit, 20- to 30-m-thick (Fig. 27), located in the lower part of Pl₁ sequence. In the Causses (Meister, 1986), Quercy (Cubaynes, 1986) and Burgundy (Dommergues, 1979; Figs. 1, 18), Pl₁ sequence is relatively thin corresponding to its short duration. There is apparently no space for a significant LST even though Rey and Cubaynes (1991) interpreted four beds of their Rangs de Pavés Formation as a LST (Fig. 22).
Pl$_4$ transgressive systems and highstand systems tracts are hard to separate within the Amaltheid-bearing marlstones in the basinal areas of the Subalps and Paris Basin and in the Argiles Grises of Quercy (Fig. 22; Rey and Cubaynes, 1991). In Burgundy, sequence Pl$_4$ is condensed within crinoidal wackestones to packstones and even amalgamated in places with the underlying and overlying sequences (Macongue sections, Fig. 21). Nevertheless, nodular fossiliferous accumulations within bioturbated wackestones in the Causses (Meister, 1989) and in Burgundy (Sologny: Dommergues and Mouterde, 1980), dated as lower Amaltheus stokesi Subzone (Protogrammoceras occidentale and P. monestieri “horizons”), are good candidates for the TST of Pl$_4$. There the MFS could be dated as Protogrammoceras monestieri “horizon”. Consistently similar observations have been made in Subalpine areas along the Serre-Ponçon lake shoreline (Turin et al., 1984). In the Yorkshire coast section, a thin shaly layer at the lower part of the Staithes sandstones (bed 6, lower part of the Amaltheus stokesi Subzone; Hesselbo and Jenkyns, 1995) and equivalents in the Hebrides and Wessex basins (Hesselbo and Jenkyns this volume) is interpreted by Hesselbo and Jenkyns (this volume) as representative of a short-duration flooding.

**Pl$_5$ Sequence**

Pl$_5$ sequence is dated as Amaltheus stokesi Subzone of the upper Pliensbachian substage. It corresponds to a 2nd-order maximum regression as shown by the Staithes Sandstones deposited in shoreface environments in the Cleveland Basin (Yorkshire coast section) and its time equivalents (Hesselbo and Jenkyns, this volume) but has no marked counterpart in more southern areas (Fig. 2).

The clearest definition of the Pl$_5$ sequence boundary is a sharp contact at the base of a coarse layer belonging to the Staithes sandstones (bed 6, lower part of the Amaltheus stokesi Subzone; Hesselbo and Jenkyns, 1995) and equivalents in the Hebrides and Wessex basins (Hesselbo and Jenkyns this volume). Within the Paris Basin offshore deposits, there is no clear equivalent of this sequence boundary except for a subtle stratigraphic trend-reversal in the upper third of the Amaltheid-bearing marls in the less subsident areas (Figs. 3 and 27). In the Causses area, the sequence boundary Pl$_5$ is placed tentatively at a lithological change at the base of the Protogrammoceras celebratum “horizon” of the A. stokesi Subzone (Fig. 25, 26; Meister, 1986). The Protogrammoceras celebratum “horizon” is marked by the ingress of Mediterranean ammonites in these areas.

The Pl$_5$ maximum flooding surface is uppermost A. stokesi in the U.K. sections (the Osmetalherley Seam in the Lower Cleveland Ironstone of the Yorkshire coast section: Hesselbo and Jenkyns, 1995). It is recorded by the finer grained deposits of the interval. There are no obvious equivalents in more southerly areas except in the Causses, where a double wackestone layer interbedded with dark-colored marlstone is dated as Protogrammoceras celebratum “horizon” of the upper Amaltheus stokesi Subzone (Meister, 1986; layers 113–144 at Rivière sur Tarn; Fig. 25). This is a good candidate for the Pl$_5$ MFS as suggested by the high concentration of fossil debris, including well preserved ammonites. This level is correlatable with a fos-
siliiferous layer towards the top of the Argiles Grises that is dated as *A. stokesi* in the Query section (Rey and Cubaynes, 1991)

**Pl₆ sequence**

The **Pl₆** sequence is dated as Upper Pliensbachian, *Amaltheus margaritatus* Zone, parts of *A. subnodosus* and *A. gibbosus* Subzones. The characteristics of sequence **Pl₆** are similar to those of sequence **Pl₅**, except that the **Pl₆** sequence boundary is more strongly expressed in the southern France than in the northern U.K. sections.

In the Query area (Fig. 22; Rey and Cubaynes, 1991) the **Pl₆** sequence boundary is marked by a sharp lithological change dated as lower *A. subnodosus* that separates silty shales (Argiles Grises) below from nodular bioclastic limestones (Marnes à taphrosequences de pente) above. In the Causses, it is dated as lowermost *Amaltheus subnodosus* Subzone (Fig. 26; Protogrammoceras depressum “horizon” XXI by Meister, 1986) at the lithological boundary between the alternating mudstones/marlstones below and black shales above (Fig. 25; layers 121/120 of Meister, 1986). In the Subalpine Zones, a regional unconformity related to sequence boundary **Pl₆** marks a phase of block tilting that records one of the Tethyan rifting episodes (Fig. 17). In the thickest sections of the basinal area around Digne, the unconformity is recorded by a thick accumulation of slumped masses of mudstone and marlstone dated as within the *A. subnodosus* Subzone (Graciansky et al., 1993). In the Paris Basin subsurface, the **Pl₅** Sb corresponds to a subtle reversal of the stratal pattern trend within the upper part of the Amaltheid-bearing marls (Figs. 3, 4, 10, 27).

The **Pl₆** lowstand systems tract is well represented in the more subsident areas of southern France that were controlled by the Tethyan rifting. More precise dates are obtained at the Causses section (Fig. 25, Meister, 1986), with calcareous shales assigned to the lower *A. subnodosus* Subzone (part of the Protogrammoceras depressum “horizon”, bed XXI, Fig. 26) and sandwiched between alternating marlstones and mudstone belonging to other systems tracts. This interval can be correlated with coeval marlstone with shell debris in the Quercy area (Cubaynes, 1986; Brunel et al., 1995). In the Digne area of the Subalpine Zones a 150-to 160-m thick accumulation of slumped mudstones and marlstones dated as *Amaltheus subnodosus* Subzone, with no more precision, is tentatively related to the **Pl₆** LST (Graciansky et al., 1993). This LST represents two thirds of the total thickness of the Upper Pliensbachian in the area, which suggests that the main Pliensbachian subsidence episode was related to one of the phases of the Tethyan rifting.

A well dated maximum flooding surface can be found in the Causses, section in a highly fossiliferous limestone (Fig. 25, Rivière sur Tarn, bed 125; Meister, 1986) at the top of a short succession of marlstone and mudstone which represents the **Pl₆** transgressive systems tract. All are assigned to the lower *A. subnodosus* Subzone (Fig. 26, horizon XXII; Meister, 1986).
Age-equivalent strata can be found in Quercy, within nodular fossiliferous marly limestones, which are not as precisely dated within the *Amaltheus margaritatus* Zone as they are in the Causses area (Brunel et al., 1995). The Pl₆ MFS is close to bed S2 of the La Boulbène section (Cubaynes, 1986). The closest age-equivalent in the Yorkshire section can be found in the Cleveland Ironstone at the Avicula Seam (bed 20, Hesselbo and Jenkyns, 1995), which is a nodular fossiliferous oolitic ironstone dated as at the *Amaltheus subnodosus/A. gibbosus* subzonal boundary. In the Paris subsurface Basin the Pl₆ TST is recorded by a well-marked retrogradational section, but with a date no more precise than the *A. margaritatus* Zone (Figs. 3 and 4).

In the Causses section, the Pl₄ highstand systems tract is represented by a marlstone layer with two limestone intercalations dated as around the *A. subnodosus/A. gibbosus* subzonal boundary (Figs. 25 and 26, beds 126–129 of the Rivière sur Tarn section; “horizons” XXIII, Fontaneillesi and XXIV pars, *Reynesoceras ragazzoni* pars; Meister, 1986). These can be correlated with fossiliferous marlstones of the *A. margaritatus* Zone in Quercy (Brunel et al., 1995) and shales dated as lower *A. gibbosus* Subzone immediately overlying the Avicula Seam of the Cleveland Ironstone in the Yorkshire section (Fig. 24; Hesselbo and Jenkyns, 1995).

**Pl₆ Sequence**

The Pl₆ depositional sequence is dated as uppermost *A. margaritatus* Zone and *Pleuroceras spinatum* Zone. It is one of the two sequences that correspond to clearly prograding deposits in both the Paris Basin and Tethyan areas. These constitute the typical forestepping phase of the Late Pliensbachian regressive half-cycle 5 (Fig. 2). Nevertheless, the Pl₆ sequence is marked by high-amplitude, short term transgression dated as *P. spinatum* Zone, *Pleuroceras solare* (= *P. apyrenum*) Subzone, which is very fossiliferous in many places.

The Pl₆ sequence boundary is extremely well defined in the Quercy area (Fig. 24, Loubressac-La Poujade; Brunel et al., 1995). There, a set of cross-beded oolitic/crinoidal limestones intersects calcareous mudstones belonging to the underlying Pl₅ HST. It cannot be placed more precisely than within the lower *A. gibbosus* Subzone. In a similar manner, a change of the stratigraphic pattern and lithology from calcareous shales to thick-bedded crinoidal wackestones to packstones is dated as lower *Amaltheus gibbosus* in the Digne area (Subalps). In the Causses area, there is no macroscopically conspicuous lithological variation within a thick stratum of black shale that is well-dated as *A. gibbosus* Subzone (Meister, 1986) and interpreted as a LST.

In the Paris Basin subsurface, Pl₅ Sb is placed at a well-marked reversal of the stratigraphic trend (Figs. 3, 4, 27, 28). This is dated as upper *A. margaritatus* Zone in several wells. A characteristic surface is recorded within the Cleveland Ironstone of the Yorkshire coast section (Fig. 24, base of bed 21; Hesselbo and Jenkyns, 1995). It is well-marked in the Yorkshire and in the Hebrides.

In the Causses area, the Pl₅ lowstand systems tract is the thickest LST among our studied outcrop sections. There, a 30-m-thick layer of homogeneous black shales represent approximately one half of total upper Pliensbachian section (Fig. 25). It is dated as most part of the *A. gibbosus* Subzone (Fig. 26, horizons XXIV to XXVIII of Meister, 1986). This suggests that the main Pliensbachian subsidence episode was dated as *A. gibbosus* Subzone and is one subzone later than in the Digne area of the Subalps. In the Quercy area, the Pl₅ LST comprises cross-bedded oolitic-crinoidal packstones on the basin margins grading to fossiliferous marlstones in deeper areas (Fig. 24; Brunel et al., 1995). In the Digne basin of the Subalps, the Pl₅ LST is composed of fine-grained crinoidal limestones that form the lower part of a persistent ledge in the area. There is no thick LST in the more northern areas which suggests that the *A. margaritatus* subsidence episodes are mainly linked to Tethyan rift development.

In the Causses area, the Pl₅ transgressive systems tract comprises black shales intercalated with sparse nodular wackestone and is dated as upper *A. gibbosus* and lower *P. apyrenum* Subzones (horizons XXIX to XXXII, Meister, 1986; Figs. 24, 25). These are capped by a fossiliferous accumulation of nodular limestone dated as *Pleuroceras solare* “horizon” (lower middle *P. spinatum* Zone). In the Digne unit of the subalpine area, thinning-upwards, alternating marlstones and fine grained packstones, have yielded rare *Arieticeras algovianum* samples (Entrages section: Graciansky et al., 1993; Barles section: unpublished data by R. Mouterde). In these different areas the

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**Fig. 27**—The prograding “Grès Médiolasiques” of the upper Pliensbachian and the Schistes Carton (Lower Toarcian) from Paris Basin well log interpretation. The siltstones and sandstones, dated as uppermost Pliensbachian, are shown to prograde from the positive area around the Bois Sainte Catherine well. See Figure 10B for location.
TOARCIAN TO BATHONIAN
SW - NE SECTION OF THE PARIS BASIN
Third - Order Framework

LEGEND

- Highstand systems tract
- Transgressive systems tract
- Lowstand systems tract

Unpublished ammonite determination by R. Mouterde (1960-1994)

Fig. 28.—Sequence stratigraphic interpretation of an upper Pliensbachian to Toarcian transect across the Paris Basin Toarcian (France). Modified from Durlet (unpubl. data).
maximum flooding surface is dated as *P. apyrenum* Subzone (*Pleuroceras solare* “horizon”) and is marked by fossil-rich argillaceous or marly layers. These are known also in the Burgundy section of the southern Paris Basin (Mouterde and Tintant, 1980, p. 96). They are documented in the subsurface on the well log record (Figs. 3, 4, 23), but they are not directly dated by ammonites.

In the Yorkshire coast section, the Pecten Seam of the Cleveland Ironstone is an unconformable fossil-rich layer included within shales sandwiched by relatively coarser layers and is interpreted as the Pl7 MFS. It is dated as close to the upper Pliensbachian section by ammonites. The total thickness of the Pl7 sequence reaches 15 m in the Yorkshire section of the Cleveland Basin. In the Paris Basin, Quercy and Tethyan marginal areas, the Pls sequence is generally condensed within coarse, crinoidal bearing, limestones and within metalliferous crusts. This is interpreted as the consequence of the very rapid transgression that follows the episode of extensional tectonics dated at the Pliensbachian/Toarcian boundary. The TST of sequence Pls marks the onset of the transgressive phase of cycle 6.

The Pls sequence boundary is well-documented in the Quercy sections (Brunel et al., 1995), and moreover it is marked by a surface of emergence and karstification (Capdenac section, Cubaynes, 1986). There, it is dated near the *Pleuroceras hawkskerense*/*P. apyrenum* subzonal boundary. In the Subalpine Zones and Paris Basin, the Pls Sb is an erosional surface (Fig. 27), the dating of which being more precise than as *P. spinatum* Zone. In the Paris Basin, ammonites belonging to the *P. hawkskerense* Subzone fortunately have been found in one well (Césarville, Figs. 7, 28), above the well log curve spikes interpreted as the Pls Sb. The *P. hawkskerense* Subzone is known also at the top of the Calcaires à Gryphées Géantès in Burgundy (Mouterde and Tintant, 1980). The *Paltarpites paltus* Subzone is known from Burgundy and Normandy (Mouterde et al., 1980, p. 97, 106). In the Yorkshire section, a surface identifiable as Pls Sb is dated as lowermost *P. hawkskerense* Subzone and marks the base of coarsening-upward silty sands close to the top of the Cleveland Ironstone (Pls LST).

The Pls maximum flooding surface belongs to the *Dactylioceras tenuicostatum* Zone, *D. clevelandicum* Subzone. It is shown by fossil accumulations dated as *D. clevelandicum* Subzone in Burgundy (H. Tintant, personal communication, 1990). The *Paltarpites paltus* and *Dactylioceras clevelandicum* Subzones have never been found in the Subalpine area, possibly from paleoecological reasons (Mouterde, pers. comm., 1992) or without being discovered of the area at this time. A good candidate for the MFS is dated as *Dactylioceras semicelatum* Subzone (*Dactylioceras tenuicostatum* Zone) in the more subsiding part of the Digne unit (Boulard section; Graciesky et al., 1993).

**The Toarcian; Cycle 6. Paris**

The 2nd-order transgressive/regressive cycle 6 comprises most of the upper Liassic series (Fig. 2). This includes a major rate of transgression dated as *Harpoceras falciferum* Subzone and coinciding with organic carbon-rich shales (Schistes Carton and related deposits) of probable world-wide extension (Jenkyns, 1984). The beginning of this cycle is dated as *A. margaritatus* Zone (upper *Alamtheus stokesi* Subzone) in the U.K. (Staithes Sandstones of the Cleveland Basin), but it is around the Pliensbachian/Toarcian boundary in the Paris Basin and within the *Dactylioceras semicelatum* Subzone of the Lower Toarcian strata in the Subalps (Fig. 29) and in the Quercy (Fig. 30).

The Toarcian sedimentation was a time of rapid subsidence and accumulation of thick successions within the major grabens in the Tethyan areas (Fig. 29) and in the Paris Basin. Toarcian strata can be divided into two parts: (a) The backstepping phase (depositional sequences To1 to To2 or To3), leading to the peak transgression and characterized by relative sediment starvation all across the European craton as a consequence of the ex-
tremely wide areal extension of the transgression; (b) the in-
filling phase, when sediments accumulated in the available
space created at the peak transgression.

In the areas bordering the Paris Basin, the Toarcian series is
attenuated in places. In the Jura, the lower Toarcian strata are
partly missing and the Toarcian transgression commenced only
during the *Harpoceras pseudoserpentinum* Subzone of the *Har-
poceras serpentinum* Zone (Contini and Lamaré, 1978). The
situation was different in the northern U.K. and North Sea areas.
For example, in the Cleveland Basin, several tens-of-meters-

thickness shales, named Grey Shales, Jet Rock and Bituminous
Shales, accumulated with high contents of organic carbon dur-

ing the transgressive phase (Hesselbo and Jenkyns, this vol-

tum). This was coeval with the condensed sequence observed
in more southern areas. The characteristic surfaces, such as se-

quence boundaries and maximum flooding surfaces, are not evi-
dent because of the relatively homogeneous lithologies. The
upper part of the section is marked by the development of sand-

stones dated as *Dumortieria levesquei* Zone (sensu anglico) and

is intersected by an erosional surface related to the To7 sequence
boundary (*Pleidellia aalensis* Zone). This is a response to the
thermal doming in the southern North Sea (Ziegler, 1989; Un-
derhill and Partington, 1993) that induced the Mid-Cimmerian
unconformity.

**To2 Sequence**

The To2 sequence is dated as upper *Dactylioceras tenuicosta-
tum* and lower *Harpoceras serpentinum* Zones. It is represen-
ted by a 160 to 180-m-thick section of alternating marlstone and
calcareous mudstone in a locally subsiding part of the Sub-
alpine Digne unit and is dated as *Dactylioceras tenuicosta-
tum Zone* through *Harpoceras strangewaysi Subzone* (*Harpoc-
eras serpentinum Zone*). This situation is exceptional. At most places, se-
quence To2 is extremely condensed; it is amalgamated with
sequence Pl6 on the adjacent outer platform areas (Fig. 29; Gra-
ciansky et al., 1993). In the Quercy area (Fig. 30), the To2 se-
quence corresponds to a less than 4-m-thick black shale in the
more basinal part of the area. The shale grades to crinoidal-
ferruginous oolitic limestone on the shallowest parts. The To2
and Pl6 sequence boundaries are amalgamated with a karstic
surface (Fig. 24) on top of the resistant uppermost Pliensba-
chin Barre à Pectens. They are overlain by shales dated as
Dactylioceras semicelatum Subzone. Fossil accumulations belonging to the Harpoceras strangewaysi Subzone (Eleganti
ceras elegantulum/Harpoceras strangewaysi “horizons” bound-
ary) are a good candidate for the To1 maximum flooding surface
(Bonnet et al., 1994). In Burgundy, the To1 sequence is amal-
gamated with the underlying and overlying sequences within
 discontinuous lenticular layers of organic-rich shales and fer-
ruginous crusts, as shown by their ammonite contents (Mou-
terde and Tintant, 1980, p. 98). In the Jura, the Pl8 and To1
sequences are both missing (Contini et Lamaud, 1978). In Nor-
mandy, the base of carbonaceous shales named as Argiles à
Poissons belong to the Harpoceras strangewaysi Subzone. The
Harpoceras strangewaysi Subzone has also been recovered
close to the base of the Schistes carton section in the Cesarville
well (central Paris Basin; Figs. 7, 28). Thus, the Harpoceras
strangewaysi Subzone is likely to be the date of the To1 maxi-
mum flooding surface (Figs 3, 4). The thinness of the To1 se-
quence in France (except in the rapidly subsiding graben at
Digne; Graciansky et al., 1993) is explained by starvation re-
sulting from the widespread and rapid transgression that fol-
lowed the latest Pliensbachian extensional episode. In conse-
quence, the sequence stratigraphic interpretation of this part of
the Toarcian series lacks of comparison points at distance, and
correlations remain ambiguous.

**To2** Sequence

The To2 sequence contains a widespread 1st-order (and 2nd-
order) maximum flooding surface (Figs. 2, 4) dated as Harpo-
ceras falciferum Subzone and located within organic-rich shales
that constitute one of the major source rocks in western Europe
(Jenkyns, 1984, 1988). The To1 sequence is represented mainly
by hightand-backstepping and foerstepping deposits.

In the subsiding parts of the Subalpine Zones (Fig. 29), the
To2 sequence boundary is expressed by a lithological change
separating carbonaceous shales dated as Harpoceras strange-
waysi below, from similar rocks above, but including nodular
wackestones dated as upper Harpoceras serpentinum Zone. On
the ramps leading to the edge of tilted blocks, the To2 sequence
boundary is amalgamated with the To1 sequence boundary and
a thin veneer of Harpoceras falciferum -bearing shales directly
overlies crinoidal limestones of the Pleuroceras spinatum Zone,
demonstrating the transgressive character of this interval Gra-
ciansky et al., 1993). In the Quercy, Rey and Cubaynes (1991),
Cubaynes et al. (1990) have shown a sharp lithological change
from the Harpoceras strangewaysi organic-rich shales to Har-
poceras pseudoserpentinum marls (Fig. 22). Subsurface in the
Paris Basin, a peak on the well log record shows the Sb To 2.I t
postdates the Harpoceras strangewaysi Subzone and predates
the Harpoceras falciferum Subzone. It is a good marker
throughout most of the Paris Basin (Figs. 3, 4).

The To2 lowstand systems tract is hard to find as a probable
consequence of starvation and rapid transgression. An excep-
tion is in the Quercy area, where less than 6 m of alternating
marlstone and mudstone are dated as Harpoceras serpentinum
Subzone and Harpoceras falciferum Zone (pars) is present in
the more subsiding part and is missing on the adjacent platform.
This is interpreted a local shelf margin wedge (Fig. 30; Rey
and Cubaynes, 1991; Cubaynes et al., 1990). The overlying
transgressive surface is marked by an accumulation of ammo-
nites dated as Harpoceras pseudoserpentinum Subzone.

The To2, transgressive systems tract and highstand systems
tract are mainly alternating shales and calcareous mudstone in
the Quercy (Bonnet et al., 1994). The highstand systems tract
comprises a set of three transgressive/regressive parasequences

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**Fig. 30.—Sequence stratigraphic interpretation of the lower and middle Toarcian succession in the Quercy area.**
dated as *Harpoceras falciferum*, *Hildoceras sublaevisoni* and *H. lusitanicum* (pars) Subzones respectively. But their organic matter content is low in contrast with the underlying sequence. In the Paris, Jura (Contini and Lamaud, 1978) and Subalpine Basins, deposits belonging to the To3 sequence are part of the organic-rich Schistes Carton wherever paleowater-depth allowed. These shales grade laterally into calcareous and/or ferruginous olites or phosphatic layers towards the submarine highs, dated as upper *Harpoceras serpentinum* and lower *Hildoceras bifrons* Zones, as illustrated from outcrop in Normandy (Rioult, 1980). The To3 TST is well dated as *Harpoceras falciferum* Subzone in the Cesarville well (Fig. 28). Equivalent of the Schistes Carton are the Bituminous shales of the Cleveland Basin and related horizons in the northern U.K. (Hesselbo and Jenkyns, this volume).

**To3 Sequence**

The peak transgression of the T/R cycle 6 (Fig. 2) is classically dated as *Harpoceras serpentinum* Zone (Jenkyns, 1984, 1988). Nevertheless, in both the Quercy and Subalpine areas (Figs. 22, 28, 29, 30), the To3 sequence unconformably overlies the earlier ones. The continental encroachment was maximum during the deposition of the To3 sequence and dated as *Hildoceras lusitanicum* “horizon” of the *H. commune* Subzone (lower *Hildoceras bifrons* Zone). Therefore, the section dated as upper *Hildoceras sublaevisoni* and *H. lusitanicum* Subzones is considered as critical one in spite of its short duration.

In both Quercy and Subalpine areas, a slightly erosional unconformity is related to both the sequence boundary and transgressive surface of To3. This rests directly on shales belonging to sequence To2 in Quercy (Figeac area, Figs. 30, 31), and it overlies directly the *Pleuroceras spinatum* crinoidal limestones of the outer Provence platform in the Digne area (Fig. 29). In the southern Subalps, the To3 maximum flooding surface is recorded by a bed that comprises nodular limestones with amonite accumulations within stromatolites and nubecularids (“bouses de vache”). The whole sequence is generally less than 2 m thick, including in the most subsiding areas, as a consequence of starvation related to the areal extension of the transgression. In the Paris Basin, the interval of sequence To3 comprises mainly calcareous shales with some calcareous mudstone beds that are located mainly in less subsiding areas (Figs. 3, 4, 28) and in marginal areas such as Normandy (Rioult, 1980).

**To4 and To4* Sequences: the Bifrons Shales**

The To4 sequence was deposited just after the time of a major peak transgression (i.e., during the time when a maximum accommodation space was available). The consequence is the existence of thick lowstand deposits with aggregating stacking patterns that are a main characteristic of sequence To4 in the Paris and Aquitaine basins. However, in parts of the Subalpine Digne thrust sheet, sequence To4 is missing or extremely condensed (Fig. 29; Graciansky et al., 1993). This results probably from starvation linked to the coeval episode of the Tethyan rifting.

In the Paris Basin, the To4 highstand systems tract in Aquitaine is divided into two parts by an erosional surface identified as sequence boundary To4* which may define the basal contact of a local To4* depositional sequence (Fig. 22). However, this To4 depositional sequence (numbered To4* in Bonnet et al., 1992) has no clear equivalent in neither the Paris Basin or the Subalps. We have not retained it on our chart.

The To4 sequence boundary is recorded in the Quercy by an erosional surface between a limestone bed dated as *Hildoceras lusitanicum* (To4 MFS) below and shales dated as *H. bifrons* Subzone above. To4 Sb is more apparent in the less subsiding parts of the area (Fig. 30; e.g., Capdenac section, Rey and Cubaynes, 1991; Cubaynes et al., 1990) where the *Hildoceras bifrons* shales rest on a Lower Toarcian attenuated sequence and comprising calcareous and ferruginous-oolite-bearing limestones. In the Subalpine area, To4 To5, To6 and To7 sequence boundaries are frequently amalgamated (Fig. 29; Graciansky et al., 1994).

In the Paris Basin, the To4 lowstand systems tract corresponds to an accumulation of *Hildoceras bifrons* shales that reach 100-m total thickness in the more subsiding areas. This represents half of the whole Toarcian section. In Quercy (Fig. 30; Bonnet et al., 1994) and along the Atlantic coast in Vendée (Gabilly, 1976), the To4 LST comprises calcareous shales with several calcareous mudstone interbeds dated as *Hildoceras bifrons* and *H. semipolitum* Subzones and deposited in the more subsiding areas. The depositional regime was mainly aggradation during this episode. The To4 transgressive systems tract is well defined in Lorraine (Hanzo, 1980) by a marker bed of marlstone that contains phosphate nodules dated as *Haugia variabilis* Zone and named Couches à *crassum*. It corresponds probably to the “Argiles à
Haugia” in Burgundy (Tintant, 1980). In the Paris Basin subsurface, a short retrograding section on top of the prograding Hildoceras bifrons marls has been dated as Haugia variabilis Zone in the Beaumont 101 well (Figs. 10, 28). In the Quercy area, the To6 TST is represented by a thin but widespread and continuous layer that is capped by a nodular bioturbated bed. This includes local accumulations of ammonites dated as Haugia variabilis/Haugia illustris subzonal boundary and is interpreted as the To6 maximum flooding surface. A similar nodular marker bed is known also in the Vendée sections, including the more condensed sections deposited on submarine highs (Gabilly, 1976).

In the central Paris Basin, the To6 highstand systems tract comprises aggrading to slightly prograding marlstones belonging to parts of the Haugia variabilis and Grammoceras thouarcense Zones (i.e., the Haugia illustris, Pseudogrammoceras bingmanni, and Grammoceras thouarcense Subzones) as documented in the Beaumont 101 well (Figs. 10, 28). The lateral equivalent is the lower part of the “Marnes à Astarte voltzi” that outcrop in Lorraine. These marls are characterized by the presence of calcareous septarian nodules and by increasing-upward contents of detrital quartz and mica (Hanzo in Mouterde et al., 1980).

In Vendée (Gabilly, 1976) and in the Quercy area (Bonnet et al., 1994), the interval dated as Haugia illustris and Pseudogrammoceras bingmanni Subzones is divided into two parts (Fig. 22). The lower one is dated as Haugia illustris Subzone sensu stricto. The upper one, dated as Bingmanni and lowermost Grammoceras thouarcense Subzones, is interpreted as a short-term depositional sequence named as Toa, by Bonnet et al. (1994) or To6 here (see above). The corresponding sequence boundary is an erosional and/or a hiatal surface in both the Quercy and proximal areas of the Northern Aquitaine (Vendée, Poitou). It is marked by a change in the stratral pattern in the Vendée basinal section (Gabilly, 1976). The flooding surface is not well documented.

**To5 Sequence**

In the Paris Basin, the To5 sequence is part of the continuation of the infilling process that commenced with the Harpoceras falciferum. MFS (Figs. 3, 4). It comprises the upper half of the “Marnes à Astarte voltzi” in Lorraine. Total thickness is less than 50 m in the Paris Basin and less than 5 m in Quercy. Detrital quartz and/or bioclastic calcareous material content increases upward in both areas (Fig. 31).

The To5 sequence is particularly well represented in the Penne section of Quercy (Cubaynes, 1986; Cubaynes et al., 1990). The sequence boundary is marked by an erosion surface dated as intra-Grammoceras thouarcense Subzone, which truncates the top of the Pseudogrammoceras marls. The lowstand deposits are represented by cross-bedded quartz-bearing bioclastic limestones dated as Grammoceras thouarcense Subzone. The transgressive systems tract comprises dark-colored marls that include sparse ferruginous oolites grading upward to bioclastic and oolitic limestones. These are dated as upper Esericeras fascigerum and lower Pseudogrammoceras fallaciosum Subzones. The thinness of this TST (less than 1–2 m thick in the basinal areas), together with the existence of fossil accumulations, indicate starvation during this time interval.

The To5 highstand systems tract shows a few layers of oolitic and bioclastic limestone, in total less than 1 m thick. These are dated as upper Pseudogrammoceras fallaciosum and lowermost Hammatoceras insigne Subzones. The To5 HST is truncated by an erosional surface interpreted as the To5 sequence boundary in the Penne section of Quercy (Cubaynes, 1986; Cubaynes et al., 1990). Similar observations are made in Vendée (Gabilly, 1976). One of the characteristics of the To5 transgression is its relatively high amplitude in Aquitaine.

**To6 Sequence**

The To6 sequence is characterized by the gradation of shales (lower To6 LST) to interbedded shales and bioclastic limestones (To5 highstand deposits) in the Quercy (Fig. 22). In the Paris Basin, sandy silts (named Grès supra-liasiques in Lorraine) constitute the To6 sequence and coarsen progressively upwards. However, their-age equivalents are truncated by an erosional surface dated as Aalenian in Normandy.

In the subsiding part of the Digne area (Subalpine Zone), sedimentation resumed with a thick section of calcareous silts and silty shales after a phase of starvation covering the Haugia variabilis, Grammoceras thouarcense and Hammatoceras insigne Zones (i.e., the missing To5 and To6 sequences). The presence of sequence To6 in the area is documented by rare ammonites dated as Dumortieria pseudoradiosa Zone (Fig. 29).

The To6 sequence boundary is well expressed in Lorraine by a regional erosional surface dated as the Grammoceras thouarcense/Hammatoceras insigne zonal boundary (Hanzo, 1980). It is post-dated by the Dumortieria levesquei Subzone of the Dumortieria pseudoradiosa Zone in the Tousson well (Fig. 28). The To6 Sb is more clearly observed in the Quercy (Cubaynes, 1986; Cubaynes et al., 1990) where it is marked by a hardground surface that intersects a 0.10- to 0.20-m-thick, condensed layer of biosparite with ferruginous ooliths. This is dated as Hammatoceras insigne Subzone. The overlying To6 lowstand systems tract comprises calcareous silts that include detrital mica and quartz together with crinoid, urchin debris and mollusks. It is dated as Hammatoceras insigne, Gruneria groeneri and Dumortieria levesquei Subzones.

The To6 transgressive systems tract comprises an interval of thinning-upward alternations of ammonite-rich shales and calcareous mudstones dated as Dumortieria pseudoradiosa Subzone. The To6 highstand systems tract corresponds to shallowing-upward beds with accumulated Gryphaea and other bivalves, dated as lower Pleidella aalensis Zone.

**To7 Sequence**

The To7 sequence corresponds to a key phase in western European Liassic development in the sense that it coincides roughly to the so-called mid-Cimmerian unconformity (Ziegler, 1988, 1989). The To7 sequence boundary is a widespread erosional unconformable surface that underlies the Brent Group in the North Sea (Jacquin et al., unpublished data, 1995). In the subsiding Digne area of the Subalpine domain, the To7 sequence marks the end of the infilling phase in the sense of Jacquin et al. (1992). Silty shales, up to several hundreds meters thick, were deposited both in the basinal areas and on the distal shelf of the adjacent platforms at the end of a period of starvation.
that continued for most early and middle Toarcian times (Fig. 29; Graciansky et al., 1993).

The To7 sequence boundary is marked in the Digne area (Subalps) by an unconformable erosional surface with a conspicuous horizon of slumped balls of calcareous mudstones included within the dark shales of the basinal parts. It is dated as lower Pleidellia aalenensis Zone (Graciansky et al., 1993). It is well marked in the more subsiding part of Quercy by a hardground surface that is particularly well dated as intra-Celtica (upper Pleidellia mactra Subzone, Lower Pleidellia aalenensis Zone).

The To7 lowstand systems tract in the subsiding part of Quercy and Causses (Causses de Severac and Larzac) corresponds to the “Calcaires à Zoophycos,” which are offshore peloidal calcareous mudstones. These may reach 100 m in thickness (Peybernes and Pélissé, 1985). In Quercy, the To7 sequence is represented only by an interval, 2- to 3-m-thick, of alternating calcareous mudstones and marlstones dated as uppermost Pleidellia mactra Subzone.

The most spectacular To7, LST is found in the Digne area (Fig. 29), where it reaches several hundred meters in thickness. The rocks are silty shales with higher quartz contents and coarser grains in the distal turbiditic layers that in the underlying units. These features are probably the consequence of rapid extension, block tilting and subsequent subsidence, that (a) allowed the accumulation of huge masses of sediments, (b) explains local erosion together with the renewal of detrital production and (c) records, at distance, the mid-Cimmerian arching and erosion of the North Sea area.

In the Subalpine basinal context, the transgressive and highstand systems tracks of sequence To7 are found within particularly fine grained shales dated as Leioeceras opalinum Zone. The best candidate for the maximum flooding surface is dated as close to the Opalinum/Comptum boundary. It coincides with the change in stratigraphic pattern in the external platform area (Caloo, 1970).

The interpretation of the To7, TST and HST is not yet as clear in the Aquitaine border areas. In Vendée, the To7 sequence is condensed to less than 0.5 m of nodular marly limestones (Gabilly, 1976). In the Quercy and Causses areas, Rey (pers. comm., 1995) indicate a MFS dated as intra-P. fluitans Subzone (upper Pleidellia aalenensis Zone). However, there is no clear equivalent known in other areas, so that the question remains open.

In the subsurface Paris Basin (Figs. 10, 28) the sequence To7 is part of an overall regressive trend that started in the lowermost Hildoceras bifrons Zone. It is represented in places by 15–20-m-thick alternations of calcareous siltstone and mudstone. Characteristic peaks on the well log curves dated as Leioeceras opalinum Zone in the Beaumont 101 well (Fig. 28), may represent the maximum flooding surface. But in many wells and in the outcrops in bordering areas (Mouterde et al., 1980), the whole To7 sequences have been eroded along a hiatal surface that corresponds to most of the Aalenian. Local accumulations of ferruginous oolites are dated as Dumortieria pseudoradiosa and Pleidellia aalenensis Zones in Lorraine and Normandy (Hanzo, 1980; Rioult et al., 1991).

AALENIAN: MAXIMUM REGRESSION OF CYCLE 6

The Aalenian was a time of tectonic uplift and erosion in northern Europe, particularly in the North Sea (Ziegler, 1988, 1989; Underhill and Partington, 1993). In the Paris Basin, the Aalenian section is attenuated in many places (Fig. 28). The only deposits that are preserved correspond to the three Aalenian maximum flooding surfaces, a consequence of the high-amplitude of the relative sea-level oscillations at this time. Lithologies vary widely; there are thin patches of conglomerates, sands and silty sands that are mainly developed in the western Paris Basin. Scattered bioclastic limestones and ferruginous oolites are more frequent than detrital deposits in the eastern Paris Basin. An exception is the accumulation of iron ore in Lorraine, which reaches 50 m in thickness (Briey-Ottange area; J. Thierry et al., 1980). Thick carbonate platform deposits developed during Aalenian times in the Bresse and the Jura (i.e., at the transition from the Paris Basin platform to the Tethyan marginal areas). In contrast the Aalenian was a time of rapid and thick sedimentary accumulation in the more basinal parts of the nascent Tethyan margin (Graciansky et al., 1993).

The infilling phase of such subsiding areas that commenced with the Hammatoceras insigne or Dumortieria pseudoradiosa Zones (upper Toarcian) graded into the forereefing phase during the early Aalenian times and ended with the maximum regression (end T/R cycle 6: see Fig. 2) dated as late Aalenian (Graphoceras concavum Zone). Comparison of well-dated sections on the eastern border of the Massif Central between Burgundy (eastern Morvan) and the Lyon area in the Saone Valley shows a similar pattern (Mouterde, 1953).

The Subalpine area provides the best outcrops for calibration of the Aalenian depositional sequences (Fig. 29; Graciansky et al., 1993). Deposits are homogenous alternating calcareous mudstones and marlstones well-dated by abundant ammonites (Caloo, 1970). The maximum flooding surfaces of sequences To7, Aa1, and Aa2 are dated as Leioeceras opalinum/Costileioceras comptum, Ludwiga haugi/L. murchisonae and Graphoceras concavum/G. limitatum subzonal boundaries, respectively, as shown in our chart. They are recorded by more argillaceous intervals that display more abundant accumulations of ammonites than the underlying and overlying ones. The more typical feature of the Aalenian in the area is the 250-m-thick Aa2 sequence that pinches out on the adjacent outer Provence platform (Graciansky et al., 1982; Fig. 29). In consequence, the top surface of the Aa2, LST is assumed to date the major regression of this time interval. It is overlain by the Bajocian aggrading depositional sequences that marks the onset of the transgressive phase (T7) of the next 2nd-order facies cycle. The data concerning the Aalenian in the Paris basin, Jura and Bresse areas have been briefly explained in the companion paper on the 2nd-order Ligurian facies cycle (Graciansky et al., this volume).

CONCLUSION

During Liassic times, active rift systems were located in the Arctic, North Atlantic, central and western Mediterranean oceans, western and central Europe. However, the rift development, including the rate of crustal extensions and the rate of subsidence, differed from one structural province to the other (Ziegler, 1988). Hence the duration and the characteristics of individual 2nd-order transgressive/regressive cycles differ also depending on the area, which is in contrast with the relative uniformity of the Rhaetian deposits over large areas.
In the nascent Tethyan margin (Subalpine Zones), one single transgressive/regressive facies cycle (4) developed during latest Norian to early Pliensbachian times (i.e., in the Tragophylloceras ibex Zone, which represents a major regression throughout southwestern Europe). The corresponding peak-transgression is dated as being close to the Lower/Upper Sinemurian boundary (Caesinites turneri Zone). In northern Europe, facies cycle 4 is subdivided into two parts. A significant regressive episode occurred in the Paris Basin and in the Viking graben (Jacquin et al., unpubl. data) during the Early Sinemurian (Amniceras semi- costatum Zone) which separates transgressive/regressive facies cycle 4a below from 4b above. This boundary is dated as close to the Lower/Upper Sinemurian boundary by Hesselbo and Jenkyns (this volume).

The transgressive/regressive facies cycle 5 is dated as Pliensbachian (Domerian) in the Tethyan marginal area and in the Paris Basin. The lower maximum regression is dated as intra-Tragophylloceras ibex Zone (Lower Pliensbachian). The upper maximum regression is uppermost Pliensbachian (P. hawskerense Subzone), except in eastern Aquitaine where it is dated as upper Dactylioceras tenuicostatum Zone. The age of the corresponding peak transgression is intra-Amaltheus stokesi Subzone in the Tethyan areas butProdactylioceras davoii Zone (Aegoceras capricornus Subzone) in the Paris Basin.

The entire cycle 5 is older in the U.K. than elsewhere. The lower maximum regression is dated as around the Sinemurian/Pliensbachian boundary (Palaechioceras aplanatum Subzone). The upper maximum regression is dated as intra-Amaltheus stokesi Subzone and is marked by the Staithes Sandstone in the Yorkshire section. The peak transgression is recorded by organic—rich shales dated as early Uptonia jamesoni Zone (Phricodoceras taylori Subzone) and is close to a maximum regression in the Tethyan realm (Hesselbo and Jenkyns, this volume). In contrast, in the Viking graben of the North Sea, Jacquin et al. (unpubl. data) consider that the Cook Sand (pars) and Burton Shale Formations, dated as latest Pliensbachian, constitute the local cycle 5 with boundaries coincident with the stage boundaries. There, the peak transgression is dated as Amaltheus stokesi Subzone as in the Tethyan area.

From northern to southern Europe, cycle 6 is more constant in age than the preceding ones. Noticeably the upper maximum regression is dated as Graphoceras concavum Zone (upper Aalenian). It is induced by the regional Mid-Cimmerian uplift in northern Europe (Underhill and Partington, 1993) and coeval rapid subsidence linked to the Tethyan rifting in the Subalpine areas (Graciansky et al., 1993). The major peak transgression, corresponding to the organic-carbon-rich Schistes Carton and related lithologies, is dated as around the Lower/Middle Toarcian boundary (Jenkyns, 1988) and generally is of the Harpoceras falciferum Subzone (depositional sequence T0). But in the Tethyan areas, the major peak transgression belongs to the Hildoceras lusitanicum Subzone (lower Hildoceras bifrons Zone, depositional sequence T0).

Comparing the ages of the 2nd-order peak transgression or maximum regression from basin to basin along our transect gives the impression of diachronity for these boundaries. For example, the early Liassic peak transgression appears to have begun first in Hettangian time in the North Sea and the Paris Basin, to have reached the U.K. in the Early Sinemurian time and then southeastern France at the Early/Late Sinemurian boundary. In a similar manner, the late Pliensbachian maximum regression (top cycle 5) which began at the earliest late Pliensbachian time in the U.K. seems to have reached the uppermost Pliensbachian time in the other areas and even later during Lower Toarcian time in the Aquitaine.

In fact, every key boundary is present everywhere and is subdivided into two or more parts. A significant regressive episode occurred in the Paris Basin and in the Viking graben (Jacquin et al., unpubl. data) during the Early Sinemurian (Amniceras semi-costatum Zone) which separates transgressive/regressive facies cycle 4a below from 4b above. This boundary is dated as close to the Lower/Upper Sinemurian boundary by Hesselbo and Jenkyns (this volume). Noticeably the upper maximum regression is dated as being close to the Lower/Upper Sinemurian boundary by Hesselbo and Jenkyns (this volume). In contrast, in the Viking graben of the North Sea, Jacquin et al. (unpubl. data) consider that the Cook Sand (pars) and Burton Shale Formations, dated as latest Pliensbachian, constitute the local cycle 5 with boundaries coincident with the stage boundaries. There, the peak transgression is dated as Amaltheus stokesi Subzone as in the Tethyan area.

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action of processes affecting other basins further away with the associated eustatic changes.

REFERENCES


SEA-LEVEL CHANGES AND EARLY RIFTING OF A EUROPEAN TETHYAN MARGIN IN THE WESTERN ALPS AND SOUTHEASTERN FRANCE

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ABSTRACT: Detailed stratigraphic and sedimentological analyses on Late Triassic-early Liassic sections in the internal and external western Alps, the southeast French basin and the southeastern part of the French Massif Central provides the opportunity to study the distribution of sedimentary bodies which were deposited during a long-term transgressive trend at the beginning of a continental encroachment cycle that extended onto the weakly differentiated Late Triassic European platform. Four short-term sequence boundaries can be identified: Late Norian or early Rhaetian, latest Rhaetian, early Hettangian and late Hettangian. They bound three sequences (S_R, S_L and S_L2) whose duration is compatible with Vail’s (1992) “sequence cycles”. In some places, the first sequence (S_R) can be divided into two minor sequences (S_R1 and S_R2), which define an additional, upper Rhaetian sequence boundary. The second sequence (S_L) is also composed locally (subalpine domain) of two minor cycles. Evidence for rapid relative sea-level fall is lacking because of overall subsidence and moderate sediment supply. The maximum flooding intervals of sequences S_R and S_L2 are of Rhaetian age, but they cannot be dated more accurately due to restricted platform facies. Two other maximum flooding intervals are dated with ammonites from the early Hettangian (upper planorbis subzone) and the middle Hettangian (upper portlocki or lower laugeux subzone), respectively. The deposition of these sequences is coeval with an increase in extensional tectonic activity whose culmination, during the middle and late Hettangian time, represents the earliest synrift event of Ligurian Tethys rifting in the study area. Differential subsidence, synsedimentary faults and paleostress indicators show that a new tectonic regime was developed, starting at the Triassic/Jurassic boundary. This can be correlated to the plate-tectonic reorganization observed, i.e., along the central Atlantic margins and may account for the long-term transgressive trend. As in many other Tethyan or Atlantic areas, this event is marked locally by short-lived intracontinental rift volcanism. The short-term sequences observed indicate some differences to Haq et al.’s (1987) chart, but the transgressive events are known in many areas inside and outside the European realm. The corresponding changes in accommodation space thus cannot be ascribed to local tectonics only. They may have been controlled by eustasy, but the widespread extent and the synchronism of some short events (volcanism around the Triassic/Jurassic boundary; middle-late Hettangian time breakup and drowning) suggest that subsidence, possibly controlled by fluctuations of intraplate stresses, may also have forced the short-term signal.

INTRODUCTION

At the Triassic/Jurassic boundary, the record in western Europe generally shows an abrupt change in the type and distribution of facies. The western Alps and their foreland permit tracing the late Triassic and early Liassic depositional sequences over long distances, since the first fully marine Mesozoic beds were laid down on a low-relief topography, in contrast to the southern or eastern Alpine realm, which was much more affected by the Late Triassic rifting events. About 50 upper Norian to lower Sinemurian field sections were analyzed (Fig. 1). Sequences and unconformities are identified using vertical transgressive/regressive trends, their lateral evolution and facies associations. The sea-level signals deduced from the sequence stratigraphic analysis is compared to European and other areas in order to determine whether they have been significantly affected by local synsedimentary extensional tectonics (i.e., lower Liassic).

STRATIGRAPHIC AND PALEOTECTONIC SETTING

All regions of southeastern France and western Alps were affected during late Triassic/early Jurassic time by a gradual change in paleoenvironments and subsidence patterns. This is due both to the beginning of the widespread Jurassic transgression and to the initiation of the Tethyan rifting, which is a special feature in the studied area. The combination of these two factors caused marine encroachment onto perialpine platforms near the Triassic/Jurassic boundary. This was followed by the breakup of most of these areas into uplifted blocks and downfaulted basins. The evolution of each part of the study area near the Triassic/Jurassic boundary is briefly outlined below.

Causses Area.—

The flat, emergent Causses area was rapidly flooded during early Liassic times, and southeastward-thickening epicontinental dolomitic facies were deposited. Northward retrogradation is shown by diachronism of the basal, fluviatile transgressive sandstones (Grignac and Taugourdeau-Lantz, 1982). Post-Hettangian extensional reactivation of Hercynian structures, related to Tethyan rifting, split the area into emergent shoals (Rouergue, Cévennes; Trumpy, 1983) and a central subsiding carbonate platform connected to the southeastern French basin.

Ardèche-Cévennes Border of the Massif Central.—

During middle and late Triassic deposition, this area was the western edge of the strongly subsiding dolomitic/evaporitic southeastern French basin (Baudrimont and Dubois, 1977). The inherited northeast-southwest “Cévenol” fault trend was active during Carnian (Bergerat and Martin, 1993) and during Liassic times (Elmi, 1985; Giott et al., 1991; Roure et al., 1992). Subsidence ceased before the end of Triassic Period, and there was a change in dynamics around the Triassic/Jurassic boundary (Colongo et al., 1979). Increased subsidence is recorded in the Hettangian hemipelagic series (Ardèche: Brunet, 1985; Giott et al., 1991; Roure et al., 1992), together with complex block faulting along two perpendicular directions (Martin, 1984).

Alpes-Maritimes and Provence.—

A very flat late Triassic sabkha was flooded by a thin widespread Rhaetian lagoonal series. The early Liassic dolomitic deposits are coeval with east-west and northwest-southeast early-rift structures and moderate vertical movements, which caused extensive stratigraphic gaps from lowermost Liassic up to Dogger time in the southern part of the area (Dardeau, 1983). This platform was separated from the central subsiding southeast French basin along northeast-southwest major faults (i.e., the Durance fault; Monleau, 1986) during the climax of rifting (middle-late Liassic). In the northern part of the area, thin platform carbonates are connected to the Subalpine early Liassic basins.
Southern Subalpine Domains.—

The restricted, evaporitic and terrigenous Late Triassic series thicken towards the central southeast basin depocenter (Courel et al., 1984). The thin transgressive Rhaetian series becomes more reduced and sandy to the north (south of the Dauphine domain) and to the east (cover of the Argentera massif). The disappearance of Rhaetian facies to the north coincides with the occurrence of alkaline volcanism. During early Liassic time, platform and hemipelagic basin environments were established in the autochtonous series and the Digne nappe, respectively. Differential subsidence with activation of east-west to north-west-southeast structural trends occurred after early Hettangian time.

Dauphine and Pelvoux Massif.—

The very thin and ubiquitous Triassic dolomitic series contains only one basal fossiliferous layer of late Ladinian to early Carnian age (Baron, 1981). The top of this wedge is assigned to Norian age. It is generally capped by thin splitized alkaline basaltic flows emplaced near sea level. The slopes of the small paleovolcanoes were flooded by neritic sedimentation during early Hettangian time. During middle and late Hettangian time, facies and sediment thicknesses remained weakly differentiated. After middle Hettangian time, waterdepth increased and the Dauphine area broke up into asymmetrical hemipelagic basins and swells in response to moderate early-rift extensional faulting. The main rifting phase occurred during late Liassic (Lemoine et al., 1986).

Subbrianconnais Domain.—

This internal Alpine area, whose remnants are presently found in a pile of cover thrust sheets along the Pennine front, was located between the Dauphinois and the Brianconnais area. The Triassic series exhibit proximal, terrigenous-rich sabkha environments with evaporites. The top of the Norian wedge is regressive and displays either Keuper or Hauptdolomit facies (probably interfingered). Both facies are capped by a Rhaetian-age widespread thin peritidal series, which is markedly transgressive at its base. This series grades upward into a transgressive, more or less condensed, early to middle Liassic sediments which recorded moderate subsidence and small-scale block faulting.

Brianconnais Domain.—

The thick Triassic section (upto 1 km) shows typical Tethyan marine facies, especially in the middle Triassic strata (Megard-Galli and Baud, 1977). It ends with the regressive Norian Hautdolomit Formation which is overlain by a thick, transgressive Rhaetian series of inner platform to tidal-flat environments. The Liassic history of this area and the age of its emergence are still a matter of discussion since most of the synrift
Liassic rocks were later eroded (Faure and Mégard-Galli, 1988; Tricart et al., 1988). However, the remnants of the lower Liassic beds found in the Peyre-Haute nappe near Briançon are markedly transgressive and are very similar to those of the nearby Subbrianconnais domain. They do not contain any detrital material coming from other parts of the Briançonnais. This suggests that at least this part of the Briançonnais domain was not uplifted during early rifting. Uplift may have begun during middle Liassic time since coeval deposits are not known in the Briançonnais stratigraphic series or in the pebbles that accumulated in karstic caves during emergence.

**Piémontais.**

The upper Triassic Piémontais series is thicker than the Briançonnais one. Liassic strata indicate a rapid drowning with evidence of extensional faulting (Dumont et al., 1984). The differentiation between the uplifted Briançonnais and the strongly subsiding Piémontais area is marked by clastic input in the latter area from middle Liassic time onward (Debelmas, 1987; Dumont, 1987). This feature can be interpreted as a rift shoulder effect due to the increase of synrift extension in the western Alpine realm.

**FACIES AND ENVIRONMENTS**

**Late Triassic Time**

Only piedmont, playa, sabkha, tidal-flats and shallow platform facies were deposited on the very flat European landmass after post-Hercynian penneplanation. **Piedmont facies** are found on the southeastern border of Massif Central (Ardèche-Cévennes), with mixed continental and tidal influences indicated by interfingerings between fluvialite sandstones and storm deposits (Perrissol, 1986). Similar clastic facies are found in Rhaetian strata of the northern Subalpine area, which suggests that the neighboring southern Dauphiné area was slightly uplifted. The mostly dolomitic **playa and sabkha facies** of the thick late Triassic series in the internal and external Alps indicate keep-up sedimentation and an arid climate (desiccation; extensive development of evaporites in the strongly subsiding central southeast French basin). Only during Carnian time did humid climate occur, as documented in the Alps and in western Europe (Megard-Galli and Baud, 1977; Simms and Ruffell, 1989). **Tidal-flat facies** are commonly found in Alpine Rhaetian formations. The high clay content is a common feature of the lower part of all these formations and is also found outside the Western Tethyan realm (Dumont, 1992). It may be related to a climatic change in late Norian time (Fabricius, 1970; Jadoul et al., 1992). The “Bone-beds”, showing evidence of condensation, are mostly found in the lower part of the Rhaetian series, whereas the upper part contains keep-up tidal-flat carbonate facies. **Shallow platform facies** are widespread in Rhaetian strata of the Internal Alps (Piémontais, Briançonnais). Evidence of restriction, probably due to the flatness of the huge European Rhaetian platform, is given by the anoxic character of lower Rhaetian shales and limestones. Environments are more oxygenated in the upper, carbonate-rich Rhaetian series, as shown by the widespread development of patch-reefs in a lagoonal setting.

**Early Liassic Time**

Fluvialite and tidal-flat sedimentation shifted landwards, causing a decrease in clastic input into the southeast French basin series. **Shorelines** are found in northern and western Causses, with basal conglomerates and sandstones resting upon a weathered Triassic erosional surface (Schmitt and Simon-Coinçon, 1985). They are fringed by coastal swamps to evaporitic/dolomitic facies containing stratified mineral precipitations resulting from continental leaching under a warm seasonal climate (Fuchs, 1978; Macquard, 1978). Dolomitic **tidal-flat sedimentation** is extensively developed in Causses and Provence, containing either carbonates (algal mud-flats with desiccation, storm layers and dinosaur footprints; Demathieu and Sciau, 1992) or mixed sedimentation (alternating dolomicticrites and continental greenish clays with paleosols). Very thick evaporites were deposited during late Hettangian time in Languedoc and in the Southern Rhône Valley (Debrand-Passard et al., 1984; Courel et al., 1984), similarly as in the Aquitaine basin (Curnelle, 1983). Middle Hettangian **karstification** occurred in the southern Causses and Cévennes (Melas, 1982; Arrondeau et al., 1985), indicating a more humid climate than late Triassic time. **Platform facies** are well represented in Hettangian formations of the Alps and southeast French basin. Widespread cross-stratified oolitic or peloidal carbonate sands mark the basal Hettangian transgression. This thin massive transgressive unit (5 to 15 m) was deposited before the main phase of extensional tectonics, and it is known in the Subalpine areas and Provence (“transition unit”; Dumont, 1988), in Ardèche (“complexe carbonaté de base”; Elmi and Mouterde, 1965) and in Cévennes (“unit 4”; Perrissol, 1986). **Lagoonal facies** with local patch-reef development were deposited at the transition between the Causses/Provence dolomitic platforms and hemipelagic basins, especially during early Hettangian time. The distribution and the more or less restricted character of the fauna of the internal platform and perireefal facies are partly controlled by block-faulting and differential subsidence (Martin, 1984; Vitry, 1982). **Rhythic (climatically controlled?) sedimentation of lagoonal lime-mudstones and dolomicticrites without any evidence of emergence occurred in the upper part of Hettangian formations of the internal Alps (Briançonnais and Subbrianconnais) and in the autochthonous formations of the northern Argentera Massif. **Open-shelf condensed facies** can be found in Western Alps and southeastern French basin lower Hettangian strata. Shell beds capping platform carbonate units and overlain by hemipelagic marls are found in Ardèche (lower Hettangian so-called “lumachelle à Mytilidés” and lower-middle Hettangian so-called “lumachelle à Ostréidés”; Elmi and Mouterde, 1965; Martin, 1984), in Dauphiné (middle Hettangian and late Hettangian Ammonites-bearing condensed surfaces) and in subbrianconnais areas (lower Sinemurian shell beds with *Gryphaea arcuata* associated with hardgrounds or burrowed surfaces which cap the Upper Hettangian platform dolomitic limestones). Condensation documented by shell beds is useful to identify rapid relative short-term sea-level rises in areas of moderate subsidence (Kidwell, 1993). Crinoid grainstones and packstones developed on flooded platforms or isolated highs. These facies occurred frequently in western Europe as early as middle Hettangian time either as “transgressive facies” during short-term relative sea-level rises (i.e., early-mid-
dle Hettangian of Ardèche; Martin, 1984) or on tectonically controlled submarine swells. In the Dauphine area, their paleo-bathymetry and their distribution were controlled by the evolution of extensional fault-blocks (Bas, 1988; Roux et al., 1988; Pinto-Bull, 1987). Hemipelagic marls with sponge spicules are common in the lower Hettangian record of the Dauphine, Subalpine and Ardèche areas, and are also found in the Maritime Alps, Northern Provence, Cévennes and Eastern Causses areas. Abundant neritic fauna and bioturbation point to moderate paleodepths. During middle-late Hettangian time, thick hemipelagic sequences of alternating micritic/marly limestones were deposited. They show a more restricted distribution (Subalpine and Ardèche areas) and significantly greater paleodepths (>200 m; Améziane-Cominardi, 1991).

SEQUENCES

Some of the studied sections are shown on Figures 2–5 with their sequence stratigraphic interpretations. The lack of high-resolution age-diagnostic fauna in shallow environments is counterbalanced by the dense distribution of sites. The Western European realm was actually weakly differentiated at the end of Triassic times, and many marker beds and sedimentological unconformities can be traced laterally over long distances.

Three sequences ($S_R$, $S_{H1}$ and $S_{H2}$), bounded by four sequence boundaries, are identified in the approximately 14-my stratigraphic interval covered here. The first sequence ($S_R$) contains, in some places, two cycles (internal Alps and Subalpine domain; $S_{R1}$ and $S_{R2}$) whose significance will be discussed below. The sequence boundaries show little or no evidence of emergence, and the transgressive surfaces are well developed. This is due to the combined effects of the long-term Late Triassic–Early Jurassic sea-level rise which is recorded on “stable” platforms (Paris Basin; Guillochon, 1991) and of regional or local tectonic subsidence, which both tend to reduce the effects of relative sea-level drops.

Sequence $S_R$ (Late Norian? to Rhaetian), Cycles $S_{R1}$ and $S_{R2}$

Basal Sequence Boundary ($S_{BR}$; Late Norian or Early Rhaetian Age).—

This sequence boundary overlies a zoned emergent platform with north-south to northeast-southwest trending mixed alluvial plain to sabkha facies of late Norian age. This platform com-
prises, from west (northwest) to east (southeast), alluvial plain to piedmont coarse siliciclastics, confined terrigenous deposits with evaporitic troughs (“Keuper” facies) and dolomitic platforms (“Hauptdolomit” facies of the internal Alps). The sequence boundary in the internal Alps is marked by the change from progradation to retrogradation without erosional unconformity. To the west (western Alps and foreland), SBR1 coincides with a sharp transgressive surface with local evidence of pedogenic alteration (internal Alps, section Asc; Ardèche, Spy-Anderson, 1980), but neither a major base-level fall nor important tectonic activity seem to have been related to this sequence boundary. However, the lack of diagnostic fauna makes it difficult to estimate the duration of the depositional gap across the boundary, which generally shows an apparent conformity. The age of SBR1 in the internal Alps, where the depositional gap is expected to be shorter, is poorly constrained between the upper Norian Dasyclad algae and gastropods (Mégard-Galli, 1972) and the lower Rhaetian Avicula contorta pelecypods (25 m above SBR1; Lebouché-Bernet-Rollandé, 1972).

**Lower Cycle (SR1)**.

The basal part of the transgressive systems tract (TSTSR1, late Norian? to Rhaetian) is a dolomitic transgressive wedge that is found in sections Cha and Asc (Fig. 2). It pinches out towards the west (Subbriançonnais and external zones), below the oolitic base of the TST which is composed of shallow-marine limestones with the bivalve Avicula contorta that reached the western areas (Subalpine, Provence, Maritime Alps). These limestones contain condensed surfaces (“Bone-beds”) and detrital quartz. They are interbedded with dark subtidal shales that are a common feature of the Alpine and peralpine lower Rhaetian sections and which must have a genetic link with the transgression (drop in carbonate production and/or climatically controlled increase in terrigenous supply). The maximum flooding surface (MFSR1) occurs in dark subtidal bioturbated micrites to the east and in shallow subtidal bioclastic limestones with storm-layers and ravinements in the western part of the profile (Fig. 2). The early highstand systems tract (HSTR1, Rhaetian) is marked by the gradual development of corals in a lagoonal setting (sections Cha and Asc; Thecosmilia sp. patch-reefs with Megalodon sp.; Dumont, 1984), coeval with a rapid increase in carbonate content. These layers contain the Rhaetian foraminifer Triasina hantkeni Majzon (Dumont and Zaninetti, 1985). They are capped by regressive dolomites with desiccation and algal matts that represent the late highstand.

**Upper Cycle (SR2).**

Within the landward sections (Subalpine and Subbriançonnais areas, Fig. 2), cycle SR2 is mostly regressive and made up of intertidal to supratidal dolomites; It can hardly be distinguished from the HST of cycle SR1. The basal, retrograding part of SR2 appears in the eastern sections (Briançonnais and Piémontais, Fig. 2). Triasina hantkeni Majzon is found at the top of SR2 in section Asc (Zaninetti et al., 1986) and in section Mor. The occurrence of two cycles within the Rhaetian series is not just a local feature as discussed below.

**Sequence S1R (Late Rhaetian? to Early Hettangian Time)**

**Basal Sequence Boundary (SB1R, Late Rhaetian or Earliest Hettangian Age).**

This sequence boundary always coincides with the first flooding surface, since no platform margin existed at that time in the study area. Very little evidence of reworking is found. In section Asc, a ravinement surface of late Rhaetian to earliest Hettangian age cutting peritidal dolomites with Triasina hankeni Majzon is marked by a meter-scale breccia containing vertebrae of a terrestrial Allosaurus-like dinosaur (Ph. Taquet, per-
The top of sequence $S_R$ is never significantly truncated. $S_{BH_1}$ flooded a much wider area than the underlying sequence $S_R$ (northern edge of the Argentera massif, sections Tp1 and Sam; Dauphiné, around section Clo) and onlapped the late Hercynian basement or red beds of the Massif Central (eastern and southern Causses; section Tri).

**Transgressive Systems Tract (TST$_{H1}$, Late Rhaetian? to Early Hettangian Age).**

The lower part of TST$_{H1}$ is made of an aggrading set of shallow parasequences (thick-bedded oolitic and pelloidial packstones-grainstones) that have a very wide distribution (Subalpine areas, Maritime Alps, Northern Provence, Ardèche and Cévennes areas). They contain detrital quartz at their base, and their top is markedly transgressive (i.e., the so-called “complexe carbonaté de base” of Ardèche; Gallien, 1985). The basal parasequences of the TST are not dated, but they are probably diachronous. The upper part of TST$_{H1}$ in many sections shows typical features associated with an increasing rate of relative sea-level rise; prominent transgressive surfaces with crinoids and shell beds, unusual microfacies with mixed faunas reworked from different types of environments (“transgressive facies”; Arnaud-Vanneau and Arnaud, 1990) and a rapid increase in paleowaterdepth and clay content. The pelycopod faunas are typical of early Hettangian fauna. The maximum clay content coincides with the occurrence of *Psiloceras planorbis* or *Psiloceras psilonotum* in the Subalpine domain (section Tau; Assenat et al., 1972), in the Dauphiné domain (Barfety, 1985) and in the Maritimes Alps (Darveau, 1983) and with the occurrence of *Psiloceras psilonotum* or *Psiloceras plicatum* in the southeastern border of the Massif Central (Elmi et Moutarde, 1965; Martin, 1984; Gallien, 1985; Perrissol and Guerin-Franiatte, 1985). These layers, which also mark the most widespread distribution of the Hettangian marine facies, correspond to the maximum flooding of sequence $S_{BH_1}$ (MFS$_{BH_1}$, planorbis zone). The diachronism of the transgression indicated by the lack of the oldest species, *Psiloceras planorbis*, in the western sections, indicates the gradual landward shift of parasequences at the base of the TST.

**Highstand Systems Tract (HST$_{H1}$, Early Hettangian Age).**

The environments are more differentiated than previously, but an overall regressive trend is observed in all the sections.
above the *psiloceras* layers. It is shown by an increase in carbonate content, shallowing upward facies (i.e., sections Aut, Vil, Asc, etc.), development of lagoonal facies with patch reefs (section Trp; sections Bos and Ver, Ardèche, Martin, 1984) and development of emergent dolomictites in the southern areas (Maritime Alps and northern Provence, section Bon; Cévennes and S. Causses, section Mia). The open-marine facies are late early Hettangian in age (*Caloceras torus* d’Orb.; section Tan; Mouterde and Coadou, 1971; *Caloceras johnstoni*; section Bos; Elmi and Mouterde, 1965; Martin, 1984).
A local transgressive event, which is documented by a sharp transgressive surface in the Subalpine domains (Diane nappe and foreland, sections Tan, Aut, Bar), occurred inside HSTH1. In these places, the sequence SH1 is composed of two cycles. The maximum flooding surface of the lower one is found in other areas and hence has a regional significance. The second maximum flooding surface occurred in the upper part of the planorbis zone (johnstoni subzone). This latter event could represent a local response to a tectonic event.

**Sequence SH1 (Middle to Late Hettangian Time)**

**Basal Sequence Boundary (SBSH1, Late Early Hettangian or Earliest Middle Hettangian Age).—**

At this time, the study area consisted of several paleogeographic areas. In the central subsiding trough of the Subalpine area, the marine sedimentary record was more or less continuous and the boundary lies between the regressive top of HSTH1 and a middle Hettangian transgressive set of parasequences. In the foreland of the Digne nappe, it corresponds to a sharp, burrowed flooding surface. In Dauphiné, it is marked by condensation and the development of crinoid limestones. In Ardèche, it is followed by the drowning of the lower Hettangian patchreefs and by hardgrounds and burrowed surfaces (Martin, 1984). In the other areas, the identification of SBH1 is problematic since the restricted platform sediments are devoid of any good paleontological markers. However, in several of these sections, one important regressive/transgressive event occurs inside the Hettangian carbonate wedge that overlies the psiloceras layers. This event can be correlated with SBH2 (Cévennes: sections Bes and Mia; Maritime Alps: section Bon). The basal flooding surface of this event seems to correspond to the karstified surfaces known in the southern Causses at the top of the lower Hettangian dolomitic wedge (Fig. 5; Melas, 1982; Aronneau et al., 1985). Such a karstified surface truncates the top of SH1 in section Rim. In some areas, the sedimentary record ceased from SBH2 until Sinemurian time or later due to emergence (southern Subalpine area; Assenat et al., 1972; “Haut-Var” swell: Dardeau, 1983; Eastern Massif Central border: Vi-try, 1982).

**Transgressive and Highstand Systems Tracts (TSTH1 and HSTH1, Middle to Late Hettangian Age).—**

The TST is best observed in hemipelagic areas. Thinning-upward marl/limestone parasequences containing (in section Tan) Waehneroceras portlocki Wright and Waehneroceras gottingense Lange (second parasequence) then Alsatites laqueus Quenst. and Alsatites playstomya Lange (third parasequence; Mouterde and Coadou, 1971, and determinations by R. Mouterde). The TSTH1 parasequence set thins towards the southern Digne nappe (section Cou) where the overlying marly interval contains Schlotheimia angulata (Corna et al., 1990), and it is condensed in the foreland of the nappe (sections Aut and Bar). A typical, well developed TST of the same age is found in Ardèche (section Bos; Elmi and Mouterde, 1965; Martin, 1984). It is composed of transgressive crinoid limestones with shell beds (so-called “lumachelle à Ostréidés”) overlain by nodular limestones and marls bearing, from bottom to top, Waehneroceras portlocki Wright, Alsatites playstomya Lange and Schlotheimia angulata. Condensation occurred in the Dauphiné area during middle Hettangian deposition as shown by crinoid grainstones with hardgrounds and, near section Clo, by a 1-dm bed containing Waehneroceras schroederi Lange and numerous small Alsatites playstomya Lange (determinations by R. Mouterde). The overlying hemipelagic marls (section Cdu) contain Schlotheimia angulata (Bas, 1985).

According to our correlations (Figs. 4, 5), TSTH2 is represented in sections Bon, Bes and Mia by a thin transgressive interval overlain by thick aggradational dolomitic wedges. In such areas, most of the keep-up sedimentary record overlying SBH2 represents the highstand systems tract HSTH2, whose aggradational character could be due to increased tectonic subsidence during late middle and late Hettangian time. Extensive, possibly mixed-water, secondary dolomitization is commonly observed near the edges of dolomitic platform areas (i.e., sections Bes and Trip). A regressive trend at the top of the Hettangian dolomitic formation is documented in southern Causses (upper unit 5 of Melas, 1982), together with the development of greenish continental clay layers. The latter are common in the southern, landward areas of Provence where they sometimes show dinosaur footprints (Arnaud and Monleau, 1979; Ellenberger, 1965). In the northern Provence area, they show a northward progradation of continental influences at the top of Hettangian strata (section Smp; Arnaud and Monleau, 1979). In the Causses area, the same intercalations of greenish continental clays occur also in the upper part of the Hettangian series (i.e., section Trip), together with an increase in confinement and the coeval development of evaporites in the upper part of the Hettangian series of the southern part of the southeast French basin (Baudrimont and Dubois, 1977; Courel et al., 1984). These features are related to the late SBH2 highstand.

**Above Sequence SH2**

The upper sequence boundary of sequence SH2 is quite visible on the platforms. In the hemipelagic sediments (Digne nappe), the continuous record makes the boundary difficult to place, i.e., in section Cou (both late Hettangian ammonites and basal Sinemurian ammonites of the rotiforme zone are present; Corna et al., 1990). In Ardèche, the boundary is marked by a condensed latest Hettangian crinoid layer and shallow bioclastic facies (Elmi and Mouterde, 1965; Martin, 1984), but no early Sinemurian sedimentary gap is documented.

On the Causses platform, the relative sea-level fall at the top of sequence SH2 is generally marked by lacustrine layers with plant debris. The basal one in southern Causses is associated with a karstic surface (Melas, 1982), and their age is latest Hettangian (Thévenard, 1994). In the eastern part of the study area, the boundary capping SH2 seems to be associated with an Early Sinemurian condensation or sedimentary gap. In western Dauphiné, the rotiforme zone is represented by a fairly continuous hemipelagic record (Barfety, 1985), but, in eastern and southern Dauphiné, it is marked by hardgrounds and condensed limestones overlain by ammonite layers of the semicostatum zone (Barfety, 1985 and additional determinations by R. Mouterde). In the Maritime Alps (section Bon), the oldest ammonite above SH2 is late early Sinemurian in age (Epophioceras sp., found by G. Dardeau). In the Internal Alps (Subbrianconnais, sections Pes and Mor) and the northeastern cover of the Argentera Massif, the upper Hettangian lagoonal wedge is capped and locally
truncated by a bored surface overlain by shell beds with *Gryphaea arcuata* and by a transgressive set of parasequences with Early Sinemurian ammonites (Fig. 6). The lowest Sinemurian *rotiforme* zone is absent, since the first ammonites found in section Mor belong to the upper *bucklandi* or *semicostatum* zone (Schneegans, 1938; Mouterde, pers. commun., 1992).

**Regional Synthesis**

Figures 7 to 9 summarize the stacking patterns in the study area along profiles whose location is shown in Figure 10. They show: (1) the shift of subsidence from the Internal Alps to the southeastern French basin between Rhaetian and early Hettangian time and the increase in tectonic activity during middle and late Hettangian time, and (2) the gradual westward encroachment on the Hercynian lands of Massif Central, which is consistent with the global long-term sea-level rise enhanced by the westward shift of extensional tectonics. A tentative paleogeographic reconstruction for late Hettangian time which corresponds to the maximum facies differentiation of the studied time span, is given in Figure 10.

**Tectonics**

Evidence of syndepositional tectonic activity is scarcer in the early-rift sedimentary record than in the interval corresponding to the climax of the rifting (middle to late Liassic time). It consists of:

1. **Block-faulting and associated synsedimentary markers.** Very late to the climax of the rifting (middle to late Liassic time). It early-rift sedimentary record than in the interval corresponding to a transgressive set of parasequences with *Gryphaea arcuata* truncated by a bored surface overlain by shell beds with *Gryphaea arcuata* and by a transgressive set of parasequences with Early Sinemurian ammonites (Fig. 6). The lowest Sinemurian *rotiforme* zone is absent, since the first ammonites found in section Mor belong to the upper *bucklandi* or *semicostatum* zone (Schneegans, 1938; Mouterde, pers. commun., 1992).

2. **Local uplifts revealed by Hettangian sedimentary gaps (Fig. 10).** In northern Provence and Maritime Alps, some Rhaetian or Hettangian sediments are unconformably overlain by late Liassic to Bathonian strata along an east-west elongated high (Assenat et al., 1972; Dardeau, 1983; Mouterde et al., 1984; Monleau, 1986). The Rhaetian strata, which are much softer than the lower Liassic dolomites, are always preserved below the unconformity, so that this feature is better explained by an early Liassic uplift than by post-Liassic erosion. This structure was more or less parallel to the northern edge of the Provence dolomitic platform. At the transition between the Ardèche basins and the Cévennes dolomitic platforms, a similar westnorthwest-eastnortheast oriented horst was active during Hettangian time (Elmi et al., 1984).

3. **Differential subsidence occurred in the easternmost domains during the deposition of sequence *S*R (Figs. 7–9).** To the west, this sequence is absent in the sedimentary cover of the whole Dauphiné Crystalline Massifs, except in their easternmost part (east of La Meije). This area corresponds to the distribution of late Triassic volcanism, which occurred during late Norian to earliest Rhaetian time (Fig. 12; see below). The absence of sequence *S*R in this area was probably due to a positive thermal anomaly following the volcanic event. Differential subsidence increased and shifted westwards during early and particularly during middle and late Hettangian time. On the southeastern border of the Massif Central, this major phase of subsidence (Brunet, 1985) is associated with large-scale block-faulting (Giot et al., 1991). Major boundary faults could also exist in areas of abrupt thickness and facies changes (i.e., to the south of the Pelvoux Massif).

4. **Hettangian barite mineralization along westnorthwest-east southeast fissural trends in the marginal dolomitic upper Hettangian platform of Causses (Macquard, 1978; Fuchs, 1978).**

5. **Paleostress history deduced from analysis of striated paleofault planes in the Dauphiné area (Grand, 1988; Grand et al., 1987).** Stress orientation shifted from north-south during the late middle and late Triassic time (pre-rift stage) to northeast-southwest or east-west during a possibly transcurrent, transitional stage (early Liassic; Hettangian and part of Sinemurian time) and then to northwest-southeast, that is, perpendicular to the rift orientation, during the main rifting phase (middle and late Liassic). The Triassic/Liassic shift is also documented in the Paris Basin (Mascl and Cazes, 1987; Bois et al., 1988), in the Swiss Prealps (Mettraux and...
Fig. 7.—Reconstructed pattern of depositional sequences along profile A (location on Fig. 10; internal Alps to Massif Central). Paleodepths not to scale. Note the increase in differential subsidence and the westward shift of extensional activity through time.

Fig. 8.—Reconstructed pattern of depositional sequences along profile B (location on Fig. 10; Dauphiné to Maritime Alps). Paleodepths not to scale. Note the increase in differential subsidence through time.
FIG. 9.—Reconstructed pattern of depositional sequences along profile C (location on Fig. 10; southeastern border of Massif Central). Paleodepths not to scale.

FIG. 10.—Paleogeographic reconstruction during the middle and late Hettangian time (sequence S H2). The subbriãoconnais sections of Ubaye (sections Mor and Pes) are restored near the Argentera massif because they closely resemble those of the eastern parautochthonous cover of Argentera massif (section Sam). This early-rift paleogeographic situation is significantly different from the synrift one (middle and late Liassic): east-west and northwest-southeast directions are comparatively better represented whereas northeast-southwest (“Cévenole”) trend is less involved than during the climax of the rifting (middle and late Liassic time). There is no evidence that the Causses and Provence dolomitic platforms were not connected during early Liassic time, but strongly subsiding intra-platform evaporitic basins (drillhole “Les Angles”) mark the future zone of breakup between the Cévennes, Nîmes and Durance northeast-southwest fault trends.
FIG. 11.—Middle to late Hettangian small-scale block faulting in the Dauphine domain (southwest from Bourg d’Oisans, near section Clo). The paleotopography at the top of the basalts created by small-scale late Triassic faulting and by the volcanic buildups is overlapped by the neritic limestones of sequence SH1. The transgressive systems tract of sequence SH2 is capped by an interval of maximum starvation containing Waehneroceras schoedleri Lange and Alsatites platysternus Lange. This series is cut by several small-scale synsedimentary normal faults and the tectonically enhanced maximum flooding surface, MFSH2, is overlain by late Hettangian hemipelagic marls. A local east-west direction of extension is documented by the orientation of normal faults and neptunian dykes in sequence SH2.

FIG. 12.—The latest Triassic volcanic event. In the Dauphine, Pelvoux and Northern Digne Nappe areas, the thickest volcanic buildups are located around the southern border of the Pelvoux massif, which is known as a transform fracture of the Jurassic paleomargin. The orientation of the dikes which fed these buildups documents an east-west extension direction compatible with the early Liassic extension (after Laurent, 1992).

Mosar, 1989), and in the French Briançonnais domain (Megard-Galli and Faure, 1988; Faure and Megard-Galli, 1988). However, there is local evidence for different stress orientation in the foreland (east-west Triassic extension in Ardèche; Bergerat and Martin, 1993).

These indications of tectonic activity clearly show (a) a tectonic shift between late Triassic and the early Liassic time, and (b) an increase in activity during early and particularly during middle and late Hettangian time. This suggests that, in the study area, the changes in paleoenvironments and depositional pattern during the deposition of sequences SH1 and particularly SH2 are partly controlled by a regional tectonic reorganization. Sequence SR was deposited before the western Alpine realm was significantly affected by this turnover. One can assume that the T/J boundary, which is regarded as the beginning of the synrift stage in the western Alps, marks the jump of extensional activity from the eastern and southern Alpine areas (Tethys) to the Ligurian Tethys area associated with central Atlantic dynamics. Some parts of the western Alpine realm had strong subsidence during Triassic time (central southeast French Basin: Courel et al., 1984), but this process ceased several million years before the T/J boundary, so that two different rifting events can be clearly separated.

Many features of the early (earliest Liassic) synrift stage do not fit the typical syn-rift, northwest-southeast Jurassic extension (Lemoine et al., 1989). Transverse faults, as described above, together with oblique facies distribution (fig. 10), show that the transverse structural grain was activated during Hettangian time, whereas the directions parallel to the rift (north-south to northeast-southwest) clearly prevailed during the climax of the rifting. This suggests that either the orientation or the mechanism of extension changed during the development of the Liassic rifting, which was polyphase (Dumont and Grand, 1987; Faure, 1990) as seen, i.e., in the modern rifting of the Gulf of Suez (Ott d’Estevou et al., 1989).

VOLCANISM

The well-known, alkaline to transitional volcanism of the Dauphiné area (“spilites”) consists of thin, continental lava flows that overlie Upper Triassic dolomites. Lateral thickness changes, from 80 m (4 or 5 flows) to a couple of meters (tuffs), illustrate the paleotopography of the volcanic buildups that
were preserved by the onlapping early Hettangian neritic limestones. Neritic fauna of Rhaetian affinity have been found locally in the first beds overlying the lava flows (section Cdu: Moret and Manquart, 1948; section Pro: Mouterde, pers. comm., 1986). The thickest flows are observed along the southern border of the Dauphiné external crystalline massifs and in the northern Digne nappe (Fig. 12). This, together with the dense pattern of dikes in the southern Pelvoux area, suggests that the southern border of the Pelvoux massif, which is believed to have acted as a transfer fracture during the Liassic rifting (Lemoine et al., 1989), was already involved during the latest Triassic volcanic event. The predominantly north-south orientation of the dikes (Buffet et Aumaître, 1979; Laurent, 1992) is more compatible with the Liassic extension than Triassic. This volcanic event may indicate the beginning of a new tectonic regime that prepared the opening of the Ligurian Tethys ocean.

Coeval, short volcanic events are known in many parts of the Tethyan or Atlantic realms: Pyrenees (Curnelle and Cabanis, 1989), Iberian chain and Majorca (Pocovi et al., 1989), Morocco and Central Atlantic margins (Manspeizer, 1988; Sebai et al., 1991; Sutter, 1989) or even northwestern Australian margin (von Rad et al., 1990). A diagenetic event of thermal origin occurred in Morocco close to the Triassic-Liassic boundary (Huon et al., 1993). These events could be linked with a global plate tectonic reorganization near the Triassic/Jurassic boundary (Dumont and Röhl, 1992).

**CORRELATION WITH EUROPEAN AND GLOBAL EVENTS**

Four regionally traceable depositional sequences occur within the late Norian to early Sinemurian time span. Only two sequences are found on the Haq et al. (1987) cycle chart (215/211 Ma. and 211/202 Ma.). The occurrence of four sequences is not surprising since Haq et al. (1987) two sequences have a significantly greater duration (5 and 9 my) than typical 3rd-order sequences (0.5–3.0 my; Vail, 1992). However, time scales are poorly constrained for this interval; the duration of the Hettangian is only 3 my according to Odin (1994) and 4.5 my according to Harland et al. (1990). In this section, the European-scale variability of the observed regional sequential organization (Figs. 7 to 9) and the possible forcing mechanisms are discussed.

**Late Triassic Time**

Although the validity of the Rhaetian stage has been a matter of debate in western Europe because of the continental influences and the scarcity of age-diagnostic fauna tied to complete basinal records, it is well known that a general transgression occurred during this period (Paris Basin: Guilloucheau, 1991; western Alps: Mégard-Galli and Baud, 1977; Dumont, 1988; Lombardy: Jadoul and Gnacolini, 1992; Eastern Alps: Tollman, 1978; Western Carpathians: Gazdzicki, 1974; Germany: Blob, 1990; Ligurian Alps: Lualdi, 1990; Betic Cordillera: Lopez-Lopez et al., 1988; Morocco: Maate et al., 1993). In western European areas, the lower Rhaetian transgression is probably diachronous, and the basal boundary generally coincides with a transgressive surface. A significant depositional gap must be suspected along this boundary, since a major late Norian sea-level drop is found in the eastern Alps (Krystyn, 1980). In the southern and eastern Alps, the more complete sedimentary sequences deposited in intraplatform troughs or along the platform edge show a marked late Norian to early Rhaetian transgressive systems tract (Zorzino limestones and Riva di Solto shales in southern Alps: Gnacolini, 1965; Masetti et al., 1989; basal limestones with coquinas and Choristoceras limestones in the northern Calcareous Alps: Karle, 1984) and a regressive evolution in the upper Rhaetian beds including reef development. Similar transgressive/regressive cycles of the same age are observed in different geodynamic settings outside Europe (Canadian Arctic, Embry, 1988; northwestern Australian margin: Dumont, 1992).

Sequence S_R thus matches this late Norian to late Rhaetian cycle recorded in different palaeotectonic settings of Europe. Very low accommodation space in the study area (especially the External Alps and southeast French basin) led to the deposition of a hiatal epicontinental sequence. This sequence is weakly affected by syndepositional tectonics, which occurred further east in Lombardy (Kalin and Trümpy, 1977; Bertotti, 1990). Two 3rd-order cycles, which could match our S_H and S_R, are identified within the very thick Rhaetian series of Lombardy (Jadoul and Gnacolini, 1992; De Zanche et al., 1992) and in the Paris Basin (Jacquin; person. commun., 1993). Thus these cycles are probably related to a Europe-wide event.

**Latest Triassic-Late Hettangian Time**

The problem of identifying the Triassic-Jurassic boundary is discussed in Hallam (1990). Below the first occurrence of the ammonite Psiloceras planorbis, many European sections contain some undated, transgressive beds (so-called “pre-planorbis beds”) which, from a sequence stratigraphic point of view, are linked to the Jurassic supercycle. The T/J boundary is placed within the base of the pre-planorbis transgressive beds in Austria, based on palynological data (so-called “boundary marls”; Karle, 1984), and thus the sequence boundary must have occurred in late Rhaetian time. This is consistent with the finding of a late Rhaetian sequence boundary on the northern Gondwanan margin (Dumont, 1992; Dumont and Röhl, 1992). A sea-level fall probably associated with this sequence boundary is documented in southern Germany by channels that cut the late Rhaetian formation and are filled with lower Hettangian fluviatile material (Blob, 1990) and in the northern Calcareous Alps by reworking on top of Rhaetian buildups (Böhm, 1992). In the dominantly carbonate series of the Internal Alps, the boundary is marked by siliciclastic input (Briançonnais and Subbriançonnais of Western Alps, this paper; Briançonnais of Prealps, Mettraux, 1989). The sequence boundary S_B is in good agreement with these data. However, no upper Rhaetian sediments could be found above it, probably because the lowstand and basal transgressive systems tracts of the overlying sequence S_H are not recorded. The basal, undated parasequence set of sequence S_H could correspond to the upper pre-planorbis beds.

The maximum flooding surface of the planorbis zone is the best documented event of Hettangian times. It is recorded in many parts of the European platform (Hallam, 1981, 1988) and on other continent margins, such as the Canadian Arctic (Frebold and Poulton, 1977; Embry, 1988), and it is clearly identified in the study area (MFS_H). A significant later transgressive event is documented on the English and German platforms and in the Paris Basin in middle Hettangian strata (liasicus zone:}
Donovan et al., 1979; Jordan, 1983; Achilles and Schlatter, 1986; Bessereau and Guillocheau, 1993). On these “stable” platforms, the middle Hettangian depositional area appears to be much larger than the early Hettangian platform, and ammonite-bearing marls of the liasicus zone directly overlie the Triassic strata in many places. This is comparable with the maximum flooding surface MFSr, and the intra-Hettangian sequence boundary SBH2 proposed in this paper (top of the planorbis zone or base of the liasicus zone) is also a European-scale event. However, no significant sea-level drop associated with this boundary is documented outside the study area, and the local emergence described here towards the end of early Hettangian time results probably from a local tectonic enhancement of this sequence boundary.

Similarly, an increase in tectonic extensional activity is recorded in middle-upper Hettangian sections in other areas such as the Austro-Alpine nappes (major fault-blocks and coeval platform drowning, Eberli, 1988); eastern Alps (the platform drowning is marked by a local angular unconformity, Böhm, 1992 and by block faulting, Blau and Schmidt, 1988) and southern England (fault-controlled thickness changes and neptunian dikes, Jenkyns and Senior, 1991). In the Ligurian Piémont series, extreme marine condensation of the middle-upper Hettangian to lower Sinemurian series (upper planorbis to rotiforme zones) results from block faulting (after Lualdi, 1986). A similar condensation of sequence S12 is observed in the internal Alpine domains of this study (Fig. 7).

**Latest Hettangian-Early Sinemurian Time**

As discussed by Hallam (1988), there is little evidence for a marked sea-level drop at the end of the Hettangian time that could correspond to the major type 1 late Hettangian sequence boundary (202 Ma) shown on the Haq et al. (1987) cycle chart. In southern Germany, Bloos (1990) indicates a possible low-amplitude sea-level fall within the angulata zone, with fluviatile channels below the transgressive base of the bucklandi subzone. In the eastern French Massif Central, the rotiforme subzone corresponds to an increase in terrigenous clastic influx indicating a possible seaward shift of coastal plain facies (Vitry, 1986). In this paper, there is a lack evidence for a major sea-level drop associated with the upper boundary of sequence S12.

**CONCLUSIONS**

The upper Triassic-lower Jurassic sedimentary series of the western Tethyan European platform in southeastern France and western Alps recorded relative short-term sea-level fluctuations superimposed on an overall long-term transgressive trend. The lack of evidence for major sea-level falls is due to the superimposed effects of regional subsidence and long-term global sea-level rise coincident with the first connection between the Arctic and Tethys seas (Ziegler, 1989), but it does not preclude the identification of regionally-traceable sequence boundaries. Figure 13 shows a schematic regional transgression/regression trend consistent with the deposition of the observed sequences and summarizes the information about the coeval extensional tectonics.

The ability to trace sequence boundaries across several palaeogeographic areas that experienced different tectonic behaviors during incipient Jurassic rifting demonstrates the occurrence of a continental- or global-scale forcing mechanism. However, the observed sequential evolution is significantly different from the Haq et al. (1987) chart: an overall transgressive trend after late Norian time, two Rhaetian cycles, one early Hettangian sequence boundary, and no major (“type 1”) late Hettangian sea-level fall. Some of these discrepancies were also noted by Hallam (1981, 1988). They suggest that the Haq et al. (1987) global chart does not apply exactly to the European realm, probably because global eustasy is not the single forcing mechanism of the observed sequential evolution.

Subsidence, which is classically the alternative explanation involved in the fluctuations of accommodation rate, is constrained both by the mechanism of extension of the continental crust and by the thermo-mechanical evolution induced by the vertical movements of the underlying mantle. Several models have been applied to the Alpine and perialpine realms, depending on the time interval and the area considered: passive-margin
evolution with a step-by-step development of large-scale basin-ment tilted blocks (French Western Alps: Lemoine et al., 1986; Swiss Austroalpine nappes: Eberli, 1988; Southern Alps: Bertotti, 1990; Castellanin and Picotti, 1990), finite-duration lithospheric stretching (Wooler et al., 1992) and detachment-controlled lithospheric thinning with emphasis on the thermal control of vertical movements (French western Alps: Rudikiewicz, 1988; Swiss prealps: Loup, 1992) or on the lateral propagation of brittle deformation in the upper crust (Swiss Austroalpine nappes: Eberli and Froitzheim, 1989; southern Alps: Sarti et al., 1991). Depending on the model chosen, the accommodation is chiefly controlled either by geometric or by thermal processes. Isostatic response is generally not involved in short-term fluctuations.

Except in Ardèche (Elmi, 1985; Giot et al., 1991), extensive brittle deformation did not affect the study area during latest Triassic-early Liassic time. The small-scale fault blocks observed (Dumont et al., 1984; Tricart et al., 1988; Grand, 1988; Fig. 11) do not indicate that the renewed subsidence that marks the beginning of the early synrift phase (Hettangian; Lemoine et al., 1986) is a result of lithospheric stretching linked to brittle extensional deformation of the upper crust. Thus, brittle deformation does not seem to be the main controlling factor of accommodation.

On the other hand, the observed changes around the TJ boundary and the coeval extensional events such as reorganization of the subsidence pattern, onset of platform breakup, volcanism and change in the paleostress orientation are surprisingly synchronous with other Tethyan or Atlantic events. These include the shift in the direction of extension in the central Atlantic (Klitgord et al., 1988) and short-lived volcanism (Sutter, 1989; Sebai et al., 1991), the early Cimmerian phase in northwestern Europe (Ziegler, 1978, 1988), and turnover in the subsidence pattern and volcanism of the northern Gondwanan margin in northwestern Australia (von Rad et al., 1990; Dumont and Röhl, 1992).

The middle Hettangian event was marked by drowning and small-scale faulting in the western Alps (this paper), drowning and large-scale faulting in the Austro-Alpine nappes (Eberli, 1988) or Ardèche (Giot et al., 1991) and decrease in subsidence rate linked to thermal relaxation in some areas of the Helvetic foreland (Loup, 1992). This indicates that a tectonic shift was rapidly transmitted through the Tethyan realm at that time, but that area had a different response to it. This can not be explained by thermal processes or by the concept of rift propagation. The model of intraplate stresses propagation (Cloetingh, 1986) could provide a better explanation for rapid, synchronous tectonic changes of different amplitudes.

Such global plate-tectonic reorganization events can produce major globally synchronous disruptions in the 2nd-order sea-level signal. A correlation between intraplate stress variations and 3rd-order cycles is a more debated question, but numerical modelling has demonstrated that short-term sequences can be generated by rapid variations in the horizontal stress field (Cloetingh and Kooi, 1989). The widespread distribution of a short tectonic event such as the middle-late Hettangian one, whose duration is compatible with the observed sequences, could support such a correlation at a European scale.

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PART V
MESOZOIC ERA
TRIASSIC PERIOD
TRIASSIC SEQUENCE STRATIGRAPHIC FRAMEWORK OF WESTERN EUROPEAN BASINS

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ABSTRACT: The Triassic western European succession comprises at least four transgressive/regressive 2nd-order facies cycles which can be followed from the Boreal siliciclastic to the Tethyan carbonate settings. The exact timing of the peak transgression and maximum regression of these cycles vary slightly from basin to basin according to local physiographic conditions, tectonic behaviour and sediment supply. Up to 23 3rd-order depositional sequences are the building blocks of the 2nd-order facies cycles. They form a good framework for correlation across basins and across facies belts. A major 1st-order transgressive/regressive facies cycle also formed during Triassic times. Its lowermost and uppermost boundaries do not necessarily match with the Triassic system boundaries but, in northern Europe areas, with the Hardegen unconformity (Late Scythian) for the lowermost one and with the Early Cimmerian unconformity (Late Norian) for the uppermost one.

INTRODUCTION

One of the major problems of the western Europe Triassic stratigraphy is the lack of adequate conventional stratigraphic tools over large areas due to the development of continental-terrigenous and/or evaporitic-facies. The sequence stratigraphic approach in that context provides a physical framework which is dependent of the respective changes of sediment supply and shelfal accommodation. High density wire-line log information across basins shows that this physical framework can be highly correlatable and used confidently for large-scale correlation between basins. Among the different orders of available physical cycles (from the 4th to the 1st-order), the 3rd-order framework is the most suitable. Their physical stratigraphic signature and position (rank) together with the 2nd-order transgressive/regressive cycles can be used as the appropriate tool. In addition, the major turning points of the 2nd-order framework (peak transgression, major flooding or drowning) may be characterized across basins by major turning points in the development of the bio-evolution which constrains the large-scale correlation.

Eight columns (Figure 1) from the Boreal to the Tethys realms summarise the most suitable published sequence stratigraphic framework for each basin or subbasin. The original published stratigraphic interpretation (3rd- and 2nd-order frameworks and lithostratigraphic nomenclature) have been maintained on Figure 1 even though alternative present-day interpretations could be proposed.

Tethys

In this volume, contributions deal with the Southern Alps (northern Italy) and the Northern Calcareous Alps (Austria) which are located in the western termination of the Tethys (Tollmann and Kristan-Tollmann, 1985; Stampfli et al., 1991; Ricou, 1994). Triassic successions comprise both shallow- and deep-marine sediments, carbonates and siliciclastics, depending on the location. The sequence stratigraphic setting has been pointed out on the basis of only outcrop observations (Gaetani et al., Gianolla et al., Rüffer and Bechstädt, this volume) and scanty well-log analysis. The dating of the sequence stratigraphic units has mainly been carried out by means of ammonoids, conodonts or palynomorphs.

Germanic Domain

The Germanic domain consists of intracratonic basins (Ziegler, 1982, 1988; Hagdorn, 1991) between the Arctic domain to the north and the Tethyan domain to the south. Depositional environments are mainly continental to shallow-marine, ranging from fluviatile, alluvial-plain to carbonate ramp settings and sabkha. The Germanotype Triassic is characterized by the classic tripartition: Buntsandstein (mainly siliciclastic and continental facies), Muschelkalk (carbonate marine facies) and Keuper (sabkha and evaporitic paralic facies). The Germanic facies extends over a wide area of the continental Europe, from Poland to Spain and from Morocco to Norway. Several seaways, such as the Burgundy Gate, Silesian-Moravian Gate, East Carpathian Gate (Hagdorn, 1991), connected the Tethys with the Germanic basins intermittently. To the north, a connection with the Arctic domain was active only during the Rhaetian (Ziegler, 1982, 1988). Sequence stratigraphic analysis has been pointed out both on the basis of regional seismic lines, integrated by well log analysis and outcrop observations (Aigner and Bachmann, Goggin and Jacquin, Courel et al., this volume). The dating of sequences and systems tracts is mainly based on ammonoids, conodonts, palynomorphs and tetrapod footprints.

Arctic Domain

The European Arctic Triassic succession was deposited within an intracratonic basin (Ziegler, 1988; Mørk et al., 1989, 1992; van Veen et al., 1992). It is characterized by mainly open-marine terrigenous deposits, ranging from offshore marls to evaporitic lagoon throughout deltic environments. The European Arctic domain includes the Svalbard Archipelago, the Barents Shelf, the Norwegian-Greenland Sea and the northernmost part of the North Sea. A rift system developed during the Triassic within the North Sea and Arctic domain (Ziegler, 1988; Steel, 1993). The Arctic Sea developed mainly with no direct connections with southern seas. Due to the high paleolatitudinal position, the outcropping Boreal Triassic successions in Svalbard are generally fully marine and bear well defined ammonoid assemblages which are difficult to correlate with their Tethyan time equivalents. Sequence stratigraphical analysis has been mainly pointed out on the basis of seismic line interpretation and well log analysis (Skjold et al., this volume).

SEQUENCE STRATIGRAPHY

Besides the differences highlighted within the Figure 1, each basin recorded many more 3rd-order cycles than the Haq et al. (1987) cycle chart was showing. This is probably due to the better and larger quality data base used for this study. Causes
for differences and similarities between basins differ from the 1st- to the 3rd-order cycles.

**Third-Order Depositional Sequences**

Generally, both the Southern Alps and the Arctic were characterized by a high rate of tectonic subsidence and open-marine environments. Their 3rd-order sequence stratigraphical setting seems to be the most complete.

Slight differences exist between the Southern Alps and the Barents Shelf Lower Triassic units. In the Germanic domain and North Sea area where the Scythian is mainly continental (Buntsandstein), a lower number of sequences have been defined. Additional work is still necessary to solve out that problem. Scanty data come from the Northern Calcareous Alps. The Anisian–Early Ladinian transgression is better documented throughout the western European basins. Therefore, few differences exist in the various authors sequence interpretation; differences result from poor age dating and/or local tectonic deformation.

The Upper Ladinian to Carnian interval shows going 3rd-order depositional sequences. Slight discrepancies still remain in the German Muschelkalk but comparisons with the neighboring eastern Paris Basin suggest that the M2 depositional sequence by Aigner and Bachmann (1992) can be split into M2-1, M2-2, M2-3 (Goggin and Jacquin, this volume). This limits the discrepancy to the single M2-2 sequence which is time equivalent to Lad1 and Lad2 in the Southern Alps. Larger discrepancies concern the south-eastern France Ardèche. They are the consequence of the poor quality data (limited outcrops and sparse datations).

Due to restricted environments and scanty biostratigraphical data, Norian sequences are hardly correlatable. However, each column in Figure 1 tentatively shows a similar 3rd-order setting. Even if the Paris Basin, western Southern Alps and Northern Calcareous Alps Rhaetian are relatively well documented, no univocal sequence stratigraphic interpretation can be pointed out.

**Second-Order Transgressive/Regressive Facies Cycles**

Striking differences appear in the 2nd-order cycle frameworks proposed by various authors. As an example Aigner and Bachmann (1992) recognized only one transgressive/regressive facies cycle for the Germany Late Scythian–Norian interval, whereas Skjold et al. (this volume) identified four transgressive/regressive facies cycles for the same interval in the Barents Sea. Such large discrepancies relate both to local features and authors’ interpretation.

On the Tethyan margin, the difference in timing relates partly to different interpretation of local tectonic and sedimentary evolution and partly to difference in biostratigraphic resolution. As an example, in the Southern Alps, Gianolla et al. (this volume) indicate one more cycle (Bithynian–Early Illyrian) than the four shown in the general 2nd-order column. Possibly, it was caused by increasing rate of tectonic subsidence and by tectonically controlled erosion of tilted blocks in a markedly extensional regime.

In south-eastern France, Courel et al. (this volume) recognize three marine ingressions within their long transgressive/regressive cycle, the first corresponding to the marls above the Grès de Rubreau (peak transgression, Lower Ladinian), the second within the Barre Carbonatée Médiane (Lower Carnian) and the third at the base of the Formation Bariolé d’Ucel (Upper Carnian). They assign a 2nd-order rank only to the first ingestion. In this sense, regaining and giving a higher rank to the ingression events in the Ardèche area, it is possible to fit them with the 2nd-order T/R facies cycles outlined in the Tethyan column. In addition, the general sequence stratigraphic framework of the Triassic in western Europe (Jacquin et al., this volume) suggests that the 2nd-order cycle by Courel et al. (this volume) fit with the 1st-order eastern Tethys cycle. Similar remarks can be proposed for the sequence stratigraphic framework by Aigner and Bachmann (1992). Their 2nd-order cycle (upper Oleneonian to uppermost Norian; peak transgression: lower Longobardian) could be ascribed to a 1st-order cycle, as it may be subdivided into several T/R cycles bounded by maximum progradations.

**First-Order Transgressive/Regressive Cycles**

The 1st-order transgressive/regressive cycles are bounded by a major subaerial unconformity, resulting from a major down-ward shift of the coastal onlap (Vail et al., 1991; Jacquin et al., 1992; Jacquin and Graciansky, this volume). However, their stratigraphic signal can be differently recorded for the same reasons mentioned above. Thus, while a general agreement on the position of the top of the 1st-order Triassic cycle (close to the Norian/Rhaetian boundary) really exists, its base is a matter of discussion (Duval and Cramez, Jacquin et al., Jacquin and Graciansky, this volume). In the Southern Alps the lower maximum regression seems to be placed at the Permian/Triassic boundary; while in the Germany and Arctic basins, it is coincident with the Hardesgen unconformity or its equivalent (Olenekian). The precise position of the peak transgression is also a matter of discussion; in France, Austria, Germany, and Barents Sea, it is placed near the Fassanian/Longobardian boundary or higher, in the Southern Alps it seems to be a slightly earlier (Fassanian, Chiesense Subzone).

**TRANSGRESSIVE/REGRESSIVE FACIES CYCLES IN WESTERN EUROPE**

The Triassic Chart (Fig. 1) includes several individual contributions regarding selected basins of western Europe. It was laid out according to the data of each article. Minor modifications, on the coordinators own responsibility, were introduced in order to obtain uniform boundaries when no constraining age indication was given by authors. Cycle positions are fitted according to the chronostratigraphical framework and the ammonoid standard scales. Two general sequence chronostratigraphical columns, respectively for the Tethyan and the Boreal domains, summarize and interpret the general setting of Triassic strata in western Europe. The differences of timing of the 2nd-order T/R cycles between individual basins may relate to local tectonics and sedimentary development. The poor biostratigraphic resolution for some time intervals may also explain the inferred discrepancies. Thus, the comparison of intracratonic, epicontinental and marginal basins, suggests that the strain field has differed with the place. In the Figure 1, four Triassic 2nd-order cycles are indicated, the only ones traceable throughout the western Europe.
TRANSGRESSIVE/REGRESSIVE FACIES CYCLE 1

Boundaries and Duration

Both in Tethys and Boreal domains the lower boundary corresponds to the Permian-Triassic boundary. In the Tethys the upper boundary falls in the uppermost Spathian, close to the Olenekian-Anisian boundary. In the Boreal domain it is a little older and it is placed in the lowermost Spathian. The duration is 5–6 my.

Lowermost Boundary

In the eastern Southern Alps, an unconformity lies within the upper part of the Permian Bellerophon Formation (Wignall and Hallam, 1992). The top of a few meters-thick series made of skeletal packstones-wackestones and terrigenous-carbonate grainstones (Permian) documents the maximum regression which seems to coincide with the P-T boundary (Assereto et al., 1973; Noé, 1987; Broglio Loriga et al., 1990). In the Barents Sea, no major change in shelfal accommodation, close to the P-T boundary, is known. However, an unconformity characterized by a not well-defined hiatus, separates Permian from markedly transgressive Lower Triassic sediments (Mørk et al., 1989; van Veen et al., 1992). A similar trend is documented in the German basins, where no major break exists between the Permian Zechstein and the overlying Buntsandstein. There, the regressive tendency of the uppermost Zechstein continues into the Lower Buntsandstein. No major break is observed in the basin center. At the basin margin, however, a regional angular unconformity exists near the base of the Lower Buntsandstein caused by non-deposition or erosion of the uppermost Zechstein cycles (Aigner and Bachmann, 1992). In the Paris Basin, seismic data suggest that the same genetic unit includes a conformable Upper Permian to Lower Triassic succession (Goggin and Jacquin, this volume).

Transgressive Phase

In the Southern Alps, the Werfen Formation laps westward on a differentiated substratum (Assereto et al., 1973; Wignall and Hallam, 1992). In the Arctic domain, lowermost Triassic marine sands and shales paraconformably or unconformably overlie Permian strata (Nakamura et al., 1990; Mørk et al., 1992; Nassichuk, 1995); sometimes the former bear lower Griesbachian ammonoids (Otoceras beds; Tozer and Parker, 1968). In the Barents Shelf, the transgressive phase is documented by a strong aggradation (Havert Formation).

Peak Transgression

In the Southern Alps as well as in Hungary and in the Dinarides, the peak transgression is placed within nodular silty limestone (Val Badia Member, Werfen Formation) bearing ammonoids (Tirolites cassianus Zone; Lower Spathian). In the Boreal domain, the peak transgression is older and is referred to the lower Smithian (Hedenstroem Zone).

Regressive Phase

Throughout western Europe, the regressive phase is documented by a rapid basinward progradation of terrigenous shorelines. In the Southern Alps, supratidal sediments covered main parts of previous basins.

TRANSGRESSIVE/REGRESSIVE FACIES CYCLE 2

Boundaries and Duration

In the Tethys this cycle ranges from uppermost Spathian to upper Longobardian; in the Boreal domain it ranges from lowermost Spathian to upper Longobardian. Its duration is 15 my.

Lowermost Boundary

In the German basin and in the Barents Shelf, this boundary corresponds to a well-defined unconformity or to a major progradational event. It is known as the Hardegsen (or “H-Diskordanz”), the most spectacular unconformity in the German Triassic (Aigner and Bachmann, 1992). According to van Veen et al. (1992), this event was caused by a change in lithospheric stress, related to a compressional event. Within the Buntsandstein in the Paris Basin, this major unconformity lies at the level of the Conglomérat Principal (Goggin and Jacquin, this volume). In the Barents Sea, the maximum regression is overlain by condensed transgressive early Spathian deposits (van Veen et al., 1992); in consequence, the lowermost boundary is considered to be near the Smithian-Spathian boundary. At that time in the Southern Alps, no major downward shift of facies occurred; the maximum regression is documented there by the widespread supratidal sediments dated as upper Olenekian.

Transgressive Phase

During Anisian and Early Ladinian times, a strong increase in subsidence was combined with a marked sea-level rise to create a widespread aggradational phase. Other consequences are the episodic connections between the Tethys and the German Basin (Pelsonian, Early Illyrian and Early Ladinian) and the migrations of ammonoid faunas in both directions (Mietto and Manfrin, 1995). Due to a marked landward shift of shoreline during the aggrading phase, a generalized decrease in siliciclastic supply was widespread throughout western Europe. The strong increase in accommodation allowed the deposition of great sediment thicknesses and forced carbonate platforms to backstep. In the German Basin, the aggrading phase allowed the deposition of the thick evaporitic Röt Formation (Aigner and Bachmann, 1992). In the Northern Calcareous Alps and in the Southern Alps, carbonate platforms drowned (De Zanche et al., 1993; Rüffer and Bechstädt, this volume). In the Paris Basin, widespread carbonate deposits (Muschelkalk) established during the latest stage of the transgression following the rapidly backstepping fluvial and deltaic sandstones of the Anisian (Goggin and Jacquin, this volume). Backstepping sequences named Grès à Voltzia, dated as Upper Anisian-Lower Ladinian, strongly lap out the crystalline basement on the borders of the Paris Basin. A similar situation is observed in the Ardèche area (south-eastern France). In the Boreal domain, the overall transgressive phase is indicated by various criteria such as: (1) onlap onto structural highs, (2) upward increase in organic productivity and content, (3) presence of widespread phosphatic nodular horizons bearing ammonoids, and (4) the upward decrease in grain size.
**Peak Transgression**

It is the major (1st-order) flooding of the whole Triassic. Nevertheless, its precise dating and its possible diachronity is still a matter of discussion among authors. In the basinal areas of the Southern Alps, the peak transgression dated as Fassanian (Chiesense Subzone) is a thin condensed muddy interval (cf. *Chiesense groove* in Brack and Rieber 1986, 1993) documenting a very slow sedimentation rate. At the same time, surrounding carbonate platforms drowned. The Tethys realm and the German basin continued to be connected at the peak transgression, dated as Lower Ladinian (“Chiesense time”). The mixture of Tethyan and Germanic ammonoids recorded in both basins (Urlichs, 1978; Mietto and Manfrin, 1995: upper *Spinusus Zone*), confirms the age of the peak transgression. However, according to Aigner and Bachmann (1992), the peak transgression in the German Basin is placed around the Middle Angolo Limestone.”
peak transgression is recorded within a widespread aggradational alluvial plain environment known as the Teist Formation (Steel, 1993). In the Barents Shelf, the analysis of seismic lines shows that the peak transgression corresponds to a major high gamma-ray log marker of interregional extent (van Veen et al., 1992; Skjold et al., this volume). In the Arctic, the peak transgression could be placed in the Tsekovites varius Zone (for age position of this zone, see Weitschat and Lehman, 1983; Weitschat and Dagys, 1989; M. Muschelkalk (Ev.)

### Regressive Phase

This phase is documented by the strong basinward progradation of terrigenous and carbonate shorelines. In the Dolomites (northern Italy), it is characterized by a spectacular progradation of Ladinian carbonate platforms (Bosellini and Rossi, 1974; Bosellini, 1984) and by an upward increase of turbidites in the basin. In the German Basin, the regressive trend continues upward into the Keuper. The stratal pattern of the upper Hauptmuschelkalk defines the regressive phase of this cycle (Aigner et al., 1996).
and Bachmann, 1992). The same trend is also shown by the forestopping sequences of the Calcaires à Tébratules in the Paris Basin. In the North Sea, the maximum regression within the continental setting coincides with the widespread development of alluvial fan and eolian deposits (Lomvi Formation). This ends with a major exposure surface (Steel, 1993).

**TRANSGRESSIVE/REGRESSIVE FACIES CYCLE 3a**

This cycle is well documented in the Southern Alps, North Sea and Barents Sea.

**Boundaries and Duration**

Both in the Tethys and in the Boreal domains, the lower boundary lies in the uppermost Longobardian. In the Tethys the upper boundary falls in the middle Tuvalian while in the Boreal area it lies in the upper Julian. The cycle duration is 4-5 my.

**Lowermost Boundary**

This maximum regression records an important worldwide documented sea-level drop (Biddle, 1984; Brandner, 1984; Haq et al., 1987). In the Southern Alps, it is represented by a regional unconformity, marked by karst surfaces on top of carbonate platforms and by the presence of huge carbonate megabreccia deposits recording the erosion of slope front and margins (Sarg, 1987; Yose, 1991). In the basinal areas, the maximum regression is placed at the top of the lowstand prograding complex of the last forestopping sequence. In conformable basinal successions, a perfect dating is provided by abundant ammonoid faunas (Regoledanus Subzone). In the German Basin, a regional angular unconformity developed at the top of the Muschelkalk (Aigner and Bachmann, 1992). Likewise, in the Paris Basin, the lower boundary of the cycle corresponds to the sharp, erosional contact of the Lettenkohle above the Calcaires à Tébratules (Goggin and Jacquin, this volume).

**Transgressive Phase**

In the Southern Alps, the Uppermost Ladinian rapid transgression is documented by the retreat of the siliciclastic shorelines and by the moderate prograding of the first Carnian carbonate platforms (Gianolla et al., this volume). In the Germanic domain, including the Rhône and Alsace grabens, the overall transgressive phase (T3) is controlled by intracratonic blockfaulting in which thick piles of salt deposits accumulated. On the margins of these grabens, Carnian clastics and dolomitic deposits quickly backstep lapping out older rocks.

**Peak Transgression**

In the Southern Alps, the peak transgression is dated as Aonoides Subzone; it is documented both on carbonate platforms and in basinal areas (San Cassiano Formation), by an ammonoid-rich level (Allasinaz, 1968; Gnaccolini and Jadoul, 1990; De Zanche et al., 1993; Urlichs, 1994). In the Northern Calcareous Alps, Rüffer and Bechstädt (this volume) have not recognized any major transgressive event. However, the first Raibl Shale, dated as Aonoides Zone, which does not document the drowning of the Wetterstein platform (Bechstädt and Schweizer, 1991), could well include the peak transgression. In the Ardèche area (south-eastern France), the peak transgression could correspond to the Barre Carbonatée Médiane. In the Germanic domain, the most landward extent of marine deposits for the whole Keuper is at the K3 maximum flooding surface (Hauptsteinmergel in Germany and Dolomie de Beaumont in the Paris Basin). This indicates that the T3 peak transgression can be slightly younger in the Germanic area. In the North Sea, the T3a peak transgression is the Lower Carnian marine incursion bracketed between aggradational continental deposits (Middle Lunde Formation). It is here also the most landward extent of marine influence for the whole Triassic. In the Barents Shelf, the peak transgression is defined by a seismic marker and corresponds to siderite-rich western shales assigned to the Stolleites tenuis Zone (van Veen et al., 1992).

**Regressive Phase**

In the Southern Alps, the regressive phase is documented by a stack of shallowing upward progradational cycles. Deposits evolve from shallow-marine mixed siliciclastic/carbonate to peritidal sabkha and continental, the last one becoming more and more areally extended with time. In the Germanic domain, the time equivalent of the regressive phase defined in the Tethys is time equivalent to the transgressive Carnian Gipskeuper Formation. In the North Sea, the regressive phase is indicated by the progradation of terrigenous clastics of the Middle and Upper Lunde Formation pars (Steel, 1993). On the Barents Shelf, the regression phase is documented by the basinward progradation of deltaic deposits.

**TRANSGRESSIVE/REGRESSIVE FACIES CYCLE 3b**

**Boundaries and Duration**

In the Tethys, this cycle ranges from middle Tuvalian to Sev- atian. In the Boreal domain, it ranges from upper Julian to Sev- atian. Its duration is 12-15 my.

**Lowermost Boundary**

The maximum 2nd-order regression documented in the Southern Alps and the Barents Sea coincides with a major and widespread sea-level fall. In the Paris Basin, this sea-level downward shift corresponds to the regional unconformity below the Grès à Roseaux (Goggin and Jacquin, this volume). In the German basin, it is at the base of the Schilfsandstein (Aigner and Bachmann, 1992). Over the Paris Basin and the Alsace and Rhône grabens, we have not interpreted this intra-Late Carnian unconformity as a 2nd-order peak regression, due to the fact that it develops within an overall transgressive phase (T3) with a peak transgression in the Uppermost Carnian (Dolomie de Beaumont and equivalent deposits; Goggin and Jacquin, this volume). In the Southern Alps, the exact dating of the maximum regression is a matter of discussion: Gianolla et al. (this volume) put it within the Raibl Formation while Gaetani et al. (this volume) prefer to place it at the base of the Dolomia Principale. In the Boreal areas, the boundary is dated as uppermost Julian. Therefore, it could appear as older than in the Tethys (Tuvalian) but this is more a problem of Tethyan/Boreal biostratigraphic correlation than a clear diachronism (van Veen, pers. commun.). In the Barents Sea, the Mid-Carnian unconformity is recognized on seismic lines (Skjold et al., this vol-
TRIASSIC SEQUENCE STRATIGRAPHIC FRAMEWORK OF WESTERN EUROPEAN BASINS

Transgressive Phase

The transgressive phase T3b has only been defined as a 2nd-order scale feature in the Southern Alps and the Barents Sea where it corresponds in fact to a single 3rd-order transgressive systems tract. In the Southern Alps, it is a major onlap of the Raibl Formation onto structural highs and inherited morphologies (Gianolla et al., this volume). In the Paris Basin, this major flooding event follows the Late Carnian (Grès à Roseaux) sea-level downward shift and is perfectly recorded by the backstepping pattern of the Dolomie de Beaumont. A similar trend is observed in Germany with the Hauptsteinmergel (Aigner and Bachmann, 1992).

Peak Transgression

The peak transgression in cycle 3b is relatively synchronous over the whole western European basins including the Germanic domain where it coincides with the peak transgression 3. However, precise documentation (i.e., at the zonal and sub-zonal scale) is still lacking, and Tethyan/Boreal biostratigraphic correlation remain uncertain to assess actual synchronity. In the Southern Alps, the peak transgression is dated by ammonoids as Tuvalian (Subbullatus Subzone) within anoxic basins located in eastern Carnia and in the Julian Alps. In the Boreal domain, on the basis of microfaunas (Hochuli et al., 1989; van Veen et al., 1992), the peak transgression is assigned to the Pentasticus Zone (Late Carnian). In the Germanic domain, the peak transgression 3 is dated as Late Carnian with no more precision. In the Barents Sea, it is placed within limestone beds or shales recording high TOC values (Skjold et al., this volume).

Regressive Phase

It can be characterized by two main features: (1) its aggradational character over wide areas which suggests high rates of sediment production and (2) the widespread development of restricted depositional environments that are dolomitic on the Tethyan margins (Haupt dolomit or Dolomia Principate type) or terrigenous continental in the Germanic to Boreal domain. Biostratigraphic information is rather scarce throughout the European basins for that time interval. Therefore, the 3rd-order sequences which build up this 2nd-order regressive phase R3 are difficult to correlate. The maximum regression is not dated by itself, but it is constrained by the following transgressive sediments in the western Southern Alps (Riva di Solto Shales) which are still dated as Upper Norian (Sevatian). The point of maximum regression may also coincide with the major, basin-scale, unconformity known as Early Cimmerian unconformity (Jacquin et al., this volume). This major event relates to the onset of the subsequent 1st-order cycle. That unconformity has been dated by magnetostratigraphy in the North Sea as uppermost Norian (Beyer, 1995) near the top of the Lunde Formation. Therefore, the Raethian represents the onset of the transgressive portion of the 1st-order Ligurian cycle (Jacquin and Graciansky, this volume).

CONCLUSION

Within the western European Triassic, four transgressive/regressive facies cycles can be recognized. They are interpreted as mainly caused by tectonic mechanisms, but the relative synchrony of the main trends would also suggest long-term eustatic influence. These four cycles developed as the basins evolved twice from extensional driven syn-rift phase through thermal and loading sag phase to relaxation and stabilization phase.

Transgressive/regressive cycle 1 (Scythian) is the first Triassic syn-rift cycle. It is characterized by marked differential subsidence resulting in laterally highly variable depocenters, hinterland rejuvenation and strong influx of coarse siliciclastics (Buntsandstein). Transgressive/regressive cycle 2 (Anisian and Ladinian) is the first Triassic thermal and loading sag phase with broad areas of uniform subsidence, resulting in laterally uniform thicknesses and facies. Carbonate platforms prevailed over most of Europe establishing northward over the Muschelkalk domain. The major (1st-order) peak transgression for the whole Triassic is reached at that stage in conjunction with a long term eustatic sea-level rise, uniform subsidence and dramatic decrease of the terrigenous supply. Time equivalent basinal facies are typically organic-rich shales. The first stabilization phase separating the two Triassic geodynamic evolution took place during the latest Ladinian resulting in reduced rate of subsidence in the basins and thin depositional sequences with proven or probable hiatuses.

The transgressive/regressive cycle 3a (mainly Carnian) is the second Triassic syn-rift cycle. It is characterized, as the first one, by marked differential subsidence, resulting in renewed coarse clastic supplies and highly variable lithologies. The major Car3 (Late Carnian) sea-level downward shift and subsequent flooding relates to the climax of the extensional phase.

The transgressive/regressive cycle 3b (mainly Norian) is the second Triassic thermal and loading sag phase giving a new laterally uniform interval of sabkha-type sediments all over Europe. Near the end of the cycle, tongues of progradation clastics merge towards the Late Cimmerian (latest Norian) unconformity that is tectonically enhanced at various places.

The difficulty to build long distance stratigraphic correlations across Triassic basins comes from (1) the poor biostratigraphical control as a result of the development of widespread terrigenous clastics sedimentation in the Germanic domain and (2) the poor correlation between the available Boreal and Tethyan cephalopod scales due to the scarcity of interbasinal communication between the Nordic and the Southern realms.

Twentythree depositional sequences are now documented from the western European basins which contrast with the eleven cycles of the Haq et al. (1988) chart. This provides an idea of the increasing resolution we can gain from the integration of the various disciplines that contribute to the sequence stratigraphic approach.

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TRIASSIC SEQUENCE STRATIGRAPHY OF THE SOUTHWESTERN BARENTS SEA

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ABSTRACT: Regional studies of Triassic sequences on the southwestern Barents Shelf (20–32°E, 71–74°N) revealed units that correlate seismically for hundreds of kilometers. Seismic character analysis identifies three Lower-Middle Triassic units with a composite progradational pattern, indicating sediment supply from the southeast (Induan, Olenekian and Anisian time). Fluvial, deltaic and marine deposits are predicted and located relative to paleo-coastlines. Shallow-marine conditions with limited vertical aggradation prevailed during the Ladininan time. A major shift in basin configuration took place in early Carnian time, and during the Late Triassic Epoch, the main sediment supply was from uplifted areas to the northeast, leaving sequences of marine to fluvial sediments on a tilted basin margin.

The predictions based on this seismic stratigraphic approach proved valuable when correlating and evaluating well information. Building a stratigraphic framework based on seismic stratigraphy, detailed analysis of well-logs and biostratigraphy combined with regional correlations in the Arctic basins led to the identification of a succession of five 2nd-order cycles, each consisting of several sequences.

Second-order cycle boundaries seem to correspond to changes in the tectonic regime/lithospheric stress, affecting the regional basin configuration; 2nd-order cycles correspond to T/R (transgressive/regressive) cycles, whereas sequences may be classified as 3rd-order cycles. A number of the sequences defined in this paper may well be of eustatic origin.

INTRODUCTION

This paper is based upon our regional studies of the Triassic sequences in the Barents Sea. The study area is located between 70° and 75°N, 18° and 32°E and includes structural elements such as the Nordkapp and Hammerfest Basins, the Bjarmeland and Finnmark Platforms and the Loppa and Stappen Highs—the last one including Björnøya (cf. Gabrielsen et al., 1990; Fig. 1). Triassic deposits are buried too deeply to be investigated in the Tromsø and Björnøya Basins.

All available seismic data have been interpreted stratigraphically, and regionally identifiable seismic units, as well as seismic character (i.e., seismic facies), have been defined. Their boundaries in general are characterized by truncation or toplap followed by onlap, or regionally persistent high-amplitude reflectors. Results of the seismic stratigraphy study were used to predict the regional distribution of lithofacies (i.e., of reservoir and source-rock-prone areas). Exploration wells and stratigraphic boreholes drilled during 1986–1990 have confirmed and modified the geologic model, resulting in a more detailed subdivision enabling the definition of a number of sequences.

Presently, we propose to subdivide the Triassic succession into five 2nd-order sequences, T1 to T5, beginning and terminating at events that are of regional importance and each of which consist of several sequences. A major aim of this integrated study is to develop a sequence stratigraphic model for the area (i.e., to divide the strata into genetically related systems tracts), thereby aiding the regional predictions of lithofacies for exploration purposes. The recognition of sequences is based on the concepts of Van Wagoner et al. (1990). We have adopted the sequence stratigraphic terminology throughout this paper (i.e., LST = lowstand systems tract, TST = transgressive systems tract, HST = highstand systems tract, MFS = maximum flooding surface).

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In this respect palynostratigraphy is of crucial importance to identify trends that indicate retrogradation, maximum marine influence, or progradation, especially in seemingly monotonous successions. Ideally, a sequence should be delimited at its base and on its top by regional unconformities. A proper qualifica-
tion of part of the sequence boundaries is still uncertain, however, due to the small number and widely spread nature of the exploration wells.

The sequences are related to the chronostratigraphical framework of the Boreal Triassic. Van Veen et al. (1993) calibrated the ammonoid zonation on Svalbard to palynozones defined in Barents Sea wells, which again are tied to our seismic stratigraphic framework. On Figure 2, the palynozonation scheme is put alongside the sequence stratigraphic subdivision (geochronology after Forster and Warrington, 1985; ammonoid zonation from Svalbard according to Weitschat and Dagys, 1989, correlated to ammonoid-zonation according to Tozer, 1984). The global onlap curve is that of Haq et al. (1988).

Only a few publications have hitherto dealt with regional aspects of the Triassic in the Barents Sea. These include Rønnevåg et al. (1982), who interpreted the Triassic from seismic data on a regional scale; Jacobsen and Van Veen (1984) described the paleogeographical development offshore Mid-Norway and the Western Barents Shelf and compared with Svalbard and East Greenland. Berglund et al. (1986) described the evolution of the Hammerfest Basin in particular, but also the general development of the Triassic in the area. A series of publications by Embry (1989), Mørk et al. (1989), Johannessen and Embry (1989), and van Veen et al. (1993) are the first to deal with regional correlations of Triassic sequences in the Arctic.

The lithostratigraphical framework follows the recommendations of Dalland et al. (1988).

The Barents Sea area was, during the Triassic Period, an intracratonic basin with the sediment source area of today’s Siberia in an easterly position (cf. paleogeography of Ziegler, 1988a).

**SEISMIC AND SEQUENCE STRATIGRAPHY**

In the following discussion, sequences are defined by the 2nd-order sequence to which they belong (e.g., T1, followed by a number, such as T1–1).

**Pre-Triassic Events**

Many of the seismic units and, hence, the sequences discussed in this chapter onlap the Loppa High. A phase of extension and rifting was most intensive at the Permian/Triassic transition, resulting in the uplift and tilting of the Loppa High (cf. Ziegler, 1988b). Subsequently, the Loppa High has acted as the western limit of the basin during the first half of the Triassic Period, and truncation/onlap relationships are particularly well developed in the vicinity of this high.

Subsidence near the basin margins resulted in the local deposition of Late Permian coarse clastics, specially at the southeastern margin of the Hammerfest Basin, where deposition was

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**Fig. 2.—**Triassic sequence stratigraphy of the western Barents Sea. Ammonoid zonation from Svalbard according to Weitschat and Dagys (1989), calibrated to composite Triassic ammonoid zonation. Chronology after Forster and Warrington (1985). Global onlap curve after Haq et al. (1988). Note uncertainty of Anisian/Ladinian boundary in Boreal vs. Tethyan Triassic. Scale change at Middle/Late Triassic boundary. Purple: Anoxic/Condensed deposits; Blue: Open Marine/Oxic deposits; Yellow: Shallow/Marginal Marine deposits; Green: Delta top/Coastal Plain deposits; Brown: Fluvial/Alluvial Plain deposits.
more or less continuous at the Permian/Triassic transition. Towards the east, near the Finnmark coast, these deposits are absent and represented by a hiatus. Our Triassic sequence stratigraphy starts above this event.

**Induan Sequences (T1–1, T1–2)**

Two Induan sequences are defined. To the southeast of the Nordkapp Basin, the oldest Griesbachian T1–1 sequence is developed as a series of clinoforms, downlapping towards the north-northwest (Fig. 3). The original slope on the clinoforms is on the order of 100 m/km (approximately 6°), and they prograded into a basin with 200- to 300-m water depth. The top of this clinoform package shows an irregular upper contact. Towards the Norwegian coastline, the sequence has been eroded by the Base Quaternary unconformity.

The next, the Dienerian T1–2 sequence, onlaps the previous sequence south of the Nordkapp Basin (Fig. 3) and increases markedly in thickness in the Nordkapp Basin. Mounded seismic patterns occur at the base. The seismic character in the remaining areas generally shows a parallel pattern, clearly thinning towards the Bjarmeland Platform in the north, and rapidly thinning by onlap on the structural highs (e.g., the Loppa High; Fig. 4).

The composite seismic facies map in Figure 5A indicates a shift in sedimentation from the Tatarian Age (latest Permian) from northward progradation in the Hammerfest Basin to Griesbachian northwestward progradation offshore Finnmark East. Finally, the Dienerian developed as a basin fill with areas of mounded seismic patterns.

**Interpretation.**

Sequences T1–1 and –2 are recognized on the basis of well information and biostratigraphy within the Induan Stage (Fig. 2). The first sequence, T1–1, has been penetrated in various wells, both in the Hammerfest Basin and further east. Marine microplankton is present in the basal parts and gives way to a terrestrial palynoflora towards the top. The TST is rather thin in the Hammerfest Basin and the western part of the study area, but seems to expand offshore East Finnmark (Fig. 6A). The Upper Griesbachian progradation represents a well-developed HST.

The next sequence, T1–2, of Dienerian age, onlaps the underlying Griesbachian sequence, and is mainly fine grained in most of the wells; marine microplankton is rather common in this unit. The irregular upper contact of the underlying highstand systems tract, interpreted as incised valleys, as well as the presence of seismic mounds, which may indicate a lowstand development, suggest a type 1 sequence boundary. In the eastern part of the study area, the MFS is recognized high in the sequence, indicating an extensive TST development (Fig. 6A, well E). In wells west of the Loppa High, as well as on Bjørnøya, the T1–2 sequence is the first to onlap the underlying Paleozoic strata.

The paleogeographical development of the Griesbachian and Dienerian sequences is shown on figure 7, with the beginning of the sea-level fall inferred at the base of sequence T1–2.

**Smithian Sequences (T1–3 to T1–5)**

Contrary to the underlying Induan sequences, Smithian deposition blankets most of the study area and consists of three
sequences. T1–3 is thinly developed and characterized by a poorly-defined progradational pattern in the southeast and generally by a parallel, onlapping pattern elsewhere.

In the overlying T1–4 sequence, a distinct area with oblique clinoforms has been mapped that downlaps towards the west-northwest (Figs. 4, 5B). Data quality worsens towards the Bjar-meland Platform, but a general thinning and more parallel internal patterns have been observed in that direction.

The upper Smithian T1–5 sequence is also defined by a well-developed clinoform belt (Figs. 4, 5B) and shows a westward onlap on the underlying deposits near the eastern flank of the Loppa High (Fig. 8).

Interpretation.

On the basis of our well information and biostratigraphical data, at least three Smithian sequences can be recognized, T1–3 to –5. The oldest one, T1–3, is only represented in a distal development in the easternmost wells in the Norwegian Barents Sea; the TST is condensed. Marine microplankton is ubiquitous, maximum flooding surfaces are marked by maximum abundances; amorphous organic matter (AOM), indicating relatively anoxic conditions of deposition, is predominant, though the TOC (total organic carbon) is often low. We tentatively correlate T1–3 to the uppermost Dienerian and the lowermost Smithian strata, partly based on palynological data. The relatively rapid succession of flooding surfaces as well as their rather distal development explain the high-amplitude, high-continuity seismic pattern at the base of the Smithian. The progradational, shingled pattern on Figure 5B is interpreted as the HST of T1–3.

The next sequence, T1–4, has been penetrated in several wells, and especially near the Nordkapp Basin a several-hundred-meter-thick succession of distal-marine to continental deposits is recorded. Marine microplankton is common in the lower 100 m of this sequence, indicating open-marine conditions. As in the underlying T1–3 sequence, the TST is condensed, and the sequence is dominated by its HST. These features indicate a sudden change in clastic input during the deposition, causing rapid lateral progradation and filling of the Nordkapp Basin. Palynological assemblages from a mangrove-like vegetation are abundant within the Lower Triassic deposits; this may indicate that Smithian coasts were of low energy. Terrestrial red beds, virtually devoid of organic matter, have been recorded from the top of this prograding package of the HST. Figure 5B illustrates progradation of the HST of T1–4. Figure 7 (Smithian) depicts the paleogeography at the final stage of this sequence.

On the seismic profiles, such as Figure 4, there is some evidence of a LST in the lower part of the T1–5 sequence. In wells, in the western part of the study area, the base of sequence T1–5 shows a rapid increase in marine microplankton, indicating a well-developed TST. A rapid lateral development can be observed, from maximum marine influence in the west to virtually nonmarine in the east. A rapid progradation of the HST concludes this sequence. In the Hammerfest Basin, this event is recognized as a level of minimum marine influence.

Spathian Sequences (T2–1, T2–2)

The Spathian strata are developed over the entire area, becoming extremely thick in parts of the Nordkapp Basin (e.g.,
FIG. 5.—Seismic facies maps of the following strata: (A) Induan, (B) Smithian, (C) Lower Spathian interval with high-continuity strong-amplitude seismic reflections, (D) Lower Spathian prograding unit. Sequences are shown (e.g., T1–1). Legend on Figure 10.
Two sequences are recognized, T2–1 and T2–2. The lower part of sequence T2–1 is developed as a wedge of local extent (Fig. 5B), onlapping the top of sequence T1–5 (Figs. 4, 8). This is overlain by highly continuous seismic reflectors with strong amplitude in the central parts (Fig. 5C). A small clinoform belt appears in the central, southern area (“1” on Fig. 5D). A second, more extensive clinoform belt with a grossly north-northeast orientation can be observed slightly further to the west (“2” on Fig. 5D). Sequence T2–1 wedges out along the eastern flank of the Loppa High.

Sequence T2–2 onlaps the underlying clinoforms from the west-northwest and subsequently forms a thin sheet covering most of the mapped area. Towards the Loppa High, the top is apparently truncated by the base Anisian marker, partly by the intra-Upper Anisian and finally, truncated by base Tertiary strata.

Interpretation.—

A major transgression marks the Smithian-Spathian transition, in the wells. There appears to be no indication of the seismically-defined LST at the base of sequence T2–1, and the TST is poorly developed (Fig. 2). Condensed deposits of Spathian age have been recognized in at least one well, confirming that the high-amplitude seismic reflections in T2–1 represent the distal development of a TST, with maximum flooding in early Spathian time. This flooding surface forms a major seismic
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The nature of the sequence boundary between T2–1 and T2–2 is uncertain. The onlapping relationship near the eastern flank of the Loppa High may be interpreted as a type 1 sequence boundary. In one well, an abrupt lithological change is recorded at this level, which may indicate the development of an incised valley.

Anisian Sequences (T2–3 to T2–6 and T3–1)

Within the Anisian Stage, a complex seismic stratigraphic configuration is discerned that allows the recognition of five sequences. Rasmussen et al. (1993) present a detailed discussion of the sequence stratigraphy of this interval.

The base of the lowermost sequence (T2–3) truncates underlying deposits along the structural highs. A seismic unit with strong amplitude reflections characterizes the lower part, the top of which forms a downlap surface for overlying clinoforms (Fig. 8). The area of strong seismic amplitude is clearly delimited towards the east and west (Fig. 10A). The layer wedges out along the eastern flanks of the Loppa High. Onlap on the...
Fig. 7.—Early Triassic paleogeographic reconstructions.
FIG. 8.—Northwest-southeast seismic profile on the eastern flank of the Loppa High showing onlap of the earliest Triassic T1-sequences, a well developed LST of T2–1, both HST and LST prograding Anisian wedges (T2–3, T2–4, and T2–5), and westward-thickening post-Ladinian sequences.

FIG. 9.—Seismic profile in the northern part of the Nordkapp Basin, showing the very thick Spathian rim-syncline deposits (sequences T2–1 and T2–2).
Fig. 10.—Seismic facies maps of the following strata: (A) Lower Anisian, (B) Middle Anisian, (C) Ladinian-Carnian. Sequences are shown.

Legend:
- Onlap
- Downlap
- Parallel
- Mounded
- Oblique
- Shingled
- Divergent
- Chaotic
- Progradational belt
- Amplitude
- Continuity

F/GEO/1014x/MacF/LJS/05.92

Fig. 10.—Seismic facies maps of the following strata: (A) Lower Anisian, (B) Middle Anisian, (C) Ladinian-Carnian. Sequences are shown.
underlying and apparent truncation of the overlying boundaries occur towards the Loppa High (Fig. 10A, B).

The basal part of the next sequence (T2–4) consists of a distinct clinoform belt, of limited extent. The overlying sequence (T2–5) onlaps the uppermost clinoforms of T2–4 near the eastern flank of the Loppa High, developing into a parallel seismic internal pattern further east. Apparently, the top of T2–5 becomes truncated by an intra-Anisian seismic marker near the Loppa High. The uppermost Anisian strata contains two sequences (T2–6 and T3–1) that show a parallel internal seismic pattern and that thin towards the Loppa High. These are recognized by well data.

Interpretation.—

The five Anisian sequences reflect the stepwise infill of the Barents Sea east of the Loppa High, before the onset of a significant transgression during latest Anisian and Ladinian times.

Well information supports the interpretation of the high-amplitude seismic data as reflecting condensed parts of the TST of T2–3, which are of source-rock quality. The base Anisian seismic reflector originates mainly from the first flooding surface of T2–3. Rasmussen et al. (1993) interpret a LST at the base of sequence T2–5. The presence of another LST is suspected at the base of T3–1, based upon the severe backwards truncating nature of the sequence boundary (Fig. 2), generating the intra-Anisian seismic marker.

The development of Early and Late Anisian paleogeographies is visualized on Figure 11.

Ladinian Sequences (T3–2, T3–3)

Ladinian deposits onlap and partly blanket the eastern flank of the Loppa High. The succession exhibits parallel to transparent and reflection-free seismic patterns throughout most of the study area. Some westward-dipping clinoforms appear on the western flank of the Loppa High (Fig. 10C). In the Nordkapp Basin, a mounded seismic facies has been observed in the rimming synclines. The Ladinian is the first deposit showing a continuous westward thickening; it onlaps and thins towards the East Finnmark Platform.

Interpretation.—

A very rapid and radical transgression, starting in Late Anisian time with T3–1, pushed back the coastline far towards the east. As a result, the Ladinian stage is dominated by fine-grained, open-marine deposits over most of the mapped area. Marine microplankton ranges from common to abundant. An important paleogeographical element: Ladinian sediments blanket the high, only slightly thinning by onlap. Yet, the presence of clinoforms west of the Loppa High indicates that its structural control still persisted during most of the Ladinian time. The paleogeographical reconstructions (Fig. 11; Late Anisian/Early Ladinian; Late Ladinian) summarizes our observations, with marine sedimentation over most of the study area. The presence of an area of anoxic sedimentation west of the Loppa High, predicted by our model, was confirmed during subsequent drilling.

A sequence stratigraphical subdivision of the Ladinian stage is tentative at present. Well-log analysis of Ladinian deposits in the Hammerfest Basin near the Troms-Finnmark Fault Zone reveals two coarsening-upwards sequences, T3–2 and T3–3 (Fig. 6B, wells A & B). The MFS of sequence T3–2 is a major high-gamma-ray log marker in central parts of the Barents Sea. The log marker coincides with the base Ladinian seismic marker and appears to be of interregional extent. The HSTs of both sequences are dominated by fine-grained clastics; coarsening-upwards trends do not reflect actual coarse grain-size.

We tentatively recognize an uppermost sequence T3–4. A sand-rich interval has been recorded in wells to the northwest of the Loppa High, which may represent a LST of this sequence and may correspond to the dipping clinoforms. In the wells near the Nordkapp basin, T3–4 may be partially eroded below the base of the Carnian strata.

Carnian Sequences (T4–1 to T4–3)

Based on well data, the Carnian Stage is divided into three sequences. The sequence boundaries have tentatively been identified on the seismic profiles. A base Carnian marker has been observed throughout the area. The lowermost Carnian sequence, T4–1, exhibits parallel seismic character throughout in the lower part and common channel-like features in the upper part; it onlaps the Base Carnian seismic marker eastwards (Fig. 8).

T4–2 is characterized by a chaotic to parallel internal seismic pattern. It thins by onlap towards the east, and a number of parallel, high-continuity reflectors occur in the middle of the sequence towards the west (Fig. 8).

The T4–3 sequence shows a parallel seismic character with some channel-like features, thinning by onlap towards the east-southeast, and an increasing number of parallel, high-continuity, high-amplitude reflectors towards the west (Fig. 8).

Interpretation.—

The most important feature of the Carnian sequences is the occurrence of a mid-Carnian unconformity, with a hiatus of increasing duration towards the southeast (Fig. 2). The recognition of this unconformity is based upon seismic data combined with well observations. Another feature is the shift in basin configuration, with subsidence mainly concentrated over and west of the Loppa High, with a northwestward thickening of the Carnian strata. Due to this shift, the drainage pattern changed towards a northeast- to southwest-oriented subaxial drainage pattern (Fig. 11).

The oldest Carnian sequence (T4–1) shows its optimal development towards the basin center, where deltaic deposits of earliest Carnian age overlie marine Upper Ladinian deposits. This lithological change is thought to reflect a sudden increase in sediment supply, resulting in a considerable fall in relative sea-level. The MFS of this sequence is marked by shales rich in siderite, which generate a seismic marker to the west of the study area (Carnian III). Microplankton and microfaunas are common around this marker. The subsequent HST is characterized by a rather thick succession of deltaic deposits, which contain thick sandstone intervals and some coals in the wells near the Loppa High. The Early Carnian paleogeography is sketched in Figure 11.

The overlying sequence T4–2 again is developed optimally in the vicinity of the Loppa High. The TST of this sequence is
FIG. 11.—Middle Triassic paleogeographic reconstructions. See Figure 7 for legend.
thin, whereas the MFS is not characterized by appreciable condensation but is only indicated by the maximum abundance of marine microplankton and microfauna. The intra Carnian II seismic marker is related to the MFS in the west, whereas the eastward continuation is determined by contemporaneous erosion along the basin margin. The HST accommodates the influx of erosional products and is characterized by repeated coarsening upward cycles (Fig. 6B, wells B and C), presumably representing prodelta deposits. In the Hammerfest Basin, sandstones at the top of this sequence were proven to be gas-bearing.

The T4–3 sequence is developed with a TST that clearly thins by onlap on the underlying unconformity towards the east. In the western, more distal part, a large number of limestones occurs; the most prominent of these coincides with the intra-Carnian I seismic marker of strong amplitude (Fig. 8). The limestones, shallow marine-oolitic banks, were formed as the sea transgressed the tilted, southeastern basin margin. Several distinct microfaunas are associated with these limestones (Hochuli et al., 1989; van Veen et al., 1993). Sediments around the MFS form a laterally persistent horizon with increased TOC values, which locally develop into source rocks. Figure 12 shows the paleogeographic development at maximum flooding. The HST, thickening markedly towards the west, is developed as a succession of paralic sediments, with abundant coals and palynological assemblages that reflect the swamp-type vegetation. Marine transgressions are occasionally recorded within this interval. In more proximal eastward settings and towards the top in the west, the HST grades into alluvial redbeds. We suspect that several more sequences are hidden in this sequence, but the present database does not allow a further subdivision.

**Norian Sequences (T5–1 to T5–3)**

The Norian succession thins rapidly towards the east-southeast by onlap. It shows a parallel seismic pattern, but with a stronger amplitude and continuity, compared with the underlying Carnian strata. Upwards, the internal pattern on the seismic data becomes less continuous. A base Jurassic marker is mapped throughout the area and terminates the Triassic sequences. The Norian becomes truncated over the Loppa High by the base Quaternary unconformity (Fig. 8). Seismically, the identification of three sequences is difficult to document.

**Interpretation.—**

The Carnian/Norian transition witnesses a profound change in depositional circumstances, as reflected in mineralogic maturity of the sandstones on Svalbard and in the Barents Sea (cf. Mork et al., 1982; Bergan and Knarud, 1993). More mature sandstones grading to pure quartz arenites are developed in the Norian and younger deposits. Palynological information suggests a change in climate, from arid and warm to temperate and humid, during Norian/Rhaetian time, which may well explain the degree of chemical weathering and, hence, the change in maturity. In addition, an overall lower sedimentation rate is interpreted.

In general, Norian deposits are not well developed and are quite condensed east of the Loppa High, except for the lowermost part, but becomes more extensive to the west. As a result, Norian sequences do not show an optimal development east of the Loppa High, except in places where accommodation space was created due to local structural events, such as in rim syn-

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**Late Carnian, (T4–3)**

**Early-Middle Norian, (T5–1)**

**Fig. 12.—** Middle to Late Triassic paleogeographic reconstructions. See Figure 7 for legend.
clines. A tentative subdivision of the Norian/Rhaetian strata into at least three sequences has been recognized in the Barents Sea.

The first sequence, T5–1, is completely developed in the Hammerfest Basin as well as west of the Loppa High. The sequence boundary is marked by a sudden influx of quartz sandstones, indicative of a type 1 sequence boundary (see Fig. 6B, wells A and B). The MFS of T5–1 is identified with the seismic base Norian marker. Figure 12 shows the early Norian paleogeography, with maximum progradation at the end of the HST.

Marine microplankton is virtually absent in the overlying sequences and recognition of systems tracts is severely hampered. Palynological information suggests the presence of a new sequence, T5–2. Presently, well data clearly show that the base of this sequence erodes into the underlying succession towards the east and coincides with the intra-Norian seismic marker. The last sequence, T5–3, with Middle Norian age palynological assemblages, is heavily condensed. The base Jurassic seismic marker marks the beginning of another, Jurassic supercycle. Thus, the available data suggest a considerable reduction of accommodation space during Middle and Late Norian/Rhaetian time.

SIGNIFICANCE OF SECOND-ORDER SEQUENCES

Following van Veen et al. (1993), the Triassic sequences have been grouped into 5 2nd-order sequences, a subdivision that partly differs from similar sequence stratigraphic studies in the Sverdrup Basin and on Svalbard (e.g., Embry, 1989; Mørk et al., 1989). Our 2nd-order sequences compare in order to sequences described by Embry (1989) from the Sverdrup Basin, Arctic Canada and to the transgressive/regressive (T/R) cycles defined by Mørk et al. (1989) on Svalbard. Our 2nd-order sequences differ in stratigraphical extent, as well as in concepts from the T/R cycles of Mørk et al. (1989).

Because the Barents Sea acted as an intracratonic basin during the Triassic, stress changes may have had a different response here compared with passive margins, and consequently, the following explanation has to be considered a working hypothesis. The boundaries between 2nd-order sequences in the Barents Sea are taken at horizons, where we recognize fundamental changes or an anomalous distribution of the systems tracts, which are mainly interpreted to reflect changes in lithospheric stress.

Permian-Triassic Boundary

The Permian-Triassic transition exemplifies the depositional response to an extensional regime (cf. Ziegler, 1988b) and its relaxation (cf. Cloetingh, 1986, 1988). Both on Svalbard as well as in the western part of the Barents Sea, the basin floor became uplifted and was subject to erosion (e.g., Mørk et al., 1989). Near the Troms-Finnmark Fault Zone, however, sedimentation continued during latest Permian and earliest Triassic times, and a thick succession of latest Permian quartz sandstones was deposited, which were transgressed during earliest Triassic deposition. A similar situation apparently existed at the northwestern margin of Spitsbergen (Steel and Worsley, 1984).

During deposition of T1, the relatively fine-grained input in the Hammerfest Basin, compared with the voluminous infill in the Barents Sea, suggests that erosion of the Hercynian moutain chains to the east (Ural and Novaja Zemlja) was the main source of sediment (cf. Ziegler 1988b). During latest Permian and Early Triassic times, basalt volcanism was widespread in the western Urals (Ulimshiek, 1982, 1985). An extensional regime east of the Ural Mountains may well have induced uplift and volcanism to the west, culminating during deposition of the Smithian T1–4 sequence. The sediments of this sequence show a remarkable proportion of volcanic minerals in the study area.

T1/T2 Boundary.—

We consider the T1/T2 boundary to be a major feature in the Barents Sea area. The earliest Spathamian T2–1 sequence has a well-defined LST in its lower part, based on our seismic studies, and is a type 1 sequence boundary. Relatively coarse-grained lithologies appear at this level in the Hammerfest Basin as well as on Svalbard. Support for a period of tectonic unrest is the simultaneous onset of halokinesis in the Nordkapp Basin (Fig. 7, early Spathian).

A comparable tectonic phase has been documented in western Europe by the “Hardegsen Unconformity.” Thus the T1/T2 boundary may well correspond to a significant change in lithospheric stress, presumably of a compressive nature (cf. van Veen et al., 1993).

T2/T3 Boundary.—

The boundary between T2–2 and T2–3 is interpreted to be a type 1 sequence boundary. The main differences between the Anisian and the underlying Spathamian sequences are the well-developed TSTs and the poor development of the HSTs in the Anisian.

During deposition of the 2nd-order sequence, T2, supply from the southeast was high in the study area, since well-defined TSTs are recognized at the base of most sequences. At the T2/T3 boundary, however, the predominantly progradational trend of T2 reverses to a retrogradational development, establishing a predominantly fine-grained deposition in T3. It is tempting to suggest a relationship to the end of spreading east of the Urals (Aplonov, 1988), inducing relaxation of the compressional stress regime west of the Ural mountains, resulting in regional onlap. On Svalbard, this marks the beginning of a period with the richest source-rock deposition (Blanknuten Beds, Botneheia Member, Mørk et al., 1982).

T3/T4 Boundary.—

In contrast, the Middle/Late Triassic transition is emphasized by a return to coarse clastic sedimentation, resulting in a considerable change in paleogeography. Several pieces of evidence point toward a major phase of compression as the cause of the observed features. These include uplift and subsequent erosion of the basin margin, culminating in the mid-Carnian unconformity and the inversion of the Loppa High, resulting in a deep Carnian basin in the western Barents Sea. The orientation of the stress field generating this compression is uncertain. An apparently oblique or perpendicular orientation of deeply incised channels in the Lower Carnian strata on northwest-southeast-oriented seismic profiles, suggests a northeast-southwest orientation of the drainage system. Mørk et al. (1982) suggest a similar northeast-southwest orientation of deltaic progradation during deposition of the DeGeerdalen Formation on Spitsbergen and Edgeøya. This suggests that uplift of areas towards the north (Franz Josef Land; Lomonosov High sensu Ziegler, 1988b) was the main source for the clastic supply.
DISCUSSION AND CONCLUSIONS

1. Application of sequence stratigraphic methods supports a subdivision of the Triassic system into five 2nd-order sequences, T1 to T5, as illustrated in Figure 2. A minimum of 21 sequences can be recognized, each with an average duration of 1–2 my.

The 2nd-order sequence boundaries are marked by inter-regional events that may reflect large scale changes in intra-plate stress. Notably here are the Permian/Triassic transition, the Smithian/Spathian boundary, the Anisian/Ladinian transition and Ladinian/Carnian and possibly the Carnian/Norian boundaries. The existence of a mid-Carnian unconformity is recognized, which represents the culmination of a compressive phase starting at the Ladinian/Carnian boundary.

2. The results of our study indicate that, whereas interregional correlation may have achieved the level of 2nd-order sequences or T/R cycles (cf. Embry 1989; Mørk et al. 1989), details are lacking for correlation at a sequence level. Sequences in the Barents Sea are difficult to recognize on Svalbard, due to differential sediment supply (cf. van Veen et al., 1993; Johannesen and Embry, 1989). During the Early and Middle Triassic Epochs, the study area received enormous quantities of sediment (e.g., the Spathian T1–4 sequence reached a thickness of several hundred meters during approximately 1 my). Both the Hammerfest Basin and Svalbard (correlated by, e.g., Jacobsen and van Veen, 1984; van Veen et al., 1993) are considered relatively distal in this respect.

3. Both on Svalbard and in the Barents Sea east of the Loppa High, accommodation space was progressively reduced during deposition of 2nd-order sequence T5.

4. On Figure 2, coastal onlap curves based on this study and that of Haq et al. (1988), are presented for comparison. The correlation is quite good at the level of 2nd-order cycle boundaries. Our Barents Sea curve shows greater detail and shorter duration for most of the sequences; an average of 2 my for each sequence, varying between 1.1 my during deposition of T2 to 5.0 my for T5. Published data from the Lower Triassic Series are hardly sufficient for comparison, and correlations between Svalbard and the Alpine Middle and Upper Triassic Series are subject to considerable debate among biostratigraphers. It is interesting to note that De Zanche et al. (1993) for example have described 3rd-order sequences from the Dolomites in Italy that are comparable both in number and magnitude to the 3rd-order sequences of this study.

The results of this study, in our opinion, do not resolve the question whether glacio-eustasy or intra-plate stress was the main causal mechanism for generating the sequences recognized in the Triassic System in the Barents Sea. Intra-plate stress may well have been responsible for generating 2nd-order sequence boundaries, corresponding to 2nd-order cycles in the sense of Vail et al. (1990), with an order of duration of 5–10 my.

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REFERENCES


A SEQUENCE STRATIGRAPHIC FRAMEWORK OF THE MARINE AND CONTINENTAL TRIASSIC SERIES IN THE PARIS BASIN, FRANCE

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ABSTRACT: A sequence stratigraphic framework has been established for the marine and continental Triassic series in the Paris Basin, France. It has been subdivided into one 1st-order major transgressive/regressive cycle, two 2nd-order transgressive/regressive facies cycles and eleven 3rd-order sequence cycles. This has been attained by well log correlation using gamma-ray-ray, sonic, neutron, density, resistivity, and SP wireline logs as the main correlation tools and calibrated core sections. Subsurface correlation was achieved by the stacking pattern analysis of genetic units recognized from well logs. Correlation with the Germany Triassic, the Southern Alps and the Dolomites has been carried out to improve accuracy in dating the succession in the Paris Basin. Each 3rd-order sequence has been subdivided into systems tracts and described with respect to its facies distribution, development and geometry. This framework has made an attempt to provide a chronostatigraphic subdivision for extensive thick fluvial sand deposits situated in the western part of the basin. By constructing a network of transects across the basin, thereby allowing three dimensional control on sequence architecture, it is possible to identify at least three stages in basin development. The depocenter migrated progressively to the west with time due to the progressive breakup of the supercontinent Pangea in the east and the subsequent opening of the Tethys seaway. The cycles are influenced by active tectonics linked to these events. Subsidence analysis across the basin supports this theory and integrates the 2nd-order cycles defined into the basin evolution.

INTRODUCTION

Triassic strata were deposited in the intracratonic Paris Basin, with a thickness of Triassic strata of up to 600 m in the basin center. Structurally, the Paris Basin is delimited by the Variscan massifs of Armorica in the west, the Massif Central in the south, the Vosges in the east and the Ardennes in the northeast (Fig. 1). Basin development was related to the opening of the Tethys sea in the southeast, due to the breakup of the supercontinent Pangea since Permian times (Ziegler, 1988). The basin did not become an independent depocenter until Carnian times, when the subsidence center was repositioned westward. Before this change in depocenter, the Paris Basin could be described as an embayment of the Germany Basin, with continental and marginal marine deposits. During limited connections with the Tethys in Ladinian and Carnian times, marine carbonates were deposited (Fig. 2).

In this study, the tripartite division of the marine and continental Triassic series of the Paris Basin has been reinterpreted using the concepts and techniques of sequence stratigraphy. The aim of this paper is to present this new subdivision, to describe each sequence up to 3rd-order frequency and to discuss their significance in the basin development. We have correlated from Germany to the English Channel, north towards Luxembourg and south to the Bresse region (Figs. 3, 4, 5). This subdivision has been correlated with other European basins, providing a large-scale regional synthesis over northwest Europe (Gaetani et al., this volume). A transect from the Paris Basin to the Bresse region (Figs. 1, 4) has been included in this publication, to illustrate the similarity in development of sequences of the Triassic succession in both regions, which until Carnian times, were separated by the Burgundy High.

METHODOLOGY

The sequence stratigraphic principles described by Vail (1987), Posamentier and Vail, (1988) and Van Wagoner et al. (1990) have been applied to the Triassic succession of the Paris Basin at the high resolution scale of wireline log data over the entire basin, to a database of 250 well logs. The succession has been subdivided into 1st- and 2nd-order transgressive/regressive cycles (Fig. 6) and into 3rd-order sequence cycles (Figs. 6, 8, 9, 10, 11). Part of the subdivision and datation of this succession has been established by correlation with the adjacent Triassic Germany Basin (Fig. 7) which lithostratigraphical subdivision is well established and dated where possible.

Sequence boundaries were identified on well logs using gamma-ray, sonic and resistivity traces as the main correlation tools. The SP, density and neutron traces were also integrated where available. Facies analysis was derived from core data and from previous outcrop analysis. Some confidential seismic data has also been integrated into the interpretation procedure. The geometry of each systems tract has been described in relation to the contributing factors of sediment supply and basin physiography. Cyclicity within the sequences, up to the scale of systems tracts and parasequences, is well developed. This is...
correlatable throughout the basin and is frequently identifiable through lateral facies changes.

**CYCLE ORDER AND DEVELOPMENT**

**Triassic Major Transgressive/Regressive Cycle**

(Scythian-Late Triassic)

In the Paris Basin, most of the Triassic succession has been interpreted as one 1st-order transgressive/regressive cycle. The lower 1st-order sequence boundary is defined as occurring at the top of the Conglomérat Principal (conglomeratic sandstones), within the upper part of the Buntsandstein Formation. This boundary is a widespread level of paleosols known as the Zone Limite Violette, overlaid by Lower Triassic Buntsandstein strata. Within this hiatus, lies the equivalent of the Hardegsen or the H-Diskordanz unconformity (Trusheim, 1961) in the North Sea (Ziegler, 1981), the Dutch and Germany Basins (Durand, 1978). Outcrop data in northeast France indicate rafts of the eroded Zone Limite Violette in a conglomeratic deposit, which previously was thought to be the equivalent of the Conglomérat Principal. However, the evidence now suggests that this conglomerate and not the Conglomérat Principal is the time-equivalent of the coarse fluvial succession overlying the Hardegsen unconformity in Germany (M. Durand, pers. comm. 1994).

Below the hiatus, the Buntsandstein form massive accumulations of braided fluvial sandstones and conglomerates. They are in connection with the Germany Basin and have a limited extension in the northeastern part of the Paris Basin. Above the hiatus, the first main transgression occurs westward into the Paris Basin (see l’Huitre and Heitz le Hutier wells, Fig. 3). This marks the beginning of the re-organization of the depocenters.

The peak transgression is defined in Ladinian Calcaires à Ceratites (Upper Muschelkalk). This is a widespread marine carbonate deposit, occurring between the continental Buntsandstein Formations and the regressive Keuper Formations. The succeeding regressive phase continues until late Norian times (top Chaunoy Formation), which is a widespread regional unconformity, with erosion being more pronounced on the basin margins. This 1st-order cycle represents gradual encroachment of sediment onto the craton, due to the first phases of the progressive disintegration and subsequent drowning of the supercontinent Pangea (Fig. 6).

**2nd-Order Cycle Development**

These cycles are caused by changes in the rate of tectonic subsidence and result in major changes in the shoreline position. 2nd-order cycles are defined on the basis of their stratal geometries and physical relationships, using objective stratal geometry and facies criteria, independent of the frequency of occurrence, nature of the cycle boundaries and depositional processes. These cycles reflect the stratigraphic signature of the basin, including long term changes in the sedimentary supply, expressed by the presence of aggrading and/or backstepping sequences and long term changes in the tectonic subsidence (Jacquin et al., this volume).

Two 2nd-order cycles have been defined within the Triassic 1st-order cycle. The oldest cycle, termed the Scythian-Ladinian cycle (T1/R1), comprises Buntsandstein and Muschelkalk groups (Fig. 6). It is present over the eastern part of the Paris Basin, but thins significantly in the Marne region, after which it is not present. The basal unconformity (B5) is the equivalent of the Hardegsen unconformity (Trusheim, 1961). The transgressive phase comprises mainly backstepping sequences, made up of the fluvial and littoral marine sandstones of the upper Buntsandstein and lower Muschelkalk. This is followed by the middle and upper Muschelkalk, where the peak transgression is reached in the Calcaires à Ceratites. This also corresponds to the peak transgression of the 1st-order cycle. The succeeding regressive phase comprises forestepping sequences of the Calcaires à Terebratules and the maximum regression occurs in the Ladinian underlying the Lettenkohle Formation (3rd-order sequence K1).

The younger cycle, termed the Ladinian-Norian cycle (T2/R2), comprises the Keuper Formations. This cycle is present throughout the Paris Basin. The lowermost unconformity corresponds to an erosional surface (K1) that erodes the fluvial sands of the upper Muschelkalk and this erosion is prominent in Lorraine and Alsace (Eastern Paris Basin; Fig. 12). In the western part of the basin, the succession comprises fluvial sands, and the K1 erosional surface by is not discernible from the wireline logs.

The transgressive phase mainly comprises a thick aggrading sequence, with little backstepping. The lower part comprises the shallow-marine dolomites of the Lettenkohle, which are uniform in thickness throughout the basin. Overlying this, a thick Carnian aggrading halite package is developed in the most subsident part of the basin located just north of the Bray fault.
The peak transgression was reached in late Carnian time and occurs within the Dolomie de Beaumont, a widespread carbonate deposit and the Argiles à Anhydrite in the west. The overlying regressive phase comprises several forestopping 3rd-order sequences of the Chaunoy alluvial sands. These forestopping sequences illustrate a reduction in accommodation space during the second order regressive phase. They are observed to downlap onto the 2nd-order peak transgression, the Argiles à Anhydrite. This geometry is clearly identified on seismic data, but not on well correlations. Maximum regression occurs near the top of the Norian in the form of a major regional unconformity (K6) that truncates the forestopping sequences on seismic data. The Germany equivalent of the Dolomie de Beaumont is the Hauptsteinmergel (Ziegler et al., 1981). During periods of tectonic movement such as in the eastern Tethys and the North Sea, and a tectonic subdivision is based on analysis of available accommodation space and the vertical and lateral control on facies distribution. Triassic strata were deposited during a synrift extensional period in the eastern Tethys and the North Sea, and a tectonic component was present in the 3rd-order cycle development (Ziegler, 1981). During periods of tectonic movement such as block uplift, thick bodies of coarse clastics spill into the basin, thinning distally. During more quiescent periods, uniform widespread marine carbonates developed across the basin.

3rd-Order Depositional Sequences

Within this 2nd-order framework, the succession has been subdivided into 11 3rd-order sequence cycles. These are the building blocks of the 2nd-order cycles (Fig. 6). Our sequence subdivision is based on analysis of available accommodation space and the vertical and lateral control on facies distribution. Triassic strata were deposited during a synrift extensional period in the eastern Tethys and the North Sea, and a tectonic component was present in the 3rd-order cycle development (Ziegler, 1981). During periods of tectonic movement such as block uplift, thick bodies of coarse clastics spill into the basin, thinning distally. During more quiescent periods, uniform widespread marine carbonates developed across the basin.

CORRELATION WITH GERMANY

The Germany Triassic series was deposited in a cratonic basin with a depocenter in northern Germany. Towards the southwest it was connected to the Paris Basin (Fig. 2). During the Triassic period, Tethyan seas advanced westward due to enhanced sea-floor spreading. In the North Atlantic realm, sea-floor spreading advanced southwards (Ziegler et al., 1981).

The Germany Triassic series is classically subdivided into three parts based on lithology: (1) continental Buntsandstein, (2) marine Muschelkalk and (3) continental Keuper. This tripartite subdivision of the Germany Triassic (Alberti, 1834), has been described in a sequence stratigraphic framework by Aigner and Bachmann (1992) and also interpreted in a chronostratigraphic context. It forms a useful connecting link between the marine succession in the Triassic Dolomites in Italy (De Zanche et al., this volume), the Southern Alps (Gaetani et al., this volume), the Barents Sea (Skjold et al., this volume) and the more continental series in northwest Europe.

Thus, the 3rd-order sequence stratigraphic framework established for the Paris Basin can be correlated to the Germany Triassic series with considerable success. In Germany, unlike the Paris Basin, the conditions for studying the Triassic are excellent, as the strata are well exposed in quarries and have also been penetrated by numerous wells. In the Paris Basin, Triassic outcrops are very limited, discontinuous and are mainly confined to eastern France in Lorraine and Luxembourg. For this reason, this study was carried out mainly with wireline log data.

The sequence stratigraphic nomenclature for the Paris Basin has been adopted from the Germany Triassic nomenclature (Aigner and Bachmann, 1992). Several sequences in the Germany domain have not been defined in the Paris Basin especially where the lower Buntsandstein succession remains undivided. As tectonic subsidence in the Paris Basin was less than in the Germany Basin, these sequences were probably not developed or are very subtle. Other sequences have been defined in the Paris Basin that are not defined in the Germany Basin (Fig. 7).

SEQUENCE DEVELOPMENT

Buntsandstein Sequences

The Buntsandstein sequences represents the first stage in the formation of the Paris Basin. The basin’s formation is first recorded in Scythian strata and continues up to the lower Muschelkalk Formations in the Anisian strata. The Buntsandstein deposition represents the first transgression westward and its maximum flooding is recorded in Anisian strata overlying the first marine deposits known as the Grès Coquiller and lateral equivalents, the deltaic deposits of the Grès à Voltzia.

Sequence B(1–4)?: Grès Vosgien, Conglomérat Principal.—

The first Buntsandstein sequence rests on preserved Permian, Carboniferous and/or Devonian strata. This boundary is poorly dated as Scythian in age. Due to inadequate well data, the nature and extent of this sequence boundary is unclear, but in wells to the east such as Meligny (Fig. 3) and Francheville, a thick conglomerate known as the Conglomérat de base du Trias (Lapparent, 1883) overlies the boundary. However, this deposit is not regionally widespread. Elsewhere, this boundary is overlain by the thick fluvial deposits of the Grès Vosgien, which chronostratigraphically remain undivided and are widespread over the eastern part of the basin. These are medium-to-coarse-grained sandstones with conglomeratic intervals, deposited in a braided fluvial regime (Fig. 8). These are overlain by the Conglomérat Principal, a coarse-grained fluvial conglomerate that can reach a thickness of up to 20–30 m. It thins towards the northeast on Pont à Mousson antiform in Moselle. Outcrop sections of the Grès Vosgien and the Conglomérat Principal provide evidence that the entire succession is a conformable, shallowing up megasequence that throughout Scythian time prograded northeastwards. The upper contact of the Grès Vosgien with the overlying Conglomérat Principal is progressive and not erosive. The conglomerate is diachronous; it is thought to be Scythian-Anisian in age (Durand, 1978).

The first sequence in the Paris Basin that can be confidently correlated with the Germany Basin lies above the Conglomérat Principal at the top of the Zone Limite Violette, a group of
palaeosols. This sequence boundary is the equivalent of the Hardegsen unconformity (sequence B5).

Due to a thicker and more developed Buntsandstein succession in the Germany Basin, three sequences are defined beneath the Hardegsen unconformity, but they are not defined in the Paris Basin. It is possible that these three sequences are represented in the Paris Basin by a series of hiatuses which form the palaeosols of the Zone Limite Violette.

**Muschelkalk Sequences**

**Sequence B5, Anisian pars.—**

Sequence B5 comprises the lithostratigraphic units traditionally named as Zone limite Violette, Couches Gréseuses Intermédiaires, Grès à Voltzia, Grès Coquiller, Dolomie à Myophoria Orbicularis and Couches Rouges, pars (Courel, 1980).

The B5 sequence boundary has been identified by a level of palaeosols known as the Zone limite violette (Perriaux, 1961). This corresponds to varicolored silts with specific alteration features present (dolocretes and cornallite) formed in a semi-arid hot climate (Durand, 1972). The well logs signature is a high, sharp gamma-ray peak.

The B5 transgressive systems tract comprises the fluviatile Couches Gréseuses Intermédiaires, the fluviatile sands of the Grès à Voltzia and the marine sands of the Grès Coquiller. The Couches Gréseuses Intermédiaires were again deposited in a semi-arid, hot climate. They contain rafts reworked from the underlying Zone Limite Violette. The sands have been derived from a different source and they are mineralogically immature compared to the underlying Grès Vosgiens. This indicates that a new depositional regime began with B5 sequence (Durand, 1978). The Grès à Voltzia is essentially a transitional deposit between the fluviatile and marine depositional regimes. It is divided into the underlying Grès à Meules at the base and the Grès à Argiles at the top. The Grès à Meules are homogenous sandstones deposited within a fluviatile system with distributary littoral mouth bars. The Grès à Argiles is a fluvial succession, deposited near marine influence (Durand, 1978). The distribution of lignite and vegetation debris in the sand and shale deposits suggests that the distributary channels flowed eastwards and the overlying littoral marine deposits of the Grès Coquiller advanced westward.

The Grès Coquiller (Lapparent, 1883) has been described as a littoral marine sandstone, feldspathic in nature, with a dolomitic cement. It is the first marine formation at the base of the Muschelkalk. The Buntsandstein-Muschelkalk limit is a facies limit and not a time limit (Durand and Jurain, 1969). It is separated from the underlying Grès à Voltzia by a transgressive ravinement surface (Fig. 9). The B5 Maximum flooding shows at the top of the Grès Coquillers in the northeast of the basin and at the top of the Grès à Voltzia in the southwest. Westward, the Grès à Voltzia extends to the area around St. Just Sauvage (Fig. 3) to Flacy (Fig. 5), overlying the Conglomé rate principal, and illustrating the retrograding facies of this first important transgression into the Paris Basin.

The Grès Coquiller has been dated by fauna identified with the biostratigraphic timescale of Kozur (1974). It is diachronous and ranges from early to late Anisian in age. The formation contains abundant marine pelecypods (Myophoria vulgaris, Lyriomyophoria elegans, Neoschizodus cardissoides), brachiopods (Coenothyris vulgaris) and crinoids (Encrius liliiformis).

The B5 highstand systems tract is represented by the Dolomie à Myophoria orbicularis (Guillaume, 1959). This is a massive, sandy dolomite or a dolomitic sand, depending on the places. It is rich in oolites, stromatolitic crusts and bone debris. It is a shallow-marine deposit of varying salinity. This dolomite is only present in the eastern part of the basin (Durand and Jurain, 1969). Moving westward, the proximal facies equivalent is found to be a restricted lagoonal shale, often anhydritic in nature and thought to be part of the Couches Rouges Formation.

**Sequence M2.1: Upper Anisian.—**

Sequence M2.1 comprises the lithostratigraphic units traditionally named as Couches Rouges, Couches Grises, Couches
**Blanches.** These are developed in the Sarreguemines syncline in the east of the basin. This represents the first stage in the development of an independent depocenter westward.

The M2.1 sequence boundary separates the lower and upper Muschelkalk. The boundary is not a sharp discontinuity, but is marks a period of renewed tectonic activity in the area. The M2.1 transgressive systems tract comprises the Couches Grises (Guillaume, 1959). This is manifested by the appearance of several thick anhydrite-halite beds. The M2.1 maximum flooding surface is interpreted as occurring at the top of the thick evaporitic beds. Further north along the basin margin, local tectonic activity produced a thick conglomeratic interval, exemplified in the Varennes well (Fig. 4).

The M2.1 highstand systems tract is composed of the Couches Blanches (Guillaume, 1959) and is a shaly formation, anhydritic in places and becoming dolomitic and often oolitic towards the top. Moving eastwards, the sequence thins dramatically as it leaves the Sarreguemines synclinal region. The sequence thins to about 20 m and is represented by a shaly deposit, that is sandy in areas. These sands are attributed to the Couches Rouges (Ricour, 1953), which thickens here at the expense of the Couches Grises. Conversely, in the Sarreguemines synclinal region, the Couches Rouges Formation does not appear to be well developed. Westward, after the Trois-Fontaines well in the Marne region (Fig. 3), the sequence is not easily defined and is represented by a condensed interval on the synclinal margins.

In Germany, the succession is represented by a thick halite deposit because subsidence was more pronounced in this area. These halite deposits are attributed to lowstand deposits (Aigner and Bachmann, 1992).

**Sequence M2.2: Anisian Pars, Ladinian Pars.—**

Sequence M2.2 comprises the lithostratigraphic units named as Calcaires à Entroques, Calcaires à Ceratites, Calcaires à Terebratules, Lettenkohle Formation (pars). In the east of the Paris Basin, the M2.2 sequence forms an open-marine carbonate ramp succession, where the facies evolves into marginal marine siliciclastics in the basin center and deltaic alluvial plain deposits in the west of the basin. This sequence is not supported by data in the Germany Basin, but interpreted to be a continuation of the previous sequence (Fig. 7 and Aigner and Bachmann, 1992).

The M2.2 sequence boundary lies at the base of the marine carbonate succession of the Upper Muschelkalk. This boundary is defined as marking the beginning of a new period of uniform subsidence over the basin, with open-marine conditions deposits.

From Lorraine to Alsace, the M2.2 transgressive systems tract comprises a retrograding parasequence set of seven parasequences, indicating a fining up trend of the limestone formations known as the Calcaires à Entroques and the Calcaires à Ceratites (Fig. 10; Dubois and Firton, 1937). The Calcaires à Entroques is described as wackestone interval, passing into crinoidal grainstones which is oolitic by places. Silty intervals occur frequently. The Calcaires à Ceratites Formation is described as a marine carbonate deposit having upward decreasing in energy levels (Dubois and Firton, 1937). Alternating limestone beds and marly interbeds are equally as thick. This formation is well dated with cephalopods (Ceratites compressus, C. evolutus, C. spinosus; Corroy and Linkoff, 1928) and towards the summit C. nodosus. Ceratites and conodonts (Biozone 2–5; Kozur, 1974) are early Fassanian/Longobardian in age.

The limit between the Calcaires à Entroques and the overlying Calcaires à Ceratites is diachronous, as shown by the ceratite fauna (Düringer and Hagedorn, 1987). It corresponds to the Ladinian-Anisian boundary for most Lorraine, but the boundary probably moves up where the Formation is abnormally thick (Schneider, 1957). West of Chevraumont (Fig. 3), both the Calcaires à Entroques and Calcaires à Ceratites becomes sandy and dolomitic. Lateral outcrop equivalents are the...
Fig. 4.—North-south transect across the eastern Paris Basin to the Bresse-Jura graben. Datum: Maximum flooding surface of the Carnian (Dolomie de Beaumont).
A SEQUENCE STRATIGRAPHIC FRAMEWORK OF THE MARINE AND CONTINENTAL TRIASSIC SERIES

Fig. 5. — Northwest-southeast transect across the western part of the Paris Basin. Datum: Final maximum flooding surface in the late Triassic (Rhaetic).
**Valerie Gogglin and Thierry Jacquin**

**Fig. 6.**—Chronostratigraphic diagram for the Triassic of the Paris Basin illustrating the main formations and the sequence stratigraphic cycle hierarchy up to 3rd-order frequency.

**Dolomie du Strombeg** and the **Dolomie de Voisey** (Durand and Jurain, 1969). The M2.2 maximum flooding surface corresponds to the 1st-order peak transgression, as previously described.

The M2.2 highstand systems tract in the east is attributed to the **Calcaires à Terebratules** (Dubois and Firton, 1937). It forms a prograding parasequence set (Vail et al., 1977), comprising five seaward stepping parasequences (Fig. 10). This formation is a limestone deposit comprising three essential elements: (a) an upper Terebratula bank (0.3–0.4 m in thickness) that is a shelly wackestone and sometimes dolomitic in nature, (b) a 4 to 5 m interval of marls with shaly interbeds and (c) a principal Terebratula bank with thin shaly intervals. The **Calcaires à Terebratules** has been dated by ceratite (**Discoceratites semipartitus**, **D. dorsoplanus**, **D. levalloisi** and **D. intermedius**; Maubeuge, 1947) of the Discoceratites Zone (mid-Langobardian age; Kozur, 1974). Other fauna include brachiopods (**Coenothyris vulgaris**) and pelecypods (**Costatoria goldfussi**; Corroy and Linkoff, 1928).

The M2.2 highstand systems tract lateral equivalent westward are the **Dolomie de Vittel** and the **Calcaires chamois**. Further in the west, these grade to fluvial deposits. In Alsace, a further sequence may be developed in the **Calcaires à Terebratules**, as observed in wells Schweighouse and Oberrodern, but this is not developed westward in the Paris Basin to the same extent and is difficult to distinguish.

A deltaic succession with thick marginal marine sands at the base is present in the center of the basin, near Lorraine. These sands are potential hydrocarbon reservoirs. They were deposited in a shallow marine setting as upper shoreface sand bodies, barrier bars and deltaic sands. The maximum flooding surface occurs in a lower shoreface bioturbated interval, as observed in the cored sections (Fig. 11). The lower part retrogrades westward during the transgressive phase and progrades eastwards during the highstand phase. However, due to lack of biostratigraphical data, it was difficult to correlate the carbonate ramp sequence into this proximal facies. This difficulty is enhanced by the occurrence of thick dolomite beds underlying the so-
east of the sands is difficult to justify. It may, however, represent progradation of fluviodeltaic sands and shales during the highstand systems tract as far west as Vulaines near the Marne valley (Fig. 3, 5), but in this area it is difficult to discern as the succession comprises fluvial sands.

Sequence M2.3: Upper Ladinian, Pars.—

Sequence M2.2 comprises the lithostratigraphic units named as Calcaires à Terebratules. This sequence has been identified in the easternmost Paris Basin and in Alsace (Fig. 3). It is developed within the Calcaires à Terebratules and is only about 20-m-thick. The sequence corresponds to mid-lower carbonate ramp, in a wackestone-grainstone depositional environment. It
Keuper Sequences

Sequence K1: Uppermost Ladinian—Lower Carnian.

Sequence K1 comprises the lithostratigraphic units named as Lettenkohle Formation, Donnemarie Formation (pars) and Couches à Pseudomorphoses (pars). The K1 sequence boundary has been identified as an erosional unconformity illustrating erosional truncation of the Lettenkohle Sands. Regional tilting occurred westward in the basin at sequence boundary K1 and the underlying Lettenkohle sands were truncated as the transgression advanced (Fig. 12). This occurs from Haute Marne to Alsace.

In the east and central part of the basin, the K1 transgressive systems tract comprises a thin transgressive dolomitic interval at the base and a shaly interval followed by three dolomitic parasequences. In Lorraine, the thin basal dolomitic interval contains teeth and bone debris (bone beds) that are thought to be transgressive lag deposits. The dolomite is a porous, friable, oolitic arenite with lumachelic intervals often impregnated with gypsum and anhydrite. Microconglomerates are sometimes present at the base (Fig. 13). This part was deposited in a marginal marine environment (Durand, 1980). The overlying shale interval appears to be restricted in nature, green-black in color and is characterized by a distinct, strong gamma-ray peak. In north Lorraine, the shale is red-green but this color disappears southwards. Sometimes the shale is rich in vegetation debris and lignite. Sandy intervals are sometimes present. Deltaic characteristics are prominent in this formation (Guillaume, 1938).

The K1 transgressive systems tract ends with a series of three dolomitic parasequences, with the maximum flooding surface being defined at the third parasequence. The three upper dol-
mitic parasequences forming the upper part of the transgressive systems tract also rarely outcrop, but have been described as a dolomitic limestone with several shaly intervals. Certain intervals are lumachelic. The depositional environment is interpreted to be marginal marine with varying salinity (Fig. 13).

The K1 transgressive systems tract essentially comprises the Lettenkohle Formation (Figs. 3, 4, 5; Guillaume, 1938). It is uniform in thickness and distribution across the basin. In the western part of the basin towards the Seine valley, there appears to be a corresponding flooding within the continental fluvial deposits of the Donnemarie Formation, tentatively identified by a shaly interval by using precise three-dimensional correlations over the area (Fig. 3, 5). This event is confirmed by Bourquin and Guillocheau (1993), who have identified a flooding period within these sandstones that is marked by an interval of bioturbated clays with occasional dolomitic and anhydritic intervals. It has been interpreted as a coastal plain environment within the fluvial succession. The Donnemarie Formation is interpreted to be a braided fluvial system. The deposits are coarse grained, conglomeratic, displaying very high gamma-ray peaks (Bourquin and Guillocheau, 1993).

The K1 highstand systems tract coincides with the Couches à Pseudomorphoses in the eastern part of the basin. This formation is predominantly shaly with anhydritic intervals, deposited in a continental setting, with restricted marine connections. Near the Seine Valley, the highstand systems tract equivalent most probably lies in the coarse fluvial deposits of the Donnemarie Formation. This formation is very thick, reaching up to 150 m in some areas (e.g. Hermes—1 well in the Marne region; Fig. 5).

**Sequence K2.2: Carnian, Pars.**

Sequence K2.2 comprises the lithostratigraphic units named as Formation salifère, Donnemarie Formation (pars), Couches à Esthères, Couches à Pseudomorphoses, Marnes à Anhydrite. The K2.2 sequence boundary lies at the base of the Formation Salifère, which comprises a thick aggrading sequence of halite (Fig. 14), interbedded with thin shale beds, and can reach up to 75 m in thickness (e.g. Meligny well; Fig. 3). In the basin, the K2.2 Sh is not erosive but the manifestation of such a halite thickness indicates a new distinct period of syntectonic deposition due to fault movement along the Pays de Bray fault and the Metz Fault in Lorraine (Figs. 3, 10, 14). On the western margin, where the succession is predominantly sandy, conglomeratic intervals have been identified at the base of the sequence, indicating that local erosion occurred.

The depocenter formed in the region of Meuse, the Marne Valley area and Lorraine accumulating a thick halite succession.
The extent of the halite is controlled by major basin faults. A second localized tectonic subsidence event provided another opportunity for the Paris Basin to attain its independence from the Germany domain. Similar subsidence occurred in north Germany at this time (Beutler, 1993), but sequence K2.2 is not supported by data in Germany in the equivalent Gipskeuper Formation (Bourquin and Guillocheau, 1993). However, proposed explanation is that this prominent dolomitic horizon is a lithological marker with no chronostratigraphic significance. It represents the carbonate ramp development nearshore and is overlain by prograding fluvial clastics during the highstand systems tract. With respect to basin development, during the Ladinian severe truncation of the Lettenkohle fluvial sands towards the east was geometrically very unlikely.

The K2.2 hightstand systems tract comprises anhydritic shales throughout the western part of the basin (Fig. 16). In some wells, such as Janvry and La Folie de Paris, just east of the Seine valley, the anhydrite development increases towards the overlying sequence boundary K3.

The halite sequence has been subdivided into a thick, aggrading transgressive systems tract. The occurrence of thick restricted deposits represents the creation of large amounts of accommodation space, typical of 2nd-order aggrading transgressive phases. To illustrate this point, the sequence has been subdivided into four halite depositional packages that retrograde into the basin from the east. Each retrograding depositional package becomes more saliferous as it advances westward (Fig 3, 16). The maximum flooding surface, which occurs after the most subsident period in the basin may be defined as lying at the top of the third parasequence because it is the most extensive across the basin. The overlying fourth parasequence has been tentatively attributed to the K2.2 hightstand systems tract because it is only developed in certain areas, possibly indicating the end of uniform subsidence in the basin and the beginning of a more regressive period, with halite “filling” the remaining subsident zones. The top of the Formation Salifère also becomes more anhydritic in places, marking the end of this subsident phase. This is known as the Marnes à Anhydrite Formation (Marchal, 1985). However, three phases of halite dissolution have been identified by Marchal (1985) and may also have caused the top of some successions to become anhydritic. Thus, the retrograding halite geometry may be exaggerated, especially in Lorraine.

**Sequence K3: Upper Carnian.**—

Sequence K3 comprises the lithostratigraphic units named as *Dolomie de Beaumont* (shallow marine carbonate), *Grès à Roseaux* (extensive fluvial sandstone; Levallois, 1867); *Argiles In-
The K3 sequence boundary is a major regional unconformity, associated with the appearance of a vast deltaic system over the whole of northwest Europe, advancing towards the southwest. This marks a major downward shift of coastal onlap. The sands are interpreted to be mainly distributary channel fills, preserved within the deltaic system during the transgressive systems tract. The sands do not appear very erosive from well log correlations (Fig. 16). However, outcrop studies in the Germany Basin indicate that they eroded up to several tens of meters of the underlying lacustrine muds and thin dolomites of the Couches à Esthérées equivalent that belong to sequence K2 (Bachmann and Wild, 1976). These sands can be very thick locally in the northwest of the Paris Basin in the Seine valley area, due to fault activity along the Bray fault. The sands are developed in a shoestring manner throughout the basin and northwest Europe (Wurster, 1963).

The lateral equivalent is a shaly, lacustrine deposit with little or no sands. Wurster (1963) has interpreted the depositional environment as a type of birdsfoot delta that developed over the lacustrine shales in the region, similar to the present day Mississippi River deltaic deposits. However, it has more recently been interpreted to be of predominantly fluvial origin by Dittrich (1989 a,b).

The transgressive systems tract continues into the widespread 2nd-order peak transgression marine deposit, the Dolomie de Beaumont. This dolomite is a dolomicrite, often shaly, hard, with a splintery fracture. Internal breccias are sometimes pres-
At this stage, relative sea level was uniform and widespread across the basin, hence the development of this shallow marine carbonate.

In the west, the transgressive systems tract is represented by a shaly anhydritic Formation called the Argiles Intermédiaires (Levallois, 1851). The base of the formation comprises three shale-anhydrite parasequences, each becoming progressively more anhydritic until the maximum flooding surface is reached. The K3 MFS lies within the Argiles à Anhydrite Formation. The anhydrite beds in each parasequence represent the greatest marine influence in this retrograding parasequence set, which is in contrast to the shale anhydrite cycles developed in the Norian sequences.

Further west, where the succession is completely fluviatile and the Grès de Chaunoy rests directly on the Donnemarie Formation, the K3 maximum flooding surface is more difficult to define. Previous studies by Bourquin et al., (1993) show evidence for a sharp gamma-ray peak between these two formations. It has been identified as an uranium peak and acts as a valuable isochronous marker in the western part of the basin. It is also present overlying the Argiles Intermédiaires. We interpret this marker to represent the sequence boundary at the base of the Grès de Chaunoy. It may have an equivalent overlying the Dolomie de Beaumont in Lorraine and Alsace, but our correlations indicate that the sequence boundary lies above this gamma-ray peak, under a thick anhydrite succession, in the Marnes Irisées Supérieures (Figs. 3, 4, 5).

To summarize, the uranium marker is interpreted to represent a surface of non deposition and sediment starvation in the west of the basin. It represents the 3rd-order sequence boundaries (K4-K(n)), underlying the Grès de Chaunoy. Therefore, the maximum flooding equivalent of the Dolomie de Beaumont and Argiles à Anhydrite in this fluviatile succession underlies this...
marker and lies near the top of the Donnemarie Formation (Fig. 3).

The K3 highstand systems tract in the east is defined by a more regressive playa type facies in the widespread formation of the Argiles Bariolées Dolomitiques and Argiles de Chanville, collectively known as the Marnes Irisées Supérieures (Mau-beuge, 1949). These formations are essentially red in color, with banks of dolomite and anhydrite. The formation thins towards the east probably due to a reduction in sediment supply and thickens and becomes sandy north of the Seine valley, as observed in the well Montreuil aux Lions.

This represents the final marine flooding into the basin within the Triassic 1st-order cycle, and it encroaches landwards as far as Villoisoin in the Seine Valley (Fig. 3). This flooding accompanies a major climatic change from an arid to a more humid phase according to Simms and Ruffel (1990). They have dated this episode as being late Julian in age based on the huge end of Carnian biotic turnover. Therefore, it is possible that this climatic change throughout northwest Europe was caused by a eustatic sea level rise that may have been due to a change in the intra-plate stress field in Ladinian or Carnian times.

**Sequence K4-K(n): Lower Norian.**

Sequence K4-K(n) comprises the lithostratigraphic units named as Grès de Chaunoy and Marnes Irisées Supérieures (pars). A return to more arid conditions occurred during Norian time, with the development of thick alluvial prisms of coarse clastics in the west of the basin, sourced by the Amorican massif.

The K4-K(n) sequence boundary is defined as a regional unconformity that occurs at the base of the Grès de Chaunoy. This unconformity is recognized to represent the base of several 3rd-order sequences, which forestep northeastwards into the basin, explaining the nomenclature used above, K4-K(n). These forestepping sequences have only been observed on seismic data, thus it is not yet possible to accurately define all 3rd-order sequences developed over the region. The K4-K(n) basal unconformity is recognized in the German Basin (Beutler, 1978, 1980). This second-order regressive phase marks the final widespread shift in the basin depocenter westward before the Liassic transgression. It is also the final phase of localized tectonic subsidence in the development of the Paris Basin as an individual entity. The alluvial episode was induced by block uplift, due to activation of the Sennely Fault along the western margins, producing a succession of forestepping 3rd-order sequences into the basin from the Amorican Massif. These alluvial deposits form the oil reservoir for the Chaunoy discovery.

The base of the K4-K(n) transgressive systems tract is characterized by thick, stacked aggradational fluvial sands (Fig. 17). The preservation of these sands indicate high accommodation rates at the beginning of the TST. The K4-K(n) maximum flooding surface, or to be more precise, the maximum accommodation surface is difficult to define in the west, but it is interpreted to occur in a shaly interval just above the thick fluvial channel sands. In core data, this interval corresponds to a bioturbated shale and silt interbedded succession that was deposited in a lacustrine environment. The K4-K(n) highstand systems tract becomes more dolomitic upward, towards sequence K6. It can be subdivided into several fining-up fluvial cycles interbedded with siltstone redbeds (Fig. 18). Two types of dolomite are found within the alluvial sandstone sequence: (a) nodular dolomitic beds, which form in mottled, overbank silts and muds and (b) a thick massive dolomite, which occurs in coarse grained channel sands and conglomerates. These dolomisation is interpreted to be due to enriched groundwater percolating from the ancient massifs. Strontium and carbon isotope data of early diagenetic dolomite cements and oxygen isotope data for diagenetic silica indicate an entirely non-marine origin for the groundwater (Spotl and Wright, 1992).

Correlating eastwards into the basin, the alluvial prism that corresponds to the Grès de Chaunoy interfingers with more playa-type, red-green shales, known as the Marnes Irisées Supérieures. A distinct cyclicity is observed within these Marnes Irisées that is manifested in an “anhydrite-shale-anhydrite” deposition and comprises a typical 3rd-order depositional sequence (Fig. 16). The sequence boundary associated with the lower anhydrite interval is erosive in nature in Lorraine and on the Burgundy High. This hiatus, however, needs further stratigraphic evaluation. The maximum flooding surface corresponds to the most argillaceous interval identified on wireline log data; shoaling up occurred progressively during the highstand systems tract to the next sequence boundary, which is marked by a very anhydritic interval. The anhydrite beds in this instance correspond to the most restricted interval in time, indicating regional emergence and evaporation in a playa-lake-type situation.

**Sequence K5: Upper Norian, Pars.**

Sequence K5 comprises part of the lithostratigraphic units named as Marnes Irisées Supérieures. Sequence K5 is preserved in the basin depocenter and southwest towards Orléans, where subsidence was most pronounced at the time. Towards the basin margins, it has been eroded due to tectonic uplift in late Norian times, although it is locally preserved as incised valley fills. Truncation is apparent in the southeast, east and north but is not as accentuated as in the west of the basin. This erosion in the west plays an important role in the development of the Chaunoy oil field reservoir. The unconformity is thought to act as an important seal for the underlying reservoir interval in addition to the overlying Sevatan shales at the base of the Rhaetic.

Sequence K5 is preserved in the north of the basin, due to enhanced subsidence by local movement along the Pays de Bray fault. In the basin center, it is again manifested by an “anhydrite-shale-anhydrite” cycle, deposited in the same playalake depositional environment as the previous sequence and forms part of the Marnes Irisées Supérieures. Eastward, in the basin where the sequence is uneroded, the deposits are also very shaly. K5 also thins noticeably perhaps due to a reduction in sediment supply in this part of the basin. The K5 maximum flooding surface is again defined as the most argillaceous interval seen on wireline log data. The K5 highstand systems tract is shaly, becoming more dolomitic in nature towards the overlying sequence boundary. Figure 17 illustrates the 3rd-order sequence developed in the alluvial clastics and the truncation of sequence K5 due to fault-block uplift and 1st-order maximum regression surface (K6Sb).

**Rhaetic Sequences**

The Rhaetic succession in France, although considered by many geologists to be part of the Liassic Series (Ricour, 1960),
succession, as observed in the Germany Basin. Tigraphically, it still has important characteristics of the Triassic regressive/regressive cycle, but lithostratigraphically and biostratigraphically, it belongs to the succeeding Liassic 1st-order transgression. Sequence K6 is included in this study. In a sequence stratigraphic cycle hierarchy, it belongs to the succeeding Liassic 1st-order transgressive/regressive cycle, but lithostratigraphically and biostratigraphically, it still has important characteristics of the Triassic succession, as observed in the Germany Basin.

The Rhaetian Stage acts as a transition between the continental Triassic deposits and the open marine deposits of the Liassic. During Rhaetian time, a different sedimentological setting was present. The main contributory factors were: (1) a reduction in the extent of mesosaline settings and a replacement by normal marine conditions, (2) a change to a more humid climate with a change in the evaporation/precipitation ratio, and (3) a decrease in evaporite deposition in the marginal marine settings in the western Tethys (Iannace and Frisia, 1994).

In this study, the Rhaetian Stage has been subdivided into three 3rd-order sequences, numbered as K6, Rh1 and Rh2. The succession comprises a shallow-marine, tidally-influenced sand and shale deltaic system, present over most of the basin. In the west, a coarse fluvial sand succession is still prominent, the remnants of the Chaunoy alluvial system. A transitional lagoonal area is present between the two areas, composed of shaly lagoonal deposits and marginal marine sandstones (intertidal), near the top of the succession.

Previous studies by Beutler (1993) indicate the presence of three tectonic events during the Rhaetic deposition of the Germany Basin, which may correspond to the three sequences defined in the Paris Basin.

Sequence K6: Uppermost Norian-Lower Rhaetian.—Sequence K6 comprises the lithostratigraphic units named as Argiles de Chalain, Grès de Chalain, Grès de Boissy, Argiles bariolées dolomitiques, Grés d’Arsy.

The K6 sequence boundary is a widespread regional unconformity, with uplift and erosion being more pronounced at the basin margins, where it appears to erode the entire sequence K5. In the west, the boundary is characterized by the development of a thick dolomite horizon (2–5 m) representing a calcite zone identified on core data and well logs. It is the result of subaerial exposure and erosion, which is thought to play an important role in the Chaunoy oil field reservoir development.

Over most of the basin, the K6 transgressive systems tract is composed of red-green shales and fluvial sandstones. The
shales are a lagoonal/lacustrine deposit, known as the Argiles de Chalain. It is a very extensive uniform deposit. It has been observed as far north as the well Pays de Bray where it directly overlies Permian strata (Fig. 5). These shales are Sevatian in age, indicating that the Liassic transgression began in late Norian times (Graciansky et al., this volume). In the west, the K6 transgressive systems tract comprises fluvial coarse grained sands composed of the lower, middle and upper Grès de Chalain. The middle Grès de Chalain is the Grès de Boissy. In the Champagne sur Marne well from this area (Fig. 5), core data indicates a conglomeratic deposit containing dolomitic rafts within the transgressive systems tract, derived from erosion of the underlying sequence. In the shallow marine basin center deposits, the K6 MFS shows just below the Rhaetic Sand and Shale Formation (Beneke, 1877) in the Marnes Bartoliées Dolomitiques.

The K6 highstand systems tract comprises the first tidal-influenced deltaic sand and shale deposits, as the marine transgression advanced. Throughout this part of the basin, the highstand systems tract is well defined and comprises a set of 3 prograding parasequence set, which are seen to be preserved in the most subsident area in the northwest. From north to south in the basin, downlap onto the maximum flooding surface is observed (Fig. 19).

Tectonic uplift in the Ardennes massif produced a supply of fine quartzose sands for the area, and a new fluvial system imposed itself on the paleogeography. A main depocenter for these sands lies in the northwest of the basin in the Marine region. This represents the initial progradation of fluvial sands into the basin and reworking by the advancing marine transgression produced tidal sand bodies within a tidal influenced deltaic system. The sediment supply produced by tectonic uplift exceeded subsidence due to the advancing transgression, thus producing progradation of the deltaic body, even though the resulting deposits are predominant-shallow marine in nature.

Sequence Rh1: Rhaetic Sands and Shales Formation.—

The Rh1 sequence boundary is a widespread unconformity produced by a pronounced base level drop. This is observed in the west of the basin by a level of paleosols overlain by channel sands. Towards the basin center, in the shallow marine deposits area, a distributary channel system was eroded during the corresponding relative sea-level lowstand. Also, fault activity became pronounced again, and thick fluvial clastics were produced. The pronounced base-level drop caused channel incision, which was subsequently infilled during the next transgression. These transgressive coarse sands and conglomerates are excellent hydrocarbon reservoirs and are developed in Les Bergers and Vért Le Grand wells (Fig. 3).

In the basin center, erosion and channel incision into the soft, lagoonal Sevatian shales are associated with the K6 Sb. The associated channel-fill is sandy and has an aggradational nature on wireline log data (see Courdemanges well; Fig. 3). As the transgression advanced towards the north of the basin, the fluvial sands (possible lowstand deposits?) were reworked. Some remnant fluvial sands were preserved at the base of the succession but are quite thin. Outside the main channel area, the K6 transgressive systems tract is composed of a retrograding parasequence set, the thickness of which is depending on the local subsidence rate. Sandbodies are predominantly tidal sand ridges. The K6 maximum flooding appears in a shelf-rich zone, often characterized by Rhaetavica contorta (Poujol, 1960).

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**Fig. 19.**—North-south correlation across the central Paris basin illustrating third-order sequence development in the shallow marine Rhaetic sands and shales. The datum is the maximum flooding surface of sequence K6. This illustrates the progradation of the deltaic sequences basinwards. Note the preservation of the lower parasequence in a subsident area. This is the only area of the basin where it has been preserved and is due to local tectonic movement of the Pays de Bray fault.
The K6 highstand systems tract is composed of a prograding parasequence set new that arrived from the north and northwest. Also, at this stage, a shift in the subsidence center is observed towards the east. The main depocenter lay between the Bray-Vittel and Metz Faults and is possibly due to renewed activation along these faults.

**Sequence Rh2: Rhaetian Sands and Shales Formation.**

The Rh2 sequence boundary is an erosional unconformity and in the western part of the basin, it is overlain by fluvial channel sands. In the east, with the transitional lagoonal facies, the boundary is clearly defined by the first appearance of marginal marine sands. In the shallow marine succession, the boundary is erosive in Alsace, eroding Rh1 and down as far as K3 south towards the Bresse region. In the west, the Seine Fault and associated faults continued to produce fluvial clastics, but fault activity was minimal and, in consequence, deposits are thin and uniform. However, thick conglomeratic sands locally overlie the Rh2 Sb (e.g., Bechevret well; Fig. 20). Towards the east, in the transitional lagoonal facies, Rh2 Sb is identified as the limit between the lagoonal succession and the first marginal marine influences of the transgressive sands, appearing at the top of the succession (e.g., the Chaunoy wells; Fig. 3). The Rh2 maximum flooding is defined as a shaly deposit and is correlatable throughout the area.

The relative sea level fall that formed the Rh2 sequence boundary may have produced a forced regressive wedge (Posamentier, 1992) in the southeast, in the region of the Forcelles well. It is represented by an isolated sediment wedge characterized by relatively thick aggradational sands (Fig. 21). Inadequate well control prevents a more detailed description of this fourth systems tract. This may be the equivalent of an inner shelf shoal or a drowned sand barrier, which often occurs due to a rapid rise in sea level and presents attractive potential reservoirs for hydrocarbons or groundwater.

As the transgression continued northwards, some of the previously deposited shallow-marine sands of the Rh1 HST were reworked and redeposited in the more subsident areas in the basin as barrier bar sands or tidal channel sands. The Rh2 maximum flooding surface is the flooding in the basin before the marine carbonates of the Hettangian. It is again characterized by *Rhaetavicula contorta* (Martin, 1983, Poujol, 1960).

The succeeding Rh2 hightstand systems tract in the basin is represented by a locally thick, sandy, prograding package. It becomes quite thick in areas (e.g., Courdemanges and Heitz le Hutier wells in Haute Marne; Fig. 3). Subsidence was relocated to the basin center once more. The sediments were deposited in a depocenter towards Lorraine in the west, with the thickest sediment deposited in an area south of Reims and north of Troyes. These thick prograding sands were deposited near the distributary channel area. Overlying these sands, a lagoonal deposit characterized by red-green shale was deposited, known as the *Argiles de Levallois* (Jacquot, 1855). The shales have been deposited in calm marine conditions, with decreasing salinity tendencies. This deposit is widespread in the eastern and central parts of the basin from Alsace to the Marne region. It is generally quite thin, up to 5 m thick.

Fauna is rare in the Rhaetic sequences, but the formation is identified as lying between the *Rhaetavicula contorta* beds and the first Hettangian faunal Zone, *Psiloceras psilonotum*. The marine transgression approached the basin in a west-northerly direction. The paleogeography of the basin changed reflecting a change in the strain direction with the onset of the Liassic cycle that is linked to the progressive breakup of Pangea.

**SUBSIDENCE ANALYSIS**

Two-dimensional subsidence analysis using the program *Basinworks* (Marco Polo Software, 1989) has been carried out using several wells situated on Figure 3.

Previous studies (Loup and Wildi, 1994) indicate that the Permo-Triassic Systems of the Paris Basin were deposited during a period of extensional fault activity in the Tethys sea, with the formation of a trough superimposed onto three fault systems derived from the Variscan structural framework and forming a triple junction (Bray-Seine, Sennelly-Loire and Vittel and Metz fault systems). Many contradictions exist between the basin-forming mechanisms (e.g., stretching factors, the present geometry and basin characteristics), but it is generally accepted that the Paris Basin originated by uniform lithospheric stretching in a transtensional stress field. Throughout the Triassic Period, this uniform stretching is seen to have been interrupted in late Ladinian and late Norian times. The subsidence patterns are observed as being discontinuous with time, with two main cycles defined. The well Meligny illustrates this cyclicity very well (Fig. 22). The two cycles correspond to the two 2nd-order transgressive/regressive facies cycles (T1/R1 and T2/R2) within the Triassic 1st-order transgressive/regressive cycle. Thus, there exists a close relationship between the tectonic evolution of the basin and the 2nd-order transgressive/regressive cycle pattern.

The transgressive phase corresponds to a progressive increase in the subsidence rate, and the regressive phase corresponds to a decrease in the subsidence rate. The transgressive

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**Fig. 20.—**Cored section through the Rhaetian fluvial sands and sequence boundary Rh2 in the western of the basin. The sequence was formed by a base level drop, and during the transgressive systems tract, it was subsequently infilled by conglomeratic sandstones.
phase of the Ladinian-Norian 2nd-order cycle (T2, aggrading Carnian salt) illustrates a very high increase in the subsidence rate in comparison with the transgressive backstepping sequences of the Scythian-Ladinian cycle (T1, Muschelkalk carbonates). It is also observed that the peak transgression (Dolomie de Beaumont, Calcaires à Ceratites) occurred after the maximum rate of subsidence in both cycles. The older cycle, the Scythian-Ladinian (T1/R1) cycle, indicates subsidence rates somewhat lower than the upper cycle (T2/R2). It is developed in the east of the basin only as far west as the well Courde-manges, in Haute Marne. The younger cycle, the Ladinian-Norian cycle, is fully developed in the western part of the basin, where erosion has not occurred. This reflects the shift in depocenters in the Paris Basin, from the east to the west with time, reflecting the basin development that is also seen in the se-quence architecture (Fig. 15).

Differences are observable in these two cycles. The 1st-order flooding of the Ladinian was a global event and this supports a eustatic component in the cycle development. This transgression is recognized in the British Isles as the Waterstones unit, as far north as offshore Arctic Norway, North Africa and the Middle East (Hallam, 1992). This, however, does not rule out the importance of a tectonic component in the development of the cycle. In the Germany basin, extensional faulting was active at the time and produced thick Buntsandstein and Muschelkalk deposits. The trend of these depocenters lies to the northeast-southwest, which coincides with the structural trend in the Paris Basin at the time (Betz et al. 1987).

The succeeding Ladinian-Norian cycle (T2/R2) exhibits a higher rate of subsidence and has a more local, pronounced tectonic influence on the cycle development. This is due to the enhanced rifting activity known as the Carnian crisis in the Tethys and the Norwegian-Greenland rift in the North Sea (Wooster et al., 1992; Ziegler et al., 1988).

However, the 2nd-order flooding event, represented in most of the basin by the Dolomie de Beaumont, may also have been a eustatic event. This idea is supported by previous studies, which propose a climatic change at this time due to an extensive fluvial system which developed over northwest Europe. The 2nd-order flooding event may have been caused by a change in eustatic sea level. This theory is supported by the Haq et al. (1987) eustatic sea-level chart which indicates a long-term rise in sea level, beginning in Early Triassic time and reaching a peak in Carnian times. Isopach mapping of the 2nd-order cycles (Fig. 23) and the 3rd-order sequences confirms that the main faults controlled in the geometry and facies development in the basin, especially during the Ladinian-Norian cycle, when the Paris Basin became an individual entity.

A SEQUENCE STRATIGRAPHIC FRAMEWORK OF THE MARINE AND CONTINENTAL TRIASSIC SERIES 685
DISCUSSION AND CONCLUSIONS

A three-dimensional sequence stratigraphic framework has been established for the entire Triassic succession in the Paris Basin. This is defined up to 3rd-order frequency over the whole basin (Fig. 6). The Triassic deposits of the Paris Basin were deposited in a cyclic manner (Fig. 24). This is expressed initially as a period of localized tectonic subsidence that established the depocenter. As block faulting ceased, uniform subsidence followed, and the main transgressive periods occurred. As deposition continued with time, the maximum flooding surfaces gradually encroached westward. This encroachment occurred in four main stages. The Ladinian (K1 sequence boundary) acts as an important hinge in time, when regional tilting in the region of the Trois Fontaines well resulted in a major westward shift of the depocenters (Fig. 23).

In the western part of the Paris Basin, thick accumulations of sand bodies, often of various genetic origin, are stacked together. They have been deposited from a continental, fluvial or alluvial regime to a marginal-marine deltaic environment. Often there is no obvious discontinuity between these packages. How-
ever, using the sequence stratigraphic framework established in the basin center where the section is more complete and facies variations are more apparent, a stratigraphic time-framework can be tentatively applied to the succession. Retrograding sand packages deposited during a transgressive systems tract are subdivided from prograding sand packages deposited during the highstand systems tract. The Donnemarie Formation, which is a thick fluvial formation (greater than 150 m thick) is thought to have been deposited from Ladinian to Carnian times.

During the transgressive phase, the sand deposits were stacked and preserved in incised valleys cut during the relative lowstand. During the highstands, the prograding sands were preserved, but often dolomitized due to reflux dolomitisation (Purser et al., 1994). An emergent surface is sometimes present at the sequence boundary, often represented by a caliche or a dolomitized horizon. Lowstand deposits are difficult to distinguish within the succession due to the continental nature of the Triassic deposits. During the relative sea level lowstand, erosion predominated and during the transgression, the incised valleys were infilled. True lowstand deposits, if present, are difficult to distinguish from the transgressive succession, but their presence in the stratigraphy cannot be ruled out.

The cause of these cycles involves an important tectonic component, but also includes variations in climate and eustasy. Separating and quantifying the role of each factor is not easy. The Triassic series was deposited during a period of extensional faulting in a predominantly transtensional stress field, and therefore fault-block movement spilling coarse clastics into the basin was common. This phenomenon is linked to progressive erosion of the surrounding ancient massifs of Armorica and Ardennes. The sediment was deposited in several sequence cycles, which on a 1st-order scale represents the beginning of encroachment of the sediment onto the craton. Marine inundation into the basin tend to have been uniform and widespread and quite thin in relation to the continental deposits where sediment supply was high. Figure 25 illustrates the variety of 3rd-order Triassic type sequences that occur throughout the succession.

It is likely that the eustatic effects may be masked in the continental fluvial cycles where climate and tectonic uplift were
the dominant factors in shaping the sequence architecture. The maximum flooding surface in the Muschelkalk carbonates represents a widespread flooding event, is well dated and, as discussed previously, was likely to have had a eustatic origin. It has been identified as a 1st-order peak transgression. This is however in contradiction to the Haq et al.'s (1987) sea-level curves.

The Haq et al.'s (1987) curve shows a peak transgression in the Carnian, as discussed earlier. The Embry (1988) sea-level curve for the Triassic is based on the Sverdup basin in Canada, which occupies an area of over 300,000 km² and 4000 metres of Triassic sediments. The basin exhibits little tectonic disturbance and has been well studied. Embry (1988) recognizes the early Anisian, early Carnian and early Norian transgressive events and also the important sea-level fall in the late Ladinian (K1), but not the Ladinian transgression equivalent to the Calcaires à Ceratites (Hallam, 1992). These three transgressive events are also recognized in the Paris Basin as the flooding of the Grès Coquillers (B5), the Lettenkohle (K1) and the Dolomie de Beaumont (K3) respectively, but the latter is dated as being late Carnian in age.

However, problems still remain in the undivided areas, such as the lower Buntsandstein in the eastern part of the basin. The biostratigraphy is still poor and correlating the succession in
continental areas is difficult. More subsurface data, detailed sedimentological analysis and magnetostratigraphic data are necessary to resolve some existing problems. Stratigraphic modeling is a possible solution for solving some of these problems, refining the existing sea level curves and quantifying the various stratigraphic factors responsible for the sequences produced.

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INTRODUCTION

The South-East France Basin formed along the western margin of the Tethyan ocean (Fig. 1). The GPF programme (Géologie Profonde de la France—The Deep Geology of France national programme) has promoted drilling operations in the Ardèche area to study sedimentary and diagenetic transformations in a Triassic reservoir at the edge of a passive margin (Steinberg et al., 1991; Sureau, 1994). To the west, marginal Triassic series composed mainly of conglomerates and sandstones ranging from 30 to 120 m thick overlie the South-East Cévennes Variscan crystalline basement (Massif Central). In the western part of the Rhône Valley, basinward series up to 300 m thick have been penetrated by numerous wells where mainly evaporitic series are representative of the basin deposits.

Genetic relationships between continental and lagoon-marine clastics, carbonates and evaporites are better understood through the concepts of sequence stratigraphy (Vail et al., 1977; Wilgus et al., 1988; Van Wagoner et al., 1990). Observation of several unconformities and major boundaries in outcrops and through the concepts of sequence stratigraphy (Vail et al., 1977; Van Wagoner et al., 1990). Observation of several unconformities and major boundaries in outcrops and synsedimentary sedimentation. Despite these features, major boundaries and successions of backstepping and foresetting systems tracts enable correlations between 3rd-order sequences in Ardèche and other domains in western Europe, in particular in the German Basin. The main characteristics of the Ardèche series are wide onlaps over the crystalline basement during the Ladinian stage, despite the thinness of the series and occurrence of thick interbedded coarse-grained clastic and evaporite highstands in the Upper Triassic strata.

2nd-order sequences have been identified following 4 marine ingestions in the transect. The proposed Ladinian peak transgression is characterized by open-marine microfossils despite a clastic facies. Eustatic variations caused broad onlaps of backstepping series over a weakly subsiding Variscan Massif Central until late Ladinian time. In the Late Triassic Epoch, from the Carnian age onwards, mainly foresetting, thick clastic/evaporitic series were controlled by a more active tectonic subsidence.

REGIONAL SETTING

Triassic strata crop out at the edge of the Massif Central crystalline basement near Aubenas (Fig. 1). In this western part, thin (<100 m thick) sandstone and conglomerate series characterize the marginal/landward, mainly clastic facies that pinch out seawards. The boundary of the Triassic depositional area lies beyond the westernmost erosional limit, although probably not very far beyond, judging from the distance between similar border facies and the visible depositional limit at the northeast edge of the Massif Central north of Lyon (Courel 1973; Aubagne et al., 1984). Sandstones were mined for Pb-Zn sulphides in the Largentière area until 1980, which provided ample opportunities for observations in boreholes and galleries. On the transect, the S53 borehole is representative of these western marginal/landward Triassic series.

Many boreholes in the Rhône Valley graben have cut Triassic series below the Jurassic-Cretaceous cover. The Triassic series thicken, especially the upper strata, from the S53,49.7 and Uzer boreholes in the west (Fig. 2) to Valvignères in the east (Fig. 3), where it attains 398 m thickness (Rhaetian excepted). The series continues to thicken eastward from Valvignères, but the Triassic basement has not been reached. Between Uzer and Valvignères, the Balazuc and Morte-Mérie boreholes were sunk as part of the G.P.F. Programme on either side of the major Uzer fault. Such synsedimentary faults (Elmi, 1990; Giot et al., 1991 a, b; Roure et al., 1992; Bergerat and Martin, 1993) con-
trolled the west to east thickening related to the development of evaporitic facies toward the basinal eastern domain. Such a transitional domain between platform and basin is a key region for studying the relative importance of eustatic events and local tectonics in controlling Triassic onlaps.

Triassic sedimentation only started along the Triassic transect in the Ardèche during the Anisian period. Older series (Lower Triassic, Buntsandstein facies) have been recognized in depocenters in the South-East France basin, in Provence (Glintzboeckel and Durand, 1984), Languedoc (Busson et al., 1984) and Bresse (Dromart et al., 1992). Scythian series (Lower Ladinian age (determination of F. Hirsch, 1984) were described in Bresse, Provence and south Subalpine ranges (Dromart et al., 1992; Glintzboeckel and Durand, 1984; Richards, 1994). A Scythian sequence 1 may be distinguished from an Anisian sequence 2, both older than the Aniso-Ladinian sequence 3 of the Grès du Roubreau (in the Balazuc well) is Aniso-Ladinian with dominantly

Sequence 3: Grès de Base and Grès du Roubreau

Sequence 3 is marked by a major onlap of several tens of kilometers of first fluvial and then lagoon-marine series, directly overlying the basement over the whole studied domain. The lagoon-marine levels are then overlain by a new fluvial series.

From the base up, in any section (Figs. 2, 3), proximal fluvial conglomerates and coarse-grained sandstones give way to distal fluvial fine-grained sandstones and then to lagoon-marine fine-grained sandstones and siltstones with vertebrate footprints (Demathieu et al., 1984) and halite pseudomorphs. Siltstones and dolomite-cemented mudstones with Chondrichthyan scales and conodonts overlie the basal clastic series (Courel et al., 1980; Finelle, 1981; Courel et al. 1984; Cula and Courel, 1987).

In the upper part of sequence 3, medium-grained sandstones and siltstones constitute a renewed distal fluvial system, capped by a pedogenetic top surface. Eastward, in the basinal domain (Valvignères borehole), anhydritic interbeds appear in the upper sandstones.

Working landward from east to west, in the basal clastic sequence, fluvial facies are increasingly proximal, while distal facies pinch out in the same direction. In the outcropping domain, near the S53.49.7 borehole, fine-grained lagoonal sandstones and siltstones directly overlie the basement. In contrast, coarse fluviatile deposits accumulated in wide troughs. Numerous field observations suggest that the trough pattern was technically controlled (Cula and Courel, 1987).

All the above observations are consistent with the interpretation of a proximal to distal and lagoon-marine TST, ending with a marine fossiliferous level MFS of dolomite-cemented mudstones with Chondrichthyan scales and conodonts. The grey color of the middle sandstones suggests preservation of organic matter (<1% TOC) in a transgressive systems tract, as suggested by Thurow (1993). The upper part of sequence 3 is marked by the progradation of a fluvial system in a regressive systems tract. In some outcrops, a bored hardground on top of sequence 3 suggests a discontinuity.

The base of the Grès du Roubreau in sequence 3 is dated late Anisian-early Ladinian by vertebrate footprints (Demathieu et al., 1984; Ischoirotherium felenci, I. delicatum, Brachychirotherium circarparvum, Sphyngopus ferox, Coelurosaurichnus largeteriensis). The lowermost palynologic association observed in the Grès du Roubreau (in the Balazuc well) is Aniso-Ladinian with dominantly Triadispora (T. aurea, T. falcata, T. plicata, T. suspecta), associated with subordinate Illinites chitonioides (Fauchonnier et al., 1996). Above the sandstones/mudstones with footprints and palynologic associations, the conodonts of the MFS 3, observed in outcrops 30 km southward of Largentière (Finelle, 1981; Courel et al., 1984) were determined to be Pseudofurnishius hudlei, which seem to be of early Lower Ladinian age (determination of F. Hirsch, 1984).
Fig. 2.—Boreholes S53.49.7 (SMMP) and Uzer-la-Courèze (SMMP) (Ardèche, France): sections.
Fig. 3.—Boreholes Balazuc 1 (GPF Programme) and Valvignères (SNPA) (Ardèche, France): sections.
Sequence 3 may correspond to the Anisian-Ladinian middle/upper Muschelkalk sequence in the German Basin. Marine limestones in Lorraine, Provence and the Jura mountains could be lateral equivalents of the foresteepr upper part of the Grès du Roubreau sandstones. Outcropping dolostones in the Crussol hills, 40 km to the north, are another marine fossiliferous lateral equivalent (Ricour, 1962; Cula, 1987; Cula and Courel, 1987).

The scale of the major Ladinian onlap, which is marked by a massive advance over the crystalline basement, is probably linked with weak subsidence of a flat paleotopography and a large sea-level rise. Basin-fill modelling (Poli, 1993) indicates that sequence 3 deposition in the Ardèche is characterized by an important eustatic rise in sea level in comparison with the total subsidence (Fig. 5).

**Sequence 4: Grès et Argilites Vertes**

Proximal fluvial conglomerates and coarse-grained sandstones (braided and ephemeral streams) in the landward domain overlie the distal medium-grained sandstones topping sequence 3. Above this granulometric discontinuity, a fining-upward trend is observed in the lower part of the sequence, followed by a coarsening-upward trend in the upper part where sandstone beds thicken and are topped by pedogenetic structures. Sulphate nodules and pedogenetic structures commonly occur throughout the sequence. In the middle part, green siltstones/mudstones contain sparse plant debris (Samama, 1969).

Basinward, in the Valvignères borehole, sequence 4 begins with fine-grained sandstones overlain by clayey anhydrite.

In this type of continental succession, clayey anhydritic series in the east and fine-clastic series in the west could mark the reversal between a lower fining-upward transgressive systems tract and an upper coarsening-upward (west) or dolomitic (east) regressive systems tract. Sequence 4 series are younger than the Upper-Anisian/Lower-Ladinian sequence 3 and older than the Upper Ladinian base of sequence 5.

**Sequence 5: Grès et Argilites Grises Silteuses and Barre Carbonatée Médiane**

In the western proximal domain, sequence 5 begins with fine sandstones, fining-upward and basinward, which are very grad-
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Fig. 5.—Proximal-distal Triassic transect along the Peritethyan margin in Ardèche.

Fig. 5—Proximal-distal Triassic transect along the Peritethyan margin in Ardèche.

clastic (fossiliferous beds, Finelle, 1981) and micritic dolostones capped by stromatolitic laminated dolostones with epi-
genetic sulphate crystals. Doubinger and Adloff (1977) recognized the Ladinian/Carnian boundary at the base of the Barre Carbonatée Médiane where Camerosporeites, Duplicipores, Paracirculina and Praecirculina first occur and rapidly increase, while the bissaccates percentage decreases (Triadispera suspecta replaced by T. aurea). The Barre Carbonatée Médiane could be a lateral equivalent of the Dolomie limite of the Lettenkohle (base of the Carnian) in the Paris Basin.

Basinward, in the Valvignères borehole, anhydritic and then halitic series are overlain by mudstones/claystones, topped by anhydrite series. The clay peak in the middle part of the sequence separates a lower evaporitic facies, which may be a lateral equivalent of the Grès et Argilites grises silteuses, from an upper evaporitic facies, the presumable lateral equivalent of the Barre Carbonatée Médiane.

In sequence 5, the boundary of the Barre Carbonatée Médiane, above lagoon-marine evaporitic mudstones, may represent a maximum flooding surface ending a transgressive systems tract. The shallowing-upward dolomite or the anhydritic series may represent a regressive systems tract.

**Sequence 6: Argilites Sulfatées Supérieures Halitiques et Anhydritiques**

Above the dolomite topping the Barre Carbonatée Médiane, sequence 6 begins with mudstones in the western margin and in the east with a few-meters-thick anhydrite level directly covered by a few-meters-thick massive halite. Judging from the occurrence of anhydritic and massive clay-free halite beds directly above the Barre Carbonatée Médiane, the very shallow (stromatolites) carbonate setting ending sequence 5 is replaced by a basinal depositional environment of up to 10 m (?) in depth.

In the western marginal domain (Fig. 2), mudstones with some slumped fine-grained sandstones and rare anhydrite interbeds (in some small depocenters in the Largentière area) overlie the dolomite. In the upper part of the sequence, coarsening-upward sandstones are cemented by dolomite. The lower part of the mudstones and their distal anhydritic equivalents may represent a transgressive systems tract. Coarsening-upward dolomitic sandstones above may form a regressive systems tract.

In the Balazuc borehole in the east (Fig. 3), a 2-m-thick anhydrite bed with thin laminated anhydrite/dolomite interbeds covering the dolomite and is directly overlain by a 1.20-m-thick massive halite unit. Such a succession characterizes a landward migration of distal basinal deposits. Backstepping of confined anhydritic/halitic deposits may appear paradoxical. However, transgressive halite series have been well-described in the Triassic Saharan platform (Busson, 1971) and the Triassic System of the Paris Basin (Bourquin and Guillocheau, 1993). Sharp thickness variations of evaporitic deposits in the Ardèche outcrops and basin-fill modelling along the transect (Poli, 1993) suggest that syndepositional tectonics may have controlled a large landward advance of anhydrite/halite regressive deposits in the more subsiding areas. An anhydrite/mudstone series overlies halite deposits, where more clayey beds may mark open basinal facies (i.e. maximum flooding surface). In the up-
per part of sequence 6, anhydrite-bearing mudstones become progressively more clastic (e.g. siltstones) upward. The general trend of this regressive upper part of the sequence 6 is interpreted as a HST.

In the eastern basinal domain, in the Valvignères borehole, mainly halite deposits form a transgressive systems tract at the base, as at Balazuc. Above, claystones are interbedded in a mainly anhydrite series. Reversal between transgressive and regressive systems tracts can be marked above a clay peak in the anhydrite series.

Sequence 6 is the first significant evaporitic unit at the edge of the Massif Central beyond the subsiding areas of the South-East France Basin in the Rhône and Bresse grabens. It is also the continuation of a strongly subsiding stage initiated during the Barre Carbonatée Médiane deposition, all along the western marginal domain of the Tethys (Fig. 5). The earliest appearance of thick massive evaporites over the margin of the domain is dated early Carnian, the base of the Carnian being situated at the base of the underlying 5-m-thick dolomite of sequence 5. Basin-fill modelling has quantified the sharp increase in the subsidence rate at the Ladinian/Carnian boundary (Poli, 1993). Local tectonic conditions may have controlled a major local transgressive onlap, characterizing marginal structuration of the western Tethys (see increase in tectonic subsidence rate, Fig. 5).

Sequence 7: Ensemble Gréso-dolomitique Gris and Base of the Formation bariolée d’Ucel

Throughout most of the Ardèche transect, sequence 7 begins with a renewed sandstone series. Basinal evaporitic series are restricted to the Valvignères borehole in the east.

In the western outcropping area (Fig. 2), sequence 7 series are composed throughout of coarse-grained sandstones and conglomerates. Fining- and thinning-upward clastic bodies at the base are progressively replaced in the upper part by large, coarsening- and thickening-upward, dolomitic cemented clastic bodies. Between the two reversed trends, several meters of grey fine-grained sandstones interbedded with grey mudstones/siltstones display facies similar to the lagoonal Grès du Roubreau or the Grès et Argilites grisiformes formations of sequences 3 and 5. Lagoonal transgressive series in which palynologic associations were preserved in the Balazuc well could have separated a basal transgressive systems tract from an uppermost regressive systems tract.

In the Balazuc borehole, in the east, 10-m-thick conglomeritic and coarse-grained sandstones are overlain by 20-m-thick grey mudstones alternating with siltstones/fine mudstones and dolostones. Many pollen-rich samples of undetermined age were observed in this grey section (Fig. 3). Grey mudstones/dolostones with preserved organic matter and pollen overlying fining- and thinning-upward coarse clastic series suggest a lagoonal onlap as in the Ladinian series described. Then the 80-m-thick mainly conglomeratic upper part of sequence 7, the basal part of the Formation Bariolée d’Ucel, covers the lagoonal grey fine-grained series. The coarse, often composite clastic bodies characterizing ephemeral streams in variegated colored floodplain mudstones fine upward and rapidly pinch out laterally (Fig. 3). The general coarsening- and thickening-upward trend of the upper part of the sequence characterizes basinward prograding parasequences, ending with dolomitic pedogenetic structures.

In the eastern basinal part of the transect, in the Valvignères borehole, a massive halite/anhydrite lower part is covered by mudstone/anhydrite series and thick aggrading-prograding siltstones with mudstone/anhydrite interbedding.

The basal part of sequence 7 seems to be characterized in the western margin of the Ardèche domain by coarse sandstones and then by mudstones and dolostones (Fig. 2) that backstep over the western border of the Peritethyan domain. These are the lateral equivalent of massive evaporitic series in the eastern, more subsiding areas. In a Carnian section of this type, the coarse clastic > mudstones > dolostones succession bears some resemblance to the Grès à roseaux > Dolomie de Beaumont pattern encountered in Lorraine (Bourquin and Guillocheau, 1993). The TST upper part of the Grès à roseaux might have onlapped the margin domain, while the LST lower part could be restricted to more subsiding depocenters in Bresse graben (Dromart et al., 1992) and in the German Basin (Aigner and Bachmann, 1992).

The upper part of sequence 7 forms a large regressive systems tract of coarsening-upward continental series. Coarse sandstones in the western domain and fine anhydritic clastics in the eastern domain prograde basinward (Vannier, 1991).

Sequences 8 & 9: Middle and Upper Part of the Formation Bariolée d’Ucel

Above sequence 7, conglomerates and sandstones together with subordinate siltstones and mudstones compose the main facies of the Upper Triassic strata. Anhydrite series are however interbedded in the basinal domain in the Valvignères borehole. No biostratigraphic data are available for the two sequences. They are younger than the Carnian sequence 7 and older than the Rhaetian sequence 10. Similar sandstone series in Bresse and the Jura mountains have been dated as Late Carnian/Norian (Adloff et al., 1984). There are obvious difficulties in defining their sequence boundaries and maximum flooding surfaces. Even the division into two 3rd order sequences is debatable. Be that as it may, two sequences, each 10–50 m thick, display the same pattern. Conglomerates and then fining-upward conglomerates and sandstones are observed along the entire transect. Above, from west to east, all along the transect, fine sandstones, mudstones/claystones and anhydrites mark a fining in clastic grain size and, in the eastern area, fine clastics vanish and are replaced by evaporites. These fine deposits are covered by coarse sandstones, coarsening- and thickening-upward, with dolomitic pedogenetic structures in the upper beds of regressive parasequences. Within the two sequences, a basal fining-upward succession could represent a TST as far as a clayey-evaporitic maximum flooding surface, covered by a coarsening-upward regressive HST. It is difficult in such main continental sequences to interpret basin-fill controls. Proximal ephemeral streams have been identified in the coarse sandstones, indicating dry climatic conditions. Differences in local tectonic conditions controlled the location and dynamics of fluvial transport (Oujidi, 1988). The proposed subdivision into two 3rd order sequences (Sequences 8 and 9), as in the German Basin (n° K4 and K5 in Aigner and Bachmann, 1992), could suggest similarities in infilling dynamics in western Europe.
Sequence 10 begins with renewed conglomerates and coarse sandstones in the west and fine sandstones and siltstones in the east. In the base of the middle part, grey sandstones/siltstones with bivalves such as *Avicula contorta* characterize the Rhaetian TST. Coarse continental clastics under the marine Rhaetian series could be either Norian or Rhaetian and represent a TST or an LST. Whatever the answer, the important event is the occurrence of a large marine Rhaetian onlap over continental Carnian/Norian deposits at the edge of the French Massif Central. Basin-fill modelling (Poli, 1993) indicates the comparative importance of a rise in relative sea level in the advance of the marine series over the weakly subsiding Variscan domain. The Rhaetian maximum flooding surface has been chosen as the upper limit of the study.

SECOND-ORDER SEQUENCES

The Ardèche transect is characteristic of a continental margin domain with marine ingression (Fig. 5). When identifying 2nd-order cycles, the relative importance of the four recognized marine ingression from the Anisian to the Rhaetian should be taken into consideration.

The earliest marine ingression is in the Ladinian sequence 3, which forms a 30-km onlap over the crystalline basement. A 3rd order maximum flooding surface has been identified at the top of a transgressive series in a mainly clastic series with evaporitic claystones. Marine fauna with Chondrichthyan scales and sepharadile conodonts with Ladinian morphology and sepharadile affinities (*sensu* Hirsch, 1977 who described them in the Mediterranean basin) have been recognized in an outcrop at Aujac 30 km from Largentière (Finelle, 1981). Dolostones have been identified at the same level associated with lamellibranch fauna and foraminifera of Ladinian age, 40 km to the north at Crussol and 50 km to the south at Valbres (Avias, 1961; Finelle, 1981; Courel et al., 1984). *Pseudofurnishius hudnelli* conodonts with Ladinian morphology and sepharadile affinities (*sensu* Hirsch, 1977 who described them in the Mediterranean basin) have been recognized in an outcrop at Aujac 30 km from Largentière (Finelle, 1981). The sequence 3 maximum flooding surface is a first candidate for a 2nd order peak transgression. It could correspond to the Fassanian main transgression in the alpine domain (Bestädt, person. commun., 1994).

The second marine ingression is that of the sequence 5 Barre Carbonatée Médiane at the base of the Carnian. It is dated by palynological assemblages, (Doubinger and Adloff, 1977) and contains marine fossils—lamellibranchia and foraminifera (Finelle, 1981). Depositional environments were shallow, sometimes oolitic or confined with epigenetic sulphate crystallization. This is the only dolomite marine level described in the transect. Dolostones disappear eastwards, as at Valvignières, and have not been observed north of Valence where the Carnian evaporite series overlies sandstones (Courel, 1973). The dolomite marine character of the Barre Carbonatée Médiane makes it a second candidate for a 2nd order peak transgression. It may correspond to the base of the Carnian (Raibl transgression in the Alpine domain, Bestädt and Schweizer, 1991).

The base of sequence 7 is clastic in the Ardèche outcrops and the Balazuc borehole. It is made up of laminated fine sandstones and grey siltstones associated with dolostones in alternating backstepping sequences. Preserved organic matter, especially in levels containing palynological assemblages, recalls the lagoon-marine facies of the Grès du Roubreau. Such backstepping associated with lagoon facies in proximal continental series has been interpreted as marginal evidence of a Carnian marine ingression. These characteristics are not sufficient, however, to make it a new candidate for a 2nd order peak transgression in the Ardèche. It should be noted, though, that it is the equivalent of the transgressive level beneath the Dolomie de Beaumont of Lorraine (Bourquin and Guillocheau, 1993).

Sequence 10 contains backstepping series with Rhaetian lamellibranch marine fauna. These are, however, only the start of a major transgression at the base of the Lias (Dromart et al., 1991; Elmi et al., 1991), which is not covered by this study.

Only the first two (Ladinian sequence 3 and Carnian sequence 5) of the four marine transgressions referred to can be retained as candidates for a 2nd order peak transgression. Basin-fill modelling of the study transect (Poli, 1993) indicates that the two ingressions occurred in different tectonic contexts. For the Ladinian sequence 3 ingression, the amplitude of tectonic subsidence relative to accommodation was less than the amplitude of eustatic variations (Fig. 5). For the Carnian sequence 5 ingression, the scale of tectonic subsidence, calculated under the same conditions, was much greater than eustatic variations. This situation observed in the transect also seems to characterize all transgressive Ladinian series over the ancient Variscan reliefs of South-West Europe. They are thin but onlap over the French Massif Central (Courel et al., 1990), les Pyrénées (Fréchhengues and Peybernès, 1991), the Ardennes (Berners et al., 1984), the Vendelician ridge (Lemoine and de Graciansky, 1988), the Iberian domain (Sopena et al., 1988; Muñoz et al., 1992) and Sardinia (Gandin et al., 1982). In contrast, in the Carnian, the series are thick, with sudden variations related to early faulting. The Ladinian sequence 3 ingression in the Ardèche is therefore characterized by a broad onlap and an open-marine environment in a context of eustatic variation without marked tectonic subsidence. The Carnian sequence 5 ingression is characterized by a marine environment with exclusively coastal fauna and a tendency towards evaporite confinement in a context of marked tectonic subsidence. Therefore it seems preferable to choose the maximum flooding surface of sequence 3 as a 2nd order peak transgression in the Ardèche transect.

The 2nd order maximum downward shift phase(s) are difficult to evaluate in the Ardèche transect. No LST has been formally identified for the evaporite series, and the halite series, which display maximum confinement, have been interpreted at Balazuc as belonging to a TST (base of sequence 6). The thickness and grain size of the sandstone series is further closely controlled by major synsedimentary faulting. When determining 2nd order trends more weight should be given to arguments of global scope than to those with regional scope. In the external clastic domain, conglomerates are well developed in coarse-grained sandstones at the bases of sequences 7, 8 and 9. The Upper Triassic sandstone series, below the Rhaetian, displays a tendency towards thicker beds and a coarsening-upward pattern in the upper regressive part (HST) of the same three sequences. The reversal of the trend between essentially regressive series and the transgressive unit that developed in the Rhaetian may be located at the conglomerate base of sequence 9, which might be the maximum fall in sea level. This suggestion should be minimized because of the difficulties in locating global trends in such proximal facies.
SEQUENCE STRATIGRAPHY ALONG A TRIASSIC TRANSECT ON THE WESTERN PERITETHYAN MARGIN IN ARDECHE
CONCLUSION

A division into 3rd order sequences is proposed for the Ardèche transect of the western margin of the Tethys, which is
characterized by mainly marginal facies. Abrupt variations in
thickness were controlled by synsedimentary faulting. Despite
these specific features, correlations are proposed with the Germanic domain in particular (Aigner and Bachmann, 1992) and
with other domains in western Europe (see in this volume).
Eustatic variations caused broad onlaps of transgressive marine series over weakly subsiding Variscan massifs until Late
Ladinian time. Although very thin, these series are useful for
establishing correlations.
In the Upper Triassic Epoch, from Carnian time onwards,
tectonic subsidence became more active in the Ardèche transect
area. This characteristic of the marginal region studied, where
faulting controlled sizeable local variations in thickness, was
perhaps a marked feature throughout the South-East European
domain. This would imply that sea-level variations in this region are essentially the result of combined tectonic and eustatic
change (Mörner, 1980).
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MULTORDER SEQUENCE STRATIGRAPHY IN THE TRIASSIC SYSTEM OF THE WESTERN SOUTHERN ALPS

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ABSTRACT: A hierarchy of depositional sequences, from long-term megasequences to transgressive/regressive sequences to Milankovitchian cyclothems, is recognized in the well-known stratigraphic record of the Southern Alps. The sedimentary succession deposited between the Hercynian and Alpine orogenies can be subdivided into three largely tectonically-controlled megasequences, with boundaries occurring in the Upper Triassic and lowermost Cretaceous successions. Patterns of sediment accumulation in the Scythian-Carnian part of the Triassic succession were primarily controlled by successive episodes of transpression and transtension with associated volcanism, indirectly linked to the opening of Neotethys in the east. In the Norian-Rhaetian stages, however, subsidence markedly increased, leading to final drowning and deposition of pelagic sediments following late Hettangian time. Transtensional movements are interpreted as related to the initial development of a narrow rift system connecting the Neotethys with the future site where the central Atlantic will open in Middle Jurassic time. Within the Triassic succession, five transgressive/regressive supersequences punctuated by major transgressive episodes are interpreted as tectono-eustatic in origin, since they are on one hand characterized by different subsidence and structural patterns, but on the other also broadly correlate with worldwide long-term changes in sea level. Besides eustasy and tectonic activity caused by Neotethyan rifting, sequence architecture was also influenced by several factors, including climate, sediment supply and changes in the productivity of “carbonate factories”. In total, 16 sequences are recognized, with an average duration of 3 my. The first supersequence (S) is confined to Lower Triassic strata. While five sequences occur in the eastern Southern Alps, only three to none of the sequences could be identified from eastern to western Lombardy. The Middle Triassic supersequence (A-L) and the Carnian supersequence (C) may be subdivided into five and four sequences, respectively. The fourth supersequence consists of one sequence only (N1), whereas the fifth one, extending into the Upper Hettangian deposition (N2-H), can be subdivided into at least three sequences.

INTRODUCTION

The Triassic succession of the Southern Alps (Fig. 1) provides an excellent case to test the application of sequence stratigraphic concepts, since stratigraphic research has been active for more than a century. The Southern Alps are also the type-area for two Triassic stages: the Carnian and the Ladinian, and southern alpine Triassic rocks provided the first evidence for basic concepts in geology, such as the intertonguing of facies (Mojsisovics, 1879).

Palaeontological tools are refined enough to establish a fairly accurate and well-controlled biostratigraphic scheme (Table 1). Good resolution is obtained for Lower Triassic and Middle Anisian to Ladinian strata, which commonly contain ammonoids and conodonts. At other times, characterized by widespread shallow-water sediments, less precise age information is provided by foraminifers, brachiopods, algae and bivalves.

Geographically, the Southern Alps (Fig. 1) can be subdivided into an eastern part, including the Dolomites as well as the Recoaro and Carnia regions, and a western part, located between Lake Garda and Lake Maggiore. A complete sequence stratigraphic analysis should consider the whole area, since Lower Triassic sediments are fully represented only in the E, whereas the Upper Triassic succession is better developed in the west. In the present paper, we will deal only with the western Southern Alps (Fig. 2).

FACTORs CONTROLLING SEDIMENTARY EVOLUTION

Besides global eustatic changes, sedimentary evolution was controlled by the interaction of several regional factors, including tectonism and subsidence, volcanism, nature and amount of sediment supply, and climate.

Tectonism

The area occupied today by the Southern Alps was submerged during the Triassic by a mostly shallow sea way linked to Neotethys (Fig. 3). Since Permian times, when the peri-Gondwanan microplate fringe detached from Gondwana, the Southern Alps “hinge zone” has been dominated by tectonic movements caused by W-ward propagation of Neotethyan spreading (Marcoux et al., 1993a). At the western corner of the Neotethyan oceanic triangle, rotations and strike-slip displacements of microplates were active at several stages (Doglioni, 1987). Transtension and associated volcanism, which possibly resulted from local oblique convergence (Marinelli et al., 1980; Garzanti, 1985b), were active from Middle Triassic to at least mid-Carnian. After a major geodynamic change occurring towards the close of the Carnian, the Southern Alps were affected by renewed extension until the Middle Jurassic time, when the region became the passive margin of the Adria microplate, facing the Piedmont-Liguria oceanic sea-way linking Neotethys to the Central Atlantic.

The impact of tectonic extension can be evaluated by calculating subsidence rates through time. Besides decompaction problems, uncertainties are due to considerable discrepancies between Triassic time scales, especially with regard to the Early and Middle Triassic. The estimated duration of the Early Triassic ranges, for instance, from a minimum of 4.1 m.y. (Harland et al., 1989), to a maximum of 10 m.y. (Haque et al., 1988), with intermediate figures of 5 and 7 m.y. given by Odin and Odin (1990) and Menning (1990) respectively. Recent efforts to calculate age figures from counting of superposed Milankovitchian cyclothem, on the other hand, are largely based on weak assumptions and circular reasoning. Moreover, data reported from the Latemar buildups in the Dolomites (Goldhammer et al., 1990), if extrapolated to the entire Ladinian, would imply a duration of over 20 m.y., which seems excessive (Brack and Rieber, 1991; 1993).

The most reasonable estimates of accumulation rates, which approach subsidence rates for this largely shallow-water succession, are those calculated according to the Haq et al. (1988) scale, which is adopted herein also because it best approaches the accurate age of 251.2 ± 3.4 Ma recently obtained for the Permian/Triassic boundary (Claué-Long et al., 1991). However, it should be noted that the estimated ages of the Triassic/ Jurassic boundary (210 Ma and 208 Ma according to Haq et al.
and Harland et al. respectively) do not match the radiometric age of 201 ± 1 Ma obtained in the Newark Group (Dunning and Hodych, 1990).

Two major stages of accelerated subsidence (Table 2), also indicated by platform dissection and deepening of intraplatform troughs, occurred in Late Anisian-Ladinian and particularly in Norian-Rhaetian time, when up to over 4000 m of mostly carbonate sediments accumulated in the western Southern Alps (Fig. 4).

Volcanism

The Southern Alps were the site of significant volcanic activity from Late Anisian to middle or even Late Carnian times. Active centres were located both within and to the south of the study area, where they are buried today beneath the Po Plain. Huge volumes of volcanic material were stored in basins and lagoons as volcanic arenites, tuffs and lava flows. Detritus rapidly filled intraplatform troughs at times of lowered sea level, as in the case of the Ladinian Wengen Group, or pushed the shoreline seaward even at times of rapid sea-level rise, as in the case of the Carnian Val Sabbia Sandstone.

Nature of Sediment Supply

Sandstones and minor conglomerates are widespread in the Lower Triassic Series, whereas in the Middle Anisian they were deposited only in proximal areas close to local uplifts. In Ladinian times, sandy turbidites were confined within intraplatform troughs. Terrigenous detritus again became widespread in the Carnian. Great volumes of clay, derived from the European-Iberian continent, accumulated in the western Southern Alps area during Late Norian-Rhaetian time.

Carbonate sediments, common throughout Triassic deposits from Lombardy to the world-famous peaks of the Dolomites, were mostly produced by the reef-building activity of organisms, after they recovered from the Permian/Triassic (P/T) crisis. The first buildups, which are lacking in the Scythian all over the world, are found in the Middle Anisian when terrigenous influx was reduced (Gaetani et al., 1981), and reached their maximum development in Upper Triassic deposits (Flügel, 1981; Fois and Gaetani, 1984). Another major biotic crisis occurred at the close of the period (Fagerstrom, 1987, p. 407).

Climate

The climatic signal is difficult to extract from the sedimentary record. It is widely held that in Triassic times the western end of Neotethys was hot and arid, with seasonal periods of higher rainfall (Tollmann, 1985; Garzanti, 1986). Widespread continental redbeds and evaporitic sediments, from dolostone to anhydrite and gypsum, indicate that dry conditions prevailed from Late Permian to Anisian time, and again during the Carnian to Early Norian, possibly with a mid-Carnian pluvial episode (Simms and Ruffell, 1990).

The Ladinian age was probably more humid (Scheuring, 1978). Then in the Late Norian-Rhaetian age, major influx of clay may have occurred during a wet stage, as deduced from the continental herbaceous origin of the kerogen (type II; Mattavelli and Novelli, 1990).

MULTIORDER SEQUENCES

The sedimentary record of continental margins typically consists of long-term sequences (Hubbard et al., 1985; Boote and Kirk, 1989; Bosellini, 1989; Gaetani and Garzanti, 1991), each made by nested sets of shorter-term sequences and cyclothems stacked in aggrading, backstepping, infilling and forestepping stratal packages. Multiorder cyclicity in the Triassic succession of the Southern Alps (Fig. 5) is documented in the present paper, which focuses expressly on longer-term sequences (for description of “Milankovitchian” cyclothems see Jadoul and Gnaccolini, 1992, and references cited therein).

Megasequences

The Upper Carboniferous to Eocene succession of the Southern Alps may be subdivided into three mainly tectonically-controlled megasequences, with duration intermediate between “Continental Encroachment” and “Transgressive/Regressive” cycles (Vail, 1992). From Westphalian to Carnian time, basin subsidence is inferred to have been largely caused by transtensional movements linked to the opening of Neotethys.

The megasequence boundary recognized within the Upper Triassic Series broadly correlates with a worldwide reorganization of plate motion and change in tectonic style (Veevers, 1989; Garzanti, 1993). Sedimentary evolution during the latest
MULTIORDER SEQUENCE STRATIGRAPHY IN THE TRIASSIC SYSTEM OF THE WESTERN SOUTHERN ALPS

Carnian to earliest Cretaceous is in fact thought to mirror active rifting in the W, finally leading to the opening of the Central Atlantic in the Middle Jurassic Epoch. Drowning with rapid transition from shallow- to deep-water sediments was recorded in Upper Hettangian strata, whereas transition from rifting to spreading might be recorded by sedimentation of radiolarites in late Middle Jurassic times. Deposition of the third megasequence, ranging from the earliest Cretaceous to the Middle Eocene times, occurred under the direct influence of oblique convergence and collision between Europe and the Adria microplate.

Supersequences

Five major transgressive/regressive sequences can be recognized within the Triassic succession: (1) Scythian (2) uppermost Scythian? to Upper Ladinian (3) Carnian (4) uppermost Carnian to Middle Norian and (5) Upper Norian to Upper Hettangian. They lasted from 8 to 13 my, and compare fairly well with supercycles UAA-1, UAA-2, UAA-3, UAA-4 and UAB-1 and 2 of the Haq et al. (1988) chart. This correspondence with long-term sea-level fluctuations does not necessarily imply dominant eustatic control, since stratigraphic information from the Southern Alps was largely used to infer the “global” curve (Haq et al., 1988, p. 108).

The Lower Triassic supersequence (S) was deposited in shallow seas transgressing from the SE. During a stage of relative tectonic quiescence, marine sediments progressively onlapped alluvial clastics derived from Hercynian basement rocks and Permian volcanics (Assereto et al., 1973). The Middle Triassic supersequence (A-L) was mainly influenced by transtensional tectonics, with widespread volcanism and local transpressional uplifts (Pisa et al., 1980; Doglioni, 1984). The effects of eustasy and tectonism are particularly difficult to separate during this cycle, which coincides with the overall eustatic rise testified by ingressions of the Muschelkalk sea onto Continental Europe. The Carnian supersequence (C) documents major regression at its top, with widespread sabkha conditions and gradual reduction in volcanic activity. The Lower-Middle Norian (N1) and Upper Norian to Hettangian supersequences (N2-H) were dominated by extensional tectonics. Strong subsidence coupled with high carbonate productivity produced very thick sections in the western Southern Alps. Shallow-water conditions predominated until the Late Hettangian time, when large parts of the Adria margin were drowned and covered by pelagic limestones.

Depositional Sequences

Within each supersequence, several 3rd-order “Vail-type” sequences can be recognized. A straightforward application of
sequence stratigraphic concepts (Haq et al., 1988; Vail, 1992), however, is commonly prevented by limited lateral continuity, local tectonic effects and lack of adequate biostratigraphic control. At the present we see no safe way to filter the eustatic signal from other factors controlling the sedimentary record. This paper only represents a first attempt at such a difficult task.

THE LOWER TRIASSIC SUPERSEQUENCE

The Lower Triassic Series in the Southern Alps is invariably represented by quartzose clastics, interbedded with mudrocks and carbonates. Stratigraphic sections are much thicker and complete in the east (Werfen Formation of the eastern Dolomites) than in the west (Servino Formation of Lombardy), where undeformed continuous sections are rarely exposed. The Servino Formation is only 35 m thick on the west bank of Lake Como, and is even missing altogether in places farther to the west; east of Lake Como its thickness rapidly exceeds 300 m.
Accumulation rates ranged up to 70 m/my in the eastern Dolomites (Broglio Loriga et al., 1990). Lower values around 15 m/my are calculated for the central Southern Alps (Adige valley to Valsassina). Exceedingly high figures (up to 180 m/my) would be calculated using the Harland et al. (1989) scale, which gives an unlikely duration for the Scythian (3.9 my instead of 10 my).

The Early Triassic is generally thought of as a stage of volcanic quiescence and reduced tectonic activity. The Servino Formation documents a major increase in quartz content and mineralogical stability with respect to the underlying Verrucano Lombardo, which resulted from greater contribution from the crystalline basement and “aporphyric” quartzose clastic sedimentary cover, possible recycling of Permian clastics and longer residence time of detritus in soils at the time of the P/T boundary, progressively more subdued relief, and prolonged reworking in high-energy coastal settings. However, the upward increase in feldspars, particularly evident in the Lake Como area, points to renewed tectonic activity close to the Smithian/Spathian boundary. Deeper erosion levels were reached, suggesting active uplift of the source areas. Deposition of pebbly fluvio-deltaic layers in the northwest, passing laterally towards the southeast to prodelta mudrocks and marine carbonates, indicates that this basement high, established in the Permian and active until the Carnian stage, was located west of Lake Como (Gaetani et al., 1987; Garzanti and Pagni Frette, 1991). In other areas of the Southern Alps, similar tectonic pulses are also recorded in the upper Scythian strata (De Zanche and Farabegoli, 1981).

Application of sequence stratigraphic concepts is by no means obvious. Biostratigraphic control is generally poor, and only the major Early Spathian transgressive episode can be dated and traced all along the Southern Alps. This event cannot be correlated with any of the condensed sections marked on the Haq et al. (1988) chart and recorded on Tethyan margins as far as the northern edge of the Indian Plate (Nicora et al., 1992). Moreover, major petrographic changes at this stage, with marked decrease of quartz content and mineralogical stability, suggest the importance of tectonic control. Therefore, not even during the Scythian was the southern alpine sedimentary record primarily influenced by eustasy. The accommodation pattern, however, suggests that depositional sequences formed under a regime of not very wide amplitude in sea level changes.

The Servino Formation records overall relative sea-level rise, as indicated by progressive coastal onlap from east to west onto the Permian substrate. The first Griesbachian depositional sequence, represented in the more complete sections of the Dolomites by the basal Triassic transgression of the Tesero Oolite and by the overlying regressive Maizzi Member of the Werfen Formation, cannot yet be recognized in the Lombardy Prealps, where the position of the Permian/Triassic boundary is still uncertain. The second sequence, Late Griesbachian to Dienerian in age and including the Andraz and Siusi Members of the Werfen Formation in the Dolomites, is represented in the Camonica valley by the lower part of the Servino Formation. The latter consists of micaceous quartzose sandstones and sideritic dolostones (up to 10 m), overlain by greenish micaceous dolostones and lutites yielding the age-diagnostic bivalve Claraia clarai (30 to 50 m; Assereto, 1973; De Donatis et al., 1991).

The overlying transgression is documented east of the Seriana valley by reddish fossiliferous oo- and bio-sparites (up to 50 m), which probably correspond to the Gastropod Oolite. In the Western Orobic Alps, the Servino Formation, characterized by several high-frequency cyclothems, displays an overall fining-upward succession, which can be subdivided into two distinct sequences (S1 and S2). The sequence boundary is set at the top of a widespread red shale horizon, in correspondence with the sharp base of lithofeldspathic sandstones pos-
sibly representing estuarine fills of scours incised by river channels while sea level was low.

The same two transgressive regressive sequences, indicating renewed erosion and deposition of coarse detritus, can be recognized in the Lake Como area.

**Sequence S1**

The basal part of the Triassic succession E of Lake Como ("Prato Solaro Member" of the Servino Formation) consists of up to 50-m-thick, reddish conglomeratic sublitharenites, locally interbedded with medium to coarse-grained white subarkoses and deposited in fan-deltas. Only occasionally have Scythian bivalves (*Myophoria laevigata, Homomya canalensis*) been reported from marine layers interbedded with these siliciclastics (Merla, 1933), which are overlain by 15-m-thick, greenish to reddish dolomitic siltstones east of Lake Como. Farther to the east, the Prato Solaro Member thins out; it is not represented in the upper Brembana valley, where the onlap of the Servino Formation onto the Verrucano Lombardo documents an unconformity with significant time gap. This lower part of the Servino Formation consists of as much as 55 m of up to very coarse-grained quartz-rich subarkoses and subordinate lithic quartzarenites with felsic volcanic rock fragments, cyclically alternating with dolostones and lutites. These sediments, deposited in coastal settings, are ascribed to sequence S1, which is capped by thin reddish shale marker horizons from the Western Orobie Alps to the Austroalpine (Livigno area). The age of this sequence is uncertain; the recent finding of *Claraia* bivalves 8 m above the top of the Verrucano Lombardo in Vallassina points to a latest Griesbachian to early Dienerian age, suggesting correlation with the Dienerian Siusi Member of the Werfen Formation (R. Posenato, pers. comm., 1994) marking a regional increase of terrigenous input (Broglio Loriga et al., 1990).

**Sequence S2**

East of Lake Como, deltaic reddish conglomeratic arkoses and lithic arkoses (about 150 m thick) are overlain by about 100-m-thick, greenish-grey siltstones and yellow dolostones. In the Western Orobie Alps, a 2- to 3-m-thick, medium-grained pebbly lithic subarkose with herringbone cross-lamination and scoured base is followed by greenish siltstones and yellowish dolomitic lithic subarkoses (up to 15 m thick), overlain in turn by grey-greenish lutites and impure yellow dolostones (60 m or more), which were deposited in coastal sabkha to open-marine environments. This upper part of the Servino Formation, displaying similar lithology and thickness all the way to the Camonica valley (De Donatis et al., 1991), contains Early Scythian ammonoids (*Tirolitites cassianus, Dinarites* sp., M. Balini, pers. commun., 1993;) and Late Scythian foraminifers (*Meandrospira pusilla*) and bivalves (*Natria costata, Costatoria costata*, Casati and Gnaccolini, 1967). Sequence S2 can thus be correlated with the Spathian Val Badia and Cencenighe Members of the Werfen Formation (Broglio Loriga et al., 1990).

**The Middle Triassic Supersequence**

The Middle Triassic Series is characterized by thick carbonate successions, commonly exceeding 1200 m or even 1500 m in thickness E of Bergamo. In the west, where the Anisian section contains major gaps, thickness is between 500 and 1000 m. According to the Haq et al. (1988) scale, accumulation rates thus ranged between 80 and 110 m/my in the west and between 110 and 150 m/my in the east. Much higher figures (up to over 400 m/my), which are unlikely for a mostly carbonate succession deposited at times of still inefficient "carbonate factories,"
The succession can be subdivided into two major composite sequences (A1 and A2-L; Fig. 5), which roughly coincide with cycles UAA-2.1 and UAA-2.2 (Haq et al., 1988; supersequences T2 and T3 of Skjold et al., 1992).

Composite sequence A1 includes in the W an apron of wave-dominated fan deltas (Bellano Formation; Gaetani, 1982), passing eastward to prodelta deposits and shallow-water carbonate bays (Angolo Limestone, up to 500 m thick in eastern Lombardy). At its top, a minor sequence (A1/II) is characterized by small prograding conglomeratic fans fed by local uplifts in western Lombardy and by the growth of small carbonate banks in central Lombardy (“Peritidal Dolomites”, Camorelli and Dosso dei Morti Formations).

Composite sequence A2-L is a 700- to 1000-m-thick undifferentiated succession representing the bulk of the Middle Triassic strata in central Lombardy. Major regional transgression is documented in the lower part by progressive deepening in the basins (Prezzo and BUSCHENHEIM Formations), while “carbonate factories” started to build the huge Esino platform and equivalent Salvatore Dolomite of the Varese area. Drowning began earlier in the Giudicarie-Brescia area (end of Middle Anisian) and progressively propagated westwards, reaching the Varese area in Ladinian time (Meride Limestone). Explosive volcanic activity is indicated by pyroclastic rocks intercalated in the Prezzo Limestone, BUSCHENHEIM Formation, BEsano Formation, Meride Limestone, Perledo-Varenna Formation and Wengen Group (Jadoul and Rossi, 1982). The combined effect of eustatic fall and huge volcaniclastic supply filling most intraplatform troughs allowed quick progradation of carbonate platforms at Late Ladinian times. Uppermost Ladinian peritidal carbonates with tepees document local emersions and repeated dissolution episodes, before most of the Lombardy area finally emerged and was karstified (Assereto et al., 1977; Frisia et al., 1989). Three distinct sequences (A2-L1, L2 and L3) can be recognized in the Upper Anisian to Ladinian successions of the Lake Como area and of the Bergamasco Alps. In central Lombardy, the middle-upper Esino carbonates recorded transgressive-regressive cycles partly controlled by volcanotectonic activity (Wengen Formation), but disconformities were recognized only at the top and not within the Esino buildup.

The composite sequence A2-L contains three elements that may be recognized not only in the whole Southern Alps, but elsewhere in Europe. They are the transgressive tract starting in the latest part of the Middle Anisian, the MFS in the Nevadites Zone within the BUSCHENHEIM Formation, and the top unconformity in the latest Ladinian. Within the composite sequence, 3rd order sequences are locally difficult to recognize.

Sequence A1/II

The A1/II sequence boundary is marked in the Brembana valley by a disconformity with “terra rossa” infilling small fractures and by intertidal-supratidal facies locally with intraformational breccia pockets recognized in the basal “Peritidal Dolomites” (upper Angolo Limestone; Jadoul et al., 1992c). The “Peritidal Dolomites” show a transgressive-regressive trend; the lower part consists of black marls and subtidal carbonates, whereas the upper part is characterized by shallowing-upward cyclothems. At this time, the Camonica-Giudicarie area was characterized by initial growth and progradation of the small Camorelli-Dosso dei Morti platforms into the Angolo lagoons (Gaetani and Gorza, 1989).

In the Lake Como area, wave-dominated fan deltas prograded onto the carbonate bay as a result of increased siliciclastic supply (Bellano Formation). Basement uplift and either active volcanism or basin inversion is indicated by lithic arkoses passing upward to feldspathic litharenites with greater abundance of feldspars and volcanic rock fragments (the latter reach as much as 60% of the main framework components in the Lugano area) with respect to the underlying Servino Formation. The bulk of this minor sequence, up to 200 m thick, is Middle Anisian (Pelsonian) in age, as indicated by a rich foraminiferal assemblage characterized by *Meandrospira dinarica*.

Sequence A2-L1

The lower boundary in western Lombardy coincides with the sharp base of the upper tongue of the Bellano Formation, which prograded for 3 to 5 km onto the underlying coastal flat. In central Lombardy, it is documented by subaerial exposure of the upper “Peritidal Dolomites” and by a usually less than 5-m-thick interval with fenestrae and desiccation features in the topmost Angolo Limestone. To the east, the sequence boundary is sharply followed by increasing clay supply, suffocation of carbonate productivity and drowning of the Camorelli-Dosso dei Morti platforms. Foraminiferal (*M. dinarica*) and dasycladacean algal assemblages (*Physoporella pauciforata, Oligoporella pilosa*) indicate that this paraconformity is still within the Middle Anisian. The sequence boundary can be best located in the upper part of the Balatonicus Zone, and thus earlier than indicated in the Haq et al. (1988) scale (local unconformities do occur in the Trinodosus Zone, like the one at the base of the Richthofen Conglomerate in the Dolomites, but they appear to be chiefly controlled by local tectonic uplift).

The upper Bellano Formation lithic arkoses, representing a wave-dominated lowstand wedge, are 50 m thick at most. The transgressive surface is marked in the west by crinoid-bearing quartzose arenites or by coarse-grained, lithoclastic to bioclastic hybrid arenites yielding crinoids and brachiopods (Brachiopod Beds, 2 to 15 m thick) and to the E by the base of the Prezzo Limestone (base of the “Pararacerties” cimegana horizon; Middle/Late Anisian boundary; Balini, 1992). Transgressive
strata are also characterized by a diversified foraminiferal fauna (Paleomiliolina judicariensis and Pilammina densa).

The transgressive systems tract, represented by the Prezzo Limestone (up to 80 m thick), corresponds to the Paracerasites Zone s.1. and Lardaroceras beds (Illyrian; Late Anisian; Balini, 1992). Rapid drowning at this stage, widespread all the way to the Dolomites, does not coincide with a major global eustatic rise on the Haq et al. (1988) curve. A few meters above the base of the Prezzo Limestone, a sharp increase in clay supply may be ascribed either to soil reworking during coastal retreat or to local uplifts, indicated in the Dolomites by deposition of the Richthofen Conglomerate. Detrital micas occur in the lower and middle parts of the unit, whereas in the upper part the first tuffs reflect the onset of explosive volcanism.

The boundary between the Prezzo Limestone and the Buchenstein Formation, marked by decreasing clay supply, corresponds to the maximum flooding. The nodular cherty limestones in the lower part of the Buchenstein Formation (up to 50 m thick) can be dated at the Nevadites Zone (earliest Ladinian; Gae
tani, 1993). This transgressive event is represented in the Varese area by the dysaerobic/anaerobic, lagoonal, fossiliferous black shales and grey dolostones of the Besano Formation, up to 15 m thick, characterized by an abundance of organic matter and well preserved vertebrate fauna.

Locally dolomitized peritidal limestones (Esino Limestone and S. Giorgio Dolomite, nomenclature according to the Lugano Sheet, Geological Map of Switzerland, 1986; Fig. 6) indicate initial growth and progradation of carbonate build-ups onto the Buchenstein basin during the highstand stage. Where the upper boundary can be recognized, sequence thickness can be estimated as 100 to 150 m.

Sequence L2

The L2 sequence boundary, which in basinal areas is para-
conformable and probably occurs in the lower part of the Buch-
enstein Formation, is marked in the Brembana valley by re-
gressive peritidal facies in the lower Esino Limestone, with local development of tepees and calcrites. Because of wide-
spread shallow-water facies and consequently poor biostrati-
graphic control, the boundary cannot be easily traced to adja-
cent areas. Biocalcarenites rich in brachiopods, bivalves and ammonoids, found 80 to 100 m above the base of the Esino Limestone in the Brembana valley (Ghegna Lumachelle, 10 to 30 m thick), indicate an earliest Ladinian age (Chieseiceras chiesense horizon; Brack and Rieber, 1986; 1993). Transgres-
sion at the top of the 60- to 120-m-thick basal Esino Limestone is recorded by transition to subtidal oncodial limestones. In western Lombardy, the lower Esino and S. Giorgio platforms are followed by thin-beded black mudstones deposited within small and anoxic intraplatform basins (Varena Limestone, Meride Limestone). Drowning may be at least partly ascribed to volcano-tectonic activity. The base of the Varena Limestone contains a late Early Ladinian conodont assemblage (Gaetani et al., 1992). Mixed calcareous-volcaniclastic turbidites were deposited within intraplatform troughs in the Grigna Mountains and farther to the east (Wengen Group, up to 400 m thick; Gaetani, 1986). This sequence is 300 to 600 m thick in carbon-
ate platform successions.

Sequence L3

This uppermost Ladinian sequence can be more easily dis-
tinguished in the Grigna Mountains, where it is represented by the topmost Esino Limestone (at least 200 m thick; Fig. 7). Within intraplatform troughs towards the west (Lake Como to Varese area), the sequence is represented by dark organic-rich mudstones (topmost Perledo-Varenna Limestone; possibly top-
most Meride Limestone), overlain by up to 130-m-thick yel-
lowish marls and grey marly limestones (Lierna Formation; “Kalkschieferzone”). Towards the east it is tentatively equated to the “Calcere Rosso” of the Bergamas Alp (up to 50-m-
 thick; Fig. 5).

In the Esino platform of the Grigna Mountains, the basal sequence boundary is locally represented by a karstified sur-
f ace. Within anoxic intraplatform troughs, the boundary is marked by a megabreccia body occurring in the upper part of the Varenna Limestone (Archelaus Zone, Late Ladinian; Gae-
tani et al., 1992) and thus quite high into the Upper Ladinian strata. In the upper part of the sequence, the Esino platform prograded into the basins, which had been largely filled by dark grey, thinly laminated marly limestones (Perledo Member, up to 100 m thick). The Calcare Rosso, recognized only in the Brembana, Seriana and lower Scalve valleys, consists of cyclic peritidal carbonates with mature tepees rhythmically alternating with “terra rossa” palaeosols showing complex diagenetic mod-
ifications (Assereto and Kendall, 1977; Assereto et al., 1977). To the east and west, the sequence pinches out against the un-
derlying unconformity, which is a karstic surface marked by carbonate breccia pockets and by a network of cavities and fractures filled by yellow marly dolostone and green tuffaceous shale. Cavities commonly reach 20 m below the Esino platform top, but huge ones may reach down to 80 m (Jadoul and Frisia, 1988).

The Calcare Rosso, capped by a thick interval with strongly deformed tepees overlain by a reddish-greenish shaly marker horizon, displays a transgressive/regressive trend (Mutti, 1992). Its inferred age is Late Ladinian, as indicated by the occurrence of dasycladacean algae (Teutoporella echinata), benthic foraminifera and rare bryozoans. A tectonic event at the top of the unit is indicated by the presence of sedimentary dykes, tensional fractures and breccias with both intraformational and extrafor-
normal clasts derived from the Esino carbonates.

THE CARNIAN SUPERSEQUENCE

The Carnian succession of the western Southern Alps is ar-
 ranged in E-W trending belts (Brusca et al., 1982). Numerous recent papers have worked out its lithostratigraphy, sedimen-
tology, petrography and paleogeography in detail (Garzanti and Jadoul, 1985; Gnaccolini and Jadoul, 1990, and references cited therein). Even though age-diagnostic fossils are mostly lacking (there is only one ammonoid occurrence: Allasinaz, 1968a), a few complete and laterally continuous sections in the Lombardy Prealps allow detailed physical correlation of strata, as for ex-
 ample the Camonica valley transect. West of Lake Como, suc-
cessions are incomplete and mostly unfossiliferous, except for Carnian pollen found in the Varese area (Scheuring, 1978); the stratigraphic framework is therefore uncertain.

In central Lombardy, the 650- to over 1000-m-thick Carnian succession begins with peritidal carbonates containing the first
appearance of the dasycladacean alga Clypeina besici (Breno Formation and Calcare Metallifero Bergamasco; Assereto et al., 1977), overlain by thick lagoonal marly limestones yielding Early to middle Carnian age (Late Cordevolian to Julian; Allasinaz, 1966) bivalve assemblages (Gorno Formation). A biooo-calcarenitic tongue in the middle part of the Gorno Formation documents a major event of carbonate progradation (Gnaccolini and Jadoul, 1988).

In the NE (Camonica and Giudicarie valleys), platform carbonates characterized by shallowing-upward cyclothems were deposited through most of the Carnian age (Breno Formation; Fig. 8). Their lower part in the Presolana Group contains a rich ammonoid fauna of the Aonoides Zone, and is thus dated as Middle Carnian (Julian; Allasinaz, 1968a). The lagoonal Gorno Formation passes laterally southward to a thick wedge of deltaic clastics derived from nearby volcanoes (Val Sabbia Sandstone).
The Carnian succession is capped by coastal sabkha deposits (San Giovanni Bianco Formation), which in the Camonica valley directly overlie the Breno carbonate platform.

In the Brescia area (Sabbia valley), the Carnian is mostly represented by a thick redbed succession (nearly 1000 m; Val Sabbia Sandstone and lower San Giovanni Bianco Formation), with intercalations of grey mudrocks, calcilithite conglomerates and calcretes. Sandstone petrography documents several volcanic episodes with rapidly changing composition, from rhyodacitic at the base (sequence C1) to mainly latitic (sequence C2). Lava flows of olivine basalts are interbedded probably in the lower part of sequence C3, and are followed by slightly more quartzose sandstones richer in felsic volcanic rock fragments (upper part of sequence C3). The redbeds pass upward to about 200-m-thick yellowish dolomitic siltstones with gypsum lenses (sequence C4; middle and upper San Giovanni Bianco Formation).

In the Varese area, well-bedded laminated micritic dolostones deposited in peritidal to supratidal lagoonal environments (Cunardo Formation, 80 to 120 m thick) are overlain by reddish-greenish dolomitic mudrocks, capped by marls, micritic supratidal dolostones and local gypsum lenses (Pizzella Formation, 40 to 60 m thick; Allasinaz, 1968b).

The boundaries of the T/R cycle are easily recognized, while within the cycle, depositional sequences are not always obvious. The lack of age-diagnostic fossils often leaves a margin of uncertainty in recognition of the sequence boundary, which is done purely on the lithostratigraphic and paleoenvironmental interpretation.

**Sequence C1**

The sequence boundary is marked by a widespread unconformable transition from the “Calcare Rosso” or directly from the karstified top of the Esino platform, to bedded peritidal carbonates locally with basal breccias or thin tuffaceous shales (Breno Formation). Light grey limestones, locally displaying small tepees in the lower part and organized in 1- to 9-m-thick shallowing-upward cyclothems, are followed by mainly subtidal and finally by intertidal to supratidal carbonates. These sediments might represent the lowstand systems tract. The overlying dark grey thin-bedded carbonates, with diffuse black chert and arranged into thin shallowing-upward peritidal cyclothems capped by stromatolitic bindstones (Calcare Metalliferio Bergamasco), could represent the transgressive and highstand systems tracts. The unit is capped in the Brembana valley and Grigne Mountains by an emergence surface containing paleokarst. Active explosive volcanism is testified by rhyodacitic tuffs commonly intercalated within the Breno Formation, and by volcanic arenites rich in felsitic volcanic rock fragments characterizing the proximal southern sections (Garzanti, 1985a). Sequence thickness ranges up to 190 m.

**Sequence C2**

In central Lombardy, the basal sequence boundary is marked either by emergence or by a sudden transition to dark grey lagoonal marls and mudstones, indicating inundation of carbonate tidal flats and subsequent drowning during a relative sea-level rise. These open lagoon to shallow bay deposits, rich in bottom-dwelling bivalves and organized in shallowing-upward marl-limestone cyclothems 1 to 7.5 m thick (lower Gorno Formation), interfinger with the Breno carbonate platform to the northeast (Camonica valley) and with the deltaic Val Sabbia sandstones to the southwest. The subaerial delta consists of reddish sandstones and siltstones organized in 1- to 23-m-thick fining-upward cyclothems; the subaqueous part consists instead of greenish sandstones and siltstones, cyclically alternating with dark grey lagoonal deposits. Active “latitic”-type volcanism supplied enough detritus to completely fill the accommodation space created by transgression, forming a triangular wedge rapidly thickening towards the volcanic arc. Strong differential subsidence, enhanced by the weight of rapidly accumulating sediments (accumulation rates reached 200 m/my in the southernmost sections), is attributed either to faulting or to flexural bending of the crust underneath the volcanic load. Grain size
shows an overall increase throughout the section (Garzanti, 1986; 1988), but the Val Sabbia fluvial-dominated delta displayed only slight progradation and remained confined to the south (Gnaccolini, 1983). The upper part of the Val Sabbia Sandstone shows rapid backstepping due to both an increasing rate of relative sea-level rise and decreasing rates of denudation and sediment supply. Maximum flooding corresponds with the upward transition to lagoonal sediments in proximal areas, marked at places by condensed arenites bearing ferruginous ooids (Garzanti and Pagni Frette, 1991). In the Camonica valley, this event is represented by a dark grey lagoonal marly limestone tongue intercalated within the Bremo peritidal carbonates.

The highstand systems tract is represented in central Lombardy by lagoonal marl-limestone cycloths, capped by foraminiferal wackestones-packstones characterized by diffuse chert at the base and by local ooid grainstones and green algal patch reefs (middle Gorno Formation between the Brembana and Camonica valleys; 20 to 85 m thick). Sequence thickness reaches 550 m in the Brembana valley.

**Sequence C3**

Deposition of open-lagoon to shallow bay marl-limestone cycloths, each 2 to 8.5 m thick, was suddenly renewed at the base of sequence C3 in central Lombardy (upper Gorno Formation). These deposits interfinger with shallow-upward peritidal cycloths, each 1 to 4.5 m thick, in the Camonica valley, where the basal boundary is not clearly recognizable because carbonate platform sediments accumulated through sequences C1 to C3 (Breno Formation).

In the regressive upper part of sequence C3 (highstand systems tract), peritidal dolostones in the NE (Campolungo Tongue at the top of the Breno Formation) and alluvial to sabkha sediments in the SW (lower San Giovanni Bianco Formation) prograded onto the Gorno lagoonal deposits. Volcanism continued in the middle part of the Carnian, with alternating episodes of mafic and felsic activity. Detritus in the upper Gorno Formation and overlying San Giovanni Bianco Formation, with abundance of pumiceous fragments, mainly testifies to rhyodacitic volcanism. In the proximal sections exposed in the Brescia Prealps, however, olivine basalts are overlain by upper delta plain sandstones and conglomerates (upper Val Sabbia Sandstone, which can be assigned to the upper part of sequence C2 or to the lower part of sequence C3). Next, renewed progradation of deltaic sandstones and calcilithic conglomerates points to a stage of tectonic uplift in late Carnian time. A distinct increase in quartz, representing a regional petrographic marker found across Lombardy (Crepaldi, 1990), testifies to deepening of erosion into the metamorphic wallrocks of the volcanic belt (Garzanti, 1985b). Relationships, if any, between this petrographic change and fluctuations of relative sea-level have not been found to date. Maximum sequence thickness is 250 m.

**Sequence C4**

The C4 sequence boundary is marked in central Lombardy by transition from an interval of reddish siltstones and sandstones (top of the lower San Giovanni Bianco Formation) to greenish siltstones and sandstones intercalated with vuggy dolostones (middle San Giovanni Bianco Formation). These sediments were deposited in coastal sabkhas influenced by both fluvial and marine processes and represent the lowstand systems tract. The occurrence of bioclastic layers with crinoids, hydrozoans, brachiopods, corals and locally nautiloids (“Nautilus brenbanus, identification by M. Balini) mark the transgressive systems tract.

In the upper San Giovanni Bianco Formation, thick gypsum lenses and yellowish dolostones interbedded with greenish/reddish dolomitic siltstones are overlain by grey, locally ooidal, amalgamated dolomitic limestones (uppermost S. Giovanni Bianco Formation, probably representing the highstand systems tract). Almost pure rhyolitic detritus may indicate either a terminal phase of explosive volcanism or erosion of older felsic volcanic products. The sequence reaches a thickness of 250 m.

**The Late Carnian-Middle Norian Supersequence**

This transgressive/regressive cycle is represented by a mostly shallow-water carbonate succession, usually at least 1700 m thick. Rapid subsidence results from renewed tectonic extension and thermal relaxation after the Carnian tectonic pulse (Garzanti and Jadoul, 1985). Widespread deposition of thick, restricted shallow lagoonal, cyclic peritidal to subtidal carbonates (Dolomia Principale) was followed by tectonically controlled deepening of anoxic intraplatform troughs (Dolomie Zonate and Zorzino Limestone; Aralalta Group of Jadoul, 1985; Fig. 9). Due to limited biostratigraphic evidence (conodonts and ammonoids are absent), correlations are mainly based on lithostratigraphy. Radiometric ages for the Norian stage have large uncertainties, but the large timespan, represented by only two depositional sequences within the Norian is striking.

**Sequence N1**

The N1 lowstand systems tract begins with vuggy limestone breccias (commonly overprinted by alpine tectonics) containing mud clasts of the underlying San Giovanni Bianco Formation. Intraformational breccias and poorly fossiliferous mudstones to packstones were deposited in marginal marine environments with temporary emergence (Castro Formation, 50 to 250 m thick), at a stage of renewed tectonic activity and climatic change (Jadoul et al., 1992b). Next, dark grey, well-bedded, laminated, mainly restricted shallow subtidal dark grey dolostones (50–250 m thick), locally with intraformational carbonate breccias, characterize the base of the Dolomia Principale (possibly representing the upper part of the lowstand systems tract). The overlying grey, thick-bedded dolostones representing a transition from peritidal to subtidal platform environments (500 to 900 m thick) point to transgression in the lower part, overlain by monotonous peritidal dolomites representing the aggradational highstand systems tract. The upper part of the Dolomia Principale is indicative of marginal platform environments and highly subsident intraplatform basins, rapidly filled by carbonate turbidites (Aralalta Group, 200 to 1000 m thick). Breccia pockets, calcere and small tepees may occur on structural highs at the top of the Dolomia Principale, pointing to local emergence and hiatuses (Jadoul and De Bonis, 1981). This very thick sequence (up to 2000 m) was deposited between latest Carnian (Conti, 1954; also suggested by occurrence of the dasycladacean alga Clypeina besici in the Brembana valley) and Middle Norian time (the upper Zorzino Limestone fish fauna is partly correlative with the Seefeld fish fauna, where
possibly Middle Norian conodonts have been found; A. Tintori, pers. commun. 1992; Wild, 1989).

THE LATE NORIAN-HETTANGIAN SUPERSEQUENCE

This complex supersequence, 700 to 2000 m thick, is represented by black shales, marls and shallow-water carbonates. At least three depositional sequences can be recognized. The first one, onlapping the Dolomia Principale and organized in well-developed shale-marl-limestone 4th-order sequences, testifies to a tectonic pulse and sudden clay supply (Riva di Solto Shale, lower and middle Zu Limestone). The second one displays progressive progradation of carbonate ramps and subtidal banks (upper Zu Limestone, Conchodon Dolomite). The third sequence is characterized by open-marine fauna and oolitic bars (Sedrina Limestone), until final drowning occurred in late Hettangian times. More sequence subdivisions could be recognized within this cycle, but the absence of precise biostratigraphic tools prevents separation of a somewhat repetitive lithostratigraphic pattern.

Sequence N2-R1

A sharp change from carbonate platform or intraplatform basin strata to dark shales, seen in all outcrops, marks the basal sequence boundary (Fig. 10). Platform crisis due to both emergence and clay suffocation is indicated by rapid disappearance of carbonate supply within the troughs (lower Riva di Solto Shale, lower and middle Zu Limestone). The second one displays progressive progradation of carbonate ramps and subtidal banks (upper Zu Limestone, Conchodon Dolomite). The third sequence is characterized by open-marine fauna and oolitic bars (Sedrina Limestone), until final drowning occurred in late Hettangian times. More sequence subdivisions could be recognized within this cycle, but the absence of precise biostratigraphic tools prevents separation of a somewhat repetitive lithostratigraphic pattern.

Black shales and marls containing common slumps and intraformational paraconglomerates were deposited on gentle slopes and in shallow basins with largely anoxic conditions at the bottom (0- to 250-m-thick lower Riva di Solto Shale, which represen the lowstand systems tract). The overlying shales, marls and marly to micritic limestones, with plenty of thin bioclastic storm layers, are organized in 4th-order sequences (upper Riva di Solto Shale and lower Zu Limestone, 300 to 1500 m thick; Masetti et al., 1989). In the Imagna valley, 5- to 30-m-thick sequences may be capped by thin limonitic hardgrounds. In the same area, thin stromatolitic bindstones and vuggy evaporitic carbonates or oolitic limestones occur at the transition between the upper Riva di Solto Shale and the lower Zu Limestone (Fig. 10). This first thick regressive carbonate interval may represent the upper part of a distinct minor sequence (Late Norian cycle of Jadoul and Gnaccolini, 1992). A Late Norian or, locally, even a late Middle Norian age for the Riva di Solto Shale is suggested by several Norian poly morph associations; Rhaetian faunas and floras occur only in the middle-upper Zu Limestone (S. Cirilli, pers. commun., 1993). In the Varese area, the Campo dei Fiori Dolomite (70 to 80 m thick) consists of dolostones and dolomitic limestones, locally yielding ooids, corals and Conchodon.

The top of sequence N2-R1 on structural highs is characterized by shallow-upward sequences of bioclastic to ooidal packstones-grainstones (middle Zu Limestone, 30–80 m thick), capped by stromatolitic dolostones and local breccia pockets. Bioturbated mudstones locally contain patch-reef boundstones capped by peritidal stromatolitic bindstones and ooidal grainstones. Marly intercalations still occur in subsiding areas, where 4th-order sequences are thicker and locally display thickening-and shallowing-upward trends. These carbonates, documenting an overall regression with episodes of carbonate ramp progradation (first “coral-ooloidal horizon” in the middle Zu Limestone), contain the rare Austrihydrichnia brachiopod of Rhaetian age (Krystyn, 1990) and other “Rhaetian” fossil assemblages (Triasina hantkeni, megalodontids, Aulotortidae, corals, algae; Lake, 1990). The Norian-Rhaetian boundary thus occurs at the top of the lower Zu Limestone, as indicated by the appearance of pleuromiid bivalves about 10 m below the first “coral-
Fig. 10.—Stratigraphic framework for sequences recognized in the Norian-Hettangian succession in the Taleggio and Imagna valleys (western Bergamasco Alps). (1) Peritidal to platform margin dolostones (2) dark intraplatform turbiditic carbonates (3) black shale-marls with slumps and intraformational paraconglomerates (4) dark subtidal bay marl-limestone cyclothems (5) subtidal ramp carbonates and (6) chert-bearing open shelf mudstones.

**oooidal horizon**. About 40 to 60 parasequences are recognized within this 400 to 1500 m thick 3rd-order sequence, which represents Late Norian to early Rhaetian time.

**Sequence R2**

In the Lombardy Prealps, the R2 sequence is documented by shallow-water carbonates (upper Zu Limestone and *Conchodon* Dolomite, 180 to 220 m thick). In the Varese area, where it may be missing altogether on structural highs, it is represented only by the *Conchodon* Dolomite (up to 100 m thick). The latter unit is underlain by a locally erosional surface marked by breccia pockets and overlain by another erosional surface with “terra rossa” palaeosols (Gnaccolini, 1964).

The basal boundary, sharp on structural highs (Mt. Albenza-Brumano), is transitional and difficult to recognize in deeper-water areas of greater subsidence. Thin-bedded, bioturbated or laminated, poorly fossiliferous mudstones, alternating with marls and unfossiliferous dark shales in 3- to 8-m-thick 4th-order sequences, are intercalated with storm-deposited bivalve-rich calcarenites. In the Imagna-Taleggio valleys, where at least two horizons of evaporitic carbonates were observed, thin breccias-paraconglomerates may occur at the base of 4th-order sequences, which are commonly capped by thin limonitic hardened surfaces. This succession, 60 to 100 m thick, may represent the lowstand systems tract.

The overlying open-marine bioclastic to ooidic packstones-grainstones are intercalated with coral and sponge patch reefs or with oncotic and megalodon-bearing wackestones-packstones (second “coral-oolitic horizon” close to the top of the upper Zu Limestone). This mainly carbonate platform succession, arranged in 10 to 20 m thick shallowing- and thickening-upward 4th-order sequences, represent the transgressive systems tract. Dark thin-bedded mudstones and fine-grained bioclastic wackestones/packstones locally with chert nodules (topmost Zu Limestone, 12 to 25 m thick) were deposited during maximum flooding. The overlying prograding cross-laminated oolitic bars, locally selectively dolomitized and intercalated with subtidal mudstones (*Conchodon* Dolomite, 80 to 200 m thick in the Lombardy Prealps), represent the highstand systems tract (Lakew, 1990).

**Sequence H**

The lower part of sequence H is made of shallow-water fossiliferous limestones (“Grenzbivalven Bank”), locally interlaminating with thin breccia-paraconglomerate lenses, overlain by well-bedded, nodular and amalgamated, transgressive open-shelf chert-bearing limestones (lower Sedrina Limestone, 40 to 70 m thick). A middle Hettangian age (Liasicus Zone) is indicated both by nannoflora in the Grenzbivalven Bank (first occurrence of *Crepidolithus crassus*; Lozar, 1993) and by ammonites at the base of the Sedrina Limestone (Bistram, 1903). The latter also contains Hettangian bivalves (Gaetani, 1970). The Triassic-Jurassic boundary thus seemingly lies within the *Conchodon* Dolomite.

The upper part of the sequence consists of progradational bioclastic (crinoids, brachiopods) and oolitic grainstones with white chert (upper Sedrina Limestone, 5 to 20 m thick). In central and western Lombardy the oolitic Sedrina platform was finally drowned during the late Hettangian Angulata Zone (Gaetani, 1970) and sharply overlain by thin-bedded pelagic limestones rich in black chert, radiolaria and sponge spicules (lower Moltrasio Formation). In southeastern Lombardy, a thick carbonate platform continued to grow during Liassic time (Corna; Fig. 5). Local tepee horizons in its lower part may be interpreted as marking the basal boundary of sequence H.
DISCUSSION

Geodynamic Scenarios

In the Triassic Series of the Southern Alps, changes in the style of tectonic activity allow separation of three successive paleogeodynamic scenarios.

In the Early Triassic time (supersequence S), a shallow sea encroached stepwise from the southeast onto a fairly flat piedmont, marking the close of an extensional-transensional episode initiated in the Early Permian. The Late Permian northeast-southwest structural lineaments were maintained during this first stage, characterized by relatively low tectonic activity and subsidence rates.

Lineaments gradually rotated in Anisian time, and were directed east-west through the Ladinian and Carnian time, when volcanism reached its climax (supersequences A-L and C; Brusca et al., 1982). Subsidence peaked in the Late Anisian to Ladinian, and was gradually reduced in the Carnian age, when it remained great only near to the southern volcanic belt (Table 2).

At the close of the Carnian time, structural trends turned again northeast-southwest, with minor east-west conjugate elements (Castellarin and Vai, 1982). Subsidence greatly increased in Norian time and remained strong in the Early Jurassic Epoch (supersequences N1 and N2-H).

These three paleogeodynamic stages reflect the progressive opening of the Neotethys. The first two steps are thought as directly connected with its successive attempts of westward propagation, the second one resulting in huge volcanic activity along east-west trends controlled by transtension. During the Late Triassic and Jurassic, evolution of Neotethys interacted with a new major paleogeographic event: rifting began in the Central Atlantic (Fig. 11; Manspeizer, 1988) and all along the future passive margin of Adria (Eberli, 1988). This stage marked the transition from aborted westward propagation of the Neotethys to a new trend, conditioned more and more by the incipient opening of the Atlantic, eventually leading in Middle Jurassic time to spreading of the Ligurian-Piedmont ocean floors.

Tectonism Versus Eustasy: An Everlasting Debate

Accommodation space and supersequence architecture are controlled by the interplay of tectonic subsidence with changes of sea level. Since the stratigraphic record seldom provides sufficient clues to distinguish between eustatic and tectonic control and eustasy is dependent on rates of oceanic spreading and global tectonic activity, only broad inferences suggested from long-term study of southern alpine rocks will be made herein.

Using the available biostratigraphic resolution, most Triassic supersequences appear correlatable worldwide. In the Barents Sea, Skjold et al. (1992) recognized five supersequences, three of which coincide with ours; even the lower boundary of their T3 supersequence is a significant feature of the southern alpine record. In the Canadian Artic, Embry (1988, 1993) also distinguished five transgressive/regressive sequences, broadly cor-
responding to the Early Triassic, Middle Triassic, Carnian, Norian and Rhaetian Epochs and times. The basal boundaries of supersequences A-L and N1, however, distinctly coincide with a changing style and intensity of tectonic activity, and significant tectonic control is documented also for the base of supersequences S, C and N2-H. Moreover, field observations show that details of sequence patterns are strongly influenced by local tectonism, peaks of which only incidentally happen to be synchronous in different sedimentary basins. Since this regional activity is related to global reorganizations of plate motion and long-term eustatic changes are largely the expression of the very same plate tectonic activity, we are not amazed that a comparable architecture of transgressive/regressive sequences, documenting major tectono-eustatic cycles, results in sedimentary basins across the world. Local tectonic events can and generally do produce peculiar signatures to each succession, superposed on these common long-term trends.

CONCLUSIONS

Multiorder cyclicity, from megasequences to high-frequency Milankovitch-like cyclothem, is recognized in the sedimentary record of the Southern Alps. The Triassic succession can be subdivided into five transgressive/regressive tectono-eustatic supersequences.

The Lower Triassic supersequence consists of cyclically interbedded quartzose siliciclastics, mudrocks and dolostones, onlapping the underlying Permian continental redbeds progressively from the southeast during a stage of relative tectonic quiescence. Basal Scythian marine sediments are represented only in the Dolomites. Transtension and increasing tectonic subsidence around the Smithian/Spathian boundary, probably induced by Neothyrian spreading in the east, led to open-marine conditions in the Spathian age.

Tectonic activity continued in the Middle Triassic Epoch, when a siliciclastic wedge prograding from Lake Como to the east documented uplift and either renewed felsic volcanism or basin inversion. Terrigenous sedimentation was instead reduced in eastern Lombardy, where increasing carbonate productivity kept pace with subsidence, thus forming the first carbonate buildups after the global P/T crisis. Explosive ryhoacitic activity characterized latest Anisian and earliest Ladinian time, when drowned areas became the sites of deep-water, starved carbonate sedimentation. During the Ladinian, intraplatform troughs were gradually filled by both carbonate and volcanioclastic turbidites.

Continuing volcanic activity with rapidly alternating felsic and mafic episodes is recorded in the Carnian supersequence, which consists of deltaic volcanioclastics passing northward to lagoonal marly limestones and peritidal platform carbonates, capped by deltaic to coastal sabkha deposits. Slightly but consistently higher quartz content in the Upper Carnian San Giovanni Bianco Formation compared with the underlying Val Sabbia clastic wedge indicates deeper erosion levels reached within Hercynian basement rocks.

The Lower to Middle Norian supersequence was characterized by widespread transgression of the peritidal Dolomia Principale, directly onlapping basement areas in the south. During mid-Norian time, platform dissection documented tectonic extension, which began to affect the future passive margin of Adria. Great amounts of clay, derived from the European continent, and huge carbonate productivity kept pace with strong subsidence during deposition of the Upper Norian-Hettangian supersequence, until the study area was finally drowned and covered by pelagic sediments towards the end of late Hettangian time.

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TRIASSIC SEQUENCE STRATIGRAPHY IN THE SOUTHERN ALPS (NORTHERN ITALY): DEFINITION OF SEQUENCES AND BASIN EVOLUTION

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ABSTRACT: A number of 3rd-order depositional sequences have been observed in the Southern Alps, far more than previously known: of these, 6 develop principally in the Scythian, 4 in the Anisian, 3 in the Ladinian, 4 in the Carnian, at least 2 in the Norian and finally 2 in the Rhaetian. Lower Anisian to Lower Carnian depositional sequences are best dated by ammonoids, while in the Scythian ammonoids and conodonts are scanty. In the Upper Carnian, Norian and Rhaetian deposits of the Southern Alps, very rare ammonoids and conodonts are available. Therefore Scythian, Upper Carnian, Norian and Rhaetian depositional sequences have been deﬁned on the basis of stratal patterns and the evolution of facies, being their chronostratiﬁgraphical position inferred from sections elsewhere in the world. The contemporaneous analysis of platforms and basins and of carbonate, mixed and siliciclastic deposits has made a good comprehension of facies migration possible. This approach was crucial in the deﬁnition of 3rd-order depositional sequences. Moreover, on the basis of the previously deﬁned 3rd-order sequences and systems tracts, a number of 2nd-order transgressive/regressive cycles have been pointed out. The sequence stratigraphic analysis compared with the tectonic history allowed the deﬁnition of the different phases of the basin evolution.

INTRODUCTION

Three recent attempts to attain a stratigraphic synthesis of the whole Triassic interval in the Southern Alps (Fig. 1) are to be mentioned: the ﬁrst by Assereto and Casati (1965), the second by Assereto (1973), the last by Pisa (1974). Contributions regarding large parts of the Southern Alps have been provided (e.g. the Triassic in Lombardy, Jadoul and Rossi, 1982; Jadoul and Gnaccolini, 1992 and the Triassic in the Dolomites, Leonardi, 1968; Assereto et al., 1977a; De Zanche, 1990; De Zanche et al., 1993).

It is clear that a major problem in these schemes was the lack of a well-deﬁned, continuous and controlled biochronostratiﬁgraphic reference scale spanning the entire Triassic interval, as has always been a major problem of Triassic succession in Italy.

The ﬁnding of a large quantity of ammonoids and the availability of the highly resolved ammonoid standard scale for the Bithynian-Early Carnian interval by Mietto and Manfrin (1995), integrated for the Carnian by the scale in Krystyn (1978), allowed us to draw time lines throughout the Southern Alps within the Anisian-Carnian interval. The ammonoid standard scale by Mietto and Manfrin (1995) is much more resolved than the scales in use (Fig. 2). When ammonoids are absent, biochronological data have been provided by conodonts.

A new synthesis of Triassic strata in the Southern Alps is proposed, made possible by an approach integrating lithostratigraphy, biostratigraphy and sequence stratigraphy. The Triassic sequence stratigraphic interpretation in the Southern Alps is mostly the result of original ﬁeld research by the writers (De Zanche et al., 1992, 1993; Gianolla, 1992, 1993), integrated

Fig. 1.—Sketch map of the Southern Alps. Arabic and roman numbers indicates successions in selected areas illustrated in Figures 4 and 5.

In this paper, two sequence chronostratigraphic schemes (Figs. 4, 5) are presented: one corresponding to a west-east section, the other to an approximately south-north section across the Southern Alps. Due to the lack of good radiometric ages, we used the subzones of the new ammonoid scale as units of measure. Therefore our “time scale” is arbitrary in so far as no relation between subzones and time is available. The result is a number of columns, corresponding to selected areas of the Southern Alps (Fig. 1), showing vertical and lateral variations, unconformities, hiatuses and platform-basin relationships. Correlations are emphasized by sequence boundaries, tops of lowstand prograding complexes and maximum flooding surfaces.

GEOLOGICAL SETTING

The Southern Alps are a part of the Alpine Chain, bounded on the north by the Periadriatic Lineament and on the south by the Po and Venetian Plains. The former separates the northern Alps s.s. from the south-vergent Southern Alps. On the whole, the Southern Alps can be divided into western and eastern Southern Alps by the Giudicarie Line, a north-northeast to south-southwest Neogene structural system. The deformation history of the South Alpine Chain is very complex; after Permo-Triassic rifting phases and the Jurassic extension (Winterer and Bosellini, 1981; Doglioni, 1987), compressional tectonics began in the Bergamasc Pre-Alps (Bersezio and Fornaciari, 1989) in the Late Cretaceous and extended into Tertiary time (Castellarin et al., 1992). Shortening was strong to the west (Lombardy) and to the east (Carnia, Julian Alps), whereas in the intermediate part (Dolomites, Verona-Vicenza Pre-Alps) it was weak.

During Triassic time, extensional conditions existed in the Southern Alps probably due to strike-slip tectonics, which is responsible for local compressional features (Doglioni, 1987). Therefore, the Triassic history of the Southern Alps was dominated by a succession of blocks with differential subsidence, controlled by roughly east-west and north-south structural trends. The subsidence history is documented by different thicknesses of the same lithostratigraphic unit within the Triassic time.

Fig. 3.—A scheme of the Scythian 3rd-order depositional sequences in the Dolomites in comparison with the cycle chart in Haq et al. (1987).
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assic sedimentary cover. The main source area of siliciclastics was to the South, as demonstrated by the supply of metamorphic clasts from the basement (cf. Assereto et al., 1977a; Viel, 1979; Brusca et al., 1982; De Zanche and Mietto, 1984).

During Anisian, Ladinian and Carnian time, a number of volcanic phases took place, both basic, intermediate and acidic in composition. Volcano-tectonic compressional structures have been recognized around the Predazzo-Monzoni (western Dolomites) intrusive complex.

LITHOSTRATIGRAPHIC NOMENCLATURE

In the Southern Alps, Triassic deposits consist of continental, lagoonal, carbonate platform and basinal units variously stacked and interfingered. It is characterized by a great variety of facies created by marked paleogeographic differences. Actual knowledge reflects the history of the different approaches proposed by authors and schools since the 19th century, which have resulted in many names and a very complicated lithostratigraphic nomenclature. Figures 4 and 5 show clearly that each part of the Southern Alps has its own terminology.

It is clear that the profound revision in lithostratigraphy, which was necessary for the sequence stratigraphic approach, caused addition, cancellation, change, enlargement or restriction of some names. However, the writers have tried to maintain the original names in each area, with the exception of the cases in which confusion might have been created. In the proposed sequence stratigraphic setting, supported by a highly resolved ammonoid scale and a great number of ammonoids, problems are minimized because lithostratigraphic correspondences can be seen clearly.

On the whole the Triassic succession in the Southern Alps can be schematized as follows.

In the eastern and western Southern Alps, the Werfen (Servino) Formation, Scythian in age, paraconformably overlies Permian units (Assereto et al., 1973). In western Lombardy and in the Lugano area, Lower Triassic units have mostly been eroded; as a consequence, the Permian substratum is directly overlain by Anisian units (De Zanche and Farabegoli, 1988).

A peritidal carbonate platform overlies the Werfen (Servino) Formation throughout the Southern Alps, except in the areas where it was eroded during Anisian time. It is known as Carniola di Bovegno in Lombardy, as Lower Serla Dolomite in the Dolomites and as Lusinizza Formation in the easternmost Dolomites and in Carnia. De Zanche and Farabegoli (1982) suggested simplifying the nomenclature in the Southern Alps by using the terms Werfen Formation and Lower Serla Dolomite. The Scythian-Anisian boundary seems to lie within the latter (Broglio Loriga et al., 1990).

The Lower Serla Dolomite is overlain by a number of Anisian terrigenous and terrigenous-carbonate units deposited in basinal, lagoonal, peritidal and continental environments. In the past, these formations were considered, on the whole, to form the Braies Group, lying between the Lower Serla Dolomite and the Livinallongo Formation and recognizable throughout the Southern Alps.

A number of Anisian carbonate platforms have been identified in the past. In Lombardy, these comprise the Camorelli Limestone and the Dosso dei Morti Limestone of middle-late Pelsonian age. In the eastern Southern Alps, two carbonate platforms are commonly differentiated: the Upper Serla Formation and the Contrin Formation (cf. De Zanche et al., 1992; Senowbari-Daryan et al., 1993). The former is late Pelsonian, the latter is Illyrian in age. Ammonoid control and sequence stratigraphic methodology allowed us to recognize two other Anisian carbonate platforms, previously confused with the Upper Serla Formation. They crop out in the northern and eastern Dolomites and in Carnia and have not yet been formally defined: one is a prograding Bithynian platform, the other corresponds to a late Bithynian-early Pelsonian backstepping carbonate body (De Zanche et al., 1993).

Viel (1979) revised the Ladinian basinal units in the Dolomites and divided them into two large units: the Buchenstein Group and the Wengen Group; the former deposited during a fundamentally transgressive phase, which caused the maximum extent of the Triassic basins, the latter characterized by important basic magmatic activity and by a regressive trend. Such a division in these two large units is no longer justifiable because the Late Ladinian regressive trend, which was thought to have characterized the Wengen Group, really began earlier. The correspondence of the Ladinian basinal units in Lombardy and in the Dolomites is shown in Figure 4.

Huge carbonate buildups, bordering the Ladinian basins, have been identified (Bosellini and Rossi, 1974; Bosellini, 1984). Their name is different in the various sectors of the Southern Alps: San Salvatore Dolomite in the Varese-Lugano area, Esino Limestone in Lombardy, Mt. Spitz Limestone in the Recoaro area, Sindech Dolomite in Valsugana and Sciliar (Schlern) Dolomite in the eastern Southern Alps. The sequence stratigraphic approach allowed us to divide the San Salvatore Dolomite, the Esino Limestone and the Sciliar Dolomite into three parts, each one ordered and marked by numbers 1, 2 and 3. This result was inferred by integrated work both on carbonate platforms and in basinal successions, in whose proximal deposits the history of the platform areas is recognizable. However, karst and erosional surfaces on the top and within “undifferentiated” buildups are known and permit one to identify different bodies. In this scheme, each one really corresponds to an independent carbonate platform belonging to different sedimentary events.

Throughout the Southern Alps in the lower part of the Ladinian buildup complex, a retrograding (give-up and/or catch-up) carbonate platform has been recognized. It partly corresponds to the “Lower Edifice” pointed out by Gaetani et al. (1981) in the Latemar massif and is thought to be of latest Anisian to Early Ladinian age. The Mt. Spitz Limestone in the Recoaro area should be divided into three bodies: one eroded and karstified at its top, corresponding to the Contrin Formation in the Dolomites, the second corresponding to the Lower Edifice and the third to the Sciliar Dolomite 1-Esino Limestone 1 carbonate platforms. The Sindech Dolomite in the Valsugana area corresponds to the Sciliar Dolomite 1.

Basin infilling began in Early Ladinian and continued during latest Ladinian-Early Carnian time due to a strong carbonate platform progradation (Cassian Dolomite 1 and 2, Breno Formation) and an intense siliciclastic supply in basinal areas. In Lombardy during Carnian time, huge deltas, fed by southern volcanioclastics (Val Sabbia Sandstones, in Garzanti, 1985), prograded northward.
FIG. 4.—Triassic sequence chronostratigraphic west-east section across the Southern Alps. Columns schematically illustrate the succession in selected areas (see Fig. 1). Ammonoid subzones, arbitrarily considered as having the same duration, are used as time units of measure.
TRIASSIC SEQUENCE STRATIGRAPHY IN THE SOUTHERN ALPS (NORTHERN ITALY)
Fig. 5.—Triassic sequence chronostatigraphic south-north section across the Southern Alps. Columns schematically illustrate the succession in selected areas (see Fig. 1). Ammonoid subzones, arbitrarily considered as having the same duration, are used as time units of measure.
<table>
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<th>STRATIGRAPHY IN THE SOUTHERN ALPS (NORTHERN ITALY)</th>
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- **TRIASSIC SEQUENCE STRATIGRAPHY IN THE SOUTHERN ALPS (NORTHERN ITALY)**

- **Acquatona FM.**

- **Bivera FM.**

- **Unnamed sbz. 2**

- **Unnamed sbz. 1**

- **Italicus Plinii Subbullatus**

- **“Oedipus” Triadicum Aonoides**

- **Daxatina cf. canadensis**

- **Regoeclusus**

- **Neumayrii Longobardicum**

- **Gredleri Margeritosum**

- **Recubariensis Curionii Avisianum**

- **Reitzi Trinodosus Abichi**

- **Binodosus Balatonicus Cucanense**

- **unnamed abz. 2 Ismidicus**

- **unnamed abz. 1 Osmani**

- **Paracrochordiceras**

- **Lithologies**

- **Note:** The diagram includes various stratigraphic units and fossils, with specific localities and time periods indicated.
Finally during Tuvalian time, the complicated paleogeographic setting was made uniform by mainly peritidal carbonate-terrigenous sediments (Figs. 4, 5). All over the Southern Alps, this Carnian peritidal complex is overlain by a set of mainly continental units that Pisa et al. (1980b, p. 999) interpreted as the “reconquest of the Alpine domain by german facies”. This event provided the morphologic conditions suitable for carbonate tidal flats of the Dolomia Principale (Bosellini and Hardie, 1988), extending throughout the Southern Alps. It is generally accepted that the base of the Dolomia Principale is Tuvalian in age. However, the age of its top is a problem: in Lombardy, it lies within the Norian, as the overlying Riva di Solto Shales still seem to be Norian in age (Gaiani et al., 1992b); in the eastern Southern Alps, the Dolomia Principale seems to cover all of Norian time. In the Recčaro area, on the basis of brachiopods, it seems to reach even the Rhaetian/Liasic boundary (Mietto, 1977).

The Norian lithostratigraphic setting is better known in Lombardy (Jadoul and Rossi, 1982; Jadoul, 1986; Stefani, 1989; Jadoul et al., 1992a) than in the eastern Southern Alps. During the Norian interval and contemporaneous to the deposition of the Dolomia Principale, there was synsedimentary extensional tectonic activity. Anoxic basins were generated, fed by breccias and microturbidites from tectonically active margins. However, a similar situation to the one in Lombardy seems to exist in Carnia and also in Slovenia where, within the Dolomia Principale, anoxic basins were recognized (Ciarpica et al., 1987).

Norian carbonate platforms, locally prograding towards anoxic basins, have been pointed out: at the western side of Lake Como (Cirilli and Tannoia, 1988), west of Lake Garda (Picotti and Pini, 1989) and in the Tarvisio area (Lieberman, 1978).

In the western Southern Alps the Rhaetian succession is well documented (Lakew, 1990; Jadoul and Gnaccolini, 1992); in the eastern Southern Alps the Rhaetian stratigraphy is not well known. Different opinions about the lithostratigraphic setting exist principally about the position and the meaning of the Dachstein Limestone (cf. Bosellini, 1989; Ciarpica and Passeri, 1990; Doglioni et al., 1990).

**BIOCHRONOSTRATIGRAPHY**

Due to the uneven vertical distribution of the ammonoids in the Southern Alps, not all the Triassic intervals have the same biochronostratigraphical resolution.

Scythian successions are extremely poor in ammonoids (Fig. 3), which have been found only inside the Val Badia Member and the Cencenighe Member of the Werfen Formation (Broglio Loriga et al., 1990; Posenato, 1992). Such associations partly testify to a Spathian age. Therefore, to define the biochronostratigraphical position of the Scythian depositional sequences, conodont data in literature (Perri, 1991; Perri and Andraghetti, 1987) have been compared with ammonoid data from extraneous sections (e.g., Guex, 1978; Pakistani-Japanese Research Group, 1985; Wignall and Hallam, 1992, 1993). As a consequence, dating of the Scythian depositional sequences is less accurate than that of the overlying ones.

The interval comprised between the base of the Anisian and the base of the Carnian is covered by the highly resolved Middle Triassic ammonoid standard scale (Fig. 2) by Mietto and Manfrin (1995). The Aegean and Bithynian *p.p.* span in the Southern Alps, not well documented with ammonoids, is integrated with data from the Aegean Sea (Assereto, 1974; Fantini Sestini, 1981, 1988). The new standard scale is well correlatable with the classic scale so far used in the Mediterranean area (cf. Krystyn in Zapfe, 1983). The new scale was arranged contemporaneously with the lithostratigraphic and sequence stratigraphic revision of Triassic strata in the Southern Alps.

Ammonoid data have been integrated with literature data from the Southern Alps (Gaiani, 1969; Assereto, 1971; Rieber, 1973; Brack and Rieber, 1986, 1993), from the Northern Calcareous Alps and Balaton area (Vörös, 1987; Vörös and Palfy, 1989; Tatzreiter and Vörös, 1991) and from the Aegean area (Assereto, 1974; Krystyn and Mariolakos, 1975; Fantini Sestini, 1981, 1988; Krystyn, 1983). Lastly, all extra-Mediterranean Tethys and likely extra-Tethyan ammonoid data in literature have been controlled and compared.

As far as possible taxonomy necessary to the definition of zonal markers was carried out by revising original material in historical collections, particularly Mojsisovics and Hauer’s collections in Vienna.

The new standard scale is based on the definition of zones, more or less corresponding to the ones previously used (cf. Zapfe, 1983), each in turn subdivided into subzones. Zones are defined by genera, whose ranges have been verified throughout the Tethys, whereas subzones are characterized by specific markers. Greater value was given to the appearance of zonal and subzonal markers, whereas their disappearance was considered much less significant because a marker frequently survives in the next zone or subzone. Therefore, according to the international stratigraphic classification, the zones in the new scale are interval-zones.

The biochronostratigraphical significance of the zonal units in the new scale allows a better appreciation of some debated themes, such as the Anisian-Ladinian and Ladinian-Carnian boundaries. The hierarchy of the chronostratigraphic units and the importance of the biohorizons, which bound the zonal units, have been considered. The appearance of taxa hierarchically higher than genus (subfamily, family, superfamly), if also controlled outside the Tethys realm, was given a progressively higher value. A complete revision of taxa from superfamsily to species, often confirmed by the recognition of relationships and phyletic lines, has allowed the identification of biochronologic surfaces (appearance of superfamilies and families) hierarchically equivalent to stage and substage boundaries (“major events” in Krystyn, 1978 and in Tozer, 1984).

Both Anisian-Ladinian and Ladinian-Carnian boundaries were originally defined on a merely lithostratigraphic basis. Due to the lack of original, or not adequately emended, definitions, in accordance with Krystyn (1983), Mietto and Manfrin (1995) propose to place the Anisian-Ladinian boundary at the appearance of the superfamly Trachycerataceae (to which the authors refer some discussed genera such as *Nevdites* Smith, *Anolcites* Mojsisovics and *Chieseiceras* Brack and Rieber, on the basis of recognized affinities and/or phylogenetic relationships) associated to a generalized turnover of fauna. Analogously, they propose that the Ladinian-Carnian boundary is marked by the appearance of the superfamilies Clydonitaceae and Choristocerataceae and by events fairly well correlatable between the Tethys and North America.
The high resolution of the new standard scale is a useful tool in studying Middle Triassic strata in the Southern Alps, as sub-zonal units were recognized both in basinal and in shallow-water successions, both in terrigenous and in carbonate settings (cf. De Zanche et al., 1993; Gianolla, 1993). When ammonoids were lacking or rare or not characteristic, the biochronostratigraphic control was provided by conodonts. In this case, the scale in Krystyn (1983) has been used as a basic framework, integrated and modified by unpublished data and/or work in progress. No other taxonomic group has provided or seems likely to provide useful chronological indications with the exception of palynomorphs and, perhaps, tetrapod footprints. In the Southern Alps, numerous unpublished ichnofaunas have been collected in almost all Spathian to Carnian continental and sometimes Norian peritidal carbonate successions (Mietto, 1988). A preliminary study suggests that several biologic events are documented, which may be of profitable biochronostratigraphical use in the future.

In regard to Late Triassic, the ammonoid standard scale in Krystyn (1978, 1982) has been used, integrated with Early Carnian data by Mietto and Manfrin (1995). In the Southern Alps, latest Julian to Rhaetian ammonoids and conodonts became progressively rare and scanty. As a consequence, it was not possible to directly date lithostratigraphic units.

3rd-ORDER DEPOSITIONAL SEQUENCES

According to the methodology of Vail et al. (1977, 1991), at least 20 Triassic 3rd-order depositional sequences have been identified in the Southern Alps (Fig. 6).

The reconstruction of the sequence stratigraphy in the Dolomites (Fig. 7) has been a fundamental step in our work (De Zanche et al., 1992, 1993). Due to often spectacular outcrops, to the weak effects of Alpine tectonics, to the excellent dating by ammonoids and lastly to the abundance of data in literature, the Dolomites are an ideal area to establish an outcrop sequence stratigraphy. The depositional sequences have been controlled and extended throughout the Southern Alps and beyond.

In an outcrop sequence stratigraphic study, it is important to work along lines roughly perpendicular to the paleoshoreline. As an example, in the Ladinian-Carnian p.p. in the Southern Alps, the main direction of shoreline migration is south-north and this is particularly easily identified in the Dolomites.

During Triassic time, the Southern Alps were affected by synsedimentary tectonism and volcanism, which tended to mask or enhance the eustatic signal. For this reason, many authors are doubtful about the application of sequence stratigraphy in this region. On the contrary, we think that, supported by an excellent biochronostratigraphic control and by observations throughout the basin and in spite of tectonism and volcanism, the eustatic record is always recognizable.

In the following, every depositional sequence will be briefly described, using the following abbreviations in the text and in the figures:

- DS = depositional sequence, SB = sequence boundary,
- LST = lowstand systems tract, SMW = shelf margin wedge,
- TST = transgressive systems tract, HST = highstand systems tract,
- LPC = lowstand prograding complex, TLPC = top lowstand prograding complex,
- TS = transgressive surface, MFS = maximum flooding surface.

Sequence Sc1

In the Dolomites, the SB is to be placed within the uppermost part of the Bellerophon Formation (Late Permian), where bioclastic packstones and wackestones and terrigenous-carbonate grainstones (unit A, in Broglio Loriga et al., 1990, Fig. 4) sharply overlie grey peritidal dolomites. In eastern Lombardy, the Tesero Horizon, the lowermost member of the Werfen Formation, unconformably overlies a Permian continental substratum (Assereto et al., 1973). The topmost part of the Bellerophon Formation, consisting of bioclastic packstones and wackestones with scanty quartz grains, could be part of a clastic wedge of a not-yet-clearly defined LST or SMW deposit. The Tesero Horizon, a widespread oolitic event throughout the northern margin of the Te- thys (oolitic grainstones and intraclastic intercalations), and the lower part of the Mazzin Member (grey marly-silty mudstones) form the TST. The MFS, assigned a Griesbachian age due to the presence of *Isarcicella isarcica* (Huckriede), as documented by Perri (1991), is to be placed half way down the Mazzin Member where maximum depth and bioturbation are reached. The HST includes the upper part of the Mazzin Member (characterized by an increase in storm layers) which shows a shallowing upward trend (Broglio Loriga et al., 1990).

Sequence Sc2

Due to a downward shift of coastal onlap, the SB is to be placed at the abrupt contact of yellowish marly-silty dolomites (Andraz Horizon) above marls and marly mudstones (Mazzin Member). Because *Neospathodus dieneri* Sweet was found in the uppermost Mazzin Member (Perri, 1991), the SB seems to lie in the early Dienerian.

The SMW deposits consist of stacked shallowing-upward cycles made up of yellowish silty dolomites and multicolored laminated marls and siltstones. The TS is to be placed at the boundary between the supratidal facies of the Andraz Horizon and the subtidal deposits of the Siusi Member. The TST includes a few metres of oolitic-intraclastic-bioclastic packstones and grainstones overlain by alternating marly mudstones and bioclastic packstones. Ravinement surfaces and conglomerate layers (*Koken Conglomerate Auctorum* p.p.) are common and suggest a backstepping pattern. The MFS can be placed at the middle of the member, where offshore facies prevail and bioturbation is stronger. The upper part of the Siusi Member, including sandstones and hummocky laminated calcarenites, and the inter-supratidal marls and siltstones of the basal part of the Gastropod Oolite Member correspond to the HST and are related to the basinward migration of facies. The Siusi Member/Gastropod Oolite Member boundary is transitional.

Sequence Sc3

The SB lies within the Gastropod Oolite Member at the sharp transition from supratidal to subtidal deposits. Its age is probably latest Dienerian (cf. Perri, 1991). The TST includes the upper part of the Gastropod Oolite Member (alternating sandstones, arenaceous limestones, oolitic-bioclastic-intraclastic grainstones and packstones). Ravinement surfaces are common at the base of intraformational breccias (*Koken Conglomerate Auctorum* p.p.). The MFS could be placed at the top of the
Gastropod Oolite Member. The Campil Member, consisting of red laminated siltstones and sandstones with an upward increasing sand content and wave influence, forms the HST. Supratidal episodes are documented by tetrapod footprints.

**Sequence Sc4**

The SB is placed at the sharp contact between the sandstones and siltstones of the Campil Member and the supratidal marly dolomites of the basal Val Badia Member. Its age seems to correspond to the Smithian/Spathian boundary because *Neospastodus triangularis* (Bender) was found at the base of the Val Badia Member (Perri, 1991). The SMW consists of peritidal yellowish-grey marly dolomites and reddish siltstones from the lower part of the member. The TST begins with a well marked TS and includes alternating biocalcarenites and marly mudstones (unit B in Broglio Loriga et al., 1990, Fig. 8). This in-
interval contains numerous ammonoids (*Tirolites* beds Auctorum), at the top of which the MFS could be placed. The HST corresponds to the peritidal terrigenous-carbonate interval (lower unit C in Broglio Loriga et al., 1990, Fig. 8) which characterizes the middle of the member.

**Sequence Sc5**

The SB is to be placed above the peritidal interval within the unit C (in Broglio Loriga et al., 1990, Fig. 8) of the middle part of the Val Badia Member (Spathian). Unfortunately, the generally poor outcrop conditions do not permit exact placing of this surface. The TST includes alternating mainly grey bioturbated mudstones, silty/sandy bioclastic packstones and grainstones. Wave influence and sand/pelite ratios decrease upwards. The MFS is to be placed in correspondence with ammonoid-rich beds (*Diaplococeras* beds in Posenato, 1992) about half way down unit D in Broglio Loriga et al. (1990, Fig. 8). The uppermost part of the Val Badia Member shows a new progra-
Fig. 8.—Relationships between shelf (Piz da Peres section) and basin (Belvedere—Río Shade section) successions of the Anisian in the Braies/Prags area (northern Dolomites). 1 = unnamed subzone; 2 = Cuccense Subzone; 3 = Balatonicus Subzone; 4 = Binodosus Subzone; 5 = (?) Abichi Subzone; 6 = Trinodosus Subzone. a = An2 SB: Piz da Peres Conglomerate unconformably overlies the Lower Serla Dolomite; b = An3 LST: skeletal grain supported calcarenites are interbedded within the lower part of the Dont Formation; c = An4 SB: the continental Richthofen Conglomerate overlies the karstified top of the Upper Serla Formation.
dational feature and testifies to a shallowing-upward trend and an upward increase in hummocky and cross-bedded sandstones (HST).

**Sequence Sc6**

In the Dolomites, the SB corresponds to the surface separating the sandstones of the Val Badia Member from the oolitic-bioclastic calcarenites of the basal Cencenighe Member. Locally this boundary coincides with a subaerial exposure surface (Broglio Loriga et al., 1990). In the Recoaro area (southernmost Southern Alps), this surface corresponds to the erosional base of the Mt. Naro Breccia, that is a fluvial deposit consisting mainly of crystalline metamorphic clasts grading upward into the Cencenighe Member (De Zanche and Farabegoli, 1981). Due to its stratigraphic position, the Mt. Naro Breccia may also correspond to the Terra Rossa Siltites Member, defined in the neighborhood of Trento and occurring between the Campil Member and the Cencenighe Member (Ghetti and Neri, 1983). The TST consists of oolitic grainstones, marls and siltstones from the lower part of the member. The MFS is placed in the ammonoid-rich beds bearing *Dinarites dalmatinus* (Hauer). The HST is made up of crinoid and oolitic grainstones and packstones, siltstones and sandstones.

**Sequence An1**

In the eastern Southern Alps, the SB is placed at the base of the San Lucano Member since its lower part shows prevalent supratidal features. However, bad and scanty exposure does not permit a good definition of the lower part of this sequence. The age of the SB is latest Spathian, as documented by the presence of *Tirolites carnii* Mojsisovics in the “Dolomite event” (corresponding to the Lower Serla Dolomite) in the Muc section (Dalmatia: Krystyn, 1974; Posenato, 1992).

The TST seems to correspond to the upper part of San Lucano Member and to the lower portion of the Lower Serla Dolomite, which are subtidal. The MFS could be placed within the Lower Serla Dolomite between its subtidal and supratidal parts. The Lower Serla Dolomite (Bovegno Carniola in Lombardy) is a peritidal carbonate platform extending throughout the Southern Alps and known also in the Balaton area (Aszófő Dolomite, Hungary). In its upper part, an increase in supratidal events (mud cracks, tepees, caliches, red surfaces) is related to a decrease in accommodation space. A regressive trend in the upper part of the Bovegno Carniola has been underlined by Gaetani (1986) in the Adamello area. Therefore, this interval corresponds to the HST. Due to Anisian erosion, this sequence is normally lacking in the western Dolomites and in western Lombardy (Figs. 4, 5).

**Sequence An2**

The SB corresponds to a subaerial erosional surface at the top of the Lower Serla Dolomite, underlying the Piz da Peres Conglomerate (e.g., Braies area, De Zanche et al., 1992; Senowbari-Daryan et al., 1993). In central-eastern Cadore, it coincides with a paraconformity between the Lower Serla Dolomite and fine-grained siliciclastics at the base of the Gracilis Formation or its correspondents. In the Recoaro area, in Val Sugana and in the neighborhood of Trento, this unconformity is emphasized by the superimposition of the Val Leogra Breccia on the Lower Serla Dolomite (De Zanche and Mietto, 1989). Generally in Lombardy, the Lower Serla Dolomite is overlain by fine-grained siliciclastics; locally terrigenous-carbonate breccias can be found (Maniva Breccia in De Zanche and Farabegoli, 1988). This sequence is deposited in a ramp setting, and the LST only consists of infilling of weakly incised valleys (e.g., Braies area). In the eastern Southern Alps, the TST corresponds to terrigenous-carbonate sediments in the upper part of the Piz da Peres Conglomerate where an upward decrease in siliciclastics is observed. The position of the MFS is uncertain; it could be placed at the base or within the Gracilis Formation, consisting of wackestones and dasyycladacean packstones-grainstones interlayered by calcisilites and siltstones (HST). Locally in the Braies area, in the Zoldo area and in Carnia, the Gracilis Formation is heteropically overlain by a carbonate platform (not yet formally defined, containing dasyycladaceans, foraminifers and *Tubiphytes*) that corresponds to the late HST in those areas. In Lombardy, the An2 DS corresponds to the lowermost part of the Angolo Limestone, is bounded above by a polymictic breccia, and is characterized by an evaporitic trend (Gaetani, 1986). In Brembana Valley, the HST is made up of peritidal limestones and dolomites bounded by an unconformity (Jadoul et al., 1992c).

**Sequence An3**

In the eastern Southern Alps, the SB is placed at an erosional surface between the Gracilis Formation and the Voltgao Conglomerate or at a karst surface at the top of carbonate platforms (De Zanche et al., 1993). In the Recoaro area, it is a surface of abrupt facies change between the Gracilis Formation and the overlying continental Voltzia beds (Barbieri et al., 1980). In eastern Lombardy, it could lie within the Angolo Limestone, above the brecciated evaporitic level at about 20 m from the base of the formation (cf. Gaetani, 1986). In the Brembana Valley, it coincides with an unconformity that subdivides the peritidal dolomites of the Angolo Limestone (Jadoul et al., 1992c). In western Lombardy, the SB is placed at the base of the Val Sassina Conglomerates (De Zanche and Farabegoli, 1988). Only indirect biochronostratigraphic data permit us to determine the age of the SB (Ismidicus Subzone). Ammonoid data from the Aegean region (Kokaeli in Assereto, 1974; Fantini Sestini, 1988) permit us to correlate the conglomerate in Kokaeli with the Voltgao Conglomerate in the Dolomites, where latest Bithynian ammonoids (unnamed subzone 2) have been found in the lower part in the An3 DS (Fig. 8).

In the emerged areas (e.g., peripheral parts of the western Dolomites), the late LST consists of the infilling of incised valleys. In the areas characterized by a ramp setting (e.g., Piz da Peres area, northern Dolomites), the LST consists of a wedge of siliciclastics and bioclastic-grainstones while in the basinal areas it is composed of sandstones, siltstones and bioclastic-grainstones containing latest Bithynian-early Pelsonian ammonoids. In the Dolomites, the TST includes sandstones, siltstones and limestones in the Voltgao Conglomerate, the lower part of the Recoaro Limestone and of the Agordo Formation p.p. because they show a fining and deepening upward trend. The MFS is placed in an ammonoid-rich bed (Balatonicus Subzone) inside the Dott Formation (Fig. 8); in the Recoaro Limestone and in the Agordo...
Formation p.p. it corresponds to the appearance of nodular, bioturbated, fossil-rich wackestones-packstones. In the eastern Dolomites and in Carnia, the TST is locally represented by a backstepping carbonate platform, rich in ammonoids (unnamed subzone 2, Cuccense Subzone, lowermost Balatonicus Subzone), not yet formally defined. This platform is drowned by red pelagic limestones or silty limestones of the Balatonicus Subzone. The HST corresponds to the upper part of the Recoaro Limestone-Agordo Formation p.p., generally rich in bioclasts, due to the growth and prograding of the Upper Serla carbonate platform. In the basinal succession, the deposits related to HST time are characterized by an increase in siliciclastics and bioclasts. In Lombardy, the An3 HST corresponds to the upper part of the Angolo Limestone, which is heterophsic with two coeval carbonate platform bodies, the Dosso dei Morti Limestone and the Camorelli Limestone. The age of HST is well defined as Balatonicus Subzone p.p. to lowermost Abichi Subzone.

**Sequence An4**

In Carnia, in the western and northern Dolomites, in the Recoaro area and in the Lugano-Lake Como area, the SB corresponds to a strong subaerial erosional surface. This well-defined unconformity is placed at the base of conglomerate-sandy-silty lithozones (Figs. 4, 5) and deeply cuts Anisian, Scythian and Permian units. In the Villaverla 1 A.G.I.P. well, the SB coincides with an angular unconformity on the pre-Permian metamorphic basement.

In the parts of the Dolomites where the previous HST carbonate platforms (Upper Serla Formation) are preserved, the SB coincides with a karst surface. In basinal areas, it is represented by a conformity lying inside the uppermost part of the Dont Formation and is placed at the beginning of a strong increase in siliciclastic turbidites.

In Lombardy, the SB is placed at the top of the Camorelli and Dosso dei Morti carbonate platforms, which are abruptly overlain by the Prezzo Limestone. In basinal areas, the SB is placed at the top of the so called “Banco a Brachiopodi”, a bioclastic wackestone-packstone interval, rich in brachiopods, which constitutes the top of the Angolo Limestone.

In shelf areas, the LST includes infilling of incised valleys and progradational clastic wedges (Ugovizza Breccia p.p., Richtofen Conglomerate p.p., Tretto Conglomerate p.p., Val Muggiasco Conglomerates p.p.), while in the basins it consists of skeletal limestones, distal turbidites and lenses of conglomerates and sandstones (uppermost part of the Dont Formation in the eastern Southern Alps and lower part of the Prezzo Limestone in Lombardy, Abichi Subzone—Trinodosus Subzone p.p.).

In shelf areas, the TST comprises the upper part of the conglomerate units while, in the basins, it includes most of the Mt. Bivera Formation and a part of the Prezzo Limestone, which are mainly made up of pelagic nodular limestones and siltstones at the top of which the MFS is placed (Trinodosus Subzone).

In shelf areas in the Dolomites, the HST is made up of the Morbiac Dark Limestones and the Contrin Formation whereas in Lombardy it includes the Mt. Albiga Dolomite (Lake Como) and the San Giorgio Dolomite (Lugano area). In basinal areas, the HST consists of the lower part of the Ambata Formation and a part of the Prezzo Limestone. Due to progradation of the carbonate platforms and to basinward migration of the source of terrigenous supply, the basal sediments of the An4 DS consist of fine silty and sandy calcarenites interbedded with marls, siltstones and sandstones. The age of the An4 HST is to be referred to Trinodosus Subzone p.p.—Reitzi Subzone p.p.

An open problem concerns the Moena Formation in the Dolomites. Due to strong extensional tectonism, coeval intraplate basinal sediments often sharply overlie the previous carbonate platforms. In these cases, the SB seems to correspond to a drowning unconformity caused by a strong increase in subsidence and in sea-level rise.

In the Dolomites, the correlative basal conformity can be drawn between calcarenites and marls (lower Ambata Formation, previous HST) and the hemipelagites and fine turbidites (LST) which characterize the upper part of the Ambata Formation (the so called “Daonella Marls”). Locally the LPC includes conglomerates and sandstones. The age of the SB is to be placed within the Reitzi Subzone and the TLPC within the Avisianum Subzone.

In central and eastern Lombardy basins, the SB is placed within the upper part of the Prezzo Limestone (Reitzi Subzone) but, up to date, no lithological evidence has been recognized, probably due to the distance of siliciclastic sources and of previous prograding carbonate platforms.

In the basins, the TST includes the Plattenkalke and/or a part of the Knollenkalke of the Livinallongo Formation (Dolomites) and a part of the Besano Formation/Grenzbitumenzone or Varenna Formation (Lombardy), whereas in the strongly subsiding shelf areas it consists of a catch-up (locally give-up) carbonate platform (Lower Edifice = Tiarfin Dolomitic Limestones in Carnia). Both the Plattenkalke and the Lower Edifice started within the Avisianum Subzone. The Lower Edifice grew only above previous Contrin-Mt. Albiga platforms and consists of a set of backstepping shallow-upward carbonate parasequences frequently bearing ammonoids (De Zanche et al., 1993).

The MFS is placed within the lower part of the Knollenkalke and is assigned to the Chiesense Subzone (Fig. 9). It corresponds to the maximum extent of the Triassic basins in the Southern Alps. At this time the backstepping carbonate platforms were definitely drowned, whereas in the basins an ammonoid-rich interval (“Chiesense groove” in Brack and Rieber, 1986, 1993) was formed. The MFS event is also documented by the migration of ammonoid faunas from the Tethyan domain.
into the Germanic basin and vice versa during a temporary connection between the two basins (Mietto and Manfrin, 1995).

During HST time, carbonate platforms (Sciliar Dolomite 1, Mt. Spitz Limestone 2, Sindech Dolomite, Esino Limestone 1, San Salvatore Dolomite 1) aggraded and moderately prograded basinward. The corresponding basal sediments are characterized by a moderate upwards increase in calcarenites and in siliciclastics (mainly epiclastic volcanic material). The age of the HST ranges between the Chiesense and lowermost Recubariensis Subzones.

**Sequence La2**

The SB corresponds to the top of the Sciliar Dolomite 1, Mt. Spitz Limestone 2, Sindech Dolomite, Esino Limestone 1, San Salvatore Dolomite 1. Where it was possible to check, we have found an erosional and karst surface. The correlative conformity in the basinal successions is one of the most difficult to be placed and is an open problem. As an example, the Ladinian basins in the eastern Lombardy (cf. Brack and Rieber, 1986, 1993) do not show marked lithologic changes as probably they were too far, or protected, from the southern terrigenous supply. However, in many sections in the Dolomites, it seems to correspond to the abrupt increase in siliciclastics that divides the cherty nodular limestones of the Livinallongo Formation into two parts.

In the basinal Varenna Formation (eastern side of Lake Como), the SB is placed at the base of an interval consisting of slumps, breccias and pebbly-mudstones which separates the “Orizzonte di Regoledo” and the “Orizzonte delle Cave” in Giannotti and Tannoia (1988).

Due to the previous strong transgression and therefore to the great shifting of the shorelines, the LST generally occupies a thin interval. It consists of distal volcaniclastic turbidites derived from erosion of volcanic belts. Near the platform margins, the LST deposits may be formed by calcarenites, rudites and megabreccias.

In the shelf areas, the TST probably consists of subtidal dolomites located at the base of Sciliar Dolomite 2 and Esino Limestone 2; in the basins, it corresponds to a portion of the upper part of the Knollenkalke (Fig. 9).

The MFS lies within the Knollenkalke and is assigned to the Recubariensis Subzone; on the top of the Sciliar Dolomite 1, it is locally documented by red pelagic limestones.
The HST is characterized by a strong progradation of the Sciliar Dolomite 2, Esino Limestone 2 and San Salvatore Dolomite 2, often masking the underlying platform which, on the contrary, seems to be much more aggrading than prograding. In the basins, the platform progradation is testified by an upward increase in neritic-derived calciturbidites (Bänderkalke, upper Livinallongo Formation). The basinward migration of the terrigenous shoreline is documented by an upward increase in siliciclastics in the topmost Livinallongo Formation. The HST is to be assigned to the Recumbiensis-Gredleri p.p. Subzones.

**Sequence La3**

In shelf areas, the SB is to be placed on the top of the eroded and sometimes karstified carbonate platforms of the previous La2 HST. In the basins, it corresponds to a strong increase in siliciclastics; in the eastern Southern Alps, it corresponds to the sharp boundary between the Livinallongo Formation and the Zoppe Sandstones, while in central Lombardy it corresponds to the base of a carbonate breccia lying within the upper part of the Varenna Formation (Gaetani et al., 1992a).

The characteristic of the Zoppe Sandstones in the eastern Southern Alps could be considered as an example of the features of the LST. This unit consists of arkosic turbiditic sandstones; its lower part is composed of massive or amalgamated turbidites (basin floor fan); a lithozone, consisting of thin-bedded turbidites, the highest of which are channelized, overlies the massive sandstones and is interpreted as a slope fan; lastly the upper part of the unit is characterized by a thickening- and coarsening-upward turbiditic succession, considered to be a LPC. Elsewhere, the LST deposits consist of distal siliciclastic turbidites.

In the basins, the TST is represented by dark calcilutites or micritic nodular limestones. Frequently at their top, a fossil-rich layer corresponds to the MFS. On the top of previous carbonate platforms, pelagic limestones or neptunian dikes bearing ammonoids have been found (Gaetani et al., 1981; Blendinger et al., 1984; Jadoul et al., 1992c; De Zanche et al., 1993; Gianolla, 1993). This MFS, well documented throughout the Southern Alps, is referred to the Longobardicum Subzone.

In the eastern Southern Alps the La3 HST is generally influenced by a strong basic volcanism that produced a great quantity of lavas and volcanioclastic. Volcanic activity seems to be strictly connected with a strike-slip faulting that is responsible for local uplifts, compressional structures, normal faults and chaotic deposits (cf. Blendinger, 1985; Doglioni, 1987; Sloman, 1989; Gianolla, 1993). Where volcanism was intense, stratal patterns could be altered, and the different systems tracts are not easily recognizable (cf. Doglioni et al., 1990; De Zanche et al. 1993).

In the areas of the eastern Southern Alps that are not or less affected by volcanism, the La3 HST is made up of marls and sandstones and is related to the progradation of the southern shoreline. In shelf areas in the Dolomites, carbonate platforms are also present (Sciliar Dolomite 3), prograding on basinal volcanioclastics of the Fernazza Formation (e.g. Schlern area, Sd II tongue 1, in Brandner, 1991) or of the Acquatona Formation (easternmost Southern Alps). In the Grigne area, the La3 HST is found in the upper part of the Esino Limestone (Esino Limestone 3) which is heteropic with the uppermost Varenna Formation (including its Perledo Member) and the Lierna Formation (Gaetani et al., 1992a). Elsewhere in Lombardy, the La3 DS is also recognizable (e.g., Brembana Valley, part of lithozones 5 and 6 in Jadoul et al., 1992c). The age of the HST is well documented as the interval Longobardicum—Neumayri Subzones.

**Sequence Car1**

Throughout the Southern Alps, the SB is a major unconformity corresponding to a strong karst surface on top of the Sciliar Dolomite 3, Esino Limestone 3 and San Salvatore Dolomite 3 or of oldest carbonate platforms.

In the basins in the western Dolomites, the SB coincides with an important submarine erosional surface at the base of the Marmolada Conglomerate (lower part of the La Valle Formation) on the Fernazza Formation or volcanics; locally the erosion cuts deeply into older underlying units. Therefore in this area, the early LST is characterized by coarse volcanioclastic conglomerates (Marmolada Conglomerate) deriving from erosion in nearby subaerial volcanic areas (Fig. 10). The late LST (Regoledanus Subzone) consists of a turbiditic complex made up of sandstones, shales and conglomerates (Civetta Conglomerate and Marmolada Conglomerate). Skeletal grain supported calcarenites and hybrid sandstones, often bearing karstified carbonate blocks of previous Sciliar 3 buildups, are widespread. In the eastern Dolomites the early LST (Neumayri Subzone) corresponds to distal turbidites and hemipelagites of the La Valle Formation. In central-western Lombardy, due to upward increase in supratidal events, we have interpreted the so called Calcare Rosso (cf. Assereto et al., 1977b; Mutti, 1992) as infilling of paleokarst cavities (late LST). Mutti (1992) and Jadoul and Gnaccolini (1992) think that the Calcare Rosso corresponds to an additional complete 3rd-order DS.

Marls, marly limestones and calcilutites, belonging to the Regoledanus Subzone, define the TST.

In the eastern Southern Alps during HST time, the Cassian Dolomite 1 prograded onto the lower part of the San Cassiano Formation (e.g., Richthofen Riff, Sasso Bianco, Mt. Coldai and Dürenstein) and consists of alternating marls, marly limestones, oolitic packstones-grainstones, biocalcarenites and sandstones (Masetti et al., 1991). In the southern areas, where there are no carbonate platforms, the HST is made up of prodelta to submarine delta deposits (De Zanche et al., 1993). In Lombardy, the Carl HST consists of the Calcare Metallifero Bergamasco and the Breno Formation 1, prograding towards the basin of the Wengen Formation and Lozio Shales. The age of the Carl HST is referred from uppermost Regoledanus Subzone to Aon Subzone p.p. (Fig. 11).

**Sequence Car2**

The SB is a karst surface located on top of the Cassian Dolomite 1, Breno Formation 1, Calcare Metallifero Bergamasco (Assereto et al., 1977b; Jadoul and Gnaccolini, 1992; De Zanche et al., 1993).

Relatively deep basins existed only in the eastern Southern Alps. In the San Cassiano Formation, the correlative conformity corresponds to the beginning of a strong increase in siliciclastics and skeletal grain calcarenites above oolite packstones-grainstones of the previous HST.
Fig. 10.—Car1 LST, Punta Grohmann section, Sassolungo massif (western Dolomites). Legend as in Figure 8. a = karstified carbonate block within sandstones and conglomerates; b = a conglomerate lense within sandstones and pelites; c = a slump involving thin turbidites; d = a chaotic body, consisting of carbonate blocks and volcanic pebbles, forms the upper part of the Car1 LPC; the carbonates seem to have been karstified in situ.
FIG. 11.—Stuores Wiesen section, Pralongià, neighbourhood of Corvara (western Dolomites). Legend as in Figure 8. a = Car1 HST: alternating clayey siltstones and oolitic biocalcarenites; b = alternating siltstones and fine-grained sandstones; c = Daxatina cf. canadensis (Whiteaves); d = Trachyceras muensteri (Wissmann).
The LST consists of prevailing siliciclastics of the San Cassiano Formation (Aon Subzone) lapping on the slope of Cassian Dolomites.

The TST consists of ammonoid-rich marls and marly limestones. The MFS is assigned to the Aonoides Subzone (e.g., San Cassiano Formation lapping on the Richthofen Riff, Urschits, 1994; Breno Formation, Camonica Valley, Allasinaz, 1968).

In Lombardy, the LST and the TST are characterized by a strong siliciclastic-volcaniclastic input (Val Sabbia Sandstones), related to a Carnian southern volcanic phase (Garzanti, 1985).

In the Dolomites, the HST is represented by Cassian Dolomite 2 strongly prograding onto the basinal sediments (oolitic grainstones-packstones, biocalcareites and marls) of the upper part of the San Cassiano Formation. The age of HST is documented by ammonoids of the Aonoides Subzone-Austriacum Zone p.p. (Bizzarini and Braga, 1987; Gianolla, 1993). In the eastern Dolomites, the northward migration of the terrigenous shorelines is documented by prograding deltas (e.g., Marmarole area, central-eastern Cadore). In Lombardy the Car2 HST includes the Breno Formation 2 and part of the Gorno Formation.

**Sequence Car3**

In the eastern Southern Alps, the SB is placed at the top of the eroded and karstified Cassian Dolomite 2. In the previous marginal areas, that were occupied by prograding deltas of the Car2 HST, it corresponds to the pedogenesis of coastal deposits. In the basins, it corresponds to the abrupt superposition of carbonate shallow-water facies (Dürrenstein Formation) above the pelagic sediment of the San Cassiano Formation (Russo et al., 1991).

In central-western Lombardy, the SB is placed within the Gorno Formation (cf. Jadoul and Gnaccolini, 1992); in the Vareso area it seems to correspond to the boundary between the Cunardo Formation and the Pizzella Marls.

In the eastern Southern Alps, this sequence starts with a carbonate SMW (lower part of the Dürrenstein Formation, Rio Conzen Limestone) lapping on the Cassian Dolomite 2 slope; the arenaceous dolomites, deposits related to this systems tract, pinch out landward to the offlap break of previous HST (Fig. 12).

Calcarenites, sandstones, and shales, overlying the shelf margin sediments, form the TST. On top of this interval, fossil-rich marls (brachiopods, thin bivalves) define the MFS. Where the Dürrenstein Formation is mainly carbonate, the TST is made up of upward thickening peritidal dolomites whose subtidal intervals are prevalent.

On the whole in the Dolomites, the HST (upper part of the Dürrenstein Formation) is characterized by upward thinning, by an upward increase in paleosols and by widespread supratidal facies. A dolomitized oolitic complex is created in the Tofana-Antelao area (De Zanche et al., 1993).

In Lombardy the Car3 HST is emphasized by the progradation of transitional and alluvial facies of the lower part of the San Giovanni Bianco Formation (Gnaccolini and Jadoul, 1990).

The age of this sequence has not yet been well defined. However, due to the presence of ammonoids belonging to the “Oedipus” Subzone in the uppermost San Cassiano Formation (Cortina d’Ampezzo area), the SB cannot be older than the latest Julian-early Tuvalian.

At the end of this time, the Triassic basins in the Southern Alps were almost completely infilled.

**Sequence Car4**

The SB is a major regional unconformity which in the Dolomites corresponds to the base of the Raibl Formation. South of the Dolomites (e.g., Recoaro area), it deeply cuts the underlying units.

According to Senn (1924), who described conglomerates within the Pizzella Marls in westernmost Lombardy, the SB could be tentatively placed inside this unit; in the western side of Lake Como, Bertotti (1991) also pointed out a 10- to 20-m-thick quartzarenite interval lying within its “Raibler Beds”.

Therefore, we think that the Car4 SB could be placed at the base of these conglomerates and quartzarenites.

The LST is recorded only in the southern Dolomites and mainly consists of the infilling of incised valleys (e.g., Pass Duran, De Zanche et al., 1993).

The well-known sandstones, siltstones, varicolored shales, aphanitic dolomites and muddy limestones of the Raibl Formation belong to the TST. The MFS is reached at the base of the overlying paralic and sabkha deposits.

The limestones, calcarenites, shales, vuggy dolomites and gypsum, forming the upper part of the Raibl Formation, can be assigned to the HST because they show an upward-thinning trend and increasing supratidal features.

In Lombardy, this DS is represented by the transgressive-regressive trend of the San Giovanni Bianco Formation (Jadoul and Gnaccolini, 1992).

The age of this sequence is hardly definable in the Southern Alps as no ammonoids or conodonts are known. Only scanty ammonoid data (Subbullatus Subzone) from eastern basinal units that correspondent to the Raibl Formation suggest a Tuvalian age (Geyer, 1900; Gianolla, 1993).

It is important to underline that by the Raibl Formation we mean only the mainly continental and paralic terrigenous-carbonate sediments which lie between the Dürrenstein Formation, or more ancient units, and the Dolomia Principale (De Zanche et al., 1993). In this sense, it must not be confused with the Raibl Group in Assereto et al. (1968) and the Raibler Schichten (or Group) in the Austrian and German geological literature, which more or less cover the entire Carnian deposition.

**Sequences No1 and No2**

During latest Carnian-Norian time throughout the Southern Alps, carbonate tidal flat and lagoon conditions were established (Bosellini, 1989). Because the shoreline shifted far eastward, coastal onlap relations are difficult.

The identification of 3rd-order DSs inside the Norian succession is therefore a real problem throughout the Southern Alps because, in an interval of more or less 10 Ma, several SBs are expected. So far however, only two DSs within the Norian sections in the Southern Alps seem to exist.

The No1 SB is placed at the base of the breccias and of the erosional extraformational fine conglomerates which in the eastern Southern Alps often characterize the base of the Dolomia Principale. In most of Lombardy, it is coincident with the...

In the Southern Alps, deposits of LST are not available, because the shoreline break shifted a great distance. Incised valley infillings have not yet been recognized.

The No1 TST deposits are made up of intraformational breccias, interlayered with micritic and pelletiferous dolomites; locally dark laminated dolomitic limestones are present. A similar setting is described in Lombardy (Jadoul et al., 1992a).

No1 HST stratal patterns also seem to exist in the Julian Alps (Tarvisio area), where the lower part of the Dolomia Principale progrades (Lieberman, 1978; Doglioni, 1988) onto pelagic terrigenous-bituminous limestones Tuvalian in age.

A possible No2 SB within the Dolomia Principale in the eastern part of the Southern Alps could correspond to the erosional base of the Passo Buse Scure Breccia, a channelized fluvial deposit bearing abundant crystalline basement clasts that crop out in the Recoaro area (De Zanche and Mietto, 1984).

Tentatively, in the Dolomites, De Zanche et al. (1993) placed this SB at the base of the shale layer, which forms a typical ledge within the lower part of the Dolomia Principale, at about a hundred meters from the base of the unit. However, a first attempt at analyzing the variation in accommodation space inside the carbonate tidal flat sediments of the Dolomia Principale, indicates that probably this SB would be better placed higher.

Due to the lack of or the rarity of significant fossils, the correlation of the Norian sequences between the eastern and the western Southern Alps is a real problem. Both in Lombardy and in the easternmost Southern Alps (Friuli, Carnia), the Dolomia Principale plateau was dissected by an extensional tectonic event. As a consequence in those regions intraplatform anoxic basins were formed (Ferasin et al., 1969; Mattavelli and Rizzini, 1974; Jadoul, 1986; Ciarpica et al., 1987; Stefani, 1989; Jadoul et al., 1992a).

In Lombardy, where the Norian succession is well developed, the stratigraphic setting is complicated owing to the tectonic control, so that quite dissimilar interpretations exist (cf. Stefani, 1989; Gaetani et al., 1992b; Jadoul and Gnaccolini, 1992).

In the basinal areas, the mainly carbonate-terrigenous deposition of the Aralalta Group (Jadoul, 1986) was abruptly interrupted by the deposition of the lower member of the Riva di Solto Shales (LST). The latter progressively lapped on the previous structural highs, above which the growth of the Dolomia Principale probably continued, as demonstrated in the basins by the calcareous intercalations within the middle member of the Riva di Solto Shales (HST).

Sequences Rh1 and Rh2

The definition of Rhaetian DSs presents the same difficulties as the Norian sequences. In the eastern Southern Alps, the Rhaetian stratigraphic setting is unclear. Doglioni et al. (1990) cite an erosional unconformity between the Dolomia Principale and the Dachstein Limestone in the Sella Group (western Dolomites).
In Lombardy, on the basis of high-frequency cyclostratigraphy, Lakew (1990) pointed out the existence of two 3rd-order DSs in the Zu Limestone. According to Jadoul and Gnaccolini (1992), the Rh1 SB should be placed between the middle and the upper member of the Riva di Solto Shales; the age of this sequence is uncertain, but it is probably Sevastian to Rhaetian in age. Furthermore, the Rh2 SB is placed at the top of the coral and oolite limestones lying within the Zu Limestone (Rhaetian).

TRANSGRESSIVE/REGRESSIVE FACIES CYCLES AND BASIN EVOLUTION

Because sequence boundaries have a chronostratigraphic significance (Van Wagoner et al., 1988; Vail et al., 1991), sequence stratigraphic units are to be considered as physical chronostratigraphic units.

We tried to use the thickness of the 3rd-order DSs as a convenient and quick tool for better understanding the Triassic tectono-sedimentary evolution in the Southern Alps. Reliable data on paleodepth of basinal facies are lacking, and in order to avoid introducing subjective data, we tentatively considered maximum thicknesses of sediments in carbonate buildup environments, that is where sedimentation occurs at or near the base level. This approach, which in first approximation theoretically does not need corrections either for paleodepth or for compaction, allowed us to express the variations in accommodation space in carbonate platform areas and to construct the diagrams in Figures 13 and 14. In reality, due to the lack of base level controlled sediments in some Triassic intervals in the Southern Alps, some approximations exist in our schemes; thus, the total thicknesses of the Upper Permian, Scythian and some Anisian units have been considered.

Supposing that in the same basin and in the same time span the biological factories of the carbonate buildups were essentially the same and environmental conditions were equal, then the buildup potential growth rates would be equal. In such a way, thicknesses of depositional sequences, bounded by “subaerial” sequence boundaries, are measured and give a speedy approximation to the total subsidence. Because in shelf areas LST deposits are absent or confined to infilling of incised valleys or exist in the form of SMW, measured thicknesses mainly include TST and HST deposits.

Diagrams of variation in accommodation space in shelf areas have been prepared by measuring thicknesses in selected areas (Fig. 1) throughout the Southern Alps. A west-east section and a south-north section are shown (Figs. 13, 14). In these thickness diagrams, neither actual distances nor palinspastic positions are considered. In fact, a certain and universally accepted paleogeographic-palinspastic frame is not yet available. However, we think that the shortening, which the Southern Alps underwent during Tertiary time, has not considerably altered the original Triassic paleogeographic framework modifying the relative position of sedimentary domains. In short, these diagrams are not necessarily an exhaustive scheme of the evolution of total subsidence of the Southern Alps during Triassic time because we have not considered the subsidence in basinal areas, which may be far different from that in shelf areas. However, we believe that they show evolutionary trends on the basis of data that we consider reliable or, at least, acceptable.

The recognition of 3rd-order DSs and systems tracts, which makes it possible to define higher order cycles, is a fundamental tool in the reconstruction of the evolutionary trends of a basin (Vail et al., 1991). A 2nd-order cycle is a transgressive/regressive facies cycle, consisting of a set of stacking 3rd-order DSs, bounded by two maximum regressions. Therefore the boundary of a 2nd-order cycle is coincident with a TLPC. Between the transgressive phase (aggrading and backstepping set of sequences) and the regressive phase (infilling and foresteping) a peak transgression is placed, which coincides with a major 3rd-order MFS (Jacquin et al., 1992). Besides eustasy, a 2nd-order cycle is markedly controlled by the tectonic history of the basin. In the Southern Alps, five T/R facies cycles has been recognized (Fig. 6).

In our opinion the Sc1 TLPC, close to the Permian-Triassic boundary, also coincides with the base of a 1st-order cycle (“continental encroachment cycle” in Duval et al., 1992) as it corresponds to a maximum basinward shift of coastal onlap (cf. problems about the Permian-Triassic boundary in Assereto et al., 1973; Noé, 1987; Broglio Loriga et al., 1990; Wignall and Hallam, 1992, 1993). Most probably the Rh1 TLPC is the upper boundary of this cycle, whose peak transgression lies in Fassanian time (Chiesense Subzone).

Considering the 3rd-order sequence chronostratigraphic framework together with the diagrams of the variations in accommodation space, the T/R facies cycles setting, and therefore the phases of the Triassic tectono-sedimentary evolution in the Southern Alps, can be schematically outlined as follows.

I—?Aegean-Earliest Bithynian

This interval seems to correspond to a 2nd-order cycle and is bounded by two maximum regressions: at its base by the TLPC placed at the Permian-Triassic boundary (cf. Assereto et al., 1973; Wignall and Hallam, 1992) and at its top by the TLPC of the An2 DS. A good candidate for the 2nd order peak transgression is the 3rd-order MFS of Sc5 DS (Tirolites cassianus Zone).

This set of DSs is characterized by prevalent shallow-water terrigenous-carbonate units, deposited in a ramp setting. It essentially includes the Scythian units and the Lower Serla Dolomite. The latter consists of a widespread peritidal carbonate platform whose deposition caused a marked uniformity throughout the Southern Alps (De Zanche and Farabegoli, 1982).

Structural trends seem to have the same style as the underlying Upper Permian succession which, on the contrary, is strongly discordant with the Lower Permian units.

A west-east polarity is evident, with reduced successions to the west, whereas rock units are thicker and more definitely sea-influenced to the east. Transgressions rise from the east, as evidenced by Scythian basins in the Dinarides (Herak et al., 1983). Regarding the south-north section, little speculation is possible due to the lack of thickness data from the units eroded during Anisian to Carnian time. However, a northward thickening trend seems recognizable. Proximal siliciclastic units, such as Mt. Naro Breccia (Recoaro area, Fig. 5, col. III) and Terra Rossa Siltstones (Valsugana, Fig. 5, col. IV), placed at the base of the Sc5 DS, suggest southern emerged source areas and seem to indicate a strong sea-level drop. On the other hand in the Braies/Prags Dolomites, proximal conglomerates, lying in the same stratigraphic position suggest a northern source.
FIG. 13.—Diagrams on variations in accommodation space in shelf areas (Upper Permian-Fassanian): west-east and south-north sections across the Southern Alps. Selected areas as in Figures 4, 5. See outline in Figure 1.
FIG. 14.—Diagrams on variations in accommodation space in shelf areas (Longobardian-Tuvalian): west-east and south-north sections across the Southern Alps. Selected areas as in Figures 4, 5. See outline in Figure 1.
Therefore, a double polarity seems to be indicated, characterized by structural and morphological highs placed to the north and to the south but always by a thickening trend toward the east.

2—Bithynian-Early Illyrian

In Bithynian time previous facies uniformity was broken and the first basins were defined. They were characterized by terrigenous-carbonate supply and by the appearance of Anisian ammonoids. At the same time, the first carbonate platforms grew (Braines/Prags area, Cadore, Carnia in De Zanche et al., 1993) much earlier than was previously believed (cf. Pisa, 1974; Assereto et al., 1977a; Pisa et al., 1979; Farabegoli and Guasti, 1980; Gaetani et al., 1981; Fois and Gaetani, 1984; Farabegoli et al., 1985; De Zanche and Farabegoli, 1988).

Structural trends (both north-south and east-west) seem to suggest strong extensional tectonics which, together with sea-level changes, controlled the distribution, the extension and the evolution of basins and carbonate platforms. On the whole, eastern basins seem to have developed earlier (Fig. 4, cols. 8, 10, 11; Fig. 5, cols. V, VII).

The differentiation trend was emphasized during the An3 DS time, when throughout the Southern Alps strongly subsiding areas were defined. A horst-and-graben setting was formed. Structural highs were often occupied by continental deposits or by terrigenous or carbonate platforms, whereas basins were strongly polluted by siliciclastic supply. The strong increase in subsidence was connected with a marked sea-level rise. As a result, a strong increase in accommodation space occurred, which made the deposition of great sediment thickness possible and which forced carbonate platforms to back-step (An3 TST). Carbonate platforms sank generally before MFS (De Zanche et al., 1993).

Diagrams on variations in accommodation space in shelf areas (Fig. 13) show that north-south and east-west tectonic structures, likely to be related to a strike-slip tectonism, cross each other and form a “chess-board” of highs and lows, further affecting the geometry of the basins (Blendinger, 1985; Doglioni, 1987; De Zanche et al., 1992, 1993).

This set of DSs seems to be arranged in a 2nd-order transgressive-regressive facies cycle, bounded at the top by a significant change in tectonic subsidence rate. The An3 MFS (Balatonicus Subzone) expresses the peak transgression; the communication between German and Tethyan basins is demonstrated by the presence of Germanic ammonoids in the Tethyan domain and vice versa (Mietto and Manfrin, 1995). Soon after the communication between the two basins was interrupted.

3—Illyrian-Longobardian

The next 2nd-order cycle includes An4 to La3 DSs bounded by maximum regressions (respectively An4 and Carl TLPC). The peak transgression corresponds to the La 1 MFS (Chiesense Subzone), well-defined throughout the Southern Alps and emphasized by the migration of Alpine ammonoids into the German Basin and vice versa (Mietto and Manfrin, 1995) when the two domains once again became connected.

In regard to basin evolution, it is useful to discuss the transgressive and the regressive phases separately. In the shelf areas, the lower boundary coincides with the strong, sometimes angular, unconformity between fluvial conglomerates (Val Mugiasca Conglomerates, Tretto Conglomerate, Richthofen Conglomerate and Ugovizza Breccia p.p.) and the underlying pre-Permian to Anisian units.

In the An4-La1 interval the Southern Alps underwent a generalized increase in subsidence that also involved previous structural highs (Fig. 13). Previously, continental areas (e.g., Lugano-Varese area, Badioto-Gardenese High and northern Julian Alps; Fig. 4, cols. 1, 7, 13; Fig. 5, cols. I, II, VII) quickly evolved into pelagic areas or into sites of very thick shallow-water sediments. This is the reason why in the shelf areas the La1 SB is often hard to recognize, it is masked by the drowning event.

In the western Dolomites, corresponding to the Badioto-Gardenese High, a marked instability along the margins of the subsiding blocks affected the growth of the Contrin carbonate platform (An4 HST) and is responsible for the great amount of breccias, slumpings and diagenetic debris-flows of the Moena Formation. In Lombardy, the Mt. Albiga Dolomite and the Mt. San Giorgio Dolomite seem to have undergone a similar event (De Zanche and Farabegoli, 1983; Farabegoli and De Zanche, 1984).

La1 transgression forced the carbonate platforms to aggrade and back-step (Lower Edifice, Tierfin Dolomitic Limestones). In unstable areas, these platforms seem to have been drowned before the MFS, whereas in less quickly subsiding areas they survived until the Chiesense Subzone. Due to the connection of a strong subsidence and a marked sea-level rise, maximum basin extent was reached during this time.

Also, the HST prograding carbonate buildups underwent strong subsidence and were essentially aggrading and only weakly prograding toward the basins. Together with the masking effect from the overlying prograding platforms, this is the reason why the La1 HST carbonate platforms are hardly recognizable in the field.

Although an Early Anisian volcanic phase has been suggested by De Zanche and Farabegoli (1988) on the basis of volcanic clasts included in breccias inside the Val Leogra Breccia or its lateral correspondents, the earliest well-documented volcanic events in the Southern Alps occurred during the An4-La1 time span (Figs. 4, 5). The most ancient, well-dated and widespread volcanics in Lombardy and the Dolomites (Prezzo Limestone, Bivera Formation) consist of distal ash-falls. The earliest subaerial intermediate lava-flows occurred on the Julian Alps (Cros, 1982; Obenholzner, 1991; Gianolla, 1992). During La1 time, basins were supplied with great amounts of acidic (rhyolitic and rhyodacitic) volcaniclastics (“pietra verde”), coeval to subaerial effusions within the southern Po Plain belt (Cassano et al., 1986), in Lombardy (Val Dezzo Volcanics: Jadoul and Rossi, 1982) and in the Tarvisio area (Rio Freddo Volcanics: Gianolla, 1992).

In the Latear area (SW Dolomites) anomalous behavior in the TST carbonate deposits (Lower Edifice) has been recognized indicating a decrease in accommodation space. This anomalous behavior, not verified in the coeval backstepping carbonate platforms throughout the Southern Alps, is probably due to the early effects on the sedimentary cover induced by the emplacement of the Predazzo intrusive bodies or to the local
effect of syndepositional strike-slip faults (Goldhammer et al., 1987; Goldhammer and Harris, 1989).

The La1 MFS is the turning point between the transgressive and the regressive phase of this T/R facies cycle. Due to the previous strong transgressive trend, shorelines and therefore the sources of siliciclastics, were moved landwards. In this interval, the paleogeographic setting was still characterized by a marked differentiation between platform and basinal areas, the latter also relatively deep.

The La2 SB (Recubariensis Subzone) is not well defined in the basins, because siliciclastic sources were distant; therefore the La2 LST deposits are relatively thin, whereas during the HST time carbonate platforms markedly prograded.

The La3 SB is characterized by a strong sea-level fall, as proved by the following LST deposits which suggest a marked northward progradation of terrigenous lobes, both siliciclastic and volcaniclastic (Zoppè Sandstones in the Dolomites). The La3 DS is also characterized by a strong, prevalently basic (andesitic-basaltic) volcanic activity, which produced a great amount of volcanics and volcaniclastics. Middle Triassic basic volcanism developed over a vast area, from Carnia to Adige Valley; however its climax occurred in the Dolomites (cf. Gaetani et al., 1980a; Bosellini et al., 1982; Castellarin et al., 1988; Sloman, 1989).

Figure 14 does not show a predominance of west-east structural trends over the north-south ones or vice versa. However, the analysis of basin and platform successions clearly proves that during the La2-La3 interval the whole basin started to undergo strong progradations both northward (siliciclastic shoreline) and southward (carbonate shoreline). Therefore, a double polarity, characterized by morphologic or structural highs placed to the north and to the south, seems to have taken place once again. Particularly the southern high (cf. “southern mobile belt” in Brusca et al., 1982), roughly east-west oriented and the main source of siliciclastic supply that filled the Triassic basins in the Southern Alps, is well documented. Figure 14 shows local anomalous behaviours concerning thicknesses (mainly volcanics) and therefore total subsidence (e.g., western Dolomites and Venetian Plain). Such anomalies could be partly related to both compressional and extensional effects of strike-slip tectonics (Blendinger, 1985; Doglioni, 1987, 1988) that is locally emphasized by volcanotectonics (Blendinger, 1985; De Zanche et al., 1993).

The Carl SB is related to a major sea-level fall (Biddle, 1984; Brandner, 1984, 1991; Haq et al., 1987). In the Southern Alps, it corresponds to the karstified top of the Esino Limestone 3 and San Salvatore Dolomite 3 in Lombardy and the Sciliar Dolomite 3 in the Dolomites and in Carnia (cf. Gaetani et al., 1992a; Jadoul and Gnaccolini, 1992; De Zanche et al., 1993). Development and depth of the karst network inside the Esino Limestone 3 suggest a sea-level fall of up to 60 m (Assereto et al., 1977b). In the basins, this SB is often characterized by major submarine erosions. The Carl LST consist of huge amounts of strongly prograding terrigenous-carbonate deposits. Therefore the Carl TLPC is a good candidate to be a maximum regression.

4—Late Longobardian-Tuvalian

The Carl TLPC defines the base of a new T/R facies cycle whose upper boundary corresponds to the Car4 TLPC. The peak transgression seems to correspond to the Car2 MFS (Aonoides Subzone). In spite of the previous strong siliciclastic supply, during the transgressive phase the basins were not filled and, at the same time, the Car1 HST carbonate platforms prograded only moderately. During the regressive phase, terrigenous and carbonate progradation completely filled the previous basinal areas. Therefore, peritidal, sabkha and continental facies became widespread throughout the Southern Alps.

During the Car1-Car2 interval, volcanic activity occurred mainly in the southwestern Southern Alps (Garzanti, 1985), although sporadic events in the Dolomites have been detected.

As far as variations in accommodation space in shelf areas can indicate, Figure 14 suggests that total subsidence values remained high, though not uniform. However, the west-east section clearly shows that the eastern Southern Alps were more subsident than the western. At the same time, the south-north section emphasizes the higher subsidence rate in the northern domain compared to the southern one. As is largely proved by the study of basinal terrigenous successions, the southern area acted as a structural and morphologic high, a source of siliciclastic supply (Figs. 5, 7).

Synsedimentary tectonics, prevalently of extensional type, were initially lighter and became stronger in the latest Julian-Tuvalian time when small pull-apart basins were formed in connection with strike-slip tectonics (Doglioni et al., 1991).

5—Tuvalian-Norian

As seen above, the Car4 TLPC corresponds to the lower boundary of a 2nd-order cycle, characterized by a change in structural trends. As a matter of fact, the shoreline and the platform-basin margins were shifted far to the east. This is the reason why in the Southern Alps it is no longer easy to check the variations of coastal onlap.

At first throughout the Southern Alps, uniform continental conditions were widespread. Only in eastern Carnia and in the Julian Alps were anoxic basins (Subbullatus Subzone, peak transgression) preserved. In our opinion, this is a major onlap event because the deposits of the Raibl Formation and its lateral equivalent extended over vast areas, including where erosion had exposed the crystalline basement (Fig. 5). Afterwards, in connection with an increase in accommodation space, carbonate tidal flat (inner peritidal platform) conditions were established (Bosellini, 1989).

During Norian time, due to a new extensional tectonic phase, extended intraplatform anoxic basins began to be defined and to subside (Jadoul, 1986; Ciarpica et al., 1987); their infilling was carbonate.

In this interval, the southern high was strongly retrenched as source area and became an accommodation area (Fig. 5). Only one case in which the southern high acted as a source is known; it concerns the Passo Buse Scure Breccia in the Recoaro region, consisting of a fluvial deposit mainly formed by metamorphic clasts, lying inside the Dolomia Principale. Moreover the discoveries of conspicuous and differentiated tetrapod footprints in the Dolomia Principale are increasing, suggesting the persistence of a southern emerged area.

A major sea-level fall seems to be placed close to the Norian-Rhaetian boundary. In the eastern Southern Alps, the maximum regression, which defines the top of this cycle, corresponds to the unconformity at the top of the Dolomia Principale.
In regard to the Rhaetian portion of the next 2nd-order cycle data in the Southern Alps are very poor and are restricted to Lombardy. The peak transgression probably lies in the Liassic stage.

CONCLUSIONS

—An approach integrating lithostratigraphy, biostratigraphy and sequence stratigraphy allowed the identification of 3rd-order DSs and of 2nd-order T/R facies cycles in the Triassic in the Southern Alps.

—The availability of a good biochronostratigraphic control is a primary condition for correlations. The ammonoid dating of the Anisian-Lower Carnian depositional sequences in the Southern Alps demonstrates the synchrony of the events throughout the basin.

—in spite of the intense synsedimentary extensional tectonism and the volcanism, that can alter stratral patterns, the eustatic signal is always recognizable.

—The 3rd-order DSs discussed in this paper are more numerous than in the cycle chart in Haq et al. (1987) and do not seem to be induced by local tectonics.

—The 3rd- and 2nd-order cycle setting is a good tool for a better understanding of the basin sedimentary history. An attempt to reconstruct the basin evolution was carried out also using diagrams of variations in accommodation space in shelf areas.

—The subdivision in physical chronostratigraphic units of the Triassic in the Southern Alps has made the recognition of the sedimentary evolution of the whole basin easier. It is possible to outline its evolution as follows:

1. (?)Latest Permian—early Anisian time (Sc1 to An2): terrigenous and terrigenous-carbonate shallow-water deposits were widespread in a prevalent ramp setting;

2. Bithynian—early Illyrian time (An2 to An4): earlier basins were formed and previous uniformity was broken by extensional tectonics. Horst-and-graben structures complicate the paleogeography. Continental deposits, carbonate platforms and basininal mixed sediments were variously stacked and interfingered. On the whole, subsidence was greater to the east;

3. Illyrian—Fassanian time (An4 to L1a1): widespread strong increase in subsidence. Huge carbonate buildups were separated by wide and relatively deep basins. Maximum extent of Triassic basinal deposits occurred near the end of this interval. Intermediate to acidic volcanic activity was strong;

4. latest Fassanian—Longobardian time (L1a to Car1): subsidence was still marked but not uniform; this may be owing to the combination of eustasy, compressional and extensional effects of strike-slip tectonics and local volcano-tectonic events. Prograding terrigenous fans document the migration of the southern shoreline. Basic volcanism reached its acme in the Dolomites;

5. late Longobardian—Tuvalian time (Car1 to Car4): basins were progressively filled by terrigenous supply and by strong carbonate platform progradation. Finally, the Southern Alps were largely occupied by continental facies and the shoreline was shifted eastward. In the western Southern Alps, intense Early Carnian intermediate volcanism is documented;

6. Tuvalian—Norian time (Car4 to Rh1): initially in the Southern Alps carbonate tidal flat conditions were widespread. Due to a new phase of extensional tectonics, Norian anoxic basins spread and sink.

7. Rhaetian time (Rh1 to Rh2): an important transgressive phase occurred, following in the Jurassic.

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REFERENCES


The sequence stratigraphic methodology can be readily applied to the cratonic basin-fill of the German Triassic System, which consists of shallow-marine to terrestrial mixed carbonate/siliciclastic rocks. The whole Triassic succession represents a second-order transgression/regression cycle, in which the continental redbeds of the Buntsandstein pass gradually upwards into Muschelkalk carbonates and evaporites and back into continental Keuper redbeds. Peak transgression occurred during the Late Muschelkalk (Ladinian). The Triassic cycle is built by at least 13 3rd-order depositional sequences, consisting of systems tracts and parasequences. Many bounding surfaces represent widely used marker beds, long used in classical lithostratigraphy.

Using a synthesis of outcrop and well-log data on stratal geometry, facies, cycle stacking patterns and paleogeography, a regional coastal onlap chart was constructed. Within the limitations of the presently available biostratigraphic data, the observed cycles appear to correlate fairly well to those in other areas, but include a number of additional sequences not included in the Haq et al. (1987) chart. Comparative analysis of regional onlap curves from different, globally separate Triassic basins, together with an improved biostratigraphy, will be necessary to relate the accommodation changes to eustatic versus tectonic controls and to produce a refined eustatic chart. The German Basin could provide a favorable calibration point for such an analysis.

REFERENCES

Fig. 1.—Regional sequence stratigraphy of the German Triassic System. Chronostratigraphy after Haq et al. (1987). Legend: * = additional sequence boundary, + = more pronounced, − = less pronounced sequence boundaries as compared to Haq et al. (1987).
ABSTRACT: During Triassic times, the Northern Calcareous Alps were situated at the northwestern margin of the Tethys ocean. Succeeding the Permian and Lower Triassic red bed stage, shallow-water carbonates were the main constituents of the Middle and Late Triassic strata, in addition to some mixed carbonate-clastic deposits and basin carbonates. Biostatigraphic data enables exact positioning of the depositional sequences. Despite many good correlations with other Alpine regions as well as with epicontinental Triassic units of central, northwestern and northern Europe, no correlative sequences exist in the Northern Calcareous Alps during late Anisian, late Ladinian, early Carnian and probably Norian times. Variations resulted from local (basin-wide) controls, mainly caused by diverging subsidence patterns. While second-order trends roughly correspond with other Triassic basins, peak transgressions and maximum regressions show some differences.

Separated by post depositional Alpine strike-slip and thrust tectonics from the stable European northern hinterland, definition of third-order sequence boundaries is often problematic in the Northern Calcareous Alps. Without unconformities, sequence stratigraphy sensu Vail and coworkers cannot be applied in a strict sense. Transgressive and maximum flooding surfaces comprised the best marker horizons for correlating sequences. A working procedure that incorporates at least parts of the genetic stratigraphic sequence model of Galloway (1989) proved extremely helpful in working with these successions.

INTRODUCTION AND REGIONAL SETTING

The Northern Calcareous Alps, situated between about the Swiss-Austrian border in the west and Vienna in the east, form part of the Alpine fold belt (Fig. 1). It covers an area of about 500 by 60 km and forms the main part of the Upper Austroalpine nappe complex, which comprises a sedimentary record from the Paleozoic (Grauwacken Zone) through the Permo-Mesozoic (Northern Calcareous Alps, built-up mainly by Middle and Upper Triassic platform carbonates) to the synalpidic Gosau sediments, deposited in basins during nappe transport.

During Triassic times, the Northern Calcareous Alps were situated at the northwestern margin of the Tethys ocean (Ziegler, 1988; Dercourt et al., 1993). This Triassic position, between epicontinental central and western Europe and more easterly and southerly Tethyan areas, make the Northern Calcareous Alps a key area for understanding the interrelationship between the Tethys ocean and the continent at the decisive time when Pangaea started to break apart. The exact location regarding Triassic strata of the Southern Alps and other Tethyan regions on one side, and central and western Europe on the other, is still questionable, as is the position of the Upper Austroalpine nappes in relation to the underlying Austroalpine nappe units.

The shelf area of the Northern Calcareous Alps was affected by a coarse clastic terrigenous input during the early Triassic (Scythian) and during short intervals in the Carnian (e.g., Raibl clastics). In strong contrast to the situation in the Southern Alps, Middle and Late Triassic clastics are extremely fine and were deposited in small, relatively shallow, intra-shelf (intra-platform) basins. Even during input of fine clastics, carbonate production flourished, forming huge carbonate platforms (we use the term “carbonate platform” according to Burchette and Wright, 1992, therefore a ramp is a particular type of platform). Only in the east, in the Hallstatt basin, were open marine conditions established during the early Middle Triassic, which continued throughout Triassic times.

STRATIGRAPHY AND THIRD-ORDER SEQUENCES

Although Triassic sediments, especially the carbonate platforms, were investigated from the last century onwards, no recent stratigraphic synthesis of the whole Triassic succession of the Northern Calcareous Alps is available. The abundance of local data, published in the last decades (see Tollmann, 1976, 1985), has not been followed up by concise stratigraphic reviews and summary papers, with the main exception of the comprehensive study by Schlager and Schönblüner (1974) concerning the entire Mesozoic sequence. Because of this absence of modern stratigraphic compilations, this sequence stratigraphic paper includes a concise review of depositional environments and biostatigraphic data (Fig. 2).

Applying the concept of sequence stratigraphy (e.g. Vail, 1987), several third-order depositional sequences have been identified in the Northern Calcareous Alps (Fig. 2). Up to now, only a few papers exist applying sequence stratigraphic procedures to parts of the succession (Bechstädt and Schweizer, 1991; Bechstädt et al., 1993; Rüffer, 1995). The sequences outlined in this paper are based mainly on our own field studies on late Anisian to Norian successions in the western part of the Northern Calcareous Alps, an area mainly lying west of the Salzach valley. Reinterpreted sections from other authors (Fruth and Scherreiks, 1982; Mostler and Scheuring, 1974; Stingl, 1987; Summesberger and Wagner, 1972; Urlichs, 1972) served as an additional database. In contrast to previous reports (Schlager and Schönblüner, 1974; Bechstädt and Mostler, 1976), the continuous Triassic deposition in the Northern Calcareous Alps was only weakly controlled by early (Triassic) tectonics. Therefore, this area is ideal for establishing sequence stratigraphic subdivisions.

Our application of sequence stratigraphy does not necessarily require the existence of an unconformity bounding two sequences. Subaerial exposures with a significant hiati are almost completely absent in Triassic units of the Northern Calcareous Alps. Being separated from its epicontinental hinterland by Alpic tectonics, only the “correlative conformities” were developed. We have interpreted sections in terms of lithofacies, environment and paleobathymetry. Regional controls on water depth have been eliminated, to produce a sequence stratigraphic model representing basinwide controls on sedimentation. Some
data on the problematic correlations with other regions, as well as an analysis of sequence controls are given, in an attempt to understand the sedimentary history and paleotectonic setting of the Northern Calcareous Alps and further to distinguish locally developed sequence stratigraphic events from regional or global ones.

**Clastic, Carbonate and Evaporite Ramp (Scythian)**

**Depositional environment.**—The analysis of sedimentary structures, lithofacies and paleocurrent directions of the Scythian Lower Buntsandstein sandstones indicates a braided river system with the main source areas of siliciclastics being in the north and probably to the west (Eisbacher, 1963; Stingl, 1987). Sporadic occurrence of gypsum, increased carbonate content, bimodal current directions and trace fossils suggest a marginal marine influence in the uppermost Lower Buntsandstein of the central and eastern part of the Northern Calcareous Alps (Stingl, 1987). Onlapping fluvial deposits of the Upper Buntsandstein transgression caused fluvially-dominated estuaries, changing later to clastic and clastic-carbonate deposits of shallow-marine to tidal flat environments (Werfen Formation). From early Induan to latest Scythian times, shallow-marine Werfen facies followed the prograding Tethys from the east and south (e.g., eastern Mediterranean and the Dolomites) to the west and north (e.g., western Northern Calcareous Alps). Therefore, the diachronous Werfen transgression was not related to any particular third-order sea-level rise, but to the (Neo-)Tethys rifting (Hirsch, 1992) or to a first and/or second-order transgression. In contrast to the eastern part of the Northern Calcareous Alps, where Werfen deposits covered the whole Scythian (Mostler and Rossner, 1984), marine deposition in the western part started only during Spithian times (late Scythian). Most of the late Scythian (? to early Anisian) Reichenhall Formation consists of unfossiliferous, sometimes bituminous carbonates with rare macrofauna in some beds (Schlager and Schöllnberger, 1974). In addition, the occurrence of gysiferous vuggy dolomites and the sulfur isotopic composition of interbedded anhydrites suggests a restricted marine depositional environment with episodic evaporitic conditions (Spötl and Burns, 1991).

**Sequences.**—In the western Northern Calcareous Alps, no sequence stratigraphy is feasible for the Griesbachian to Smithian substages of the Scythian. The sequence boundary S2 can be placed at the contact between Lower and Upper Buntsandstein, in correspondence to a sharp boundary between the marine influenced highstand S1 and the fluvial deposits of lowstand S2. Spithian transgression from east to west caused the development of estuaries which continued into the succeeding marine clastic-carbonates of the Werfen Formation, forming transgression S2, also recognized by Brandner (1984). Highstand S2 comprised the intertidal to supratidal, increasingly restricted and partly evaporitic Reichenhall Formation (late Scythian to questionable early Anisian).

**Biostratigraphy.**—Due to the lack of stratigraphically significant fossils, no biostratigraphic data have been obtained from the Alpine Buntsandstein (up to now no investigations on palynology were conducted). Therefore, the exact age of sequence boundary S2 is unclear, but it can be placed somewhere into the late Scythian. Using conodonts and other microfossils, the Werfen Formation of our study area has been dated as Spathian by Mostler and Rossner (1984). The top of the Reichenhall Formation has been placed by most authors at the Scythian/Anisian boundary, although an early Anisian age cannot be excluded (Brandner, 1984 mentioned dasyclads attributed to the Anisian, unfortunately without naming species).

**Carbonate Ramp Deposits (Anisian)**

**Depositional Environment.**—During Anisian times a wide shelf area with peritidal deposits existed, comprised of mainly pure calcareous homoclinal ramp deposits of the Steinalm Formation and strongly bioturbated shelf deposits of the Virgloria Formation, which were made up of carbonates with a low clastic influx (Fig. 3). In late Anisian time, the clastic input gradually...
vanished, resulting in a predominance of the Steinalm Formation with respect to the Virgloria Formation. Hardly any terrigenous clastic material reached the extensive ramp during the Illyrian. Compared to the Southern Alps, the succession is rather uniform, a distinction expressed in the different number of lithostratigraphic units in both areas.

The Steinalm Formation comprises inner to outer ramp deposits including shallow intrarramp basin sediments (Rüffer, 1995). Due to almost absent early diagenesis, along with the predominance of micritic facies in all depositional zones, an unstable muddy substrate evolved. Encrusters prevailed in shallow-marine environments, and organisms capable of building reefs were missing completely. Anisian “reefs” of the Mieming Range described by Miller (1965) and quoted later by Tollmann (1976), Flügel (1982) and Senowbari-Daryan (1993) are definitely mid- to inner ramp deposits with few microbial stromatolites. In the Northern Calcareous Alps, reef building organisms and also high-energy shoals did not exist during Anisian times (the first reefs may have started development in the late Avisianum zone of the latest Illyrian). No time equivalent counterparts for the Pelsonian (early Late Anisian) reefs of the Southern Alps (e.g. Fois and Gaetani, 1984; Senowbari-Daryan et al., 1993) existed in the Northern Calcareous Alps. Especially during the final stage of the homoclinal ramp, tempestites were intercalated into usually mud-supported carbonates.

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**Fig. 2.**—Triassic depositional environments and sequences in the western and central part of the Northern Calcareous Alps. Timescale after Gradstein et al. (1994).

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**Fig. 3.**—Middle Triassic depositional environments in the central western part of the Northern Calcareous Alps. Due to strong subsidence during latest Ladinian to early Carnian times, no sequence stratigraphy could be established for this part of the Northern Calcareous Alps.
Sequences.—The most striking features of the Anisian are third-order transgressive intervals, causing a ramp stacking with repetitions of depositional environments in the same systems tracts of different sequences. During the Anisian, rates of accommodation and accumulation were similar and only slightly prograding depositional environments can be recognized. High rates of sedimentation within the outer ramp during low sea level compensated for low rates during high sea level, whereas within the inner ramp rates of sedimentation were substantially higher during early highstands than during lowstands. Therefore, within a complete sequence of the homoclinal ramp, the average rate of carbonate production and sedimentation was almost independent of the depositional environment. Due to the low depositional relief, third-order sea-level fluctuations caused extended lateral shifts in facies, but did not change the fundamental mechanism of sediment production, reworking and transportation. This implies, that a certain sedimentary facies was not restricted to a particular systems tract, for instance crinoidal bioclastic wacke- to packstones were characteristic for transgressive inner ramp settings as well as for the mid-ramp during highstands.

Due to the absence of erosional surfaces and supratidal facies in most areas, often no distinct sequence boundaries nor lowstands were recognized. Only sequence boundary A4, at the base of a dolomite horizon, and the following lowstand, made up of microbial stromatolites both in inner and in mid-ramp settings, could be traced throughout the western part of the Northern Calcareous Alps.

Transgressions began with a well-marked transgressive surface consisting mainly of crinoidal wackestones. After a decrease in particles, late transgressions and early highstands comprised crinoids, fecal pellets, brachiopods, and especially within the maximum flooding interval microbes. During highstands, microbial packstones characterized mid-ramp positions and stromatolites the inner ramp settings (Fig. 4).

Biostratigraphy.—Whereas biostratigraphic data are not available for the early Anisian, the late Anisian depositional sequences are fairly datable by conodonts occurring within the transgressive systems tracts. Pelsonian transgression A4 is characterized by the conodonts Gondolella bifurcata and Nicoraella kockeli. Late Anisian systems tracts between transgressions A4 and A5 have been interpolated.

Rimmed Platforms and Basin to Platform Transitions (Latest Anisian to Early Carnian)

Depositional environment.—At approximately the Anisian/Ladinian boundary (top of Avisianum zone / base of Nevadites zone), the late Illyrian homoclinal ramp was drowned (Reifling event, Schlager and Schöllnberger, 1974). In local areas, carbonate production recovered following a short initial lag phase (incipient drowning) and distally steepened ramps evolved gradually, but quickly from the Anisian homoclinal ramp. Thus, on the one hand, the gradual deepening of the outer ramp led to the formation of a deep ramp slope, and on the other hand, high-energy shoals and buildups separated the increasingly restricted lagoons from the more open-marine mid-ramp. In the Northern Calcareous Alps, no indication of any synsedimentary tectonics breaking up Anisian platforms has been recognized. Late Anisian and Fassanian rates of subsidence had the same order of magnitude (20–30 mm/ky, according to almost all existing time scales).

A distinct increase in subsidence during late Fassanian to middle Longobardian times (some 50 to 300 mm/ka, Rüffer, 1995) caused the transformation from distally steepened ramps to rimmed platforms. Progression and widening of the Tethyan seaway characterized (late) Ladinian to early Carnian times (Hirsch, 1992). Sedimentation in the Northern Calcareous Alps was controlled mainly by the instability of the Tethys realm, causing rates of subsidence several times larger than during other Triassic stages. Incessant platform development with aggradation and progradation characterized this interval, resulting in increasingly larger platforms and in over 1000-m-thick platform deposits (Fig. 3). No basal intercalations separated the platform succession, as reported from the Southern Alps (e.g., Sciliar Dolomite 1 to 3, De Zanche et al., 1993).

Sequences.—The late Anisian to early Ladinian transgression A5, the most significant event during Middle Triassic times, caused the transformation from homoclinal to distally steepened ramps. This transgression is mostly characterized by intra-platform basin and basin margin deposits (Reifling Formation), implying, that the Anisian carbonate ramps experienced a drowning event (Fig. 4, section Laliderer Tal). In some areas, carbonate production recovered after incipient drowning, and platform sedimentation continued without basin intercalations during Middle Triassic times (Fig. 4, section Laliderer Wände). In these cases, transgression A5 was characterized by extensive migration of ammonoids onto the platform, by pelmicritic deposits during early transgression and by the evolution of high-energy shoals and reefs (Wetterstein Formation). In the few areas where late Anisian deposits are represented by deeper marine deposits, bituminous nodular marls (lowstand A5) were succeeded by cherty limestones with pelagic ammonoids, conodonts and bivalves (Großreifling area in the central-eastern Northern Calcareous Alps; Summesberger and Wagner, 1972).

The depositional sequence L1 is not recognized throughout the Northern Calcareous Alps. Situated in the second-order peak transgression interval, no distinct lowstand was developed (Fig. 4). Whereas, the recognition of the early Longobardian lowstand L2 is somewhat problematic at the platform margin, lowstand wedges are characteristic of basin to lower slope areas. Compared to the Anisian, late Ladinian and early Carnian transgressions had a minor impact on sedimentation, and sequences are difficult to recognize. The main information for late Longobardian to early Julian sequences come from distributional patterns of the terrigenous influx, but up to now no well founded sequence stratigraphy has been established. An analysis of the variation in accommodation space of lagoonal cycles may close this gap. A basinal section in the eastern part of the Northern Calcareous Alps, published by Mostler and Scheuring (1974), allows a preliminary sequence stratigraphy for this interval (Fig. 5). A third-order sea-level rise for at least the upper 140 m (representing a few hundred thousand years) of the Wetterstein Formation can be assumed from high-frequency cycles in lagoonal deposits (Zeeh, 1994).

Biostratigraphy.—Due to the widespread distribution of conodonts, biostratigraphic control is rather good in this interval. Data including other microfossils and ammonoids have been published by many authors (Bechstädt and Mostler, 1974, 1976; Donofrio et al., 1980; Krystyn, 1978; Mandl, 1984). Conodont
Mixed Carbonate-Clastic Shelf Deposits (Carnian)

Depositional environment.—In Middle Carnian times (Aonoides subzone), terrigenous clastics of the Raibl Group abruptly covered carbonate platforms as well as carbonate and shale basins. The Raibl Group consists of three repetitive carbonate-clastic sequences, clearly separated into carbonate and clastic intervals (Fig. 6). Each of these intervals (often even single marker beds) can be correlated for hundreds of kilometres.

Within the western Northern Calcareous Alps, each carbonate interval ranges between 30 and 80 m in thickness, with clastic intervals between 20 and 40 m thick (Bechstädt and Schweizer, 1991). The carbonate intervals mainly represent restricted marginal marine, tidal and evaporitic environments, while the clastic intervals were primarily formed on the inner and outer shelf.

Fig. 4.—Middle Triassic platform (“Laliderer Wände”, Karwendel) and basin section (Laliderer Tal”, Karwendel).
This depositional model is in contrast to many other carbonate-clastic depositional settings, where clastic intervals represent the sea-level lowstands and carbonates the highstands. As Bechstädt and Schweizer (1991) illustrated, the western Northern Calcareous Alps were influenced by only one deltaic point source. During sea-level lowstands, this delta was situated near to or at the shelf margin, allowing clastics to bypass the sedimentation area of the Raibl Group. A subsequent sea-level rise caused a distinct northward delta retreat. The earlier deposited delta sands were reworked and transported along the shore, taken up by westerly directed, shore-parallel currents and strongly affected by the fourth-order cyclicity, favoring sand waves in the upper (shallower) parts of these fourth-order cycles.

In summary, each of the three Raibl cycles corresponds to a third-order sequence consisting of 50- to 100-m-thick carbonate-clastic depositional sequences (C2 to C4 in Fig. 6).

**Biostratigraphy.** Within the study area, the Raibl Group is nearly void of index fossils, probably because of prevalent re-
triassic sequence stratigraphy in the western part of the northern calcareous Alps

restricted conditions within the area of deposition. As relevant biostratigraphic data only comes from outside the Northern Calcareous Alps or from basinal sections in the eastern Northern Calcareous Alps (cf. Fig. 5), the exact ages of the depositional systems remain unclear. The cephalopod *Carnites floridus*, indicative of the *Aon, Aonoides* and *Austriacum* subzones (Krytyn, 1978), has been found in the first and second clastic interval (Bechstädt and Schweizer, 1991). Mandl (1984) dated conodonts in carbonates overlying the upper shale horizon (Tuvalian 1–2) and Krytyn (pers. commun., 1992) put the base of the first Raibl shale within the Aonoides subzone.

**Shallow-Marine Carbonates (Late Carnian to Early Rhaetian)**

*Depositional environment.*—In the western Northern Calcareous Alps, the Raibl Group was capped by a continuous record of about 2000- to 2500-m-thick shallow-marine sediments of mainly Norian age (Hauptdolomit Formation), ranging in facies from oolitic and bioclastic mounds of a barrier bar complex to lagoonal and tidal flat deposits (Fruth and Scherreiks, 1982). Towards the east and south, the Hauptdolomit Formation is overlaid and replaced by platform margin deposits of the Dachstein Formation, for example in the Lofer area, where Fischer (1964) established the relationship between high-frequency cyclicity and fluctuations in eustatic sea level. Dachstein facies range from slope through central platform and back margin (Fruth and Scherreiks, 1982). In the southern part of the central and eastern Northern Calcareous Alps, basinal facies developed adjacent to carbonate platforms. Reijmer et al. (1991) analyzed calciturbidites from this transition to investigate the record of sea-level fluctuation in deeper water.

*Sequences.*—Defining of Norian sequence boundaries are a problem throughout the Alps. In the over 2000-m Hauptdolomit Formation of the western part of the Northern Calcareous Alps and also in the less thick Dolomia Principale of the Southern Alps, which spans about 8 to 10 my comprising tidal flat and lagoonal successions, only two or three sequences were recognized so far. In the Bergamasc Alps of Northern Italy, Jadoul et al. (1994) have observed five Norian sequences in carbonate successions of platform and intra-platform basin facies (including the lower part of the “Rhaetian facies”).

Marking the transition between the underlying Raibl Group and the Hauptdolomit Formation, a matrix-rich breccia characterized the locally transgressive surface C5 (Bechstädt and Schweizer, 1991). Fruth and Scherreiks (1982) supposed a nearshore environment for the breccia indicating longer periods of reworking during a transgression (the base of conglomerates in a similar lithostratigraphic position in the Southern Alps were interpreted by De Zanche et al., 1993 as a sequence boundary). In a more landward position, towards the north and west, transgressive surface C5 is placed between the evaporitic uppermost Raibl Group and the peritidal Hauptdolomit facies.

The most striking feature of the Norian Hauptdolomit is a clearly definable maximum flooding interval in the middle part of the formation (“Nautica Beds”), which is also used for separating the Lower from the Upper Hauptdolomit Member (Fig. 7). Intercalating peritidal microbial mudstones and stromatolites, this gastropod-bearing maximum flooding interval comprises inter- to subtidal bioclastic mud- to packstones with a remarkably high noncarbonate content (Fruth and Scherreiks, 1982). Another maximum flooding interval could be established in the early Lower Hauptdolomit Member, which is used by some authors to subdivide the Lower Hauptdolomit Member into two members (Lower and Middle Hauptdolomit Member). The cause of this wide ranging transgression starting at the diachronous base of the calcareous Plattenkalk Member and continuing to the Rhaetian Kössen Formation, is still disputed (Marcoux et al., 1993).

The relationship between systems tracts and noncarbonate content show the same trend as in the Carnian, which implies that the highest rate of terrigenous input occurred during maximum flooding intervals. However, for Norian time it is still unknown if pollution, with fine terrigenous clastics, controlled the depositional facies and the bathymetry or if relative sea-level fluctuations primarily controlled deposition of the shales, as in the Carnian Raibl Group.

*Biostratigraphy.*—No biostratigraphic data are available from the lagoonal and tidal flat deposits of the Hauptdolomit. The well-dated overlying Kössen Formation can be used for dating the upper limit, but the ages of the maximum flooding surfaces remain unclear. This problem may be solved by sequence stratigraphic interpretation of basin sections in the eastern Northern Calcareous Alps.

**Carbonate Platforms and Mixed Cretaceous-Carbonate Intraplatform Basins (Rhaetian)**

*Depositional Environment.*—In contrast to the uniform Norian deposits, different carbonate platforms and mixed carbonate-clastic basins characterized the Rhaetian. The platform carbonates of the Dachstein Formation were persistent since Norian times, while platform margin carbonates in northern areas evolved during Rhaetian times (Oberrhätirrfalkalk or Steinplattekalk). Piller (1981), and more recently Stanton and Flügel (1989), have described this platform, which passes into an intraplatform basin in the north (Kössen Formation). This basin comprised mixed carbonate-clastics deposits, which can be used for biostratigraphy as well as for sequence stratigraphy. Terrigenous clastic input originated from northern to northwestern areas and decreased towards the south.

*Sequences.*—The sequence stratigraphy is based on basin deposits of the Kössen Formation, due to its biostratigraphic valuable fauna. A landward section (“Weißbloferbach”), with a continuous mixed carbonate-clastic succession, published by Urluchs (1972), shows a strong relationship between faunal diversity and systems tracts. Both, the highest diversity of species and the maximum quantity of ostracodes occurred during the maximum flooding interval in the uppermost *Suessi* zone (early Rhaetian). In respect to the sequence stratigraphic analysis of the basin sediments, the Steinplatte reef platform, was built during a highstand. Intense vadose diagenesis suggests a sea-level fall in the “capping facies” of the uppermost platform (Stanton and Flügel, 1989).
FIG. 7.—Norian inner platform section (Lechtal Alps), with sequence stratigraphic interpretation based on the data from Fruth and Scherreiks (1982). Due to intense dolomitization and the lack of detailed investigations, two alternate depositional sequence models are indicated for the Northern Calcareous Alps, both in terms of regression/transgression trends.

Suessi and Marshi zone lies in the upper part of the Kössen Formation (Mostler et al., 1978). Conodonts and ammonoids in the Kössen Formation underlying the Steinplatte platform (Kachroo, 1989) show that the platform carbonates were deposited contemporary to the upper part of the Kössen Formation within the Marshi zone (Middle to Late Rhaetian). No data are available for the Triassic/Jurassic boundary, as a depositional and/or erosional hiatus characterizes this interval in the Northern Calcareous Alps.

LOCAL, REGIONAL AND GLOBAL DEPOSITIONAL SEQUENCES

Comparisons with other European basins are essential in the recognition of local versus regional controls of third-order sequences. Triassic sequences of the Northern Calcareous Alps correspond reasonable well with depositional sequences of the other Alpine regions as well as the epicontinental seas of central, northwestern and northern Europe. However, a few sequences do not correlate with other European basins (cf. Fig. 2).

1. Sequence A4 is interpreted as the result of very low rates of subsidence during the late Anisian, leading to restricted conditions (this sequence is also recognized in those parts of the Southern Alps, which were affected only by minor synsedimentary tectonics, M. Gaetani, pers. commun., 1994).

2. Mid-Longobardian sequence L3 results from increased rates of subsidence; therefore, this sequence is also restricted to the Northern Calcareous Alps.

3. Early Carnian C2 sequence comprises the uppermost Wetterstein platform with its local subaerial emersions and the early transgressive Raibl shales. With one exception (western Southern Alps, Gaetani et al., this volume), no other sequences have been recognized in European basins, which could be correlated. The principle controls of the C2 sequence are still unknown. The facies development of this sequence as well as of its stacked fourth-order cycles is almost identical to the other two Raibl sequences (C3, C4). Probably, strong subsidence of the investigated area governed the formation of these three Raibl cycles.

Most of the other depositional sequences can be correlated with other Alpine regions and the central and northern European basins. Some of them reflect sea-level fluctuations active only in the (north)western Tethys area, while a few might reflect a truly global control (for the Scythian to Carnian time interval, see Rüffer and Zühlke, 1995). However, during latest Ladinian to earliest Carnian times, tectonic subsidence overprinted regional and/or global sea-level fluctuations. The widely distributed latest Ladinian sequence boundary, which was reported from different tectonic settings (e.g., the Southern Alps, the Arabian peninsula, the Candian Arctic, and probably the Himalayan; M. Gaetani, pers. commun., 1994), cannot be recognized in the Northern Calcareous Alps. Due to the lack of biostratigraphic data in some intervals, especially the Norian, sequence correlation is somewhat questionable.

SECOND-ORDER SEQUENCES

Schlager and Schöllnberger (1974) originally subdivided the Triassic into four depositional units, bounded by four “stratigraphic turning points” in the late Scythian, late Anisian, early Carnian and at the Triassic/Jurassic boundary. Based on this still valid approach of recognizing and interpreting the main phases of sedimentary evolution, Brandner (1984) proposed a
similar division with cycle boundaries in the late Anisian, the early Carnian and at the Norian/Rhaetian boundary. However, these earlier studies, did not apply the concepts of sequence stratigraphy.

The first marine ingestions reached the western part of the Northern Calcareous Alps in late Scythian time. After a short-termed third-order regression, the late Scythian to early Anisian evaporites and carbonates marked the onset of continuous marine sedimentation, followed by Anisian homoclinal ramps and early Ladinian distally steepened ramps. All of this development belonged to the second-order transgression T2, which was also characterized by a decreasing input of terrigenous clastics (Fig. 8). The second-order peak transgression at the Fassanian (Curtinii zone/Longobardian (Gredleri zone) boundary) corresponds with the German Basin (Aigner and Bachmann, 1992) and the boreal Barents Sea but not with the Southern Alps, where the peak transgression was already achieved during the Fassanian (latest Nevadites zone; Fig. 8).

Due to increasing rates of subsidence during the second-order regression R2, definition of the maximum regression is somewhat problematic in the western part of the Northern Calcareous Alps. The only interval with emersions and subaerial hiatus can be found in the uppermost few hundred metres of the Wetterstein platforms (Bechstädt, 1975; Bechstädt and Döhler-Hirner, 1983; Zeeh, 1994), dated as Carnian in age from late Longobardian conodonts in the underlying basin deposits and dasy-clad algae from the inner platform (Poikiloporella duplicata instead of Diplopora annulata in underlying successions). Therefore, the maximum regression R2 can be dated as early Carnian, which is definitely younger than the concurrent maximum regression reported from the Southern Alps and the Barents Sea (Fig. 8). The pelagic deposits of the eastern part of the Northern Calcareous Alps help to better define the maximum regression R2. Assuming that the only shale interval in the classic Ladinian to Carnian “Zonaukogel” section (Schlager and Schöllnberger, 1974) reflected a maximum regression, the second-order cycle boundary can be attributed to the early Aon subzone.

The transgressive trend during Raibl deposition, constrained by facies analysis and fourth-order stacking patterns (Bechstädt and Schweizer, 1991), fully matches the second-order transgressive peak for late Carnian time reported from the Southern Alps and the Barents Sea (Fig. 8). Additional evidence for a second-order transgressive trend and a maximum flooding interval during mid to late Carnian times is the absence of emersions within the fourth-order cycles. Fourth-order cycles can clearly be delineated in the Ladinian to Norian carbonates, but clear evidence of emersion and karstification features can only be found in cycles from the Ladinian to early Carnian Wetterstein successions (Bechstädt, 1975; Bechstädt and Döhler-Hirner, 1983; Zeeh, 1994) and in the Norian formations (Fischer’s, 1964 Lofer cycles).

The Raibl Group was capped by latest Carnian to Norian lagoonal and tidal flat deposits (Hauptdolomit and Dachstein Formation), belonging to the second-order R3 regressive cycle. No distinct major regression can be concluded from the late Norian deposits of the Northern Calcareous Alps. Subsequently, the interval starting at about the Norian/Rhaetian boundary comprised a (major) transgressive succession continuing into the Liassic.

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**Fig. 8.**—Comparison of Triassic second-order sequences in the Northern Calcareous Alps with data from the Southern Alps and the Arctic Sea. According to Gaetani et al. (this volume) the maximum regression in late Triassic (R3/T4) time of the Southern Alps is more in agreement with the Northern Calcareous Alps.
CONCLUSIONS

This study deals with a marine succession from a depositional basin with moderate to high rates of subsidence, separated by postdepositional strike-slip and thrust tectonics from the stable European northern hinterland. Due to this, (a) shorelines as well as distinct shelf breaks were outside the area, and therefore coastal onlap relations could not be used for defining sequences; (b) no major erosional hiatus is known from the whole succession, emersions are solely related with tops of high-frequency cycles, which are found during second-order regressive intervals; (c) deposition mostly remained in subtidal conditions, rates of subsidence matched or even exceeded rates of sea-level fall; and (d) a continuous regressive trend was frequently formed during highstand and lowstand systems tracts. Definition of the third-order sequence boundaries is therefore often difficult. In the depositional sequence model (e.g., Duval et al., 1992), the second-order transgressions and regressions are bounded by a major third-order maximum flooding surface and a third-order transgressive surface following a major third-order sequence boundary. Due to this, third-order sequence boundaries have only a minor significance for defining second-order cycles.

Working in an area without distinct relief but with strong subsidence, the application of depositional sequence stratigraphy needs some modifications, especially because sequence boundaries are problematic to define. In such areas, without being able to correlate “conformities” with unconformities in more proximal areas, sequence stratigraphy cannot be applied in a strict sense. The best discernible, datable, traceable and correlative features were the transgressive surfaces and the maximum flooding surfaces. A working concept that incorporates at least parts of the genetic stratigraphic sequence model of Galloway (1989) is extremely helpful to deal with such successions. Maximum flooding surfaces (taken as sequence boundaries in Galloway’s, 1989 stratigraphic sequence model) are always recognizable in the Northern Calcareous Alps, in contrast to the “correlative conformities”. Moreover, the maximum flooding intervals contain the best biostratigraphic information. This allows a good biostratigraphic dating of the stratigraphic sequence boundaries, which can be much better defined than the depositional sequence boundaries.

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TRIASSIC SEQUENCE STRATIGRAPHY IN THE WESTERN PART OF THE NORTHERN CALCAREOUS ALPS 761
CENOZOIC ERA

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A new geomagnetic polarity time-scale for the late Cretaceous and Cenozoic (Cande and Kent, 1992: CK92) was based on an analysis of magnetic anomaly profiles from the world’s ocean basins. It is the first time since Heirtzler et al. (1968) published their time-scale that the relative widths of the magnetic polarity intervals for this entire interval have been systematically determined from magnetic profiles. A composite geomagnetic polarity sequence was derived based primarily on data from the south Atlantic where anomaly spacings were constrained by a combination of 9 finite rotation poles and averages of 61 stacked profiles distributed over the 9 finite rotation pole intervals. Fine scale information was derived from magnetic profiles on faster spreading ridges in the Pacific and Indian Oceans and inserted into the south Atlantic sequence. Based on the assumption that spreading rates in the south Atlantic were smoothly varying but not necessarily constant, a time-scale was generated using a spline function to fit a set of 9 age calibration points plus the zero-age ridge axis to the composite polarity sequence. The selected tiepoints (see also Berggren et al., 1992) reflect a preference for those data which can be tied to the magnetic anomaly sequence via marine magnetostratigraphic correlations and constraints from biostratigraphic correlation of sediments overlying oceanic basement.

The new time-scale has several significant differences from previous time-scales. For example, Chron Cn5n is ~0.5 my older and Chrons C9 through C24 are 2–3 my younger than in the chronologies of Berggren et al. (1985) and Harland et al. (1990). Many additional anomalies that may represent reversals of the global geomagnetic field were also identified, for example, between Anomalies 3A and 4A. On the other hand, an essentially continuous pattern of small scale anomalies or tiny wiggles was documented between Anomalies 24 and 27 that appear to be an “earth-filtered” record of short period (2 to 20 ky) intensity variations of the dipole field. This type of dipole field behavior, previously recognized within Anomaly 5 and between Anomalies 12 and 13, may have characterized the geomagnetic dynamo throughout the Cenozoic (Cande and Kent, 1992b).

Geomorphic polarity chron nomenclature is based on the long-standing numbering scheme (sometimes with lettered additions) for magnetic lineations in which prominent anomalies (generally positive and corresponding to predominantly normal polarity) have been designated from 1 (youngest) to 34 over the Cenozoic and to the younger end of the Cretaceous Quiet Zone or Cretaceous Long Normal. A chron corresponds to the interval from the younger boundary of the eponymous anomaly to the younger boundary of the preceding anomaly and has the prefix C. (e.g., Chron C3A). However, each of these chron is usually divided into the two constituent intervals of predominantly normal and reversed polarity which are designated by adding to the chron name the suffix n for normal polarity and r for the preceding reversed polarity interval (e.g., Chron C3An and Chron C3Ar). When these polarity chron are further subdivided into shorter polarity intervals they are referred as subchrons and identified by appending, from youngest to oldest, a .1, .2, etc. to the polarity chron name, and adding an n for a normal polarity interval or an r for a reversed polarity interval (e.g., Chron C3An.1r). Finally, the designation .1, .2, etc. is used following a chron or subchron name to denote apparently very short polarity intervals corresponding to the tiny wiggles which, upon calibration, convert to durations of less than 30 ky. In view of their uncertain origin, these globally mapped geomagnetic features are referred to as cryptochrons and have not been included in any of these charts.

Cande and Kent (1995) generated an adjusted geomagnetic reversal chronology for the late Cretaceous and Cenozoic using the same tiepoints and anomaly distances as CK92 except in two instances: a) a consensus age of 65 Ma (rather than 66 Ma in CK92) was used for the Cretaceous/Paleocene boundary in Chron C29c; and b) a tiepoint at 5.23 Ma for the older boundary of Subchron C3n.4n was used rather than 2.60 Ma for the younger boundary of Chron C2An. The latter modification allowed the direct incorporation of the astrochronologically calibrated polarity time scale for practically all of the Pleistocene and the Piocene that was developed by Shackleton et al. (1990) and Hilgen (1991) and thereby avoided the promulgation of separate timescales over this interval (see discussion in Berggren et al., 1995a). The revised geomagnetic polarity time scale (CK92/95, or sometimes just CK95) was used as the chronological framework for the integrated Cenozoic time scale of Berggren et al. (1995b).

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PLANKTONIC FORAMINIFERA

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Calibration of planktonic foraminiferal datum events/zonal boundaries to the GPTS has been made essentially using the same DSDP and ODP sites/holes as reviewed below by Aubry. All datum events compiled in Berggren et al. (1985) have been reviewed and updated as well as all datum events identified and correlated to magnetostratigraphy in the 10 year interim to 1995. A major advance has been made in the compilation, and calibration, of Pliocene-Pleistocene datum events. The Achilles heel of this scheme remains, as for the calcareous nannoplankton, the middle Eocene, where lack of continuous, temporally complete stratigraphic sections precludes accurate magnetostratigraphic correlations. The Paleogene planktonic foraminiferal zonation follows that established by Berggren and Miller (1988); the Miocene zonal scheme is taken from Berggren et al. (1995), and the Pliocene Pleistocene is taken from Berggren et al. (1995b).

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CALCAREOUS NANNOFOSILS

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The magnetobiostratigraphic/chronologic framework presented here draws from the recent revision to the Cenozoic time scale by Berggren et al. (1995a) for the Paleocene to Miocene and by Berggren et al. (1995b) for the Pliocene and Pleistocene. Progress in Cenozoic magnetobiostratigraphic correlations has been uneven since the publication of the work of Berggren et al. (1985a,b), and the number of sections reliable for magnetobiochronologic calibration remains very small, even for the Neogene. This is largely due to the lack of quality of magnetobiostratigraphic correlations in many sections, due to poor recovery in some, and to the ambiguity or insufficient quality of the magnetic polarity signal in others. We are just recognizing that the deep sea record is less complete than once thought and that unconformities may account for discrepancies previously attributed to diachrony (Aubry, 1995). As a consequence, temporal interpretation of stratigraphic sections must be conducted to establish that the sections used to calibrate datums to magnetochronology are continuous.

The calibration of Paleogene datums herein relies primarily on DSDP Holes 384 (Aubry in Berggren et al., 1995a), 527 (Shackleton et al., 1984) and 577 (Monecchi et al., 1985) for the Paleocene; on DSDP Holes 516 (Berggren et al., 1983a; Wei and Wise, 1989), 522,523 (Poore et al., 1982,1983), 527,528 (Shackleton et al., 1984), 530,531,532,533 (Steinmetz and Stradner, 1984), 550 (Aubry et al. 1995; Berggren and Aubry, 1995), ODP Holes 689,690B (Wei and Wise, 1990), 703A (Wei, 1991), 744 (Wei and Thierstein, 1991), 748 Aubry, 1992), and on the Contessa Highway, Massignano and Bottacino sections (Napoleone et al., 1983; Coccioni et al., 1988; Monecchi and Thierstein, 1995; Nocci et al., 1983; Premoli Silva et al., 1988) for the Eocene; on DSDP Holes 516F (Berggren et al., 1983a; Wei and Wise,1989), 522, (Poore et al.,1982), 558, 563 (Miller et al., 1985), ODP Hole 703A (Wei, 1991), 774A (Wei and Thierstein, 1991), 748A (Aubry,1992; Wei et al., 1992), and the Massignano section (Premoli Silva et al., 1988) for the Oligocene; and on DSDP Holes 558, 563 (Miller et al., 1985; Wright and Miller, 1993), 608 (Gartner,1992; Olafsson,1991), 516 (Berggren et al.,1983b), ODP sites 844, 845, 846, 852 and 853 (Raffi and Flores, 1995; Raffi et al., 1995) and the Buff Bay section, Jamaica (Aubry,1993 Berggren,1993 and Miller et al., 1994) for the Miocene. The reader is referred to Berggren et al., (1995a,b) for the details on magnetobiostratigraphic correlations in these sections. The calibration of Pliocene and Pleistocene calcareous nannofossil datums in Berggren et al. (1995b) is based on the studies of Backman and Shackleton (1983), Backman and Pestiaux (1987), Berggren et al. (1983). It appears there remain two main problematic stratigraphic intervals. Middle Eocene datums are poorly tied to the magnetic reversals event due to the lack of continuously recovered and (temporally) complete sections. Upper middle and lower upper Miocene datums (NN7-NN10 zonal interval) are unsatisfactorily tied to the magnetic polarity pattern because of unprecedented inconsistent correlations between calcareous microfossil (calcareous nannofossil and planktonic foraminifera) datums and magnetozones in different sections (see also Aubry, 1997). For this reason, two sets of magnetobiostratigraphic correlations are given for the NN7-NN10 zonal interval. Oligocene diachrony between high and mid-low latitudes is now well-established as a result of drilling in the Southem Ocean, and this is reflected in the magnetobiostratigraphic correlations as well.

REFERENCES CITED CAN BE FOUND IN


DINOFLAGELATES

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The diversity of Tertiary dinoflagellates makes them ideal zonation microfossils for most marine deposits: the one exception is in abyssal sediments where the organic-walled species are rare. The increasing climatic differentiation between tropical and polar regions during this time, however, is reflected in the increasing provinciality of the assemblages. Consequently, any plots of the first appearance datums (FADs) and the last appearance datums (LADs) must include information on the source.

The most detailed studies of Tertiary dinoflagellate assemblages are based on the type sections of Europe, where there is calibration with the foraminiferal and nannofossil zonations. This is especially true of the Paleogene. The Miocene dinoflagellate assemblages from the type sections, however, are poorly preserved when compared to other regions. For this reason, much of the data for this epoch has been derived from sections in the Salisbury embayment, eastern United
States of America. The Pliocene-Pleistocene records are primarily European and North Atlantic.

Our compilation has benefited from several comprehensive reviews of Tertiary biostratigraphy. These include Costa and Manum (1988), Powell (1992), Stover et al. (1996), Williams and Bujak (1985) and Williams et al. (1993), Costa and Manum (1988) and Powell (1992), published Paleocene-Miocene zonations for northwest Europe, with reference sections from surface and subsurface locations in Denmark (especially Jutland), southern England (primarily the Isle of Wight), France, Belgium, Germany, the North Sea and the North Atlantic Basin (Rockall Plateau and Bay of Biscay).

Many of the horizons plotted for the individual epochs are based on original research. In the Paleocene, a few of the FADs and LADs are based on the El Haria section, Tunisia, where there is continuous deposition across the Cretaceous/Tertiary boundary. Brinkhuis and Leereveld (1988) and Brinkhuis and Zachariasse (1988) describe the dinoflagellate assemblages from this section and correlated them with the planktonic foraminiferal zonation of Blow (1969) and the calcareous nannofossil zonation of Martini (1971). Northwest European Paleocene assemblages have been described by Hansen (1977, 1979); Heilmann-Clausen (1985, 1988); Huitberg (1986) and Powell et al. (1996).

The Eocene FADs and LADs for southern England are derived from Bujak et al. (1980), de Coninck (1990), plus personal knowledge of the Hampshire Basin sequences. Ranges of dinoflagellates from southern European sections are from Brinkhuis and Biffi (1992). This paper fills a gap in our knowledge of Priabonian assemblages.

The major source of Oligocene data has been Stover and Hardenbol (1994), Benedek and Müller (1974) and Brinkhuis et al. (1992). Stover and Hardenbol (1994) studied the dinoflagellates from the type and other sections of the Boom Clay of Belgium. The Boom Clay provides the lithostratigraphic basis for the lower Oligocene Rupelian Stage. Control in the late Oligocene is also based on Brinkhuis et al. (1992) who studied sections from the Piedmont and Marche basins in Italy. This paper also provided control for the early Miocene.

Miocene dinoflagellate FADs and LADs are based primarily on de Verteuil and Norris (1992, 1994, 1996). De Verteuil studied the diverse assemblages from the Chesapeake Group of the Salisbury Embayment, a basin occupying the coastal areas of New Jersey, Delaware, Maryland and Virginia and extending out into the North Atlantic. The surface and subsurface sections are keyed to the planktonic foraminiferal zonation and the calcareous nannofossil zonation. Zevenboom (1995) examined Oligocene-Miocene surface sections of central and northern Italy and wells from the Netherlands. Other important papers utilized were Powell (1986a, 1986b, 1986c).

The former paucity of data on Pliocene dinoflagellates is being rectified through such studies as Head (1992, 1994), Head et al. (1989a), de Vernal and Mudie (1989b, 1992) and Mudie et al. (1990). These studies have been the basis for the plots of FADs and LADs.

**SELECTED REFERENCES**


**OSTRACODES**

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During the last two decades, numerous detailed studies undertaken on Cenozoic European ostracode faunas have provided a good understanding of the stratigraphical distribution of a great number of species. In many cases planktonic foraminifera, nannoplankton and larger foraminifera (Nummulites, Alveolina) zones have been proposed.

**Paleogene**

For the Paleogene of northwestern Europe, the most comprehensive study can be found in the works of Keen (1977, 1978). This author proposes a 14-fold ostracode zonation for the marine environment based on the distribution in the Paris Basin, Belgium and England. Each zone is tentatively correlated with planktonic foraminifera and nannoplankton zones often through Nummulite zones. This author also proposes a brackish and a freshwater ostracode zonation.

For the southern North Sea Basin, additional information is provided on the stratigraphic value of Oligocene ostracodes by the works of Gramann and Spiegl (1986), Uffenorde (1986); and Uffenorde et al. (1979) who proposed bio-ecosтратigraphical zonations.

For southern Europe, our data essentially come from the Aquitaine Basin and the Pyrenees (Ducasse, 1969; Tambareau, 1972; Ducasse et al., 1985). For the Pyrenees, good correlations have been established with larger foraminifera.

**Neogene**

Data on Neogene ostracodes from northern Europe are scarce. The most comprehensive works are those of Uffenorde (1986); and Uffenorde et al. (1979) on Miocene ostracodes from the southern North Sea Basin.

In southern Europe, important works have been carried on the Miocene ostracodes from the Aquitaine Basin (Carbonel, 1985), the Miocene and Pliocene of the Rhône Valley by Carbonel (1969), Carbonel and Ballesio (1982) and Carbonnel and Martini (1976), and the Miocene-Pliocene on the central and eastern Mediterranean Basin by Sissingh (1976, 1982). This last author subdivided the middle Holoocene to Holocene interval into 19 ostracode zones characteristic for the successions in brackish, infralittoral, circalittoral to upper bathyal and deeper environments. Carbonel and Jiricek (1977) proposed tentative correlations based on ostracode bioevents between the Rhône Valley and the paratethys.

The stratigraphic distribution of Plio-Pleistocene ostracodes is fairly well known in southern Europe, essentially in Italy by the various works of Colalonghi (1968), Colalonghi et al. (1972), Colalonghi and Russo (1974) and in the eastern Mediterranean Basin (Sissingh, 1976, 1982). Correlations with planktonic foraminifera are generally well established.

**SELECTED REFERENCES**

Laagland, 1986; Laagland, 1990) appear in the Mediterranean towards the end of that zone. The late Rupelian sees the appearance of Lepidocyclinids (first in the northern area and numerously recorded in the Lower Miocene shelf facies of tropical seas. A biozonation scheme based on these forms is proposed: (1) biozone (SB 24): Rivianian. Index fossils: Lepidocyclinidae spp., Operculina spp., and P. escornebovensis. The assemblages also include S. blankenhorni, G. assilioides, Heterostegina spp., O. complanata, V. aquitanica, B. pygmaea, B. inflata, P. delicata, Austrotrilina spp. (e.g. A. paucalveolata); M. septentrionalis seems to be restricted to the upper part of the zone (Germany, Aquitaine, Italy: de Bock, 1976). The last, rather rare N. fichtelli and H. maxima die out in the lower part of the zone (Cahuzac and Poignant, 1993a). N. morgani and E. dilatata are frequent and the latter is said to disappear towards the Oligo-Miocene boundary in many areas; Eulepidina has been frequently reported from the Aquitanian, although quite often deposits dated as Aquitanian by authors are known to be Chattian in age.


It is characterized by the development of the Miogypsinoidea anagenetic lineage (M. complanatus, formosensis, bantensis, lateralis; Drooger, 1963; Cahuzac, 1984), and at the bottom by the appearance of C. eidae and P. escornebovensis (Cahuzac and Poignant, 1993a). Larger Foraminifera are abundant and diversified from throughout Aquitanie and the whole Mediterranean area and some taxa are known up to Germany and the Paratethys (N. morgani, Miogypsinae). N. bouillei is still present in many areas (Mediterranean, Aquitaine up to the peri-Armorican domain: Cahuzac and Poignant, 1988).

The assemblages also include S. blankenhorni, G. assilioides, Heterostegina spp., O. complanata, V. aquitanica, B. pygmaea, B. inflata, P. delicata, Austrotrilina spp. (e.g. A. paucalveolata); M. septentrionalis seems to be restricted to the upper part of the zone (Germany, Aquitaine, Italy: de Bock, 1976). The last, rather rare N. fichtelli and H. maxima die out in the lower part of the zone (Cahuzac and Poignant, 1993a). N. morgani and E. dilatata are frequent and the latter is said to disappear towards the Oligo-Miocene boundary in many areas; Eulepidina has been frequently reported from the Aquitanian, although quite often deposits dated as Aquitanian by authors are known to be Chattian in age.


The distributional pattern changes due to several disappearances at the top of the Chattian and as a result Larger Foraminifera are reduced both in number and diversity. This zone is characterized by the M. gunteri-tani lineage. Some other species are also present: N. morgani, P. escornebovensis, O. complanata, Heterostegina spp., likewise M. dehaartii at the base. The first “Miopleidocyclina” (M. socii group: de Bock, 1977) appears during that interval.


The M. globalina-intermedia-mediterranea lineage is the essential element of the Burdigalian. A. howchini (Italy, Turkey) and P. escornebovensis are recorded in the lower part of the zone just as M. bardi-galeensis-negrii (Mediterranean, Aquitaine; Adams et al., 1983; Cahuzac and Poignant, 1993b). N. tournoueri still persists, and disappears according to the different areas at the latest towards the N6-N7 boundary (Drooger, 1979; Adams, 1992). At this limit, M. casalmanni-mediterranea just appears in the southernmost basins (Portugal, Spain, Italy: Wildenberg, 1991). The co-occurrence of the latter with B. melo group is not reliable anywhere in the European basins, in which Miogypsina does not reach zone N8.


The taxa diversity strongly diminishes. H. spp. (for instance H. granulatata, occurring in the Langhian, Papp and Kupfer, 1954) and B. melo group, are the only rather common Larger Foraminifera. B. melo curvicauda seems to occur first in N8, while the last B. melo melo reach the lower Messinian in some areas of the Mediterranean (Bizon et al., 1973). In that interval, Planorbulinella spp., D. italica are observed, just as the last Operculina in the Tortonian (Mediterranean; Drooger, 1979; Adams, 1992), and some Neorotalia in the mid-Miocene.

SELECTED REFERENCES


Other cited references can be found in:
The shallow benthic foraminiferal biozones (SB) presented on Chart 3 are, in part the result of the project “Early Paleogene Benthos” (IGCP Project 286). These SB biozones cover the Paleocene and Eocene time from the eastern shores of the Atlantic (Paris-Pyrenean Basins) to Assam (India). Basically, they are derived from species ranges as observed in many outcrop sections in the Pyrenees, Swiss Alps (Schlieren- and Gurnigelflysch and various sections in the Hel- vetic units), northern Italy (Verona, Vicenza), Adriatic and Gargano platforms, Crimean Peninsula and Haymana Basin (central Anatolia).

The integrated numbered biozonations with the prefix SB is independent from the standard plankton zonations, but correlated with them. It is directly correlated with sedimentary sequences as observed in many outcrop sections in the Pyrenees, Swiss Alps (Schlieren- and Gurnigelflysch and various sections in the Hel- vetic units), northern Italy (Verona, Vicenza), Adriatic and Gargano platforms, Crimean Peninsula and Haymana Basin (central Anatolia).

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Although separate Miocene to Quaternary diatom zonations exist for the North Pacific, low latitudes, and Southern Ocean (Barron, 1985), only the North Pacific zonation is provided, because it is the most widely applicable zonation and can often be used at higher latitudes. The Oligréic to late early Eocene zones are those of Fenner (1984) with secondary calibration to the magnetostatigraphy mainly through the correlation with calcareous nannofossil zones suggested by Fenner and Mikkelsen (1990). The base of the Triceratium kanayae zone, however, is placed in the middle part of calcareous nannofossil subzone CP 12a based on Barron’s unpublished studies of DSDP Hole 390A. The only direct correlation to magnetostatigraphy for these zones is for the bases of the Rocella gelida and R. vigilans zones which are taken from Gladenkov and Barron (in press).

The early Eocene to early Paleocene zones are those of Fourtanier (1991) who also provides correlation to calcareous nannofossil zones and limited calibration with magnetostatigraphy at ODP Site 752. In order to fill out the diatom zonation for the Cenozoic, an earliest Paleocene zone, the Hemiaulus rossicus -Trinacria heibergiana assemblage zone is included, in part after Strelnikova (1990). Here, this basal Cenozoic zone is informally recognized as the interval from the last occurrence of Gladiaux spp. at the Cretaceous/Tertiary boundary to the first occurrence of Hemiaulus periterus.

**APPENDIX**

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**REFERENCES CITED CAN BE FOUND IN**


**DIATOMS**

John A. Barron


Although separate Miocene to Quaternary diatom zonations exist for the North Pacific, low latitudes, and Southern Ocean (Barron, 1985), only the North Pacific zonation is provided, because it is the most widely applicable zonation and can often be used at higher latitudes. The Oligréic to late early Eocene zones are those of Fenner (1984) with secondary calibration to the magnetostatigraphy mainly through the correlation with calcareous nannofossil zones suggested by Fenner and Mikkelsen (1990). The base of the Triceratium kanayae zone, however, is placed in the middle part of calcareous nannofossil subzone CP 12a based on Barron’s unpublished studies of DSDP Hole 390A. The only direct correlation to magnetostatigraphy for these zones is for the bases of the Rocella gelida and R. vigilans zones which are taken from Gladenkov and Barron (in press).

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MESozoic ERA

GEOmagnetic POLARITY TIME-SCALE

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Introduction

The Mesozoic portion of the magnetic polarity time scale was compiled from selected publications. Magneto-biostratigraphic studies published prior to 1993 were compiled by Ogg (1995). A version of that magnetic polarity scale with modifications derived from publications through early 1994 was incorporated in the Mesozoic time scale of Gradstein et al. (1994, 1995) after rescaling to the durations of ammonite zones or subzones. In cases where the ammonite-zonal control is less complete (e.g., Sinemurian), the observed pattern is scaled within the stage. This Gradstein et al. (1994) version has been used on the chronostratigraphic charts of this volume. The following review briefly summarizes revisions of the compilation of Ogg (1995) incorporated on the chronostratigraphy charts and indicates a few additional magnetostratigraphy studies of late 1994 through 1996 that are not included on the charts.

The magnetic polarity time scale for the Mesozoic is well-documented in the Cretaceous and latest Jurassic where the seafloor magnetic anomaly pattern provides a guide for scaling the polarity sequence. The polarity pattern is known in partial detail for two-thirds of the Triassic and Jurassic ammonite zones. The major stages with ill-defined, inadequately calibrated or unresolved magnetic polarity patterns are the Carnian, Hettangian-Sinemurian, and late Bathonian-Callovian. This magnetic polarity time scale will continue to be enhanced with further high-resolution magnetostratigraphy research.

PRINCIPAL REFERENCES


cretaceous PERIOD

GEOmagnetic POLARITY TIME-SCALE

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The calibration of the magnetic polarity scale to Cretaceous stage boundaries remains uncertain due to lack of agreement for placement of international stage boundaries by the Subcommission on Cretaceous Stratigraphy (e.g., Rawson et al., 1996). The magnetic time scale shown on the chronostratigraphic charts is according to pre-1993 ‘‘common usage’’ biostratigraphic markers for stage boundaries (reviewed in Ogg, 1995, and Gradstein et al., 1994).

There have not been any precise ammonite or nannofossil markers for the Valanginian/Hauterivian boundary in magnetostratigraphic sections, and the observed variability in a dinoflagellate marker for the boundary (last appearance datum of Scripinodiunm dictyotum) brackets polarity zone M10N. However, Channell et al. (1994) have reported a possible occurrence of Acanthodiscus radiatus in an Italian section that would place the ammonite-defined Valanginian-Hauterivian boundary near the base of polarity zone M11n. The regional Purbeck stage of southern England has yielded a magnetostratigraphy consistent with an age assignment to polarity chron M19r through M14r, indicating correlation to latest Tithonian through earliest Valanginian stages of the Tethyan realm (Ogg et al., 1994). The underlying Portland appears to span only polarity zones M21r through M19n, implying a middle and late Tithonian age correlation (Ogg et al., 1994).

AMMONITE ZONATIONS

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Introduction

Ammonite biostratigraphy is a key element in the organization of Cretaceous stratigraphy. Ammonite zones and subzones are used to define stage and substage boundaries. The current Cretaceous ammonite zonation, which is continuously improved, reflects an evolution towards a consensus scheme. The “Colloque sur le Crétacé” in Lyon, France (1963, published in 1965), the Symposium on Cretaceous stage boundaries held in Copenhagen, Denmark (1983, published in 1984), the International Symposium on Cretaceous Stage Boundaries in Brussels, Belgium (1995, published in 1996), the meeting on “Tethyan and boreal Cretaceous” Maastricht, the Netherlands (IGC.P Project n° 362, 1995) and the 5th International Cretaceous Symposium in Freiberg, Germany, (1996), are important milestones in this process.

The zonations used on the Cretaceous Charts originate from the most recently published synthesis (Hancock, 1991, Bulot et al., 1992 and Hoedemaeker et al., 1993) or from publications devoted to specific
Cretaceous subsystems or stages (Robaszynski and Amédro, 1980; Hancock and Kennedy, 1980; Rawson, 1980; Owen, 1985 and Amédro, 1992). The zonation adopted here is a simplified scheme and very likely a provisional one. Certainly it will be modified and/or partly ratified during subsequent meetings on Cretaceous stratigraphy.

As in other Mesozoic systems, ammonite zones and subzones were selected in order to maximize the relative time resolution of the biographical reference framework, preserving the correlations between the faunal realms (boreal or northwestern Europe and Tethyan or southwestern Europe). For the Upper Cretaceous, special attention was given to recently proposed correlations between U.S.A. and Western Europe zonal schemes (Cobban, 1994; Hancock et al., 1994; Kennedy et al., 1992). The selection retains both the up-to-date species names and some obsolete or no longer used ones, in order that non-ammonite specialists would not be lost.

Different philosophies for calibrating ammonite zonal schemes to stages were used for the lower and upper Cretaceous. In the upper Cretaceous, many radiometric data are correlated with ammonite zones in the U.S.A. (Obradovich, 1994; Gradstein et al., 1994); and the subdivision of each stage from Cenomanian to Maastrichtian reflects the selection of Cretaceous stage and many substage boundary strata.

The International Symposium on Cretaceous Stage Boundaries in Brussels, Belgium 1995 (published 1996), made recommendations for the selection of Cretaceous stage and many substage boundary stratotypes (GSPP). These Global Stratotype Section and Points depend on the selected boundary markers. Most of these proposals require further investigation and ultimately the acceptance by the Commission on Stratigraphy. Recommendations made in Brussels postdate the preparation of the Cretaceous Charts and are thus not included. Differences with stage boundaries on the charts are small.

Proposed boundary markers:

Tithonian/Berriasian = Jurassic/Cretaceous = base Berriasella jacobii zone
Berriasian/Valanginian = base Calpionella zone E
Valanginian/Hauterivian = FAD Genus Acanthodiscus
Hauterivian-Barremian = base Spitidiscus hugii zone
Barremian/Aptian = base Magnetic Chron MO
Aptian/Albian = FAD Leymeriella schrammi
Albian/Cenomanian = FAD Rotalipora globotruncanoides
Cenomanian/Turonian = FAD Watinoceras devonense
Turonian/Coniacian = FAD Cremnoceramus rotundatus
Coniacian/Santonian = FAD Cladoceramus unduloplicatus
Santonian/Campanian = LAD genus Marsupites
Campanian/Maastrichtian = FAD Pachydiscus neubergicus

UPPER CRETACEOUS AMMONITES (J. M. Hancock)

Ammonite zonations in use ten years ago for upper Cretaceous successions have already been changed in many details. The proposed scheme is mainly based on zonations established by Hancock and Kennedy (1980), Owen (1984, 1988a, b).

There is every expectation that the zonation shown on the Cretaceous charts will be modified further. The most up to date information since the completion of the charts, and thus not included on the charts, can be found in the proceedings of the “Second International Symposium on Cretaceous Stage Boundaries” in Brussels Belgium, (1995, published in 1996). Provisional basis for the stratigraphic correlation of the upper Cretaceous are still in discussion (Kennedy, 1994).

Improvements in recent years are dominated by the research of W. A. Cobban in the United States of America, C. W. Wright and W. J. Kennedy in the United Kingdom, the late J. Wiedmann in Germany, A. A. Atabekyan in Russia, M. Matsumoto in Japan, H. C. Klinger in South Africa, and until the 1970’s the late M. Collignon in France.

Summary papers in recent years, already out of date, include Cobban (1994), Hancock (1991) and Hancock, Cobban and Kennedy (1994). Some more recent developments are given by Amédro in Robaszynski et al., 1990, Chancellor et al. (1994), Kennedy and Cobban (1991), Kennedy, Cobban and Scott (1992), Thomel (1993), Ward and Kennedy (1993). Several of these papers are focused on correlations between Western Europe and the United States of America.

PRINCIPAL REFERENCE


LOWER CRETACEOUS AMMONITES (Ph. J. Hoedemaeker).

Standard ammonite zonation for southern Europe.

The “Colloque sur le Cretace inferieur” (B.R.G.M., 1965) accepted, albeit with minor changes, the old subdivisions of Kiiian (1910). Since then the standard ammonite zonation for southern Europe (Tethyan) was drastically improved. In Digne, France (1990, IGCP Project 262), results of the investigations of Bogdanova (1978), Busnardo (1984), Company (1987), Delanyo (1990), Hoedemaeker (1982), Le Hégaret (1971), Kakabzde (1983), Owen (1979), Moullade and Thieuloy (1967), Thieuloy (1972, 1977, 1979), and other unpublished data were used to construct a consensus standard ammonite zonation for the Mediterranean region (Hoedemaeker and Bulot, 1990).

In Mula, Spain (1992, IGCP Project 262) agreement was reached on several improvements in the standard ammonite zonation for the Mediterranean region (Hoedemaeker et al., 1993). In Piobbico, Italy (1994, IGCP Projects 362 and 343), the “Mula zonation” was confirmed. This zonation is used on the Cretaceous chart with some additional subzones and horizons proposed subsequently. The scheme is also based on new data provided recently by Bulot et al. (1992, 1993a, b), Blanc et al. (1992, 1994), Bulot and Thieuloy (1993), Atrops and Reboulet (1994).

Standard ammonite zonation for northwestern Europe.

Lower Cretaceous standard ammonite zonations for northwestern Europe have not benefited as much from international agreement. The zonation on the chart is a mixture of German and English zones. The standard ammonite zonation for the Albian, mainly shaped by Spath (1923–1943), Breistroffer (1947) modified by Owen (1979, 1988a, 1988b) and Casey (1961), includes elements of the phylectic zonation constructed by Amédro (1980, 1992) and by Robaszynski and Amédro (1986). The zonation of the Aptian is the English one of Casey (1961). The zonation for the Barremian is in fact the German one introduced by Koenen (1902, 1908) updated by Kemper (1976). The Hauterivian zonation on the Cretaceous chart is based on the Speeton Clay succession (Rawson 1971) and accepted with minor modifications for Germany by Kemper (1976). The Valanginian zonation was developed for Germany by Kemper (1961, 1976, 1978), Kemper et al. (1981), Jelletzy and Kemper (1988), and Quensel (1988). The zonation for the uppermost Portlandian and Ryazanian is from Casey (1973) described in England and applied to Greenland by Birkelund et al. (1984). The Upper Volgian and Ryazanian zonations for Russia and Siberia have been in use for many years.

Tethyan—Boreal correlations

Correlations between the tethyan and boreal ammonite zones for the Aptian and Albian are not particularly problematic as is the case in...
the uppermost Jurassic. However, correlations between tethyan and boreal zones for the Berriasian to Barremian stages are extremely tentative and based on very few genera and species the areas have in common due to extreme provinciality. Correlations for the Berriasian and Valanginian were published by Kemper et al. (1981) and Hoedemaker (1987, 1991). Hauertavian and Barremian calibrations are from unpublished sources.

**PRINCIPAL REFERENCES**


**APPENDIX**


DINOFLAGELLATES, Boreal: Berrisian-Aptian, Tethyan: Berrisian-Turonian

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All Cretaceous FADs and LADs are first-order correlations with boreal or tethyan ammonite zones, except for the Cenomanian-Turonian interval, where first-order correlations are with Planktonic Foraminiferida zones. Boreal and tethyan FADs and LADs are presented on the charts in four columns. Only those publications documenting a selected bioevent are listed below.

Each bioevent (FAD, LAD or acme) is identified by a numerical age (my) and a bibliographic citation associated with that bioevent. These numerical biostratigraphic datums have been related to the boreal or tethyan ammonite zones and subsequently correlated to the chronostratigraphy and absolute time scale of Gradstein et al. (1994, 1995); decimal numbers are intended only as a place holder to help determine the relative position of bioevents. Uncertain stratigraphic positions for zonal boundaries, FADs and LADs are shown with dashed lines. Taxonomy follows Lentin and Williams (1993).

**REFERENCE**


Boreal Dinoflagellate Cysts

Upper Cretaceous Albain-Maastrichtian (see Foucher, in appendix).

**LOWER CRETACEOUS** (Berrisian-Aptian)

Entries on chart: FADs, 124.23/ 124.14; LADs, 125.88/ 125.36/ 125.22/ 124.44/ 123.61/ 123.30

**REFERENCE**


Entry on chart: LADs, 143.83

DINOFLAGELLATES, Boreal: Berrisian-Aptian, Tethyan: Berrisian-Turonian
APPENDIX

OSTRACODES

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Cretaceous ostracodes have been extensively studied in Europe and numerous synthesis on their biostratigraphic value and distribution published (Babinot et al., 1978, 1982, 1983, 1985a,b; Damotte et al., 1981; Neale, 1978)

Upper Cretaceous

In the Boreal realm, ostracode datums can be correlated to a certain degree with ammonite zones in the Cenomanian, echinoids, inoceramid and belemnite zones in the Turonian to Maastrichtian interval (Neale, 1978; Clarke, 1983). Late Campanian and Maastrichtian ostracodes from the Netherlands (Maastrichtian stratotype) have been extensively studied by Deroo (1966).

In the Tethyan Realm, all the ostracode works have been undertaken in carbonate platform environments especially in southern France (Babinot, 1980; Babinot et al., 1985a), and Spain (Rodriguez-Lazor, 1985) which very seldom contain ammonites. Correlation with standard ammonite-zones are therefore purely tentative.

Lower Cretaceous


For the Tethyan realm, ostracode datums were selected from the work of Babinot et al. (1985) on southern France. Correlations with ammonite zones are also rather accurate.

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ACKNOWLEDGEMENTS

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Northern Tethyan area (northern Mediterranean margin)

Data are from Portugal (Berthou), Spanish Pyrenees (Caus, Peybernès); French Pyrenees (Peybernès); Provence and Subalpine Chains (Masse and Anneau-Vanneau); Hungary (Bodrog); Romania (Bucur, Dragastan); Slovenia-west Carpathians (Köhler and Salaj). From the base of the Valanginian to the lower Hauterivian platforms were drowned and deposition is dominated by marls, crinoidal-bryozoan limestones or oolitic limestones. Larger benthic foraminifers are usually missing in these types of environments.

Adriatic area

Data are from northern Italian areas, Karst, Gorizia, Veneria-Giulia (Longo Salvador, Pirini Radazzani, Pugliese) and Friuli (Sartorio), central Italy (Arnaud-Vanneau) southern Italian areas Apulia-Apenines, (Sartorio and Gargano-Murge; Luperto-Sinni, Masse), Croatia, Dinaric Karst area, (Velic and Radoicic), Kosovo (Peyberne`s), Albania (Sadushi), Greece (Decrouez, Peybernès, Skoursis-Coroneou, Carras). From the base of the Albain to the middle Alban, carbonate platforms emerged and were karstified. Carbonate sedimentation took place on the platform margin. The larger benthic foraminifer commonly present in these lowstand systems tracts is Orbitolina (Mesorbitulina) texana.

SELECTED REFERENCES


SMALLER BENTHIC FORAMINIFERA


CHAROPHYTA


CALPINELLIDS

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There are only minor paleobiogeographic variations in the composition of calpionellid faunas. Regional differences in the relative frequencies of species or genera do exist: The genera Calpionellopsis, Calpionellites and Calpionella elliptica are more frequent in the southern part of the Mediterranean basin than in southeastern France. On the other hand, only in the central part of their domain, corresponding to the Mediterranean basin and Cuba, the succession of calpionellid faunas is documented completely. More marginal areas such as different parts of Mexico and the northeastern Caucasus were invaded by calpionellids in the Berriasian only.

Despite these regional differences, there is general consensus about the chronologic succession of the main events and their calibration with ammonite zones. Statements postulating a diachrony of calpionellid zonal boundaries are not supported by factual arguments. The standard zones of Rome 1971 (Allermann et al., 1971) and the standard subzones of the 1984 Symposium meeting in Hungary (Remane et al., 1986) provide the basic frame for interregional correlations. Within this framework, various finer subdivisions have been developed: Remane (1963, 1964) for southeastern France, which is used on the chart, or the subdivisions by Pop (1974, 1994, 1997) for the Roumanian Carpathians and Cuba, Grün and Blau (1996, 1997) for the southern Alps, Lakova et al. (1997) for the Balkan, Rehakova (1997) for the western Carpathians.

Observations:

1. The occurrence of calpionellids in the basal Hauterivian was confirmed by Blanc (pers. comm. 1995) who discovered Tintinnopsella carpatica in a borehole in Neuchâtel. Together with the finds of calpionellids in the Hauterivian of the Slovak Carpathians this justifies the establishment of a Tintinnopsella Zone. The problem with this zone is, however, that both its boundaries are defined by extinction events so that it does not posses truly diagnostic species. In certain regions calpionellids disappear already in the Valanginian, or at least there are intervals without calpionellids from the middle Valanginian upward.

2. A subdivision of the Calpionellites Zone is possible due to the appearance of new species of Calpionellides shortly after Ct. darderi, but more data are necessary to be sure of the exact position of these events due to the rarity of these forms. Taxonomy of the various species may also still need some clarification.

3. At the International Symposium on Cretaceous Stage Boundaries in Brussels, Belgium 1995, the Valanginian working group decided to equate the base of the Valanginian Stage with the base of the Calpionellides Zone, a proposal to become official with the definition of a boundary stratotype. The boundary formerly used by ammonite workers in France was at the base of the Otopeta Zone, corresponding to the base of the Praecalpionellites murgicani Subzone or the middle of the Vocontian subzone D3.

4. The first appearance of Tintinnopsella longa in the upper part of Zone C, confirms the observation in the Vocontian Basin but the precise level may still be subject to further refinement.

5. There is a certain confusion as to the scope of a Calpionella elliptica Zone or Subzone. Its base should correspond to the first appearance of C. elliptica, in the uppermost Zone B but some authors have also used it as a synonym of Zone C of Remane (1963).

6. Calpionellides have originated in the central Tethys. Only there the transition from Chitinoidella can be observed. Several successive waves of faunal migration originate from the central Tethys region. In the eastern Sierra Madre of Mexico, calpionellids appear only in the lower part of zone B; in central Mexico they appear in Zone C (Adatte et al., 1996. Another important migration occurs in Zone D, (perhaps two closely spaced events), documented in the state of Oaxaca (Mexico) and the northeastern Caucasus (Remane, in press). It is of course very tempting to relate these faunal migrations to marine highstands. In any event, on the carbonate platform of the Jura mountains, marine transgressions could be dated by calpionellids as middle to higher Zone D and as Zone E and a carbonate platform in the northeastern Caucasus was drowned at the beginning of Zone D.

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APPENDIX


RUDISTS

Upper Cretaceous

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During the late Cretaceous (Cenomanian to Maastrichtian) rudists extend widely on the shelf areas of southern Europe. According to the paleogeographical evolution of the western Tethyan area the rudist provinciality increases (Philip 1985). Two main rudist provinces can be distinguished: the Periadriatic (Apulian) province and the western European province. Thus cosmopolitan species (recorded with asterisks on the chart) can be found in both provinces and constitute an accurate basis for correlations. Three rudist families contribute to the biozonation of the upper Cretaceous: Caprinidae (mainly for the Cenomanian), Hippuritidae (from the lower Turonian to the Maastrichtian), Radiolitidae for the entire upper Cretaceous.

Rudist biozones in the upper Cretaceous have in general been interpreted as coenozoones, each zone separated by horizons where rudists are scarce or absent.

Calibration of the rudist zonation has been established mainly in the western European domain (southeastern France, northern Spain), areas where basinal facies, bearing ammonites or planktonic foraminifers, are interbedded with rudist carbonate banks. Strontium isotope calibration has been carried out only on the Campanian-Maastrichtian rudist beds of Bulgaria (Swinburne et al., 1992).

Cenomanian

In western Europe, the lowermost transgressive Cenomanian is characterized by the first appearance of Ichthyosarcotites triangularis, an eurytopic species represented both in carbonate and siliciclastic littoral facies (Philip 1978; Bilotte 1985). In the periadiatic area there is in general no hiatus between the Albian and the Cenomanian. The Cenomanian being characterized by the first appearance of genera like Caprina, Neocaprina, Orthopychus, etc. (Polsak 1965; Carbone et al., 1971; Sliskovic 1971; Sirna 1982).

The upper Cenomanian coenozoone contains cosmopolitan species (Caprinula boissyi, Sauvageaia sharpei) allowing correlations between western European and Periadiatic regions (Philip 1978; Jancone and Laviano 1980; Polsak et al., 1982).

Turonian

Due to complex paleogeographic events, a strong rudist renewal occurs at the Cenomanian-Turonian boundary (Philip and Airaud-Cruvére 1991). In sections without hiatuses (i.e. Provence, Philip 1978) the first appearance of Hippuritids takes place in the lowermost Turonian.

In western Europe, Hippuritids (Vacconites, Hippurites) provide a zonation of the Turonian calibrated to ammonite zones (Devalque et al., 1982; Platel 1982; Bilotte 1985), while in the periadiatic area the Turonian is poorly documented (Polsak 1962).

Coniacian and Santonian

In western Europe, three coenozoones, well calibrated to ammonite zones, characterize this interval (Philip 1970; Pons 1977; Bilotte 1983, 1985; Floquet et al., 1982; Floquet 1990). In the periadiatic area only the Santonian displays rich and well differentiated rudist coenozoone (Polsak 1965; Laviano and Sirna 1979).

Campanian and Maastrichtian

In western Europe and the periadiatic area, rudist coenozoones are well exposed in this interval. A much debated problem concerns the Campanian-Maastrichtian boundary in the rudist carbonate platforms where ammonites are scarce. In western Europe, the appearance of Hippurites radiosus (Des Moulins) was either considered coeval with the base of the Maastrichtian (Philip and Bilotte 1983; Pons 1977; Platel 1987), or with the uppermost Campanian (Neumann et al., 1983). Lower Maastrichtian ammonites were described by Kennedy et al. (1986) above the Hippurites radiosus biozone of Maurens in the Aquitaine Basin. The ammonite bearing layer of Maurens contains Nostoceras hyatti now considered as the uppermost zone of the Campanian (Kennedy et al., 1992). Cosmopolitan species of rudists occur in the upper Campanian-lower Maastrichtian of both southwestern Europe (Philip 1983) and the periadiatic area (Sladic-Trifunovic 1972 and 1979–80).

The uppermost Maastrichtian rudist coenozoones are found in Sicily (Camoin 1983; Cestari and Sirna 1987), in the Dinardis (Pejovic 1987; Plenicar et al., 1992) and in Limburg (Philip and Bilotte 1983).

Acknowledgments

Many thanks are due to Michel Bilotte for constructive remarks about the rudist biozonation of the Campanian and Maastrichtian.
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RUDISTS

Lower Cretaceous

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Investigations on rudists, as biostratigraphic markers have demonstrated the chronological value of these bivalves. As members of the shallow water carbonate platform biota lacking ammonites and pelagic indices, rudist biostratigraphy is less precise and well calibrated than those of deep water organisms. Nevertheless, studies performed during the last decades have improved their biostratigraphic resolution at stage, substage or even ammonite zone level.

The lower Cretaceous Rudist stratigraphic distribution is mainly known from western Europe where these typical mesogean fauna is recorded subcontinuously throughout the whole corresponding time interval. The perigebiotic record (Apulian domain) is more limited and essentially documented for the Aptian-Albian interval (Masse 1976, 1992, 1995). Three main periods are to be distinguished:

Berriasian to Barremian

The dominant groups are the requeniids and the monopleurids. Among the requeniids only one genus; Matheronia is recorded in the Berriasian-Valanginian while this taxon is followed by Requenia and Lovetchenia during the Hauterivian. Toucasia appears in the upper Hauterivian. Primitive caprinids (Pachytraga) are recorded in the Hauterivian both in European and some African areas; some conical to tubular shell monopleurids such as Agriopleura and Petalodontia develop in the Barremian.

Lower Aptian

Advanced caprinids develop near the Barremian-Berriasian Boundary and spread rapidly in different areas with a distinctive biogeo- graphic distribution. Thus, during the lower Aptian; Ofceria and Praecaprina are known both in western Europe and perigebiotic regions, with distinctive species assemblages whereas Caprina seems to be restricted to western Europe.

At the same time caprinoids and monopleurids are also increasing in diversity with some generic provincialism, e.g. Himeractides-Glos- somyophorus and Biconornicopina are restricted to the perigebiotic domain.

Among the requeniids, Matheronia is found in western Europe while Lovetchenia is recorded in the perigebiotic area.

Upper Aptian-Albian

Caprinids record a mass extinction in the whole peri-Mediterranean area at the lower-upper Aptian boundary.

Thus the upper Aptian and Albian are marked by the development of radiolitids (Eoradiolites—Praruadiolites) and Polyconitids (Hor- topleura—Polycomites) recorded in the whole Mediterranean area; Sel- laea, a typical Arabo-African Albian Taxon, is only present in the periadriatic area. The uppermost Albian is marked by the first appearance of advanced radiolitids (Durania) and the restoration of caprinids (i.e. Caprina).

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CALCAREOUS ALGAE

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As components of the warm shallow Tethyan Cretaceous biota, dasycladale algae are mainly restricted to the present perimediterranean domain where distinct European and Apulian (i.e. perigebiotic) assemblages are found. A number of taxonomic works and regional syntheses, performed during the two last decades now permit to have a comprehensive overview on the chronological distribution and the paleogeography of this group.

Western Europe

With data from southern France, Spain and Switzerland, a synthetic biostratigraphy has been proposed dealing with 27 genera and 44 species (Masse, 1993). The chart only takes into account the most representative taxa. Two major breaks are recorded corresponding with: (1) the mid Valanginian turnover when the majority of Berriasian-Valanginian species disappear, followed by a slow progressive restoration of the species diversity during the Hauterivian and a diversity peak during the late Barremian and early Aptian and (2) the mid Aptian turnover marked by a mass extinction event with a limited recovery during the late Aptian-Albian.

Perigebiotic domain

Luperto-Sinni and Masse (1993) summarized the wealth of data obtained from authors working in Italy, Croatia, Bosnia-Herzegovina and Montenegro. The majority of western European species are also recorded from these areas, some of them with a distinctive chronological meaning while endemic species are also found (especially Hensonella dinarica) or even some genera (e.g. Humiella). The Berriasian-early Aptian assemblages are less well defined than their western European analogs, there is no clear evidence of a mid-Valanginian break whereas the mid-Aptian break is well marked. The percentage of endemic taxa increases with increasing diversity which shows its maximum during the early Aptian.

As a whole for the two regions, a remarkable turnover in the species of Salpingoporella is observed which gives this genus a high biostratigraphic potential. Similarly the Berriasian-Valanginian assemblage and the Hauterivian-early Aptian assemblage are different in composition.
Appendix

Jurassic Period

Geomagnetic polarity time-scale

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The magnetostratigraphy scale in the chronostратigraphy charts is mainly from the compilation by Ogg (1995). Later studies in the Callovian-Oxfordian-Kimmeridgian indicate that this portion of the scale requires modification.

The Oxfordian/Kimmeridgian boundary in the Tethyan realm (base of the Sutneria platynota ammonite zone) was provisionally assigned to the top of polarity chron M25n in the time scale of Gradstein et al. (1994), but later studies suggest an assignment within the older polarity chron M25r (Ogg and Gutowski, 1994; Ogg and Atrops, in prep.). A synchronous Oxfordian–Kimmeridgian boundary between the tethyan and boreal realms, at the base of the Pictonia baylei ammonite zone, is consistent with sequence stratigraphic and magnetic polarity patterns (Gygi et al., this volume; Ogg and Coe, 1997). However, Matyja and Wierzbowski (1997) present biostratigraphic arguments that the Oxfordian-Kimmeridgian stage boundary defined in the boreal realm is equivalent to the middle of the “upper Oxfordian” as defined in the tethyan realm. Further work is required to resolve this discrepancy. Deep-tow magnetic surveys by Sager et al. (1998) have acquired a complex signature of magnetic anomalies M26 through M41 in the Pacific crust of early Callovian-Oxfordian age. Portions of this magnetic anomaly sequence have been correlated to Oxfordian ammonite zones (e.g., Ogg and Gutowski, 1996; Juarez et al., 1994, 1995; Ogg and Coe, 1997, and in prep.), and the base of the Callovian appears to correspond to polarity subchron “M36A” of the deep-tow pattern. Magnetostratigraphic polarity successions for the late Bathonian through Callovian stages have not yet been verified, and this interval represents one of the longest gaps in our knowledge of the Mesozoic magnetic polarity scale.

The Hettangian and Sinemurian stages have not yielded a verified magnetostratigraphy. The Sinemurian appears to be dominated by normal polarity (Yang et al., 1996; Kent et al., 1995), whereas the Hettangian may be dominated by normal polarity (Yang et al., 1996; Kent et al., 1995).

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Ammonite zonations

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Introduction

Ammonite biostratigraphy plays a central role in the definition of Jurassic stratigraphy. Stages and their boundaries are primarily expressed in ammonite zones, subzones and horizons. Jurassic ammonite zonations have been under constant revision during the last two decades. Subdivisions used here refer to the most recently published syntheses with data from Callomon, Cope, Duff, Getty, Howarth, Ivimey-Cook, Parsons, Sykes, Torrens, Willemston, Wright (in Cope et al., 1980a,b), Atrops, Cariou, Contini, Corna, Dommergues, Elmi, Emay, Gably, Geyssant, Hantzpergue, Mangold, Marchand, Meister, Mouterde, Riolut, Rulleave and Thierry (in Cariou and Hantzpergue, 1997). The zonal scheme adopted here is somewhat schematic because ammonite zones and subzones on the charts are selected in order to maximize the relative time resolution of the biostratigraphic reference framework while preserving the correlations between faunal realms. However, where possible, the selected zones and subzones follow the most current species index, but retain, where possible, the outdated zones no longer in use to avoid confusing non-ammonite specialists.

Ammonite resolution and calibration of the Jurassic system

Recent advances in Jurassic ammonite biostratigraphy concern the increased precision in ammonite subdivisions and the improved correlation of these subdivisions between the different faunal realms in Europe. The Jurassic system, depending on the faunal realm, is subdivided into about 70 or 80 zones and 160 or 170 subzones (an additional 350 horizons are not listed on the charts). Considering the entire Mesozoic, the number of ammonite subdivisions within the 61.5 my Jurassic, represents the highest resolution currently attainable by combining the most detailed records. Correlation of zones and subzones has greatly improved between faunal realms such as boreal (arctic areas and northern Europe), sub-boreal (northwestern and northeastern Europe), sub-Mediterranean (southwestern and southeastern Europe) and tethyan (southern Europe and Tethys margins). However, in the
Bajocian-Bathonian and Tithonian of northern Europe major regressive events isolate ammonite faunal realms which results in very different faunas that cannot be correlated directly. Palaeocological constraints such as water depth differences between the epi-cratonic platforms in northern Europe and the tethyan ocean margins in southern Europe also constrain direct calibration of ammonite faunas.

Because of the scarcity of radiometric data, all ammonite zones or subzones within a stage are arbitrarily assigned equal duration except in the Kimmeridgian and Tithonian where magnetic polarity data are available. Gradstein et al. (1994). However, it must be noted that several ammonite zones are well calibrated with radiometric data and these “tie points” are integrated in the Gradstein et al. (1994) timescale.

Boundaries of the Jurassic System


The basal ammonite zone in the Jurassic and the Hettangian Stage where the first Psilocerataceae ammonites appear is the Planorbis zone (Planorbis Subzone), Mouterde and Corna (1991). However, the base of the Jurassic System coincides in western Europe with a major flooding event and there is a tendency to include beds with the first marine invertebrate faunas (hivalves, gastropods, echinoids, foraminifers, etc.) in the Hettangian although they lack the ammonite index fossil.

Opinions strongly diverge for the top of the Jurassic System (Jurassic/Cretaceous boundary). In the tethyan area the final zone of the Tithonian is the Durangites Zone (Mediterranean realm: southern Spain, Italy, Carpathians and Balkans, Enay and Geyssant, 1975) or its traditional equivalent the Transitorius/Microcanthum Zone (sub-mediterranean realm, southeastern France and southern Germany). The Berriasian begins with the Jacobi Zone. In the boreal/sub-boreal scheme (Casey, 1973; Cope 1984) the top of the Lamplugh Zone is correlated with the top of the Jacobi Zone. Below the Lamplugh Zone, no correlation is possible between boreal and tethyan areas because of the total absence of common taxa due to a major regressive event near the Jurassic/Cretaceous boundary. Moreover, opinions diverge on the position of the top of the Portlandian Stage (sensu anglico), it is placed either at the top of the Lamplugh Zone (Wimbledon, 1980) or at the top of the Oppressus Zone (Birkelund, Callomon and Fursich, 1984). At the present day, there is no consensus on the Jurassic/Cretaceous boundary (Hoedemaeker, 1987, 1991). Jurassic stages on the chart reflect what we believe to be a majority view (Geyssant and Enay, 1991), although alternative solutions are entered as well.

Subdivisions of the Jurassic Subsystems

The main subdivisions of the Jurassic follow the decisions of the “Colloques du Jurassique” in Luxembourg (1962, 1967) albeit with minor modifications.

The Liass/Dogger and Dogger/Malm boundaries can be correlated between the faunal realms with relative ease due to a homogenization of faunas. The Liass/Dogger boundary (= Toarcian/ Aalenian boundary) is placed between the last Toarcian ammonite zone (Aalensis/ Fluitans Zone ) recognizable in both the boreal and tethyan domains (Elmi et al., 1991), and the basal zone of the Dogger (Opalinum Zone, Opalinum Subzone) of the Aalenian Stage (Contini et al., 1991). Such a decision eliminates the traditional boundary established by Haug in the last century and discussed by generations of biostratigraphers but has the advantage of coinciding with the boundary of the “Brauner Jura” in Germany.

The Dogger/Malm boundary (= Callovian/Oxfordian boundary) coincides in the sub-boreal and sub-mediterranean realms with the top of the Lamberti Zone, Lamberti Subzone (Thierry et al., 1991). The Oxfordian begins with the Mariae Zone, Scarburens Subzone (Cariou et al., 1991), which can be recognized in both areas as well.

Boundaries between stages and stage-subdivisions

Stage boundaries and their subdivisions into zones, subzones and horizons, have been continuously refined over the last twenty years. Subdivisions in the Hettangian (Mouterde and Corna, 1991), Sine-murian (Corna et al., 1991) and Pliensbachian (Dommergues et al., 1991), are essentially based on the subdivision defined in northwestern Europe (sub-boreal realm: England, France, northern Spain, Portugal and Germany). Prominent differences in ammonite faunas appear only southward in the tethyan realm (Italy and southern Spain). Provincialism begins to be noticeable in the Toarcian (Elmi et al., 1991), Aalenian (Contini et al., 1991) and Bajocian (Contini et al., 1991), but the northwestern European areas (sub-boreal realm) remain the basic reference for the zonal scheme. Alternative units have been plotted alongside the standard divisions, as well as for zones or subzones in the two realms.

The basic scheme for the Bathonian is based on Mediterranean areas (Mangold, 1991) where the ammonite faunas are more diverse and better known than in northwestern Europe. However, there is no complete agreement on a standard scheme of stage zonal subdivisions and on correlations between boreal and tethyan realms, therefore some of the alternative zones and subzones currently used are referred to on the chart.

For the Callovian (Thierry et al., 1991) and lower Oxfordian (Cariou et al., 1991) when boreal influences extend southward to Spain and Portugal, the standard reference is based again on the northwestern European sub-boreal realm. Correlation of biostratigraphic units are rather easy during the Callovian.

Problems arise with middle-upper Oxfordian, Kimmeridgian (Cariou et al., 1991; Hantzpergue et al., 1991) and especially during the Tithonian (Geyssant and Enay, 1991) when increasing provincialism complicates correlations. Connections between southern Europe and northern Europe are episodic and endemic faunas settle in northern Aquitaine, the northern Paris Basin and Germany (“Biome franco-germanique”, Hantzpergue et al., 1991). The late Jurassic regression influences ammonite diversity and distribution. On the chart, it was impossible to represent all subtleties, therefore the boreal scheme for the Oxfordian through Tithonian Stages was selected as the standard.

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OSTRACODES

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Numerous detailed analyses of European Jurassic ostracode faunas during the past 20 years have provided a good knowledge of the stratigraphical and palaeogeographical distribution of a great number of species. Ostracode bioevents can be generally correlated to ammonite zones, especially in the lower Jurassic.

Most data come from the boreal and sub-boreal realms: Great Britain, Paris Basin, Denmark. Information from the European tethyan realm (southern France, Spain, Portugal, Italy) is still scarce.

**Late Jurassic**


As noticed by Kilenyi (1978) “The correlation of British upper Jurassic marine ostracode faunas with those described from Germany, France, Poland and Denmark present no problems. Very often the ranges of individual species agree over long distances within one or two ammonite zones”.

**Middle Jurassic**

Aalenian ostracodes are only well known from Germany (Plumhoff, 1963). Because of the strong condensation of the Bajocian sections, especially in southern England and northern France, relatively little is known on Bajocian ostracodes although Bate (1975) proposed a zonation for the early Bajocian (Discites to Humphresianum ammonite Zone). To the contrary, Bathanian ostracodes are well known, especially from southern England and northern France, and several workers have proposed zonations: Bate (1978), Sheppard (1981a,b and Depéche 1984). However, as stated by Bate (1978), the rare occurrence of ammonites in the Bathanian makes correlations of the varied lithologies very tenuous.

**Early Jurassic**

The most comprehensive study of early Jurassic ostracodes can be found in Michelsen (1975) who studied a large number of wells in the Danish Embayment. Additional useful data on the lower Lias of northern Europe can also be found in Lord (1978), Sivshed (1980), Donze (1985) and Boomer (1991).

For the upper Lias Toarcian, references are made essentially to the works of Bate and Coleman (1975) in Great Britain and of Knitter (1984) in southern Germany.

**SELECTED REFERENCES**


**CHAROPHYTES**


**RADIOLARIANS**

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Numerous Italian and German authors (Parona, 1890, 1892; Panatelli, 1880; Rust, 1885, 1889) were among the pioneers in publishing papers dealing with Jurassic radiolarians. Since then successive workers published new information on the Jurassic fauna of Europe (Cayeux, 1891, 1896; Deflandre, 1953; Dumitracha, 1970; Baumgartner, 1980; De Wever, 1982, 1986). However, in the last two decades the number of papers published increased dramatically and the first biozonation for Europe was published by Baumgartner, De Wever and Kocher in 1980. Unfortunately, most papers dealt with the tethyan area and very little work was done on the boreal area.

Information on Jurassic radiolarians is abundant for the folded tethyan terranes and radiolarians are rock forming (radiolarite) in numerous localities. The present set of tethyan marker species is based on recent publications by Gorican (1987, 1994) and on the synthesis published by the InterRad working group.

There is essentially no literature describing well preserved boreal or sub-boreal faunas. Only occasional species are mentioned from scattered localities in Scotland, (Dyer and Copestake 1989) and Russia (Khabakov 1973; Bragin in press). Illustrations are, however, marginal and of limited use. Therefore, since no reliable datums exist for the boreal province none were entered on the Jurassic chart.

In western Europe faunal differences between warm and cold depositional environments has not yet been demonstrated with certainty. This is mainly do to the fact that most information comes from radiolarite-type rocks which were deposited under the most active parts of upwelling (De Wever et al., 1994). In settings of upwelling mixtures of species representing warm, cold, shallow and deep water faunas often co-occur. Before a distinction between provinces can be attempted, it will be necessary to identify from the appropriate data which markers are indicators of boreal or tropical environments.

Most of the Jurassic datums are calibrated with other biozonations such as ammonites, calcareous nanofossils or foraminifera. First order calibration is to biozones of other fossil groups or in absence of such data directly to the stage. There is no first order calibration with the numerical scale in Ma.

**SELECTED REFERENCES**


**BRACHIOPODS**


**TRIASSIC PERIOD**

**GEOMAGNETIC POLARITY TIME-SCALE**

James G. Ogg
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The magnetostratigraphy scale in the chronostratigraphy charts is mainly from the compilation by Ogg (1995). Later studies in the Anisian through Rhaetian indicate that this portion of the scale requires modification.

Two independent magnetic polarity scales are shown for the Upper Triassic. The first column is derived from continental sediments in North America and is scaled according to the placement of stage boundaries in published stratigraphic studies, whereas the second column is derived from marine sediments in southwestern Turkey and has been rescaled to correspond to individual ammonite zones within each stage (reviewed in Ogg, 1995; see also Gallet et al., 1996). Drilling of lacustrine deposits in the Newark Basin of eastern U.S. has yielded a complete upper Triassic magnetic polarity pattern scaled to Milankovitch cycles of eccentricity (Kent et al., 1995), but the correlation to standard geological stages and associated magnetostratigraphy scales derived from macrofossil- and conodont-bearing sediments remains uncertain.

The boundaries of the Anisian have now been calibrated with magnetostratigraphy (Muttoni et al., 1995, 1997). The polarity pattern from ammonite-zoned Griesbachian through Spathian substages in the Canadian Arctic can be correlated to the magnetostratigraphy of the Werfen Formation of marginal-marine deposits in the Dolomites of Italy using constraints from biostratigraphy and sequence-stratigraphy (Graziano and Ogg, 1994).

PRINCIPAL REFERENCES


AMMONOIDS


CALCAREOUS NANNOFÓSSILS

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Since we know little about the modes of life of the few Triassic nannofossils—be they planktic, nektic or benthic—and since, in the present context, the emphasis is on the stratigraphic distribution of fossils and not on their systematic position, calcispheres and other calcareous forms of unknown affinity are included. Note that calcareous nannofossils from the Triassic have so far only been found in a few areas. Di Nocera and Scandone (1977) and Wiedmann et al. (1979) first illustrated late Triassic calcareous nannofossils from the Triassic of Greece, the southern Alps and southern Germany, Jafar (1983), Janofske (1987, 1992) and a few other authors (see Janofske, 1992) augmented our knowledge with reports of calcareous nannofossils from the alpine upper Triassic. The Triassic calcareous nannofossils from other regions were treated by Bown (1992; British Columbia, and Timor) and Bralower et al. (1991; ODP Leg 122 NW Australia).

Direct correlation of calcareous nannofossil findings with ammonite zonations are very rare. The positions of most events on the chart are thus chosen more as educated guesses than after any criterion.

Carnian

In the lower Carnian (Cordevolian; ammonite zone), the assemblage is dominated by “calcispheres”, namely Orthopithonella misurinae and O. prasina. Also found are Carniclyxia tabellata and Casinaopsis curvata (Janofske, 1992).

Norian


Rhaetian

Richer assemblages are reported from the Rhaetian of the Northern Calcareous Alps (stuerezbau to marshi ammonite zones) by Janofske (1992); P. triassica, O. geometrica, Obliquipithonella rhombica, Eococusphaera zlambachensis, C. minutus and Archecoccolithus koessenensis. The upper Triassic of west Timor yielded P. triassica, E. zlambachensis and C. minutus according to Kristan-Tollman et al. (1987) and Bown (1992). From the Wombat Plateau, Bralower et al. (1992) reported the FO of E. zlambachensis together with the FO’s of C. minutus and A. koessenensis while all species found in the Norian continued into the Rhaetian.

Rhaetian/Hettangian boundary (= Triassic/Jurassic boundary)

According to Bown (personal communication 1994), the last occurrence (LO) of P. triassica best approximates the Triassic/Jurassic boundary, Bown (1992) found no calcareous nannofossils in the section in the New York Canyon, Nevada, a section which had been proposed for the Triassic-Jurassic System boundary.

SELECTED REFERENCES


APPENDIX

OSTRACODES

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Relatively detailed zonations based on ostracodes have been provided only for the Germanic realm although ostracode records and diversity are much greater in the Alpine realm. Zonations based on ostracodes have been established in the Germanic and pre-Caspian Basins by Will (1969), Kozur and Mostler (1972), Kozur (1975), and in Great Britain by Anderson (1964) and Bate (1978). Since these zonations have been established in lagoonal and marginal marine environments, correlations with conodonts and/or ammonites zones are very tentative.

Detailed information on Triassic ostracodes from the Germanic realm are available from two main stratigraphic intervals: the Ilyrian (late Anisian) to Julian (early Carnian) and the Sevatian (late Norian) to Rhaetian.

PRINCIPAL REFERENCES


RADIOLARIANS

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Records of radiolarians are still relatively rare from Triassic sedimentary rocks and most of the information available is concerning Alpine faunas.

Triassic radiolarian are known since a long time but comprehensive studies are rather recent. A preliminary note by Rust (1887) was followed by a more comprehensive study, Rust (1892) which recorded 29 species from 28 Triassic samples of central European hornsteins and calcareous limestones. Parona (1892) figured a dozen poorly preserved forms. More recent studies on the same levels are from De Wever et al. (1979); De Wever (1982); Kozur and Mostler (1972, 1978, 1979); Dumitrca (1978a,b); Gorican and Buser, (1990) and Lahm, (1984).

Samples yielding radiolarians are rare from scattered localities in Europe, mainly concentrated in the tethyan area (Austria, Italy, Slovenia, Serbia, Montenegro, Albania, Greece and Turkey). Most of the Triassic FADs and LADs were calibrated with conodonts, ammonites or peliecyops.

Few well preserved boreal and sub-boreal faunas have been described to date and therefore no FADs or LADs of boreal markers are entered on the chart. Only recently some faunas were recorded from Russia (northern Siberia, Egorov, 1995).

PRINCIPAL REFERENCES


DINOFLAGELLATES

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Records of dinoflagellates are relatively rare in the Triassic and are essentially restricted to its upper part. The oldest unequivocal representative of this group was described by Stover and Helby (1987) from the middle Triassic of Australia. Although palynomorphs of marine origin are abundant in the lower and middle Triassic of the Arctic as well as in the Muschelkalk of the Alpine/Germanic realm, no dinoflagellate cysts were identified.

Southern hemisphere

The most complete succession of dinoflagellates is known from Australia where Helby et al. (1987) subdivided the upper Triassic into five zones. Most of the FADs and LADs on the chart are based on this publication. However, the calibration of these assemblages is relatively poor because of the absence of independent control. The assemblages described by Brenner et al. (1992) from the Wombat Plateau, offshore northwestern Australia which are calibrated with ostracodes and magnetostratigraphy confirm the stratigraphic interpretation given by Helby et al. (1987).

Arctic

In the Arctic, Norian and Rhaetian sections are characterized by regular occurrences of dinoflagellates. Assemblages calibrated with amonites are known from the middle Norian (N. columbianus zone) of the Canadian Arctic Islands (Bujak and Fisher, 1976). Based on spore-pollen evidence, lower Norian and upper Carnian ages are suggested for the oldest occurrences of the Sverdrupiella/Noricyusta assemblages.

Alpine/Germanic

In the Alpine/Germanic realm, distribution of dinoflagellates is restricted to the uppermost Triassic. Diverse assemblages were described from the Rhaetian in the Alpine Kendelbach Graben section in Austria (Morby, 1975). Most of the assemblages from the Rhaetian German facies are not diverse. No dinoflagellates older than Rhaetian are known from the Alpine/Germanic realm.

PRINCIPAL REFERENCES

APPENDIX


Spore-Pollen

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In Europe two distinct provinces—Germanic/Alpine and Arctic—can be distinguished based on the distribution of spore pollen. The events from the Alpine/Germanic realm are plotted in the same column despite the fact that not all of the recorded species are known from both areas. In this province three, probably climatically induced, cycles are reflected in the flora which essentially correspond to the classical threefold lithological subdivision of the Triassic (Bunstandstein, Muschelkalk and Keuper).

Based on available data, some of the widely used markers of the Alpine/Germanic realm seem to have different ranges in the Arctic, although the most distinctive differences between the two provinces are in the quantitative distribution of taxa. Compared to southern localities, palynological successions are more complete in the Arctic.

Germanic/Alpine
In the Germanic realm, calibration with the stratigraphic standard is poor because the Germanic sediments are marginal marine or non-marine whereas the standard is marine. Consequently the ranges of many species are constrained only by the inferred age of the formation they occur in, whereas others are tied to major stratigraphic events. First-order correlations can be worked out for the upper part of the lower Muschelkalk (upper Wellenkalk) and for the uppermost part of the Muschelkalk. So far the most complete palynological records were published by Orlowska-Zwolinska (1977, 1983) from Poland. Van der Zwan and Spaak (1992) proposed a zonal scheme for the lower and middle Triassic.

In some parts of the alpine Triassic, the ranges of palynomorphs are constrained by the occurrence of ammonites. Well-dated assemblages are known from the interval between the Anisian and the lower Carnian and from the upper part of the upper Triassic.

The most complete palynological records in the Alpine realm have been published by Brugman (1986). This author proposes a series of phases in the evolution of the palynological assemblages between the Spathian and the lower part of the Carnian based on both first and last occurrences and the abundance of specific taxa. However, these successions are based on isolated outcrop samples or relatively short stratigraphic intervals. The palynological records from the upper part of the Carnian and the Norian are incomplete as a result of poor preservation of palynomorphs.

Arctic
A first overview Triassic palynostratigraphy has been published by Hochuli et al. (1989) based on some dated core and outcrop samples from exploration wells of the Barents Sea area. Although not complete, reliable independent stratigraphic control from ammonites exists for the interval between the upper part of the lower Triassic (Smithian) and the base of the Carnian. A few well controlled records are also known from the Griesbachian and the Norian. New evidence from a well-calibrated Smithian interval of the mid-Norwegian shelf was recently published by Vigran and Mangerud (1991). The ranges of several species on the chart are taken from a more complete account of the palynological records from dated cores of Smithian to Ladinian age of the Barents Sea (Vigran et al., in prep.) Some events on the chart are based on unpublished data of P. van Veen.

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