Chapter 7
Late Cretaceous and Paleocene Atlantic Sea-Floor Spreading and Alpine Collision

INTRODUCTION

In the North Atlantic domain, sea-floor spreading continued during the Cenomanian to early Campanian along axes established during the Aptian and Albian. During the Campanian, crustal separation was achieved between the Labrador Shelf and Greenland and the southern margin of the Rockall–Hutton Bank. With this, the North Atlantic sea-floor spreading axis rapidly propagated northward into the Labrador Sea (Plate 16).

At the same time, the sea-floor spreading axes in the Bay of Biscay and the southern part of the Rockall Trough became extinct (Kristoffersen, 1977; Srivastava, 1978; Olivet et al., 1984). Following this reorganization of sea-floor spreading axes, the North Atlantic–Labrador sea-floor spreading system dominated the late Senonian to Paleocene evolution of the Arctic–North Atlantic rift systems.

Sea-floor spreading in the North Atlantic domain and continued crustal distension in the Baffin Bay and the Norwegian-Greenland Sea rifts (see *Opening of the Labrador Sea* and *Norwegian–Greenland Sea Rift System,* this chapter) was accompanied by the rotation of the Eurasian Craton relative to Greenland and also a rotation of Greenland relative to the North American Craton. This induced transpressional deformations in the northeastern parts of the Sverdrup Basin and possibly also in Svalbard (Kerr, 1981a, 1981b; Price and Shade, 1982; Hanisch, 1983). In the Canada Basin, in which clearly identifiable magnetic sea-floor anomalies are lacking, sea-floor spreading is thought to have terminated during the Late Cretaceous (Sweeney, 1985).

Despite the rotation of Eurasia relative to North America, the sinistral translation between Europe and Africa continued during the Late Cretaceous. This was accompanied by the transtensional opening of oceanic basins along the South Anatolian fracture zone (Whitechurch et al., 1984). Late Cretaceous opening of the South Atlantic–Indian Ocean, on the other hand, induced a change in the drift pattern of Africa, which now began to converge gradually with Laurasia (Olivet et al., 1984; Livermore and Smith, 1985; Savostin et al., 1986; Westphal et al., 1986).

This caused the progressive closure of oceanic basins in the Central and Western Tethys. The Albian to Turonian closure of the south Penninic Ocean and the ensuing collision of the Alpine subduction system with the southern, passive margin of the European Craton was accompanied by the transmission of compressive stresses into the latter. These stresses induced during the Senonian and mid-Paleocene major intraplate compressional and transpressional deformations at distances up to 1300 km to the north of the Alpine collision front (Plate 17; Ziegler, 1987e). Furthermore, there are indications that the rate of crustal extension in the North Sea and Norwegian–Greenland Sea rifts decreased during the Late Cretaceous. Termination of sea-floor spreading in the Bay of Biscay coincides with the onset of convergence of Iberia with the southwestern margin of Europe, their collision, and the early phases of the Pyrenean orogeny (Olivet et al., 1984; Mirouse, 1980).

The coincidence of these phenomena suggests that the northward drift of Africa induced a change in the regional stress patterns affecting the Tethys domain and Western and Central Europe.

The Late Cretaceous is characterized by a major tectonoeustatic rise in sea level, presumably resulting from a global acceleration of sea-floor spreading and a commensurate reduction of the ocean basin volume (Pitman, 1978; Donovan and Jones, 1979). At the end of the Late Cretaceous, sea levels had risen to a maximum high stand of some 110-300 m above the present level, according to some authors (Hays and Pitman, 1973; Vail et al., 1977; Bond, 1978, 1979; Hancock and Kauffmann, 1979). This caused a worldwide overstepping of the Early Cretaceous basin margins and the reopening of seaways linking the colder water Arctic, Barents Shelf and West Siberian Platform seas with the warm water North Atlantic and Tethys oceans. This facilitated a renewed, extensive faunal exchange (Matsumoto, 1973; Tröger, 1978; Wiedmann, 1979). For instance, establishment of a new seaway linking the Arctic Basin and the Norwegian–Greenland Sea is dated as early Cenomanian by the reappearance of boreal faunas in the basins of Northwest and Central Europe from which they were lacking during the late early Aptian and Albian (Stevens, 1973b).

During the Late Cretaceous transgression, extensive carbonate platforms occupied the northern and southern Tethys shelves (Plate 16). In Western, Central, and Eastern Europe much of the Early Cretaceous land areas became inundated. Consequently, the clastic influx into its basins became drastically reduced. This gave rise to the prevalence of clear water conditions and the deposition of the Chalk series (Ziegler, 1982a; Vinogradov, 1969). Detail facies analyses of the chalk of the North Sea area indicate that the depositional water depth of these carbonates ranged from less than 100 m to possibly 1000 m (Hancock and Scholle, 1975; Watts et al., 1980; Hancock, 1984). Upper Cretaceous chalks form an important reservoir for major hydrocarbon accumulations, particularly in the Norwegian and Danish sector of the Central North Sea (Ziegler, 1980; D'Heur, 1986; Sørensen et al., 1986).

In the cold water dominated areas of the Sverdrup Basin, on the Barents Shelf, in the West Siberian Basin, and in the Norwegian–Greenland Sea area, Upper Cretaceous series are represented by open marine shales (Plate 26).

In the Arctic–North Atlantic realm, the mid-Paleocene corresponds to a period of regression (Vail et al., 1977), which was probably induced by regional lithospheric deformations (Cloetingh et al., 1985; Cloetingh, 1986a, 1986b; Plates 17, 26, 27). Particularly in Western and Central Europe, this regression was accompanied by major intra-plate deformations. During the late Paleocene, sea-levels rose again, causing renewed transgressions (Vail et al., 1977).

EARLY ALPINE OROGENY

The early Alpine orogenic cycle is here defined as spanning Late Cretaceous to mid-Paleocene times (Trümpy, 1980).

During the Late Cretaceous, the Italo-Dinarid promontory continued to rotate counterclockwise relative to Europe as a consequence of the sinistral translation between Africa and Fennosarmatia (Westphal et al., 1986). At the same time, the counterclockwise rotation and northward drift of Africa caused the progressive closure of the South Penninic–Piedmont–Ligurian–Alboran Ocean along a system of south-plunging subduction zones (Tollmann, 1980; Homewood et al., 1980; Trümpy, 1980; Debelmas et al., 1983; Vegas and Banda, 1983). Similarly, gradual closure of the Tethys in the Eastern Mediterranean domain was associated with the eastward propagation of the collision front between the Italo-Dinarid block and the Dinarid–Hellenic–South Pontides subduction system (Michard et al., 1984; Bonneau, 1984; Plate 16).

In the Dinarides and Hellenides, internal nappes became stacked (Aubouin, 1973; Richter, 1978; Channell et al., 1979; Cadet et al., 1980; Bonneau, 1982), and in the southern Pontides Tethyan ophiolites became obducted during the Senonian onto the Anatolid–Tauride Platform (e.g., Bozkir nappes) (Sengör and Yilmaz, 1981; Okay, 1984). In the North Pontides-Transcaucasian arc, the Late Cretaceous corresponds to a major orogenic cycle that was punctuated by widespread calc-alkaline volcanism. This was accompanied by compressional foreland deformations in the Crimean and the Caucasus (Adamia et al., 1981; Borsuk and Sholop, 1983; Khain, 1984a).

In the intra-Carpathian domain, Albian and Late Cretaceous compressional deformation of the internal Dacides was associated with the closure of the oceanic Transylvanian–Pienid basin and the imbrication of the external Dacides. The deformation front of the Dacides looped southward around the stable Moesian platform and linked up to the east with the North Pontide–Transcaucasus thrust front (Burchfiel, 1980; Sandulescu, 1982, 1984). During the Late Cretaceous, extensive synorogenic flysch series were deposited in the Carpatho-Balkan–Pontides foredeep. Laterally, these gave way to carbonate platforms covering the stable foreland (Vinogradov, 1969; Sandulescu, 1984; Borsuk and Sholop, 1983).

In the Alpine domain, progressive closure of the Penninic–Piedmont Ocean culminated during the Albian to Turonian in the collision of the Austro-Alpine subduction systems with the Central Penninic Block and the development of a new subduction zone along its northern margin in the area of the Central and Eastern Alps. Following the imbrication and partial subduction of the North Penninic Basin, which apparently was characterized by a drastically attenuated continental crust, the orogenic front reached probably during the late Turonian to early Senonian the passive Helvetic shelf margin of the Eastern Alps and the Northern Carpathians (Frisch 1979; Geyssant, 1980; Trümpy, 1980; Tollmann, 1980; Debelmas et al., 1983; Winkler and Bernoulli, 1986). This was presumably accompanied by the development of a true A-subduction zone at the Penninic–Helvetic boundary, marking the southern limit of the North European Craton.

As a consequence of the collision of the Alpine orogenic system with the fractured and weakened European foreland, compressional stresses were transmitted into the latter. These stresses induced the intra-Senonian, Sub-Hercynian reactivation of preexisting intraplate discontinuities in the Alpine foreland (see next section of this chapter). In particular, the Mesozoic rifts and wrench-induced basins with their thinned crust, acting like shear-pins in an otherwise rigid craton, started to collapse and became inverted to varying degrees. At the same time, the Permo-Carboniferous and Late Jurassic–Early Cretaceous fracture zones, which transected the Bohemian Massif, became reactivated and governed the gradual uplift of major basement blocks along wrench and steep reverse faults (Malkovsky, 1987; Schröder, 1987). These compressional foreland stresses caused during the Senonian intraplate deformations at distances up to 1200 km to the north and northwest of the Alpine collision front (Fig. 41).

In the west Central and Western Alps, the Late Cretaceous closure of the Penninic–Piedmont Ocean was accompanied by the obduction of oceanic crustal material onto the eastern flanks of the Briançonnais High, the so-called Piedmont domain. Development of an A-subduction zone in the North Penninic Trough did not, however, progress further westward than the Central Alps (Debelmas et al., 1983). Correspondingly compressional stresses were not yet exerted onto the foreland of the Western Alps where intra-Senonian compressional intraplate deformations are conspicuously absent (Ziegler, 1987c).

In the Ligurian and Alboran oceans, the development of intraoceanic subduction zones was followed by the Senonian obduction of ophiolites onto southern Piedmont and onto the margin of the Corsica–Sardinia block (Debelmas et al., 1983; Harris, 1985). This was accompanied by the compressional deformations of Mesozoic graben systems of the Lower Rhone Valley (Baudrimont and Dubois, 1977). Furthermore, there is evidence that Senonian compressional deformations, giving rise to metamorphism, also affected the margins of the Alboran–Kabylia block corresponding to the internal zones of the Betic Cordillera of southern Iberia and the Maghrebian fold belt of northern Algeria and Tunisia (Wildi, 1983; Vegas and Banda, 1983; Dercourt et al., 1986).

In the Northern Carpathians and in the Alpine domain, the Late Cretaceous evolution of the Austro-Alpine and Penninic nappe systems was accompanied by the accumulation of synorogenic flysch series in the North and South Penninic troughs and in the Piedmont Basin. From the latter, basin axial currents transported clastics into the northern parts of the Lagonegro Trough (Ksiaskiewicz, 1965; Gwinner, 1971; Schwab, 1981; Homewood, 1983; Debelmas et al., 1983). Progressive stacking of the crystalline cored Austro-Alpine and Penninic nappes was associated with high-pressure metamorphism but only a very low level of synorogenic volcanism. In this aspect, the Alps and Northern Carpathian differs from the Southern and Eastern Carpathians in which Late Cretaceous andesitic volcanism played an important role (Burchfiel, 1980; Homewood et al., 1980; Milnes and Pfiffner, 1980; Trümpy, 1980; Tollmann, 1980; Sandulescu, 1982, 1984).

During the late Campanian to Danian, little compressional deformation can be observed in the Alpine foreland and in the Helvetic domain (Trümpy, 1980; Ziegler, 1987c). This may be interpreted as reflecting a decrease or even a pause in the dextral oblique convergence between the Austro-Alpine and Penninic nappe systems and the European foreland. These movements accelerated/resumed again, however, during the mid-Paleocene and gave rise to the second, "Laramide" phase of foreland compression which was even more pervasive than the intra-Senonian, "Sub-Hercynian" one (Fig. 42). Northernmost deformations are recorded from the Central North Sea Egersund Basin, located some 1300 km to the north of the present Alpine deformation front (Pegrum, 1984).

Since the foreland of the Western Alps was also affected by intra-Paleocene compressional deformations, it must be assumed that the collision front between the Penninic nappe system and the Helvetic Shelf had propagated westward. Stratigraphic evidence suggests, however, that the north Penninic Valais Trough of the Western Alps became closed only during the Eocene (Homewood et al., 1980; Laubscher and Bernoulli, 1982; Debelmas et al., 1983). From this it can be inferred that the crust and upper mantle underlying the Valais Trough were sufficiently rigid to facilitate the transmission of tangential stresses into the foreland.

The strong Laramide deformation of the Alpine and North Carpathian foreland indicates that the Central and Eastern Alpine nappe systems were partially mechanically coupled with the foreland at their respective A-subduction zones.

It is likely that during the late Senonian and Paleocene crustal shortening persisted, on a moderate scale, also in the Ligurian–Alboran Sea and along the southern margin of the Alboran–

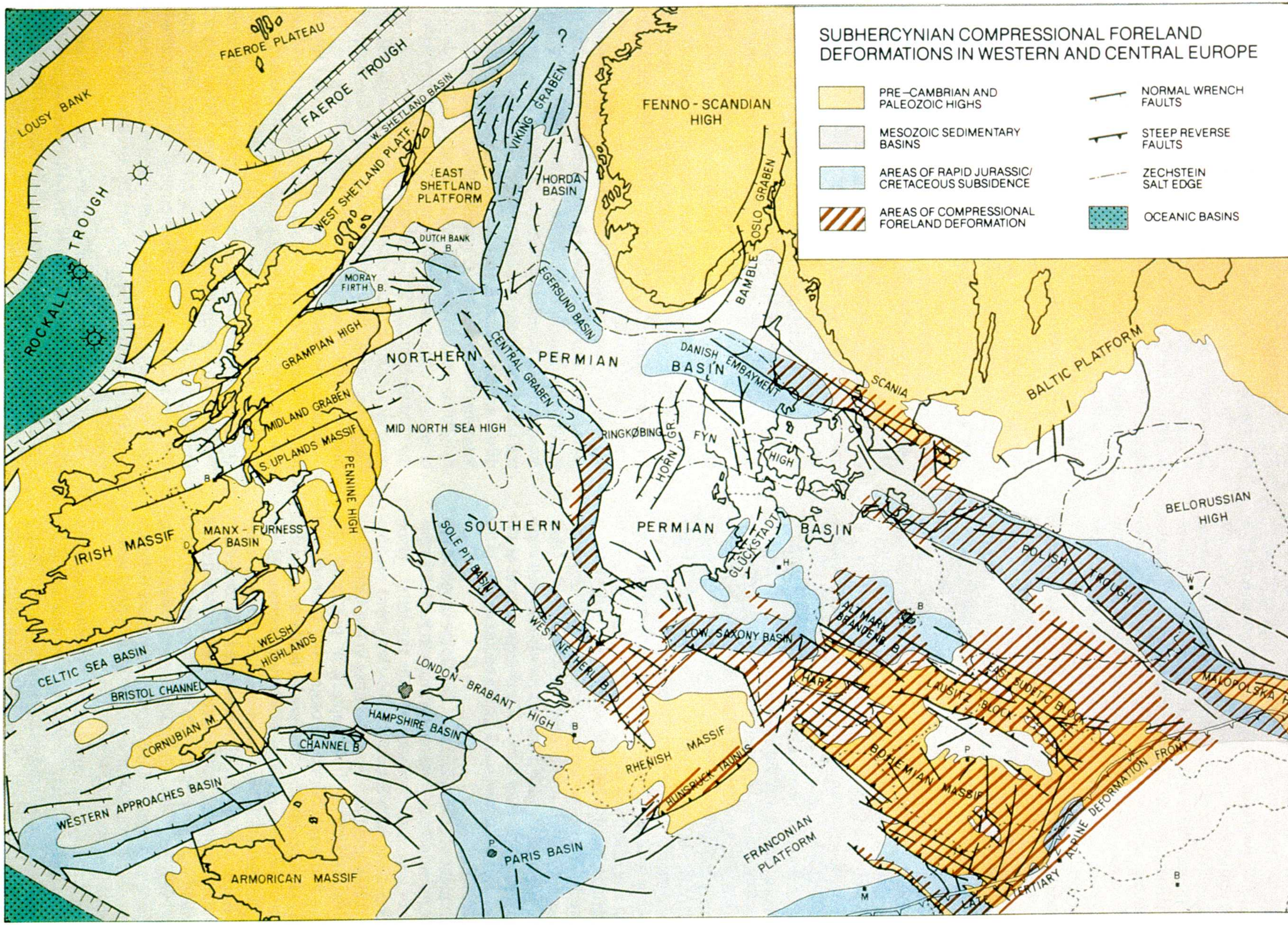


Figure 41—Sub-Hercynian compressional foreland deformations of Western and Central Europe.

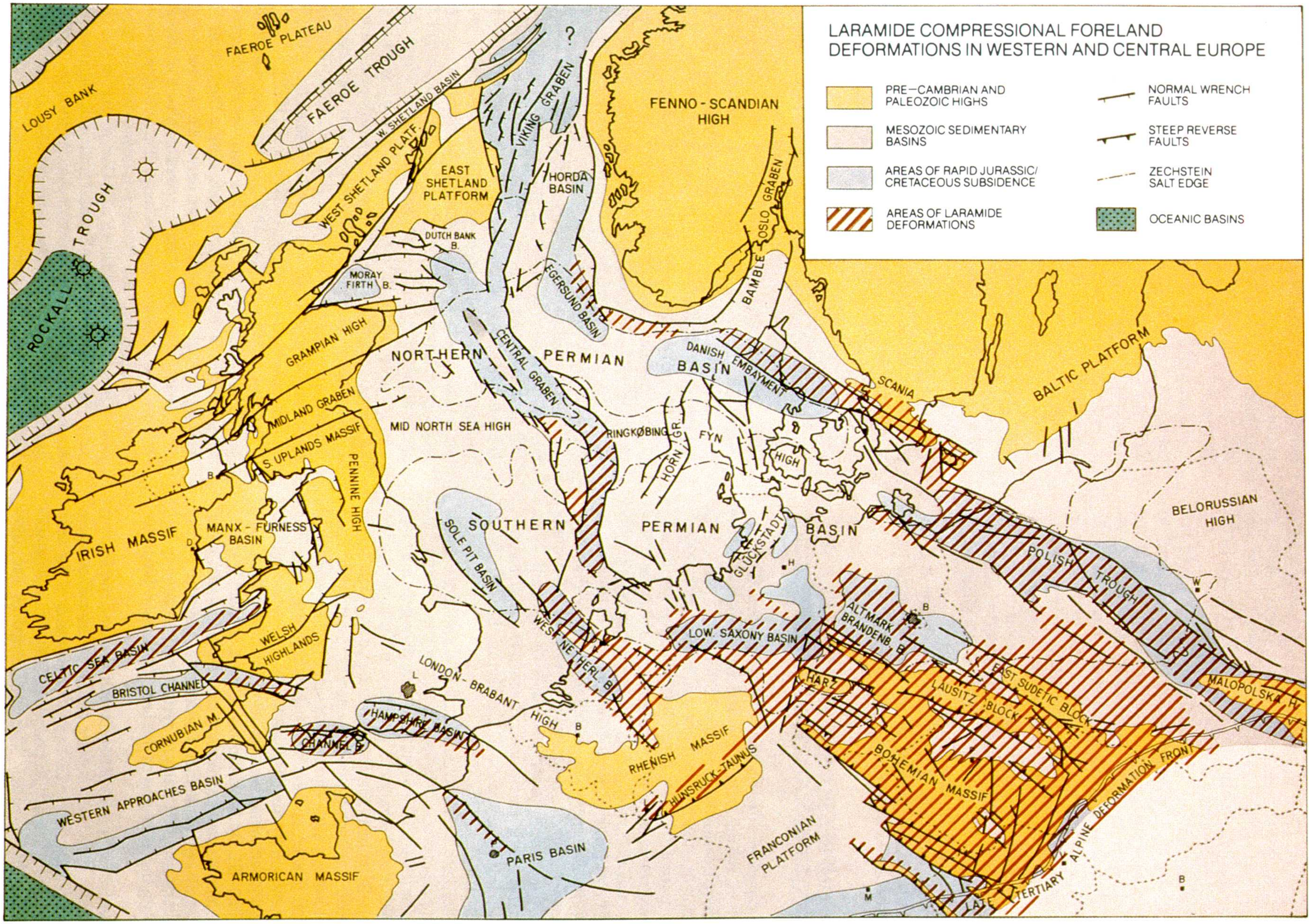


Figure 42—Laramide compressional foreland deformations of Western and Central Europe.

Kabylia block (Plate 17). This concept is supported by limited stratigraphic data from the Rif and Tellian nappes of northern Algeria and Tunisia (Wildi, 1983) and the Betic Cordillera of southern Iberia (Diaz de Federico et al., 1980; Durand-Delga and Fontboté, 1980; Vegas and Banda, 1983).

In the Eastern Carpathians, crustal shortening continued during the Paleocene, as evident by the accumulation of flysch in the foredeep basin paralleling the deformation front of the external Dacides (Sandulescu, 1982, 1984). Also in the Hellenides and Dinarides, orogenic movements, paralleled by the deposition of extensive flysch series, gave rise to local metamorphism and the stacking of the most internal isopic zones (Aubouin, 1973; Bonneau, 1982). In Turkey, the Anatolide and Bozkir nappes were emplaced while the Levantine oceanic basins began to close with ophiolites being obducted during the Maastrichtian (Sengör and Yilmaz, 1981; Okay, 1984; Whitechurch et al., 1984; Ricou et al., 1984).

On the other hand, late Senonian to Paleocene back-arc extension induced the opening of two oceanic basins in the Black Sea area (Zonenshain and Le Pichon, 1986).

COMPRESSIONAL INTRAPLATE DEFORMATIONS IN NORTHWESTERN EUROPE

With the decrease in the rate of crustal extension in the North Sea rift system during the Albian and Late Cretaceous, the wrench-induced basins flanking the northern margins of the London–Brabant and Rhenish–Bohemian massifs became tectonically quiescent (Fig. 35; Plate 30). Similarly, the Polish–Danish Trough ceased to subside differentially and its early Late Cretaceous evolution reflects regional downwarping, possibly in response to progressive lithospheric cooling and contraction.

Following crustal separation in the Bay of Biscay, the rifts of the Celtic Sea–Western Approaches area also became tectonically inactive and the area started to subside regionally (Plate 29). With this, tectonic activity in the wrench-induced basins in the area of the English Channel and also in the Paris Basin abated (Ziegler, 1982a, 1987a).

During the Late Cretaceous, eustatically rising sea levels caused the progressive overstepping of the Early Cretaceous basin margins (Plates 15, 16). By early Senonian time, much of the Irish, Welsh, and Armorican massifs were inundated while the Rhenish–Bohemian Massif remained a positive feature. To the southeast, the Paris Basin was open to the Helvetic Shelf seas. It is, however, uncertain to what extent the area occupied by the Rhine Graben and the South German Franconian Platform were covered by the Late Cretaceous seas since in these areas Mesozoic sediments became deeply truncated during the Cenozoic (Fig. 45). On the other hand, there is clear evidence that the Cretaceous seas encroached onto the Bohemian Massif along the Southwest Bohemian Borderzone. Tectonic activity along this long-standing fracture system had apparently also abated during the early part of the Late Cretaceous (Schröder, 1987).

In the northern foreland of the Carpathians and the Eastern and Central Alps, tectonic activity gradually intensified again during the late Turonian to Campanian as a result of the collision of the Alpine–Carpathian nappe systems with the passive margin of the Helvetic Shelf. This is reflected by the intra-Senonian gradual uplift of major basement blocks along steep reverse and wrench faults in the Bohemian Massif and in southern Sweden and by the inversion of major Mesozoic grabens and wrench-induced basins in the southern North Sea area, in northern Germany, in Denmark, and in Poland. This phase of intraplate compressional/transpressional deformation is referred to as the Sub-Hercynian tectonism (Plate 30; Fig. 16; Ziegler, 1987c).

Basin inversion is defined as the reversal of the subsidence patterns of a sedimentary basin, which had developed under a tensional or transtensional tectonic regime, in response to compressional or transpressional stresses (Fig. 80). Basin inversion generally involves uplift of the basin floor and deformations of the basin fill, whereby the throw on tensional faults, controlling the original structural relief of the respective graben or trough, becomes partly or totally reversed (Bally, 1984). This implies that during basin inversion a commensurate amount of crustal shortening occurs, which is responsible for the deformation of the basin fill. This is reflected in the buckling-up of major anticlinoria, the crestal parts of which generally become subjected to erosion while in adjacent, secondary basins sedimentation is often continuous. Owing to the greater preinversion sedimentary thickness in basinal areas, the structural relief created by inversion movements is in many inverted basins considerably smaller at depth than at shallower levels (Voigt, 1962; Ziegler, 1982a, 1983b, 1987c; Bally, 1984).

The structural style of inverted basins is very variable and depends on the preinversion configuration of the respective basin, the lithologic composition of its sedimentary fill, the degree of its inversion (amount of strain), and the orientation of the basin axis relative to the greatest principal stress that induced its inversion. In this respect, the presence of major halite and/or shale layers, acting as detachment planes, can play a significant role in the disharmony of the structural style observed at shallow and at deeper levels (e.g., Lower Saxony, Broad Fourteens and Sole Pit basins; van Hoorn, 1987b; van Wijhe, 1987; Betz et al., 1987).

In the northern Alpine and Carpathian foreland intra-Senonian, Sub-Hercynian, compressional/transpressional deformations related to the Alpine collision are restricted to areas between the Polish Trough in the east and the Broad Fourteens–West Netherlands Basin to the west. They are also recognized along the Fennoscandian borderzone in southern Sweden and northern Denmark and in the Danish and Dutch sector of the North Sea Central grabens (Fig. 41; Plates 29, 30).

In the area of the Bohemian Massif, the uplifting of major basement blocks along Permo-Carboniferous and Late Jurassic–Early Cretaceous fracture zones also started during the late Turonian and early Senonian as illustrated by the increased clastic influx into the adjacent basins (Malkovsky, 1987; Nachtmann and Wagner, 1987; Bachmann et al., 1987).

During the Maastrichtian and Danian, compressional foreland deformations abated, at least in the more distal parts of the Alpine foreland, but resumed again during the middle Paleocene, giving rise to the even more pervasive "Laramide" phase of inversion. This is the main inversion phase of the Polish Trough, the Altmark–Brandenburg, the Sub-Hercynian, the Lower Saxony, and the Broad Fourteens–West Netherlands basins as well as of the Dutch and Danish part of the Central North Sea Graben and the Fennoscandian borderzone. Northernmost traces of basin inversion occur in the Egersuud Basin, offshore southern Norway. Mild mid-Paleocene inversion movements have also been recorded in the Celtic Sea Trough, in the Fastnet Basin, the Channel area, and in the Paris Basin. In contrast, the Western Approaches Trough was hardly affected by the Laramide movements (Ziegler, 1987a, 1987b, 1987c), (Fig. 42; Plates 29, 30).

Although not closely bracketed by a corresponding stratigraphic record, the main phase of deformation of the Bohemian Massif and of the areas occupied by the Swiss, German, and Austrian part of the Cenozoic Molasse Basin, is probably also of a mid-Paleocene age. Subsurface data from the Vienna Basin (Wessely, 1987) and the Carpathian foredeep (Kolarova and Roth, 1977; Heller and Moryc, 1984) clearly indicates that the area affected by the Late Cretaceous and mid-Paleocene compressional foreland deformations extends at least some 50 km southward in the autochthonous substratum under the East Alpine and Carpathian nappes (Fig. 45). Similarly, the southern parts of the strongly inverted Polish Trough appear to extend under the Carpathian nappe systems (Pozaryski and Brockwicz-Lewinski, 1978; Pozaryski and Zytko, 1979). In the Central Alps, similar foreland deformations are considered responsible for the latest Cretaceous-Early Tertiary unconformity recognized in the area of the Aare Massif and the sub-Helvetic and Helvetic nappes (Herb, 1965).

The degree to which the different rift- and wrench-induced Mesozoic basins in the northern Alpine and Carpathian foreland became inverted during the Sub-Hercynian and Laramide phases of foreland compression is very variable. In general it can be observed that the intensity of inversion decreases with increasing distance from the Alpine collision front.

This is particularly evident in the case of the Polish–Danish Trough (Fig. 43). The southern parts of the Polish Trough have been inverted to the degree that the graben floor, forming the Malopolska Massif, is now exposed in the Holy Cross Mountains. Its northern parts are characterized by gentle anticlinoria that have been truncated to Jurassic levels and display a structural relief of some 2000 m. Along the Fennoscandian Borderzone, major basement blocks have been tilted up, presumably along steep reverse faults, while in northern Jutland inversion movements induced only gentle undulations at the Late Cretaceous Chalk level (Fig. 44; Norling and Bergström, 1987; Liboriussen et al., 1987).

The scope and effects of the Sub-Hercynian and Laramide phases of compressional deformation in the Alpine foreland can be assessed by a comparison of Fig. 42 with the pre-Tertiary geological map given in Fig. 45 (see also Ziegler, 1982a).

Considering that during the Paleocene the Alpine thrust front was probably located some 75 to 100 km further south than at present, the total area affected by these intra-Paleocene foreland deformations and the structural relief created by them is quite spectacular and can be readily compared with the Lara-

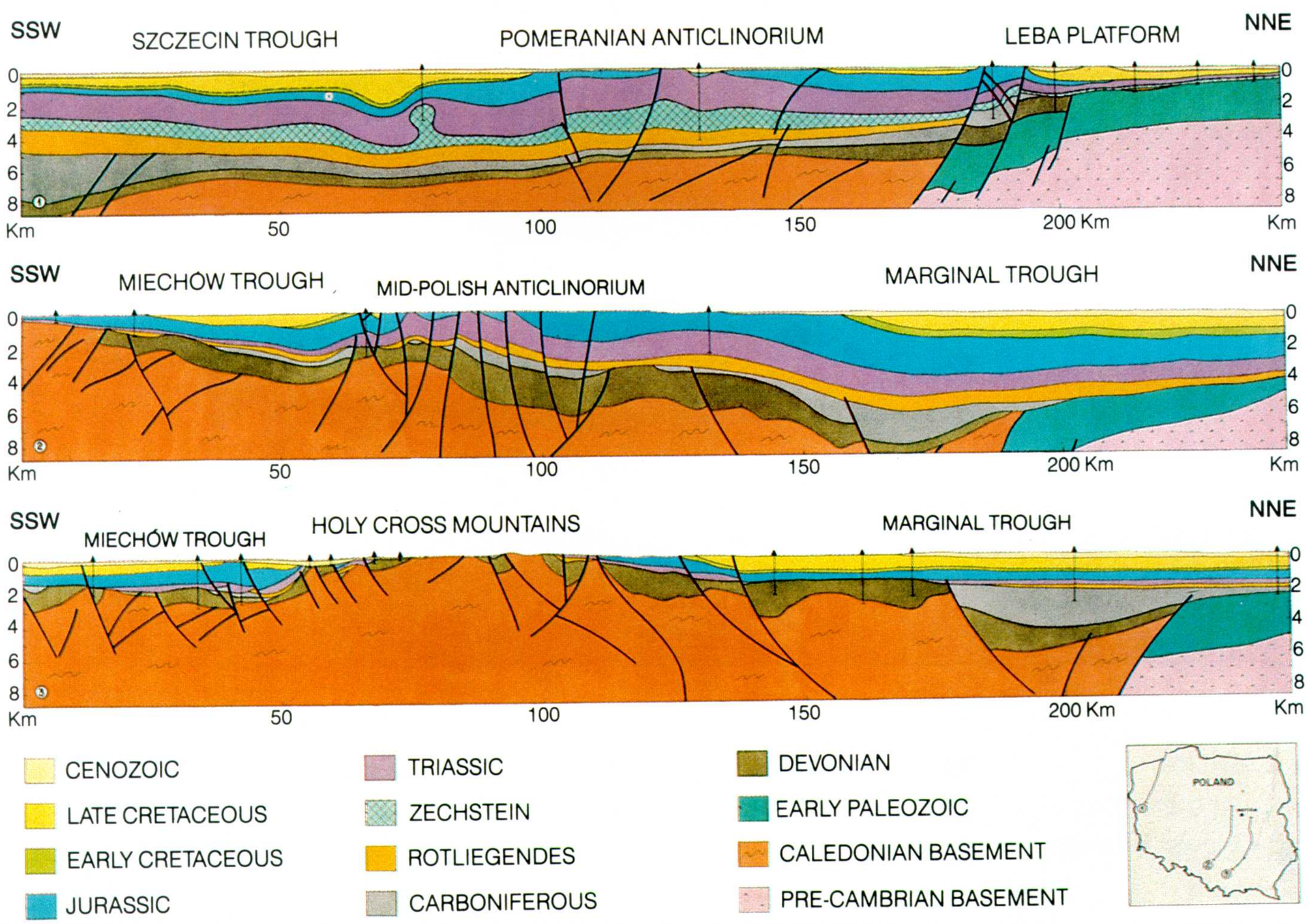


Figure 43—Structural cross sections through Mid-Polish Anticlinorium. After Znosko and Pajchlowa (1968).

mide foreland tectonics of the Rocky Mountains (Brewer et al., 1980) and the Permian foreland structures of the Timan–Pechora area (Ulmishek, 1982; Matviyevskaya et al., 1986). While the Rocky Mountains consist essentially of basement blocks carried by reverse faults, to which the Harz Mountains and the Lausitz Block in the Bohemian Massif can be compared, the Timan–Pechora foreland folds, which represent inverted Devonian rifts, are analogs, for example, for the Polish Anticlinorium (Pozaryski and Brochwicz-Lewinski, 1978).

The crustal configuration of the Polish Trough, which is controlled by refraction and deep reflections surveys, indicates that beneath its axial, inverted part the crust–mantle interface is not pulled up but is somewhat depressed and forms the so-called Guterch Graben (Fig. 46; Guterch et al., 1976, 1986). Reconstructions of preinversion profiles across the Polish Graben indicate, however, that in its deepest parts Mesozoic and late Paleozoic sediments attained thicknesses of up to 10 km (Pozaryski and Brochwicz-Lewinski, 1978). In order to accommodate such a thick sedimentary sequence, isostatic considerations require that a significant amount of crustal thinning through mechanical stretching and possible physicochemical processes must have taken place during the Late Permian and Mesozoic rifting stages of the Polish Trough. Its preinversion crustal configuration presumably resembled that of the present-day Central North Sea (Fig. 61).

These features suggest that during the inversion of the Polish Trough, its crust became mechanically thickened again by transpression-induced crustal shortening. As a consequence of its intense inversion, the Polish Trough achieved isostatic and thermal stability during the Early Tertiary.

The amount of crustal shortening associated with the Sub-Hercynian and Laramide inversion of the different basins in the Alpine foreland and the deformation of the Variscan massifs is at present still difficult to estimate but is unlikely to exceed a few tens of kilometers. These displacements require a decoupling within the crust of the foreland and/or between its crust and the underlying upper mantle or even within the upper mantle. On the other hand, as these intraplate deformations are apparently related to collisional events in the Alpine and Carpathian fold belts, a certain amount of mechanical coupling is required between the foreland plate and the nappe system of these orogens at the respective A-subduction zones in order to facilitate the transmission of tangential stresses from collision zones into the foreland. A working model for the transmission of tangential stresses from collision zones into foreland areas has been developed by Neugebauer and Gao (1985).

Crustal shortening induced by these compressional intraplate deformations requires the subduction of a commensurate

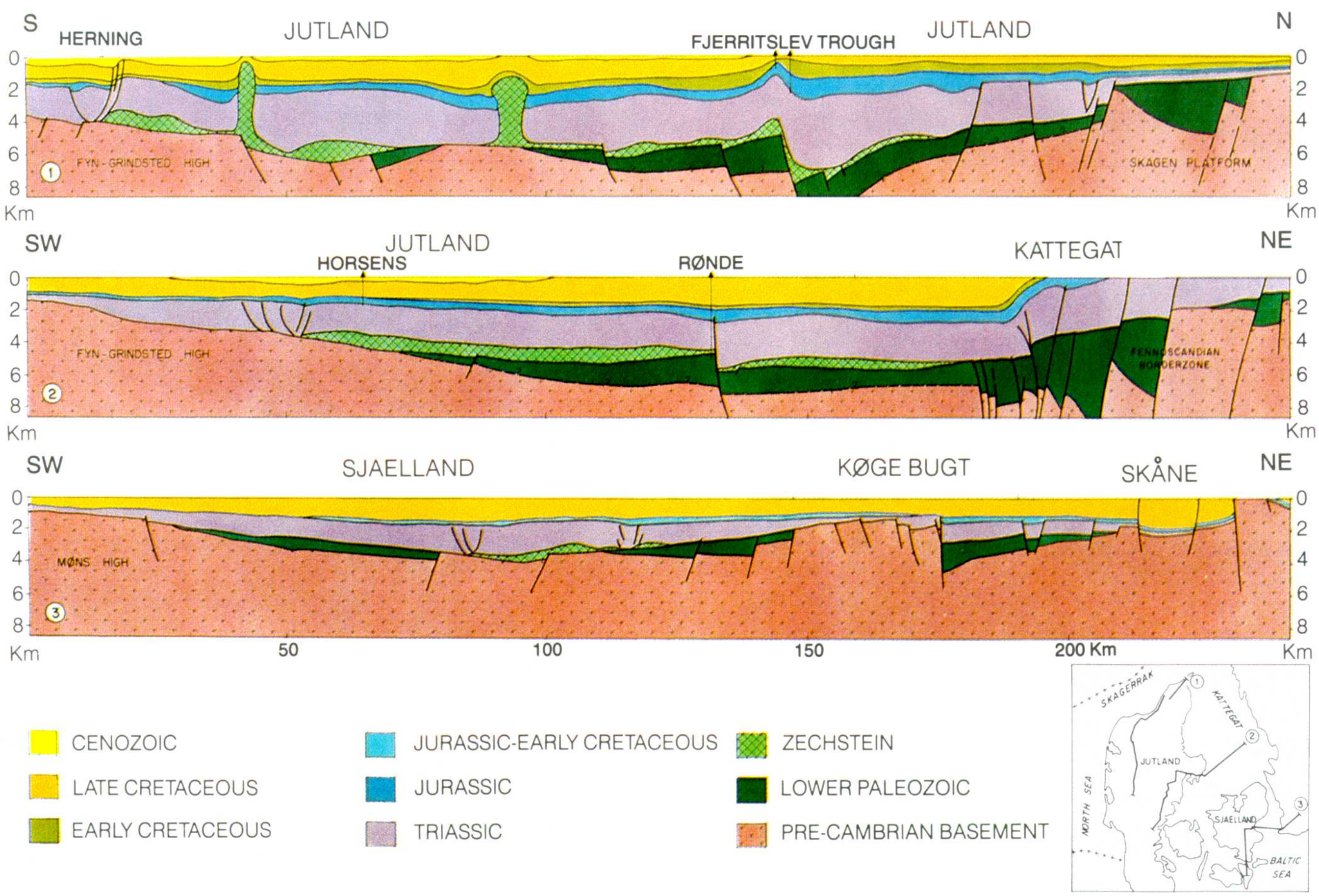


Figure 44—Structural cross sections through northern Denmark and southern Sweden.

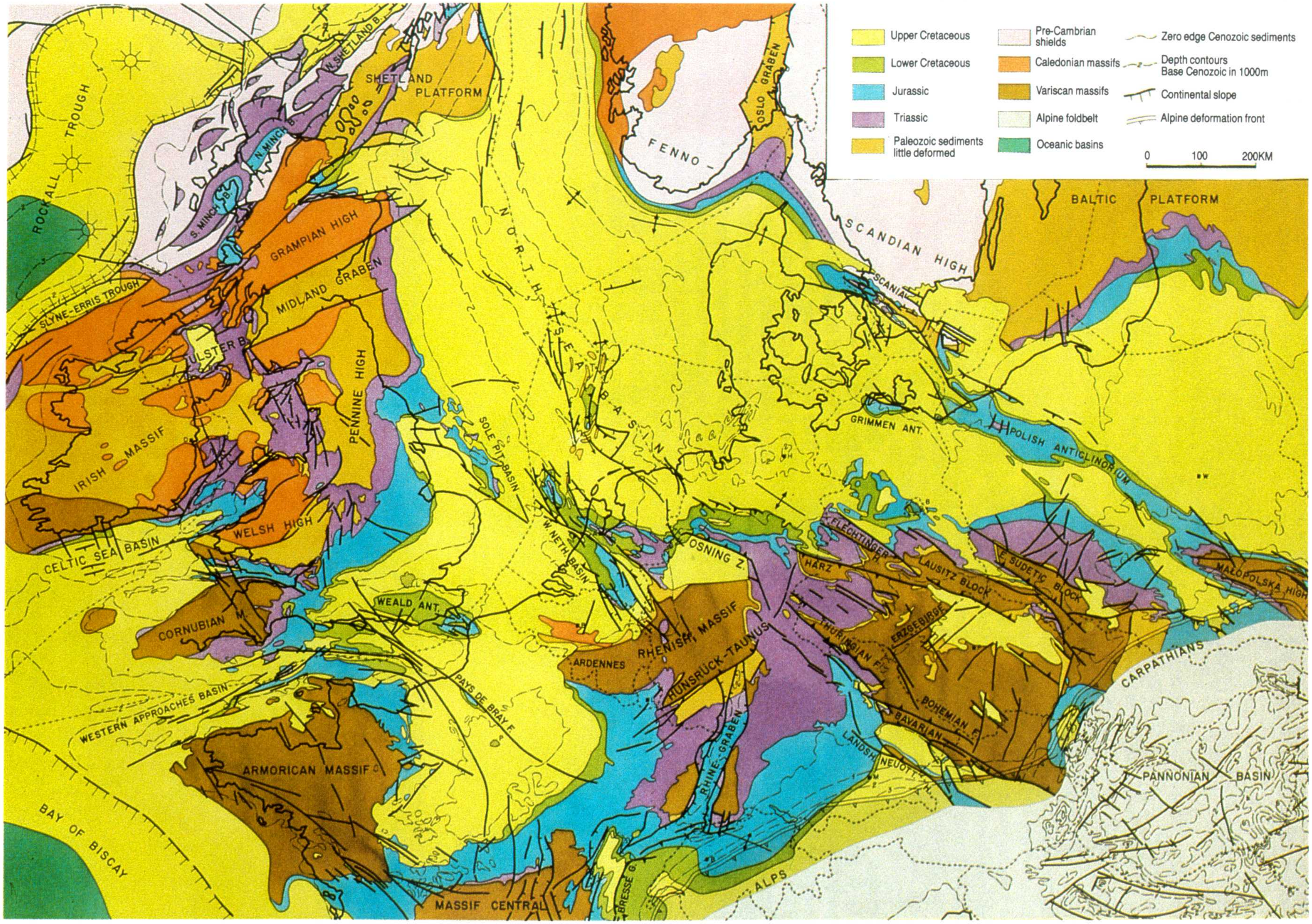


Figure 45—Geological map of Western and Central Europe with Cenozoic sediments removed.

amount of lithosphere, either in the Alpine orogen, if a decoupling occurred within the foreland plate, or locally, if the entire lithosphere is involved in these deformations (see Chapter 10).

IBERIAN MICROCONTINENT

With the late Aptian onset of sea-floor spreading in the Bay of Biscay, Iberia began to rotate counterclockwise away from the Armorican margin of France (Olivet et al., 1982, 1984). As a consequence of continued sea-floor spreading in the North Atlantic, this motion was accompanied by transtensional shear in the Pyrenean and Cantabrian domain, giving rise to metamorphism, an alkaline magmatism, and the tectonic emplacement ultrabasics during the Albian to early Senonian (Plate 27; Choukroune and Mattauer, 1978; Dubois and Séguin, 1978; Brunet, 1984; Vielzeuf and Kornprobst, 1984; Goldberg et al., 1986; Boess and Hoppe, 1986). The southern limit of the Iberian plate was formed by the Azores and Alboran fracture zones. As such, the Iberian microplate occupied an intermediate position between the Eurasian and the African plates.

During the early Campanian, sea-floor spreading ceased in the Bay of Biscay and Iberia began to converge in a clockwise, oblique sense with the southwestern margin of Europe. In the eastern Pyrenees and in Cantabria, first compressional deformations are evident during the Santonian and Campanian. These probably marked the onset of convergence and the subsequent collision between Iberia and the southern margin of Europe, and with this the beginning of the Pyrenean orogeny during which oceanic crust along the southern margin of the Bay of Biscay became subducted (Bilotte, 1978; Boillot et al., 1979, 1985; Alvarado, 1980; Mirouse, 1980; Olivet et al., 1984; de Luca et al., 1985; Sancho et al., 1987). During the late Senonian and Paleocene, continued convergence of Iberia and Europe resulted in the westward propagation of their mutual collision front. This was accompanied by the accumulation of late Senonian and Paleogene synorogenic deep water clastics (flysch) in the rapidly narrowing Pyrenean trough. In southern France, contemporaneous intraplate compressional deformations are evident by the uparching of the Durance axis and the Toulouse Peninsula (Baudrimont and Dubois, 1977; Plaziat, 1975, 1981).

In southwestern Portugal there is evidence for late Senonian transpressional deformations along northeast-striking faults. Late Senonian and Paleocene transtensional dislocations along north–northwest-trending fracture systems induced the intrusion of alkaline and ultrabasic magmas in the area of Lisbon and also offshore on Gorringe Bank (Plate 28; Antunes et al., 1980; Mougenot, 1981; Feraud et al., 1982).

During the Cenomanian to Santonian, much of eastern Iberia was occupied by an extensive, shallow marine carbonate platform. The upper Senonian to Paleocene series is developed in a regressive, partly continental facies. Deeper water conditions dominated the southeastern margin of Iberia, which had not yet collided with the Alboran block (Plates 16, 17; Alvarado, 1980; Azema et al., 1974, 1979; Garcia-Hernandez, et al., 1980; Plaziat, 1981).

Transpressional deformations along the margins of the Alboran block were also initiated during the Santonian (Wildi, 1983; Vegas and Banda, 1983).

This was probably a consequence of the northward drift of Africa, which can also be held responsible for the convergence and collision of Iberia and Europe and the reactivation of fracture systems in southwestern Portugal. In this context, it should be noted that the first compressional deformations evident in Cantabria appear to be contemporaneous with the last phases of sea-floor spreading in the Bay of Biscay. The subsequent termination of sea-floor spreading in the Bay of Biscay may be a direct consequence of the build-up of regional compressive stresses in response to the convergence of Africa and Europe. In this respect, it may be surmised that sea-floor spreading in the Bay of Biscay was not directly associated with a deep

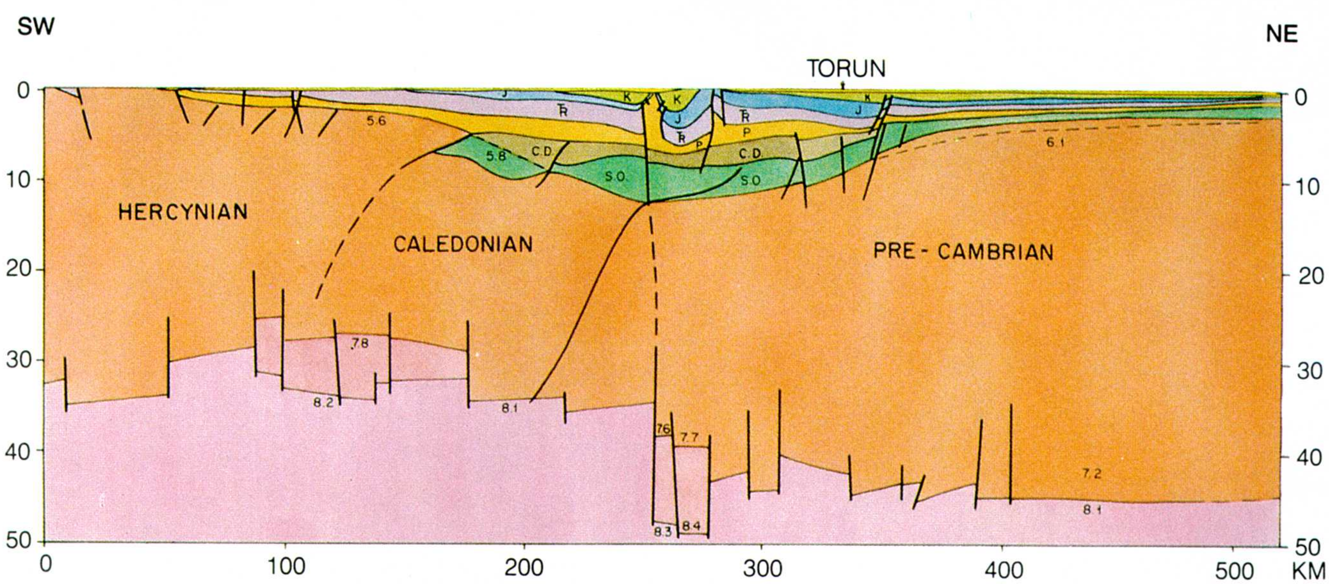


Figure 46—Crustal configuration of the Polish Trough along DSS profile VII. Interval velocities given in km/sec. After Guterch et al. (1976).

asthenospheric upwelling convective system, but was rather the result of asthenospheric advection induced by the oblique translation and rotation of Iberia relative to the southern margin of Europe. This motion of Iberia was presumably governed by the North Atlantic sea-floor spreading system, which most likely was associated with deep asthenospheric upwelling convection currents.

OPENING OF THE LABRADOR SEA

Rifting activity in the Labrador Sea culminated during the Santonian–Campanian in crustal separation and the onset of sea-floor spreading. With this, the Labrador Shelf became tectonically inactive and its further development followed the pattern typical for passive margins (Fig. 47; Umpleby, 1979; Grant, 1980; Gradstein and Srivastava, 1980).

In the southeastern parts of the Labrador Sea, the oldest magnetic sea-floor anomalies recognized are anomalies 34 and 33. In its northwestern parts, where magnetic anomalies are poorly defined, the oldest one identified is anomaly 31 (Srivastava, 1978; Srivastava et al., 1981, 1982; Tucholke and Fry, 1985).

The magnetic sea-floor anomalies of the Labrador Sea show that sea-floor spreading patterns established during the Senonian persisted until late Paleocene time. To the north, the Labrador Sea sea-floor spreading axis terminated in the Davis Strait at the Ungava transform fault zone (McMillan, 1979; Srivastava et al., 1981, 1982; Klose et al., 1982). Seismic and well data indicate that movements along this fault system were of a sinistral transpressional nature and involved the uplifting of high trends (Fig. 52).

In the Baffin Bay, crustal distension continued through Late Cretaceous and Paleocene time. At its northern termination, there is evidence for tensional tectonics in Devon Island and in the southern parts of Ellesmere Island (Mayr, 1984). This suggests that the bulk of crustal stretching taking place in the Baffin Bay was taken up by extension in the Arctic Archipelago and only to a minor degree by strike-slip movements along the Nares Strait transform zone (Plate 16). There is stratigraphic evidence that the Eclipse Trough and probably also the Lancaster

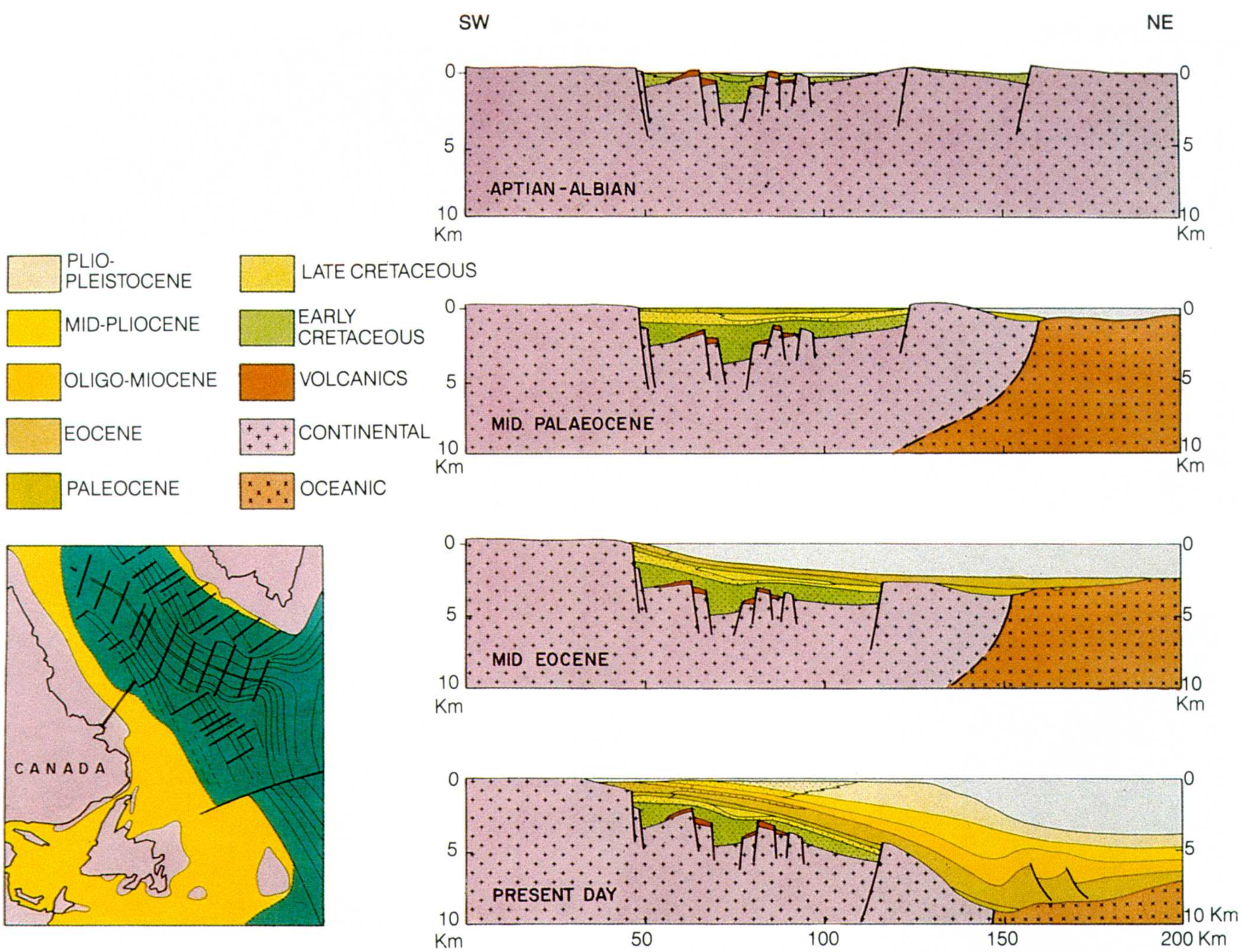
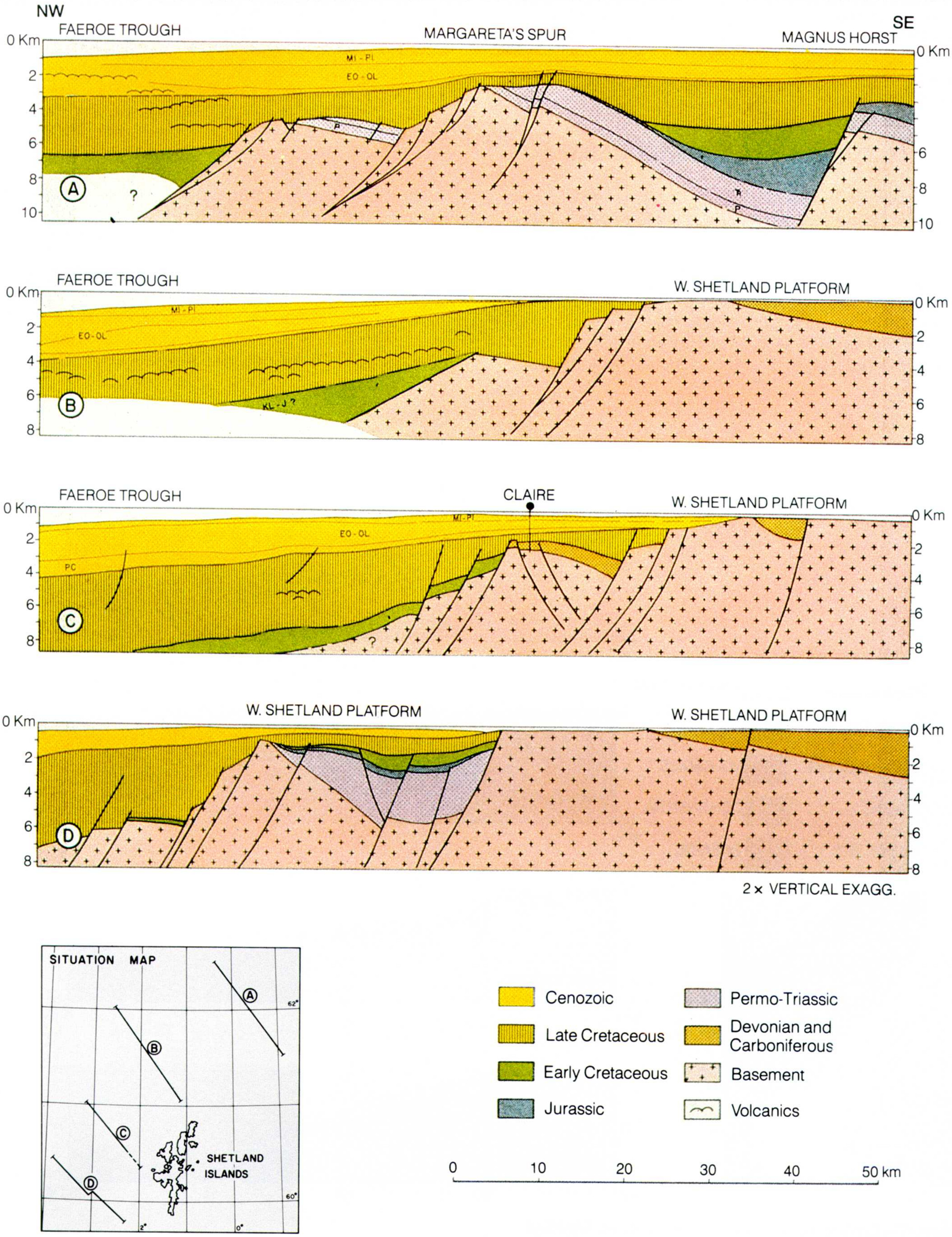


Figure 47—Evolution of the Labrador Shelf Basin. Modified after Umpleby (1979).


NW
SE
FAEROE TROUGH
MARGARETA'S SPUR
MAGNUS HORST
0 Km
2
4
6
8
10
MI - PL
EO - OL
A
B
C
D
W. SHETLAND PLATFORM
KL - J ?
CLAIRE
PC
2 × VERTICAL EXAGG.
SITUATION MAP
SHETLAND ISLANDS
62°
60°
2°
0°
Cenozoic
Late Cretaceous
Early Cretaceous
Jurassic
Permo-Triassic
Devonian and Carboniferous
Basement
Volcanics
0
10
20
30
40
50 km

◀ *Figure 48—Schematic structural cross sections of the northern part of the Faeroe-West Shetland Trough. After Shell UK Expro.*

Sound Rift continued to subside during the Late Cretaceous and Paleocene (Beh, 1975; Miall et al., 1980; McWhae, 1981; Price and Shade, 1982). The Lancaster Sound Graben may extend westward into the Parry Channel fracture zone (Kerr, 1981b). It is also likely that the aeromagnetically defined Melville Bight Graben on the Northwest Greenland Shelf continued to subside differentially during this time (Henderson, 1976). The late Cenomanian to Turonian extrusion of tholeiitic basalts in northern Axel Heiberg Island (Sweeney, 1985), and the occurrence of time equivalent bimodal intrusives and extrusives in northern Ellesmere Island (Trettin and Parrish, 1987) in the onshore prolongation of the Alpha Ridge, may be related to a last phase of transform movements along the latter.

Sea-floor spreading in the Labrador Sea and crustal extension in the Baffin Bay involved a counterclockwise rotation of Greenland relative to the North American Craton (Fig. 55; Beh, 1975; Kerr, 1981b; Price and Shade, 1982). This induced compressional deformations in northeastern Ellesmere Island and with this the onset of the Eurekan orogeny (Balkwill, 1978; Kerr, 1981a, 1981b). In the Sverdrup Basin, this is reflected by the outbuilding of deltaic fans during the Campanian from the gradually uplifted eastern part of Ellesmere Island. During the Maastrichtian and Paleocene, compressional deformations intensified and affected much of Ellesmere Island and also Axel Heiberg Island. At the same time, deltaic clastics spread over much of the northern part of the Canadian Arctic Archipelago (Miall, 1981, 1984a, 1984b). Contemporaneous compressional deformations are also evident on the north coast of Greenland (Håkansson and Pedersen, 1982).

During the Turonian, and possibly even earlier, the Arctic seas ingressed the Baffin Bay Rift as indicated by the occurrence of corresponding marine faunas on Núgssuag Peninsula (Disko Bay area, central West Greenland; McKenna, 1983). Arctic faunas appeared on the Labrador Shelf for the first time during the Campanian while Atlantic faunas reached the Disko Bay area at the same time. In the latter, Maastrichtian faunas reflect a mixture of Arctic and Atlantic waters (Gradstein and Srivastava, 1980; McKenna, 1983).

This suggests that during the Turonian and early Senonian, the Davis Strait High (Srivastava, 1983) formed an effective barrier between the Arctic-dominated Baffin Bay sea arm and the Atlantic-dominated Labrador Sea (Plate 16). This barrier apparently broke down during the Campanian early phases of sea-floor spreading in the Labrador Sea. This new connection between the Arctic and the Atlantic Seas became constricted, however, and finally interrupted during the Paleocene and Eocene as a consequence of the uplift of the Eurekan fold belt in the eastern Arctic Archipelago (Plates 17, 18).

The occurrence of tuffaceous material in the Late Cretaceous sediments of central West Greenland reflects volcanic activity, possibly associated with the Ungava transform fault system (Plate 28; Henderson et al., 1981; McKenna, 1983; Rolle, 1985).

NORWEGIAN-GREENLAND SEA RIFT SYSTEM

During the Late Cretaceous, crustal extension across the Norwegian-Greenland Sea Rift was concentrated on its axial parts. Overall it appears, however, that the rate of crustal stretching decreased during the Late Cretaceous and that dike intrusion along the zone of future crustal separation became increasingly important (Bukovics and Ziegler, 1985).

It is also likely that during this time the Iceland Sea Rift, located between the Rockall-Hatton Bank and southeastern Greenland, was active. In view of the limited geophysical data available from this area and its pervasive cover with uppermost Paleocene and lower Eocene volcanics, this hypothesis cannot, however, be supported by concrete evidence (Roberts et al., 1984). On the other hand, the occurrence of Danian shallow marine shales and clastics in the south-central East Greenland Kangerdlugssuaq area lends some support to this concept (Soper et al., 1976; Higgins and Soper, 1981).

Rockall-Faeroe-West Shetland Trough

Although sea-floor spreading had apparently ceased in the southern Rockall Trough during the Campanian (Roberts et al., 1981a), there is evidence for minor Late Senonian igneous activity on the Rockall Bank (Harrison et al., 1979) and along the north coast of Ireland (Slyne Head, Killala Bay, and southwest Donegal; M.D. Max, personal communication). It is uncertain whether in the northern part of the Rockall Trough, which is probably underlain by a drastically thinned continental crust, the Rosemary Bank, Anton Dohrn Bank, and Hebrides Terrace seamounts had already developed to their full dimensions during the latest Cretaceous or only during the Paleocene-Eocene Thulean volcanic period (see Chapter 8).

During the Late Cretaceous and early Paleogene, the Rockall Trough linked up northward with the Faeroe-West Shetland Trough from which it is now separated by the Wyville-Thomson Ridge. This ridge is formed by predominantly Paleocene-Eocene volcanics (Fig. 51; Roberts et al., 1983).

In the Faeroe-West Shetland Trough, many faults, controlling the base Cretaceous block-faulted relief, die out upwards in Upper Cretaceous shales. Master faults associated with the margins of this rift remained active, however, into Paleocene times. Particularly in the northern parts of the Faeroe-West Shetland Trough, reflection seismic data indicate the occurrence of volcanics and sills within the Upper Cretaceous series (Fig. 48; Plate 27). Some of these have been calibrated by wells and found to consist of dolerites yielding a Campanian to Maastrichtian age. Toward the northwest, the Faeroe-West Shetland Basin is separated from the Faeroe Basin by a geophysically defined major lineament that is partly overlain by Upper Cretaceous shales (Fig. 36). Previously, this lineament was interpreted as an intrusive body, possibly of Campanian-Maastrichtian or Paleocene-Eocene age (Ridd, 1983). Recent seismic evidence suggests, however, that this lineament consists of a heavily intruded basement block (Duindam and van Hoorn, 1987).

Upper Cretaceous shales and marls attain thicknesses of 3-5 km in the Faeroe-West Shetland Trough. The mid-Paleocene low stand in sea level was accompanied by the progradation of deltaic systems over the flanks of this trough and the accumulation of widespread deep water clastic fans in its axial parts (Fig. 48).

Although Late Cretaceous and Paleocene crustal extension in the Faeroe–West Shetland Trough was accompanied by an increasing level of igneous activity, it is doubtful whether locally crustal separation was achieved in this rift as postulated by Hanisch (1983, 1984b), Price and Rattey (1984), and Bott (1984).

Central and Northern North Sea

In the Central and Northern North Sea, Upper Cretaceous chalks and marls progressively infilled the sea-floor topography of the Viking and Central grabens and onlapped against intrabasinal highs and the graben flanks (Fig. 18; see also Ziegler et al., 1986). Many of the faults that control the Early Cretaceous graben relief die out within the Upper Cretaceous strata, and only a few master faults show continued displacement growth during the Late Cretaceous. Intra-Senonian block faulting is locally evident, however.

It is inferred that in the Central and Viking Graben, Late Cretaceous sedimentation rates somewhat exceeded subsidence rates. Thus, despite generally rising sea levels, water depths decreased gradually at least until Maastrichtian time. However, true shallower water conditions were never established within the axial parts of the North Sea graben system (Hancock and Scholle, 1975; Watts et al., 1980; Hatton, 1986). This is further supported by the occurrence of Paleocene sandstone turbidites in these grabens, which were deposited in response to the intra-Paleocene low stand in sea level (Morton, 1982). Outside these grabens, Late Cretaceous and Paleocene sedimentation rates apparently remained in step with subsidence rates and cyclically rising sea levels. During the Late Cretaceous, the area of sedimentation expanded substantially and encroached on the Fennoscandian Shield and the Scottish Highlands. Late Cretaceous series attain maximum thicknesses of 1000 to 2000 m in the Central and Viking Graben and also in the Danish Basin (Ziegler, 1982a).

Crustal stretching played only a minor role during the Late Cretaceous and Paleocene evolution of the North Sea Basin. Its regional downwarping can be related to lithospheric cooling whereby water loading in response to rising sea levels and sediment loading of the lithosphere had an overprinting effect (Sclater and Christie, 1980; Wood and Barton, 1983).

Mid-Norway Basin and Central East Greenland

In the Mid-Norway Basin, tectonic activity abated during the Late Cretaceous. Minor syndepositional faulting, affecting Late Cretaceous strata, is evident on the Trøndelag Platform but is more important along the shoreward margin of the Vøring Plateau (Figs. 37, 38). In the Vøring and Møre basins, Late Cretaceous pelagic shales and marls attain thicknesses on the order of 2–4 km while on the Trøndelage Platform they hardly exceed a thickness of a few hundred meters. The Nordland Ridge forming the outer margin of the Trøndelag Platform remained emergent during the Late Cretaceous. As the Faeroe–Shetland and Vøring escarpment is approached, reflection seismic data indicate the occurrence of volcanics at increasingly shallower levels and also at an increasing frequency. Although not calibrated by well data, these volcanics are likely to be of Late Cretaceous to Paleocene–Eocene age (Plate 26).

Late Cretaceous and Paleocene series onlap and pinch out on the flanks of the Vøring Plateau and thin toward the Faeroe–Shetland escarpment (Fig. 37). These features presumably formed a system of highs that marked the outer margin of the Møre and Vøring basins. In the area of these highs, seismic resolution is, however, impeded by nearly continuous Late Paleocene–Early Eocene basalt flows (Fig. 38; Bøen et al., 1984; Bukovics et al., 1983; Bukovics and Ziegler, 1985).

In Central East Greenland, limited outcrop data suggest that rifting activity continued during the Late Cretaceous as reflected by the accumulation of turbiditic clastics along fault scarps (Surlyk et al., 1981). There is evidence for an Albian–Cenomanian and a Santonian–Campanian transgression (Haller, 1971). Cenomanian faunas are almost exclusively of the European–Atlantic type whereas younger Cretaceous faunas reflect a mixture of Arctic and Atlantic waters (Birkelund and Perch-Nielsen, 1976).

To what extent the coastal areas of East Greenland were inundated by the Late Cretaceous seas is difficult to determine owing to the Paleogene uplift of the area and the corresponding deep truncation of Mesozoic strata.

Barents Shelf and Wandel Sea Basin

In the Southern Barents Sea area, the Tromsø and Bjørnøya grabens continued to subside rapidly during the Late Cretaceous while the Hammerfest Basin subsided only slowly (Figs. 10, 15, 39). Seismic data indicate that in the Tromsø Basin Upper Cretaceous shales reach a thickness of 2000–3000 m while in the Hammerfest Basin they are generally thinner than 400 m. This suggests that in time, rifting activity concentrated further on grabens subparalleling the zone of future crustal separation.

Late Cretaceous regional subsidence patterns of the Barents Shelf are, however, difficult to determine owing to the Cenozoic truncation of the Upper Cretaceous strata and also owing to their transpressional deformation along the Senja–Svalbard fracture zone (Rønnevik and Jacobsen, 1984; Faleide et al., 1984). Outcrop evidence from Svalbard indicates that a first phase of deformation, involving folding and upthrusting, took place during the Late Cretaceous. This deformation is the expression of space constraints in the motion between Greenland and the Barents Shelf that developed in conjunction with sea-floor spreading in the Labrador Sea and differential crustal extension in Baffin Bay and the southern part of the Norwegian–Greenland Sea Rift (see above). These motions involved the rotation of Eurasia relative to Greenland around a pivot point located in the area of the Senja Ridge (Hanisch, 1984a).

During the Late Cretaceous, the northern parts of the Barents Shelf and probably also the area of the Lomonosov Ridge became uplifted forming together the Lomonosov High (Plate 16). This was accompanied by the intrusion of northwest–southeast-striking dike systems on Franz Joseph Land. Uplift of the Lomonosov High may be related to the inception of the Nansen Rift, which separated the area of the future Lomonosov Ridge from the northern margin of the Barents–Kara Sea Platform.

In the area of the Wandel Sea Basin, dextral strike-slip faulting caused the differential subsidence of a series of pull-apart grabens and their subsequent transpressional deformation. These deformations, which were associated with volcanic activity in northernmost Greenland, can be related to wrench movements along the Senja–De Geer fracture zone (Batten, 1982; Birkelund and Håkansson, 1983; Brown and Parson, 1981; Håkansson and Pederson, 1982).