CHAPTER 6

LATE JURASSIC–EARLY CRETACEOUS CENTRAL ATLANTIC SEA-FLOOR SPREADING, CLOSURE OF NEO-TETHYS, AND OPENING OF CANADA BASIN

INTRODUCTION

The Mid-Jurassic development of a sea-floor spreading axis in the Central Atlantic marked the onset of a new kinematic regime in the Atlantic–Tethys domain, and with this a new chapter commenced in the break-up history of Pangea.

As discussed in Chapter 5, the initial phase of the Pangea break-up, spanning Permian to Mid-Jurassic times, was governed by the southward propagation of the Arctic–North Atlantic and the westward propagation of the Tethys rift systems. It peaked in the development of the Tethys and the Central Atlantic sea-floor spreading axes that linked-up via the Western Mediterranean transform fault system. In contrast, the post-Mid-Jurassic phases of the Atlantic–Tethys plate reorganization were governed by the evolution of the Central Atlantic sea-floor spreading axis and its stepwise northward propagation; this was paralleled by the stepwise opening of the South Atlantic Ocean. The evolution of the Central Atlantic sea-floor spreading axis is suggestive of its association with a major upwelling asthenospheric convective cell that had slowly developed during the Triassic and Early Jurassic.

During the Late Jurassic and Early Cretaceous, the progressive opening of the Central Atlantic caused a major sinistral translation between Africa and Laurasia (Olivet et al., 1984; Livermore and Smith, 1985). This was coupled with the transtensional opening of oceanic basins in the Western Mediterranean domain and a gradual closure of the Dinaric–Hellenic ocean, culminating during the earliest Cretaceous in the collision of the leading edge of the Italo-Dinarid promontory with the southern margin of Fermosarmatia (Plate 13). This collision marked the onset of the Alpine orogeny. During the Early Cretaceous, continued opening of the Central Atlantic and commensurate sinistral translations between Africa and Laurasia went hand in hand with the eastward and westward propagation of the early Alpine collision front, the counterclockwise rotation of the Italo-Dinarid promontory and its partial decoupling from Africa (Plate 14; Chapters 6 and 7).

In the Arctic–North Atlantic rift zone, crustal extension persisted until crustal separation was achieved in its different parts during the Cretaceous and early Cenozoic. In the North Atlantic area, increased tectonic activity, accompanied by rift propagation into the Labrador Sea, preceded the northward propagation of the Central Atlantic sea-floor spreading axis and the intra-Aptian separation of the Iberian microcontinent from Laurasia (see *Opening of the North Atlantic,* this chapter).

Furthermore, the stress system governing the evolution of Northwest Europe changed fundamentally as a consequence of Mid-Jurassic crustal separation in the Western and Central Mediterranean. This is expressed by the abandonment of northeast–southwest-oriented grabens and troughs and the development of new northwest–southeast-trending wrench systems. This late Middle to Late Jurassic "polarization" of the Northwest European rift system indicates that its further evolution was now exclusively governed by stresses related to continued crustal extension across the Arctic–North Atlantic megarift (Ziegler, 1982a; see *Norwegian-Greenland Sea Rift System*, this chapter).

In the Arctic domain, a relatively short rifting phase preceded the Valanginian crustal separation between Laurasia, the Alaska–Chukchi–Chukotka and New Siberian Islands blocks; this was followed by the opening of the oceanic Canada Basin (see *Opening of the Canada Basin*, this chapter).

During the Late Jurassic, the Arctic and Tethys seas were in open communication with each other via the Norwegian–Greenland Sea Rift and the basins of Western and Central Europe (Plate 13). This facilitated a faunal exchange that was only restrained by the vast, shallow carbonate platforms occupying the Tethys shelves (Enay and Mangold, 1982; Enay et al., 1982; Cariou et al., 1985). Although there is paleogeographic evidence for a Neocomian constriction of seaways linking the Canada Basin and the Barents Shelf, the exchange of faunas between the Arctic and Atlantic–Mediterranean provinces continued, mainly via the rift systems of the North Atlantic and the Norwegian–Greenland Sea (Plate 10; Rawson, 1973; Birkelund and Perch-Nielsen, 1976). These tectonically induced constrictions became, however, intensified during the Aptian and Albian, and by late Albian time boreal faunas were completely confined to the Arctic domain (Plate 15; Casey and Rawson, 1973; Owen, 1973; Stevens, 1973b).

The strong tectonoeustatic sea-level fluctuations, which occurred during the Late Jurassic and Early Cretaceous (Vail et al., 1977; Hallam, 1978; Haq et al., 1987), had an important overprinting effect on sedimentation patterns, and it is often difficult to distinguish between their effects and those induced by local tectonics. Glacioeustatic contributions to the observed relative sea-level changes can, however, not be excluded in view of the sedimentary record of the Arctic areas (Epshteyn, 1978; Brandt, 1986; see also Bird, 1986).

TETHYS SHEAR AND EARLY ALPINE OROGENY

Following crustal separation at about 180 Ma the Central Atlantic Ocean opened during the Late Jurassic at the rapid rate of some 3.4 cm/year. During the Early Cretaceous, spreading rates decreased, however, to some 2.3 cm/year (Emery and Uchupi, 1984; Olivet et al., 1984; see also Savostin et al., 1986).

Late Jurassic Evolution of the Tethys Domain

The rapid opening of the Central Atlantic was accompanied by a commensurate sinistral translation of Africa relative to Europe whereby during the Late Jurassic the bulk of these movements was taken up by transtensional deformations in the Alboran and Ligurian–Piedmont-South Penninic Ocean (Plate 13). Paleomagnetic data indicate that the Italo-Dinarid promontory, including the Lucania-Campania block and possibly also the Alboran-Kabylia block, moved essentially in unison with Africa (van den Berg, 1979; van den Berg and Zijderveld, 1982; Horner and Freeman, 1983; Westphal et al., 1986). This suggests that during the Late Jurassic only minor movements occurred along the Maghrebian transform zone. Contemporaneous wrench deformations, however, reactivated the fault systems of the Atlas troughs and induced the intrusion of basic plutons (Plate 28; Harmand and Laville, 1983) while alluvial fans prograded from the Sahara Platform onto the carbonate shelves flanking the Tethys (Stets and Wurster, 1982; Wildi, 1983). In the external Betic Cordillera, transtensional deformations associated with minor volcanic activity accompanied the foundering of the southeastern shelves of Iberia (Azema et al., 1979). In the Alpine domain, the Valais-North Penninic-Intra-Carpathian Trough subsided differentially during the Late

Jurassic in response to transtensional movements that paralleled the rapid opening of the Ligurian–Piedmont–South Penninic Ocean. In this basin the formation of oceanic crust is stratigraphically dated as Late Jurassic (Trümpy, 1980; Durand-Delga and Fontboté, 1980; Lemoine, 1985; De Wever and Dercourt, 1985; Weissert and Bernoulli, 1985; Lemoine et al., 1986). At the same time, limited sea-floor spreading took place in the northern part of the Lagonegro Trough (Beccaluva and Piccardo, 1978), suggesting that differential movements occurred between the Lucania-Campania Platform and the composite Latium–Abruzzi–Apulia–Karst Block.

During the Late Jurassic, the northern shelves of the Tethys were occupied by vast, upward-shallowing carbonate platforms. These flanked the deep water troughs of the Dauphinois and the Penninic zone in which pelagic series accumulated, containing mass flow deposits derived from adjacent carbonate platforms such as the Briançonnais High. Also the Lucania-Campania, the Latium–Abruzzi, Apulia, Karst, and Pelagonia-Golija blocks continued to be occupied by stable carbonate platforms while the Trento and Julia blocks became drowned. Submarine plateaux, such as the Austro-Alpine domain, and intervening troughs were generally sediment starved as reflected by the deposition of condensed series, in part containing radiolarites (Abbate et al., 1970; Kälin and Trümpy, 1977; Channell et al., 1979; d'Argenio et al., 1980; Trümpy, 1980; Winterer and Bosellini, 1981; De Wever and Dercourt, 1985; Lemoine et al., 1986).

The Late Jurassic translation of Africa and the Italo-Dinarid promontory induced a gradual narrowing of the Dinaric–Hellenic Ocean in which sea-floor spreading presumably ceased during the Late Jurassic. Compressional deformations are indeed evident along the western and eastern margin of the Pelagonia–Golija microcontinent onto which ophiolites were obducted during the Late Jurassic and at the transition from the Jurassic to the Cretaceous, respectively (Smith and Spray, 1984). Furthermore, there is evidence for the initiation of a subduction zone along the Balkan–Rhodope (Serbo-Macedonian) margin of the Dinaric–Hellenic Ocean (Burchfiel, 1980; Bonneau, 1982; Dercourt et al., 1985; Mountrakis, 1986). At the same time, compressional deformation of Cimmeria (sensu stricto) and its foreland continued with flysch being deposited in the foredeep flanking the Moesian carbonate platform (Foose and Manheim, 1975; Sandulescu, 1978; Sengör, 1984; Sengör et al., 1984; see also Milanovsky et al., 1984). During the later part of the Late Jurassic, alkaline volcanic activity in the Lesser Caucasus area testifies to a phase of back-arc extension that persisted through Early Cretaceous times (Zonnenshain and Le Pichon, 1986).

This suggests that the Late Jurassic rapid opening of the Central Atlantic and the ensuing translation of Laurasia relative to the African plate, including the Italo-Dinarid promontory, induced space constraints in the Central and Eastern Mediterranean domain where they were compensated by the gradual closure of the Dinaric–Hellenic Basin and the subduction/partial obduction of its oceanic crust. This indicates that in the Dinaric–Hellenic Ocean compressional stresses developing as a consequence of sinistral translations between Laurasia and Africa were able to overpower the western parts of the Tethys sea-floor spreading axis, causing its rapid decay. Conversely, it may be speculated that the decay of the western parts of the Tethys sea-floor spreading axis facilitated the rapid opening of the Central Atlantic Ocean. (This is a classical "chicken and egg" question.)

Early Cretaceous Collision of the Italo-Dinard Promontory with the South European Margin

During the Early Cretaceous, the Pelagonia–Golija Block and the leading edge of the Italo-Dinarid promontory of Africa collided with the Balkan–Rhodope subduction zone (Plate 14). Closure of the Dinaric–Hellenic Ocean along the Vardar suture zone was accompanied by the emplacement of major nappe systems in the internal Dinarides and Hellenides. At the same time, the sub-Pannonian oceanic zone became closed, and deformation of the Dacides Block commenced (Aubouin, 1973; Debelmas et al., 1980; D'Argenio and Alvarez, 1980; Burchfiel, 1980; Mountrakis, 1986; Radulescu and Sandulescu, 1980; Sandulescu, 1984; Dercourt et al., 1986). This collisional event marked the onset of the Himalayan-type Alpine orogeny in this area.

During the Early Cretaceous, the Italo-Dinarid Block, which, due to its collision with Europe, was now no longer able to follow the relative eastward drift of Africa, apparently became partly decoupled from the latter and began to rotate in a counterclockwise direction (van den Berg, 1979; Horner and Freeman, 1983; Westphal et al., 1986). This was accompanied by the gradual closure of Late Jurassic–earliest Cretaceous oceanic basins and the propagation of the Alpine collision front into the area of the Pontides and also into the Carpathians and the Eastern and Central Alps. This is evident by the occurrence of the compressional deformations in the frontal parts of the Austro-Alpine and Penninic nappes and also in the central East Carpathian nappes as well as by the accumulation of extensive synorogenic flysch deposits. In combination with the occurrence of high pressure–low temperature metamorphics in the Central and Eastern Alps and in the Carpathians, this suggests that transtensional opening of the South Penninic Ocean had ceased during the Early Cretaceous. During the Aptian–Albian, development of a south-dipping subduction system accompanied the gradual closure and subduction of the South Penninic Ocean and the development of the first nappe systems (Plate 15; Slazka, 1976; Tollmann, 1978; Frisch, 1979; Faupl et al., 1980; Geyssant, 1980; Homewood et al., 1980; Trümpy, 1980, 1982; Gillet et al., 1986; Debelmas and Sandulescu, 1987).

In the Ligurian–Piedmont Ocean, the production of new oceanic crust under a transtensional tectonic regime decreased at the transition from the Jurassic to the Cretaceous. The extrusion of ophiolitic material continued, however, into the Aptian and Albian (Durand-Delgas and Fontboté, 1980; Trümpy, 1980; Lemoine et al., 1986) and thus illustrates that the Alboran–Piedmont transform system had remained active during this time.

The model given in Plates 14 and 15 for the Early Cretaceous evolution of the Mediterranean area suggests that movements along the Alboran–Piedmont transform zone had drastically decreased during the later parts of the Early Cretaceous and that the rotation of the Italo-Dinarid promontory was accompanied by its partial decoupling from northern Africa along a complex sinistral wrench-fault system extending from Gibraltar eastward through the Ionian Sea into the Eastern Mediterranean area, referred to as the Gibraltar–Maghrebian–South Anatolian fracture zone. For geometric reasons, it is assumed that during its counterclockwise rotation the Italo-Dinarid block became internally deformed by wrench movements, presumably along such preexisting fault systems as those of the Lagonegro and Molise troughs. Such deformations are, however, difficult to establish in view of the severe tectonic overprinting of these areas during the Alpine orogeny (see Chapters 7 and 9).

During the Early Cretaceous, carbonate platforms continued to occupy the stable Lucania–Campania, Latium–Abruzzi, Apulia, Karst, and Anatolia–Taurid platforms with pelagic sedimentation characterizing the intervening submarine plateaux and troughs (Channell et al., 1979; D'Argenio et al., 1980; D'Argenio and Alvarez, 1980).

The Azores fracture zone, delimiting the Central Atlantic Ocean to the north, extended eastward into the Gibraltar–Maghrebian–South Anatolian fracture zone. With the Italo-Dinarid promontory, and probably also the Alboran–Kabylia block, lagging behind the continued relative eastward drift of Africa during the Early Cretaceous, this fracture zone developed into an important transform plate boundary (Plates 14, 15). In the Eastern Mediterranean, this boundary was probably diffuse as reflected by the transtensional opening of limited oceanic basins in the Ionian Sea and along the South Anatolian zone (Michard et al., 1984; Poisson, 1984; Delaloye and Wagner, 1984; Dercourt et al., 1986).

In the Western Mediterranean area, latest Jurassic and Early Cretaceous wrench deformations along the Gibraltar–Maghrebian transform zone were accompanied by the extrusion of pillow lavas in the Maghrebian Trough and alkaline volcanic activity in the Atlas (Plate 28; Durand-Delgas and Fontboté, 1980; Laville and Harmand, 1982; Harmand and Laville, 1983). Wrench-induced reactivation of the Atlas troughs and of the Oran and Morocco Mesetas is reflected by the Neocomian shedding of clastics into the Maghrebian Trough from southern sources. Transpressional deformations may be held responsible for the Aptian–Albian uplift of the Alboran–Kabylia block, the erosion of its sedimentary cover, and the shedding of clastics into the Maghrebian Trough from northern sources (Plate 15; Durand-Delgas and Fontboté, 1980; Stets and Wurster, 1981, 1982; Wildi, 1983).

In summary, the Late Jurassic and Early Cretaceous evolution of the Mediterranean area was governed by the sinistral translation between Africa and Laurasia that was caused by the opening of the Central Atlantic. The resulting earliest Cretaceous oblique, passive collision of the Italo-Dinarid promontory, including the Austro-Alpine domain, with the southern margin of Europe and its ensuing partial decoupling from Africa and subsequent rotation, marked the onset of the Alpine orogenic cycle. This reflects a fundamental change in the plate tectonic setting of the Mediterranean domain.

OPENING OF THE NORTH ATLANTIC

Following the Mid-Jurassic crustal separation between Africa and North America, sea-floor spreading remained confined to the Central Atlantic for some 50 Ma while rifting activity persisted and accelerated in the North Atlantic domain.

The analysis of sea-floor magnetic anomalies indicates that the Central Atlantic sea-floor spreading axis propagated during the Neocomian into the southern part of the North Atlantic. This involved a first phase of counterclockwise rotation of the Iberian block. With the early Aptian crustal separation between Galicia Bank and Flemish Cap and in the Bay of Biscay, the Iberian microcontinent became isolated from Laurasia (Plate 15). During the middle Aptian to Santonian–early Campanian, sea-floor spreading in the Bay of Biscay and in the North Atlantic induced a second phase of counterclockwise rotation of Iberia and at the same time its sinistral translation relative to mainland Europe (Plate 16; Masson and Miles, 1984; Olivet et al., 1984).

In middle Albian time, crustal separation was achieved between the Goban Spur and Flemish Cap, the Porcupine Bank and Orphan Knoll, and probably also in the southern parts of the Rockall Trough (Roberts et al., 1981b; Olivet et al., 1984; Masson et al., 1985).

The stratigraphic record of sedimentary basins flanking the North Atlantic, the Bay of Biscay, and the Labrador Sea is summarized in Plates 27 and 28. These diagrams illustrate the correlation of Late Jurassic and Early Cretaceous depositional cycles and tectonic events recognized in these basins.

North Central Atlantic Shelves

The Late Jurassic and Early Cretaceous evolution of the Moroccan and Nova Scotia shelves, flanking the gradually opening Central Atlantic, essentially followed the pattern of classical passive margins. There is also evidence, however, that these areas were marginally affected by the transform movements along the Azores fracture zone and subsidiary fracture zones in the northern parts of the Central Atlantic.

On the Moroccan Shelf, reef fringed carbonate platforms, established already during the Early Jurassic, kept balance with subsidence rates and eustatically rising sea levels during the Middle and Late Jurassic (Fig. 24). During the late Berriasian–Valanginian to late Albian, a strong influx of clastics from the hinterland dominated sedimentation on these shelves (Jansa and Wiedmann, 1982; Jansa et al., 1984; Winterer and Hinz, 1984). This is probably the combined effect of their thermal downwarping in response to lithospheric cooling of the intra-Berriasian and early Aptian low stands in sea level (Vail et al., 1977) and also of wrench-induced deformation of the Atlas troughs (see above). The reported contemporaneous tensional reactivation of the fault systems of the Mazagan Plateau (northwest Moroccan shelf, Ruellan and Auzende, 1985) may be related to transtensional stresses developing between the Azores and South Atlas fracture zones. In the Essaouira and Tarfaya basins, the Jurassic and Early Cretaceous sedimentary outbuilding of the shelf was accompanied by diapirism of the Late Triassic–earliest Jurassic halites (Fig. 20).

On the Nova Scotia Shelf, the carbonate depositional regime of the Late Jurassic became suppressed during the Berriasian by a massive influx of clastics that persisted until the late Albian (Fig. 25; Jansa and Wade, 1975; Wade, 1981; Jansa and Wiedmann, 1982; Grant et al., 1986). This change in the paleogeographic setting of the Nova Scotia Shelf can only partly be related to its subsidence pattern and the intra-Berriasian and Aptian low stand in sea level. Early Cretaceous basaltic extrusions, ranging in age between 125 and 102 Ma, occur in the Orpheus Graben, which is superimposed on the Chedabukto fault zone. This indicates that this long-standing geofracture became repeatedly reactivated in conjunction with movements along the Azores transform fault zone (Jansa and Pe-Piper, 1985) and intensified rifting activity in the North Atlantic domain. Early Cretaceous alkaline intrusions and dikes occur also in Newfoundland (Helwig et al., 1974; Strong and Harris, 1974), on Anticosti Island (Poole, 1977), in the Monteregian Hills of Quebec (Eby, 1984, 1985; Foland et al., 1986), and in the White Mountains of New England (McHone, 1978). This suggests that during Early Cretaceous opening phases of the north Central Atlantic Ocean large areas of the adjacent Laurentian

Craton became tectonically destabilized whereby intraplate stresses caused the reactivation of the preexisting fracture zones (Bédard, 1985) and triggered a generally short-lived intraplate magmatism. This, combined with a possible thermal doming of the affected areas, could be in part responsible for the strong clastic influx onto the northern Nova Scotia and the southern Newfoundland shelves (Plates 14, 15). Contemporaneous "hotspot" activity has also been recorded from the oceanic domain of the New England seamounts in the western parts of the north Central Atlantic (Swift et al., 1986) and attests to the fact that tectonic instability was not confined to continental cratonic areas only.

The Late Jurassic and Early Cretaceous sedimentary outbuilding of the Nova Scotia Shelf was accompanied by flowage and diapirism of the Late Triassic–Early Jurassic salts and associated growth faulting at shallower stratigraphic levels (Wade, 1981; Friedenreich, 1987; Fig. 21).

North Atlantic Areas

During the Late Jurassic, sedimentation in the grabens on the Newfoundland shelf, in those of the Celtic Sea–Western Approaches area and in the Bay of Biscay rift zone was dominated by a shallow marine mixed carbonate–clastics depositional regime. Deeper water conditions were essentially confined to the axial parts of the Bay of Biscay Rift and possibly also the Tagus Abyssal Plain area (Plates 13, 27, 28).

A regional rift-induced unconformity, straddling the Callovian–Oxfordian boundary, is evident in the Celtic Sea, Bristol Channel, and Western Approaches troughs, in Cantabria, and in the Lusitania Basin. In the Jeanne d'Arc Basin and in the Porcupine Trough, this tectonic pulse was less intense and is reflected by a regressive–transgressive clastic cycle. In the basins of the Southern Grand Banks (Horseshoe, North and South Whale basins), shallow marine sedimentation was continuous across the Middle to Late Jurassic boundary.

During the Late Jurassic, differential subsidence of the individual grabens and troughs of the North Atlantic rift system resumed/continued. Crustal extension across the Celtic Sea and Western Approaches troughs was compensated by sinistral movements along a system of wrench faults extending from their northeastern termination into the Paris Basin. Similarly, crustal extension between Iberia and Newfoundland was probably accompanied by sinistral wrench movements in the Bay of Biscay Rift. The occurrence of Late Jurassic coast parallel dikes in southwestern Greenland (Watt, 1969) and diatremes on the southeastern coast of Labrador (King and McMillan, 1985) suggests that rifting propagated into the southern parts of the Labrador Sea which, by this time, was presumably occupied by a shallow marine basin (Umpleby, 1979). Although it is likely that the Rockall Trough continued to subside differentially during the Late Jurassic, it is questionable whether the Iceland Sea Rift, delimiting the Rockall–Hatton Bank to the west, was already active during this time, as indicated on Fig. 17. In the Lusitania Basin several intra-Late Jurassic rifting pulses are evident; these were coupled with the influx of clastics from western sources and minor alkaline volcanic activity in onshore areas (Antunes et al., 1980).

At the transition from the Late Jurassic to the Early Cretaceous, a major rifting pulse combined with an important, presumably tectonically induced lowstand in relative sea level (Hallam, 1978; Vail and Todd, 1981) affected the entire North Atlantic rift system. The resulting "Late Kimmerian" unconformity is of regional significance and can be recognized in most basins flanking the North Atlantic and the Bay of Biscay (Plates 31, 32). In the North Atlantic area, this unconformity can be related to a phase of regional thermal doming presumably in response to progressive lithospheric thinning preceding crustal separation and the northward propagation of the Central Atlantic sea-floor spreading axis.

Earliest Cretaceous uplift of the southern Grand Banks area (Fig. 28) induced the deep truncation of Jurassic strata in the Horseshoe and North and South Whale basins in which diapirisms of the Upper Triassic–lowermost Jurassic halites had caused considerable deformation of the Middle and Late Jurassic series (Fig. 26). In the South Whale Basin, which is limited to the south by the Azores fracture zone, erosion cut down into Lower Triassic series indicating that uplift of this area was very strong. Uplift of the southern Grand Banks area was associated with the extrusion of basaltic flows in the South Whale Basin and on the flanks of the Grand Banks High (Gradstein et al., 1977). Following its late Kimmerian uplift, the area became tectonically quiescent and started to subside regionally. Lower Cretaceous shallow marine clastics are relatively thin in the Horseshoe and North Whale Basin but expand rapidly in the South Whale Basin toward the continental margin. The intra-Aptian hiatus coincides with crustal separation between Flemish Cap and Galicia Bank and in the Bay of Biscay (Plates 15, 28; McWhae, 1981; Wade, 1981).

The stratigraphic record of the Jeanne d'Arc–Avalon Basin (Fig. 29, Plate 28), located to the southwest of the Flemish Cap, shows clear evidence for a later Kimmerian, an intra-Aptian and an intra-Albian unconformity. The later Kimmerian rifting pulse induced rapid basin subsidence whereby in the axial parts of the basin deeper water sedimentation was continuous across the Jurassic–Cretaceous boundary. As a result of the progradation of major cyclical deltaic fans the basin shallowed on during the Barremian. The intra-Aptian unconformity and, to a lesser degree, also the intra-Albian unconformity, were associated with tensional tectonics. The subsequent evolution of the basin was governed by salt-induced tectonics and its regional downwarping to the north, toward the East Newfoundland Basin into which the Jeanne d'Arc–Avalon Basin opens up. Cretaceous sands form the principal reservoirs of the hydrocarbon accumulations occurring in the Jeanne d'Arc Basin (Fig. 24; McKenzie, 1981; Arthur et al., 1982; Hubbard et al., 1985; Meneley, 1986).

Also on the Atlantic margin of Iberia, a major unconformity separates Jurassic and Lower Cretaceous strata. During the latest Jurassic and particularly during the late Kimmerian rifting pulse, the outer margin of the Lusitania Basin became progressively uplifted, causing the complete erosion of its former sedimentary cover (Fig. 27). Restored to the late Kimmerian unconformity level, the Lusitania Basin has essentially the geometry of a half-graben, controlled by fault systems forming its landward margin. During the Early Cretaceous, clastics were shed from the uplifted offshore high eastward into the Lusitania Basin (Plate 28). At the same time, major alluvial fans prograded into this basin from its eastern, landward margin. Southward, these deltaic complexes interfinger with shallow marine carbonates (Plate 14). Minor volcanic activity persisted into the Valanginian (Antunes et al., 1980). To the west of the outer basement high, delimiting the Lusitania Basin, reflection seismic data indicate the occurrence of several half-grabens, containing apparently deeply truncated Jurassic–Triassic series. These are buried beneath essentially unfaulted Cretaceous and Cenozoic strata (Fig. 27). This indicates that the Atlantic margin of Iberia, similar to the Grand Banks–Flemish Cap area, became thermally uplifted during the earliest Cretaceous rifting stage preceding crustal separation in the North Atlantic. At the same time, Western Iberia as a whole became uplifted and tilted to the east. This is reflected by the Early Cretaceous shedding of continental clastics onto its eastern shelves facing the Tethys on which shallow marine carbonates and shales had accumulated during the Late Jurassic (Plates 13–15).

To the north of the Lusitania Basin, the area of the complex Galicia Bank horst is characterized by an intensely block-faulted structural relief involving Neocomian and older shallow marine synrift series. These fault blocks are covered by unfaulted Albian and younger pelagic series that represent the postrifting sequence (Mauffret et al., 1978; Group Galice, 1979). Detailed seismostratigraphic analyses indicate that this area also became progressively domed up prior to crustal separation, resulting in subaerial erosion of Jurassic and Neocomian

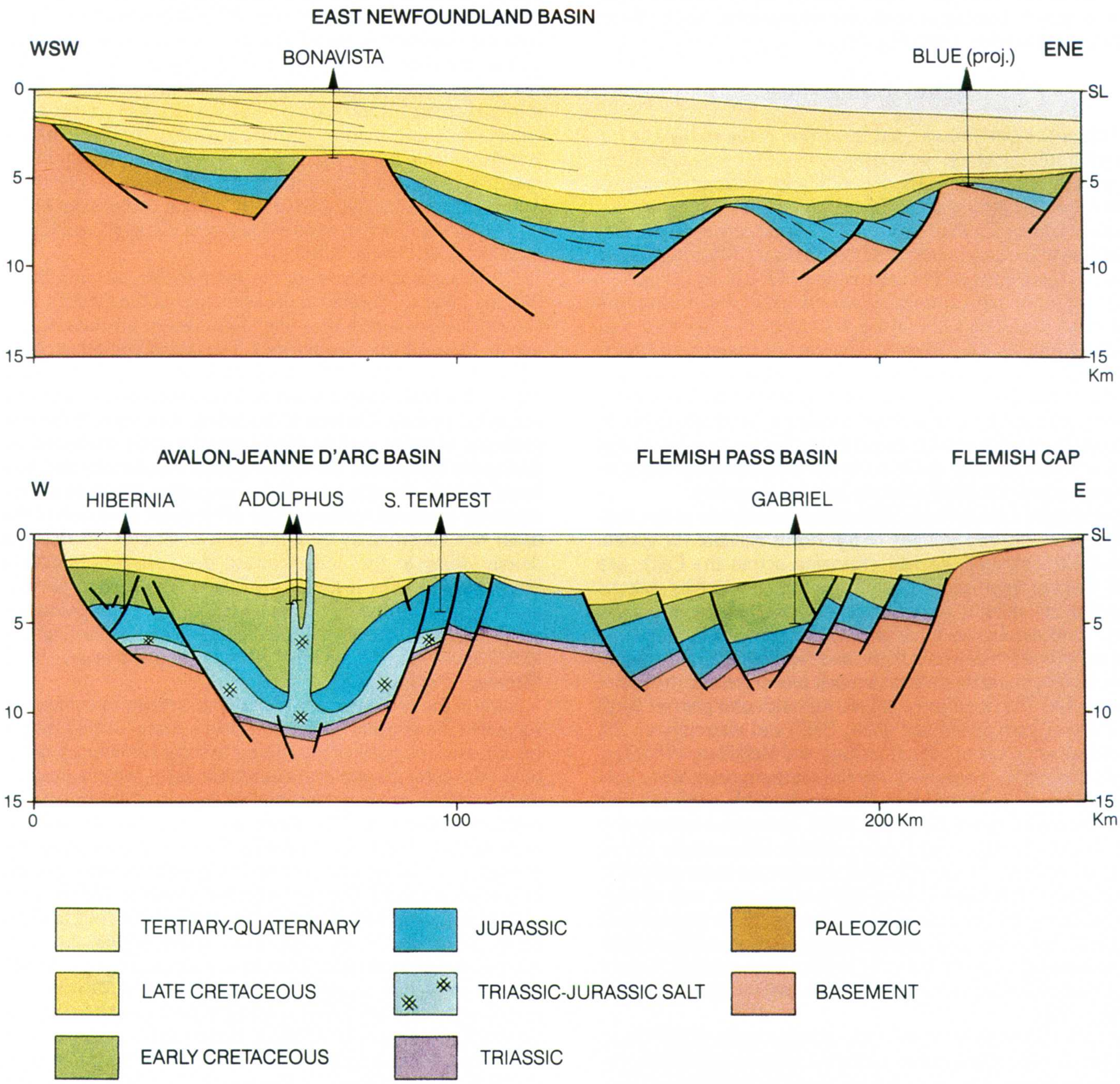

Figure 29—Schematic structural cross sections of the Jeanne d'Arc and East Newfoundland basins. For location see Figure 20. After Levesque (1985) Shell Canada (top); Faught (1986) Shell Canada (bottom).

sediments over the crestal parts of rotational fault blocks (Maufret and Montadert, 1987). Following crustal separation, the North Atlantic margin of Iberia subsided rapidly, largely under sediment-starved condition (Fig. 26).

The continent-ocean transition in the North Atlantic is thought to coincide with the J-anomaly. This suggests that the Flemish Basin and the Tagus Abyssyal Plain are underlain by strongly attenuated continental crust (Boillot et al., 1980; Sullivan, 1983; Masson and Miles, 1984).

Bay of Biscay

In the stratigraphic record of the sedimentary basins flanking the Bay of Biscay, the late Kimmerian rifting pulse is reflected as a major regression as evident by a regionally correlative unconformity (Plate 27). In the area of the Aquitaine Basin, repeated transtensional deformations caused the rapid subsidence of the Parentis and Adour subbasins in which Lower Cretaceous clastics and carbonates reach a thickness of 4000 and 5000 m, respectively (BRGM et al., 1974; Curnelle et al., 1982). In northern Spain, a counterpart to these wrench-induced basins is the Duero Basin in which 4000 m of continental Lower Cretaceous sediments accumulated (Choukroune and Mattauer, 1978; Salomon, 1983). Contemporaneous wrench deformations are also evident in the Celt-Iberian Ranges transecting the Iberian Craton (Vegas and Banda, 1983). Even after the mid-Aptian crustal separation between Iberia and Southern France, wrench deformations persisted into the Albian and Turonian, by which time they were accompanied by volcanic activity (BRGM et al., 1974; Brunet, 1984).

In Cantabria, several phases of transtensional deformations accompanied the Late Jurassic and Early Cretaceous subsidence of the complex, half-graben-shaped Cantabrian offshore basin that was limited to the north by the Le Danois basement horst. Repeated clastic cycles advanced onto the narrow shelves of this basin, in the axial parts of which deeper water conditions were already established during the Late Jurassic. Following the accumulation of thick halites during the latest Jurassic–earliest Cretaceous deeper water conditions were reestablished. Similar to the Aquitaine Basin, wrench deformations, accompanied by repeated volcanic activity, continued in Cantabria and in the Pyrenean domain, even after crustal separation, into Late Cretaceous times (Boillot et al., 1985; see also Emery and Uchupi, 1984).

Intersection of the North Atlantic and Bay of Biscay Rifts

The area at the intersection of the North Atlantic–Bay of Biscay, Celtic Sea, Western Approaches, Porcupine and Flemish Pass rifts apparently became thermally domed up during the late Kimmerian rifting pulse.

In the Celtic Sea and the Western Approaches troughs, the deposition of shallow marine, in part evaporitic carbonates and shales, ceased at the transition from the Jurassic to the Early Cretaceous. A regional erosional hiatus, associated with important wrench deformations, corresponds to the late Kimmerian unconformity (Plate 29). This is illustrated by the palinspastic reconstructions of the Western Approaches Trough, given in Figure 70. Contemporaneous wrench and rift deformations are also evident in the Bristol Channel and the Channel basins and also in the basins of the Irish Sea (Plate 29; Fig. 20). Following the late Kimmerian tectonism, the Celtic Sea and Western Approaches troughs subsided rapidly under a tensional regime (Fig. 22, 23, 75). During the Berriasian to early Aptian, thick continental to lagoonal "Wealden" clastics accumulated in these basins. There is only very limited evidence for syndepositional volcanic activity. During this time, the Celtic Sea Trough was largely isolated from the marine shelves of the Bay of Biscay Rift. On the other hand, the Western Approaches Trough was open to the Bay of Biscay as indicated by the transition of the Wealden clastics into a shallow marine carbonate and shale series near the present-day shelf edge (Ziegler, 1982a, 1987a, 1987b; van Hoorn, 1987a; Robinson et al., 1981; Tucker and Arter, 1987).

During the Early Cretaceous, the Armorican Massif became regionally uplifted and tilted to the northeast. This gave rise to the shedding of the "Wealden" clastics into the Paris Basin (Mégnien, 1980).

Similarly, the area of the Goban Spur, occupying the triangular block at the junction between the Bay of Biscay, Western Approaches, and Porcupine rifts, apparently became uplifted during the earliest Cretaceous. This caused the partial erosion of its Jurassic and Triassic sedimentary cover and a restriction of the marine connection between the Celtic Sea–Fastnet Trough and the Bay of Biscay Rift. At the same time, intense crustal extension along the southern flank of the Goban Spur, locally accompanied by volcanic activity, caused the differential subsidence of a complex array of rotational fault blocks controlled by listric normal faults. Tensional tectonics persisted until crustal separation was achieved between the Goban Spur and the Flemish Cap during the mid-Aptian. Subsequently these fault blocks were covered by relatively thin pelagic series ranging in age from early Aptian to Cenozoic (Fig. 30). Fault block geometries suggest that the crust of the Goban Spur was extended by some 50 km. On the other hand, if it is assumed that during rifting the crust-mantle boundary is not being disturbed, the crustal configuration of the area suggests that the total amount of extension would be as high as 150 km (Fig. 31; Dingle and Scrutton, 1979; Roberts et al., 198lb; Avedik et al., 1982; Masson et al., 1985; Sibuet et al., 1985; de Graciansky et al., 1985; de Graciansky and Poag, 1985; Pinet et al., 1987). This presents a challenge to the pure shear extensional model (McKenzie, 1978) but could be compatible with the simple shear extensional model (Wernicke, 1981; Fig. 74). Alternatively it may be assumed that other processes also contributed to lower crustal attenuation (see Chapter 10).

Mid-Aptian crustal separation in the Bay of Biscay was associated with further wrench deformations and regional uplift of the Celtic Sea and Western Approaches troughs, the Channel area, and probably also the basins of the Irish Sea. With the onset of sea-floor spreading in the Bay of Biscay, these long-standing rifts became tectonically quiescent and their Late Cretaceous evolution followed the pattern of passive margins and aborted rifts (Ziegler, 1982a, 1987a, 1987b; van Hoorn, 1987a; Tucker and Arter, 1987).

The Late Jurassic and Early Cretaceous evolution of the Porcupine Trough differs considerably from that of the rifts of the Celtic Sea–Western Approaches area. In the axial parts of the Porcupine Trough, deeper water conditions were probably established already during the Late Jurassic. The late Kimmerian rifting pulse, inducing further subsidence of this graben along a set of listric normal faults and the progradation of minor clastic fans from its margins, was accompanied by the development of an elongate major volcanic edifice in its south-central parts (Fig. 32). This apparently submarine chain of volcanoes is evident on reflection seismic records and correlates to a major magnetic anomaly (Max et al., 1982). Intra-Aptian, Albian, and Late Cretaceous tensional tectonics were of relatively minor importance and affected mainly the master faults delineating the Porcupine Trough. These rifting pulses accompanied crustal separation between the Porcupine Bank and Orphan Knoll and in the Southern Rockall Trough. Regional southward tilting of the Porcupine Trough during its postrift evolution probably reflects differential crustal thinning in its southern parts. Reflection seismic data refute the earlier held notion that limited sea-floor spreading had occurred in the southern Porcupine Trough (Roberts, 1975; Riddihough and Max, 1976; Naylor and Shannon, 1982; Ziegler, 1982a; Masson and Miles, 1986a).

The structural framework of the East Newfoundland Basin is characterized, at pre-Late Cretaceous levels, by a complex, block-faulted relief involving the basement, Triassic(?), Jurassic, and Early Cretaceous synrift sediments. This relief is overlain by thick Late Cretaceous and Cenozoic postrift sediments,

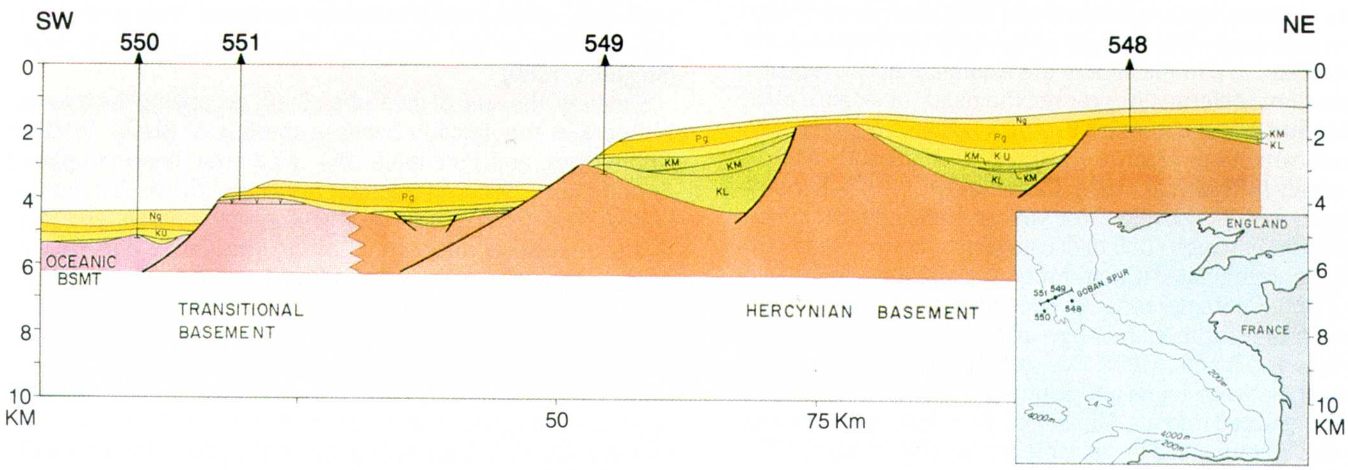

Figure 30—Goban Spur transect, northern Bay of Biscay. After de Graciansky et al. (1985).

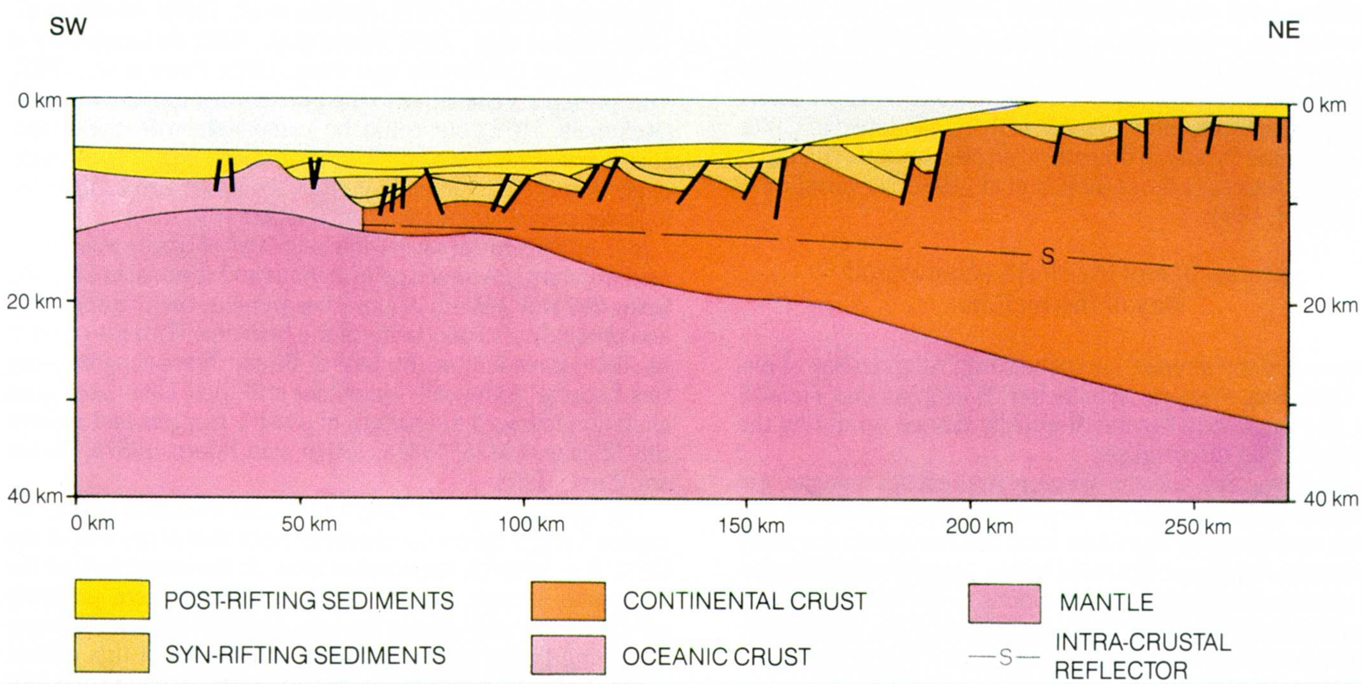

Figure 31—Crustal configuration of the Goban Spur continental margin, northern Bay of Biscay. Location similar to Figure 25. After Avedik et al. (1982).

corresponding to the drift sequence (Fig. 33; Grant, 1975; Keen and Hyndman, 1979; Keen and Barret, 1981). The outer margin of the East Newfoundland Basin is marked by the Flemish Cap and Orphan Knoll basement highs (Fig. 28). There is only limited control available on the age and composition of the synrift series. Upper Triassic to Lower Jurassic sediments, if at all present, are devoid of major halites as indicated by the regional absence of diapiric structures. In analogy with the Jeanne d'Arc–Avalon Basin, Jurassic sediments probably consist in the East Newfoundland Basin of shallow marine carbonates, shales, and minor clastics. The JOIDES No. 111 well, drilled on the Orphan Knoll, bottomed in nonmarine Bajocian sands preserved in a synclinal depression (Laughton, 1972; Umpleby, 1979). The late Kimmerian rifting pulse sharply accentuated the structural relief of the East Newfoundland Basin. Differential subsidence and rotation of individual fault blocks was associated with the erosion of their sedimentary cover. It is unknown whether this erosional phase was of a subaerial or a submarine nature. Rifting activity apparently continued into Aptian time under deeper water conditions. A regional hiatus separates Aptian–Albian(?) strata from the Upper Cretaceous and younger drift sequence.

Reflection and refraction seismic data indicate that the East Newfoundland Basin is underlain by strongly attenuated continental crust (Fig. 33). Under the assumption that during rifting the crust is thinned by stretching only (mass conservation), the crustal configuration of the East Newfoundland Basin suggests that its crust has been extended by over 130% or by some 280 km. The available reflection seismic data suggest, however, that the amount of crustal extension at the base of the Mesozoic sedimentary sequence is of the order of only 100 km. Thus, there is a considerable discrepancy between the amount of extension postulated for this basin on the basis of its crustal configuration and on the basis of fault block geometries at top basement level. In this context, it should be noted that an extension of the crust of the East Newfoundland Basin by some 100 km is required if the tight Permo-Triassic fit of the continents in the North Atlantic domain, as shown in Plates 7–10, is to be accepted.

The northern margin of the East Newfoundland Basin coincides with a fracture zone which, on a predrift fit of the continents, can be traced from the northern end of the Porcupine Trough across the Porcupine Bank (Riddihough and Max, 1976) and onto the shelf of Newfoundland (Grant, 1975). This fracture zone could be considered as the precursor of the intraoceanic Charlie Gibbs fracture zone (Fig. 28).

North of this fracture zone, in the area south of the Cartwright Arch and under the Labrador Shelf, the zone of extended continental crust is much narrower than in the East Newfoundland Basin (Keen and Hyndman, 1979; Grant, 1980). From Figure 34 it is evident that in areas north of this fracture zone the continental crust became extended during the rifting phase by a considerably smaller amount than in the East Newfoundland Basin. Moreover, the available stratigraphic and igneous record suggests that rifting started in the Labrador Sea area only during

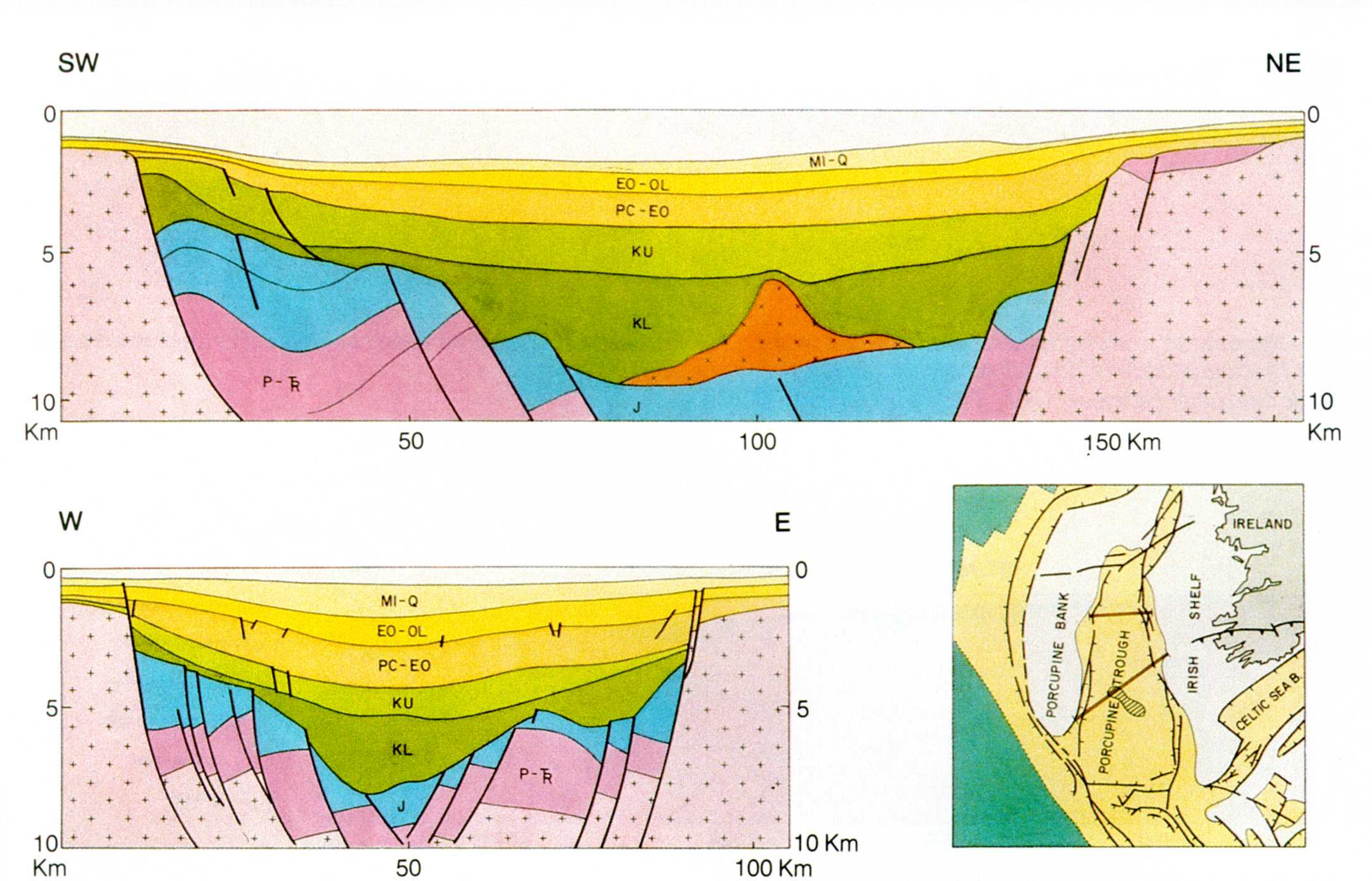

Figure 32—Schematic structural cross sections of the Porcupine Trough, West Irish Shelf.

the Late Jurassic whereas in areas of the East Newfoundland Basin and the Grand Banks area, rifting commenced presumably during the Middle to Late Triassic.

Labrador–Baffin Bay Rift

Wells drilled on the Labrador Shelf indicate that the later Kimmerian rifting pulse strongly affected the Labrador Sea Rift. Differential subsidence of individual graben segments started during the Late Jurassic to Barremian whereby the accumulation of continental clastics was accompanied by intense volcanic activity (McMillan, 1979). Contemporaneous dike intrusions are evident in southwestern Greenland and on the coast of Labrador (Watt, 1969; Wanless et al., 1974; Umpleby, 1979; King and McMillan, 1985). By Barremian time, rifting had propagated northward into the area of the Davies Strait and probably also into the southern Baffin Bay (Henderson et al., 1976, 1981; McWhae, 1981; Srivastava et al., 1981; Klose et al., 1982; Plates 13, 14, 28).

Volcanism abated gradually at the beginning of the Aptian, and the area of sedimentation expanded. Aptian and Albian sediments are coal-bearing, and there is evidence for marine

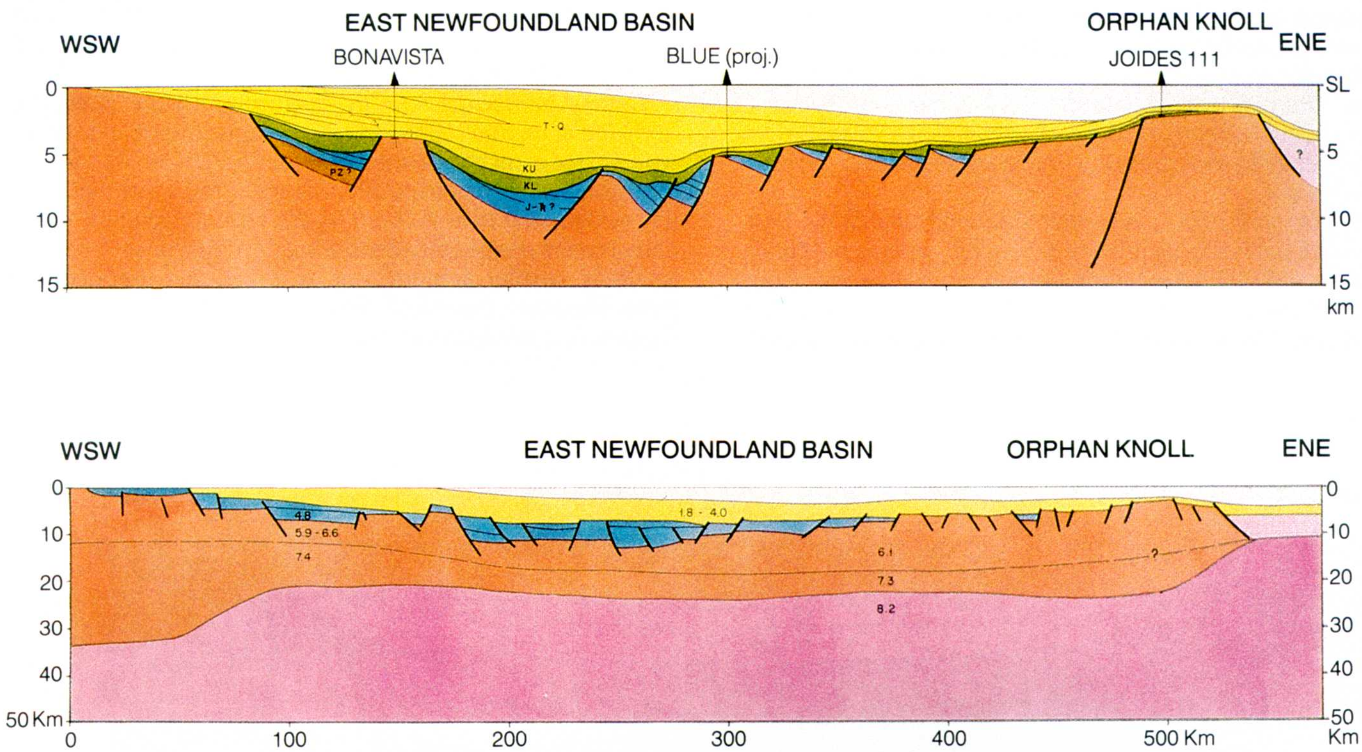

Figure 33—Regional structural cross section and crustal profile of the East Newfoundland Basin based on multichannel reflection seismic data (after Levesque, 1985, Shell Canada) and crustal profile based on refraction data (after Keen and Barrett, 1981).

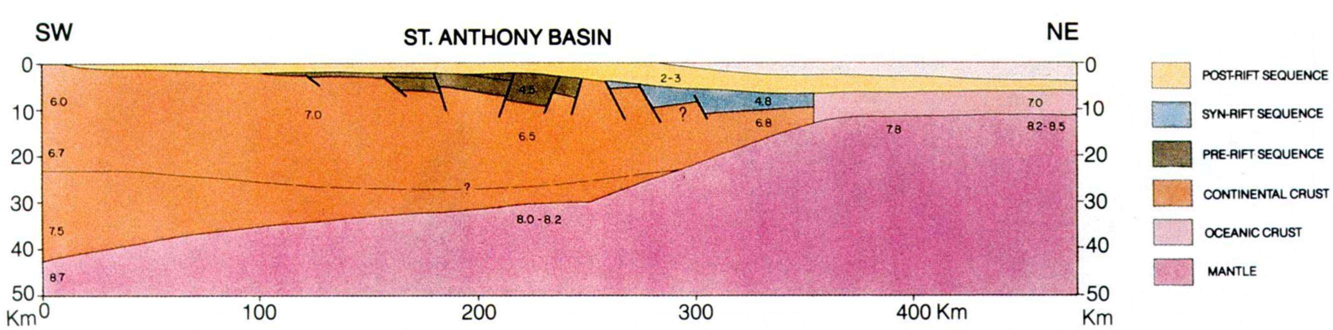

Figure 34—Crustal structure of the St. Anthony Basin–South Cartwright Arch, Labrador Shelf. After Keen and Hyndman (1979).

incursions from the North Atlantic area into the southern Labrador Sea Rift. At the same time, rifting propagated into the northern Baffin Bay and via the Lancaster Sound into the Canadian Arctic Archipelago (Plate 15). This is borne out by the onset of differential subsidence of the Eclipse Trough on northern Baffin Island (McWhae, 1981; Kerr, 198lb; Rice and Shade, 1982). Furthermore, there is evidence that in the northward prolongation of the Baffin Bay Rift, Devon Island and the southern parts of Ellesmere Island became affected by crustal extension (Mayr, 1984 and personal communication). Moreover, the purely geophysically defined Melville Bight Graben, located on the northwestern Greenland Shelf may also have started to subside during the Aptian–Albian (Henderson, 1976).

Assuming a constant rate of rift propagation, this suggests that the over-3000-km-long Labrador–Baffin Bay Rift propagated itself during the Late Jurassic and Early Cretaceous from the North Atlantic area into the southern parts of the Canadian Arctic Archipelago at a rate of some 8 cm/year.

NORWEGIAN–GREENLAND SEA RIFT SYSTEM

During the Late Jurassic and Early Cretaceous, the rate of crustal extension increased in the Norwegian–Greenland Sea rift system. This was coupled with the rapid subsidence of its axial parts in which deep water conditions were established. At the same time, the tectonic evolution of the Northwest European rift systems became increasingly dominated by the stress systems governing the development of the Norwegian–Greenland Sea Rift. Crustal distension in the latter was compensated by sinistral transform movement along the fracture systems of Northeast Greenland and Western Svalbard. In combination with tectonic activity related to the Early Cretaceous opening of the Canada Basin (see *Opening of the Canada Basin,* this chapter) this resulted in the constriction and ultimately the closure of the seaways linking the Arctic Basin with the Norwegian–Greenland Sea area.

North Sea Area

In the North Sea, the Viking and Central grabens developed during the late Middle and Late Jurassic into the dominant rift systems while the importance of the Horda–Egesund Halfgraben diminished and the Horn Graben became essentially inactive (Fig. 35). In the Viking and Central grabens, deeper water conditions were established during the Late Jurassic as a result of continued crustal distension and the progressive foundering of the Central North Sea dome. By mid-Kimmeridgian time,

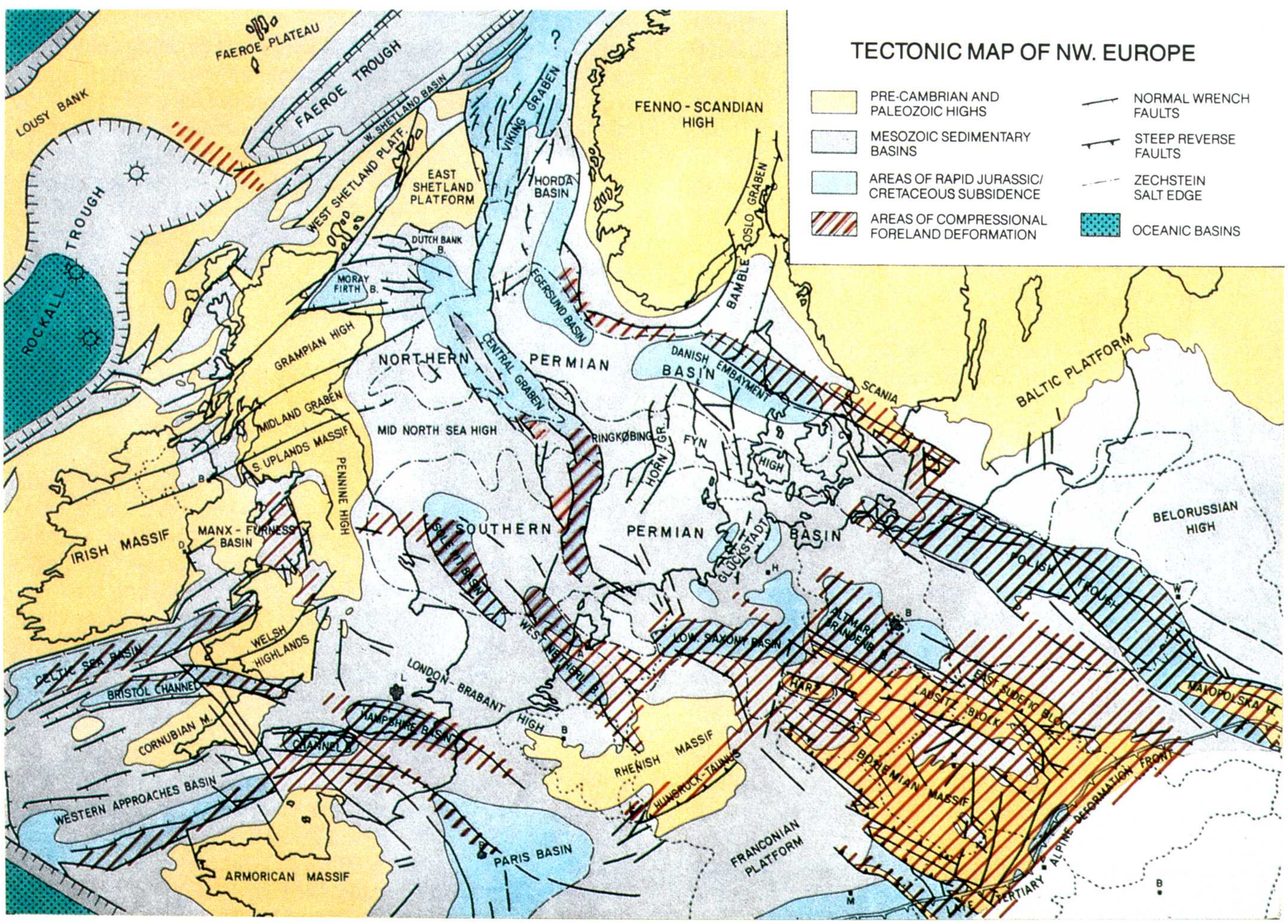

Figure 35—Mesozoic tectonic units of Western and Central Europe.

marine connections were established between the Central and Southern North Sea area across the Mid-North Sea High.

In much of the North Sea area, Kimmeridgian and Tithonian sediments are represented by highly organic shales that form the principal source rock for the hydrocarbon accumulations occurring in the Viking and Central grabens. Upper Bajocian and Bathonian continental clastics and Callovian to lower Kimmeridgian transgressive sands, resting unconformably on deeply truncated Triassic red beds (mid-Kimmerian Unconformity), contain a number of important hydrocarbon accumulations in the Central North Sea. Where these clastics are absent by onlap, Triassic sandstones form an important reservoir. These reservoirs rely for their hydrocarbon charge on Kimmeridgian–Tithonian source rocks (Ziegler, 1980, 1982a; Barnard and Cooper, 1981; Cornford, 1984; Baird, 1986).

Crustal extension across the North Sea rift system was taken up at its southern termination in the Dutch offshore by a system of northwest–southeast-striking wrench faults. Dextral displacements along these induced the differential subsidence of the Sole Pit, Broad Fourteens, West and Central Netherlands, Lower Saxony, Sub-Hercynian, and Altmark–Brandenburg basins (Ziegler, 1982a, 1987c). These basins flank the northern margin of the London–Brabant, Rhenish, and Bohemian massifs. The localization of these wrench-induced basins probably involved the reactivation of Permo-Carboniferous fracture systems. Rapid subsidence of these basins, locally accompanied by repeated igneous activity, went hand in hand with the wrench-induced uplift of the London–Brabant, Rhenish, and western Bohemian massifs (Plates 29, 30; Fig. 12). With this, the long-standing seaway, linking the North European Basin with the Tethys shelves via the area of the Rhenish Massif (Hessian Depression), became closed during the early Late Jurassic (Plates 12, 13). On the other hand, the Bohemian Massif became transected during the Callovian by the Saxonian Strait (Malkovsky, 1976, 1987), which provided a new link between the Tethys and the Lower Saxony Basin. The latter started to subside rapidly during the late Oxfordian, while areas offsetting it to the north became uplifted, thus forming the Pompeckj Swell (Fig. 12, 31). At the same time, a number of north-northeast–south-southwest-trending troughs in northern Germany became inactive (Glückstadt, Emsland, Gifhorn troughs; Betz et al., 1987).

Late Middle to Late Jurassic tectonic activity along the Fennoscandian borderzone is evident by the intrusion of basalt necks in southernmost Sweden (Klingspor, 1976) and by the differential subsidence of half-grabens in northern Denmark and southern Sweden (Liboriussen et al., 1987; Norling and Bergström, 1987). The Polish Rift also continued to subside rapidly during the Late Jurassic. During the Kimmeridgian, marine connections were established between the Tethys shelves and the Central North Sea via the Polish–Danish system of grabens.

Overall, a progressive reorganization ("polarization") of the rift systems of Northwest Europe can be observed during the late Middle and Late Jurassic. In the process of this, northeast–southwest-trending grabens and troughs became inactive (e.g., Horn and Glückstadt grabens, Emsland Trough, Weser Depression; Fig. 35) while new northwest–southeast-striking wrench systems came into evidence (e.g. Sole Pit, West-Netherlands, Lower Saxony, and Altmark–Brandenburg Basins, Saxonian Strait; Fig. 35; compare also Plates 11–13). This is thought to reflect changes in the regional stress systems affecting Northwest Europe. These changes were the consequence of mid-Jurassic crustal separation between Europe and the Italo-Dinarid promontory, the resulting relaxation of tensional stresses associated with the Tethys rift system and continued lithospheric stretching in the Arctic–North Atlantic area (see above; Ziegler, 1982a, 1987c).

At the transition from the Jurassic to the Cretaceous, a major rifting phase affected the graben systems of the North Sea. In the Viking Graben, rapid differential subsidence of rotational fault blocks in response to accelerated crustal distension resulted in the development of a submarine relief of some 1000 m. Highs were swept clean by contour currents while pelagic shales accumulated in the intervening lows in which sedimentation was more or less continuous across the Jurassic–Cretaceous boundary. In the Central Graben, contemporaneous rift tectonics are less obvious since much of the tensional deformation at the pre-Permian level was taken up by plastic deformation in the Zechstein salts (Fig. 18).

In the North Sea area, the largely submarine and in many places composite late Kimmerian unconformity is regionally correlative (Rawson and Riley, 1982). Its origin is probably related to drastic changes in the current regime of the area that were caused by the combined effects of rapid subsidence of the Viking and Central grabens and an important drop in relative sea level (Vail et al, 1977; Vail and Todd, 1981). This change in current regime was apparently associated with a stronger oxygenation of the bottom waters. This is reflected by the regional termination of kerogenous shales deposition during the earliest Cretaceous and the subsequent accumulation of shales having a low organic content.

From a geodynamic point of view, it is interesting to note that in the North Sea, the late Kimmerian rifting pulse, which spanned perhaps less than 10 Ma, was not associated with significant volcanic activity nor with a renewed doming of the Central Graben. In this respect, it differed greatly from the mid-Kimmerian rifting pulse at the transition from the Early to the Middle Jurassic, which was coupled with considerable volcanic activity and the uplift of a major rift dome (see Chapter 5). It can only be speculated that the thermally induced ductility of the lower crust and upper mantle had considerably increased as a consequence of the mid-Kimmerian thermal surge. The upper mantle and the lower crust may therefore have been able to yield by ductile deformation during the late Kimmerian phase of rapid crustal extension. This could have impeded a second phase of lithospheric failure and the intrusion of asthenospheric material to the crust–mantle boundary.

Following the late Kimmerian rifting pulse, the rate of crustal extension across the North Sea graben system decreased gradually. Within the Viking and Central grabens, the throw on many faults decreases and fades out upwards in the Lower Cretaceous shales. On the other hand, master-faults delineating these grabens remained active through Early Cretaceous into Late Cretaceous times (Fig. 18; see also Glennie, 1984b, Fig. 2.14). During the Early Cretaceous, continued differential subsidence of this graben system was accompanied by the gradual infilling of its relief with deep water shales and minor pelagic carbonates ranging in age from Berriasian to Albian (Hancock, 1984).

The Fennoscandian Borderzone remained tectonically active through Early Cretaceous times as reflected by repeated volcanic activity in southernmost Sweden (Plate 30; Prinzlau and Larsen, 1972; Klingspor, 1976) and the shedding of clastic from the Fennoscandian Shield into the Danish Basin (Liboriussen et al., 1987; Bergström and Norling, 1987). Similarly, the Polish Trough continued to subside, in such a way that subsidence and sedimentation rates kept balance; correspondingly, marine connections between the Tethys and the North Sea remained intermittently open (Marek and Raczynska, 1972; Raczynska, 1979; Kemper et al., 1981) (Plates 14, 15).

Early Cretaceous high rates of the crustal extension in the North Sea were coupled with the sharp accentuation of the wrench-induced basins along the northern margin of the Rhenish–Bohemian Massif and a further uplift of the latter. During the Neocomian and Barremian, this high linked up to the northwest with the Pennine High of northern England, thus forming a barrier between the North Sea Basin and the Channel–Paris Basin (Plate 14). During this time, only occasional transgressions originating from the North Sea crossed the English Midlands and advanced southward into the Channel area (Allen, 1976, 1981; Rawson et al., 1978). During the late Aptian and Albian, however, permanent marine communications were reestablished between the Channel area and the North Sea via the English Midlands (Plate 15). Uplift of the Rhenish–Bohemian Massif was accompanied by the Early Cretaceous closure of the Saxonian Strait and wrench deformations along its southwestern margin (Malkovsky, 1987; Schröder, 1987; Nachtmann and Wagner, 1987). Clastics shed northwards from the uplifted Rhenish–Bohemian High were deposited in the adjacent continuously subsiding wrench-induced basins. Contemporaneous deep crustal fracturing gave rise to repeated volcanic activity in these basins, and in the Lower Saxony Basin to the Aptian intrusion of felsic and mafic laccoliths (Stadler and Teichmüller, 1971; Deutloff et al., 1980; van Hoorn, 1987b; van Wijhe 1987; Betz et al., 1987). Tectonic activity in the wrench-induced basins flanking the northern and southern margins of the London–Brabant and Rhenish–Bohemian massifs abated gradually during the Albian and Cenomanian.

Early Cretaceous sands contain important oil accumulations in the West Netherlands and Lower Saxony basins (Ziegler, 1980; Boigk, 1981; Bodenhausen and Ott, 1981).

Rockall and Faeroe–West Shetland Troughs

In the southern and central parts of the Rockall Trough, sea-floor spreading is thought to have commenced during the Albian and to have continued into the Senonian. Subdued magnetic anomalies evident in this area have been related by Roberts et al. (1981a) to anomalies 34 and 33 (see also Bott, 1978). An alternative interpretation, supported by reflection seismic data, has recently been advanced by Megson (1987), who suggests that these anomalies are associated with tilted fault blocks (involving attenuated continental crust?) and interspersed volcanic edifices. The width of oceanic or partly oceanic crust in the southern and central parts of the Rockall Trough could range between 100 and 140 km. Refraction and gravity data indicate for the axial parts of the Rockall Trough a crustal thickness ranging from 5 km in the south to 10 km in the north (Roberts et al., 1983; Scrutton, 1986; Megson, 1987). In the northern parts of the Rockall Trough, seismic definition of its structural configuration at Cretaceous levels is impeded by the presence of extensive uppermost Cretaceous-lower Eocene volcanics that are overlain by up to 3000 km of Cenozoic clastics.

To the north, the Rockall Trough finds its continuation in the Faeroe–West Shetland Trough from which it is now separated by the Wywill–Thomson Ridge, a Paleocene–lowermost Eocene volcanic edifice (Fig. 51; Roberts et al., 1983). Reflection seismic data from the southern parts of the Faeroe Trough, calibrated by wells, indicate that it is floored by highly stretched continental crust (Duindam and van Hoorn, 1987; Fig. 36). This refutes the postulate that crustal extension in this rift had proceeded to crustal separation during the Albian to early Senonian as suggested by Hanisch (1983, 1984a, 1984b) and Price and Rattey (1984; see also Bott, 1984; Bott and Smith, 1984).

In view of the above, it must be assumed that the width of oceanic or para-oceanic crust in the Rockall Trough decreases northward. This implies that limited Cretaceous sea-floor spreading in the southern and central Rockall Trough was accompanied by a counterclockwise rotation of the continental Rockall–Hatton–Faeroe Bank, and that the northern parts of the Rockall Trough are probably floored at least to a large extent by highly attenuated continental crust as suggested by Plates 15–17.

The Faeroe–West Shetland Trough grades northward at its intersection with the Viking Graben, into the Norwegian–Greenland Sea rift system. Seismic and well data from the triple junction between the West Shetland–Faeroe Trough, the Norwegian–Greenland Sea Rift, and the Viking Graben indicate that this area became domed up during the Kimmeridgian. From this high, over which Jurassic and Triassic strata became deeply truncated, clastics were shed southward into the northernmost parts of the Viking Graben where they were deposited as turbiditic sands; these are productive in the Magnus oil field (Plate 27; Fig. 48; De'Ath and Schuyleman, 1981). These sands contain shallow marine faunas which indicate that erosion in their source area was of a subaerial nature. This domed-up area, the dimensions of which are difficult to determine, subsided apparently rapidly during the late Kimmerian rifting pulse.

Similarly, the northwestern margin of the West Shetland Basin, the Sula–Sgeir Platform, and the Faeroe–Shetland Basin became apparently uplifted during the Late Jurassic as erosion cut down into Permo-Triassic red beds and even onto the basement (Fig. 36). Thick Permo-Triassic and Jurassic sediments are only preserved in the relatively narrow half-grabens paralleling the margins of the Shetland Platform and in the Minches Basin (Ridd, 1981; Duindam and van Hoorn, 1987). Uplift of the Faeroe–West Shetland Trough and of the Shetland Platform was accompanied by apparently only minor volcanic activity (Knox, 1977; Ziegler, 1982a) During the Late Jurassic, clastics were shed eastward from the uplifted Shetland Platform into the Morray Firth and southern Viking grabens where they were deposited in submarine fan complexes that form the reservoirs of such oil and gas/condensate fields as Brae, Miller, Tony, and Thelma (Brown, 1984).

The uplifted parts of the Faeroe–West Shetland Trough began to subside during the Kimmeridgian and foundered rapidly during the Early Cretaceous as a consequence of a high rate of crustal distension. Movements along listric normal faults involving the basement continued into the Late Cretaceous (Fig. 36). The marginal half-grabens of this rift system subsided only little during the Early Cretaceous while the Shetland Platform stayed elevated. During the Early Cretaceous, clastics were shed westward from this high into the axial parts of the Faeroe–Shetland Trough where they were deposited as deep water fans along major fault scarps; similarly, clastics were shed eastward into the Morray Firth Graben (Hancock, 1984; Plates 14, 15).

Norwegian–Greenland Sea Area

During the Late Jurassic, the subsidence of the Norwegian–Greenland Rift was governed by continued crustal extension. In combination with eustatically rising sea levels, this led to the establishment of deep water conditions in its axial parts during the Oxfordian and Kimmeridgian (Plates 13, 26). Similar to the North Sea area, Kimmeridgian to early Berriasian shales are highly kerogenous in large parts of the basins of Eastern Greenland, Mid-Norway, and on the Barents Shelf. These shales form the principal source rocks for the oil and gas accumulations of the Mid-Norway Basin (Heum et al., 1986) and the gas fields of the southwestern Barents Sea (Berglund et al., 1986).

In the Mid-Norway Basin, Late Jurassic extensional and transtensional faulting appears to have been more intense in the western parts of the Trøndelag Platform than in its eastern parts

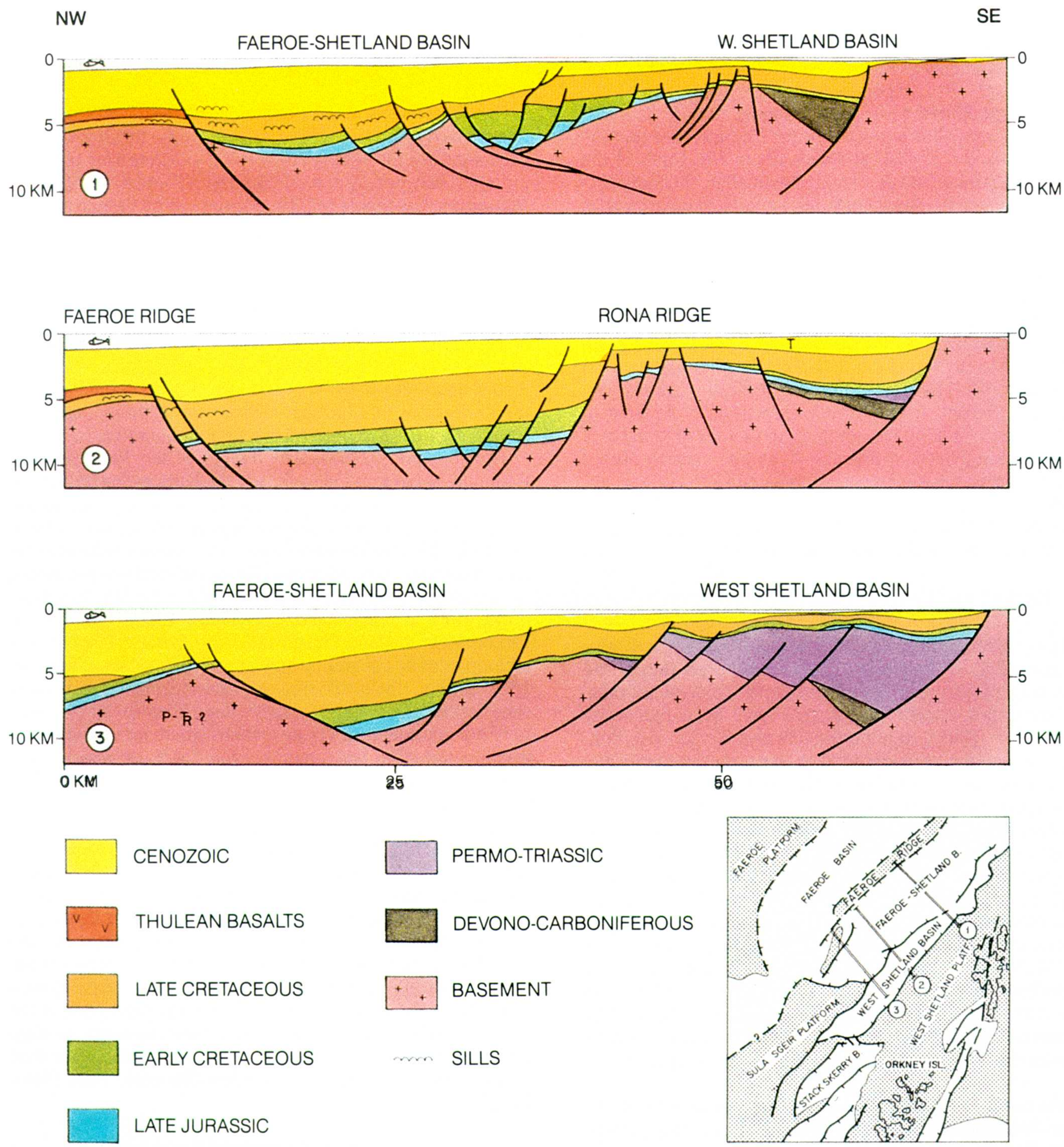

Figure 36—Schematic structural cross sections of the southern Faeroe–West Shetland Trough. After Duindam, Shell UK Expro.

(Figs. 37, 38). During the late Kimmerian rifting pulse the Vøring and Møre basins, forming the axial parts of the Norwegian-Greenland Sea Rift, began to subside rapidly while the Trøndelag Platform was little affected by faulting and subsided only little. At the same time, its western margin became uplifted, presumably in isostatic response to progressive lithospheric thinning. This illustrates that rifting activity was now concentrated in the axial parts of the Norwegian-Greenland Sea Rift (Gowers and Lunde, 1984; Bukovics et al., 1984; Buckovics and Ziegler, 1985). Apart from a thin Kimmeridgian tuff layer observed on Andøya (Dalland and Thusu, 1977; Dalland, 1981), there is no evidence for volcanic activity associated with the Late Jurassic–Early Cretaceous rifting pulses in the Mid-Norway Basin.

In the Møre and Vøring basins, the base of the Cretaceous corresponds to a regional, often angular unconformity and it is unknown whether its origin is related to subaereal or submarine erosion. These basins subsided very rapidly during the Early Cretaceous such that faults offsetting the base of the Cretaceous die out upwards in pelagic Lower Cretaceous shales attaining thicknesses of 3–4 km and more. This sedimentary wedge thins progressively by onlap and condensation toward the zone of future crustal separation along the Faeroe–Shetland and Vøring Plateau escarpments (Fig. 37).

The Late Jurassic and Early Cretaceous evolution of the East Greenland rifted basins is comparable to the development of the Mid-Norway Basin. Continued differential subsidence of major rotational fault blocks led to the establishment of deeper water conditions during the Oxfordian to Kimmeridgian and the accumulation of organic-rich shales. Crustal extension accelerated during the Tithonian to Berriasian with tectonic activity being concentrated on the eastern coast parallel fault zones as indicated by the occurrence of coarse submarine clas-

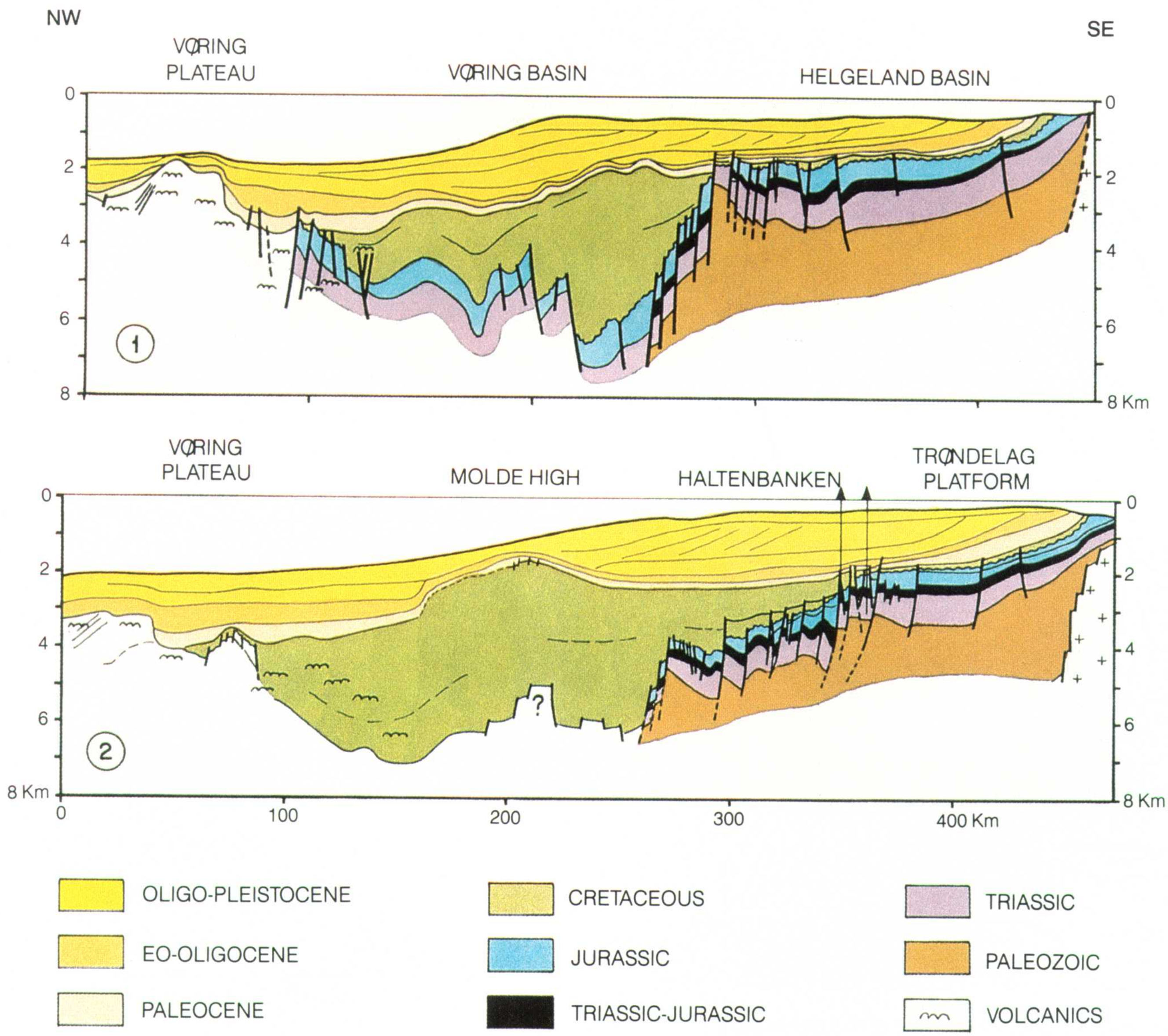

Figure 37—Schematic structural cross sections of the Mid-Norway Basin. For location see Figure 38. After Bukovicz, Norske Shell.

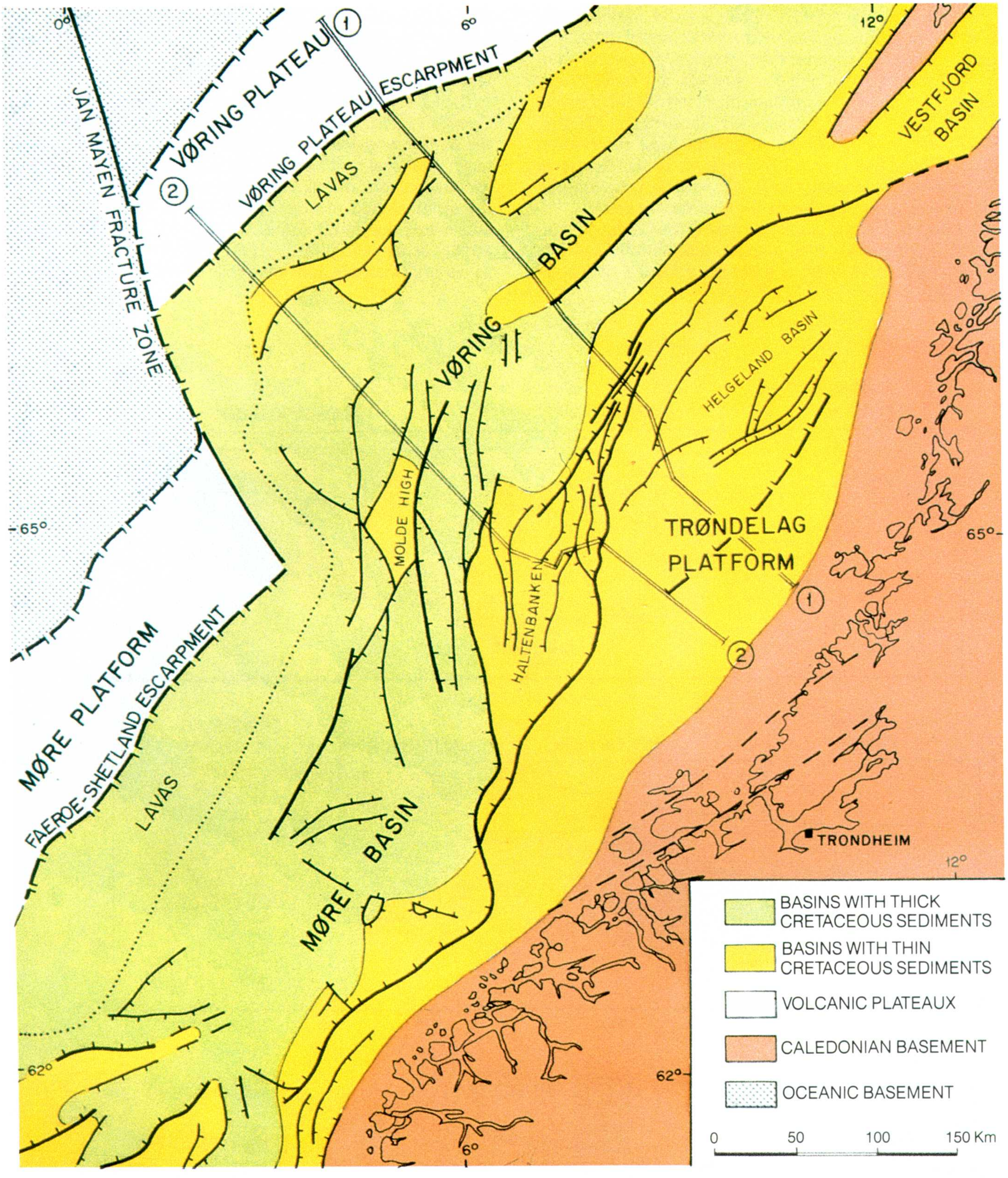

Figure 38—Generalized tectonic map of the Mid-Norway Basin. (After Norske Shell.)

tic fans associated with deep water shales. On the other hand, the Jameson Bay Basin, located further to the west, subsided little during the latest Jurassic and earliest Cretaceous. This is reflected by the rapid progradation of deltaic sands over the upper Oxfordian and Kimmeridgian deeper water shales (Surlyk, 1977a, 1977b, 1984; Surlyk et al., 1981; Surlyk and Clemmensen, 1983; Birkelunde et al., 1984). This also suggests that in Central East Greenland rifting activity gradually concentrated toward the zone of future crustal separation. In this context, a similarity can be seen between the Late Jurassic and Early Cretaceous evolution of the Trøndelag Platform of the Mid-Norway Basin and the Jameson Bay Basin of Central East Greenland (Figs. 37, 50).

In Eastern Greenland, there are only limited outcrops of lower Cretaceous series. These consist of deeper water shales containing in part coarse debris-flow deposits along fault scarps (Surlyk et al., 1981). This suggests that the tectonic and depositional regime, which was established during the late Kimmerian rifting pulse, persisted into the Aptian–Albian (Plate 26). It is likely that during this time the large, aeromagnetically defined basins underlaying the northeastern shelf of Greenland subsided differentially (Henderson, 1976). It is uncertain whether the East Greenland rift system that extends southward between Greenland and the Rockall–Hatton Bank, as shown in Plates 14–16, was already active during the Early Cretaceous. Some support for this concept is, however, lent by the occurrence of thin upper Albian to Danian shallow marine shales and clastics in the Kangerdlugssuaq area of south-central East Greenland (Soper et al., 1976; Higgins and Soper, 1981).

Barents Shelf

On the Barents Shelf, Late Jurassic sediments consist of kerogenous shales as indicated by well results from its southwestern parts and outcrops from the Svalbard Archipelago. Also in the Timan–Pechora area and in the West Siberian Basin, Kimmeridgian to Tithonian (Volgian) sediments are developed in a highly organic, starved basin facies (Rudkevich, 1976; Ulmishek, 1982, 1984; Artyushkov et al., 1986). On Franz Josef Land, Upper Jurassic shales and limestones attain a thickness of some 500 m (Ulmishek, 1982). In the northern parts of the Svalbard Archipelago, Upper Jurassic sediments consist of shallow marine, distal prodelta clays that grade southward into deeper marine, partly kerogenous shales (Smith et al., 1977; Steel and Worsley, 1984; Dypvik, 1984). Overall, it appears that during the Late Jurassic large parts of the Barents Shelf were sediment starved and had subsided well below wave base. These shales form an important hydrocarbon source rock.

In the southwestern Barents Sea, Late Jurassic syndepositional faulting is evident on reflection seismic lines. Rifting activity accelerated sharply at the transition from the Jurassic to the Cretaceous (Rønnevik and Beskowv, 1983; Faleide et al., 1984; Rønnevik and Jacobsen, 1984; Berglund et al., 1986). This is exemplified by the renewed sharp uplift of the Loppa Ridge, which was coupled with downfaulting of the Byørnøya Graben (Figs. 3, 15). At the same time, the Hammerfest Graben became accentuated and the Tromsø Basin began to subside rapidly (Fig. 39). The thickness of the Lower Cretaceous sediments contained in these grabens clearly indicates that the Tromsø Graben, which grades southward into the Harstad Basin, paralleling the coast of northwest Norway, subsided considerably faster than the Hammerfest and Byørnøya grabens. This suggests that also on the Barents Shelf, tensional tectonics gradually concentrated on the axial zones of the Norwegian–Greenland Sea Rift.

In the Svalbard Archipelago, the late Kimmerian pulse of rift and wrench deformation, giving rise to local unconformities, was associated with the intrusion of dolerite dikes and sills (Plate 26). A second phase of volcanic activity is dated as Barremian (Burov et al., 1977). This igneous activity was probably induced by wrench movements along the Svalbard–Northeast Greenland oblique-slip zone. Basaltic flows of a Hauterivian to Albian age cover large parts on Franz Josef Land (Tarakovsky et al., 1983). This volcanic activity was associated with the uplift of the northern margin of the Barents Shelf, the Lomonosov High. The Franz Josef Land hotspot came into evidence during the early phases of sea-floor spreading in the Canada Basin and became extinct with the termination of sea-floor spreading in the latter during the early Late Cretaceous (see *Opening of the Canada Basin,* this chapter). The location of the Franz Josef Land center of igneous activity may coincide with the projection of one of the major sea-floor spreading axes of the Canada Basin into the area of the northern Barents Shelf (see Plates 14, 15). As such, this hotspot could be interpreted as an abortive attempt at splitting the crust of the Lomonosov High at right angles to the strike of the postulated Devonian–Early Carboniferous Lomonosov fold belt. Since the Canada Basin is, however, devoid of clear sea-floor magnetic anomalies, this concept cannot be verified (see *Opening of the Canada Basin,* this chapter).

From the uplifted Lomonosov High, deltaic complexes prograded southward into the area of the Svalbard Archipelago during the Barremian and Aptian. Following a Late Aptian transgression, clastic fans probably built out again over the area and preceded the Late Cretaceous regional uplift of the Lomonosov High and the northern parts of the Svalbard Platform (Birkenmajer, 1981; Steel and Worsley, 1984; Smith et al., 1977; Dypvik, 1984). On much of the Barents Shelf, Early Cretaceous sediments consist of distal prodelta and pelagic clays that have a low organic carbon content.

Wandel Sea Basin

In the Wandel Sea Basin of Northeast Greenland, sedimentation resumed during the Oxfordian after a hiatus spanning Late Triassic to mid-Jurassic time. Transgressive, shallow marine Oxfordian to Tithonian shales are conformably overlain by shallow marine and continental sandstones and conglomerates ranging in age from Berriasian to early Albian (Plate 26). This is taken as evidence for Early Cretaceous increased tectonic activity in the Hinterland. From northeastern Greenland, clastics apparently prograded during the Barremian into the area of Western Svalbard (Håkansson et al., 198lb; Håkansson and Pedersen, 1982; Birkelund and Hakansson, 1983). On the north coast of Greenland, the peralkaline intrusive and extrusive Kap Washington complex came into evidence during the Aptian–Albian and remained active until early Paleocene time (Dawes and Soper, 1979; Brown and Parson, 1981; Batten, 1982). This igneous activity was probably triggered by wrench movements along the Svalbard–Northeast Greenland oblique-slip zone compensating for continued crustal extension in the Norwegian–Greenland Rift.

OPENING OF THE CANADA BASIN

During the Late Jurassic to earliest Cretaceous, the Sverdrup Basin subsided relatively slowly. Cyclical deltaic clastics derived from southern sources shale out basinward where prodelta clays, in part with source rock characteristics, reach a thickness of some 600 m (Plate 26). Late Jurassic and Early Cretaceous basin subsidence was associated with the intrusion of basic dikes indicating renewed tectonic instability (Balkwill, 1978; Trettin and Balkwill, 1979; Balkwill and Fox, 1982). At the same time, increased tectonic activity along the northern margin of the Sverdrup Basin and the Alaska North Slope is reflected by the differential subsidence of fault blocks in the area of the present-day continental shelf and slope and the gradual uplift of the adjacent stable platforms. Reflection seismic data from the northern margin of Alaska indicate its rifted nature and suggest that tensional tectonics were initiated during the Middle Jurassic, intensified during the Late Jurassic, and persisted until Valanginian time (Grantz et al., 1979, 1981; Grantz and May, 1982, 1984; Hohler and Bischoff, 1986; Hubbard et al., 1987). On the Canadian Arctic margin, similar features are less obvious and have been overprinted by Late

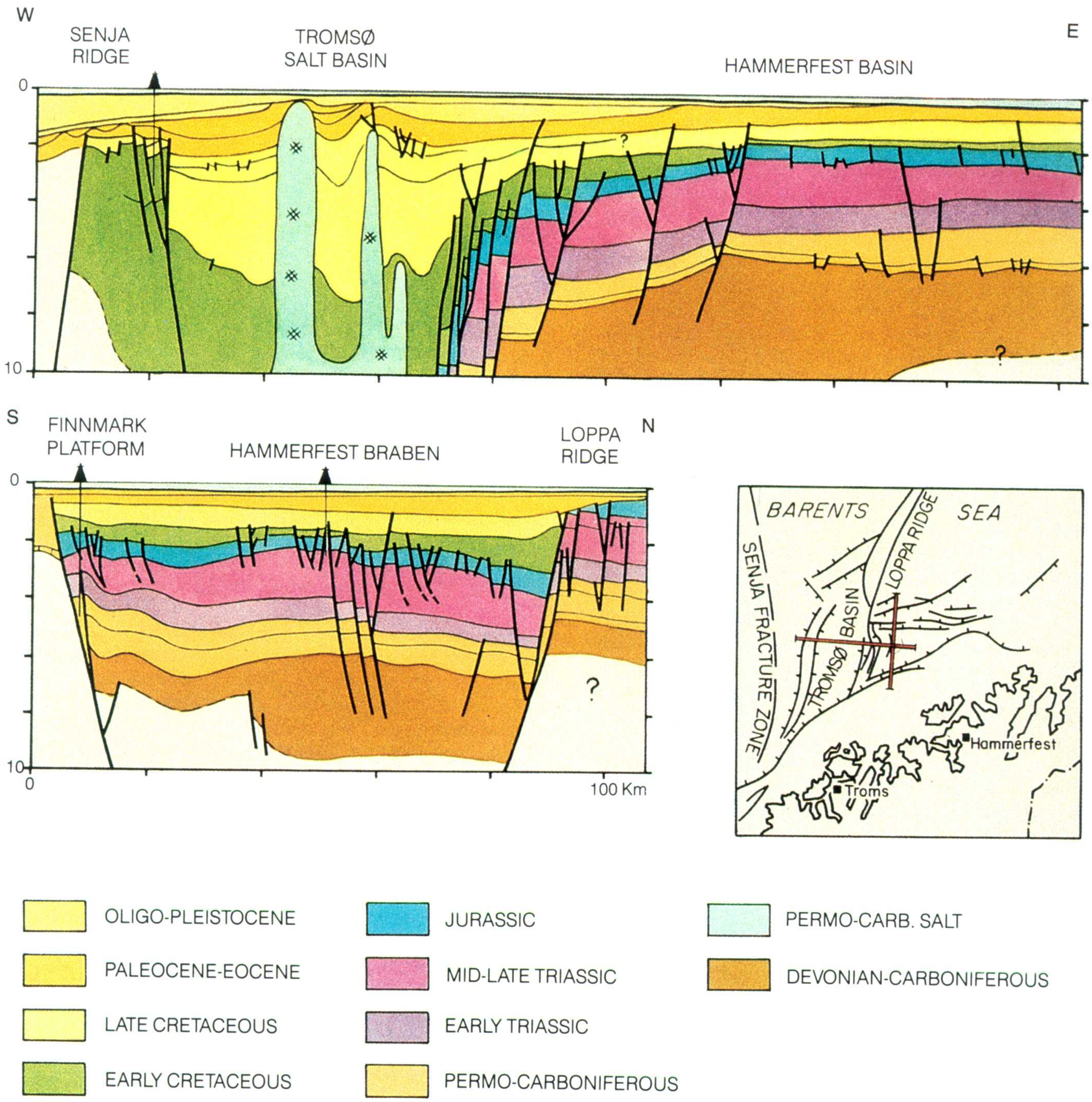

Figure 39—Schematic structural cross sections of the Tromsø and Hammerfest basins, southwestern Barents Sea. After Norske Shell.

Cretaceous tensional and wrench tectonics. This is exemplified by the graben system of Banks Island, which subparallels the present-day shelf edge and which started to subside during Middle and Late Jurassic time (Meneley et al., 1975; Miall, 1975; Kerr, 1981a; Norris and Yorath, 1981; Jones, 1982). On the westernmost New Siberian Islands, turbidites of a Middle to Late Jurassic age were probably deposited in a rifted basin (Fujita and Cook, 1986). This basin, referred to as the Makarov Rift, separated the Lomonosov High from the New Siberian Island block.

Overall, it appears that Late Jurassic and earliest Cretaceous rifting along the Canadian and North Alaskan margins progressed rapidly and culminated during the Valanginian in crustal separation between the northern margin of the Sverdrup Basin, the Alaska–Chukchi–Chukotka, and the New Siberian Islands blocks. Doming of areas flanking the incipient Canada Basin gave rise to the development of a pronounced, regional "break-up unconformity" that is clearly evident on the Alaska North Slope (Fig. 40; Bird and Molenaar, 1983; Magoon and Claypool, 1983; Bird, 1986; Hohler and Bischoff, 1986; Hubbard et al., 1987) and also along the northern margin of the Canadian Arctic Archipelago (Balkwill, 1978; Kerr, 1981b; Balkwill and Fox, 1982; Sweeney, 1985; Sobczac et al., 1986; see also Figs. 13, 14).

Following crustal separation, the Sverdrup Basin continued to subside rapidly with Valanginian to Aptian series attaining a thickness of 1500 m. Tectonic instability continued, however, as indicated by the intrusion of dikes and sills and the extrusion of basaltic flows. During the Valanginian to Aptian, major deltas advanced from the southeast and west into the Sverdrup Basin and to a lesser degree from its northern uplifted margin (Plates 14, 15). During the Aptian and Albian, the northeastern, outer margin of the Sverdrup Basin remained a positive feature. This may be related to persistent transform movements along the Northeast Greenland–Svalbard fracture zone.

In the model proposed here, opening of the Canada Basin involved the counterclockwise rotation of the Alaska–Chukchi block (comprising the Alaska North Slope, the Brooks Range, the Chukchi Sea, and Chukotka) and of the New Siberian Island block, probably representing separate microcratons, away from the Canadian margin (Fig. 7). Sea-floor spreading rates were apparently high with the bulk of the opening of the Canada Basin occurring during the Aptian to Santonian period of single, normal magnetic polarity (see Harland et al., 1982). Furthermore, it is likely that spreading ridge geometries were complex and that several jumps of the sea-floor spreading axes occurred

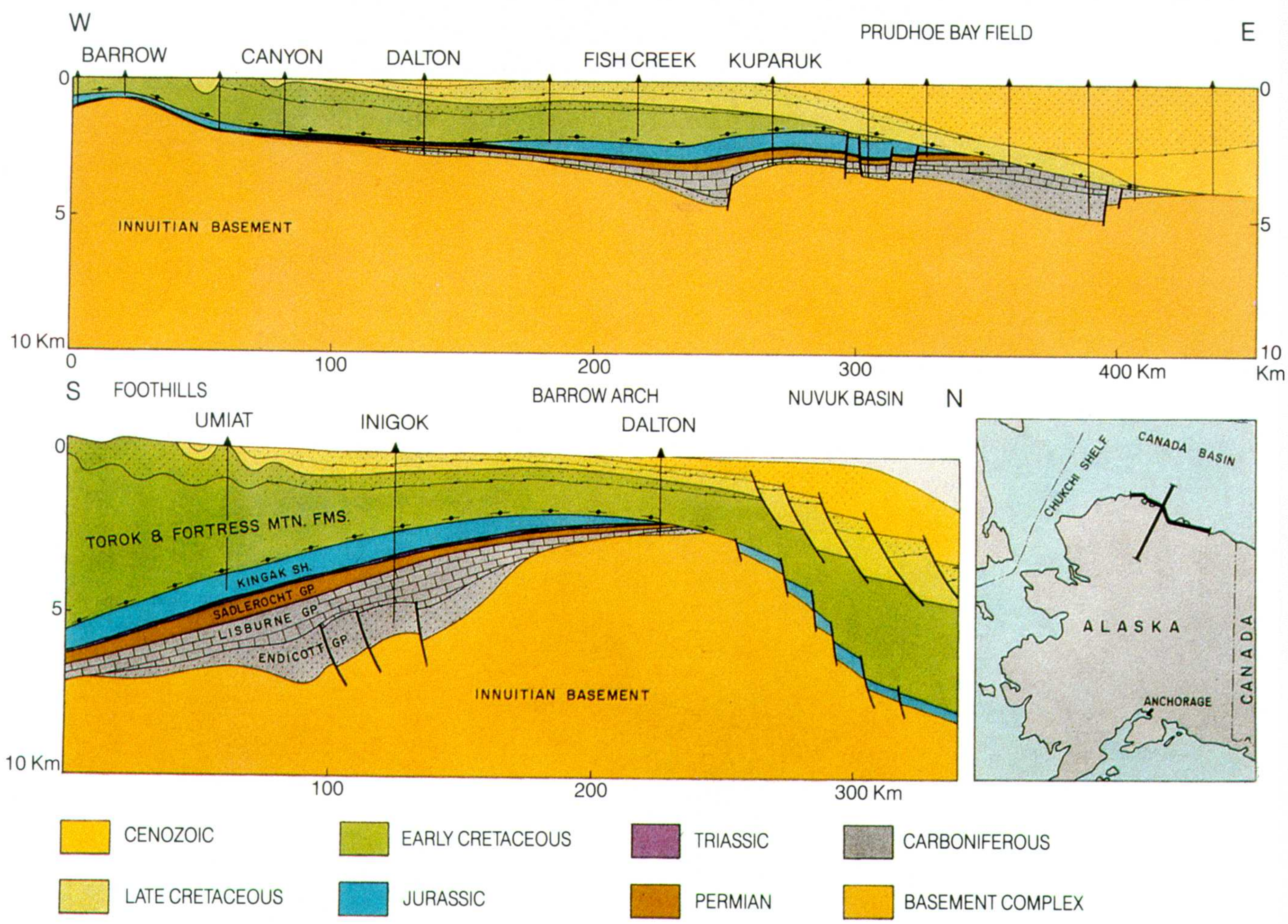

Figure 40—Schematic structural cross sections of the Alaska North Slope. After Bird and Molenaar (1983).

during the northwestward drift and rotation of the Alaska–Chuckchi and the New Siberian Islands continental fragments. This could explain the absence of clear magnetic sea-floor anomalies in the Canada Basin (Taylor et al., 1981; Vogt et al., 1982, 1984; Sweeney, 1985). In this model, the Alpha Ridge is interpreted as representing a hotspot track superimposed on a transform fault that can be compared to the Cenozoic Iceland Ridge (Jackson and Johnson, 1984; Forsyth et al., 1986).

During the early phases of opening of the Canada Basin, the northern margin of the Barents–Lomonosov Platform was probably governed by transform faulting. This, in combination with the development of the Franz Josef Land hotspot, may have induced the sharp intra-Valanginian uplift of the Lomonosov High from which deltaic complexes prograded southward onto the Barents Shelf and into the Svalbard Archipelago during the Hauterivian to Albian (Steel and Worsley, 1984).

The Early Cretaceous rotation of the Alaska–Chukchi block cannot be documented paleomagnetically owing to repeated thermal and diagenetic overprinting of the critical strata (Churkin and Trexler, 1981). The proposed predrift reconstruction of the borderlands of the Arctic Ocean (Fig. 7) is, however, supported by the late Paleozoic and early Mesozoic stratigraphic and tectonic framework of the Alaska–Chukchi block, which is fully compatible with that of the Canadian Arctic Islands. In this respect, the trace of the Innuitian deformation front is a major argument in favor of the intra-Cretaceous rotation of the Alaska–Chukchi Block. The Innuitian deformation front leaves the Canadian Arctic Archipelago in the western part of Prince Patrick Island but is recognized again in the northernmost part of the Richardson Mountains (southwest of MacKenzie Delta; Bell, 1973). In northwestern Alaska, the trace of the Innuitian deformation front loops to the south around the eastern Chukchi Sea block and Point Barrow; for this area, seismic data indicate the presence of north-verging thrusts (Grantz et al., 1981). In northeastern Alaska, the Innuitian fold belt underlies the North Slope and is exposed in the British Mountains straddling the Alaska–Canadian border. Detailed structural analyses in the northern Brooks Range suggest, however, that Devonian thrusting was south-vergent (Oldow et al., 1987). This cannot be easily reconciled with the rotational model of Alaska proposed here for the opening of the Canada Basin (for an alternate model for the opening of the Canada Basin, the reader is referred to Hubbard et al., 1987).

According to the model presented here, the Alaska North Slope–Chukchi and the New Siberian Island blocks passed during the opening of the Canada Basin through a polar position and may have been the locus of grounded ice sheets. It is suspected that the Lower Cretaceous Pebble Shale of the Alaska North Slope is a glaciomarine deposit (Bird, 1986). Furthermore, evidence for cold water deposits during the Early Cretaceous comes from the Sverdrup Basin, Western Spitsbergen, and northeastern Siberia (Epshteyn, 1978; Kemper and Hermann, 1981; Brandt, 1986).