Chapter 8

Crustal Separation Between Eurasia and North America–Greenland: Opening of the Arctic–North Atlantic Ocean

INTRODUCTION

During the late Paleocene, volcanic activity increased sharply in the area of the Rockall–Faeroe Trough, in the southern parts of the Norwegian–Greenland Sea Rift, between the Rockall–Hatton–Faeroe Bank and Greenland, and in the Davies Strait. Additional volcanic centers developed to the north of Ellesmere Island at the junction between the Nansen Rift and the Senja–De Geer fracture zone (Plate 17).

This regional volcanic surge, during which extensive plateau basalts were extruded, is referred to as the so-called Thulean Volcanism. It is the surface expression of the final rifting phase that preceded crustal separation between Greenland and the Rockall–Hatton–Faeroe Bank and Norway, between the Barents–Kara Sea Shelf and the Lomonosov Ridge, and between Greenland and Baffin Island (Talwani and Eldholm, 1977; Srivastava and Falconer, 1982). Following crustal separation and the beginning of sea-floor spreading in these areas, volcanic activity generally abated quickly but persisted till the present in the Iceland hotspot (Vogt, 1983).

In the Labrador Sea, a major change occurred in the location of its sea-floor spreading axis during the late Paleocene between anomalies 25 and 24 (Tucholke and Fry, 1985). This was paralleled by the development of the Reykjanes sea-floor spreading ridge in the northward prolongation of the North Atlantic spreading axis, and of the Aegir and Mohn's ridges in the Norwegian–Greenland Sea. In the Arctic–North Atlantic, the oldest magnetic sea-floor anomaly recognized is anomaly 24. This suggests that crustal separation between Eurasia and Greenland was achieved around 56 Ma during the earliest Eocene (Talwani and Eldholm, 1977; Vogt et al., 1981; Bott et al., 1983; Olivet et al., 1984). With this, the evolution of the Norwegian-Greenland Sea rift system, which had begun at the transition from the Early to the Late Carboniferous, came to a close after some 275 Ma of intermittent crustal extension.

In the Eurasian Basin, sea-floor spreading probably also began between anomalies 25 and 24 or slightly before (Vogt et al., 1981; Srivastava, 1985).

During the Eocene to earliest Oligocene, simultaneous sea-floor spreading in the Labrador Sea–Baffin Bay, in the northern North Atlantic, in the Norwegian–Greenland Sea, and in the Eurasian Basin caused a northward displacement of Greenland. This induced intensified dextral strike-slip movements between northern Greenland and the Barents Shelf along the Senja–De Geer fracture zone and sinistral movements between Greenland and Ellesmere Island along the Nares fracture zone. Northward movement of Greenland was accompanied by its continued counterclockwise rotation relative to North America in response to northward-decreasing rates of sea-floor spreading in the Labrador Sea and Baffin Bay (Fig. 60). This caused intense deformation of the eastern parts of the Sverdrup Basin, corresponding to the main phases of the Eurekan orogeny, and also the deformation of the western margin of the Barents Shelf where the "Alpine" fold belt of Svalbard came into evidence (Eldholm and Thiede, 1980; Rice and Shade, 1982; Srivastava and Falconer, 1982; Miall, 1984a, 1984b; Srivastava, 1985).

Sea-floor spreading along the Mid-Labrador Sea Ridge ceased during the early Oligocene, prior to anomaly 13 ($\pm$35 Ma) (Tucholke and Fry, 1985) in rough coincidence with the termination of the Eurekan orogeny. From then on, Greenland, which had formed an intermediate plate between North America and Europe during the late Senonian to earliest Oligocene, permanently joined the North American plate (Fig. 78). Correspondingly, the Labrador Sea–Baffin Bay represents now an aborted arm of the Arctic–North Atlantic sea-floor spreading system and as such can be compared to the equally aborted southern Rockall Trough and Bay of Biscay oceanic basins.

In the Norwegian–Greenland Sea, activity along the Aegir Ridge stopped during the late Oligocene around anomaly 7 (27–26 Ma), and a new spreading axis, the Kolbeinsey Ridge, developed some 300 km to its west. This important ridge jump entailed the separation of the Jan Mayen microcontinent from the margin of Greenland (Vogt et al., 1981). The Reykjanes and the Mohn's Ridge, on the other hand, continued to be active and maintained their mid-oceanic position (Plate 18).

The onset of sea-floor spreading between northeastern Greenland and the transpressionally deformed western margin of the Barents Platform coincides also with anomaly 13 (36–35 Ma). It was followed by the late Oligocene development of the Knipovich Ridge (Vogt et al., 1981; Myhre et al., 1982). This coincides with the termination of the sea-floor spreading in the Labrador Sea and the translation of activity from the Aegir to the Kolbeinsey Ridge in the Norwegian–Greenland Sea.

With the early Oligocene change in sea-floor spreading patterns and plate boundaries in the Arctic–North Atlantic domain, the earlier constraints in the movement between Greenland and the northwestern tip of Eurasia were removed and the "Alpine" fold belt of Svalbard, similar to the Eurekan fold belt of the Canadian Arctic Archipelago and northern Greenland, became inactive. Intra-oceanic transform motions in the northernmost parts of the Norwegian–Greenland Sea continued, however, along the Spitsbergen and Molloy fracture zones.

From anomaly 7 onward, sea-floor spreading axes in the Arctic–North Atlantic domain had stabilized to their current median position and lithospheric cooling and contraction governed the progressive deepening of the flanking basins (Thiede, 1979; Vogt et al., 1981).

Only by mid-Miocene time had sea-floor spreading in the Norwegian–Greenland Sea and in the Eurasian Basin progressed to the point where the northeastern point of Greenland finally cleared the northwestern shelf edge of Svalbard (Plates 9–21).

During the late Paleocene to early Eocene Thulean volcanic surge, the connection between the North Atlantic and the Norwegian–Greenland Sea became interrupted by the development of the volcanic Iceland–Faeroe land bridge between Greenland and Scotland (Plate 17). Combined with the tectonically induced barriers in Northwest Europe, caused by intraplate compressional deformations, and a low stand in relative sea level, this resulted in the isolation of the North Sea Basin from the Atlantic and Tethys seas. Although temporary connections were reopened during the Eocene and Oligocene between the North Sea Basin and the North Atlantic, the Tethys, and the Donets basins (Plate 18), these were too restricted to permit a significant faunal exchange (see Chapter 9). Correspondingly, the North Sea Basin was dominated by the colder waters of the Norwegian–Greenland Sea until the mid-Miocene reestablishment of a major crossflow between the North Atlantic and the Norwegian–Greenland Sea across the Greenland–Shetland Ridge (Plate 19).

During the Paleocene, compressional deformation of the Canadian Arctic Archipelago caused the interruption of the marine connection between the Arctic Ocean and the North Atlantic via the Baffin Bay and Labrador Sea (Chapter 7). Moreover, marine connections between the Arctic Basin and the

Norwegian–Greenland Sea apparently also became constrained in the area of the De Geer fracture zone by clastic influx from the rising Eurekan fold belt, the development of the volcanic Morris Jessup Rise and the Yermak Plateau, and possibly also by transpressional deformations. This resulted in the nearly total isolation of the Arctic Basin during the Paleocene (Plate 17). This is reflected in the development of endemic faunas that were only able to migrate southward into the Norwegian–Greenland Sea and the North Sea Basin during the late Paleocene and early Eocene (Marincovich et al., 1985). With the early Oligocene transtensional opening of the Svalbard–Greenland Strait, communications between the Arctic Basin and the Norwegian–Greenland Sea became progressively derestricted and permitted the southward flow of the colder Arctic waters. By late Oligocene–early Miocene times, a deep water connection was finally established between these basins (Eldholm and Thiede, 1980).

During the earliest Eocene, a broad interchange of European and North American terrestrial biotas took place across the Canadian Arctic Archipelago, Greenland, and the Greenland–Shetland Ridge. This land-bridge between North America and Europe was largely formed as a direct result of the Thulean development of the Iceland hotspot. This important migration route became, however, permanently interrupted again during the later part of the early Eocene, possibly because of the opening of a marine channel via the Faeroe–Rockall Trough (McKenna, 1983; Hoch, 1983). Thus, a connection between the North Atlantic and the Norwegian–Greenland Sea was reestablished (Berggren and Schnitker, 1983). Substantial southward flow of waters from the Norwegian–Greenland Sea into the North Atlantic across the Iceland–Faeroe Ridge commenced only at the transition from the Eocene to the Oligocene (Nilsen, 1983). By middle to late Oligocene time, a considerable influx of Atlantic warm waters into the Norwegian–Greenland Sea is evident by the occurrence of calcareous oozes on the Vøring Plateau (Berggren and Schnitker, 1983).

By mid-Miocene time, thermal subsidence of the Iceland–Faeroe Ridge had progressed to the point where a massive crossflow between the cold waters of the Norwegian–Greenland Sea and the warm Atlantic waters occurred (Thiede and Eldholm. 1983). This contributed to the development of the ancestral Gulf Stream as recorded by the early middle Miocene ingression of warm water faunas into the North Sea Basin.

In contrast, the vegetation on Iceland retained a distinctly North American affinity until the late Miocene (Friederich and Simonarson, 1981). This suggests that the Denmark Strait segment of the Greenland–Shetland Ridge became permanently submerged only during the latest Miocene. This is in conflict with calculated subsidence curves which suggest that by late Miocene time a water depth of some 400 m was reached in the Denmark Strait (Thiede and Eldholm, 1983). On the other hand, seismostratigraphic analyses of reflection seismic lines extending from Greenland into the Denmark Strait indicate that water depths in this gradually widening seaway increased during the late Miocene (Larsen, 1984).

THULEAN VOLCANISM

The Paleocene to early Eocene Thulean volcanism is unique in the geological history of the Arctic–North Atlantic rift systems. Comparable widespread plateau basalt extrusions did not precede the opening of the Canada Basin and North Atlantic Ocean, nor crustal separation in the Western and Central Tethys. The Early Jurassic volcanism preceding the opening of the Central Atlantic was significant and affected wide areas around future plate boundaries but was apparently not of the same dimension as the Thulean volcanic surge.

Thulean volcanic activity, although concentrated on areas flanking the zones of latest Paleocene–earliest Eocene crustal separation, affected wide areas around these incipient plate boundaries. This is particularly evident on the Rockall–Hatton–Faeroe Bank, which is covered by extensive plateau basalts, in the Rockall–Faeroe Trough where several seamounts came into evidence during the Paleocene, and specifically in the British Isles. Scotland and Northern Ireland are the sites of important extrusive and intrusive centers from which dike swarms extend in a southeastward direction to the shores of the North Sea and across the Irish Sea into Wales and the Midlands. The glaciers of Greenland conceal a possible connection between the volcanic fields of Eastern Greenland and those of central West Greenland (Plate 17).

In the coastal areas of East Greenland, Thulean volcanics and intrusives extend over a distance of some 1300 km (Kent-Brooks, 1980) and, if the data from shelf areas and those from the southwestern margin of the Rockall–Hatton Bank are taken into account (Larsen, 1978; Roberts et al., 1984), over a distance of about 2200 km. On the other hand, considering a Paleocene continent assembly, as given in Plate 17, manifestations of Thulean igneous activity extend from coastal Baffin Island to the shores of the central North Sea over a distance of some 2500 km. The Iceland hotspot is located approximately in the center of this area of volcanic activity.

Regional reviews of the Thulean volcanism have been given by Kent-Brooks (1980), Hall (1981), and McKenna (1983). Summaries of the Early Tertiary igneous activity have been presented for Ireland by Preston (1981); for Scotland by Steward (1965), Walker (1975), Vann (1978), Meighan (1979), and Thompson (1982); for Western Greenland by Clarke and Pedersen (1976); and for Eastern Greenland by Haller (1971), Noe-Nygaard (1976), and Deer (1976). Roberts et al. (1983, 1984) discuss the occurrence of Thulean volcanics on the Rockall–Hatton–Faeroe Bank, and Ridd (1983) and Smythe (1983) review those in the Faeroe–West Shetland Trough. The distribution of Paleocene volcanics in the offshore of Mid-Norway is given by Bukovics and Ziegler (1985; Fig. 38).

The Thulean volcanic surge gave rise to the extrusion of extensive flood basalts that attain thicknesses of 2000–3000 m, the injection of regional basaltic dike systems, and the intrusion of gabbros, syenites, granophyres, and granites. The ascent to shallow crustal levels of felsic intrusives, representing late stage differentiates, is generally confined to the last phases of igneous activity. Thulean extrusives are distinctly mafic-felsic bimodal. Tholeiitic and picritic basalts generally predominate. There is evidence for considerable crustal contamination of extrusive and intrusive rocks, particularly in the British Isles and on the Faeroes (Moorbath and Welke, 1968; Bell, 1976; Dickin et al., 1981; Thompson, 1982; Hald and Waagstein, 1983; Watson, 1985). The regional, earliest Eocene ash-marker in the basins flanking the Norwegian–Greenland Sea and also in the North Sea Basin indicates that the late phases of the Thulean magmatism were accompanied by an intensive explosive volcanism (Plates 26, 27, 30; Jacqué and Thouvenin, 1975; Roberts et al.. 1984).

In Western Greenland, volcanic activity commenced during the late Danian and persisted until the early Eocene. The last dikes are dated as ± 30 Ma (Clarke and Pedersen, 1976; McKinna, 1983). In central East Greenland, the extrusion of plateau basalts commenced around 55.5 Ma and probably ceased during the early Eocene; intrusive activity continued, however, into the Oligocene. On the Faeroe Islands, the age range of basaltic extrusion is uncertain with paleontological criteria and radiometric data, suggesting that volcanic activity spanned early Paleocene to early Eocene times (Tarling and Gale, 1968; Lund, 1983). On the British Isles, the age range of Thulean igneous activity is 66–50 Ma (Bell, 1976; Preston, 1981). Volcanic activity in the Faeroe-Shetland Trough commenced during the Late Cretaceous, intensified during the Paleocene, and abated during the early Eocene (Ridd. 1983). Similarly, volcanic activity began in the igneous province of Northwest Scotland at the transition from the Cretaceous to the Tertiary and persisted into the early Eocene (Curry et al., 1978). Along the southwestern flank of the Rockall–Hatton Bank, a 6 km thick pile of basalt was extruded during the late Paleocene and early Eocene (Roberts et al.. 1984).

Overall, the main phases of the Thulean volcanic pulse lasted for some 10 Ma and spanned late early Paleocene to mid-Eocene times (60–50 Ma). Magnetic sea-floor anomaly 24, the earliest recognized in the northern parts of the North Atlantic and in the Norwegian-Greenland Sea, is dated as about 56 Ma and thus indicates that these oceanic basins began to open during the early Eocene. This indicates that the peak of the Thulean volcanism coincides with crustal separation between Greenland and Eurasia.

Following crustal separation in the Arctic–North Atlantic, hotspot activity was concentrated on Iceland where it persisted to the present. Igneous activity in areas flanking the newly developing Reykjanes and Aegir-Mohn's ridges abated rapidly and became extinct during the early Eocene.

The persistence of intrusive activity in north-central East Greenland into the Oligocene (Rex et al., 1978) can be related to the separation of the Jan Mayen microcontinent from Greenland during the late Oligocene. Similarly, intrusive activity in West Greenland lasting into the early Oligocene reflects transform movements along the Ungava fracture zone during the Eocene to earliest Oligocene phases of sea-floor spreading in the Labrador Sea and Baffin Bay.

The occurrence of major igneous centers on the Rockall–Hatton–Faeroe Platform, in the Rockall–Faeroe Trough, and in Scotland and northern Ireland illustrates that during the Thulean thermal surge, profound mantle disturbances were not confined to areas immediately adjacent to the incipient plate boundaries. Development of these intrusive centers was accompanied by regional doming of the respective areas as indicated by the erosion of the formerly extensive Chalk cover of the shelves flanking the Rockall and Faeroe–West Shetland troughs (Vann, 1978; Preston, 1981; Naylor and Shannon, 1982). It should, however, be kept in mind that the regional deep truncation of the Late Cretaceous series in the area of the British Isles can be attributed to Oligocene broad-scale deformations in response to compressional intraplate stresses. This is evident by the relief of the top Cretaceous surface as extrapolated by the distribution of Upper Cretaceous chert nodules representing erosional remnants (George, 1967; D. G. Roberts, personal communication, 1987; see Chapter 9).

Thermal uplift of areas affected by the Thulean volcanism was associated with transtensional deformation of preexisting fracture systems along which basalt dikes were injected over

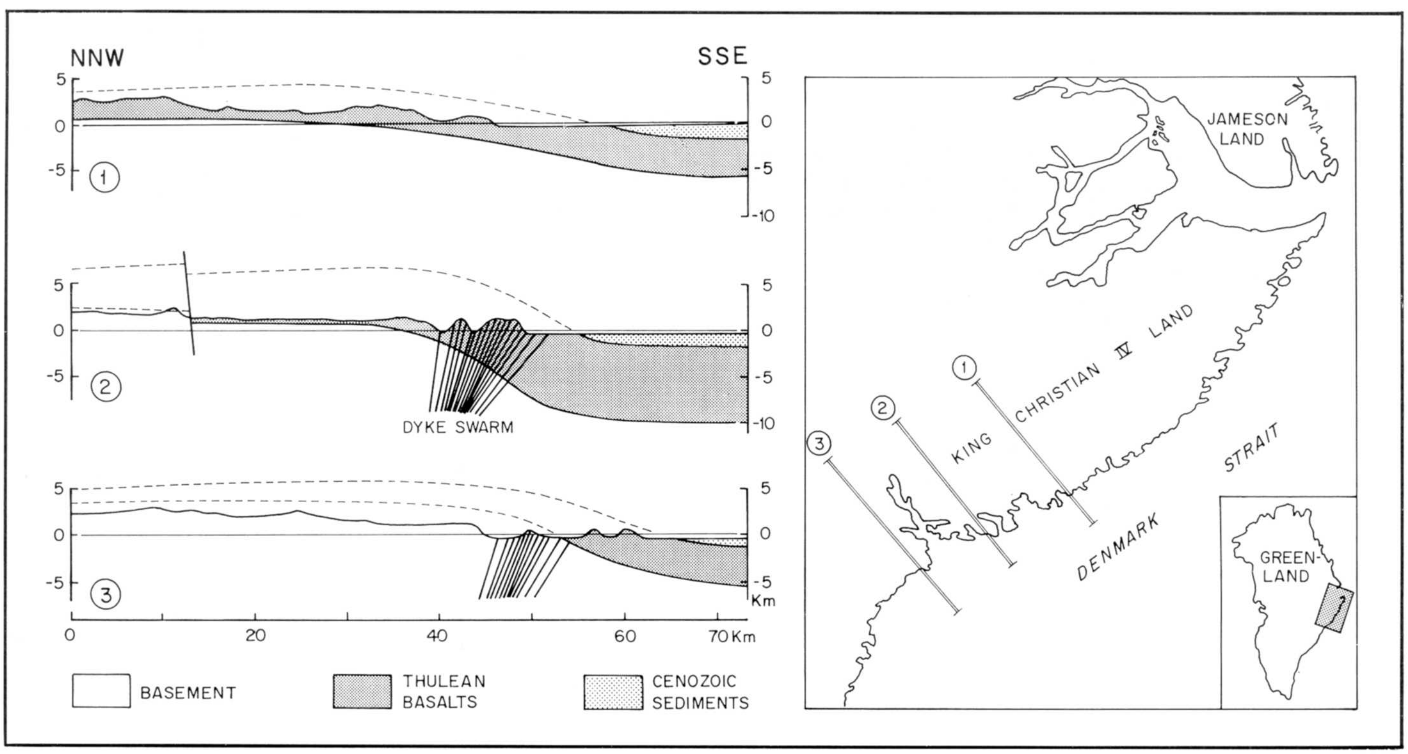

Figure 49—Coastal flexure of King Christian IV Land, Central East Greenland, showing Eocene radiating dyke systems. After Haller (1971).

great distances from the respective intrusive centers (Vann, 1978; Ziegler, 1982a). The doming of these areas, in conjunction with the initiation of volcanic activity, is probably the consequence of a heat surge in the mantle causing the diapiric ascent of hot asthenospheric material to the crust–mantle boundary where it spread out laterally while part of the magmas intruded to shallower crustal levels and ultimately reached the surface as alkali olivine basalts. Progressive permeation of the lower and middle crust by mantle-derived magmas, leading to its partial melting, is reflected by the late stage extrusion of contaminated silicic basalts and the intrusion of felsic plutonic differentiates, containing reactivated crustal material (Thompson, 1982). At the same time progressive thermal upward displacement of the asthenosphere–lithosphere boundary induced a second phase of long wavelength doming of the areas affected by the Thulean volcanism (Vann, 1978).

A similar two-phase doming is evident in the coastal areas of central East Greenland, where Thulean flood basalts, partly extruded near sea level, unconformably overlie truncated Cretaceous and older strata and/or the basement (Noe-Nygaard, 1976). The spectacular coastal flexure of the King Christian IV Land area, which involved the intrusion of fan-shaped dike systems and an uplift of the basalt-covered mainland areas by several kilometers, is dated as mid-Eocene to early Oligocene (Fig. 49; Haller, 1971; Kent-Brooks, 1980). This late uplifting phase, which mainly affected areas with a thick continental crust, was probably induced by the thermal upward displacement of the asthenosphere–lithosphere boundary during the early phases of sea-floor spreading, possibly as a consequence of small-scale convection in the upper mantle (Fleitou and Yuen, 1984). Development of such convective systems presumably preceded the Oligocene separation of the continental Jan Mayen Ridge from the margin of Greenland and the spectacular sea-floor spreading axis jump from the Aegir Ridge to the nascent Kolbeinsey Ridge. The location of the East Greenland coastal flexure may coincide with the transition from little to strongly attenuated continental crust. This cannot, however, be ascertained for want of refraction seismic control. Thus, it can only be speculated that the wavelength and amplitude of this coastal flexure is presumably a function of the width of this crustal transition zone, the crustal thickness of the downflexed block, and the thermal state of the crust during its deformation. Yet the lack of substantial postseparation subsidence of King Christian IV Land and of the Mesozoic Jameson Land half-graben (Fig. 50) suggests that these areas are now upheld by a thickened crust, development of which possibly involved lateral ductile transfer of lower crustal material away from the rift zone as proposed by the model of Moretti and Pinet (1987).

In areas underlain by thick continental crust, the bulk of the Thulean sheet flows were extruded under subaerial conditions, close to sea level (Roberts et al., 1984). The accumulation of several-thousand-meters-thick lava piles exerted a load on the lithosphere and induced concomitant subsidence as evident, for instance, from marine sediments capping the flood basalts of Central East Greenland (Deer, 1976). This area became, however, uplifted by several kilometers during the mid-Eocene–early Oligocene in conjunction with the development of the coastal flexure. From plateau areas, flows prograded into adjacent deeper water basins, such as in the Disko Bay area of West Greenland (Henderson et al., 1981) and the Rockall Trough (Roberts et al., 1983), giving rise to foresetting geometries. Within the Rockall Trough, submarine extrusions resulted in the construction of volcanic edifices that built up to sea level and prograded laterally.

This is evident on the seismic line drawings, given in Fig. 51, which crosses the Rosemary Bank Sea Mount. Similar Upper Cretaceous to Lower Tertiary volcanic buildups are the Hebrides Terrace and Anton Dohrn seamounts in the Rockall Trough, the Wyville–Thomson and Ymir Ridge, separating the latter from the Faeroe–West Shetland Trough (Roberts et al., 1983) and the Erland and Brendan igneous complexes in the Faeroe–West Shetland Trough (Ridd, 1983; Smythe, 1983).

In areas immediately adjacent to the developing plate margins, Thulean volcanics were extruded partly under shallow marine to subaerial conditions and partly in a deeper marine environment. In the Mid-Norway Basin, Paleocene sills and dikes are lodged at increasingly shallower stratigraphic levels as the Vøring and Shetland escarpments are approached. The latter are covered by continuous flows (Fig. 37; Bukovics and Ziegler, 1985). On the outer margin of the Vøring Plateau, and also on the southwestern margin of the Rockall–Hatton Plateau, Thulean volcanics are associated with seaward dipping reflection patterns that coincide with magnetic lineations 24a and 24b. These reflection patterns have been interpreted as resulting from the co-magmatic subsidence of seaward-located extrusive centers that built up to sea level (Hinz and Schlüter, 1978; Mutter et al., 1982; Roberts et al., 1984; Roberts and Ginzburg, 1984; Eldholm et al., 1984). During the early phases of volcanism, fissure and explosive eruptions apparently took place over wide areas, as evident in East Greenland (Larsen, 1978), but in time concentrated on the actual zone of crustal separation. During this phase of impending crustal separation, dike injection probably played a greater role than mechanical stretching of the crust (Royden et al., 1980; Dewey, 1982). This was probably coupled with thermal uplifting of the zone of incipient crustal separation, culminating in subaerial volcanism (see *Faeroe–West Shetland Trough and Mid-Norway Basin*, this chapter).

The latest Cretaceous to mid-Paleocene development of the giant Thulean hotspot, centered on Iceland, presumably reflects accelerated diapirism of lower mantle material to the asthenosphere–lithosphere boundary where it reached its density equilibrium and spread out laterally (Artyushkov, 1983). From this reservoir of hot upper mantle, individual plumes ascended to the crust–mantle boundary and later into the crust and to the surface. The location of these secondary plumes was probably predetermined by anomalies at the base of the lithosphere that developed during the preceding rifting phases (Watson, 1985). During the late Paleocene, continued material transfer through the now firmly established asthenospheric heat conduit resulted in rapid thermal attenuation of the lithosphere, the injection of mantle-derived magmas into the crust, and finally in crustal separation.

The onset of sea-floor spreading in the northern North Atlantic and the Norwegian–Greenland Sea coincides roughly with the relaxation of compressional stresses related to the Laramide phase of foreland deformation in Northwest Europe (Chapter 7). This raises the question of whether compressional stresses, originating from the Alpine collision zone, impeded the tensional separation of North America–Greenland and Eurasia and their drifting apart during the Paleocene. In such a model, the restriction in the movement of these megacratons may at least in part be indirectly responsible for the build-up of the Thulean hotspot. Under unrestricted conditions, crustal separation presumably would have been effected considerably earlier with the thermally upwelling asthenospheric material being accreted at the new plate boundary as oceanic crust.

LABRADOR SEA AND BAFFIN BAY

In the southern parts of the Labrador Sea, sea-floor magnetic anomalies are well defined and readily permit the reconstruction of its spreading history, which terminated around anomaly 13 in the early Oligocene. In the northern parts of the Labrador Sea, and particularly as the Davis Strait is approached, sea-floor magnetic lineations are poorly defined. This is largely due to complex wrench deformations along the Ungava transform zone (Srivastava, 1978; Srivastava et al., 1981, 1982; Tucholke and Fry, 1985).

The western margin of the Labrador Sea is characterized by a Late Cretaceous to recent passive margin prism that prograded onto oceanic basement (Fig. 47). This clastic wedge attains maximum thicknesses of some 10 km under the present shelf edge and partly obscures the oldest sea-floor magnetic anomalies (Umpleby, 1979; McMillan, 1979; Grant, 1980, 1982; McWhae, 1981). In contrast, the Greenland margin of the Labrador Sea is sediment starved. However, a major deltaic complex prograded from Greenland further northward into the Davis Strait during the Eocene to Pliocene. Cenozoic clastics reach a maximum thickness of some 4000 m near the present shelf edge at latitude 67 °N (Henderson et al., 1981; Rolle, 1985).

In the Davis Strait, wrench deformations, compensating for sea-floor spreading in the Labrador Sea, led to the development of a complex system of block-faulted and transpressional anticlinal and synclinal structures (Fig. 52). These features involve oceanic and continental basement. Seismic data, calibrated by wells, indicate that the main phase of deformation occurred during the Eocene and that tectonic activity abated during the early Oligocene (Klose et al., 1982; MacLean et al., 1982). Late dike intrusions in coastal West Greenland are dated as 30.6–

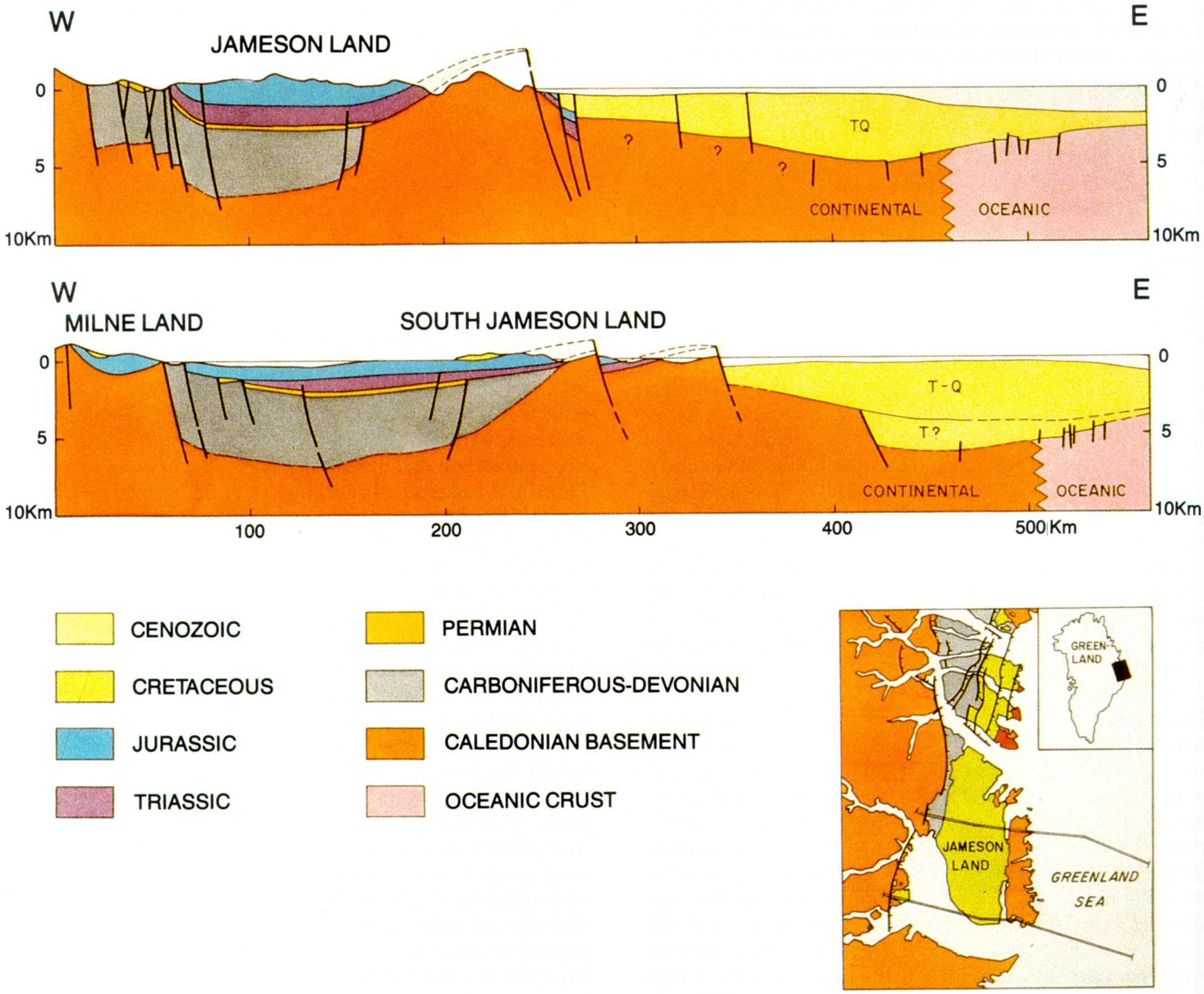

Figure 50—Schematic structural cross section through Jameson Land Halfgraben and Central East Greenland shelf. After Cleintuar (1985), SIPM.

30.2 Ma (McKenna, 1983). Termination of tectonic activity along the Ungava fracture zone is thus in keeping with the termination of sea-floor spreading in the Labrador Sea, as derived from its sea-floor magnetic anomalies.

It is uncertain to what extent the Baffin Bay is underlain by oceanic crust. Gravity and refraction data show that under its axial parts the crust–mantle boundary is pulled up to about 10 km, that the crust is thinned to 3–5 km and that it is overlain by a substantial thickness of sediments. Magnetic data indicate the presence of small-amplitude anomalies that are difficult to correlate from line to line; however, a case can be made for their linearity, subparalleling the basin axis. These anomalies cannot be calibrated in terms of the magnetostratigraphic time scale. The subdued nature of these anomalies is probably the result of the blanketing effect of the Cenozoic sediments. These attain thicknesses of 3000–4000 m in its axial zone of the Baffin Bay (Keen et al., 1974; Jackson et al., 1977, 1979; Srivastava et al., 1981; Rice and Shade, 1982; Menzies, 1982).

The regional cross section through the Baffin Bay, given in Fig. 53, is based on published refraction and gravity data (Srivastava et al., 1981; Menzies, 1982) and information derived from a multichannel reflection line. It illustrates that a young clastic sequence up to 3.5 km thick, deposited under a tectonically quiescent regime, overlies a regional unconformity characterized by a considerable relief.

In the west-central parts of the basin, this relief is upheld by a system of volcanic build-ups. These were presumably extruded under submarine conditions since they neither are associated with foresetting geometries nor display seaward-dipping reflection patterns. The area underlain by volcanic edifices has a minimum width of 160 km but could possibly be as wide as 210 km. The relief of the volcanic edifices between shotpoints C-3 and 37/38 displays a rough symmetry.

In the remainder of the profile, the young clastic sequence overlies a block-faulted relief upheld by older sediments. Near shotpoint C-9 a small volcanic mound is evident on the dipslope of one of these fault blocks. There is some uncertainty whether volcanics are involved in the block-faulted structures between shotpoints 37/38 and C-8. Refraction data indicate, however, that this segment of the profile is underlain by 2–3 km thick sediments having an average velocity of 3.7 km/sec. Under the shelf and slope of Baffin Island, possible sub-unconformity sediments have a velocity of 5.1–5.4 km/sec.

The age of the pre- and post-unconformity sediments and of the volcanics can only be inferred from regional correlations.

Seismic and well data from central West Greenland show that tensional faulting affected Cretaceous to Eocene series, that the area became tectonically quiescent during the early Oligocene, and that mid-Oligocene and younger sediments accumulated under a passive margin setting (Henderson et al., 1981; Rolle, 1985). Volcanic activity started during the late Danian and persisted into the early Eocene. Last volcanic manifestations are early Oligocene dikes on Nugssuaq Peninsula (McKenna, 1983).

Seismic data from the shelf of Baffin Island indicate that tensional faulting affected mainly Late Cretaceous sediments, persisted into the Early Tertiary, and that Mid-Tertiary to recent series were deposited under a passive margin regime. Further-

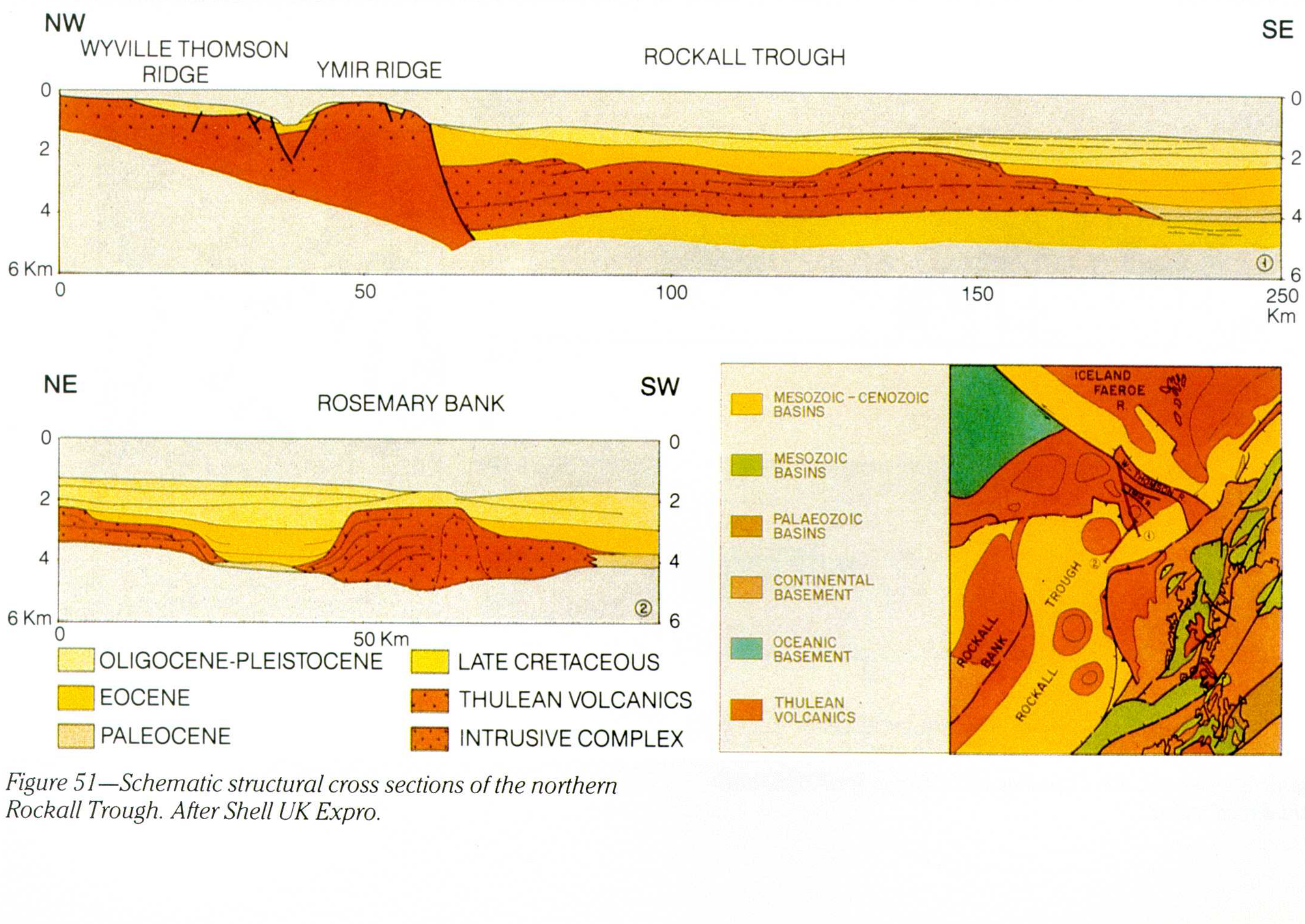

Figure 51—Schematic structural cross sections of the northern Rockall Trough. After Shell UK Expro.

more, there is evidence for Early Tertiary crustal extension at the northern end of the Baffin Bay in the Lancaster Sound and in the Eclipse Trough. Contemporaneous rifting may also have affected the Parry Channel fracture zone (McWhae, 1981; Kerr, 1981a, 1981b; Rice and Shade, 1982; see next section of this chapter).

In conjunction with the late Paleocene reorganization of the sea-floor spreading axis in the Labrador Sea (Tucholke and Fry, 1985), this suggests that crustal separation was achieved in the Baffin Bay during the late Paleocene, that the regional unconformity evident on Fig. 53 has a corresponding age, and that sea-floor spreading persisted into the early Oligocene. The width of the oceanic crust generated during this period is, however, difficult to define on the basis of the available data because the boundary between oceanic and continental crust generally is ill defined.

In view of the magnitude of the latest Cretaceous and Paleogene motion of Greenland relative to the North American Craton, as indicated by the sea-floor magnetic anomalies in the Labrador Sea and by the transect through the Baffin Bay given in Fig. 52, it is unlikely that the Baffin Bay is entirely underlain by highly stretched continental crust, as suggested by Kerr (1967) and van der Linden (1975). On the other hand, the area occupied by oceanic crust in the Baffin Bay is probably not as wide as assumed by Srivastava (1978; see also Keen et al., 1974).

Yet, palinspastic reconstructions, taking into account the sea-floor magnetic anomalies of the Labrador Sea (Fig. 55), indicate that during the opening of the Labrador Sea the distance between Cape Dyer on Baffin Island and Disko Island in central West Greenland increased by some 350 km; of this, 150 km were presumably achieved by crustal extension and sinistral translation during the Cretaceous and Paleocene and some 200 km by sea-floor spreading during the latest Paleocene to early Oligocene.

Although the width of oceanic crust underlaying the Baffin Bay probably decreases northward, as suggested by the rotational movements of Greenland (Menzies, 1982; Rice and Shade, 1982), the above-postulated values for crustal extension and sea-floor spreading are difficult to reconcile with the transect given in Fig. 53. In this context, it should be noted that the available data do not permit estimations of the amount of

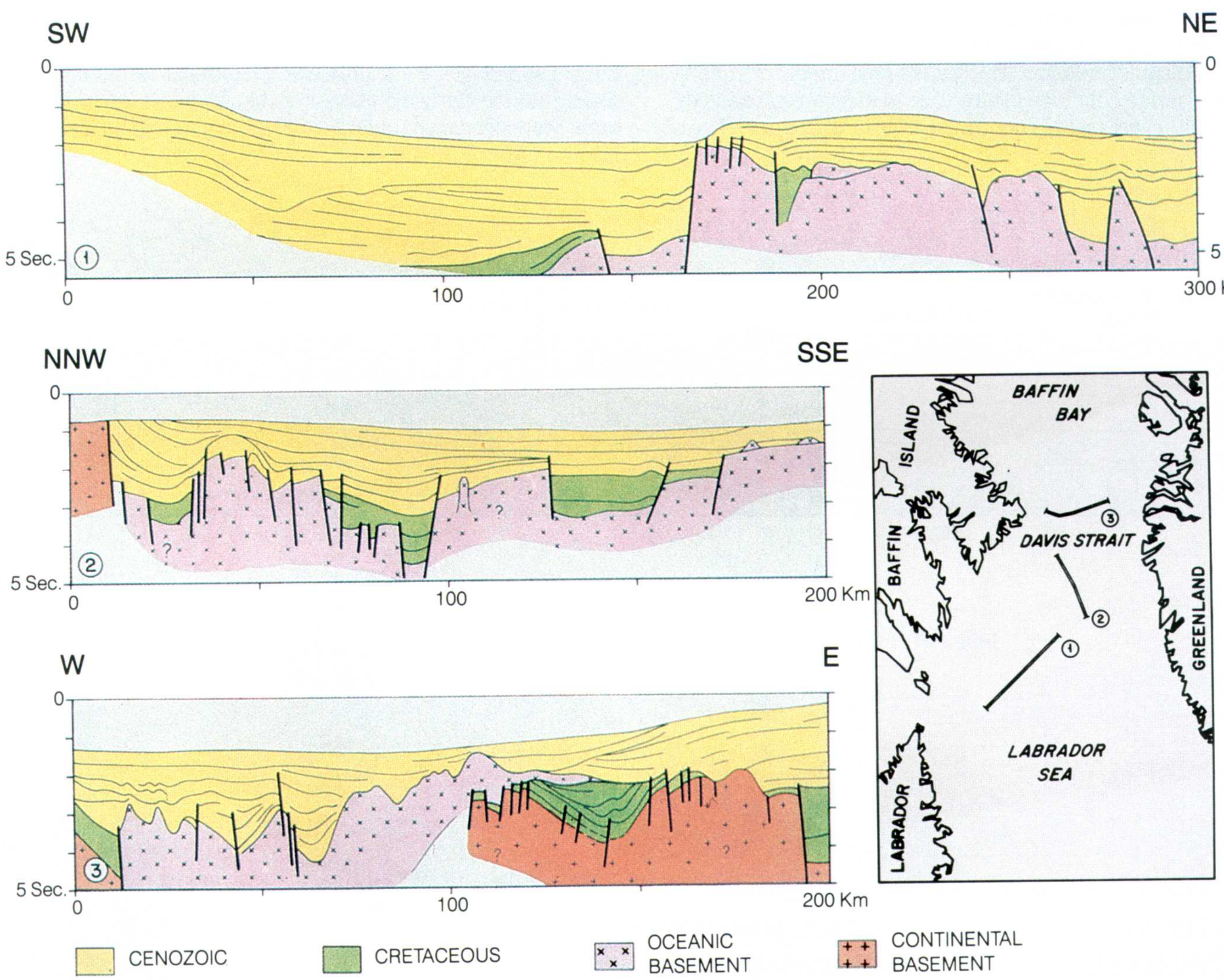

Figure 52—Seismic line drawings of the Davis Strait and northern Labrador Sea.

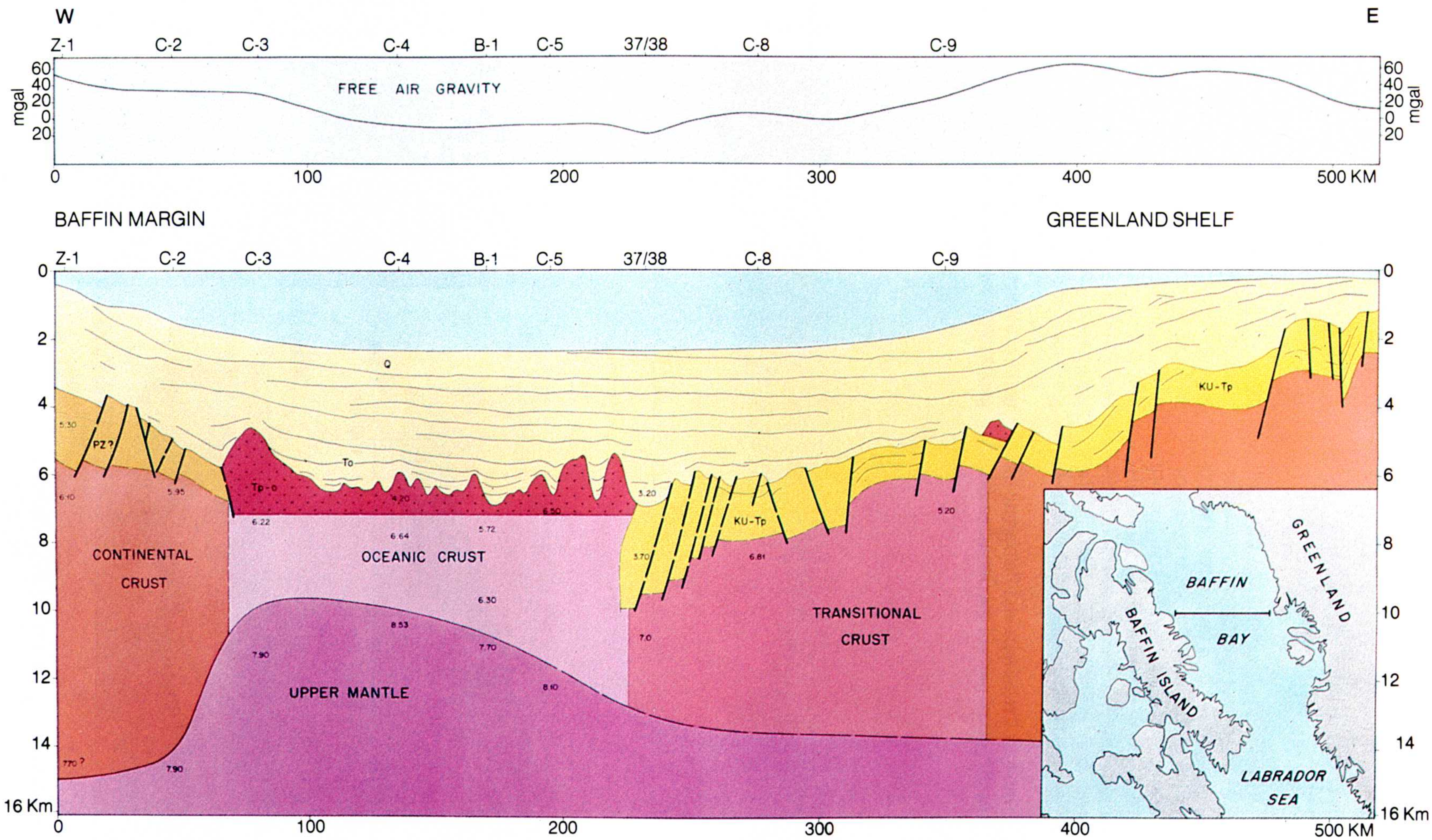

Figure 53—Crustal structure of Baffin Bay, based on multichannel reflection, refraction, and gravity data. Interval velocities given in km/sec. After Srivastava et al. (1981) and Menzies (1982).

crustal extension that took place in the shelf and slope areas of Western Greenland and Baffin Island.

Clearly, additional combined reflection and refraction seismic traverses are needed to constrain the rifting and sea-floor spreading history of the Baffin Bay and of the drift patterns of Greenland.

GREENLAND-ARCTIC ARCHIPELAGO DIFFUSE PLATE BOUNDARY

Cenozoic displacements between Greenland and the North American Craton, caused by sea-floor spreading in the Labrador Sea and in the Baffin Bay, were taken up by crustal extension in southern parts of the Canadian Arctic Archipelago, by sinistral strike-slip movements along the Nares Strait fracture zone, and mainly by compressional deformations in the eastern Sverdrup Basin; the latter resulted in the consolidation of the Eurekan fold belt (Trettin and Balkwill, 1979; Kerr, 1981a, 1981b; Miall, 1981; Srivastava and Falconer, 1982; Srivastava, 1985).

The main element of the Late Cretaceous and Paleogene rift system transecting the southeastern parts of the Canadian Arctic Islands is the Parry Channel fracture zone, which is linked via the Lancaster Sound Graben with the Baffin Bay (Kerr, 1981a). From the Lancaster Graben, the Eclipse Trough branches off to the south, and additional grabens may underlay the westward adjacent Admiralty and the Prince Regent inlets (Plate 17). Moreover, there is evidence that Devon Island and the southeastern part of Ellesmere Island were also affected by tensional tectonics and that the Jones Sound, separating them, is underlain by a latest Cretaceous to Cenozoic graben (Kerr, 1981a; Mayr, 1984). Apart from the Eclipse Trough, the age of the sedimentary fill of these grabens is, however, unknown (Miall et al., 1980; Kerr, 1981b; McWhae, 1981). Yet, the occurrence of a mid-Eocene alkaline bimodal volcanic suite on southeastern Bathurst Island (central parts of Canadian Arctic Archipelago; Mitchell and Platt, 1984) indicates that this rift system was active during the tensional evolution of the Baffin Bay. The amount of crustal extension, which was achieved within this Arctic rift system, is unknown, and, correspondingly, it is uncertain to what extent it compensated for crustal dilation and sea-floor spreading in the Baffin Bay.

In the northernmost parts of the Baffin Bay (North Water Bay, at the transition from the Baffin Bay to the Nares Strait), marine geophysical data indicate the presence of a Cretaceous–Tertiary

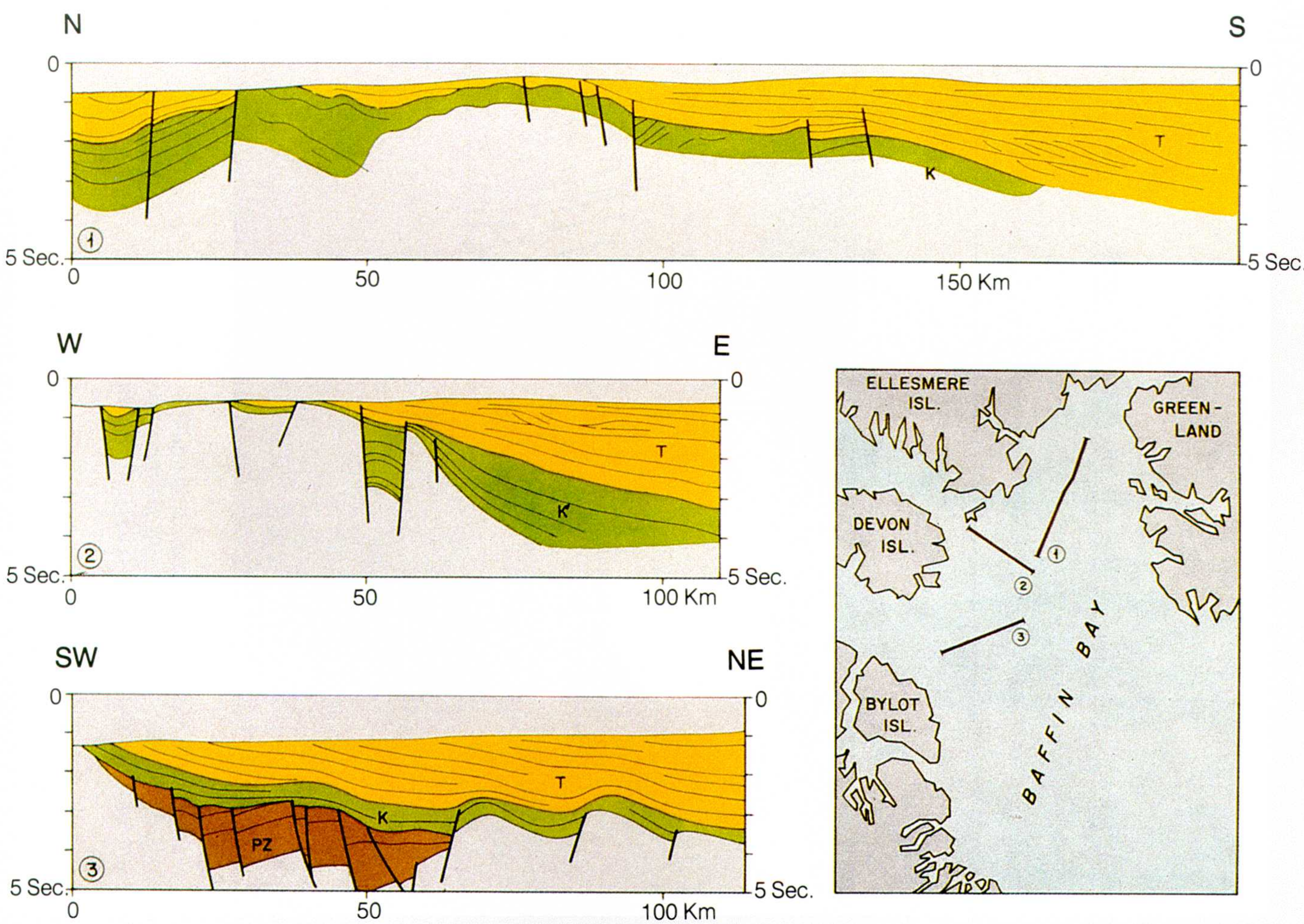

Figure 54—Seismic line drawings of North Water Bay, northern Baffin Bay.

sedimentary sequence, which is involved in synclinal and complex anticlinal structures (Fig. 54; Newman, 1982). The development of these features was probably caused by transpressional movements along the intracontinental Nares Strait fracture zone. The magnitude of these indicated sinistral movements is, however, still being debated with estimates ranging from a low of 25 km or less (Dawes and Kerr, 1982) to some 125 km (Miall, 1983; Srivastava, 1985).

The third mechanism compensating for movements between Greenland and the North American Craton was crustal shortening in the Eurekan fold belt. This intracratonic orogen developed during Maastrichtian to early Oligocene time (Miall, 1981, 1984a, 1984b). Compressional and transpressional Eurekan deformations are evident along the northern margin of Greenland (Håkansson and Pedersen, 1982), in northwestern Ellesmere Island and along its coast facing the northern part of the Nares Strait (Mayr and de Vries, 1982), and on Axel Heiberg Island. To the west, Eurekan deformations fade out on Amund Ringnes and Corwall Island. It is possible that the north–south-trending Boothia Arch also became reactivated during the Eurekan diastrophism (Trettin and Balkwill, 1979; Balkwill and Bustin, 1980; Kerr, 1981a; Hugon, 1983).

On Ellesmere and Axel Heiberg Island the Eurekan fold belt is characterized by folds, evaporite-cored diapiric structures, steep reverse and thrust faults involving Ellesmerian deformed sequences, and the late Paleozoic to Mesozoic series of the Sverdrup Basin (Osadetz, 1982; van der Berkel et al., 1983). Northwest- and southeast-verging thrust faults are evident with southeastward thrusting and overturning prevailing (Fig. 13). The amount of crustal shortening achieved during the consolidation of the Eurekan fold belt is difficult to determine for want of reflection seismic control and due to extensive ice cover. On individual thrust faults, horizontal displacements of the order of 10 km have been estimated by Osadetz (1982), while other structures are interpreted as steep reverse faults or up-thrusts (Higgins and Soper, 1983). Overall deformation patterns are suggestive of left-lateral ductile shear (Hugon, 1983).

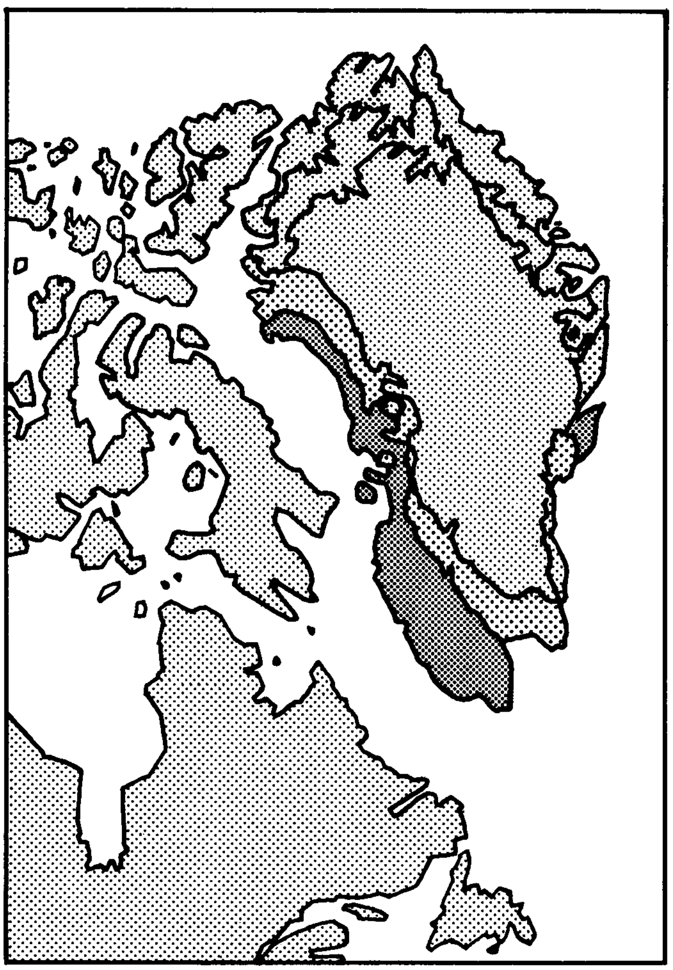

Figure 55—Palinspastic reconstruction of Labrador Sea–Baffin Bay area illustrating relative motions between Greenland and North America. Dark shading: anomaly 31 (68 Ma); medium shading: anomaly 24 (56 Ma); light shading: present position.

The timing of the Eurekan diastrophism is constrained by the accumulation of the Eureka Sound synorogenic molasse sequence that ranges in age from Maastrichtian to early Oligocene (Miall, 1981, 1984a, 1984b, 1986). This series is unconformably overlain by the mid- to upper Miocene Beaufort clastics that were deposited after a major erosional hiatus, corresponding to a phase of postorogenic uplifting (Reidiger et al., 1984; Miall, 1984b).

The stratigraphically dated termination of the Eurekan diastrophism coincides with the cessation of sea-floor spreading in the Labrador Sea (Miall, 1984a, 1984b). A genetic link between these two megatectonic processes is therefore plausible (Kerr, 1981a, 1981b; Pierce, 1982; Srivastava, 1985).

Palinspastic reconstructions, based on the closure of the Labrador Seas as dictated by its magnetic sea-floor anomalies, suggest that a major gap should have existed between Ellesmere Island and Greenland prior to the Eurekan orogeny, if the present shape of Canadian Arctic Islands is retained (Fig. 55). Such a gap, which would have to be occupied by oceanic crust, obviously never existed as is evident from the late Paleozoic and Mesozoic paleogeography of the Sverdrup Basin and the Svalbard area. In order to close this gap, the Eurekan fold belt has to be restored palinspastically to the end that the southeastern coast of Ellesmere Island was always located adjacent to Greenland. This entails a significant enlargement of the northeastern parts of the Sverdrup Basin. On the paleogeographic–paleotectonic maps given in Plates 4–16, this was taken care of in a simplistic way by spreading out the eastern part of the Arctic Islands over the available space.

Detailed analyses of the generally accepted drift patterns of Greenland relative to Baffin Island and the Arctic Archipelago indicated that the late Senonian and Paleocene rotation of Greenland induced a first phase of compression of the eastern parts of the Sverdrup Basin. During the late Paleocene to early Oligocene phase of sea-floor spreading in the Labrador Sea and in the Baffin Bay, Greenland underwent a second phase of rotation and at the same time a sinistral translation relative to Ellesmere Island. This gave rise to the second phase of the Eurekan orogeny. The total amount of crustal shortening achieved in northern Ellesmere Island should therefore be of the order of 300–350 km (Srivastava, 1985). In this context, it should be kept in mind that the Eurekan deformation front passes through southeastern Ellesmere Island and only touches the northwestern promontory of Devon Island. These undeformed areas thus formed part of the North American Craton, although they were affected by crustal distension during the evolution of the Baffin Bay and the consolidation of the Eurekan fold belt.

Paleomagnetic data suggest that during the Eurekan diastrophism, northern Ellesmere Island had undergone a $\pm 36°$

counterclockwise rotation but no detectable latitudinal translation (Wynne et al., 1983). As the late Senonian to early Oligocene counterclockwise rotation of Greenland amounts to only 10°, it must be assumed that the bulk of the paleomagnetically detectable rotation in northern Ellesmere Island is of a local nature, probably involving large-scale structural rotations (oroclinal bending).

The dimensions and timing of sinistral translation along the Nares Strait fracture zone (Senonian-Paleocene or Eocene-early Oligocene) depends essentially on the amount and timing of crustal extension and sea-floor spreading in the Baffin Bay. The latter is difficult to quantify on the basis of the available data, and information from the Eurekan fold belt is too imprecise to provide further constraints (see above).

The Nares Strait fracture zone, combined with the Eurekan fold belt, represent a diffuse plate boundary between Greenland and the North American Craton (Hugon, 1983; Miall, 1983, 1984b). The Eurekan fold belt can be considered as an inverted basin and its evolution is probably the effect of intraplate compressional stresses. Under the impact of the northwestward-rotating stable Greenland Craton, inversion of the eastern parts of the Sverdrup Basin was presumably predetermined by its basement configuration and composition. In this respect, it should be kept in mind that the "basement" of the Sverdrup Basin is largely formed by the externides of the Ellesmerian fold belt and that it was subjected to considerable extension during the late Paleozoic and Mesozoic subsidence of the Sverdrup Basin (see Fig. 14 and Chapter 4). This immature, tensionally attenuated basement complex was prone to deform under the influence of tangential stresses. It is, however, uncertain whether the total amount of crustal shortening implied by the drift of Greenland was taken up in the presently exposed Eurekan fold belt or whether imbrication of the continental margin prism facing the Arctic Ocean and possibly even subduction of oceanic crust also played an important role (e.g., comparable to the Cantabrian margin of Iberia, Chapter 9).

In view of the substantial amount of crustal shortening that was achieved during the consolidation of the Eurekan fold belt, a commensurate amount of crustal thickening and subduction of upper mantle material must have occurred. With the termination of sea-floor spreading in the Labrador Sea, compressional stresses relaxed in the Eurekan fold belt and the thickened lithosphere of the eastern part of the Canadian Arctic Archipelago became isostatically uplifted. This uplifting phase is probably responsible for the late Cenozoic reactivation of the Arctic graben system and, in particular, of the Parry Channel fracture zone (Kerr, 1981a, 1981b; Miall, 1984b).

In conclusion, it appears that opening of the Labrador Sea and Baffin Bay entailing a 10° counterclockwise rotation of Greenland away from the Canadian Shield (Fig. 55) caused rifting in the southeastern Canadian Arctic Archipelago (Lancaster and Jones Sound, Parry Channel rifts), important sinistral translations between Greenland and Ellesmere Island, and major crustal shortening in the Eurekan fold belt.

OPENING OF THE EURASIAN BASIN

Sea-floor magnetic anomalies indicate that crustal separation between the Barents-Kara Sea Shelf and the Lomonosov Ridge was achieved at the transition from the Paleocene to the Eocene between anomalies 25 and 24, or slightly earlier (Plate 17). With the onset of sea-floor spreading in the Eurasian Basin along the mid-oceanic Nansen-Gakkle Ridge, the Lomonosov Ridge became part of the North American plate (Vogt et al., 1981; Sweeney, 1985; Srivastava, 1985).

The Lomonosov Ridge has a flat-topped sea-floor relief of some 3000 m. It is about 1800 km long, and its width ranges between 20 and 100 km. On geophysical evidence, this ridge represents a clearly block-faulted, partly sediment-covered continental fragment that is offset by the oceanic Makarov and Fram basins (Blasco et al., 1979; Mair and Forsyth, 1982; Sweeney et al., 1982; Weber and Sweeney, 1985). Separation of such a long narrow piece of continental crust from the northern margin of Eurasia was presumably preconditioned by the structural grain of its basement. Such a basement fabric was probably provided by the hypothetical latest Devonian-earliest Carboniferous Lomonosov fold belt, forming the eastern continuation of the Innuitian fold belt (see Chapter 2, Innuitian Fold Belt).

Little is known about the rifting stage preceding the onset of sea-floor spreading in the Eurasian Basin. Repeated uplift of a land mass (Lomonosov High) in the area of the future Eurasian Basin is reflected by the influx of clastics into Franz Josef Land and Svalbard during the Late Triassic to Middle Jurassic (Plates 9-12) and again during the late Early Cretaceous (Plate 15). This may be related to tectonic activity, including possible shearing, along the ancestral Makarov and Nansen rifts. During the Late Cretaceous the northern parts of the Barents Platform became uplifted and remained emergent during the Cenozoic opening of the Eurasian Basin (Plates 16-21).

In northern Svalbard, volcanic activity resumed during the Paleocene and persisted into the early Miocene (Burov and Zagrusina, 1976; Prestvik, 1977). Similarly, there is evidence for Paleocene volcanic activity in northeastern Greenland (Dawes, 1976; Larsen, 1982). This volcanism can be related to Paleocene hotspot activity at the junction of the Nansen Rift and the Senja-De Geer fracture zone, which led to the development of the submarine volcanic Morris Jessup Rise and the Yermak Plateau (Plate 18; Crane et al., 1982; McKenna, 1983).

Volcanic activity in the area of the Yermak Plateau and the Morris Jessup Rise apparently persisted during the early phases of sea-floor spreading along the Nansen-Gakkle Ridge. Progressive opening of the Eurasian Basin ultimately resulted in the separation of the Yermak Plateau and the Morris Jessup Rise and their drifting apart. Because of their young, volcanic nature and their location at the continent-ocean transition, the Yermak Plateau and the Morris Jessup Rise can be suppressed in predrift assemblies of the Barents-Kara Shelf and the Lomonosov Ridge (Crane et al., 1982).

NORTHERN NORTH ATLANTIC AND NORWEGIAN-GREENLAND SEA

Following crustal separation between Greenland and the Rockall-Hatton-Faeroe Bank and Norway, the northern North Atlantic and the Norwegian-Greenland Sea oceanic basins opened during the early Eocene at sea-floor spreading rates on the order of 1.3-1.5 cm/year. After anomaly 20, spreading rates decreased gradually to about 0.7 cm/year during the Oligocene but subsequently accelerated again to an average rate of 0.9 cm/year.

Throughout the opening history of the northern North Atlantic and the Norwegian-Greenland Sea, the Reykjanes and Mohn's ridges maintained their mid-oceanic position. Along the Aegir Ridge, normal sea-floor spreading occurred during anomaly 24B to 20 time. Between anomalies 20 to 7, sea-floor spreading rates decreased more rapidly in the southern parts of the Aegir Ridge than in its northern parts. This resulted in the development of a northward-divergent magnetic anomaly pattern (Plate 18). At the same time, the Jan Mayen microcontinent became gradually separated from Greenland as a consequence of northward-decreasing, slow sea-floor spreading along the nascent Kolbeinsey Ridge. With anomaly 13, sea-floor spreading began in the southern parts of the Greenland Sea, and with anomaly 7, normal sea-floor spreading began in the prolongation of the Mohn's Ridge along the Knipovich Ridge, while the Aegir Ridge became extinct (Plates 19, 20; Eldholm and Thiede, 1980; Vogt et al., 1981; Myhre et al., 1982; Nunns, 1983; Nunns et al., 1983; Eldholm et al., 1984; Skogseid and Eldholm, 1987).

The segment of the Norwegian–Greenland Sea, which formed by sea-floor spreading along the Aegir Ridge and later along the Kolbeinsey Ridge, is limited to the north by the Jan Mayen fracture zone and to the south by the Iceland Ridge transform zone. These fracture zones apparently projected beneath the European Craton where important wrench deformations occurred during the Oligocene period of sea-floor spreading ridge reorganization in the Norwegian Basin. The onset of this reorganization coincides with the termination of sea-floor spreading in the Labrador Sea and its completion coincides with the beginning of sea-floor spreading between the Barents Shelf margin and northeastern Greenland along the Knipovich Ridge.

The latest Cretaceous to mid-Oligocene evolution of the Western Barents Shelf margin reflects a complex sequence of transpressional and transtensional deformations that accompanied the early sea-floor spreading phases in the Norwegian–Greenland Sea (Faleide et al., 1984; Spencer et al., 1984).

East Greenland Margin

The structural and stratigraphic framework of the sedimentary basins underlaying the vast continental shelves of East Greenland is known in rough outlines only because sea-ice and icebergs impede seismic surveys and even more so the drilling of exploratory wells.

Aeromagnetic and limited reflection seismic surveys indicate that major sedimentary basins occupy the northeastern and the south-central shelves of Greenland. These contain up to 10 km of sediments. Little is known, however, about the age of these sediments and the structural configuration of the respective basins (Henderson, 1976; Featherstone et al., 1977; Larsen, 1980, 1984).

Reflection seismic data from Central East Greenland show that its shelf and continental rise are underlain by a 3–5-km thick, seaward prograding clastic wedge that was deposited under a passive margin setting (Fig. 50). These clastics overlie a partly block-faulted acoustic basement (Hinz and Schlüter, 1978, 1980; Larsen, 1984).

Because this is the area from which the Jan Mayen Ridge was separated during the mid-Oligocene, it is likely that this passive margin prism overlies partly downfaulted and downflexed continental crust and older sediments covered by Paleocene–Eocene volcanics, and partly Oligocene and younger oceanic crust (Fig. 50). The age of this passive margin wedge probably ranges from late Oligocene to Recent and its dimensions suggest that after crustal separation between the Jan Mayen microcontinent and Greenland was achieved, a major eastward-directed drainage system had developed in the latter.

In Central East Greenland, development of the coastal flexure during the mid-Eocene to early Oligocene was accompanied by syenitic and alkali-granitic intrusions. These range in age between 40 and 28 Ma (Haller, 1971; Brown et al., 1977; Rex et al., 1978). This magmatic activity coincides with the gradual development of the Kolbeinsey Ridge during the mid-Eocene to mid-Oligocene and the separation of the Jan Mayen Ridge from Greenland. Moreover, mid-Eocene basalt flows have been encountered at DSPD site 350 on the Jan Mayen Ridge. Reflection seismic lines crossing this continental fragment indicate the presence of a mid-Oligocene separation unconformity that truncates Eocene and early Oligocene clastics that were presumably derived from Greenland (Talwani and Udintsev, 1976; Hinz and Schluter, 1978a; Nunns, 1982).

Aeromagnetic data over the Northeast Greenland Shelf suggest the presence of a system of grabens containing 8 to 10 km of late Paleozoic and Mesozoic sediments that are covered by a 2–4 km thick Cenozoic passive margin wedge (Larsen, 1984).

Along the East Greenland margin, the continent–ocean transition is only loosely defined in view of the widespread occurrence of Thulean volcanics and the progradation of a thick post-separation clastic wedge over oceanic basement. Furthermore, the amount of crustal extension that occurred during its late Paleozoic to early Cenozoic rifting stage is essentially unknown. Correspondingly, palinspastic reconstructions have to contend with considerable uncertainties.

Rockall–Hatton–Faeroe Bank, Rockall and Porcupine Troughs

The continental Rockall–Hatton–Faeroe Bank subsided only slowly during the opening phase of the northern North Atlantic. Water depth over this large, sediment-starved submarine plateau vary from less than 500 m to 1500 m.

Deep-Sea drilling results show that pelagic sediments consisting of shales, spiculites, and chalks attain thicknesses of up 1000 m in the Rockall–Hatton Basin (Roberts, 1975; Roberts et al., 1984; Bott, 1983).

The post-Thulean subsidence of the Rockall Trough is difficult to quantify. For instance, the present water depth over the crests of the Anton Dohrn and Rosemarie Bank seamounts is 500 and 600 m, respectively. These volcanic edifices had built up to sea level during their Paleocene–early Eocene development. It is uncertain whether the present water depth over the crestal parts of these structures reflects the total post–Thulean subsidence of the Rockall Trough, since it is unknown whether these seamounts were originally capped by subaerial volcanic cones that were later beveled off by wave action. On the other hand, the relief and internal configuration of the Rosemarie Bank and the morphology of the stepwise retreating constructional lava scarps along the flanks of Ymir Ridge (Fig. 51) indicate that during Paleocene–early Eocene time, water depths in the northern parts of the Rockall Trough increased from some 1000 m to 3000 m. At present, water depths in the Rockall Trough range from 1500 m in the north to 3000 m in its southern parts. On the other hand, the thickness of Cenozoic deep water clastics contained in the Rockall Trough decreases from some 2500 m in its northern parts to about 1000 m in its southern parts. This suggests that lithospheric cooling and sedimentary loading played an important role in the late Eocene to Recent subsidence of the Rockall Trough. The clastics that accumulated in the Rockall Trough and that only partially bury the Rosemary Bank, Hebrides and Anton Dohrn seamounts were derived from the Hebrides and northwest Irish Shelves. These shelves and their steep, fault-controlled slopes are characterized by considerably less intense Mesozoic crustal extension than, for instance, the West Shetland Shelf. Following their Paleocene uplift, these shelves subsided only little and were largely bypassed by clastic. With the progressive inundation of these shelves during the Neogene, the Rockall Trough became sediment starved (Roberts, 1975; Bott, 1978; Scrutton, 1986).

In the Porcupine Trough, Cenozoic sediments attain a thickness of some 3500 m and consist predominantly of deep water clastics (Fig. 32). On the Porcupine Bank and the Irish Shelf, only thin Cenozoic sediments accumulated (Max, 1978; Roberts et al., 1981b; Naylor and Shannon. 1982). Following the Thulean volcanic outburst, subsidence of the Porcupine Trough was probably governed by continued lithospheric cooling and sediment loading of its thinned crust. Particularly during the late Cenozoic, clastic supply to the Porcupine Trough lagged behind its subsidence; consequently, present water depths over much of the basin are in excess of 1000 m. Some of the border faults of the Porcupine Trough remained active, however, through much of Cenozoic time; this may be only partly an effect of differential crustal loading (see Beaumont and Sweeney, 1978). On the other hand, the occurrence of late Oligocene doleritic tholeiitic sills in the northern parts of the Porcupine Trough (Seemann, 1984) and of similar aged dikes on Dingle Peninsula (Western Ireland, Horne and MacIntyre, 1975) is probably related to small-scale wrench deformations in the onshore prolongation of the Charlie Gibbs fracture zone.

Time-equivalent wrench deformations are also evident at the northern termination of the Rockall Trough. Here the northwest-southeast-trending Ymir and Wyville Thomson ridges, which are upheld by Thulean volcanics, are bounded by faults offsetting Eocene and lower Oligocene sediments (Fig. 51; Roberts et al., 1983). These deformations may be the expression of transform movements along the Iceland–Faeroe fracture zone during the Oligocene reorganization of sea-floor spreading axes in the Norwegian Basin.

It is also possible that the Oligocene sinistral wrench deformations in the Irish Sea, the Cardigan Bay, and the Bristol Channel area, which were associated with the subsidence of local transtensional basins (Lake and Karner, 1987), are genetically related to transform movements along the Greenland–Scotland Ridge (see Chapter 9, *Intraplate Deformations in Northwest Europe*).

Faeroe–West Shetland Trough and Mid-Norway Basin

The Cenozoic subsidence patterns of the Faeroe–West Shetland Trough are difficult to assess, since much of its sedimentary fill from the Late Cretaceous and the Cenozoic consists of deep water clastics. Regional downwarping of the area is indicated, however, by the geometry of deltaic lobes prograding northwestward from the West Shetland Shelf. These foresetting units provided the source for extensive deep water clastic fans accumulating in the axial parts of the Faeroe–West Shetland Trough where Paleocene series attain a thickness of some 2000 m and Eocene to Oligocene sediments range between 2000 and 3000 m. With the progressive inundation of the Shetland Platform, clastic supply diminished and basinal areas became sediment starved; Mio-Pliocene series are generally only a few hundred meters thick (Fig. 48). There is only minor evidence for Cenozoic syndepositional faulting, much of which is related to gravitational instability of depositional slopes. From the Faeroe Platform, which is upheld by Thulean volcanics, only minor amounts of sediments were shed into the Faeroe–West Shetland Trough.

It is uncertain whether Paleocene igneous activity along the axial zone of the Faeroe–West Shetland Trough was of a subaerial or submarine nature. This trend of intrusive centers and basalt flows, of which the Faeroe Ridge forms part (Fig. 36), may be associated with an incipient, albeit abortive Paleocene sea-floor spreading axis (Ridd, 1983; Duindam and van Hoorn, 1987).

In the Mid-Norway Basin, the Cenozoic series forms a prograding wedge that attains a thickness of 3 km beneath the present shelf edge offshore Kristianssund in the Møre subbasin and further north, in the Vøring subbasin, some 2 km (Figs. 37, 38; Bøen et al., 1983). The toe of this wedge onlaps against and partly oversteps the volcanic Møre and Outer Vøring plateaux, the crests of which have subsided to a present depth ranging between 1.5 and 3 km. DSDP drilling results from the Vøring Plateau indicate that it became submerged during the early Eocene (Caston, 1976). This is in keeping with the assumption that the seaward-dipping reflection patterns associated with the outer margins of these plateaux are related to an initial phase of subaerial sea-floor spreading (Mutter et al., 1982; Skogseid and Eldholm, 1987). The present depth of the crests of these plateaux is, however, considerably smaller than would be expected from subsidence curves of normal oceanic crust but is roughly compatible with the subsidence curve of a-seismic oceanic ridges and also of attenuated continental crust (Detrick et al., 1977; Sawyer et al., 1983).

The sedimentary cover of the Møre Platform and the Vøring Plateau is generally less than 1 km. This suggests that they are upheld either by anomalously thick oceanic crust or by thinned continental crust that was covered and permeated by mantle-derived material immediately prior to and during the crustal separation stage (Skogseid and Eldholm, 1987). In general, the latter interpretation is favored, and it is therefore assumed that the continent–ocean transition lies to the west of the Møre Platform and in the western part of the Outer Vøring Plateau (Fig. 38; Bukovics and Ziegler, 1985). This concept is supported by the chemical composition of volcanics drilled during Leg 104 on the Vøring Plateau, which show evidence of crustal contamination (Taylor et al., 1986).

The lack of a regional separation unconformity in the Mid-Norway Basin indicates that crustal separation between Greenland and Norway was not associated with a major tensional event and regional thermal doming, but was rather accompanied by a phase of intense dike intrusion along the axis of impending crustal separation.

Following crustal separation, deep water clastics, ponded behind the gradually subsiding Faeroe–Shetland and Vøring Plateau highs, accumulated in the Møre and Vøring subbasins during the Eocene and Oligocene. In these basins, Paleogene sediments reach maximum thicknesses of 1 km.

During the Oligocene, a system of anticlinal structures became upwarped in the Vøring Basin (Figs. 37, 38). Development of these compressional features, which are superimposed on deep Cretaceous sedimentary troughs, is probably related to transform movements along the Jan Mayen fracture zone during the Oligocene rearrangement of sea-floor spreading axes in the Norwegian Basin. The largest of these structures, referred to as the Molde High, has at base Tertiary level a structural relief of 1200 m (Bukovics and Ziegler, 1985).

Miocene and Pliocene sediments onlap the flanks of these inversion structures and ultimately overstep them. During the Pliocene, an increase in clastic influx resulted in a rapid outward building of the shelf margin. Neogene sediments attain maximum thicknesses of the order of 2–2.5 km.

During the Plio-Pleistocene, the shoreward margin of the Mid-Norway Basin became uplifted and truncated as erosion cut down into the Triassic series as evident by the erosional truncation of Mesozoic series (Bukovicz and Ziegler, 1985; Bukovicz et al., 1984; Bugge et al., 1984). The mechanism governing this uplift is uncertain but may be related to isostatic adjustments of the crust to glaciation and deglaciation of the Fennoscandian Shield.

The crustal configuration of the Mid-Norway continental margin is given in Fig. 56. This profile, which crosses the shelf 80 km to the south of Andøya, is based on refraction and gravity data and illustrates that the continental crust thins from some 43 km beneath the eastern Caledonides to some 16 km beneath the Vøring Plateau. Significantly, westward thinning of the lower crust sets in some 100 km to the east of the shoreward margin of the Mid-Norway sedimentary basin. Within the Mid-Norway Basin, available reflection seismic data do not permit quantification of the amount of crustal extension at the base of the synrift sediments. A wider zone of lower crustal attenuation than at shallow crustal levels is compatible with the "depth dependent extension" model of Beaumont et al. (1982a, 1982b) and Hellinger and Sclater (1983; in their paper referred to as "non-uniform stretching model"). This model was introduced to approximate the inhomogenous response of the lithosphere to stresses, thought to be caused by depth-related changes of the rheological properties of the crust and lithosphere (see Chapter 10).

Although in the case of the Mid-Norway margin attenuation factors at shallow and deeper crustal levels cannot be established, it is suspected that similar to the North Biscay margin (Fig. 31) and the East Newfoundland Basin (Fig. 33), subcrustal erosion contributed substantially to the observed lower crustal thinning, particularly under the Western Caledonides and possibly also in the area of the Mid-Norway Basin.

Western Barents Shelf and Spitsbergen Orogen

The stratigraphic record and the structural framework of the western margin of the Barents Shelf and of Svalbard provide an account of tectonic processes that accompanied the early phase of sea-floor spreading in the Norwegian Greenland Sea and of the dextral oblique separation between northeastern Greenland and the northwestern tip of the Eurasian Craton (Plate 26).

The main structural elements of the southwestern Barents Sea are summarized in Fig. 10.

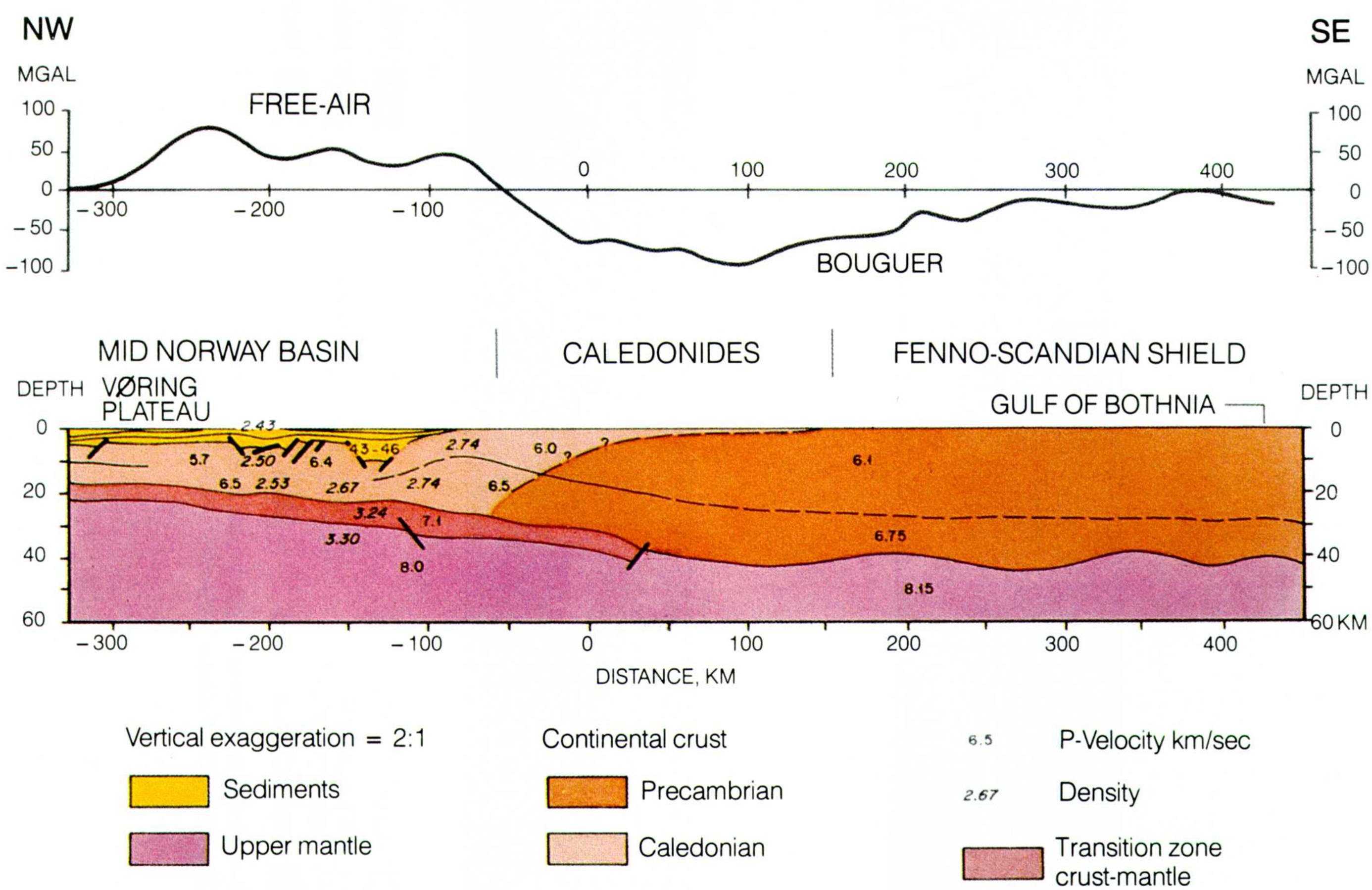

Figure 56—Mid-Norway continental margin, crustal configuration, based on "Blue Norma" refraction profile. After Meissner, personal communication.

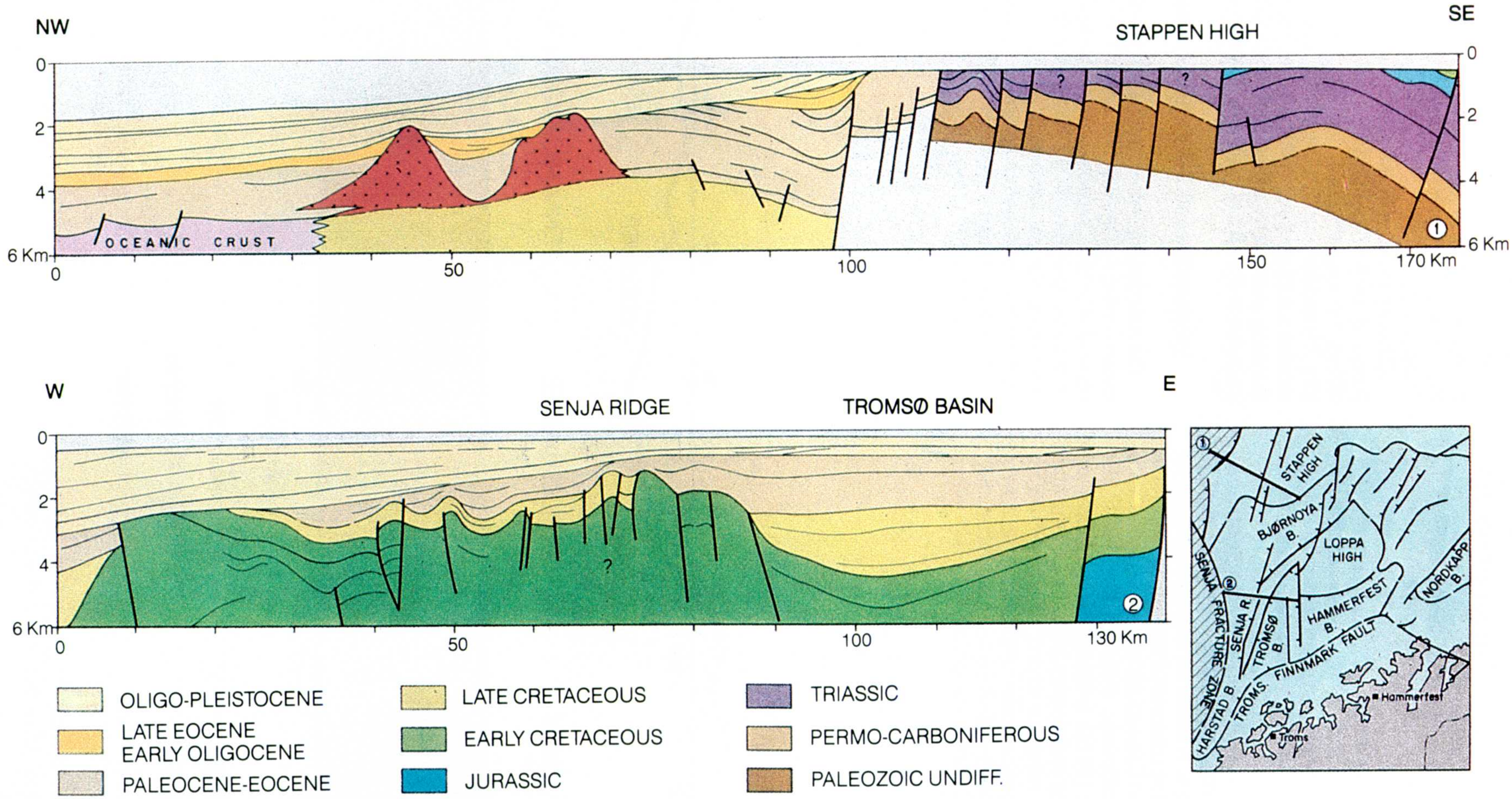

Figure 57—Schematic structural cross-section of the southwest Barents Shelf continental margin. After Norske Shell.

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Latest Cretaceous to Paleocene transpressional deformation of the Senja Ridge, and probably also of the Stappen High (Fig. 15), coincides with the termination of differential subsidence of the Tromsø and Bjørnøya basins (Fig. 39). Latest Cretaceous to Paleocene compressional deformations have been reported from Western Svalbard (Hanisch, 1984a). Late Cretaceous uplift of the Svalbard Platform may be related partly to these transpressional deformations and partly to tectonic activity along the Nansen Rift.

In the area of the Senja Ridge, a major unconformity separates deformed and deeply truncated Cretaceous strata from the late Paleocene and younger series (Spencer et al., 1984; Fig. 57). This early Paleocene erosional event was followed by late Paleocene regional subsidence of the area. In the Tromsø Basin, on the other hand, sedimentation was continuous across the Cretaceous–Cenozoic boundary (see also Fig. 39). However, compressional structures appear to have continued to grow in the western parts of the Senja Ridge as suggested by onlap patterns of the Paleocene to Eocene series. Tectonic activity intensified during the late Eocene to early Oligocene. This was accompanied by the upthrusting of the western margin of the Senja Ridge where a major, regional unconformity marks the base of the Oligocene and younger series. These were deposited under a passive margin regime and formed the Barents Sea submarine clastic fan that prograded over the newly generated oceanic crust of the Norway Basin (Lofoten Basin). This clastic wedge is up to 4 km thick (Hinz and Schluter, 1978b; Faleide et al., 1984; Eldholm et al., 1984).

The Stappen High represents a major inverted Mesozoic basin that probably formed part of the Bjørnøya Basin (Fig. 57). Uplift of this large positive feature, which culminates in Bjørnøya, probably occurred during the late Eocene to early Oligocene with possible precursor phases during the latest Cretaceous and early Paleocene (Faleide et al., 1984). During the development of the Stappen High, the northern parts of the Bjørnøya Basin also became uplifted and deeply truncated (Fig. 15). In its southern parts, where a more complete stratigraphic record is preserved, the main phase of deformation can be dated as late Eocene to early Oligocene. Moreover, seismic data indicate that the Loppa Ridge also became transpressionally reactivated and uplifted during this time.

Structural features associated with the Stappen High show that it developed in response to transpressional stresses. The western margin of the Stappen High is marked by a major normal fault zone to the west of which a thick Paleogene sedimentary sequence is preserved near the continent–ocean transition (Fig. 57). Furthermore, there is evidence for the occurrence of two major Paleocene(?) submarine volcanic buildups in this half-graben. An important erosional unconformity marks the base of the Oligocene and younger series that were deposited under a passive margin tectonic regime (Spencer et al., 1984).

During the late Oligocene and Mio-Pliocene, much of the Barents Shelf and probably also the North Kara Sea area was emergent, thus forming the clastic source area for the Barents Sea fan.

In the Svalbard area, major Cenozoic basins are the Central Tertiary Basin, located to the east of the Spitsbergen "Alpine" fold belt, the Forlandsundet Graben paralleling the west coast of Spitsbergen, and the continental margin prism (Steel et al., 1985).

The Central Tertiary Basin contains up to 2.5 km of Danian to early Eocene, brackish marine and deltaic sediments (Plate 26). Carbonization analyses indicated that an additional 1-1.7 km of sediments has been eroded from the present basin surface (Manum and Throndsen, 1978). This suggests that sedimentation had continued into the middle to late Eocene.

During the early Paleocene, the basin subsided presumably under a transtensional regime with clastics being derived from the north and east. Subsidence exceeded sedimentation rates, and deeper water conditions developed along its western, sheared margin. From the latter, conglomerates were shed during the late Paleocene, suggesting transpressional uplift of a marginal high. During the early Eocene, clastic influx into the basin increased, particularly from the west, and resulted in its shallowing out. The early Eocene series consists of 1000 m of sandy, coastal plain deposits (Kellogg, 1975; Steel et al., 1981, 1985; Steel and Worsley, 1984). Differential subsidence of the basin probably continued into the late Eocene, by which time dextral, transpressional stresses induced the uplift of the West Spitsbergen "Alpine" fold belt. This fold belt, which is up to 40 km wide, is characterized by east-verging, basement-involving, en echelon thrust faults and upthrusts and folds (Birkenmajer, 1972; Lowell, 1972; Kellogg, 1975; Lepvrier and Geyssant, 1984; Steel et al.. 1985; Maher et al., 1986).

The Forlandsundet Graben on the west coast of Spitsbergen contains a 3–5 km thick succession of clastics; this basin probably developed during Eocene and Oligocene as a transtensional feature. Subsidence of this graben is at least partly contemporaneous with the deformation of the Spitsbergen fold belt (Steel and Worsley, 1984; Steel et al., 1985).

The continental margin prism of Svalbard consists of a 6–7-km-thick sequence of presumably Oligocene and Neogene clastics that prograde over oceanic crust. This clastic wedge was dammed up by the Knipovich sea-floor spreading ridge. There is little evidence for syndepositional deformation. The continent–ocean transition is rather sharp and coincides with the Hornesund fault zone (Schlüter and Hinz, 1978; Myhre et al., 1982; Eldholm et al., 1984). This fault probably evolved out of a Late Cretaceous to Eocene shear zone into an early Oligocene rifted margin at the time sea-floor spreading commenced between the Barents Platform and northeastern Greenland. During this phase of crustal separation, the Hovgaard continental fragment, now located in the northern parts of the Greenland basin, became isolated (Myhre et al., 1982).

In the Wandel Sea Basin of northeastern Greenland, there is evidence for post-Paleocene graben formation associated with a substantial heat flow increase (Håkansson and Pedersen, 1982). These deformations could be interpreted as being equivalent to the Paleocene to early Oligocene wrench and rift activity along the margin of Svalbard. Whether a counterpart of the Spitsbergen "Alpine" fold belt occurs along the submerged northeastern shelf margin of Greenland is unknown, but probable.

Overall, the latest Cretaceous and Paleogene evolution of the western margin of the Barents Sea–Svalbard platform reflects a sequence of transpressional deformations that can be considered as equivalents to the far more dramatic Eurekan orogeny of the Eastern Sverdrup Basin (see above).

CENOZOIC NORTH SEA BASIN

The North Sea Basin forms part of the large Cenozoic Northwest European Basin, which extends from the Atlantic shelves of Norway and the Shetland Isles to the Carpathians and the Ukraine. Within this megabasin, the North Sea area stands out

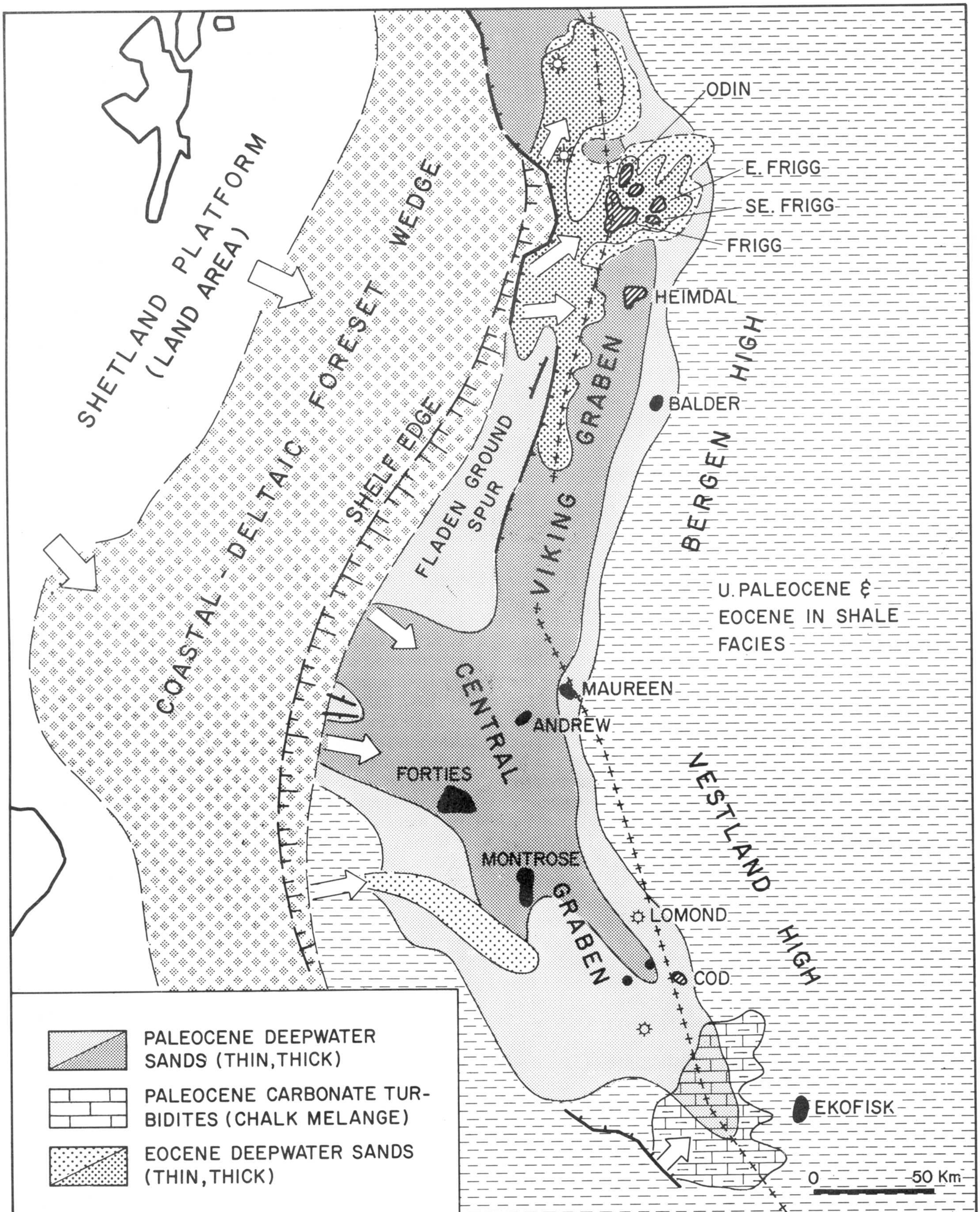

Figure 58—Schematic facies map of late Paleocene and early Eocene series, central and northern North Sea. Ziegler and Louwerens (1978).

because of its great thickness of Tertiary and Quaternary sediments (Ziegler, 1982a).

Paleogene Deep Water Stage

In the North Sea, chalk deposition gave way to a clastic regime in the Viking Graben during the Danian and in the Central and Southern North Sea at the onset of the late Paleocene. This shows that denudation of the Shetland Platform, in response to thermal uplift and falling sea levels, began earlier than the erosion of the Chalk series on the shelves flanking the central and southern parts of the North Sea Basin.

The late Paleocene–earliest Eocene crustal separation stage in the Norwegian–Greenland Sea corresponds in the North Sea to a last, albeit mild, rifting phase. This is reflected by the reactivation of the eastern border faults of the Shetland Platform, the Great Glen Fault, and parts of the Moray Firth fault system. Furthermore, some of the faults marking the northeastern margin of the Mid-North Sea High also became reactivated; this triggered mass-flow of Chalk debris from the Mid-North Sea High into the Central Graben (Fig. 58).

Paleocene uplift and eastward tilting of the Shetland Platform gave rise to the development of an eastward-directed drainage system and the outbuilding of deltaic complexes along the margins of the Viking and Central grabens; in the latter, water depths were of the order of 500–900 m (Parker, 1975). At the intersection of these deltaic and barrier-bar complexes with the tectonically still active graben margins, recurrent slope instability triggered density currents, which transported sands into these troughs where they accumulated as extensive deep water fans (Rochow, 1981; Knox et al., 1981; Morton, 1982; Enjolras et al., 1986). The geometry of these fans illustrates that the Viking and Central grabens formed an elongate deep-water trough during the late Paleocene and early Eocene, while somewhat shallower seas occupied the remainder of the North Sea Basin (Fig. 58). Progressive subsidence of the North Sea Basin, combined with rising sea levels, induced a regional transgression during the latest Paleocene and early Eocene. This is reflected by a gradual reduction of the sand influx into the Central Graben and somewhat later, also into the Viking Graben. In much of the North Sea area, late Paleocene and Eocene strata are developed in a pelagic shale facies, and there is only limited evidence for clastics being derived from the Fennoscandian Shield. Paleogene shorelines were probably located much further inland than today's coastlines of Norway and Sweden (Ziegler and Louwerens, 1979). Upper Paleocene and lower Eocene deeper water sands contain major oil and gas accumulations in the Central and Northern North Sea (Ziegler, 1980).

Over large parts of the North Sea, Oligocene strata consist of pelagic, partly deeper water clays containing minor carbonate intercalations. These contain Arctic cold-water faunas that do not permit a reliable determination of depositional water depths. In the Central North Sea, Oligocene clays reach a thickness of 800 m and represent, on the basis of seismostratigraphic criteria, a basinal facies. During the early Oligocene, sedimentation and subsidence rates probably were more or less in balance so that no appreciable shallowing of the basin took place. The Shetland Platform continued to act as a major source of sand as evidenced by a fringe of delta complexes. During the mid-Oligocene low stand in sea level, a major foreset unit prograded into the Viking Graben. This coal-bearing, sandy sequence reaches a thickness of 500 m and is laterally offset by 200 m of pelagic, deep water clays. Clastic influx from eastern sources remained also during the Oligocene at a very low level. On the other hand, thin, deltaic sheet sands prograded during the Oligocene from the southern margin of the North Sea Basin (Ziegler, 1982a).

Neogene Shallowing of the Basin

During the Miocene and Pliocene, repeated sea-level fluctuations strongly influenced sedimentation patterns in the North Sea, especially in its shallower parts where regionally correlative disconformities are evident, for instance, at the base and top of the Miocene.

The Neogene uplift of the Rhenish Shield was coupled with the development of deltaic conditions in the Southern North Sea and in Northern Germany during the Miocene and Pliocene. Similarly, deltas began to prograde from the Fennoscandian Borderzone through Denmark. This was accompanied by an increase in clastic influx into the Central North Sea where Neogene and Quaternary open-marine clays are some 2000 m thick (Fig. 59). From mid-Miocene time onwards, warm water microfaunas indicate a progressive shallowing of the central parts of the North Sea Basin. This shows that during the Miocene and Plio-Pleistocene, sedimentation rates outpaced subsidence rates. Also in the northern North Sea, Neogene strata reflect a progressive shallowing of the basin whereby progradation of deltaic complexes from the Shetland Platform into the Viking Graben continued to play an important role.

Throughout the Neogene, clastic influx from the Norwegian coast remained, however, at a rather low level. This is suggested by continued open marine connections between the Central North Sea and the Norwegian–Greenland Sea.

In the glacially scoured Norwegian and Skagerrak trenches, and also in Northern Denmark, Mio-Pliocene and older strata, forming a gently dipping monocline, are deeply truncated. This illustrates that during the Pleistocene the eastern margin of the North Sea Basin became uplifted and subjected to erosion in conjunction with the isostatic adjustment of the Fennoscandian Shield to cyclical Pleistocene glaciation and deglaciation (Figs. 18, 59).

Subsidence Patterns and Mechanisms

The regional structure map of the top Chalk series, given in Fig. 60, describes the geometry of the Cenozoic North Sea Basin and indicates that it developed by broad downwarping (Ziegler and Louwerens, 1979).

Cenozoic series reach a thickness of 3.5 km in the central parts of this saucer-shaped basin. The axis of this basin is aligned with the trace of the Viking and the Central grabens.

Most faults affecting the base of the Tertiary strata die out rapidly in Paleogene sediments. There is no evidence for a Neogene reactivation of the Mesozoic North Sea Graben system, and the Tertiary Ruhr Graben, forming the northwestern branch of the Rhine Rift, dies out in the Dutch onshore areas near the coast of the southern North Sea.

Diapirism of Permian salts caused local subsidence anomalies and faulting that interrupt the otherwise smooth Cenozoic isopachs. In the southern North Sea, additional anomalies are related to the Paleogene inversion of the West and Central Netherlands Basin, of the Sole Pit Basin, and of the southern parts of the Central Graben (see in Chapter 9, *Compressional Deformation of the Alpine Foreland*).

In general, Paleogene to Quaternary strata increase in thickness from the margins of the North Sea Basin toward its center. Notable exceptions are, however, the Oligocene deltaic complexes that prograded from the Shetland Platform into the deeper waters of the Viking Graben and also the Paleocene and Eocene series, in which both sandy shelf and turbiditic basinal facies stand out by their thickness (Fig. 59).

Refraction and deep reflection data indicate that the Moho Discontinuity is located in the coastal area of Norway at a depth of 33–34 km, under the Viking and the Central grabens at a depth of 22–24 km and 20 km, respectively, and under the Shetland Platform and the British Isles at a depth of 30 km and 32 km, respectively (Fig. 61; Solli, 1976; Christie and Sclater, 1980; Wood and Barton, 1983; Barton and Wood, 1984). Under the axial parts of the Central and Viking Graben, in which sediments reach a thickness of 8–10 km, the continental crust is 60–70% thinner than beneath adjacent Britain, the Shetland Platform, and the Fennoscandian Shield. Throughout these refraction profiles, the upper mantle displays a normal velocity of 8.1–8.2 km/sec. Deep reflection profiles show that the lower crust is highly reflective and diffractive under the Viking Graben and that its reflectivity decreases toward the Norwegian Coast and the Shetland Platform where a fairly discrete Moho reflection is evident (Beach, 1985; Fig. 62).

If it is assumed that the crust of the Viking and the Central grabens was thinned by mechanical stretching only (McKenzie, 1978), then the amount of extension across the Viking Graben would be on the order of 80–100 km and for the Central Graben about 110 km (Wood and Barton, 1983; Barton and Wood, 1984; Sclater et al., 1986).

Multichannel reflection seismic data acquired by the petroleum industry show, however, that in the northern North Sea extension of the synrift sediments by faulting is, at the Jurassic level, on the order of only 10–15 km. In the Viking Graben, Triassic series transgressed over Caledonian basement and Devonian clastics. The base of the Lower Triassic sediments, representing the earliest synrift deposits, can be mapped only locally. Thus, the extension at the base of the synrift sediments cannot be readily determined from reflection seismic data. Yet, to judge from the geometry of the individual fault blocks and the lack of significant divergence of Triassic reflectors, it seems unlikely that the total amount of Mesozoic crustal extension across the Viking Graben exceeds 20–30 km.

In the Central North Sea, where the Zechstein Salt forms part of the prerift sequence, its base corresponds to a regionally correlative stratigraphical and seismic marker. Thus, the extension mapped at this level gives a reasonable constraint to the amount of extension that can have occurred during the Mesozoic rifting stage. Amounts of extension by faulting at the base Zechstein Salt level are in the range of 25–30 km. It is very unlikely that a multitude of additional, small faults, undetectable on multichannel seismic reflection data, could account for a doubling or even a quadrupling of this amount as suggested by the stretching model (McKenzie, 1978; Wood and Barton, 1983; Barton and Wood, 1984; Ziegler, 1983a; Sclater et al., 1986).

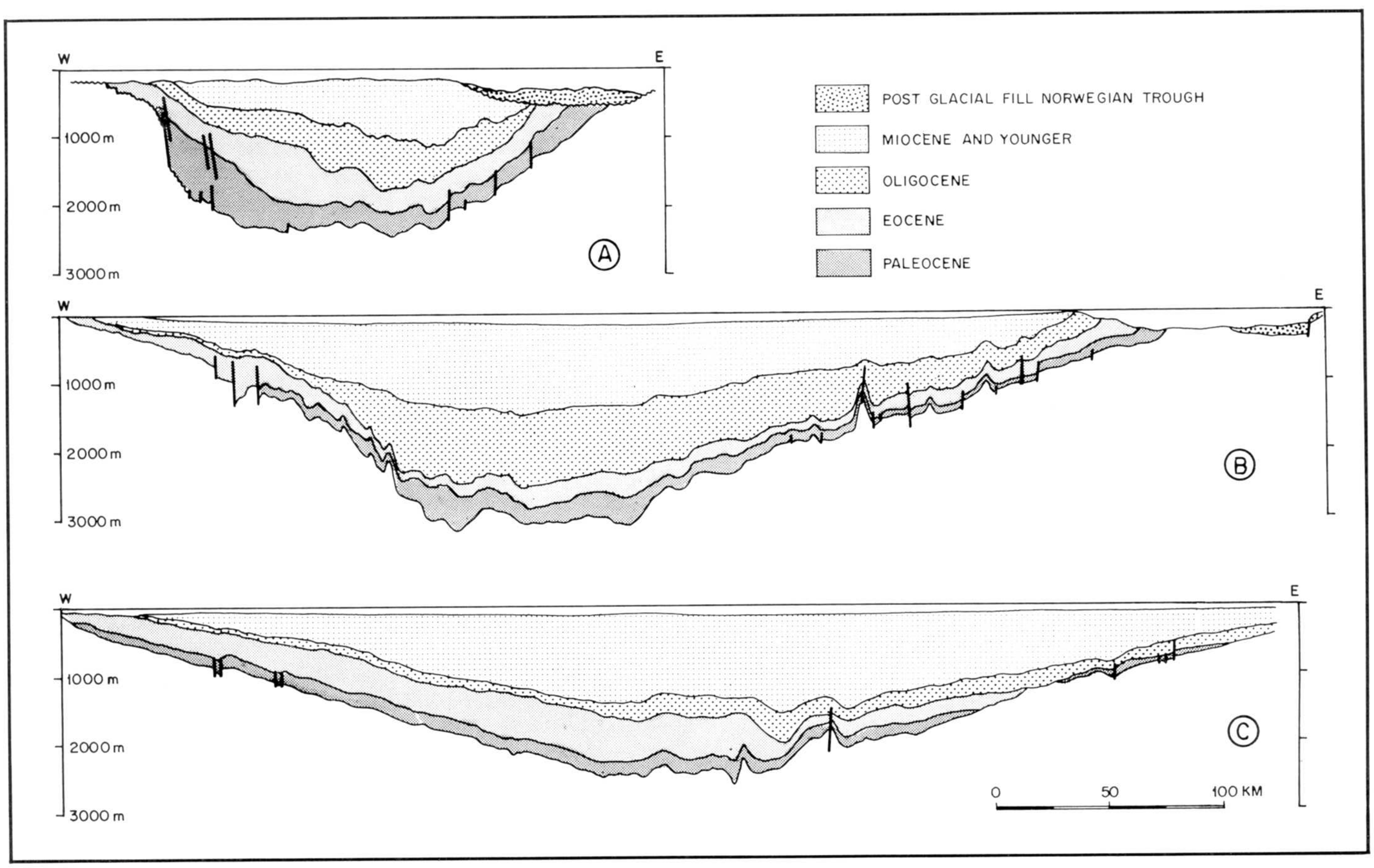

Figure 59—Structural cross sections through Cenozoic North Sea Basin. For location see Figure 60.

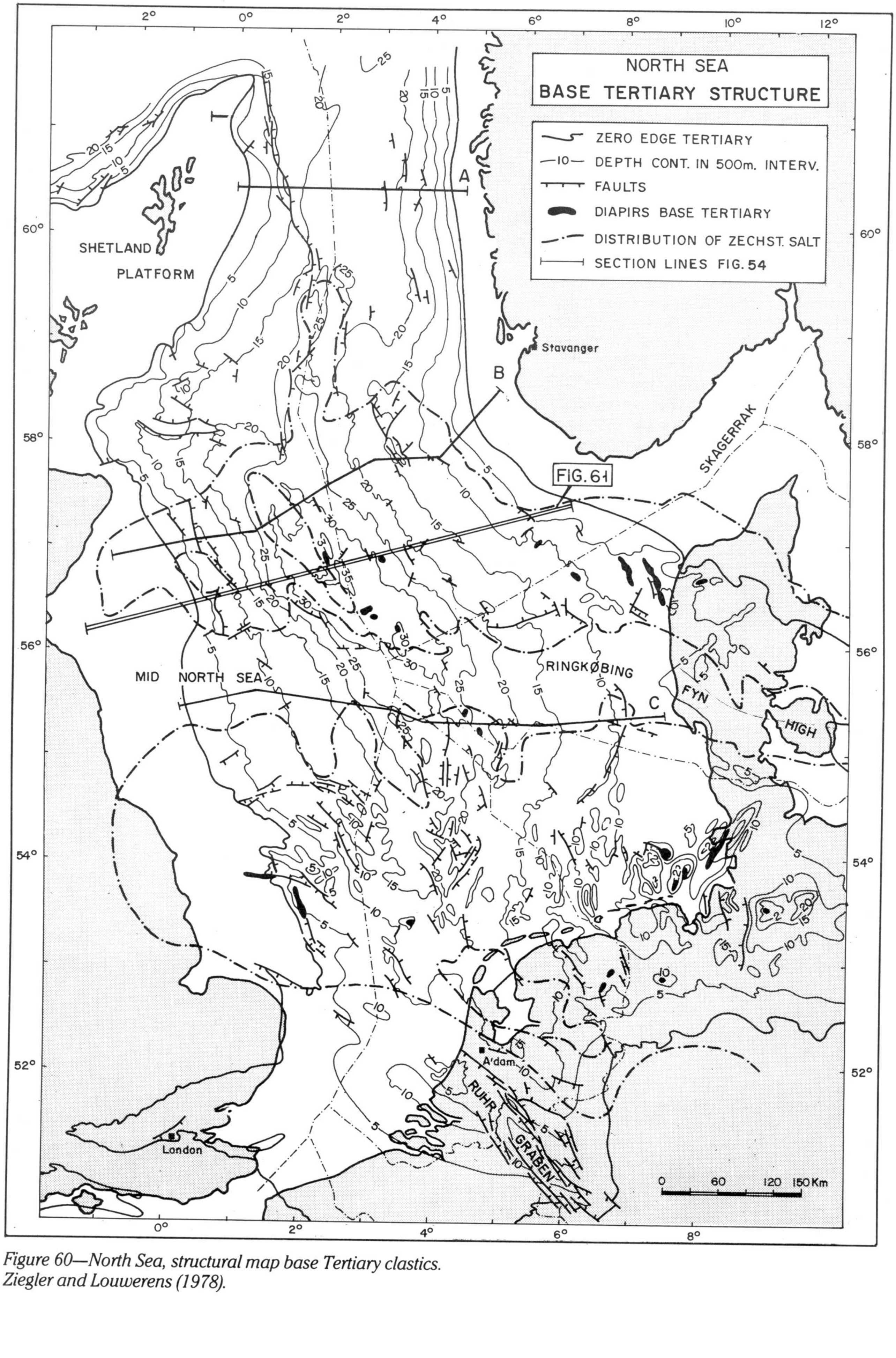

Figure 60—North Sea, structural map base Tertiary clastics. Ziegler and Louwerens (1978).

Similarly, the amount of crustal extension postulated by the stretching model for the Witchground and Buchan grabens in the outer Moray Firth area of the northwestern North Sea (Christie and Sclater, 1980) is at variance with the amount of extension observed on reflection lines (Smythe et al., 1980).

For the Danish part of the Central Graben and for the Permo-Triassic Horn Graben, reflection seismic data indicate that at the base Zechstein Salt level, the combined amount of crustal extension ranges from about 10 km to a maximum of 15 km. This is less than the values determined for the Central Graben. Therefore, it must be assumed that part of the crustal dilation that occurred in the Central North Sea was taken up in the North Danish Basin (Fig. 44; Sorensen, 1985).

Gravity data show the presence of a mass excess beneath the low-density sediments of the Viking and Central grabens. The deepest parts of the Cenozoic North Sea Basin coincide with the largest gravity anomaly in the area of the Central Graben (Donato and Tully, 1981; Hospers et al., 1986b).

The Cenozoic subsidence patterns of the North Sea Basin can be explained by the decay of the thermal anomaly that had developed during its Mesozoic rifting stage, spanning some 175 Ma. Although not yet quantifiable, the rate of crustal extension appears to have accelerated during the Triassic to reach a peak at the transition from the Jurassic to the Cretaceous. A major subcrustal thermal anomaly was probably introduced during the Mid-Jurassic doming of the Central North Sea area (Chapter 5). During the Cretaceous, the rate of crustal extension decreased rapidly. The last tensional events occurred during the late Paleocene and earliest Eocene. In this context, it should be kept in mind that the maximum thermal disturbance of the lithosphere of the North Sea area probably did not coincide with termination of rifting activity during the early Eocene but with the Mid-Jurassic and earliest Cretaceous tensional events. Thus, cooling and contraction of the lithosphere started already during the earliest Cretaceous; the subsequent minor rifting events, as well as the Thulean volcanic surge, interfered with or even temporarily reversed this cooling process. Only after the late Paleocene Laramide rifting pulse was the evolution of the North Sea Basin exclusively governed by progressive lithospheric cooling and contraction and its isostatic adjustment to sedimentary loading and sea-level fluctuations (Watts and Steckler, 1979; Royden et al., 1980; Watts et al., 1982; Watts and Thorne, 1984).

For the Central North Sea, Wood and Barton (1983) postulated, on the basis of quantitative subsidence analyses, a β-value of 1.5 for its stretched crust, while reflection data indicate a β-value of 1.1–1.2. On a basin-wide scale, this translates into an extension by 50–80 km, as derived from thermal subsidence analyses versus 110 km as derived from crustal configuration and 20–30 km as derived from reflection data (Ziegler, 1982a, 1983a; Beach, 1985; Sclater et al., 1986). Similar discrepancies between stretching factors derived from subsidence analyses, crustal configuration, and reflection data are also observed in the eastern Moray Firth Rift (Witch Ground–Buchan grabens; Fig. 35; Christie and Sclater, 1980; Smythe et al., 1980) and also in the Viking Graben (Sclater and Christie, 1980; Ziegler, 1982a).

This has led to the development of the "linked tectonics" concept that stresses oblique extension across the Viking and Central grabens to the end that the amount of extension measured in cross sections normal to these rifts may not give a true value of the actual amount of crustal extension achieved in them. This concept requires major sinistral displacements along the Viking Graben and the Fennoscandian Borderzone (Fig. 35; Beach, 1985), the magnitude of which is difficult to reconcile with the available data. Although there is evidence for minor lateral displacements along the fault systems of the Viking and Central grabens, these are unlikely to account for the major discrepancies between extension factors derived from subsidence analyses and reflection and refraction data.

The available data point toward an important discrepancy between upper and lower crustal "stretching" factors and the possibility that the mass of the lower crust is not being conserved during rifting. This leads to the assumption that crustal attenuation of the Central and Northern North Sea, as defined by gravity, refraction, and reflection data was achieved not only

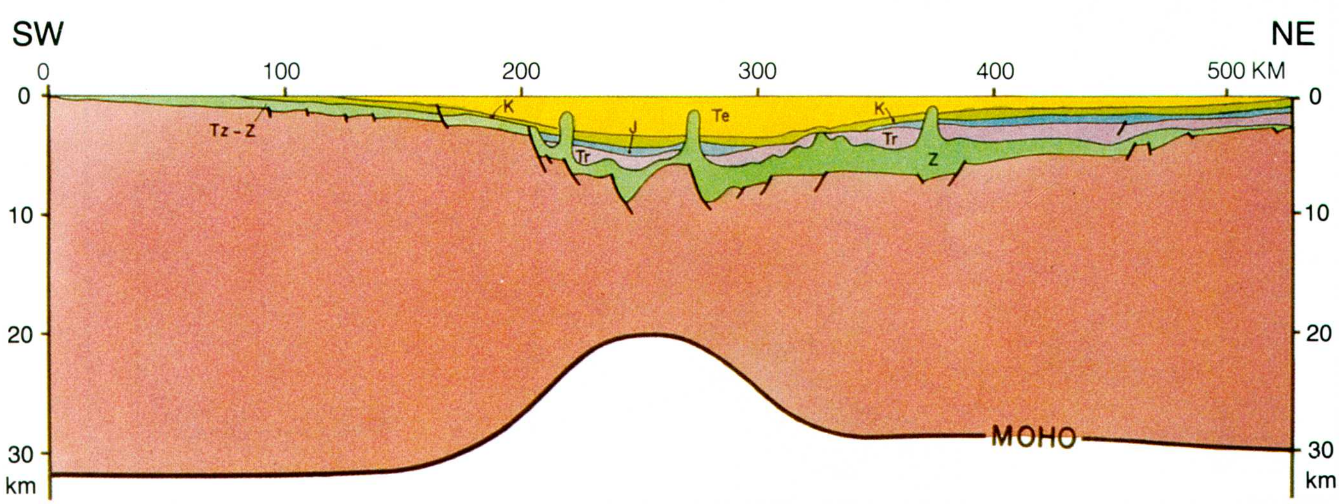

Figure 61—Crustal configuration of Central North Sea. After Barton and Wood, 1983. For location see Figure 60.

by mechanical stretching of the crust but also by thermally induced physicochemical processes that affected the lower crust and caused an upward displacement of the crust–mantle boundary. These processes, which are still poorly understood and which are apparently irreversible, are here referred to in generalized terms as "subcrustal erosion." It is speculated that the high reflectivity of the attenuated lower crust, as evident in the Viking Graben (Fig. 62), may be in some way related to these processes. It is, however, realized that intracrustal shear zones can also give rise to deep, intracrustal reflections (Brewer and Smythe, 1984; Gibbs, 1984; Beach, 1985; see in Chapter 10: *Rifting Processes*).

Lower crustal attenuation processes appear to have been more effective in the area of the Mid-Jurassic Central North Sea dome (see in Chapter 5: *North Sea Rift Dome*) than in the Northern Viking Graben where sedimentation was continuous throughout the Jurassic. This suggests that the Cenozoic Central North Sea subsidence maximum is probably related to an area of maximum crustal attenuation and the decay of a maximum thermal anomaly, resulting from the Mid-Jurassic intrusion of asthenospheric material to the crust–mantle boundary and further crustal extension during the late Kimmerian rifting pulse.

It is obvious that additional refraction and deep reflection data are required to evaluate the validity of these concepts.

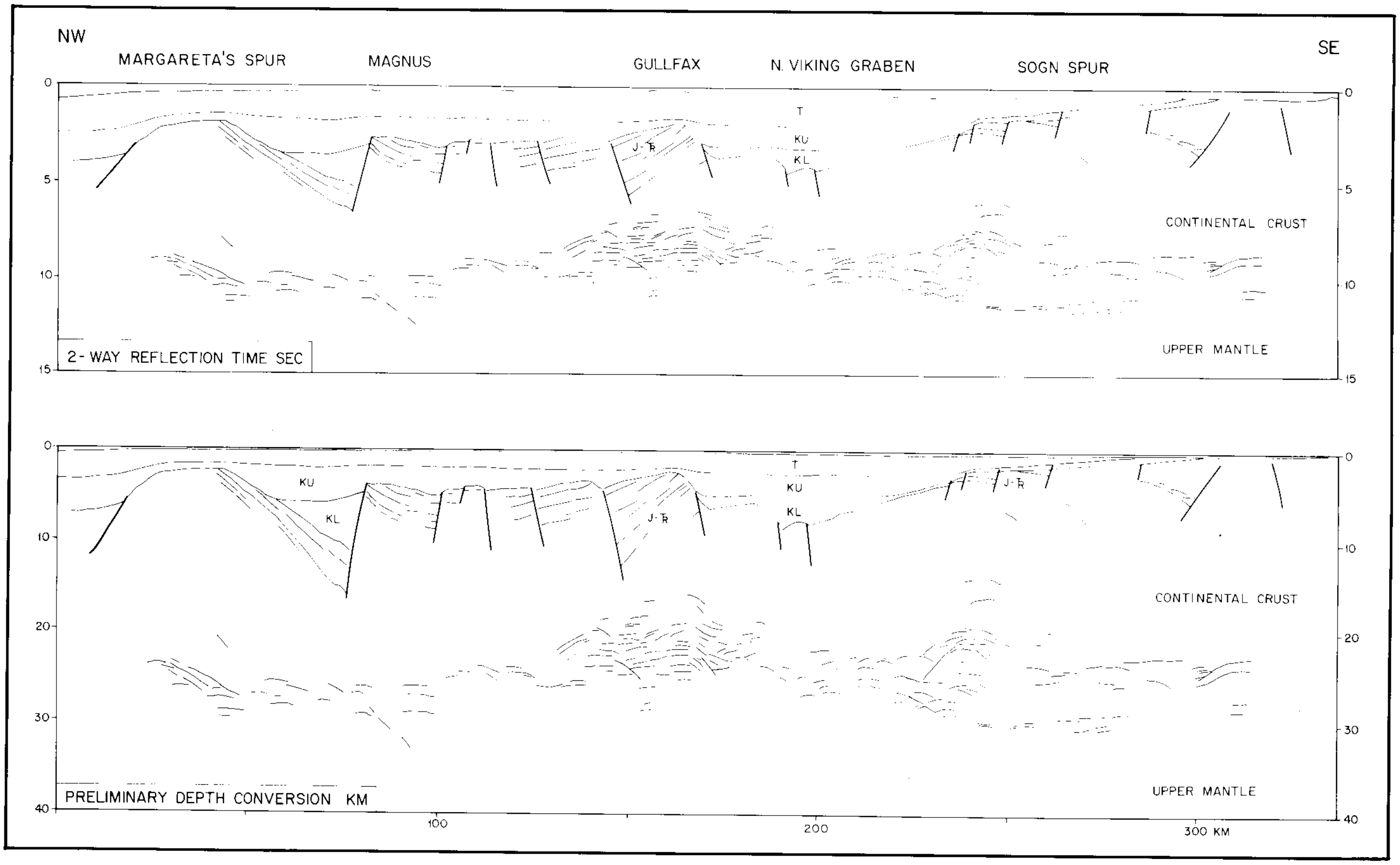

Figure 62—Seismic line drawing of a deep sounding line crossing northernmost Viking Graben, showing patterns of lower crustal reflectivity. See also Beach (1985).