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Reconstructing the tectonic history of Fennoscandia from its margins: The past 100 million years

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EQE International Limited

December 1995

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This report concerns a study which was conducted for SKB. The conclusions and viewpoints presented in the report are those of the author(s) and do not necessarily coincide with those of the client.

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Prepared by

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6 December 1995

ABSTRACT

In the absence of onland late Mesozoic and Cenozoic geological formations the tectonic history of the Baltic Shield over the past 100 million years can most readily be reconstructed from the thick sedimentary basins that surround Fennoscandia on three sides. Changing patterns of sediment thickness accompanying active tectonics, as observed on high resolution commercial multichannel seismic reflection lines, record the boundary conditions of deformation internal to the Baltic Shield. Tectonic activity around Fennoscandia through this period has been diverse but can be divided into four main periods: a) pre North Atlantic spreading ridge (100-60Ma) when transpressional deformation on the southern margins of Fennoscandia and transtensional activity to the west was associated with a NNE-SSW maximum compressive stress direction: b) the creation of the spreading ridge (60-45Ma) when there was rifting along the western margin; c) the re-arrangement of spreading axes (45-25Ma) when there was a radial compression around Fennoscandia, and d) the re-emergence of the Iceland hot-spot (25-0Ma) when the stress-field has come to accord with ridge or plume 'push'. Since 60Ma the Alpine plate boundary has had little influence on Fennoscandia. The highest levels of deformation on the margins of Fennoscandia were achieved around 85Ma, 60-55Ma and 15-10Ma, with strain-rates around 10^{-9}/year. Within the Baltic Shield long term strainrates have been around 10^{-11}/year, with little evidence for significant deformations passing into the shield from the margins. Fennoscandian Border Zone activity, which was prominent from 90-60Ma, was largely abandoned following the creation of the Norwegian Sea spreading ridge, and with the exception of the Lofoten margin, there is subsequently very little evidence for deformation passing into Fennoscandia. Renewal of modest compressional deformation in the Vøring Basin suggests that the 'Current Tectonic Regime' is of Quaternary age although the orientation of the major stress axis has remained approximately consistent since around 10Ma. The past pattern of changes suggest that in the geological near-future variations are to be anticipated in the magnitude rather than the orientation of stresses.

SAMMANFATTNING

Baltiska sköldens tektoniska historia under de sista 100 miljoner åren kan, i avsaknad på sen-mesozoiska och keneozoiska bildningar på land, bäst rekonstrueras utifrån de djupa sedimentbassängerna som i tre riktningar omger Fennoskandia. De förändringar i sedimenttjocklek som uppkommit i samband med aktiv tektonik har registrerats med hjälp av högupplösande kommersiell flerkanals-reflexionsseismik längs profiler och ger begränsningarna även för Baltiska sköldens interna deformation. Under denna tidsperiod har den tektoniska aktiviteten varit av olika karaktär runt Fennoskandia men kan indelas i fyra huvudperioder; a) perioden före det att Nordatlantiska ryggen bildades (100 - 60 miljoner år sedan) då "transpressiv" deformation i Fennoskandias södra randområde och "transtensiv" aktivitet i väster var förbunden med kompression i huvudspänningsriktningen NNO - SSV; b) bildandet av en spridningsrygg (60-45 miljoner år sedan) då den västra randzonen sprack upp; c) omfördelning av spridningsaxlarnas lägen (45 - 25 miljoner år sedan) då radiell kompression rådde runt Fennoskandia; och d) reaktiveringen av den isländska mantel plymen (25 - 0 miljoner år sedan) då spänningsfältet överensstämde med det som uppkommer vid en spridningsrygg eller mantelplym.

Från och med 60 miljoner år sedan har den alpina plattkanten haft liten påverkan på Fennoskandia. Fennoskandias randområden deformerades starkast för ca 85, 60-55 och 15-10 miljoner år sedan med töjningshastigheter runt 10^{-9}/ år. I ett längre tidsperpektiv har denna varit ca 10^{-11}/ år inne i Baltiska skölden med små bevis för att signifikanta deformationer propagerat från randområdena in i skölden.

Aktiviteten i den Fennoskandiska gränszonen, vilken var betydande för 90-60 miljoner år sedan, upphörde i stort efter det att Norska havets spridningsrygg utbildats och med undantag för Lofotens randområde finns det därför mycket få bevis för deformation inne i Fennoskandia. Förnyad måttlig kompressions-deformation i Vöring bassängen visar att den rådande tektoniska regimen är av kvartär ålder även om orienteringen hos huvudspänningsriktningarna har varit likartade alltsedan 10 miljoner år sedan. Den geologiska historien visar att de förändringar, vi har att förvänta oss inom den geologiskt sett närmaste framtiden, är variationer i storlek snarare än ändringar i riktning hos spänningsfältet.

SUMMARY

This project, entitled, "Reconstructing the tectonic history of Fennoscandia from its margins: the past 100 million years", follows an earlier study to review data of the Seismotectonics of Sweden (SKB Report 93-13). That review showed that current seismotectonics must reflect interference between two strain-fields: one tectonic, the other associated with post-glacial rebound. In consequence present-day observations of seismicity, stress or strain anywhere across Fennoscandia are likely to reflect a compound of both processes, with rebound being the dominant influence on crustal strain and consequently seismicity. In order to discover the tectonic deformation that will eventually prevail once the glacial rebound has fully recovered it is necessary to search back in time, beyond the influence of individual ice-sheets, or the collective Ice-Age.

The absence of recent marine geological formations across Sweden, makes the area particularly problematic for determining tectonic deformation over the past tens of millions of years. This absence of evidence has encouraged speculation and a number of proposals exist in the literature claiming significant levels of tectonic activity within the Baltic shield through the Tertiary and even Quaternary. However, surrounding Fennoscandia on three sides there are major sedimentary basins, that have continued to subside, accumulating sediments that record contemporary tectonic activity. As the chief influences on tectonics in the region have come from plate boundaries located to the south and west, the North Sea and Norwegian continental margin basins have acted as sensitive recorders of all strain-fields and stress-fields that have influenced the crust of Sweden over the medium term geological past. Hence, the study of tectonic activity in the surrounding basins provides an alias for the reconstruction of stress and strain within the craton.

The geology and tectonics of these sedimentary basins is today known in enormous detail, as a result of the huge investment in collecting multichannel seismic reflection profiles and borehole studies, undertaken in the search for hydrocarbons. Although all this data starts off proprietary, it has now generated numerous publications. It is now possible to access much of this data for the purposes of exploring tectonics. This resource of interpreted seismic reflection profiles provides the foundation for the present study.

The period of time chosen for this study is that of the past 100 million years, as this encompasses the creation of the North Atlantic spreading ridge plate boundary. The report follows a historical narrative, subdividing the period into a series of six phases which have distinct patterns and styles of tectonics. A number of these phases have in turn distinct pulses or episodes of tectonic activity. For each phase, having established the plate tectonic context, the regional distribution of tectonic activity is described.

During the late Cretaceous (from 85-65Ma), in the lead up to the creation of the spreading ridge through the Norwegian Sea, there were repeated episodes of dextral transpression along the Fennoscandian Border Zone and through the Central North Sea, with cumulative displacements of ca 2-3 km. At the same period there was rifting and transtensional strike-slip motion along the Møre and Vøring basins and along the south-west margin of the Barents Sea.

In the Palaeocene (65-55Ma), as an enormous mantle plume emerged beneath southern Greenland, an incipient spreading ridge developed in north-west Britain. Significant rifting accompanying further transtensional faulting occurred along the Faeroes-Shetland

Basin and the outer Norwegian continental margin. This was accompanied by transpressional deformation across much of central and Eastern Europe as well as the Fennoscandian Border Zone.

For the first ten million years as the North Atlantic spreading ridge arose after 54Ma, the new continental margin was initially passive. However, as a result of the formation of overthickened ocean crust, and the underplating of the neighbouring stretched continental crust, the outer margin remained raised.

For the twenty million years after 45Ma, as the orientation of spreading in the Norwegian Sea adjusted to the ending of spreading in the Labrador Sea, significant deformation passed into north-west Europe. As the orientation of plate separation rotated anticlockwise, the Iceland-Faeroes Fracture Zone passed into compression and became locked. A new spreading ridge began to form to the west of the Jan Mayen Block, which became rifted and rotated, and zones of sub-plate boundary deformation passed around the British Isles connecting with the Africa-Eurasia plate boundary in central Europe. The North Sea became subject to NE-SW compression to the south and E-W compression to the north leading to mild reversal (inversion) on a number of rift-bounding faults.

Following 45Ma, as the rate of sea-floor spreading declined down to less than 50% of its initial value, the tectonics of Western Europe passed into rifting and extension first on the Rhine Graben and other rifts in south-east France and later on a series of rifted basins through western Britain. However, in the mid-Miocene (ca 12-14Ma) there was a re-invigoration of the Iceland hot-spot and significant compressional deformation, that had first appeared on the mid-Norway continental margin around 25Ma, became manifest all around the continental margin of north-west Britain, as well as the northern Rockall Bank. At this same period, volcanic activity and minor plume upwellings appeared in three areas in western Europe, and the major phase of uplift of western Scandinavia was initiated. Throughout western Europe, rifting was replaced by compressional deformation.

As the rate of North Atlantic spreading speeded up again, this compressional deformation faded all along the continental margin and by the Pliocene (ca 6Ma) rapid subsidence had been initiated in the Vøring and Møre Basins of mid Norway. This subsidence involved some extensional collapse of structures previously active in compression. In the Quaternary, along at least two structures in the mid-Norway outer continental margin, minor compressional deformation has been resumed, but at a rate far less intense than that prevailing in the mid-Miocene. Around the Lofoten margin however there is also some suggestion of extensional deformation while some strike-slip displacement is seen passing into the outer Barents Sea margin, apparently along a south-easterly continuation of the Greenland Fracture Zone. Meanwhile the southern North Sea is involved in slow extension, which may also include the Fennoscandian Border Zone.

Through this period the stress-regime in Sweden has rotated from a NNE-SSW to NW-SE orientation of the maximum horizontal compressive stress direction. As indicated by the intensity of deformation along the margins, levels of stress have also varied, the highest magnitude of NW-SE stress was at around 12Ma, and having declined between 8-2Ma, has now risen again.

Strain-rates on the continental margins reached ca 10^{-9}/year at the periods of most intense deformation, probably culminating around 12Ma. Strain-rates within the Baltic Shield are considered to have been around 1% of these levels.

There is no evidence for significant zones of deformation passing into Fennoscandia through this period, and the overall configuration of transpressional and transtensional tectonics during the late Cretaceous demonstrates that deformation has had to skirt around the thick, cold and rigid lithosphere of the Baltic Shield. The past pattern of tectonic changes suggests that in the geological near-future variations are to be anticipated in the magnitude rather than the orientation of the tectonic stress regime.

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1.0 INTRODUCTION

1.1 Scope of the Report

This report takes as it time frame the past 100 million years over which to explore the 'recent' tectonic development of the Baltic Shield. This period encompasses the major change in tectonic context of the region, from being at the heart of a massive supercontinental plate, through the creation of a new plate boundary, to once again being set within a plate interior on a continental margin. The period encompasses episodes of compressional deformation and episodes of extension, through which the state of stress within the Baltic Shield has undergone significant changes. The second half of this period includes the establishment of the current plate tectonic context of western Europe, from which it is possible to constrain the style of tectonic processes to be expected into the geological near-future.

While Fennoscandia has itself remained consistently intraplate through this whole period, it has not been tectonically inert. Prior to the opening of the North Atlantic Ocean, attempts to split the supercontinent had already been proceeding for two hundred million years. Even once the new oceanic ridge had been initiated the new continental margin did not become a classic 'passive-margin'. Rates of sea-floor spreading seem to have diverged at different periods from rates of plate displacement, leading to deformation within the plate. Throughout the past fifty-five million years, western Europe has existed as a promontory of continental crust between two unstable plate boundaries: an extensional boundary to the northwest in the mid-Atlantic, and a zone of collision and shear to the south through the Mediterranean. During this period, there has been widespread deformation distributed through parts of the intervening region linking these boundaries.

The present document sets out to analyse the degree to which the pattern of tectonics, and in consequence the prevailing stress-and strain-fields, have varied through the late Cretaceous and Tertiary. The discussion also explores the causes of the tectonic activity at different periods. From such an analysis it becomes possible to place constraints on which faults may have been active within the Baltic Shield through this period, and also to estimate upper bounds of regional strain-rates. The boundary conditions of Fennoscandia also create a fundamental context within which to judge alternative models for tectonics within the shield proposed in the literature. It also becomes possible to identify the current tectonic state (as differentiated from observations of stress and strain that may reflect the ongoing effects of glacial unloading: Muir Wood, 1993). By exploring the date from when the current tectonic regime (CTR) can be said to have become established, one can gauge the time-period over which evidence for fault displacement is likely to be relevant to near-future geological hazards (Muir Wood, and Mallard, 1991).

1.2 The Tectonic Record

The function of the present document is to explore the tectonic history of

Fennoscandia from its margins. The logic of this exercise needs to be stated. Across the whole of Fennoscandia (ie Norway, Sweden and Finland) but neglecting Denmark) with the exception of a small area of Skåne in southernmost Sweden, the surviving outcrops of Mesozoic and Tertiary sediments are too small and scattered to enable any, but the most meagre, tectonic reconstruction. However to the south, west and north of Fennoscandia there are thick Mesozoic and (to the west) Cenozoic sedimentary basins, that have been subject to continuous or near-continuous sedimentation. Such basins record prevailing tectonic regimes, with local deformation in response to changing stress-fields, and changes in sedimentation in response to local uplifts and subsidence.

The existence of these sedimentary basins is not in itself accidental but reflects the localisation of strain that occurred during the intraplate extension that prevailed across north-western Europe throughout almost all the Mesozoic. Strain became localised because the lithosphere was weak (probably as a result of its thermally dominated rheology) in comparison with neighbouring areas. Like breaking a stick, once failure has been initiated, it becomes localised along a single lithospheric zone, rather than distributed on many parallel, widely separated zones. The current shape of Fennoscandia remains outlined by these zones of lithospheric weakness. In consequence, these basins have remained regions prone to concentrating strain: any new pattern of stress imposed on the region is first likely to provoke strain on its rifted margins. Hence the tectonic history of these rifted margins can in turn be seen to be an exaggerated or amplified version of the strain that was relieved within Fennoscandia itself through the same period.

The sedimentary record preserved in these peripheral basins is itself chiefly the result of the original phases of rifting. Crustal extension was accompanied by lithospheric thinning that led to a rise in the asthenosphere in compensation. Thermal cooling of the lithosphere and isostatic loading in response to the rifting subsidence has led to continued sinking for many of these basins (McKenzie, 1977). Even after these rifted margins had ceased extending, and in some cases become subject to compressional reverse faulting (and consequently mild crustal shortening), at least to the west there was sufficient prevailing subsidence to continue to preserve the tectonic record.

Suitable sedimentary basins for preserving the 100Ma tectonic record are found all along the western margins of Fennoscandia. However to the south and north, as a result of late uplift, these basins contain a less complete record in passing to the east, and to the east itself there are only platform sediments on the Russian and Ukrainian shields. Although incomplete, the breaks in this ring correspond with directions in which there has been little if any plate tectonic influence through the last 100 million years. There is no evidence for significant deformation arriving from further into Russia, at least since the end of the Palaeozoic, and for the past 100 million years the nearest plate boundary to the east has been in the Caucasus and Siberia. The geological record from the Barents Sea confirms that tectonic activity through the Mesozoic became increasingly concentrated to the west. To the south of Fennoscandia, while an important phase of tectonic activity, interlinking the Alpine and North Atlantic plate boundaries, occurred along the Fennoscandian Border Zone in the late Cretaceous, throughout the Tertiary these sub-plate boundary zones of interlinkage have shifted to the west, passing through Britain and the western North Sea.

1.2.1 The Tectono-Sedimentary Record

The principal elements of data relevant to an analysis of tectonics are geological formations that have been affected by some local deformation. By far the most important of these formations are those that can be shown to have developed contemporaneously with the deformation. These include: sedimentary deposits that have accumulated concurrently with fault-related uplift or subsidence, and igneous intrusives that provide evidence of crustal dilation within the prevailing stress field.

As the Mesozoic and Tertiary sediments that formerly covered parts of Fennoscandia have almost all been lost, most of the information contained in this report is derived from offshore. However, all offshore dates require: a borehole that has sampled the relevant formation; a geologist who considers that the sample is worth investigating, and a company or organisation that makes the data available. The structures themselves are best imaged on multi-channel seismic reflection profiles almost all of which have been collected in the course of hydrocarbon exploration. While such commercial data starts off confidential, fortunately the culture of oil exploration in Scandinavia has always been relatively open. Inevitably much more information is available in those areas in which hydrocarbon exploration is mature, such as the North Sea, than in frontier areas awaiting drilling, such as the Møre Basin. For fields for which production is well underway detailed information is often released, while for fields that await development this is often held back for reasons of commercial sensitivity.

In the beginning seismic reflection data was focused on the anticipated depth range of likely hydrocarbon reservoirs, reducing the resolution of younger and shallower reflectors. More recently late-stage tectonic evolution has become increasingly recognised as a major factor in determining hydrocarbon migration and reservoir development and in consequence seismic reflection acquisition has been 'tuned' to investigate the higher parts of the profiles. This present report reflects access to the full volume of published data, as well as some seismic reflection data that remains confidential and hence cannot be presented. Wherever possible published (although sometimes necessarily inferior) examples are displayed.

The tectonic history presented here will inevitably undergo refinement as new dates become available. The logic guiding this history is that, at a regional scale, the tectonics of any given period is constrained by two actualistic rules: that the contemporary stress field is regionally consistent, and that zones of concentrated deformation cannot exist in isolation. These are no less than the fundamental tenets of plate tectonics.

1.3 Time-frames

In order to explore the tectonic development of northwest Europe through the Tertiary, it is necessary to establish a detailed chronology, made possible as Tertiary stratigraphic stages have been refined and correlated with an 'absolute' time-scale based on radio-isotope dating. However, most sedimentary geological formations have not been directly dated and, hence, dates have had to be obtained through correlation.

There are two principal routes for such correlation:

- Fossils in marine sediments provide the opportunity for detailed stratigraphic correlation with formations that have been directly dated, perhaps through the occurrence of contemporary volcanic materials.

- For igneous rocks themselves, and particularly long-lasting sequences of lavas or shallow intrusives, there may be a possibility for magnetostratigraphy employing the polarity of the Earth's magnetic field recorded in a rock as it cools through the Curie temperature. This technique is of greatest importance in the oceanic crust where the age can be reconstructed through identifying the sequence of recorded magnetic anomalies imprinted as the crust has been generated at a spreading ridge.

There are two or three time-scales in current use, that show small variations to each other. The time-scale employed in this report is that of Harland et al., (1989) (see Figure 1.1).

1.4 Geographical Scope

The geographical scope of this Volume does not remain constant with time as, at every period, the area is intended to be large enough to encompass the major controls of tectonic activity. For most periods, this requires a consideration of the region up to, and including, the nearest plate boundary. As the relevant plate boundaries receded from Fennoscandia through the Tertiary, the area of coverage expands. For example, in the Palaeocene, the area is largely restricted to the surrounding continental shelf because tectonic activity was dominantly associated with the incipient formation of a new constructive plate boundary. Subsequent to the beginning of the Eocene it is hard to find any significant influence of the Eurasia-Africa plate boundary on deformation in Fennoscandia (in marked contrast to the situation in the British Isles).

These general principles apply for all azimuths to the north, west and south of Fennoscandia but cannot so readily be applied to the east, where there was no plate boundary within several thousand kilometres. A general eastern limit has, therefore, been taken approximately along the boundary of the former USSR.

In order to make some sense of the tectonic development it is necessary to employ the names of the key structural features. A key map is provided in Figure 1.2 to assist comprehension, and a gazetteer of all the named geological features is found in Appendix A.

1.5 Plate Tectonic Context

To understand the driving forces and causes of Tertiary deformation in western Europe, it is necessary to place this region in the context of the plate boundaries that have surrounded it.

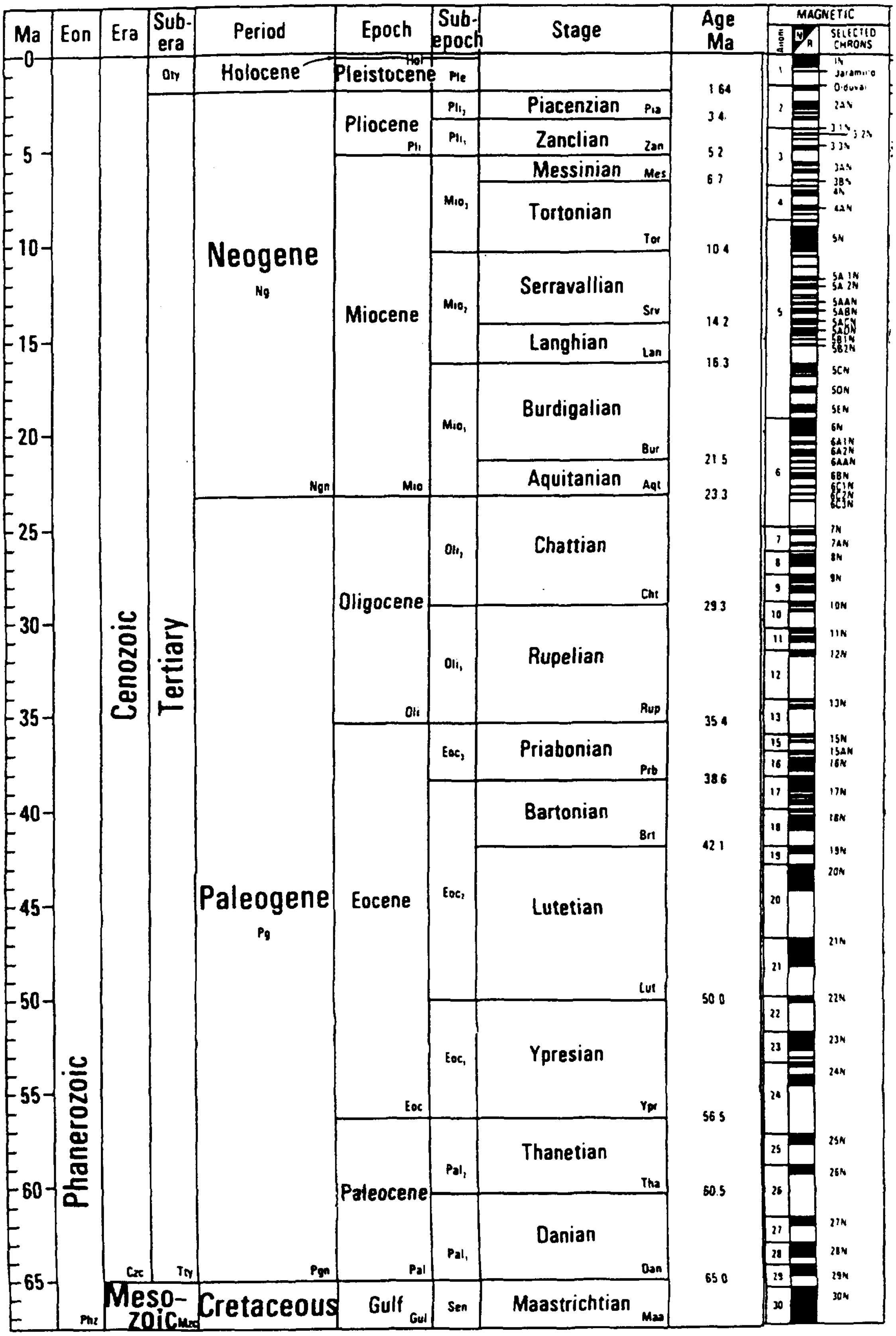
The changing nature of the plate boundary to the west of Fennoscandia is well preserved in the pattern of sea floor spreading mapped in the Norwegian Sea. For the Africa-Eurasia plate boundary to the south, the pole of relative rotation through the Tertiary can only be reconstructed from the configuration of sea floor magnetic anomalies that define the North America-Eurasia and North America-Africa spreading histories. Because the position of this pole is calculated from the difference between two reconstructed spreading histories, it is liable to errors. Nevertheless, all authors agree it has persisted close to the western end of the plate boundary, where the plate motions have been constrained to pivot by the difficulty of forcing a compressional boundary through the Atlantic ocean floor.

1.6 Structure of this Volume

The overall architecture of this Volume is in the form of a historical narrative, starting from the late Cretaceous, subdivided according to a tectonic 'logic'.

Section 2 is a description of the tectonic context of the late Cretaceous preceding the northward expansion of the North Atlantic spreading ridge. Section 3 provides a detailed discussion of the Palaeocene, up to the initiation of spreading in the North Atlantic. The early history of spreading, and the transform fault activity along the Barents Sea margin through the Eocene is considered in Section 4. Section 5 concerns the Oligocene deformation that accompanied the re-arrangement of spreading directions in the North Atlantic. By the mid to late Oligocene transtensional tectonics dominated across much of northwest Europe. During the Miocene period (Section 6), there was the reinvigoration of the Iceland hot-spot, that was accompanied by compressional tectonic activity along the continental margin of the Norwegian Sea. Section 7 discusses the Plio-Quaternary period when the compressional margin showed signs of returning to extension, and then (at least locally) being subject to renewed compression.

Each of the Sections begins with a description of the prevailing plate tectonic context during the relevant period. This is followed by a geographical tour around Fennoscandia to explore the detailed picture of contemporary tectonic activity. The present document concludes (Section 8) with a review of the tectonic development in terms of the consequent stress-history and potential for internal deformation within the Baltic Shield through this period. The implications of this tectonic activity lead on to consider the role of present-day and future tectonics.



TIME-SCALE FOR THE TERTIARY
(FROM HARLAND ET AL., 1990)

Figure 1.1 **Time-scale for the Tertiary (from Harland et al., 1990)**

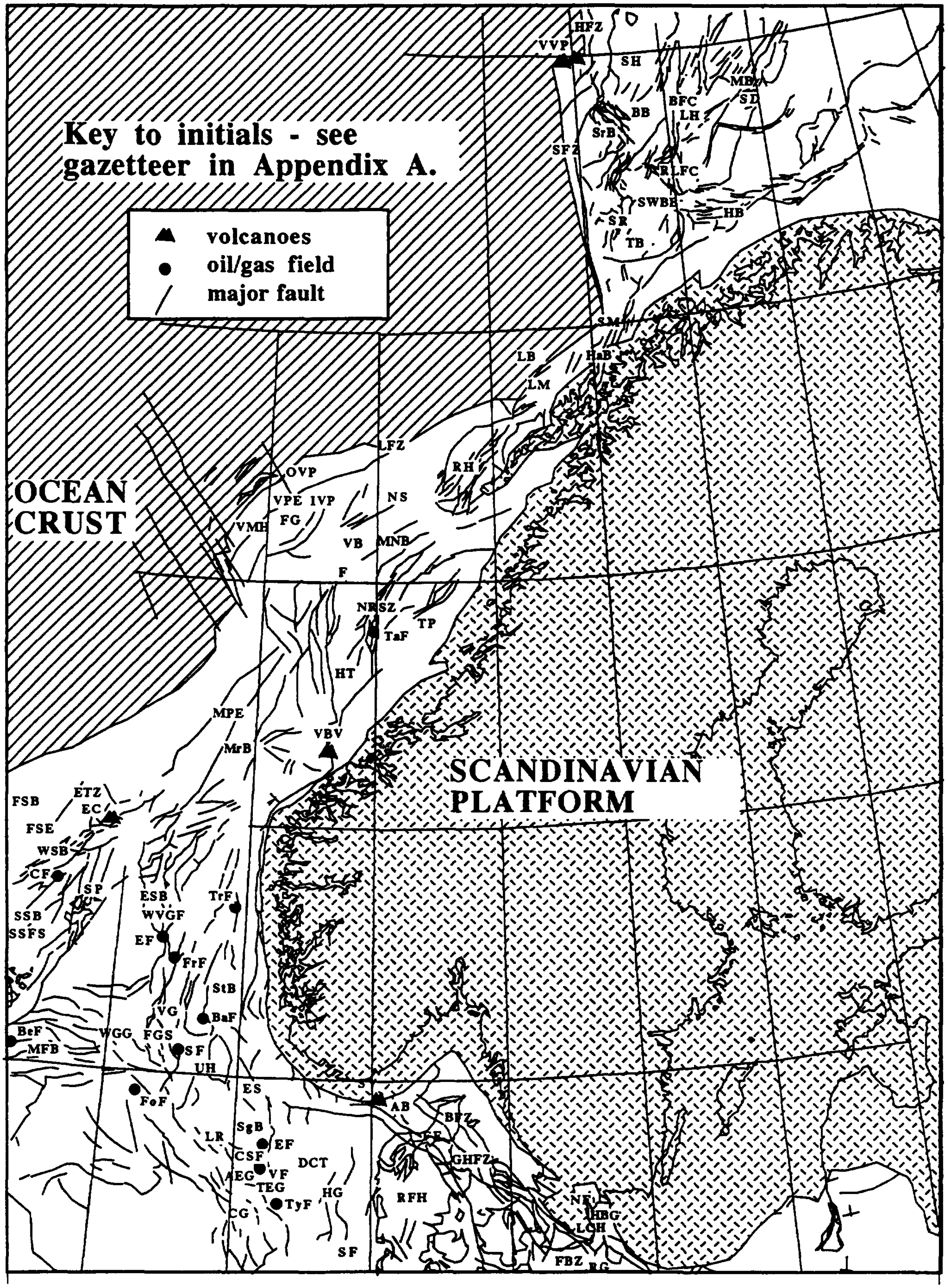


Figure 1.2 Map of named structures

2.0 LATE CRETACEOUS

2.1 Plate Tectonic Context

The North Atlantic spreading ridge developed in the early Cretaceous, having propagated from the south where it had been established in the Jurassic. Masson and Miles (1986) dated the initial spreading between Iberia and Newfoundland at 112-118Ma, in the Bay of Biscay at 108Ma, and between Ireland and Newfoundland at 100Ma. The creation of this spreading ridge effectively ended the major Mesozoic episodes of rifting and crustal extension found across the North Sea. To the north of the advancing Atlantic spreading ridge rifting became concentrated to the west. During the early Cretaceous there was pronounced rifting along much of the present western continental margin of Scandinavia, including the Møre, Vøring and SW Barents Sea areas.

2.1.1 Rockall Trough

The Rockall Trough, located to the west of Ireland, reflects about 120km of WNW-ESE extension accumulated in a series of episodes through the Mesozoic. The last of these occurred in Coniacian to Santonian times (88-5 - 83Ma) (Knott et al., 1993), with a stretching factor of 1.3, implying ca. 40km overall opening, decreasing to the north-east. In the late Santonian the Rockall Trough became abandoned as a site of prominent rifting when the North Atlantic spreading ridge expanded into the Labrador Sea separating Greenland from North America (Masson and Miles, 1986). This westward jump in the spreading axis created the Charlie Gibbs Fracture Zone with its major left-stepping offset (Kristofferson, 1978). From this time, the plate boundary to the north of the Labrador Sea became transpressional through the Canadian Arctic islands (Miall, 1984).

2.1.2 Shetland-Faeroe Trough

The major episode of rifting in the Rockall Trough can be traced along the Faeroes-Shetland Trough where NW-SE extension resumed in the Turonian (ca. 90Ma) (Duindaan and van Hoorn, 1987), leading to the accumulation of up to 3000m of sediment fill associated with ENE-WSW trending normal faulting developed along the northwestern flank of the Rona Ridge (Turner and Scrutton (1993). The Solan and Shetland Spine Faults were reactivated as normal faults at this time, downthrowing to the north with up to 1500m of Upper Cretaceous sediments being deposited in the Solan and West Shetland basins. Seismic evidence suggests late Cretaceous sediments also formerly extended across the West Shetland Platform. Fault movements continued sporadically into Maastrichtian times (74-65Ma) with a pulse of Campanian (83-74Ma) faulting and footwall uplift (Hitchen and Ritchie, 1987).

Between the Faeroes and Rockall Basins there was some prominent transpressional deformation accompanying this episode. In the E-W trending Mesozoic Solan Basin, there was partial basin inversion involving the extrusion of the Permian-Lower Cretaceous sedimentary fill towards the northwest onto the dipslope of the half-graben with reversal of pre-existing ENE-WSW normal faults and the development of thrusts. Along the southern basin margin there was high-angle reverse faulting and folding. This shortlived but intense episode is dated stratigraphically as postdating the latest Cenomanian limestone, while sediments that postdate the fold are of earliest Turonian to Santonian age: ie ca 90Ma (Booth et al., 1993). The localisation of this compression indicates transpression probably accompanying the transfer of rifting at the northern end of the Rockall Trough.

2.1.3 *Mid-Norway Basins*

Further to the north, in both the Møre and Vøring Basins, Early Cretaceous rifting was followed by thermal subsidence giving rise to thick Upper Cretaceous sedimentation. Relatively minor late Cretaceous extensional faulting is seen along NE-SW trending structures in the Haltenbanken and Traenabanken areas on the eastern edge of the Vøring Basin (Gowers and Lunde, 1984). Skogsheid and Pedersen (1992) claimed the initiation of more significant fault-defined extension (in the Fenris Graben), occurred at the Campanian-Maastrichtian transition (ca 75Ma), associated with the initiation of the forthcoming Atlantic spreading ridge. This coincides with the resumption of rifting and volcanism in both the Faeroes-Shetland Basin and Rockall Trough. Volcanic centres formed around the northern margins of the Rockall Trough, and in the Erlend Centre north of Shetland, through the Campanian and Maastrichtian.

2.1.4 *Barents Sea*

Further to the north the southwestern Barents Sea had developed deep early Cretaceous basins at Harstad, Tromsø, Bjornøya and Sørvestsnaget, now separated by intra-basinal highs (Senja Ridge, Veslemøy High and Stappen Highs) (Gabrielsen et al., 1990). (The eastern platform area of the SW Barents Sea was chiefly affected by Jurassic rifting without significant post-rift subsidence. Rifting ceased in the Hammerfest Basin in the early Barremian when tectonic decoupling occurred along the Ringvassøy-Loppa Fault Complex.) Deep late Cretaceous subsidence occurred in the Sørvestsnaget and Harstad basins located at rift-shear intersections. The Tromsø and Sørvestsnaget Basins continued to subside during the late Cretaceous while areas to the east have a condensed sequence. Local late Cretaceous compressional deformation (Gabrielsen et al., 1990) indicates minor wrench components along some of the larger faults.

A number of the intrabasinal highs reflect the impact of late Cretaceous tectonics (although overprinted by the more dominant early Tertiary deformation). The generation of the Senja Ridge and the Veslemøy High (that offsets the Tromsø and Bjornøya Basins), may reflect strike-slip movements (both transtensional and transpressional) along the Bjornøyrenna Fault Complex: Riis et al., (1986) proposed sinistral shear during the early Cretaceous, while Gabrielsen and Faerseth (1988) claimed dextral transpression. Local deformation at the transition between the Senja Ridge and Veslemøy High, during the post-Cenomanian - pre-late Palaeocene period, was considered by Faleide et al., (1993) to reflect a small lateral component, although the western boundary faults of both highs are indicative of normal faulting during the late Cretaceous. At the Veslemøy High and in the Bjornøya Basin, distributed late Cretaceous extension is seen with numerous normal faults with throws less than 300m. An ENE trending normal fault with a throw of 750-1500m down to the northwest runs along the northwestern border of the Veslemøy High and a similar fault marks the western border of the Senja Ridge (Faleide et al., 1993).

The eastern extensional boundary faults of the Loppa High had their main activity in the early Cretaceous, but there is evidence for some reactivation in the late Cretaceous, locally with a slight compressional component (Gabrielsen et al., 1990).

In summary the late Cretaceous tectonics of the SW Barents Sea margin was dominated by NW-SE extension although there was significant accompanying strike-slip activity, some of which has been locally transpressional.

2.2 Africa-Eurasia Collision Zone

In the Bay of Biscay, between Anomalies 34 and 31 (84-69Ma), there was fan-shaped spreading in association with anticlockwise rotation of the Iberian block. This spreading accompanied the initiation of collision along the line of the Pyrenees where, since the late Cretaceous, there has been an estimated cumulative 147km of shortening (Munoz, 1991). Between Magnetic Anomalies 31 and 25, (69-58Ma) Roest and Srivastava (1991) define a pole of relative rotation between Iberia and Eurasia at 28.7°N 7.43°W, while Dewey et al. (1989) give 31.52°N 10.73°W at Anomaly 30, both locations implying slow extension along an E-W zone through the outer Bay of Biscay and dextral transpressive motion along the North Pyreneean Fault Zone. An estimated 75km offset along this boundary is projected at this period.

2.2.1 Alpine Collision Zone

Detailed stratigraphic studies of the basins distributed along the flanks of the mountains indicate that the initiation of compressional deformation along the zone of Lower Cretaceous rifts was in the Upper Santonian (83Ma) (Puigdefabregas et al., 1991). Santonian to Campanian age turbidites, which developed in advance of the Pyreneean thrusts, are up to 6000m thick to the north and 2000m in the south. It was at this period that compressional deformation was also transferred into the basins around the southern North Sea. Thrusting in the southern Pyrenees continued at a slower rate from the uppermost Turonian to the Maastrichtian. The region to the north was largely unaffected by tectonic activity through this later period.

The Alpine plate boundary across the centre of Europe was already in existence in the Cretaceous, involving a complex of transcurrent faulting passing through what has now become the Alpine mountain chain and the eastern Mediterranean. Until the mid-Cretaceous, deformation in central and northern Europe to the north of the Alps was dominated by extensional tectonics but, once the Atlantic sea floor spreading had passed to the north of Iberia, this was replaced by compressional or transpressional deformation.

In the late Cretaceous, the orogenic front reached the Helvetic shelf of the central and eastern Alps and the Carpathians. Initially collision did not impact into the foreland of the western Alps (Ziegler, 1987a) and the first episode of such deformation (known as the 'sub-Hercynian') was during the late Turonian to early Senonian (ca. 89Ma) (Ziegler, 1987a) when transpressive deformation was channelled along major NW-SE trending Variscan fault systems, into the foreland to the north. Such deformations (involving partial basin inversion) are found from the Polish Trough in the east to the Sole Pit, Broad Fourteens and West Netherlands Basins in the west and as far north as the Fennoscandian Border Zone. Closer to the Alpine Front, around the Bohemian Massif, major basement blocks became uplifted along high-angle Hercynian reverse faults. This episode of compressional tectonics was contemporary with rifting in the Rockall Trough. Deformation in the Alpine foreland was of lesser consequence between the late Campanian and Danian.

2.2.2 *Fennoscandian Border Zone*

The dominant influence of transcurrent fault motions in the region to the north of the Alps is indicated by areas of significant basin inversion located adjacent to zones of greatest subsidence. Over many of the basinal areas, the Chalk is more than 500m thick, and along the axes of the Viking and Central Grabens there is more than 1000m of Chalk. The greatest thickness (2200m) is, however, located in a trough passing NW-SE through Denmark, along the line of the Fennoscandian Border Zone.

The Fennoscandian Border Zone marks the northern edge of the broad zone of shear passing through central Europe, connecting the Alpine plate boundary with the Rockall Trough rift-zone. (The FBZ demarcates the south-western edge of the Baltic shield, and was the site of pronounced transtensional rifting, in particular in the Triassic when the Rønne and Ålborg Grabens accumulated up to 6000m of sediments.) An extensional tectonic regime persisted through the Jurassic and early Cretaceous, but switched to dextral transpression in the late Cretaceous (Liboriussen et al., 1987). In the Polish Trough there is up to 2000m of late Cretaceous sediment deposited in linear furrows running parallel with the flanks of the major late Cretaceous NW-SE inversion uplift, that is typically 50-80km wide (Pozaryski and Brochwicz-Lewinski, 1978).

The Fennoscandian Border Zone undergoes a step north across the NNE-SSW Rønne Graben to the northwest of Bornholm. In the Hanø Bay

Graben, located to the north of the main axis of inversion, no sediments younger than Campanian are preserved. (There is however strong evidence for post-Campanian displacements on the Navlinge Flexure and the northern scarp of the Linerodsåsen-Chrstiansø Horst (Kumpas, 1980)). During the late Cretaceous the axial zones of the FBZ became uplifted to form elongated islands or a peninsula along the margins of the Fennoscandian shield. Clastic sediments from off this landmass are found in Bornholm (Surlyk, 1980), and Øresund (Larsen et al., 1986) and debris flows are seen on seismic reflection profiles across the flanking basins in the Kattegat (Figure 2.1). The deepest part of the late Cretaceous basin lies adjacent to the Kattegat inversion, with more than 2000m of Upper-Cretaceous -Danian sediments beneath the north-east corner of Sjaelland (Libiorussen et al., 1987) (see Figure 2.2). To the north of Sjaelland the depocentre is chiefly controlled by the WNW trending Grena-Helsingør Fault Zone to the south and the Borglum Fault Zone to the north (Aubert, 1988). On seismic lines, flower and palm structures reveal the important of dextral transcurrent displacements along this system. The intensity of this inversion decays to the north-west, fading out in the Skagerrak.

Faults involved in transpression imply a north to northeasterly principal compressive stress orientation, and the uplifts reflect right-stepping jogs of the controlling WNW-ESE trending faults. Well-calibration of seismic data in the Kattegat shows that inversion began in the Coniacian and accelerated rapidly in the Santonian-Campanian (ca. 83Ma) (Liboriussen et al., 1987), concurrent with Rockall Trough rifting. Uplift decreased in the Maastrichtian-Danian (but was renewed in the Palaeocene, see Section 3).

2.2.3 *Central North Sea*

The inversion seen along the Fennoscandian Border Zone is echoed along the Danish Central Trough (Vejbaek and Andersen, 1986) around the NW-SE trending Lindesnes Ridge and the Tail End Graben. Thermal subsidence in the North Sea has preserved a complete record of late Cretaceous sedimentation from which it is possible to date phases of activity with considerable precision. Late Cretaceous reverse faulting is dominantly found along a series of NW-SE to NNW-SSE trending structures bounded by WNW-ESE faults, that appear to reflect the bounding strike slip faults. The overall configuration of faulting suggests a braided western continuation of overall dextral shear transferred from the Fennoscandian Border Zone, in particular along faults associated with the Rynkøbing Fyn High. Deformation appears to converge to the south of Sweden and the Polish Trough. As in the Fennoscandian Border Zone depocentres are located adjacent to the reverse faults with an overall vertical relief of more than 1000m (Michelsen, 1982).

In the eastern North Sea, deformation was relatively intense in the Turonian-Santonian period (90 -85Ma), while from the Late Santonian to Early Campanian there was little activity. Inversion was however intensified once again in the Late Campanian to Early Maastrichtian, (ca 75Ma). Another uplift event can be specifically dated to the mid-

Maastrichtian: ca 70Ma. These individual events, although sharing a common tectonic style, affected different regions: the Turonian-Santonian deformation being concentrated to the north and west of the Danish Central Graben while the Arne-Elin Graben was active at all periods. Structures related to this pattern of deformation control a number of oil-field reservoirs, including the Valhall, Tor, Edda and Ekofisk Fields (Munns, 1984; d'Heur, 1984; d'Heur et al., 1985)

This same pattern of deformation can be found to the north in the Egersund Basin (Oie, 1990). Reversal is found on two series of en echelon fault segments trending overall NW-SE to NNW-SSE but made up from individual WNW-ESE fault segments, located on either side of the Egersund Basin. The northeasterly line runs for almost 100km. Reverse faulting has been dated to two periods: first in the Turonian and second in the early Maastrichtian, contemporary with movements mapped to the south.

This dextral transpressional shear is clearly related to phases of rifting seen on what became the North Atlantic continental margin. Transpression seems chiefly to be concentrated around the margins of the Baltic shield. The concentration of deformation has the makings of a 'sub-plate boundary', active with a similar character and course at several episodes during the late Cretaceous. On each occasion the transfer of displacement through to the Alpine plate boundary must accompany an increase in rifting that passes along the Norwegian continental margin. The overall amount of crustal shortening across the shear-zone implies ca. 3-5km of overall horizontal displacement.

2.3 End-Cretaceous Emergence

The Chalk seas represent a period of high global eustatic sea-levels, general submergence in northern Europe, and an absence of significant quantities of detrital sediment input. Upper Cretaceous detrital sedimentation persisted only in the northernmost part of the Viking Graben north of 59°20', with a source presumably in Greenland. Those tectonically-induced elevations that can be identified were relatively subdued and generally provided only earlier Chalk to be redissolved, leaving almost no detrital remnant.

Sedimentation was continuous from the Cretaceous into the Tertiary in relatively few areas: most importantly, in and around the Central and Viking Grabens in the North Sea and across parts of northern Denmark (where the Chalk seas survived into the Palaeocene and up to 700m of Maastrichtian chalks developed along the tectonically-active Fennoscandian Border Zone). The Upper Maastrichtian shows a distinct shallowing over the marginal parts of the basin, probably recording the regional regression, and the Maastrichtian-Danian boundary is almost everywhere marked by a marl layer about 10cm thick corresponding to a break in carbonate deposition. Chalk sedimentation was resumed following a revival of coccolith production in early Danian times, although there was continued regression with considerable submarine topographic relief around bryozoan reefs.

2.4 Constraints on Late Cretaceous Deformation in the Baltic Shield

During the late Cretaceous there was active deformation on the rifted basins located on both the south-west and western margins of the Baltic Shield. This deformation everywhere involved a component of strike-slip displacement with dextral movements and partial transpression along the Fennoscandian Border Zone and rifting and transtension along parts of the Norwegian continental margin. The concentration of deformation along the Fennoscandian Border zone relative to other lesser fault displacements in the Central North Sea must reflect the marked contrast in crustal rheologies that exists on this margin of the thick cold lithosphere of the Baltic Shield. This border zone has acted as a 'crumple-zone' within the regional deformation.

There is no evidence to indicate that any significant deformation passed into the shield itself at this period. Structures almost everywhere run parallel with its margins. The regional style of deformation could however be consistent with minor reactivation of appropriately oriented rifted structures, such as the Oslo Graben.

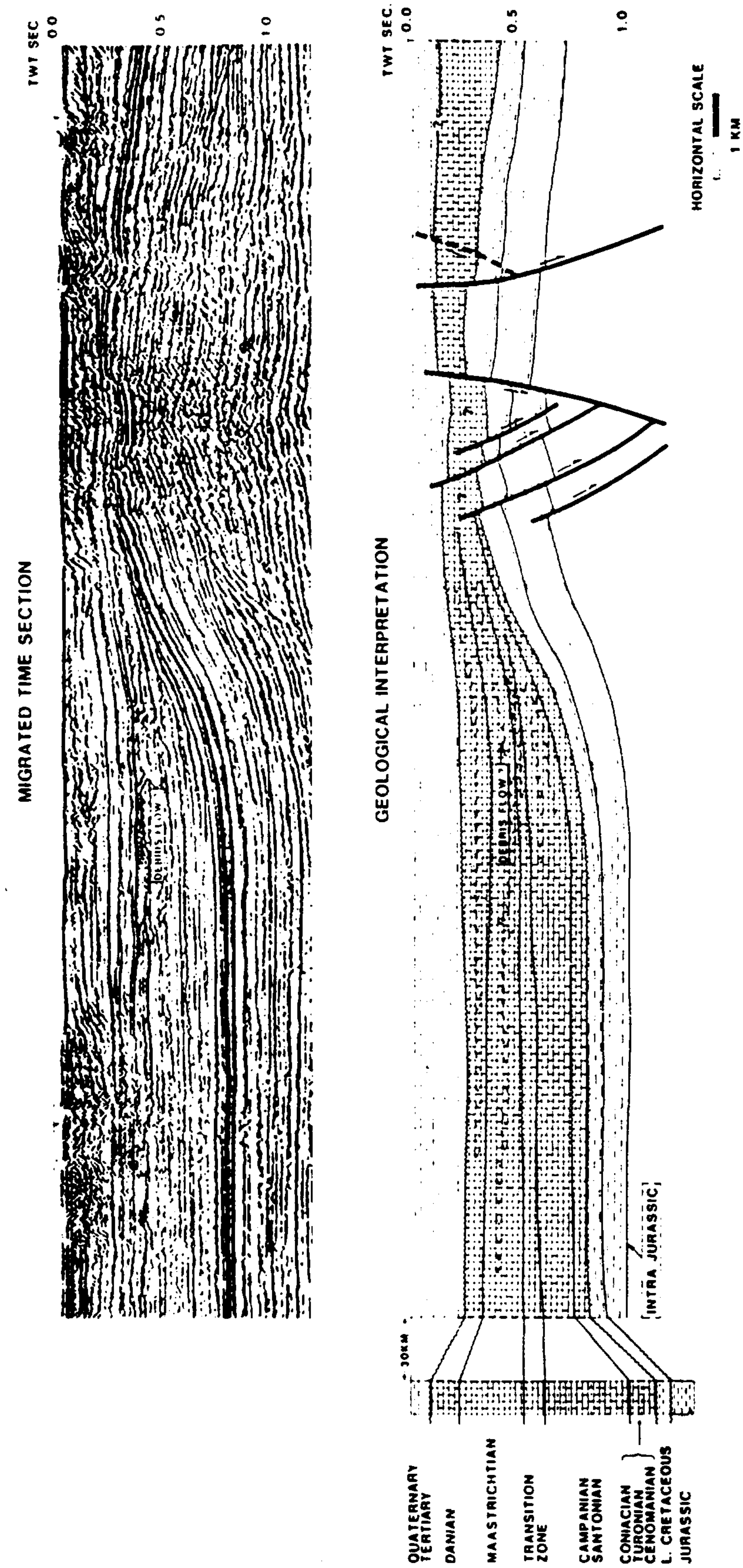


Figure 2.1 **Seismic section and interpretation across Late Cretaceous and Danian Inversions in the western Kattegat (from Liboriussen et al., 1987)**

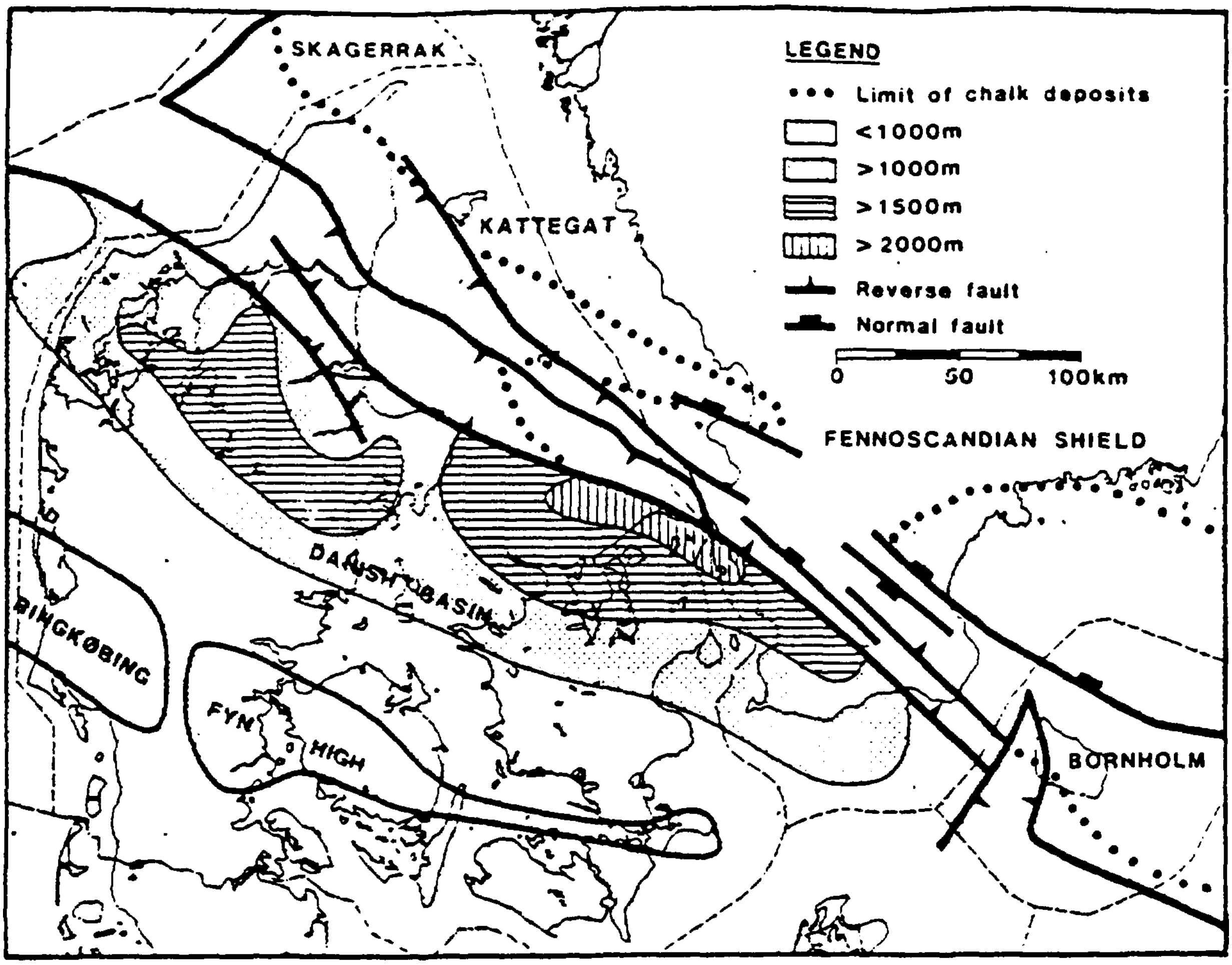


Figure 2.2 Isopach map of the Upper Cretaceous - Danian sequence around the Fennoscandian Border Zone (from Liboriussen et al., 1987)

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3.0 PALAEOCENE

3.1 Plate Tectonic Context

At the beginning of the Palaeocene, the North Atlantic spreading ridge, first established in the Campanian, was offset by two pronounced left-stepping transforms into the Labrador Sea. However, as a result of the arrival of a massive mantle plume diapir hot-spot, centred over eastern Greenland, but extending all the way from Ireland to western Greenland, the pattern of spreading became temporarily displaced with an attempt to create a new constructional plate boundary along the line of the North Atlantic, involving a short spreading ridge passing through the north-west British Isles (see Figure 3.1).

3.1.1 The Hebridean Spreading Ridge

The Hebridean volcanic province in the NW British Isles, represents the arrival of a major pulse of upwelling asthenospheric material, coinciding with a short-lived but rapid episode of crustal dilation, to create a nascent spreading ridge. Combined palaeomagnetic and radiometric age-dating studies demonstrate the short duration of volcanism (Musset, 1984): the Rhum and Skye central complexes and lavas all fall within the single episode of reversed polarity around 60Ma, correlated with the Interval 26R, see Figure 1.1. Magma was released into a series of intense dyke swarms some of which fed rifted and subsiding lava basins. The main igneous centres lie on a major N-S zone passing from the Faeroes through the west Scottish centres down to Lundy, the orientation of an underlying ridge of upwelling.

The northernmost section of 'spreading ridge' is found in the lava-basin at Skye, where the dyke swarms provided around 1.5km of ENE-WSW extension (Speight et al., 1982). The next lava-basin to the south is at Mull, offset to that at Skye, and overlying a wall of lower crustal intrusions (Hipkin and Hussein, 1983). In Mull, the lava base is located 400m below sea-level, while in Skye the projected depth is -1500m. The southernmost lava basin is in Antrim, northern Ireland. Such en-echelon extensional basins have operated like offset mid-ocean spreading ridges, comparable to the current Reykjanes Ridge seen onland in south-east Iceland.

A complex set of Palaeocene dykes runs across Ireland, bending around to run E-W, or even WSW-ENE, on the west coast around Connemara. This zone represents a connection between the Hebridean igneous province and a largely separate 'Porcupine Igneous Province' (Tate and Dobson, 1988) around the northern end of the Porcupine Basin, to the southeast of Ireland. Palaeocene crustal dilation in the Porcupine Basin was not dominantly related to dyke intrusion but was also accomplished by subsidence and some extensional rifting involving the reactivation of N-S faults along both sides of the Porcupine Basin converging at the northern end of the basin close to the Slyne Fissures (Naylor and Anstey, 1987). White et al. (1992) estimated a Beta value of 1.05 for the Palaeocene extension in the

Porcupine Basin which, across the approximately 80km wide basin, would represent ca. 4km of E-W elongation. Palaeocene rifting in the Porcupine Basin is itself bounded to the south by the prolongation of the Clare Lineament, the landward continuation of the Charlie Gibbs Fracture Zone. Hence, the extension in the basin represents a short-lived or partial switch of some of the North Atlantic spreading to the east. The extension seen in the Porcupine Basin and the contemporary extension associated with the dyke swarms of the main Hebridean Province (from Antrim to Skye) are clearly inter-related and must have communicated through some intervening transform fault zone or system of faults passing WSW-ENE through the centre of Ireland.

The pattern of NNW-SSE trending Hebridean dyke swarms reveals the contemporary regional stress field. At the heart of the province, the dyke orientations are regularly aligned. Close to the major volcanic centres on Skye and Arran, a second minor set of dykes is oriented ENE-WSW. At St. Kilda, an earlier NW-SE set of dykes is cut by a later NE-SW system.

3.1.2 East Shetland Platform

Tectonic activity related to this episode can also be traced to the north of the Hebridean volcanic province. At the end of the Danian there was an episode of rapid uplift of the Scotland Shetland platform, heralding the initiation of Hebridean volcanism accompanied by subsidence on the eastern margin of the Shetland Platform (Morton, 1979). Slump and debris flows began to accumulate in the formerly sediment starved basin in the late Danian, and by the early Thanetian the river system draining uplifted Scotland had become focused onto the NW Moray Firth Basin. During Phase 1 (Unit B of Knox et al., 1981, Sequence 2 of Stewart, 1987), two en echelon segments of NNE-SSW faulting on the southern end of the Viking Graben appear to have bounded half-graben style basins to the east, which filled with more than 500m of clastic sediments, mainly derived from Upper Jurassic sands, eroded off the uplifted Shetland Platform. Contemporary subsidence and deposition was minimal to the north and east where only thin shale equivalents were deposited. Within the outer Moray Firth, sedimentary units extend towards the southeast and 200km from the contemporary land margin are still 300m in thickness.

Phase 2 (Unit C of Knox et al., 1981, Sequence 3 of Stewart, 1987) formed in response to subsidence, by up to 800m, along the NNW oriented northern half of the western Viking Graben margin. The half-graben style basins became infilled with sands interbedded with shales, commonly with sudden facies variations both laterally and vertically. A shallow shelf environment is again indicated. The Fladen Ground spur and outer fringes of the Shetland Platform also became covered with 100m or more of Unit C sediment. Most critically for the regional correlation, these Unit C sediments are rich in volcanic tuffs correlated with the main phase of Hebridean volcanism. At the end of the deposition of these sediments there was a change in sedimentation from sand to shale, reflecting a relative transgression over the Shetland Platform shelf. Foundering of the Shetland

Platform appears to have occurred at the end of the Palaeocene contemporary with the staunching of Hebridean volcanism and the initiation of Atlantic spreading. This cut off the major sediment supply into the East Shetland Basin and allowed the formation of the ash-rich Balder formation.

There is ongoing debate as to the degree to which this episode of Palaeocene subsidence and sedimentation was accompanied by extensional faulting. While Joy (1992) could find no evidence for renewed extensional tectonism, Stewart (1987) and Milton et al. (1990) mark a number of faults "implicated" in this episode. Along the western Viking Graben, distinct increases in thickness of Palaeocene units across the eastern marginal faults of the Shetland Platform are seen on seismic reflection sections, although displacement on these faults does not appear to pass into the sedimentary cover. In the Balder Field on the Utsira High numerous faults are seen to cut the top Palaeocene reflector, many of them soling out within the Palaeocene section, but some of which are basement related. Dominant trends are N-S and NW-SE (Jensen et al., 1993). There is little doubt that there was active tectonism and extension accompanying this episode

The pattern of subsidence along the western Viking Graben suggests that, by the time of the first eruptions (at around 60Ma), the locus of extension had switched to the north and also changed orientation. The new extension was parallel with the dyke swarms formed in association with the Hebridean volcanism and appears to have been a northeastern continuation of rifting. By the time of Unit C sedimentation, the uplift had gathered pace, and erosion had exposed metamorphic rocks beneath the original Mesozoic cover.

On the eastern side of the northern North Sea, following the early Tertiary regression, there was no fan unit sedimentation equivalent to the thick Unit B and C deposits associated with subsidence of the western Viking Graben margin. This suggests that there was no significant uplift of western Norway at this time, uplift being restricted to the region surrounding the Hebridean volcanic province. However, a late Palaeocene-early Eocene deltaic marginal marine unit is present on both sides of the palaeobasin, reaching thicknesses of more than 250m to the east (before being truncated by the late Tertiary uplift unconformity; Lilleng, 1980). This suggests a new sediment source in a mildly emergent western Norway.

3.1.3 *West Shetland Platform*

To the northwest of the Shetlands, there is evidence for Palaeocene transtensional faulting on the platform margin accompanying rifting in the Faeroes Basin. On the margin between 60°N 3°15'W and 60°30'N 2°30'W, there runs the 90km NNE-SSW Shetland Spine Fault system, which underwent early-mid-Palaeocene movement creating the Scarvister sub-basin containing up to 600m of clastic sediment (Hitchen and Ritchie, 1987). The configuration and narrowness of the basin suggest a strike-slip component of displacement. Other, more irregular, patterns of faulting can also be found to the southwest between 59°'N 4°50'W and 60°10'N

3°40'W, again trending generally NE-SW, passing along-strike into the Outer Isles Fault system (Hitchen and Ritchie, 1987).

In the adjacent Faeroes Basin there was a marked divergence in subsidence with a sudden acceleration in the basin centre during Thanetian times (60.5-56.5Ma) while subsidence on the basin flanks slowed dramatically. In the early Palaeocene a 125-200m thick sand unit was deposited in the eastern part of the Solan Basin and flanks of the Rona Ridge, transported by mass-flows (possibly earthquake-induced?) from the West Shetland Platform (Booth et al., 1993). This was followed by a late Palaeocene unconformity, with submarine sands accumulating in the western Solan Basin and Faeroes Basin, succeeded by Upper Thanetian marine shales (Booth et al., 1993). The maximum thickness of early Palaeocene sediments (66Ma-55Ma) in the Faeroes-Shetland Basin is 3600m (Mitchell et al., 1993). Flank uplift of around 100m is indicated with a marked Danian unconformity across much of the area. This unconformity coincides with the accelerated subsidence in this and other nearby basins including the northern North Sea and Porcupine Basin (White et al., 1992, White and Latin, 1992, Joy, 1993). Uplift and subsidence coincides with the oldest ca. 60Ma lavas found in the lower series of the Faeroes (Waagstein, 1988), and the oldest Faroes-Shetland Basin sill complexes (Hitchen and Ritchie, 1987). Water depths reached a minimum during late Thanetian times, probably correlated with the coals found between the lower and middle-upper lavas on the Faeroes. Towards the end of the Thanetian, between 57 and 55Ma, subsidence rates dramatically increased and igneous activity resumed creating the entire thickness of upper/middle series on the Faeroes, the bulk of the Faeroes-Shetland sill complex and the Balder tuffs.

The northeastern and southwestern boundaries of the Faeroes-Shetland Basin, are the WNW-ESE trending Erlend Transfer Zone and the Judd Fault respectively. In the Late Danian there was an unconformity, combined with localised inversion in the southwest of the basin. Transpressional reactivation of many of the major ENE trending (previously normal) fault zones in the Solan Basin caused intense localised basin-margin deformation, with uplift of at least 500m. The most spectacular 'palm-tree' transpressional deformation is seen where the Solan Fault steps about 4km to the northwest, indicating that the transpression was dextral (Booth et al., 1993).

The north-eastward continuation of Palaeocene rifting in the Faeroes-Shetland Basin seems likely to have become the new North Atlantic spreading ridge along the Møre Basin margin, but further to the northeast beyond the Jan Mayen Fracture Zone can be traced through the Vøring Basin. In the outer Vøring basin there is a broad region of Palaeocene extension which resulted in the accumulation of up to 2000m of Palaeocene sediments in a relatively isolated hollow to the northeast of the basin, within the Nyk Syncline. The overall Palaeocene depocentre trends ENE-WSW and has more typical sediment thicknesses of around 1000m. The western Vøring Basin experienced regional uplift in the late Palaeocene or early Eocene, accompanying the initiation of spreading and leading to more than 2400m of subsequent erosion (Skogsheid et al., 1993). As with the Faeroes-Shetland Basin Palaeocene dextral strike-slip faulting is seen along the eastern margin of the Vøring Basin, in particular along the NNE-SSW trending Nordland Ridge Shear Zone separating the basin from the Trondelag Platform.

This pattern of tectonics can be continued up to the SW Barents Sea margin. Following a hiatus around the Cretaceous-Tertiary boundary, a widespread sheet deposit crossed the entire western Barents Sea in late Palaeocene times. Towards the end of the Palaeocene the Loppa High became a source area for sediments passing into the Tromsø Basin and local faulting took place along the western flank of the Senja Ridge, apparently reflecting rifting, and possibly strike-slip faulting, prior to the opening of the North Atlantic.

This episode of rifting and dextral transcurrent faulting along the continental margin basins of NW Europe, appears to have connected with the North Atlantic spreading ridge both through the Hebridean volcanic province and also further to the west. On the Rockall Platform Tertiary sediments are generally restricted to the western margin, where a number of restricted basins contain large thickness of locally derived clastic sediments of Danian age (Hitchen and Ritchie, 1987, Mudge and Rashid, 1987), probably associated with strike-slip movements.

3.2 Africa-Eurasia Collision Zone

Around the Cretaceous-Tertiary boundary until Anomaly 24, the relative convergence between Eurasia and Africa became slowed and erratic with a pole of relative rotation at 33.04°N 14.59°W (Dewey et al., 1989). This boundary continued to pass through the Pyrenees where, during Palaeocene times, compressional deformation finally exceeded the original Lower Cretaceous extension (Puigdefabregas et al., 1991).

3.2.1 Alpine Deformation

In the mid-Palaeocene, deformation related to the Alpine collision zone once again affected a broad area of central Europe (see Figure 3.1), apparently as a result of mechanical coupling of the foreland and A-subduction zones of the central and eastern Alps and Carpathian nappe system (Ziegler, 1987a). However, it is also possible that this deformation once again reflects a sub-plate boundary or boundaries interconnecting the Alpine and proto-North Atlantic plate boundaries. Inversion was intense in eastern Europe along the Polish Trough, and also, according to Ziegler, affected basins inverted in the late Cretaceous sub-Hercynian phase, including: the West Netherlands and Broad Fourteens Basins, the Dutch and Danish parts of the Central North Sea Basin and the Fennoscandia Border Zone. The pattern of transpression was generally located further to the south than that found in the late Cretaceous.

In the Danish Central Graben minor post-Danian deformation along the Arne Elin trend was claimed by Ziegler (1982) to be of late Palaoecene age. A renewed phase of Palaeocene inversion can also be found affecting the Fennoscandian Border Zone around the Kattegat (Liboriussen et al., 1987). There was however no inversion around the Egersund Basin in the Palaeocene, all transpression being restricted to the south of the Viking Graben.

In particular around the coast of Holland, the Broad Fourteens and adjacent basins were inverted once again in this mid-Palaeocene Laramide phase. Small-scale thrust faults developed along the margin of the NW-SE trending Broad Fourteens and west Netherlands Basins while, in the axial parts of the basins, fault geometries suggest transpressive motions (van Wijhe, 1987). Inversion was significant in this basin, although the relative magnitude of uplift in comparison with that in the late Santonian to Campanian 'Neo-Hercynian' phase, is difficult to gauge because of the accompanying marine regression that led to deep erosion across the anticlinal inversion structure. In the N-S oriented Dutch Central Graben, to the north of the Netherlands, inversion was relatively mild and tapered out to both north and south. However, in the southeastern parts of the central Netherlands Basin and the central parts of the west Netherlands and Broad Fourteens Basins, inversion was intense.

3.3 Summary

The overall configuration of active tectonics at this period heralds the creation of the new North Atlantic plate boundary at the beginning of the Eocene. Rifting and transtension dominates the western seaboard, with a short-lived section of spreading ridge passing through NW Britain. Extension involved the movement of Greenland to the north and north-west relative to Western Europe. However across central Europe, as far north as the southern boundary of the Baltic shield, there was transpression at this period, reflecting the north westerly movement of central Europe relative to Fennoscandian.

3.4 Constraints on Palaeocene Deformation in the Baltic Shield

During much of the Palaeocene transpressional deformation along the Fennoscandian Border Zone shared a number of features in common with that found through the late Cretaceous. As in the late Cretaceous the pattern of deformation indicates the concentration of strain along the margins of the shield.

Along what is now the mid-Norway continental shelf and also in the northern North Sea there was some pronounced subsidence, with evidence for some rifting during the Palaocene, with evidence for strike-slip displacements on the Nordland Ridge Shear Zone. It is possible that some component of this deformation passed into the coastal areas of western and northern Norway. Although no sediments reflecting such subsidence have survived, these would anyway have been removed by erosion long-ago, following uplift of this margin through the Cenozoic. This episode marks the youngest phase of significant extensional tectonics through these marginal Mesozoic basins.

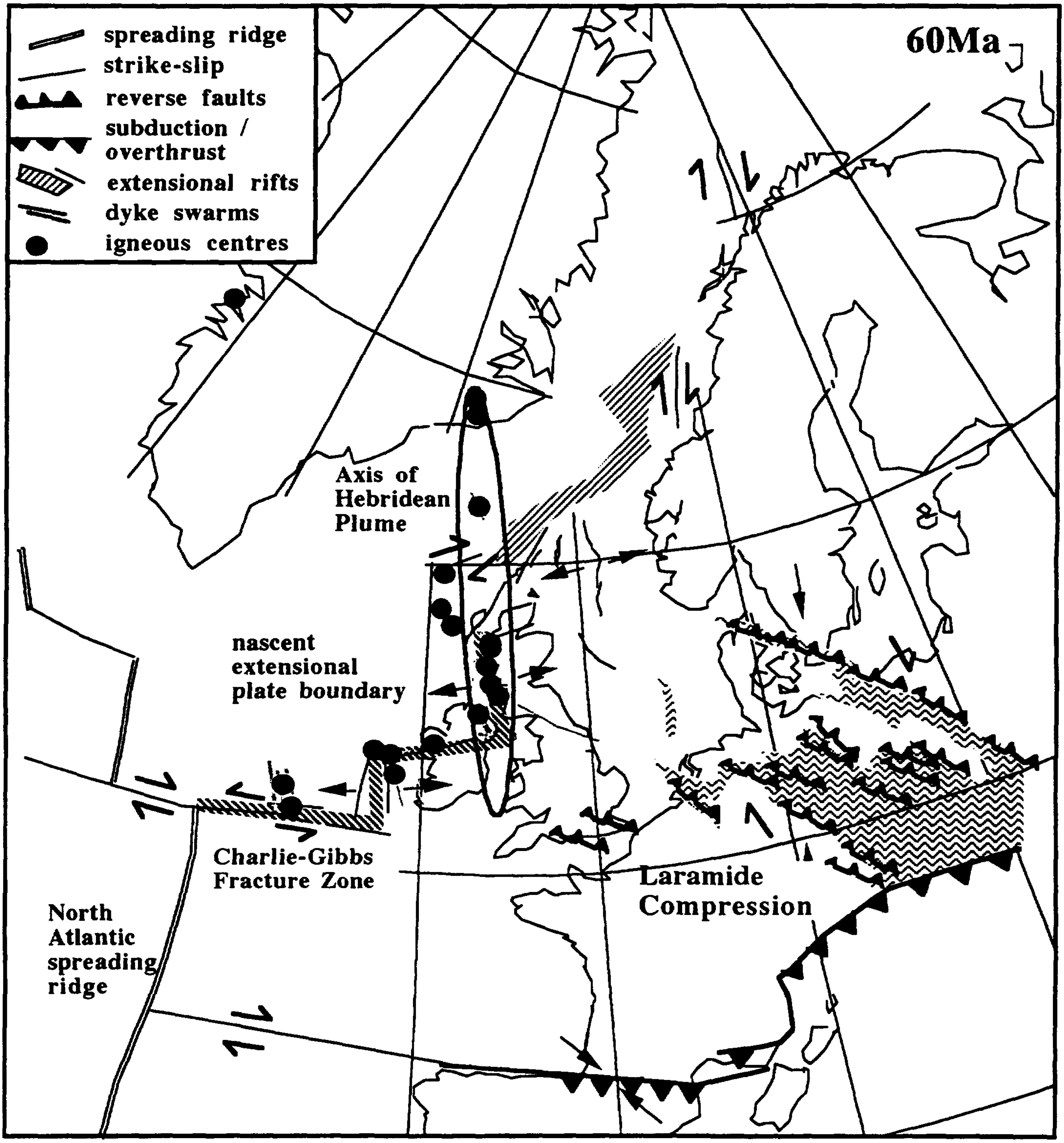


Figure 3.1 Palaeocene tectonics around the North Atlantic

4.0 EARLY EOCENE

4.1 Plate Tectonic Context

Srivastava et al. (1990) proposed an E-W trending strike-slip plate boundary passing through the axis of the Bay of Biscay prior to 44Ma. At Anomaly 21, Dewey et al. (1989) give an Africa-Eurasia rotational pole at 35.08°N 15.54°W, little changed from that at Anomaly 24.

The Pyrenees continued to undergo some compressional deformation although there is little evidence for active tectonics in the forelands of the Pyrenees, or the Alps, in this period. Hence, all the tectonic activity to have affected north-west Europe accompanied the northward expansion of the North Atlantic spreading ridge. Having failed in its attempt to drive a path through Ireland, and northern Scotland at Anomaly 24N times, the spreading ridge found a new path from the southern end of the Labrador Sea spreading ridge, diagonally between the Rockall Bank and Greenland, to link with the western edge of the pre-existing Mesozoic rifts located between Greenland and Norway.

4.1.1 Thulean Magmatism

A great outburst of basaltic volcanic activity characterised the opening of the North Atlantic spreading ridge. Continental rifting began with an extraordinary phase of plateau basalt formation and dyke intrusion over the formerly contiguous regions of east Greenland, the Faeroes and the West Shetland Basin, see Figure 4.1. Up to 15km thickness of new igneous crust was added to the rifted margins (see Figure 4.2). Dipping reflectors, characteristic of repeated sub-aerial lava flows, can be traced all around the margins of the northern Atlantic, terminating sharply to the south, to the west of Rockall at 57°N. Further to the south, there was no major addition of igneous melt to the lower crust during rifting (White, 1988).

In marked contrast to the NE-SW dilation of the Hebridean volcanic outburst, which had ended only three or four million years previously, this volcanic activity accompanied NW-SE dilation. This episode of volcanism produced enormous outpourings of magma very suddenly, about a sub-aerial volcanic axis (Mutter, 1984) (see Figure 4.1). After a few million years, the rate of crustal accretion decreased and the original spreading axis subsided below sea-level. At first the new spreading ridge appears to have taken a more easterly path.

4.1.2 Faeroes-Shetland Spreading Ridge

Between the Faeroes and the Shetland Platform runs the westerly-facing Faeroes-Shetland escarpment, the buried feather-edge of a thick pile of early Eocene flood basalts, overlying a thinner, but more widespread, layer of late Palaeocene basalts. Water depths reached a minimum during late Thanetian times, probably correlated with the coals found between the lower and middle-upper lavas on the Faeroes. Towards the end of the

Thanetian, subsidence rates and igneous activity resumed. That phase between 57 and 55Ma was by far the most volumetrically important in the basin's history creating the entire thickness of upper/middle series on the Faeroes, the bulk of the Faeroes-Shetland sill complex and the Balder tuffs. On the edge of the Faeroes-Shetland Basin, the Erlend volcanic centre (about 100km NNE of Shetland) was a major basic igneous complex in the late Palaeocene (Gatliff et al., 1984). A zone of basic intrusions (each 2-100m thick) oriented NE-SW, runs at least 250km from 61°N, affecting Palaeocene and Cretaceous sediments. A similar linear intrusive complex is presumed to run through the Axial Opaque Zone of the Faeroes-Shetland Basin apparently associated with premature sea floor spreading and extension between the Faeroes and Scotland. The Magnus Transfer Zone and the Erlend Transfer Zone (that together define the Erlend volcanic centre) controlled the north-eastern end of the Faeroes-Shetland Basin. The ensuing subsidence of the Erlend complex, beneath early Eocene sediments, suggests that there was crustal thinning in this region.

4.1.3 *Faeroes*

The transition between the Hebridean and Thulean volcanic episodes can be traced in the Faeroes. The Lower and Upper Basalt series on the Faeroes are dated, by correlation along reflection lines to the North Sea and West Shetland wells, as late Palaeocene and early Eocene respectively. The Faeroes coal sequence, between the Upper and Lower Basalt Series, has been dated as latest Palaeocene (Smythe, 1983). The first phase of basaltic volcanism was initiated in the late Danian-early Thanetian (Anderson, 1988), approximately contemporary with the Hebridean activity and was fed from fissures trending NNW-SSE to NW-SE. Beneath the lavas it is now believed lie marine sediments. Lavas have subsided more than 3000m in places. In contrast to the Hebrides magmatism was continued in parallel with the Thulean outburst that was initiated in the early Eocene in Greenland.

It is likely that the gap between the upper and lower series corresponds to the initiation of spreading along a WSW-ENE trending axis to the north of the islands. Volcanism continued for 4-7 Myr, in the Thanetian, with the highest lavas distributed adjacent to the emerging ridge. During this first phase of spreading, the Faeroes plateau was cut by hundreds of dykes, most of them clustered into a belt running WSW-ENE across the northern group of islands. The Faeroes, with 3000m of exposed lavas, were located immediately to the south of the Jan Mayen Plateau but the axis of spreading, offset by a NW-SE transform fault passing along the south coast of the Blosseville Peninsula, began to separate the Faeroes from Greenland.

4.1.4 *East Greenland*

In eastern Greenland, the oldest fossiliferous sediments, intercalated with the 9km thick pile of tholeiitic basaltic lavas of the Blosseville Group, contain marine shales with Sparnacian dinoflagellates. Faller and Soper (1979) placed the whole magmatic episode in geomagnetic polarity epoch

24-25R, dated at 59-54Ma. This volcanism was associated with the formation of a major step and subsequent flexure, whereby the continental crust became attenuated, and high crustal strain rates were accommodated by dyke intrusions. Along sections of the Greenland coast the dykes account for more than 50% of the outcrop. The moncline runs for some 600km between 66° and 70°N and, along the Blosseville coast, affects the plateau basalt lava pile. Dykes were intruded, both before and during the flexure, almost all of them striking NE-SW. Similar flood basalts must have formed much of the upper crust of the submerged Jan Mayen Plateau, which was located adjacent to the Blosseville Peninsula through to Anomaly 20 time (about 44Ma).

The earliest oceanic magnetic anomaly which can be detected in the regions to the north and south is the double Anomaly 24A and 24B (Hagevang et al., 1983), dated at 54Ma, which must mark the submergence of the spreading axis. Following the tholeiitic outburst, the Greenland trailing edge was the site of continued normal faulting, with the emplacement of differentiated dykes of both alkaline and tholeiitic type (Nielsen, 1978), of mixed magnetic polarity, suggesting intrusion over an extended period.

4.1.5 Norwegian Continental Margin

To the northwest of the mid-Norway Basin, offset 150km from the Faeroes-Shetland escarpment by the Jan Mayen Fracture Zone, great outpourings of basaltic lavas developed along the Vøring Plateau Escarpment (Bukovics and Ziegler, 1985). The intensity of igneous activity increases towards the west and includes numerous dykes and sills intruded into earlier sediments. An alkali basaltic volcano at Vest Brona, offshore close to Kristiansund (63°15'N, 07°E) and located on a southern prolongation of the Jan Mayen Fracture Zone, provided a date of 55.7±0.9Ma (Bugge et al., *1980*), at the same period as the lower lava series in the Faeroes. A major tholeiitic basaltic volcano was active at the same period in the Skaggerak (see Figure 4.2) (Norin, 1940; Åm 1973), apparently exploiting a major lithospheric fracture system. The parallelism of fault trends in the Skaggerak with the fissure magmatism in the Faeroes-Shetland Basin suggests that dilation occurred over a wide region around the North Atlantic. The magmatism in the Skaggerak implies significant mantle melting, at considerable distances from the main axis of spreading, and implies that the mantle plume extended under western Scandinavia.

4.1.6 Barents Sea

The Barents Sea is bounded by the Lofoten Basin, floored by Cenozoic ocean crust, that from the evidence of accompanying linear magnetic anomalies becomes younger northwards along the margin. To the north the new plate boundary connected with the Arctic Ocean spreading ridge through a lengthy and complex transform fault system, part of the De Geer Zone (Faleide et al., 1991). Between anomaly 25/24 and 13 the direction of early opening was close to the orientation of the Senja Fracture Zone. Further to the north the Hornsund Fault Zone was in transpression as is manifest in western Svalbard.

At the southern end of this system, along the Senja Fracture Zone, the continent-ocean transition is generally narrow and distinct (Faleide et al., 1991) and the oceanic crust can be followed to within 10km of a prominent continental boundary fault. East of the fault a steeply dipping reflector can be traced down to 20km correlated with the base of the crust (Faleide et al., 1991).

At the northern end of the 250km long Senja Fracture Zone, lies the Vestbakken volcanic province, that masks the oceanic continental boundary. This marks a step in the transform, associated with a short section of intervening spreading ridge. By analogy with the Vøring Marginal High it is considered to reflect sub-aerial flow basalts extruded during break-up in earliest Eocene times. The landward margin of the province is a listric fault complex formed in the early Tertiary. Locally it is possible to map flows that reveal major growth along these faults through the Eocene and Oligocene (Faleide et al., 1988). The greatest magnitude of downfaulting of this pull-apart basin exceeds 2km along the northern rifted segments. As a result, the wider northern part was formed as a pull-apart basin at a releasing bend in the margin, causing extensional faulting and deposition of a relatively thick Palaeogene succession. Immediately to the south of the Vestbakken volcanic province the narrow southern part of the Sørvestsnaget Basin was uplifted and eroded. West of the Senja Fracture Zone there was uplift of the continental crust associated with transform displacement. The main deformation took place in the Eocene although continued through to the Oligocene leading to about 1km of Palaeogene sediments being eroded from an elevated outer marginal high. Controlling faults in the northern basin are sub-parallel to the rifted and sheared margin segments oriented NE-SW and NW-SE.

The Stappen High that surrounds Bjornøya, forms part of a north-south trending zone of elevation that was uplifted and eroded during the Tertiary. An estimated 3.5km of uplift at Bjornoya decreases eastward to 1-2km (Wood et al., 1990). This uplift was probably initiated with early Eocene sea-floor spreading.

4.1.7 North Sea

Sub-aerial volcanism all around the new spreading ridge culminated at 54Ma, contemporary with the climax of volcanic tuff accumulations in the North Sea Basin Balder Formation (see Jacque and Thouvenin, 1975). The main Balder tuff zone decreases in both the number of internal layers and their individual thicknesses in passing to the south, with around 500 separate layers at well 30/2-1, in the north of the Viking Graben (Malm et al., 1984), and 333 at well 25/10-1, 200km to the south, while in Denmark, despite a local Skaggerak source there are only 122 layers.

Throughout the central and northern North Sea, extending into Denmark, there is a marked disconformity above the ash series, with erosion and sub-aerial weathering providing a red shale. The northern North Sea appears to have suffered less emergence than the central North Sea. However, at the Frigg Field (Heritier et al., 1979; 1981), the uppermost ash layers are found to be rich in a continental microflora. Sedimentation, largely with continental derivation, continued through at least some of the dinoflagellate Zones missing from the central North Sea. At the period of the ensuing central North Sea transgression, the Frigg Formation was deposited (Lilleng, 1980): massive coarse grained carbonaceous and micaceous sands form a large mound up to 250m thick emerging basinward from the earlier sandy units developed on the western side of the northern North Sea Basin. The formation infills the deepest part of the Balder surface, centred at around 59°30'N 2°E, and its localisation and NNW-SSE axis suggests potential tectonic control. Following the Frigg Formation, the entire northern North Sea region was covered by muds of the Hordaland Group.

For about ten million years in the early Eocene, northwestern Europe appears to have approximated to a passive margin, relatively free from deformation away from the newly created North Atlantic spreading ridge. Apart from passive uplift as in the central North Sea, there is little evidence for any deformation during this period. The chief exception to this may be along the landward continuation of the newly created Iceland-Faeroes Fracture Zone where an intense development of foreland thrust folds developed to the south of the Ymir Ridge in the early Eocene, at the initiation of spreading (Booth et al., 1993).

In the Solan Basin, at the southern end of the Faeroes - Shetland Basin the Balder formation is succeeded by a thin unit of continental to paralic coal-bearing strata. Subsidence slowed in the late Palaeocene - late Eocene but rates increased in the late Eocene. Between the Lower Eocene through to Lower Oligocene there was a transition from sedimentation accompanying rifting to that associated with thermal subsidence.

4.2 Constraints on Early Eocene Deformation in the Baltic Shield

The initiation of seafloor spreading through the Norwegian Sea appears to have concentrated all the deformation that had persisted through the rifted basins around the margins of the Baltic Shield. There is little evidence of ongoing inversion of the Fennoscandian Border Zone subsequent to the creation of this new plate boundary. Any minor deformation that may have persisted in the Baltic Shield as a consequence of this marginal deformation through the late Cretaceous and Palaeocene is likely to have all but disappeared at this time. An episode of regional uplift of the flanks of the new spreading ridge, known to have accompanied the plume volcanic activity of the North Atlantic, is likely to have included the western edge of the Baltic Shield.

The most realistic present-day analogue of the impact of the North Atlantic continental splitting on tectonic deformation internal to the Baltic Shield is probably to be found around the Red Sea spreading ridge, where continental separation was initiated about 15 million years ago. While there is some active deformation around the raised edges of the Red Sea rift (as in coastal Yemen and Saudi Arabia) inland within the Arabian shield, 300-500km from the spreading ridge, there is effectively no seismicity nor active deformation. It is reasonable to assume that at a period when regional plate tectonic strain is concentrated on the spreading ridge, a comparable situation applied in the Baltic Shield.

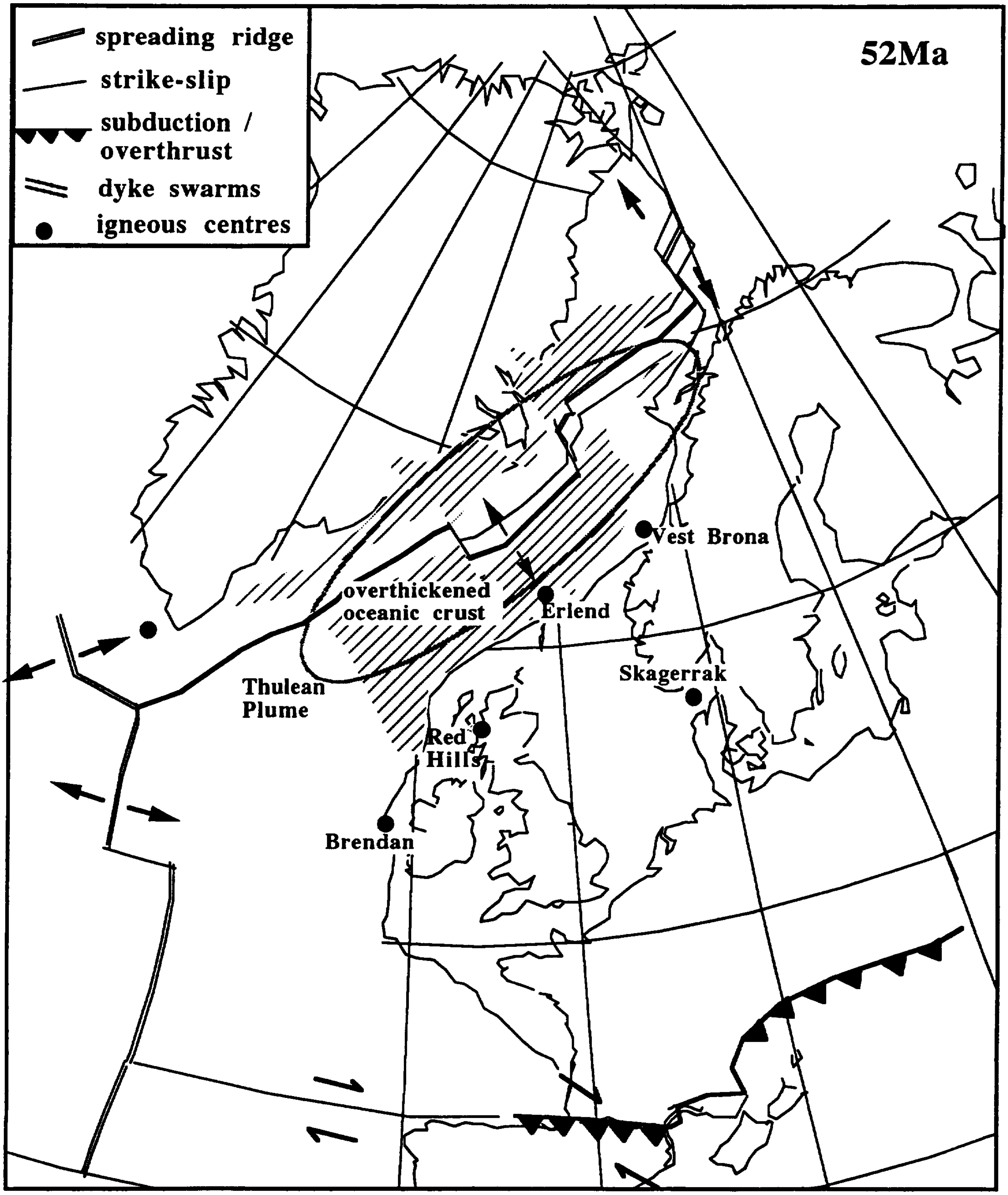


Figure 4.1 Early Eocene tectonics around the North Atlantic

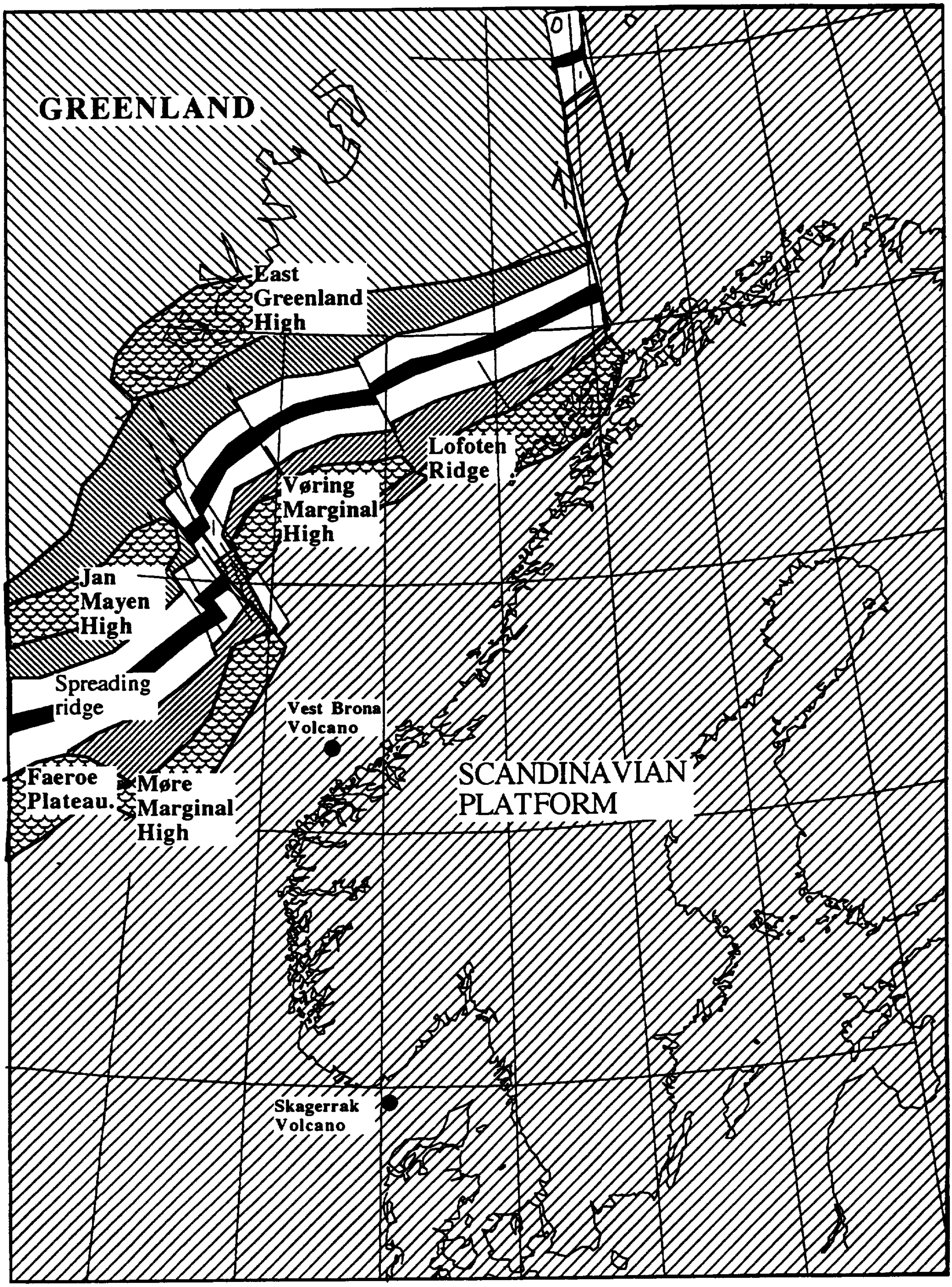


Figure 4.2 Tectonics around Fennoscandia at 50 Ma

5.0 LATE PALAEOGENE

The period from the mid-Eocene to the end of the Oligocene saw a major change in the orientation of spreading in the Norwegian Sea that accompanied a complex and interrelated series of sub-plate boundaries passing through northwest Europe. The deformation within this region was extremely varied in tectonic style and the overall configuration of tectonics was complex, with compressional and extensional styles co-existing. While along the Norwegian continental margin the tectonic configuration of the late Eocene was closely comparable to that of the Oligocene, to the south-west of Fennoscandia there were marked variations through this period, and one can discern distinct mid-Eocene, Eocene-Oligocene boundary, and late-Oligocene tectonic episodes.

5.1 Plate Tectonic Context

Around Anomaly 20 (44Ma, mid-Eocene), spreading in the Labrador Sea slowed and by Anomaly 13 (35Ma, earliest Oligocene boundary) had ceased (Kristofferson and Talwani, 1977) when Greenland became welded to the North American plate. Consequently, through this period the opening vector for the Norwegian Sea had to become consistent with the pole of relative rotation between the North American and Eurasian plates (see Figure 5.1). This change in relative plate motions had an important impact on both the plate boundary and the neighbouring continental margins (see below).

While the direction and velocity of extension at the spreading axis could adjust instantaneously to any change in relative plate motions, a problem arose in the re-orientation of transform faults. The anticlockwise rotation in opening vectors (to nearly E-W) in the northern Atlantic placed pre-existing left-stepping transforms in partial extension: the Senja and Hornsund Fracture Zones becoming the Knipovich spreading ridge separating Greenland from Spitsbergen. However, where pre-existing transforms were right-stepping, they became the site of oblique compression.

By far the most important right-stepping transform in the northern Atlantic lay along the modern Iceland-Faeroes ridge, which offset the Reykjanes spreading ridge to the south from the Aegir spreading ridge to the north by about 150km. The oblique compression at this transform was difficult to resolve through local overthrusting, as the oceanic crust to either side of the transform was equally young, hot, strong and buoyant. As the spreading in the new orientation persisted, several tens of kilometres of oblique crustal shortening became concentrated at this transform and had to become absorbed in the surrounding region (Nunns, 1983).

The compression at the transform locked its motion as a pure strike-slip fault. This prevented symmetrical spreading along the Aegir Ridge which, instead, began to spread in a fanshaped pattern, with the relative motion vector located at the locked transform. As this spreading no longer accorded with the overall Greenland-Eurasia plate motion, a new spreading ridge (the Kolbeinsey Ridge) had to develop to the north of the Reykjanes spreading ridge, offset by a left-stepping transform. This spreading was adjacent to the Greenland continent and, at first, seems to have been sub-aerial, or distributed, producing no recognisable magnetic anomalies (Nunns, 1983). Accompanying this new spreading axis alkaline plutonic activity was resumed along the Greenland coast at around 35Ma, twenty million years after the commencement of the original basaltic volcanism (Rex et al., 1979). Between the old and the new spreading ridges, the original marginal crust of Greenland became separated to form the Jan Mayen microplate, which underwent a 29° anticlockwise rotation and suffered internal rifting, as fanshaped spreading occurred to either side. The Jan Mayen microplate was in existence until Anomaly 7 (25Ma), after which all spreading ceased along the Aegir spreading ridge to the east.

Ideally, between Anomalies 20 and 7 (44 to 25Ma), the cumulative spreading of the Aegir and Kolbeinsey ridges should have been equivalent to the anticipated motion between Greenland and Eurasia. However this assumes that no internal deformation passed into the plates on either side. Deformation is however known to have radiated to the south of the blocked transform, where a new series of fracture zones appeared along the Reykjanes spreading ridge, after Anomaly 20 (Vogt and Avery, 1974), as far south as the original triple junction, (Voppel and Rudloff, 1980). These fracture zones acted as transform faults to short (50-100km) segments of spreading axes, themselves oriented almost N-S as compared with the earlier and present-day NE-SW linear anomalies. Several of these fracture zones show a kink in their orientation, varying from WNW to more nearly E-W around Anomaly 13, reflecting the most profound change in spreading direction. These fracture zones, therefore, record the re-orientation of spreading directions. Nunns (1982) has suggested that these fracture zones in the Reykjanes Ridge formed as a response to stresses coupled through from the locked NW-SE transform fault along the Scotland-Greenland ridge.

5.1.1 East Greenland

As a new rift developed there was pronounced faulting along what is now the coast of East Greenland. The northwestern end of the Jan Mayen Fracture Zone passes up the Kejser Franz Josef's Fjord, to form the northern boundary to a band of post-Eocene reverse faults and folds only found along the edge of the Greenland landmass between latitudes 71°30' and 72°40' (Haller, 1970). One reverse fault, displaces a dolerite sill by 350m. The NNW trending thrust block is only 5km wide and is cut off to the south by a transverse fault separating Devonian from Carboniferous sediments, with a post-Devonian throw of up to 3,000m, some of it of Tertiary age. A zone of shallow post-basaltic folding, apparently overlying Upper Permian evaporites, is also found in northeastern Jameson Land. The configuration of inversions implies that the 'fossil' NNW-SSE Jan Mayen Fracture Zone was active as a dextral transcurrent fault, on both sides of the Atlantic.

To the south, on the northern part of the Blosseville Coast there are the only two localities of mid-Tertiary age sediments known from the eastern coast of Greenland, both probably in basins formed as step-overs on major coast-parallel strike-slip faults. At Kap Dalton, there is a NE-SW trending downfaulted block about 5km wide, bounded on both sides by major faults with crush zones up to 500m wide. Within this graben, the downfaulting exceeds 600m although, when traced further to the northeast, the vertical component of displacement along the coast-parallel fault zone is relatively minor (Watt et al., 1976). The Kap Dalton Palaeogene sediments are located adjacent to a major N-S fault, at a left-stepping offset in the controlling NE-SW faults, and comprise Lower Eocene volcanogenic sediments overlain by 45m of Lower Oligocene sandstones and flags (Soper and Costa, 1976).

At nearby Kap Brewster, Oligocene and Miocene sediments are found in a graben defined by a major NNE-SSW trending fault, with a downthrow of more than 1000m to the east, at an intersection with a prominent NW-SE trending fault (Birkenmajer, 1972). The steeply-dipping synclinal structure formed at the junction of these faults is less than 1km in width and is suggestive of a pull-apart basin, formed at a left-stepping offset in the controlling NNE-SSW trending near-vertical (strike-slip?) fault. Late Eocene to early Oligocene sediments are up to 130m thick and developed in a low-relief topography, while the overlying Miocene sediments developed adjacent to a fault-defined cliff line.

Both these structures indicate that coast-parallel left-lateral strike-slip faulting, along NE-SW to NNE-SSW trending faults, was of significance during the period of re-arrangement of spreading axes in the North Atlantic, just as it was on the other side of the Atlantic through western Britain.

5.1.2 Barents Sea Margin

Along the western margin of the Barents Sea the rotation in spreading direction caused a change from a transform fault to a constructional plate margin. The original oceanic spreading prior to anomaly 20 had created a transform bounded continental margin up against oceanic crust along much of the Senja Fracture Zone (up to 72°N). Between anomaly 20 and anomaly 13, the northern section of the Senja Fracture Zone, along with all the Hornsund Fracture Zone became first an extensional rift that developed into a spreading ridge. During this period renewed faulting and volcanism took place in the Vestbakken volcanic province and faults were reactivated and subsidence continued in the northern Sørvestsnaget Basin, where there is the maximum depth to the base Tertiary found along the SW Barents Sea margin. The creation of the new spreading ridge is likely to have been accompanied by uplift of the adjacent continental margin, a significant component of the cumulative Tertiary uplift demonstrated along the northern edge of the western Barents Sea. The transition from transform to spreading ridge is unlikely to have been simple, and it is possible that rifting was initiated on a range of structures, before becoming concentrated to the west. An example of this may be the volcanism at Vestbakken that did not proceed into a spreading ridge. Reknes and Vagnes (1985) suggested that the Greenland Fracture Zone was generated by volcanism along a leaky transform, and that there was not room for the SE part of the transform at anomaly 13.

Once oceanic crust was being generated all along the margin, sinking following thermal uplift and extension, accompanied by the transfer of large quantities of sediments from the uplifted Barents Sea, caused profound subsidence.

5.2 Africa-Eurasia Plate Boundary

For the Eocene, Savostin et al. (1986) and Klitgord and Schouten (1985) agree on an Africa-Eurasia pole located to the northwest of Portugal. Prior to 25Ma, Iberia was either attached to Africa or maintained very similar relative motions, the boundary between Eurasia and Africa being located to the north. Estimates for the position of the pole of relative Africa-Eurasia rotations around Anomaly 13 are 24.42°N 15°W (Srivastava et al., 1990) and 29.26°N 20.67°W (Dewey et al., 1989). Both are consistent with compression to the east (in the Pyrenees) passing into extension to the west.

5.2.1 Alpine Collision Zone

In the mid-Eocene, around the time when spreading in the Labrador Sea began to slow down, N-S compressional deformation occurred throughout central southern France, becoming more pronounced to the south (Rouire and Rousset, 1980; Gezé, 1979). At this same period in the German Molasse Basin, pronounced transpressional tectonic activity related to dextral strike-slip movements on Hercynian NW-SE faults, separates the Upper Cretaceous from the late Eocene when sedimentation was resumed. Major reverse faults have vertical throws of up to 1500m. The initial phase of deformation can be dated from accompanying detrital sedimentation as Santonian to Campanian (Bachmann et al., 1987). Ziegler (1987a) considers that the second phase is of mid-Palaeocene age but a significant component appears also likely to be mid-Eocene.

In southern Britain, pronounced mid-Eocene reverse displacements along E-W trending faults (many of them blind) are found in both the Weald and Wessex basins and includes structures such as the Isle of Wight Monocline. In SW England dextral NW-SE faulting, also reactivating late-Variscan structures, accompanies pull-apart basin creation and there is more mid-Eocene reverse displacement on E-W trending faults observed to the west in the North Celtic Sea and South West Approaches Basins. Similar deformation is also seen around the southern end of the Rockall Trough associated with the Clare Lineament.

All these structures lie within a broad WNW-ESE Variscan (Lefort, 1973) shear zone, passing from the Alps and the Pyrenees into the North Atlantic spreading ridge, that was subject to a ca. 3 kilometres of dextral displacement at this period. This has the characteristics of a sub-plate boundary separating the oceanic crust of the Bay of Biscay and the adjacent Atlantic margins from western Europe. Such an Eocene plate (the "Porcupine Plate") was proposed by Srivastava and Tapscott (1986) in order to match Eocene magnetic anomalies (in particular 24 to 20) to the south of the Charlie Gibbs Fracture Zone, where there is an increasing overlap, implying clockwise rotation of the Bay of Biscay.

Compressional tectonic deformation has not been traced to the west of the Clare Lineament, suggesting that, as with the Africa-Eurasia plate boundary, the pole of relative rotation lay close to the continental margin. The Clare Lineament itself lies at the eastern end of the Charlie Gibbs Fracture Zone, allowing motion to become transferred to the North Atlantic plate boundary.

5.2.2 *Foreland Extension*

The mid-Eocene transpressional shear zone that affected much of France and southern England was relatively short-lived. From the late Eocene through the Oligocene there was no continent-continent collision impacting into western Europe from the south and throughout this period, the foreland region of central western Europe passed into extension. To the south, in central and southeastern France, this extension was concentrated on a series of deep fault-bounded basins. Similar extension is also found in the northern margin of the eastern and central Alps in Upper Austria (Nachtmann and Wagner, 1987) and through the Molasse Basin of Germany (Bachmann et al., 1987). Further to the north, it was largely concentrated along the single NNE-SSW trending Rhine Graben rift, where up to 1600m of Lower Oligocene sediments were deposited. For the whole late Eocene to early Miocene period (ca.45-20Ma) of Rhine Graben rifting, total extension was 6-8km.

Although the Alpine foreland was in extension, the Alps themselves appear to have been zones of convergence and nappe formation through the Oligocene, with ophiolite emplacement (Dewey et al., 1973). In the eastern Alps, there is evidence for a wedging of overthrusts formed in the Eocene and Oligocene whereby the magnitude of crustal shortening decreased from around 2-300km in Greece to almost nothing in northern Italy (Burchfiel, 1980).

5.3 North European Compressional Province

While central Europe was in extension (reflecting the influence of subduction 'suction' to the south) a zone of compression began to develop to the north (Figure 5.3) reflecting the influence of the blocked Iceland-Faeroes Fracture Zone on the North Atlantic plate boundary. These compressional structures appear typically as blind thrusts reactivating and partly inverting pre-existing rift-bounding normal faults. The most southerly trend of such inversion structures runs in a WNW-ESE band across central Holland, where uplift of around 200m can be fairly precisely dated as early Oligocene (Keizer and Letsch, 1963). Renewed uplift of the Broad Fourteens and West Netherlands Basins also occurred during the late Eocene, around the transition to the Oligocene, although this was relatively mild in comparison to the late Cretaceous (sub-Hercynian) and Palaeocene (Laramide) phases. There is no evidence for contemporaneous deformation in the Dutch Central Graben to the north (van Wijhe, 1987).

5.3.1 North Sea

Inversion structures of the same general orientation can, however, be followed towards the northwest, where the Sole Pit Basin, a transtensional structure developed along a NW-SE strike-slip fault complex, was inverted for the second time (having already been partly uplifted to the southeast in the Upper Cretaceous). The main anticlinal axis has a cumulative 1500m of inversion (Van Hoorn, 1987a) and lies about 20km to the northeast of the major NW-SE Dowsing Fault Zone, separated from the Viking Monocline another inversion structure by the Collapse Zone, attributed by Walker and Cooper (1987) to decollement on evaporite horizons, associated with the inversion to the southwest. Where the controlling structures bend to the west into the Cleveland Hills, on the coast of Yorkshire, there has been up to 2500m of uplift (Hemingway and Riddler, 1982). The presence of Eocene sediments, preserved in collapse lows, indicates that the major phase of inversion occurred in the Oligocene and possibly in the Miocene. Uniform subsidence returned to the area, following the truncation of the fold structures during the Pliocene.

In the region to the northeast of the Sole Pit Basin, a number of NNW-SSE to N-S oriented North Sea anticlines (overlying blind reverse faults) formed at this same early Oligocene period. Several of these structures in the northern North Sea (with uplifts of around 200m) formed hydrocarbon reservoirs, as at Sleipner Gamma (Pegrum and Ljones, 1984), see Figure 20, and at Troll, where the inversion partly collapsed in the Miocene (Muir Wood, 1987a). Top Chalk anticlines, trending NNW-SSE, are also mapped in the Norwegian blocks 15/5 and 15/6 to the north of Sleipner (Fagerland, 1983). Further to the north, on the border between UK Blocks 2 and 3, along the East Shetland Platform boundary fault, transpressional faulting along a NNW-SSE trending structure, defines the Emerald Field (Wheatley et al., 1987) close to where renewed faulting in the Oligocene caused uplift resulting in extensive sand deposition.

Another line of such structures is seen to the southwest of Norway. In the Danish Central Trough, to the southeast of the main Central Graben, along the NNW-SSE trending Arne-Elin Trend in the Danish Central Trough, to the southeast of the main Central Graben, across a band almost 10km wide, the early Oligocene sediments are about 100m thinner than in the adjacent region (Clausen and Kortsgoard, 1993), while to the north, in the Tail End Graben, there is a small NNE-SSW trending early Oligocene 'basin' which is about 100m thicker than the surrounding region. This may represent a transtensional stepover partly defined by a NNE trending segment of the

Coffee Soil Fault. Another structure also subject to inversion at this period is the NNW-SSE Rynkobing Fyn High located to the north of the Arne-Elin Trend. Closer to Norway in the Stord Basin, two prominent compressional structures with vertical displacements of around 100-200m are seen both within the basin and towards its eastern margin (Biddle and Rudolph, 1988), that by analogy with dated structures to the south and west are probably of the same age.

Closer to Scotland the Forties Field gained some of its structural closure at this period, the structure persisting elevated through to the mid-Miocene. In the nearby Nelson Field, NNW-SSE trending periclines were active in the late Eocene (Whyatt et al., 1992). There was also no late Eocene sedimentation in the Witch Ground Graben of the Outer Moray Firth within which mild compressional deformation is indicated on sections across the basin (Boote and Gustav, 1987).

Sinistral deformation along the Great Glen Fault, possibly related to this phase, can be evidenced within the inner Moray Firth Basin, although Palaeocene uplift makes it hard to date. An original rifted basin associated with extension along NE-SW trending normal faults (principally the Helmsdale Fault), during the Jurassic, became partially inverted in the Tertiary. Inversion, as a result of sinistral displacement along the Great Glen Fault (Bird et al., 1987) can be seen strongly affecting the fault-controlled margins of the Inner Moray Firth, with the culmination of uplift in the north-west corner of the basin where the Great Glen Fault passes into the E-W trending basin bounding fault. A cumulative Tertiary uplift of 900m was estimated by Bird et al. (1987) from over-compaction of Beatrice Field sediments.

In considering the overall configuration of active tectonics in the North Sea at this period a number of generalisations can be made. Although the Sole Pit Basin suffered by far the greatest amount of mid-Tertiary inversion, much of this may have been later in the Oligocene and early Miocene, when it appears that deformation was largely, but not solely, concentrated around the western edge of the North Sea. However, a number of other structures indicate that compressional deformation was most broadly distributed across the North Sea Basin around the Eocene-Oligocene boundary. However, such compression was sustained, if at a lower level, through to the mid-Miocene.

The reverse faults underlying these inversions are typically oriented NNW-SSE across much of the North Sea, although rotating to N-S between the Viking Graben and Norway (see Figure 5.3). It is likely that individual structures are interconnected through some form of strike-slip transfer faulting and the wide separation of these structures suggests that they are linked as confining bends associated with NW-SE to N-S trending strike-slip faults. At least three distinct lines of structures can be traced: one along the western side of the Central Graben; a second along the eastern margins of the Central and Viking Grabens and a third closer to the boundary with Norway from the inner Stord Basin through the Troll field.

It is not possible to trace these structures further to the south-west but it seems possible that similar deformation may have continued into the Fennoscandian Border Zone. However across the whole North Sea, individual structures show no more than about 200m of uplift, the cumulative WSW-ENE crustal shortening across the North Sea Basin amounting to no more than a few hundred metres.

5.3.2 *Iceland Faeroes Fracture Zone*

This distributed province of compressional or transpressional faulting appears to become concentrated to the north-west along the continuation of the Iceland-Faeroes Fracture Zone onto the continental margin. Faulting and folding of the lavas, accompanying profound compressional crustal shortening, is seen in the region around the northern part of the Rockall Trough - Faeroes-Shetland Basin and the Faeroes-Rockall Platform. The most impressive compressional structure is the Wyville Thomson Ridge Complex southwest of the Faeroe Islands, which consists of two anticlines, the Wyville Thomson Ridge and the Ymir Ridge, as well as a small intervening Tertiary basin, which to the ESE is connected to the Rockall Trough. The complex extends as a whole for 150km NW-SE by 50km NE-SW, separating the Faeroes-Shetland Basin from the Rockall Trough. The southern flank of the Ymir Ridge is expressed for a considerable distance as a high and steep escarpment in the surface of volcanic rocks. Many reverse faults are seen on the multichannel seismic data around this southern flank. Whereas Boldreel and Andersen consider that the structure is almost purely compressional on one of their sections thickening of Eocene sequences can be seen to have preceded uplift and inversion, strongly suggestive of strike-slip movements. In places thickness variations are seen in the uppermost part of the volcanic sequence, along the two most prominent anticlines: the NW-SE trending Wyville Thomson Ridge and the Munkegrunnur Ridge, trending NNW-SSE, that heads directly for the southernmost of the Faeroe islands.

The deformation is assumed to have been initiated in the late Palaeocene - Early Eocene in a relatively narrow zone including the Wyville Thomson ridge complex and Munkegrunnur Ridge. In the early Eocene, an intense development of foreland thrust folds developed to the south of the Ymir Ridge, at the initiation of spreading. Relative structural stability ensued in middle to late Eocene times, although some structural movement occurred in the middle Eocene. Renewed compression during the Oligocene was located in the same geographically narrow zone. In the Wyville Thomson ridge complex this can be seen as erosional truncations on the late Oligocene unconformity. Further to the east there was structural inversion of the West Lewis Basin. Reflectors in the lower and intra-mid Miocene sequence decrease in thickness towards the thrust complex although the early Mid-Miocene is missing. The upper mid-Miocene and younger sequences do not appear to be affected.

5.3.3 Faeroes - Holland Sub-plate Boundary

From the mid-Eocene this compressional deformation formed as spreading came to an end in the Labrador Sea, and appears to be an accommodation to the re-orientation of spreading axes in the Norwegian Sea, causing lithospheric shortening in the continent to the east. An early Oligocene zone of active extension can be traced from the Gulf of Lyon to the NNE into the Rhine Graben. At the northern end of the Hessen Depression, it passes into the northwesterly-oriented zone of dextral shear and transpression seen in the Netherlands (see Figure 5.4).

Thus, it appears that the Faeroes-Holland compressional shear zone absorbed crustal shortening to the east of the Atlantic that was previously accommodated as crustal dilation to the west. Dextral motion along this boundary is consistent with the continued extension along the Aegir ridge, and the blocked Iceland-Faeroes transform. Consequently, for periods of the late Eocene - early Oligocene (ca.35 Ma), France, Belgium, England and Ireland, with the adjacent North Atlantic and Rockall Bank, were located on a separate 'sub-plate' while Fennoscandia remained attached to Eurasia.

5.4 Late Oligocene and Foreland Tectonics

The main phase of back-arc sea-floor spreading in the western Mediterranean began in the Chattian (Cohen, 1980), and was accompanied by a re-invigoration of extension across central France. The French Upper Oligocene Basins show that the crust was being extended in an approximately NW-SE direction (ie parallel with that accomplished at the North Atlantic spreading ridge).

The overall pattern of Palaeogene basin development across Europe suggests that there were two major phases of development in the Upper Eocene and Upper Oligocene, with the zone of rifting and extension migrating towards the northwest. The main phase of back-arc sea-floor spreading in the western Mediterranean began in the Chattian (Cohen, 1980), and was accompanied by a re-invigoration of extension across central France. The French Upper Oligocene Basins show that the crust was being extended in an approximately NW-SE direction (ie parallel with that accomplished at the North Atlantic spreading ridge). The overall pattern of extension must reflect some complex regional zone of lithospheric shearing. In the Rhine Graben, the centre of subsidence shifted north, so that by the Middle and Upper Oligocene it was around the centre of the Graben where more than 1000m of sediments were deposited. In the late Oligocene, subsidence came to an end in the northern continuation of the Rhine Graben (the 'Hessen Depression') which became dry land.

At the beginning of the Oligocene (around 36Ma) a new northwesterly 'branch' of the Rhine Graben became active, including the small Neuwied Basin within the Rhenish Massif and the NW-SE trending Mesozoic Roer Valley Graben. On the Peel Horst on the northeast graben margin, the sediment thickness accords with an initial rifting event involving 2-5% extension completed by about 30Ma, followed by passive subsidence. The main graben is about 30-50km wide and the adjacent marginal horst about 25km wide, implying that around 1-2km of extension may be of Oligocene age (Zijerweld et al., 1992).

5.4.1 Southern North Sea

Similar extension and subsidence cannot be traced further to the north into the North Sea Basin. The Sole Pit Basin, sharing a similar orientation, was subject to inversion at this period. This suggests that the Roer Valley Graben was an extensional stepover structure, developed at a right handed offset, in NW-SE trending dextral transcurrent faults that had to negotiate the eastern end of the Anglo-Brabant Platform. On the northeast margin of the Platform, in the Sole Pit Basin, the stepping is to the left, hence causing transpression.

When, in the late Eocene-early Oligocene, the Rhine Graben continued as far north as the Hessen Depression, the transfer zone of strike-slip faulting passing towards the northwest was transpressional (ie left-stepping) both in central Holland and in the Sole Pit region. Following the mid-Oligocene abandonment of the Hessen Depression, however, this transfer zone appears to have become transtensional (right-stepping) in the Neuwied and Roer Valley Grabens while remaining transpressional in the Sole Pit Basin. The inter-relation of pronounced rifting with inversion demonstrates the importance of strike-slip faulting throughout this period.

5.4.2 Celtic Basins

At this same period, deep basins, similar to those found in southeastern France, began to develop in western Britain, along the lithospheric weakness associated with the 600km long N-S Hebridean igneous province. From south to north such basins include Cardigan Bay, Kish Bank, Lough Neagh, Blackstones, Canna and South Harris. The basins accumulated up to 1000m of late Oligocene (and early Miocene?) sediment and are bounded by reactivated NE-SW to N-S Caledonian fault zones. They appear to be interconnected by NW-SE or NNW-SSE dextral and NE-SW sinistral strike-slip faults. The northern most of these continental basins is located close to the southeasterly continuation of the Wyville Thomson Ridge. Hence the west British zone of rifting, and the E-W zone of compression and transcurrent fault motion around the English Channel, represent a more westerly successor to the Holland-Faeroes compressional shear boundary which was active at the Eocene-Oligocene boundary (Figure 5.5). This westerly sub-plate boundary divides Europe, including England, from the westerly relative motion of Ireland, France, Rockall and the adjacent North Atlantic.

5.4.3 Norwegian Continental Margin

In the Halten Terrace and the Outer Vøring Plateau, on the edge of the mid-Norway continental margin facing the Norwegian Sea, studies of the Cenozoic succession obtained from boreholes show that, following a thin (55-178m) sequence of Middle to Upper Eocene sediments (spanning from 38/42Ma to 36/37Ma), there was a major Oligocene unconformity lasting for 7-14My, which appears to correspond with the early phase of the re-organisation of the spreading axes in the North Atlantic. Slow sedimentation was renewed in the late Oligocene and persisted into the early Miocene spanning from 30/24Ma to 21/20Ma but only achieving a thickness of 45-247m (Goll and Hansen, 1992). This Oligocene unconformity appears to have accompanied the initiation of compressional deformation along the Norwegian continental margin, which was to persist, intermittently, through to the present.

A number of major domal structures, associated with compressional deformation, are identified from the 4,000m thick Mesozoic and Palaeocene sediments of the Vøring Basin (see Figure 5.6). These structures generally trend NNE-SSW, or even NE-SW, although in the centre of the basins appear as broad antiformal upwarps whose underlying structural control is not always evident. The largest of these structures is the Molde High, running N5E through the Vøring Basin, which can be traced for almost 400km and which acted as a barrier to sediment transport for much of the mid-Tertiary (Mutter, 1984). The structure was active from middle Oligocene to early Miocene times (Eggen and Vollset, 1984) and achieved a total uplift of about 1500m. Within this structure, individual anticlinal axes of inversion are internally complex and the Mesozoic stratigraphy cannot be traced across them (Mutter, 1984). DSDP holes 339 and 340 were drilled on this complex anticlinorium in the Voring Basin and revealed diapiric Oligocene and Eocene shales (Bukovics et al., 1984). This inversion complex is bounded to the south by the prolongation of the Jan Mayen Fracture Zone, where around the continental margin, the base of the Eocene shows a marked step, from 2km depth to the north down to 3.5km to the south.

Further to the east, a well-studied line of inversion structures is found along the NE-SW trending Nordland Ridge, which was also inverted from the mid-Oligocene through to the Miocene (Gowers and Lunde, 1984). At Traenabanken, located 100km off Helgeland, at around 10°E 66°30'N, a series of en echelon N-S Triassic faults, dipping to the east and arranged along an 80km NNW-SSE trending line, were inverted by an estimated 250m (Larsen and Skarpnes, 1984). Faulting had ended by the late Miocene after which there was general subsidence and tilting towards the WNW. Compressional effects are found along a distinct NNW-SSE trend, and can be followed both north and south of the Nordland Ridge as flexures in the basinal area. This NNW-SSE trend lies on-line with the prolongation of the Lofoten Fracture Zone transform that has acted to transfer a few hundred metres of sinistral displacement into the continental margin.

Further north, an (undated) outer-margin NNE-SSW trending anticlinal inversion structure is also known from the poorly investigated Lofoten margin (Rokoengen and Saettem, 1983) (see Figure 5.7). Similar compressional structures have not been identified to the north of the islands, although at Andøya, the only outcrop of Mesozoic (Lower Cretaceous) sediments onland in Norway is within a basin whose southern margin is inverted (Dalland, 1975). Thus, all along the continental margin between the Senja and Jan Mayen Fracture Zones, there is evidence for southeasterly to easterly directed compression, which, where it can be dated, was probably initiated in the mid-Oligocene. While this deformation began during the re-organisation of spreading ridges in the Norwegian Sea it continued far into the Neogene suggesting that its ulterior cause lay in the disparity between motion at the spreading ridge and the motion of the continents embedded in the plates.

5.5 Summary

Although there were profound differences in tectonic style across North-West Europe through the Oligocene there is a common theme of dextral shear on NW-SE oriented structures.

The change from the tectonics of the late Eocene to those of the late Oligocene reflects a transition from NW-SE dextral transcurrent faulting, associated with Pyrenean compressional tectonics, to transcurrent faulting of similar trend associated with E-W rifting. This rifting, in effect, reflects some component of Atlantic extension becoming transferred into western Europe.

Therefore, while the first stage was controlled by NW-SE strike-slip faulting passing off the Pyrenean plate boundary, the second stage involved both NW-SE dextral and NE-SW sinistral strike-slip faulting within the context of the regional N-S maximum horizontal compressive stress which pertained in northwest Europe at this period.

5.6 Constraints on Late Palaeogene Deformation in the Baltic Shield

The evidence of widespread, but relatively mild, compressional reactivation of appropriately oriented structures across the North Sea around the late Eocene and early Oligocene reflects a regional stress field that must also have impinged the Baltic Shield. The pattern of minor compressional deformation structures that can be mapped along the eastern edge of the North Sea basin imply no more than a few tens of metres of displacement on specific faults: the majority of the deformation being concentrated in the south-west of the basin.

The profound rheological differences between rifted and shield continental lithosphere, as are apparent, for example, where the Arabian Shield moves towards the Iranian collision zone, imply marked differences in levels of internal deformation. In the face of the major zone of continental collision in Iran, the Arabian Shield remains effectively aseismic.

Levels of deformation internal to the Baltic Shield during the far more subdued partial basin inversion of the late Palaeogene, are similarly likely to have been very restrained. Plausible levels of strain would have been more nearly 1%, and quite possibly less than 1%, of those seen on the adjacent Norwegian continental margin: equivalent to perhaps 10m -100m across the whole shield over a period in excess of 10 million years (a strain-rate of 10^{-11} - 10^{-12} per year).

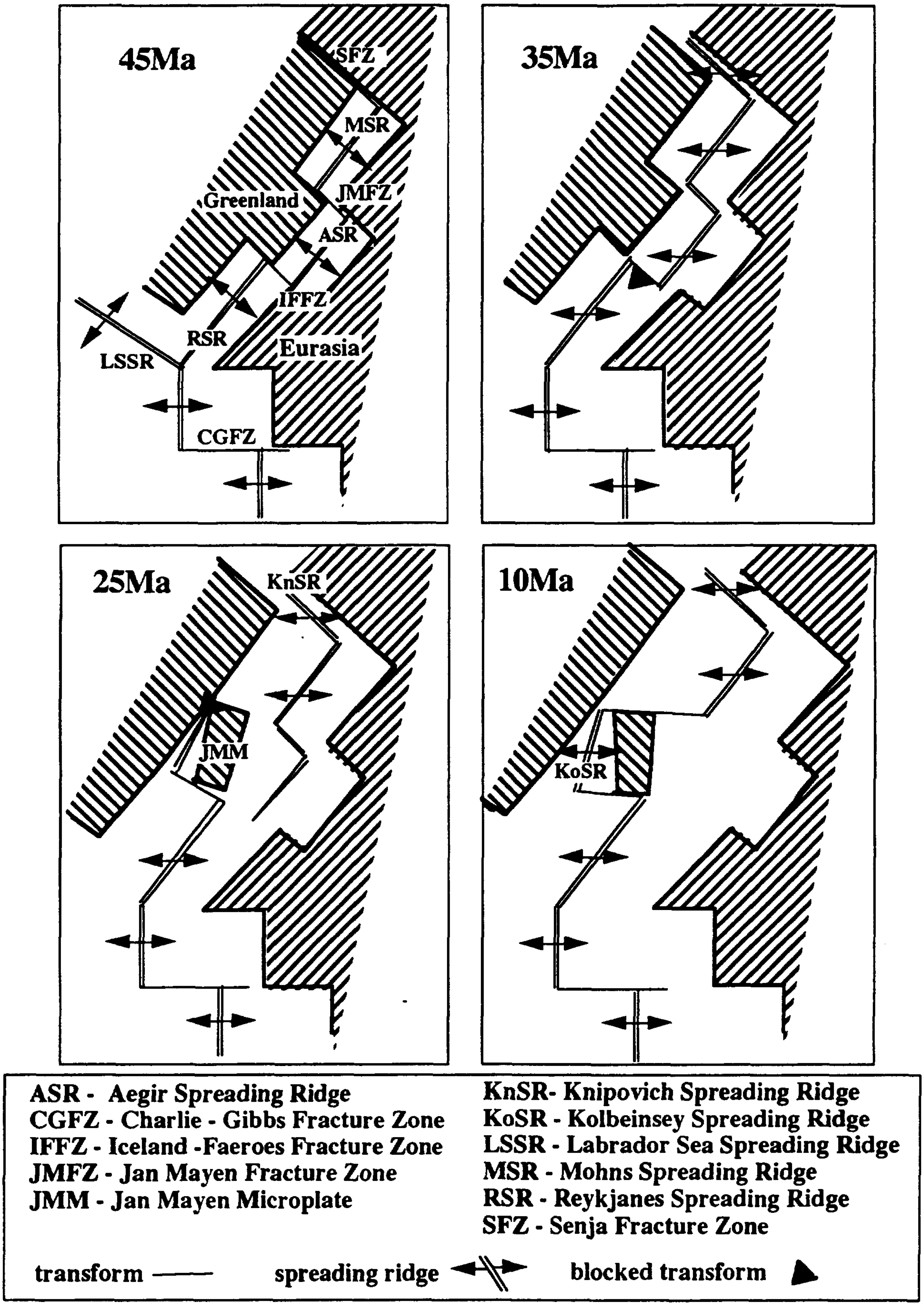
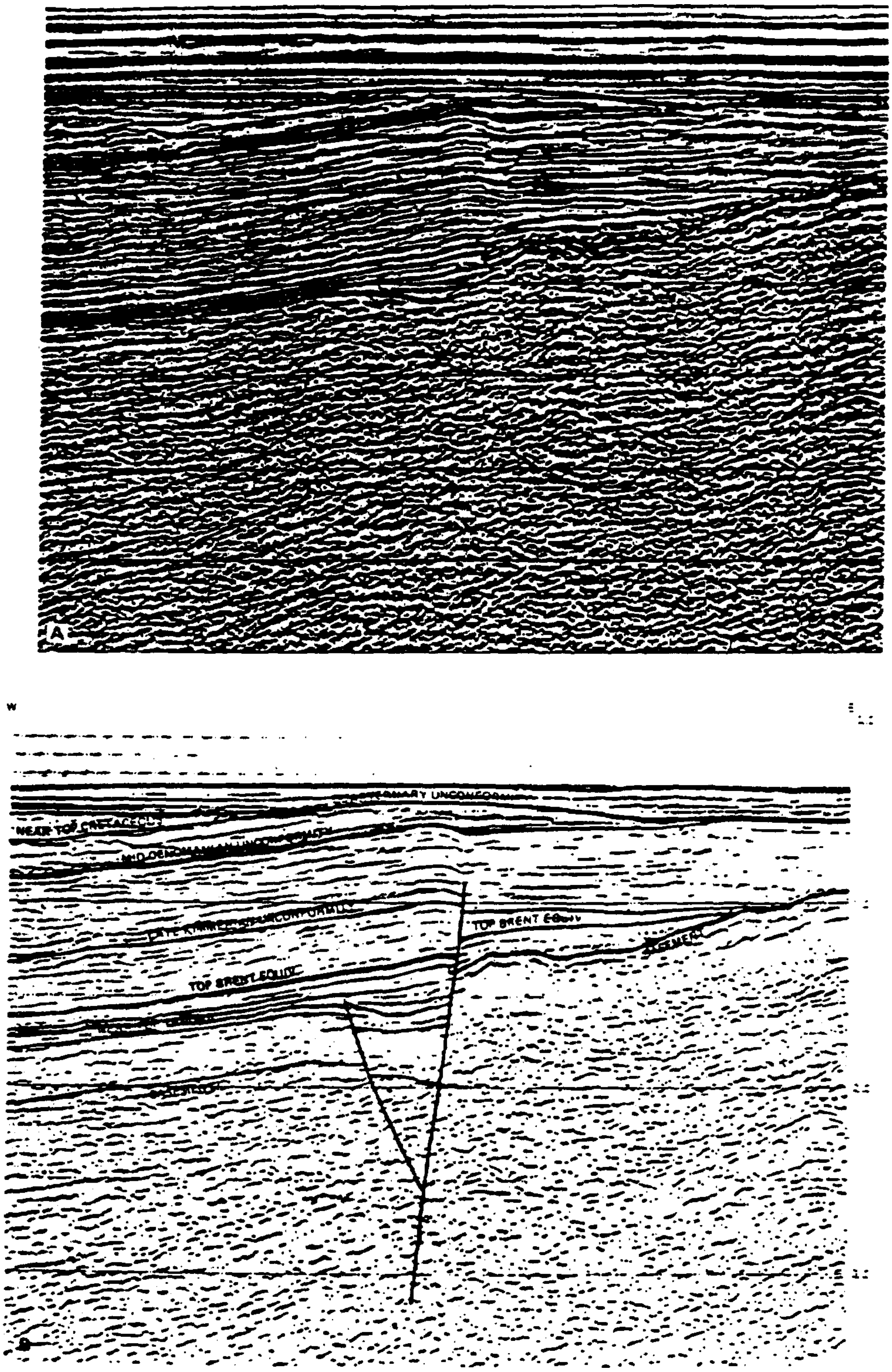


Figure 5.1 The Mid-Tertiary re-arrangement of North Atlantic spreading axes



UNINTERPRETED (A) & INTERPRETED (B) SEISMIC REFLECTION PROFILE OF AN INVERTED STRUCTURE IN THE SOUTHERN STORD BASIN (FROM BIDDLE & RUDOLPH 1988)

Figure 5.2 **Seismic reflection profile across an inverted rift margin fault in the southern Stord Basin, offshore SW Norway (from Biddle & Rudolph, 1988)**

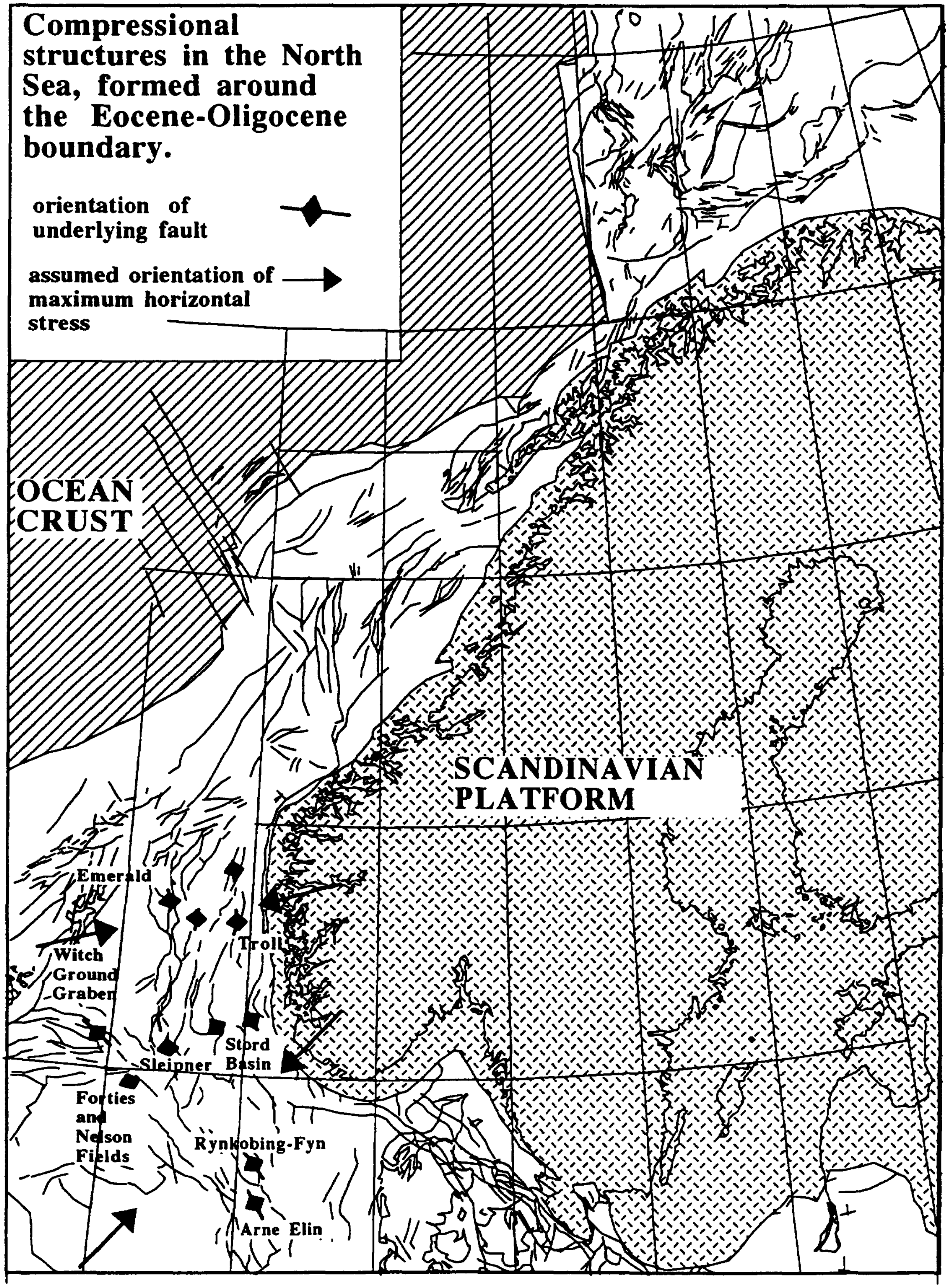


Figure 5.3 Early Oligocene tectonics around Scandinavia

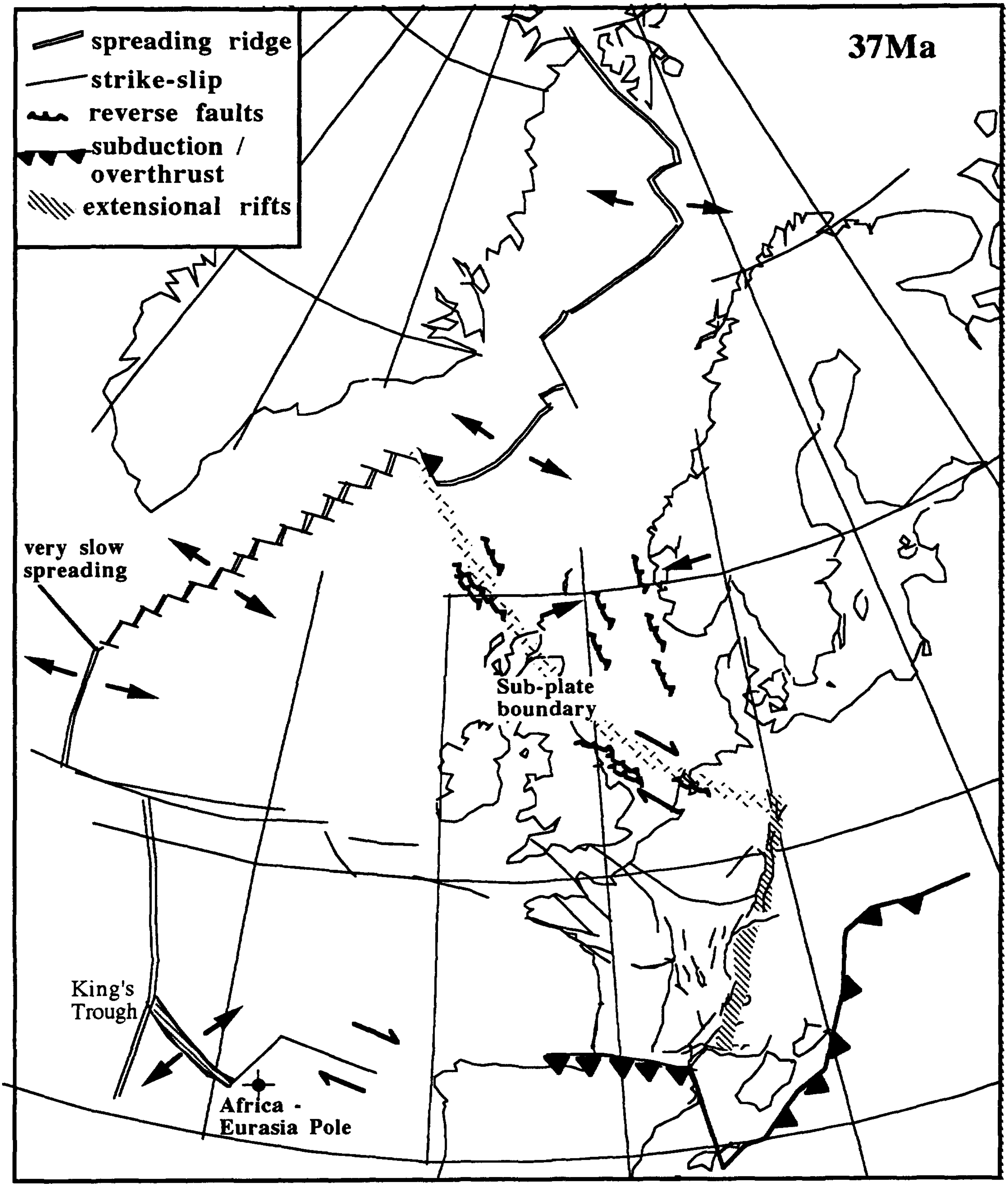


Figure 5.4 Early Oligocene tectonics around the North Atlantic

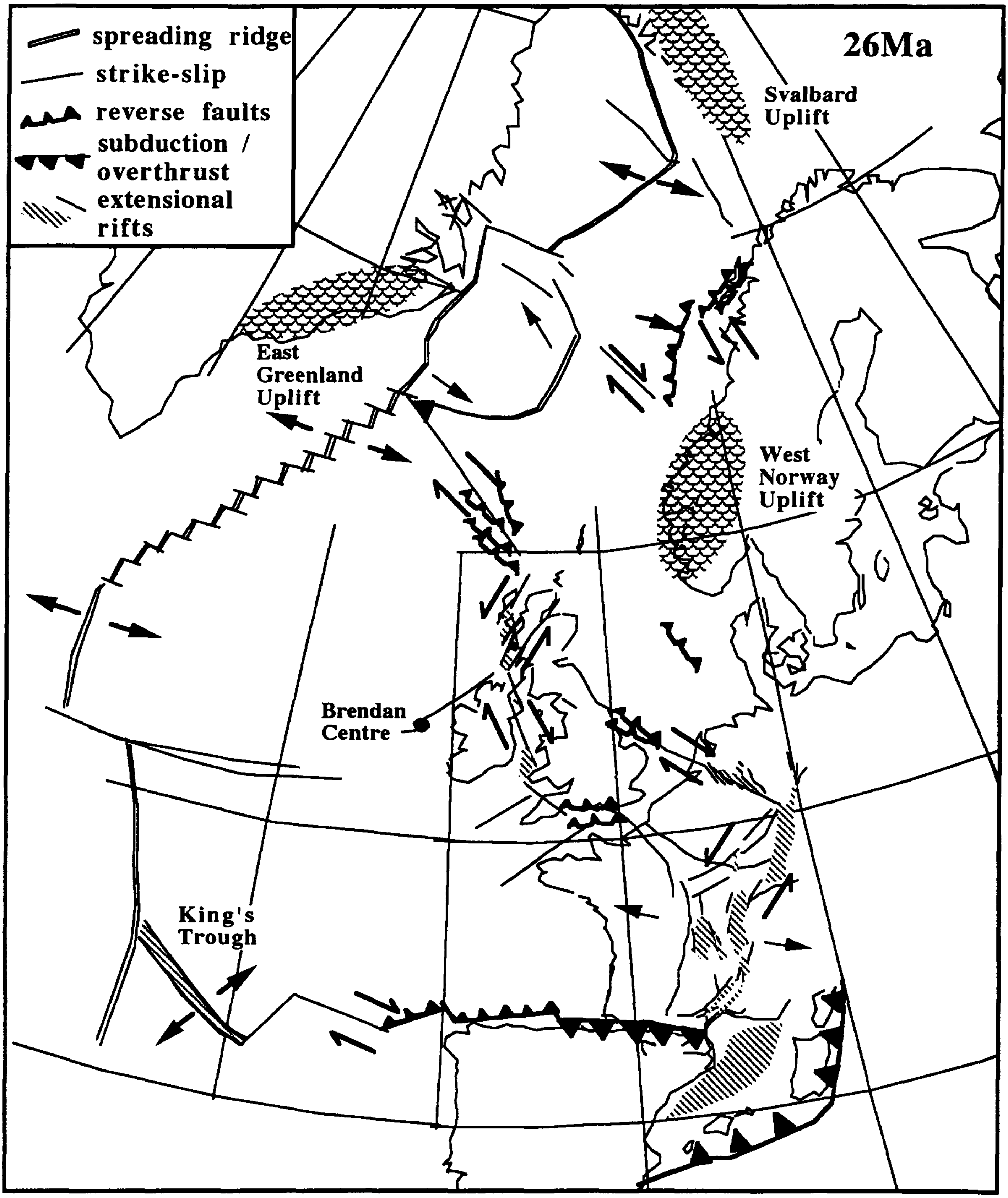


Figure 5.5 Late Oligocene tectonics around the North Atlantic

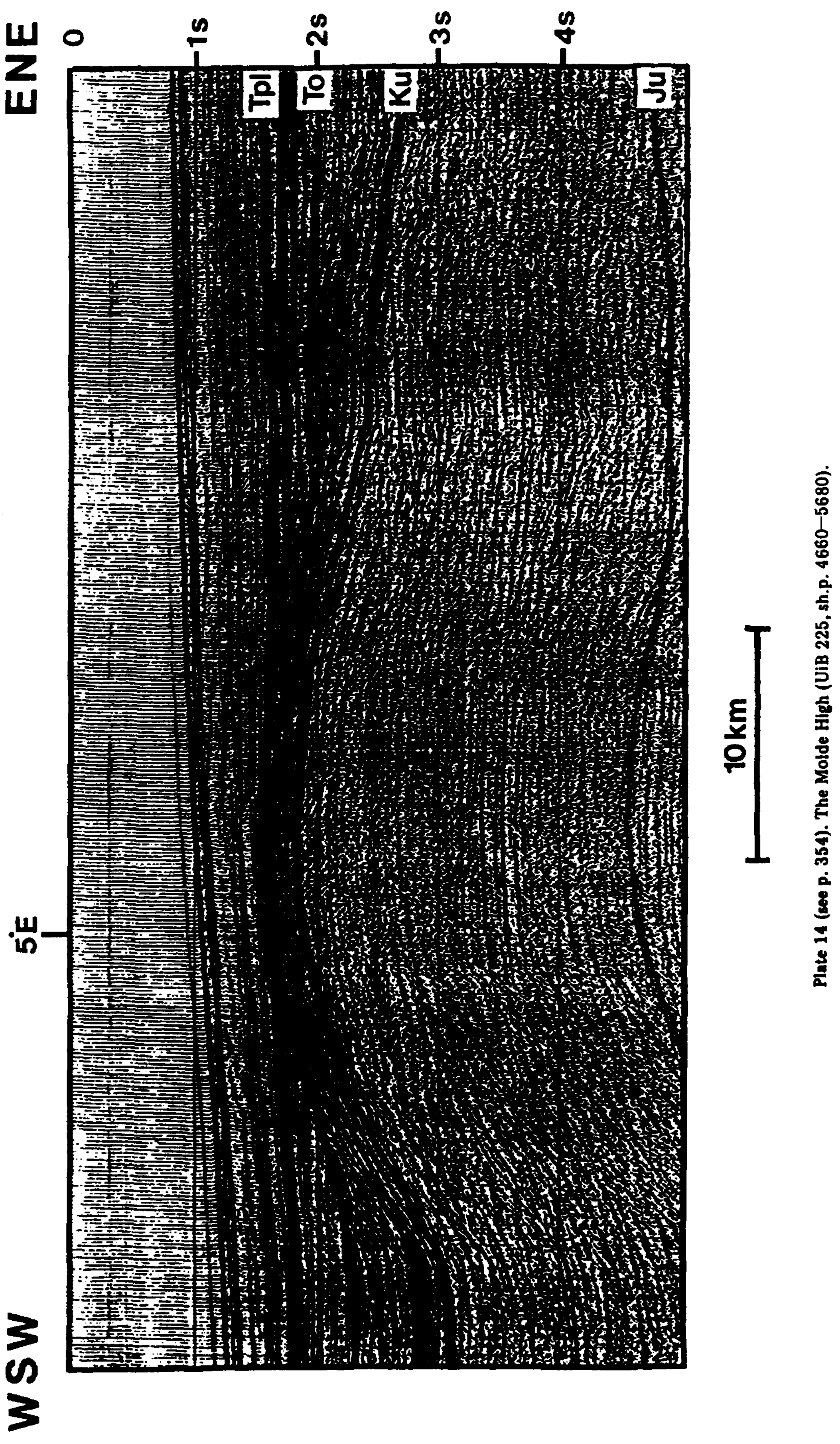
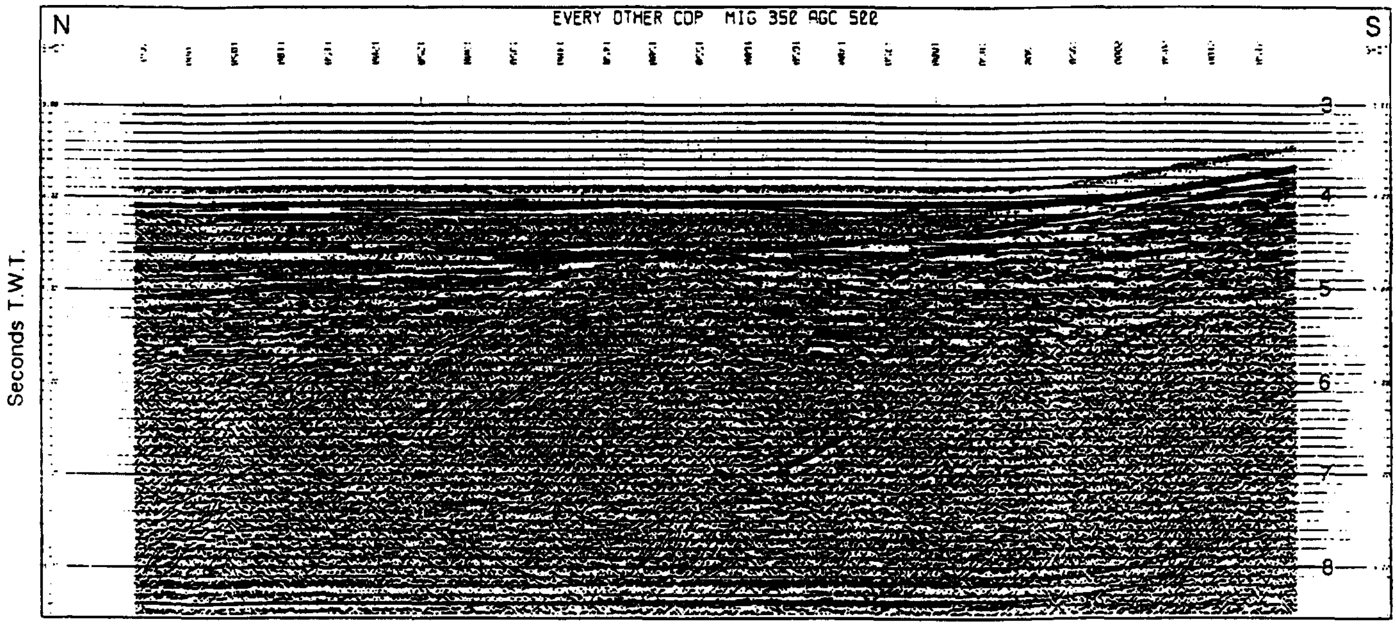


Figure 5.6 Seismic section across the Mølde High showing prolonged late Oligocene to Miocene anticlinal uplift above a blind reverse fault (from Hamar and Hjelle, 1984)

Seismic Line: UB 82 350



Line drawing: Line UB - 82 - 350

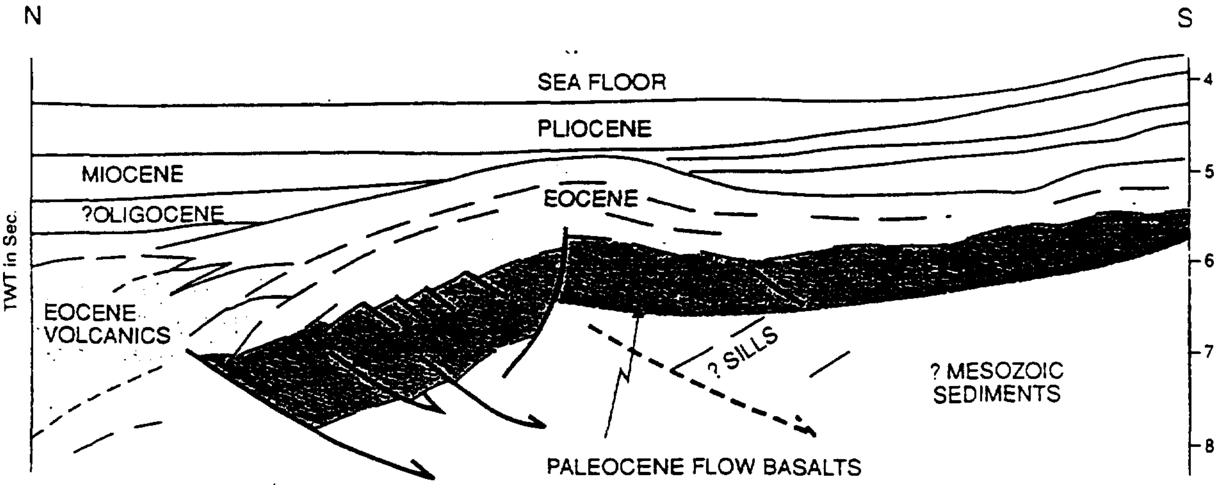


Figure 5.7 Seismic section across anticlinal uplift above a blind reverse fault on the outer Lofoten Margin (from Mokhtari and Pegrum, 1992)

6.0 MIOCENE

6.1 Plate Tectonic Context

During the Miocene, the Africa-Eurasia pole of relative rotation switched to the south, close to where it is today, at around 28°N and 20°W (Savostin et al., 1986; Klitgord and Schouten, 1985). From the pattern of Atlantic magnetic anomalies, Dewey et al. (1989) estimated that 200km of N-S convergence between Africa and Eurasia has occurred in the western Mediterranean region since Anomaly 13 (38Ma) and 50km of northwest-directed shortening since Anomaly 5 (9Ma). Relative Africa-Eurasia convergence remained N to NNE until Anomaly 5 although small changes in the Africa motion are found at Anomaly 6 (20Ma) (Dewey et al., 1989).

In the North Atlantic, once spreading had entirely passed to the west of the Jan Mayen microplate (around Anomaly 7), freeing the blocked Iceland-Faeroes Fracture Zone, anomalous ridge activity began (or become amplified) at Iceland, where asthenospheric 'hot-spot' diapirism has built up an enormous volcanic pile and made the mid-ocean spreading ridge sub-aerial (Bott, 1985). Further north, at the Knipovich Ridge, spreading has also been anomalous with asymmetric oceanic crustal accretion (faster to the west than to the east) as the ridge has continued to migrate towards the northeastern ocean margin around Spitsbergen.

6.1.1 Changes in Spreading Rate

The Miocene marks a most important period of transition in the tectonic development of the North Atlantic margins. Through the Tertiary, the rate of sea floor spreading in the North Atlantic between the Azores and Iceland declined, according to Vogt and Avery (1974), from an opening rate of 34 mm/yr in the early Tertiary down to about 14 mm/yr at 30Ma, rising again to around 22 mm/yr by 10Ma (see Figure 6.1). The increase from 15 to 20 mm/yr occurred at about 12-14Ma (in the mid-Miocene), and can be correlated with the development of time-transgressive basement ridges found in V-shaped configurations (expanding out from the centre of the hot-spot) both to the north and south of Iceland (Vogt et al., 1980). Between Iceland and the Azores new transform faults, had appeared around 44Ma as spreading ceased in the Labrador Sea. These faults disappeared with this renewed pulse of plume activity passing out from beneath the Iceland hot-spot (Vogt and Avery, 1974).

On the Greenland margin Hinz et al., (1993) report an 120 km wide area of abnormal upper oceanic crust parallel to the Kolbeinsey Ridge comprising internally divergent to planar patterns of reflections having dips towards the spreading centre. This pattern is very similar to that observed accompanying the earliest stage of sea-floor spreading along the Norwegian and East Greenland continental margins that accompanied the original Thulean plume. Such oceanic crust implies an episode of excess volcanic activity along the spreading ridge that reflects raised temperatures in the mantle. This abnormal oceanic crust formed between chrons 7-6 (25-19Ma late Oligocene) and 5A-5 (11.5-9.8Ma late middle Miocene) (Hinz et al., 1993). This is consistent with the resurgence of Iceland, and the increase in the spreading rate at this same period. Hence, the increase in spreading appears to correspond with, and be driven by, the surge in plume activity.

6.1.2 Iceland

The island of Iceland itself records changes in the rates of volcanic production and hence plume activity. The oldest lavas in Iceland date from the mid-Miocene, extending back to 16Ma in western Iceland and 13Ma in eastern Iceland. In the past 16Ma, spreading has been symmetrical along the spreading ridge to the north and south of Iceland, but not on the island itself, where since the mid-Miocene, the E-W dimension of Iceland is 40% (or 210km) wider than the separation of corresponding oceanic magnetic anomalies north or south (Helgason, 1985). This disparity can most easily be explained through shifting spreading axes, overprinting older crust.

The great mountain plateaux (reflecting overthickened oceanic crust) that are found right up to the eastern and western edges of Iceland do not pass into a series of eroded remnants of former 'Icelands', as is the case with the long-lasting series of hot-spot volcanic centres seen in the Hawaiian islands. Prior to the mid-Miocene period, the lava production rate that existed at the present site of the Iceland hot-spot was probably much reduced. From the lava sections preserved on Iceland it appears that the rate of magma production was at its culmination at about 10Ma.

In northern Iceland, a lava section, detailed by Saemundsson (1979), spanning from around 12Ma to 9Ma, yields an accumulation rate of about 1000m/My for the lower 2.5km but 4000m/My for the upper 2.5km. A 3.5km thick section in western Iceland, spanning from 6.5 to 2Ma, yields a more or less constant rate of accumulation of 780m/My. Walker (1982) detailed the geology of northeast Iceland and found that, between the major accumulations of plateau lavas at 12 and 9Ma and the renewed activity at 6.5Ma, the constructional surface (as deduced from the elevation of zeolite zones) dropped from 1200 to 700m. This implies a reduction in crustal thickness of at least 10%. Even today, the mountain plateaux of northeast and northwest Iceland testify to the thicker crust that was accreted during the late Miocene, and consequently the greater quantity of partial melting (and hence mantle upwelling) than that which prevails today.

Vogt et al. (1980) described a diachronous V-shaped basement escarpment, flanking the mid-Atlantic ridge to the southwest of Iceland, which they attributed to an abrupt increase in discharge from the Iceland mantle plume around 7Ma. Watkins and Walker (1977) also reported a short-lived 7.3-6.4Ma pulse of increased lava production in a continuous 9km section of lavas in eastern Iceland spanning from about 13 to 2Ma. Onland in Iceland, Jancin et al. (1985) identified a tectonic reorganization of spreading ridges also dating from 7Ma. Prior to 7Ma, Iceland had multiple-branched crustal accretion zones, including a proto spreading rift along the present N-E Iceland axial rift zone and a main rift along the

Snaefellsness Peninsula which became concentrated in a single zone at 7Ma.

Hence, the pattern of magma production of the Iceland hotspot appears to show a pronounced mid-late Miocene surge accompanied by an increase in spreading rates. As the increase in spreading rates appears to be driven by the plume, it is likely that spreading-rates have for a time exceeded the rate at which the plates are separating. Such a situation is likely to create intraplate compression. The re-awakening of plume activity in the North Atlantic appears to have been the trigger for a remarkable tectonic crisis whose affects can be found deep into Europe. Intraplate extension became replaced by compression across the whole of western Europe, accompanied by volcanic activity and subsidiary plume activity in central Western Europe. Plateau uplift of up to 2.5km has occurred in East Greenland since the new spreading axis developed adjacent to the coast in the Oligocene (Brooks, 1985). This parallels the major orientation and timing of uplift in western Scandinavia (see below), and is almost certainly related to the reinvigoration of the Iceland hot-spot. Tectonic activity around the continental margin to the south-east of Iceland also shows a clear geometric and chronological relationship to the hot-spot.

6.2 Continental Margin Compressional Tectonics

6.2.1 Shetland Continental Margin

Smooth post-rift subsidence of the Faeroes Shetland Basin was interrupted in Oligo-Miocene times by inversion and uplift in the southwest of the basin (Earle et al., 1989). A major mid-Tertiary erosional unconformity extends across the West Shetland Platform and the Solan and West Shetland basins, becoming less pronounced and eventually conformable towards the centre of the Faeroes-Shetland Basin. In the Solan Basin Pliocene sediments rest directly on early Oligocene or Eocene strata. The Rona Ridge was uplifted and deeply eroded at this time losing an estimated 1250m of cover (Booth et al., 1993). In deeper parts of the Faeroe Basin large scale low amplitude folds developed. At the same time many of the large normal faults systems of Late Cretaceous or early Tertiary age suffered some minor reverse motion. This important unconformity is assumed to be of mid-Miocene age although this is not well constrained (Booth et al., 1993).

The Faeroe Rockall plateau comprises a number of large banks 200-300km^2 in area, in which Tertiary volcanic rocks outcrop are covered by thin Quaternary or Neogene sediments. On the Rockall Platform a number of almost E-W oriented structures on the northern margin of the Faeroe-Rockall Plateau were reactivated as reverse faults affecting middle Miocene and older sediments (Boldreel and Andersen., 1993). Compression in middle or late Miocene times, is also seen around the Wyville Thomson Ridge Complex (see Figure 6.2) and along WNW-ESE to SW-NE oriented compression structures within and on the margins of the Faeroe-Shetland Channel. The NW-SE Clair Transfer Zone marks the north-eastern limit of the Clair Ridge inversion, suggestive of sinistral movement.

6.2.2 *British Isles*

Miocene tectonics in Britain involved compressional deformation which is particularly apparent around northwest Britain. The most consistent feature of this tectonic phase comes from the identification of compressional structures (reverse faults and overlying monoclines) striking NE-SW. Within western Britain, such compressional structures are most readily demonstrated from the margins of the Oligocene continental basins. These basins were the youngest (and apparently the weakest) basins affecting the lithosphere. Compressional deformation, post-dating basin development, can be demonstrated from the Canna, South Harris, Lough Neagh, Cardigan Bay and Bovey Tracey Basins.

Bailey et al. (1974) presented a series of profiles across the Hebridean and Irish margins bordering the Rockall Trough, drawn from seismic reflection profiles between 57° and 54°N. At the northern end of these traverses, at around 56°30'N, faulting apparently cuts most of the sedimentary sequences and is associated with a steepening of the continental slope. To the northwest of Ireland, folded Mesozoic strata are unconformably overlain by Miocene sediments. The margin is also very steep at around 55°. Around the Geikie igneous centre to the west of Lewis, between 58° and 59°N and 80km NW of St. Kilda, the configuration of the Geikie and Hebridean escarpments appears to show the influence of Neogene tectonics. The Geikie escarpment marks a major transection of thick Oligocene and Miocene sediments and appears to reflect slumping or erosion off the continental margin (Evans et al., 1989). The underlying horizons beneath the escarpment are continuous, however, showing that there is no fault control. In contrast, the Hebridean escarpment, located generally a few kilometres to the west, appears to be controlled by N-S trending or E-W trending faults. Plio-Pleistocene sediments are continuous across the steep scarp but all earlier horizons are truncated, suggesting that deformation and erosion may have occurred in the Miocene, followed by subsidence.

Overthrusting is seen along the continental slope north of St. Kilda (Roberts, 1989) and involves uplift of as much as 500m. This was considered by Roberts to be of Oligocene age, but south of St. Kilda the linear slope is controlled by Miocene faulting (Roberts, 1989), probably associated with underlying reverse displacements. Compressional structures of the same NE-SW trend are also found along the continental margin, as far north as 61°, on the southeast flank of the Faeroes-Shetland Basin, where Palaeocene normal faults sustained minor (100m) inversion.

6.2.3 *Vøring Margin*

In the Halten Terrace and the Outer Vøring Plateau, slow early Miocene sedimentation continuing until 21/20Ma was followed by a mid-Miocene unconformity that lasted 7My -15Myr. Compressional deformation persisted along a number of structures along the Mid Norwegian continental margin (see Section 5.4.3 and Figure 6.3). The mid-Miocene unconformity was followed by rapid subsidence creating 1400-1955m of Middle Miocene to Holocene sediments (Goll and Hansen, 1992). This ('Savian') unconformity reflects a regional event which preceded the reawakening of plume magmatism beneath Iceland. This mid-Miocene unconformity is also found around the northern North Sea indicating regional uplift at this period. The subsidence history of the North Sea Basin can be separated into a pre- and post- mid-Miocene period.

6.2.4 West Scandinavia

The major episode of uplift of western Norway had been initiated by the beginning of the Miocene as recorded in the sedimentation of the adjacent basins such as the Central Graben (see Jensen and Schmidt, 1992). The first phase of uplift in the early Eocene, at the time of the opening of the North Atlantic spreading ridge, amounted to only a few hundred metres and the more impressive broad domal uplift (of up to 2,000m) began in the late Oligocene. In the Stord Basin, offshore Oligocene sedimentary sequence on the eastern flank of the basin is the first post-Palaeocene unit to be progradational and the overlying Miocene sequence is characterized by large westward-prograding clinoforms (Ghazi, 1992) which continued through the Pliocene. In Denmark, the first sedimentary unit to contain terrigenous clastic material in quantity is also of Upper Oligocene age (Spjeldnaes, 1975). In the Miocene, there was a major drainage system from the east and northeast and sediments were dominated by an alternation between terrestrial/limnic sands with lignite, and marine glauconitic and micaceous clays.

Along the eastern margin of the North Sea basin, up to 1500m of Mesozoic and Tertiary sediments have been lost to erosion following Neogene uplift. The total relief from the base of the Neogene depocentre to the top of the mountains is today about 4,000m. The uplift of western Scandinavia must have reflected some change in sub-lithospheric densities but was not accompanied by any significant rifting or volcanism. This change in densities most plausibly reflects the replacement of asthenosphere (or the base of the lithosphere) with lighter plume mantle.

6.2.5 North Sea

There is some evidence for relatively minor tectonic activity in the eastern North Sea, concurrent with this uplift. In the Danish sector of the North Sea, latest early Miocene extension along the NNW-SSE Tail End Graben-Sogne Basin trend involved movement along the Coffee Soil Fault, with about 100m more subsidence than in the surrounding region (Clausen and Kortsgård, 1993). The whole Oligocene sequence in this region is relatively thin (typically about 300m) and the early Miocene sediments increase in thickness towards the east and pinch out on the Arne-Elin Trend, reflecting progradation of sediments from the northeast.

A short N-S trending normal fault is seen to cut the Oliogocene sequence to the east of the Troll Field, related to a releasing bend in an underlying N-S fault subject to minor (tens of metres?) sinistral strike-slip displacement at this period. Around the mid-North Sea there was minor structuration and inversion in the mid-Miocene on structures also uplifted around the beginning of the Oligocene including the Forties Field and the NNW-SSE trending Nelson Field located immediately to the southeast (Whyatt et al., 1992). In the Witch Ground Graben, a phase of extensional faulting occurred at the end of the Mid-Miocene. This appears on seismic profiles to have involved faults which primarily detach at or above the Eocene section (Harding et al., 1990). In the Tail End Graben, compressional deformation persisted through the mid-Miocene and the structural highs of the Roar, Tyra, Gorm and Skjold oil fields remained elevated at this period (Megson, 1992). In a detailed study of sediment isopachs, Clausen and Kortsgard (1993) noted that the Arne-Elin trend ceased to be a structural high during the mid-Miocene, while the Coffee Soil Fault continued to control differential subsidence through to the late Miocene.

The short-lived and minor episode of mid-Miocene compression is contemporary with the far stronger deformation along the continental margin. As with the compressional deformation seen along the Norwegian continental margin, this coincides with the 14-12Ma increase in opening of the North Atlantic spreading ridge and the re-invigoration of the Iceland plume. This episode marks the final episode of compressional tectonic deformation that can be identified within the North Sea Basin.

6.2.6 Barents Sea

In southwestern Barents Sea erosion estimates range between 0-1000m in the Tromso Basin (Riis and Fjedskaar, 1992), 100-1500m in the Hammerfest Basin and Loppa High (Berglund et al., 1986; Wood et al., 1990), 1750-2050 m in the Maud basin adjacent to the Svalis Dome (Loseth et al., 1992) and more than 3000m at the Stappen High (Wood et al., 1990). Most of the wedge of sediments in the adjacent Lofoten Basin appears young and related to glacial erosion. Subsidence of the Senja margin has led to the formation of 5-7km of middle and upper Cenozoic sediments. Removing subsequent isostatic recovery, uplift in the southwestern Barents Sea is only 0-200m while it is around 1000m in the northwestern Barents Sea and Svalbard (Vøgnes et al., 1990). Positive features in the Sorvestsnaget basin were related to compression by Riis et al., (1986) and Brekke and Riis (1987) but they were interpreted as salt structures by Faleide et al., (1993). Salt mobilisation was considered to be a response to deep burial beneath the Upper Tertiary sedimentary wedge along the ocean margin.

6.3 Alpine Collision Zone

In the earliest Miocene Aquitanian (23Ma), extensional tectonics were in evidence throughout the western Mediterranean, involving the rotation of

Corsica-Sardinia and Mallorca-Menorca, while compressional tectonics had been initiated in the northern Apennines. At the end of the Miocene, there was a major change in boundary conditions at Anomaly 5 (9Ma), according to Dewey et al. (1989), accompanied by a concentration of compressional deformation in Sicily and Calabria, as well as along the North African margin. This was accompanied by back-arc extension in the Tyrrhenian Sea.

A cumulative total of approximately 250km of European lithosphere shortening was involved in the Alpine deformation. The mountain range represents dextral transpression along the northern boundary of the Alpine chains and sinistral transpression along the N-S sector of the western Alps. There was progressive deformation through time along the Alpine plate boundary.

6.3.1 Alpine Foreland

In the late Oligocene and early Miocene, the southern parts of the Molasse Basin in Switzerland and Germany became overridden and partly scooped out by nappes. In the foreland area of Germany, some of the fault systems active in the late Cretaceous and early Palaeogene were reactivated at this period although only with displacements of some tens of metres (Nachtmann and Wagner, 1987). North of the Danube, thrust faults cut the Molasse series: west of the South Bohemian Basement Spur, these thrusts continued into the Eggenburgian (earliest early Miocene), and east of this high, they continued up to the Carpathian (late early Miocene) (Wessely, 1987).

In the late Miocene, there was a change in the fundamental foreland tectonics of central western Europe, when the Apulian promontory of Italy collided with Europe in the western Alps and allowed the force of the Africa-Eurasia collision zone to pass into western Europe. It was at this period that the Helvetic nappes were thrust over the former Molasse Basin in Switzerland and the Jura mountains were formed from the complete inversion of a Mesozoic basin. The Jura mountains reflect a crustal shortening of around 25km (Burkhard, 1990). Latest mid-Miocene sediments are found beneath thrusts in the Bresse Graben. However, the timing of the ending of deformation in the Jura remains in dispute: Laubscher (1988) infers an end in the Tortonian while some other authors have considered that deformation still continues at a moderate pace today (e.g. Burkhard, 1990).

6.3.2 The Rhine Graben

In the early Miocene, 1600m of sediments accumulated in the northern section of the main Rhine Graben. During the late Miocene and early Pliocene subsidence was discontinued and there was fluvial erosion of the Graben floor (Illies, 1972).

From observations of sediment thicknesses, Meier and Eisbacher (1991) noted that, by early Miocene time (22Ma), the direction of extension had shifted from ESE-WNW to NE-SW. In the southern half of the Graben, there are no Aquitanian age sediments. During late Miocene and early Pliocene times, subsidence was interrupted for about 15My and much of the graben fill was subject to erosion (Illies and Greiner, 1978).

6.3.3 Volcanic Activity

Domal uplifts began to form in the Massif Central of France in the early Miocene associated with marginal rifting, and alkali volcanism continued through to the Holocene. A domal uplift over the Black Forest and Vosges mountains, which flank the southern Rhine Graben, began in the mid-Miocene and led to some large alkali volcanoes within the rift. The domal uplift of the Rhenish Massif, on the Belgium-German border, which began in the late Miocene, has to-date reached an elevation of almost 500m and has been associated with active rifting through the Lower Rhine (Roer Valley) Graben, as well as alkali volcanic activity.

The Vogelberg volcano, the largest neovolcano of central Europe, developed in the Upper Miocene at the northern end of the Upper Rhine Graben within the narrower Hessen Rift. Further south in the Upper Rhine Graben, there is another major Miocene alkali basaltic Swabian volcano.

At this same period, other late Miocene grabens began to form, the most prominent being the northern Bohemian Graben which was accompanied by ENE trending basaltic dykes (Malkovsky, 1976). Such a trend of extension is also typical of active 'back-arc' rifting in the inner Carpathians and suggests marked variations in the stress regime to the north of the Alps at this period.

6.4 Summary

The Miocene period reflects a phase of transition between the active sub-plate boundary tectonics which dominated north west Europe in the late Palaeogene and the return, for the first time since the early Eocene, to a passive margin. However, this transition became interrupted by a mid-Miocene 'convulsion' which appears to have involved widespread compressional tectonics along much of the continental margin of NW Europe, with the accompanying volcanic activity suggesting the active role of the mantle. These features are likely to have been inter-related.

The early Miocene tectonics can probably most easily be considered as a continuation of those of the late Oligocene. Plate collision continued through the Pyrenees into the early Miocene and subsidence and rifting continued along the northern Rhine Graben. However, from the mid-Miocene, throughout much of Europe, the dominant tectonic style once again became compressional. Compressional deformation also dominated along the Norwegian continental margin at this period and the significant crustal shortening and basin inversion observed in western Britain is almost certainly of the same age (see Figure 6.4). However, in the centre of western Europe, compressional tectonics did not come to dominate until the late Miocene.

The accompanying Miocene mantle activity has the appearance of a series of independent mantle upwellings, causing doming and, across the Hercynian terranes of Europe, volcanic activity. One can almost discern a mantle wave spreading out from the Atlantic margins where the uplift of western Scandinavia, and probably northern Britain, was initiated in the late Oligocene, in concert with renewed igneous activity at the Brendan Centre to the west of Ireland. Igneous activity accompanying domal uplifts arrived in the Massif Central in the early Miocene, in the Black Forest and Vosges in the mid-Miocene, and in the Rhenish Massif in the late Miocene.

The simplest explanation for this combination of mantle activity and compressional tectonics is some change in the relative motion between the Eurasian plate and the underlying asthenosphere. This, in turn, is suggested by the evidence for an acceleration of the speed of Atlantic opening which occurred around the mid-Miocene, accompanied by a massive re-awakening of the Iceland hot-spot. An acceleration of plate movement accompanied intraplate compressional deformation just as the deceleration of relative plate movement through the Eocene and Oligocene accompanied intraplate extensional rifting.

As the magnitude of the inversions seems to be greater towards the northwest, they cannot simply reflect Alpine foreland compression, but instead a dominance of deformation transmitted from the mid-Atlantic spreading ridge around Iceland. At this same period, similar easterly-directed compression continued along the whole of the mid-Norwegian continental margin in particular to the north east of the Jan Mayen Fracture Zone.

6.5 Constraints on Miocene Deformation in the Baltic Shield

Throughout the Cenozoic, the most significant episode of deformation marginal to the Baltic Shield occurred in the Miocene. The evidence for pronounced basin inversion along the rifted margin of mid-Norway demonstrates however that almost all crustal shortening was concentrated on the weakest rifts, rather than on other parallel structures also seen on the margin. This argues for relatively little deformation being accommodated within the shield itself. However the Lofoten margin is itself much steeper than to the south and the absence of major rifted basins to act as a buffer for deformation does argue for some component of crustal shortening passing into the Caledonides on the shield's western edge. The Lofoten islands themselves lie on-strike with the continuation of some of the compressional structures seen to the south-west.

Hence it is probable that levels of deformation adjacent to the Lofoten continental margin along the coastal Caledonides of northern Norway may have involved displacements of specific faults involving tens or hundreds of metres, consistent with those seen on the continental margin to the southwest. However no faults have yet been found inland whose displacement has been dated from this period.

Beyond the zones of the most intensive deformation, often along the axis of the major rifted basins along the western Norwegian continental margin, many other parallel structures have little if any evidence for activity suggesting that even within the thinned crust of the margins, strain was highly localised. Further into the cold rigid lithosphere of the Baltic Shield, deformation is likely to have been subdued, with regional strainrates expected to have been no higher than 10^{-11} per year, perhaps 1% of those of the margins. Some flexural deformation is also likely to have accompanied regional crustal doming in western Scandinavia, but a dome 500m high and 500km in radius produces a strain-field comparable to the cumulative strain of post-glacial rebound, but distributed over a time period 100 to 1000 times as long. This again implies horizontal strain rates only of the order of 10^{-11} per year.

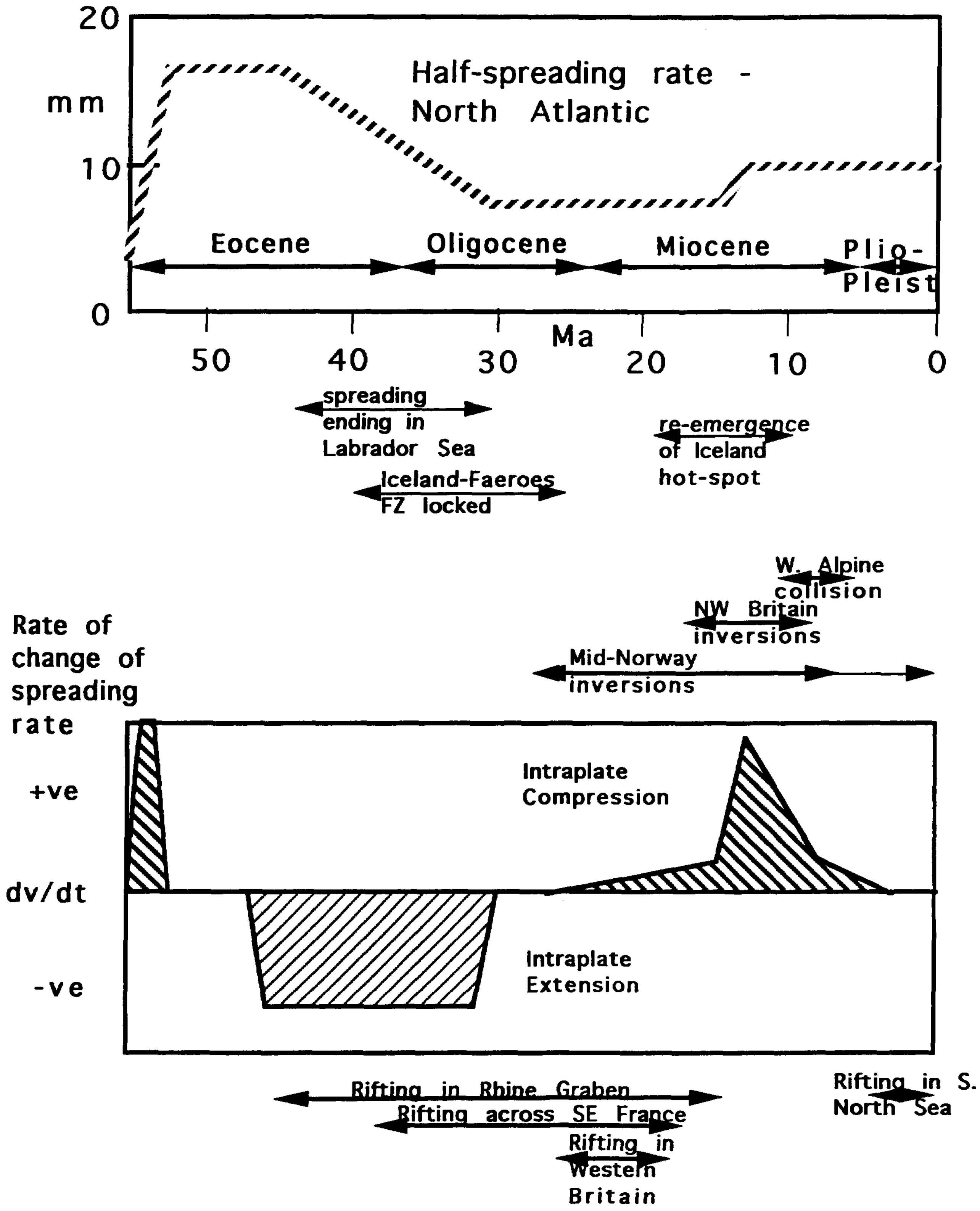


Figure 6.1 Changes in spreading rate along the North Atlantic spreading ridge through the Tertiary

Figure 6.2 **Compressional tectonic deformation along a NE-SW section across the Ymir Ridge south of the Faeroes islands (from Boldreel and Anderson, 1993)**

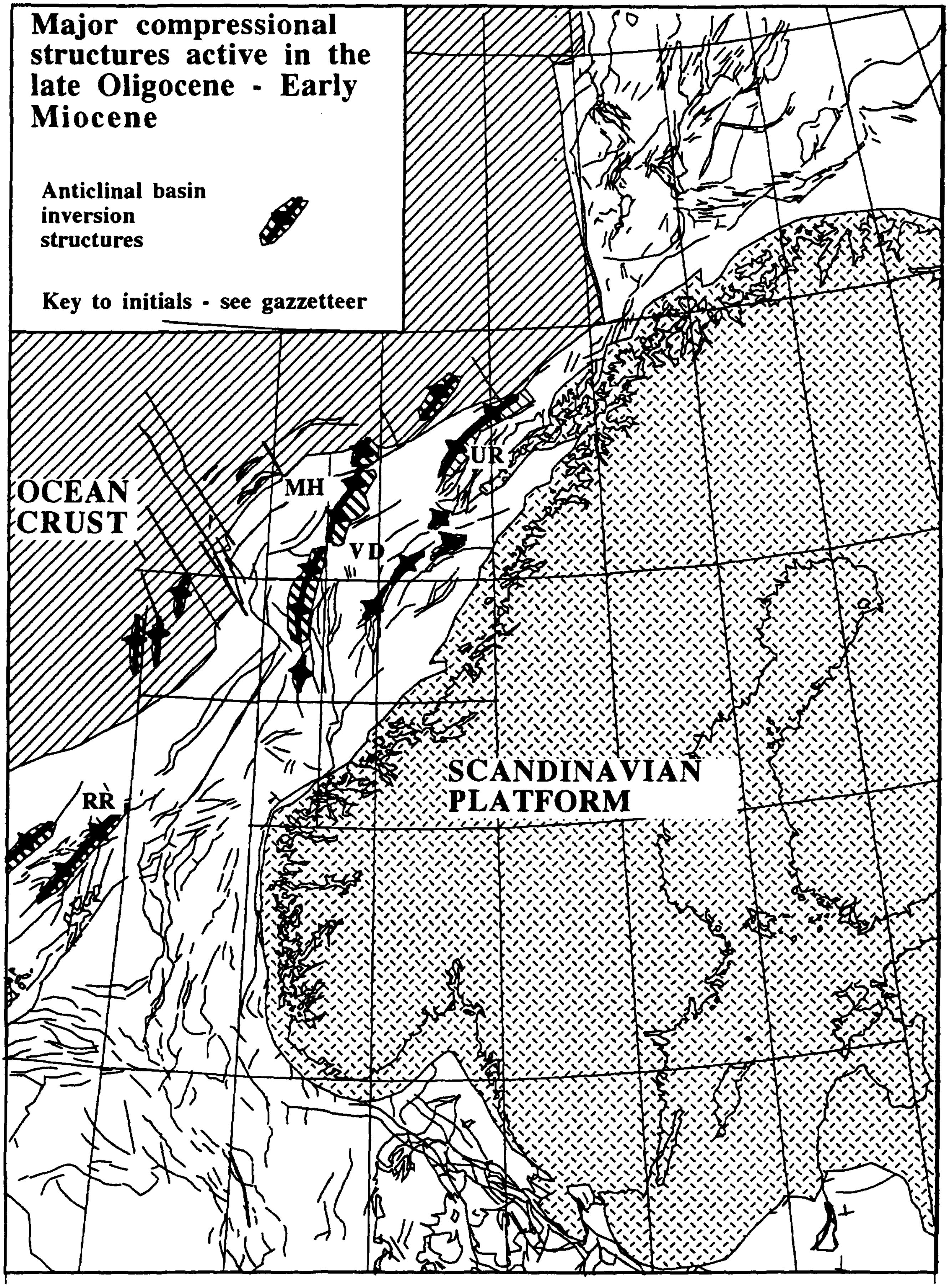


Figure 6.3 The disposition of compressional structures along the Norwegian continental margin active in the Miocene

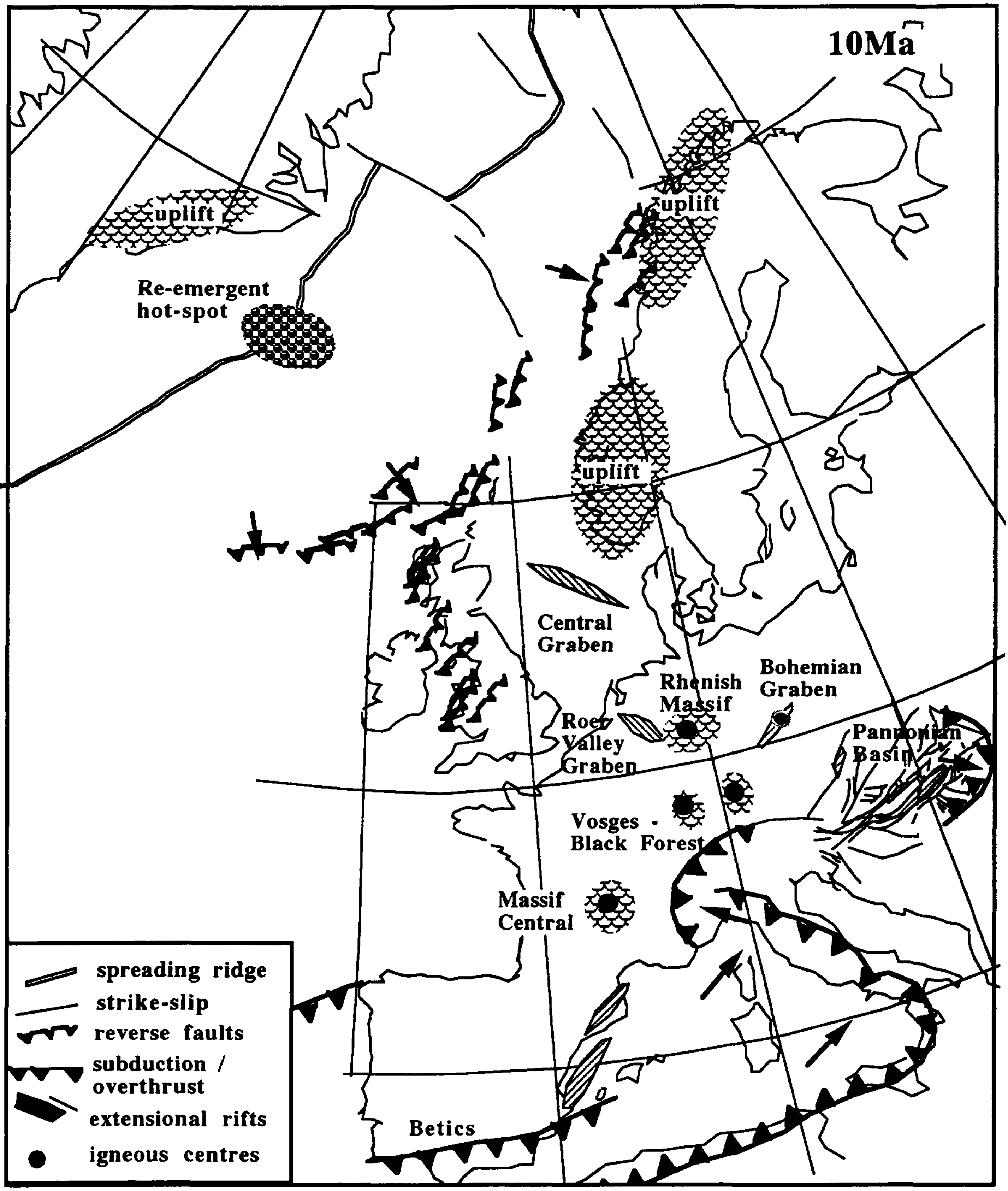


Figure 6.4 Mid Miocene tectonics around the North Atlantic

7.0 PLIO-QUATERNARY

7.1 Plate Tectonic Context

Throughout the Plio-Quaternary, as determined by the separation of magnetic anomalies, sea-floor spreading in the North Atlantic has been at an approximately constant rate, with little evidence for deformation radiating into the Atlantic continental margins. The 7Ma lavas on Iceland are now separated by 140km, a distance which is consistent with a half spreading rate of 1cm/yr (Helgason, 1985). Since 7Ma, it appears from the thickness of plateau lava fields that the rate of magma production and spreading has remained approximately constant.

7.2 Continental Margin Deformation

7.2.1 Mid-Norway Margin Extension and Collapse

Along most of the continental margin the mid-Miocene compressional deformation can also be seen from seismic reflection profiles to have stopped before, or by, the end of the Miocene. This can be seen on profiles across the continental margin of west Shetland and also along much of the margin of mid-Norway where such reverse faulting was most prominent. The end of this compressional deformation appears to have coincided with a period in the early Pliocene when rapid subsidence was initiated on the mid-Norway continental margin (Larsen and Skarpnes, 1984; Goll and Hansen, 1992). Rapid Pliocene subsidence of greater than 300 m per million years caused significant overpressuring of Haltenbanken oil fields (Jensen and Fore, 1993). This subsidence episode was dated as chiefly being in the late Pliocene to Early Quaternary, when 1 km of claystones were deposited in 1-1.5 m.y. (Vik and Hermanrud, 1993). Such rapid subsidence in the late Pliocene and early Quaternary is also observed along the Møre Basin margin and around the southern North Sea (Thorne and Watts, 1989), but not to such an extent in the northern North Sea.

Accompanying this collapse there is little general evidence of tectonic extension. However, on one high resolution seismic reflection profile (line VB-2-87) (Granberg, 1992) across the southern end of the Fulla Ridge inversion in the central Vøring Basin indicates some extensional collapse. This can be seen on a NW-SE seismic profile across the Fulla Ridge, where on the south-east side of the ridge, along the line of the Fles Fault, an extensional fault with a cumulative Pliocene displacement of ca. 150 m can be seen dipping to the north west. This fault is an antithetic structure to the main south east dipping reverse fault involved in the uplift of the Fulla Ridge in the Miocene (see Figure 7.2). However, such extension is not seen on a comparable seismic profile across the Fulla Ridge further to the north east. Hence, rapid subsidence accompanied a significant reduction in the level of horizontal stress that had prevailed during the Miocene. Locally this reduction in horizontal stresses led to the partial collapse of some inversions.

There is also a suggestion of more recent extensional faulting along the Lofoten continental margin where on Line LM-2-876 to the north west of the Røst high, Mokhtari (1991) indicates a recent fault scarp in about 2000 m of water on the continental slope. This fault is interpreted by Mokhtari as an extensional structure although the dip angle given to the fault suggests it could in fact be the head-scarp of a massive rotational slide, as along the Storregga structure on the Møre margin.

7.2.2 *Mid Norway Margin - Compressional Reactivation*

If extension was the dominant mode of tectonics in the Pliocene, within the Quaternary, two of the mid Vøring Basin structures, the Fulla Ridge and part of the Mølde High, active in the Miocene have become reactivated once again in compression. Across the Fulla Ridge the episode of Pliocene extension can be seen to have become replaced once again by compression, in the mid-late Quaternary involving further modest reversal of some faults active in the Miocene with an estimated 50 m uplift across the structure (see Figure 7.2 from Granberg 1992).

Internal diapirism and mud volcanoes are also evident from recent mud diapirs identified by Mutter (1984) on the sea floor of the Inner Vøring Plateau around 67°N 6°W, suggestive of continuing deformation. A prominent diapir province, measuring about 40 km N-S 25 km E-W at 57°N, 50-60 km east of the Vøring Escarpment contains individual structures rising more than 70 m above the sea floor. The diapirs rise from the centre of the Vema Dome around the Nyk High.

However, the majority of the compressional structures active in the Miocene show no evidence of renewed compressional reactivation, at least from available seismic reflection profiles. Such structures include those in the Outer Møre Basin illustrated in Hamar and Hjelle, 1984) and those along the continent-ocean transition off the outer edge of the Vøring Margin (Mokhtari and Pegrum, 1992).

7.2.3 *Barents Sea Margin*

The south west margin of the Barents Sea shows very little evidence of Plio-Quaternary faulting with one exception. At around 72.1°N 14.7°E, a zone of extensional faults, probably associated with strike slip displacement is seen (Fiedler, 1992). On one seismic section (Line BGR-37) two faults spaced about 2 km apart form a graben that passes up to the sea floor, with the dominant fault dipping to the east, while on a seismic line located about 10 km to the north west a series of parallel easterly dipping faults is seen with less evidence for passing up through the full Quaternary section. These faults appear to trend NW-SE and may be a continuation of the Greenland Fracture Zone heading into the continental margin. Their displacement suggests some transtension along this line. Similar displacements have not been followed beyond the continental margin.

7.2.4 Northern North Sea

Some minor evidence for faulting is seen in the northern North Sea, on faults associated with the West Viking Graben Boundary Faults, possibly passing up into the Neogene section, although it is always difficult to separate faulting from compaction-related movements, associated with differences in the thickness of sediments across some major faults.

7.2.5 Uplift of Western Scandinavia

There is also some evidence to indicate continuing 'tectonic' uplift of western Scandinavia. In two regions along the exposed western coast of Norway there are wave-eroded caves that lie at elevations above the Late Weichselian marine limit. On the Møre-Romsdal coast these cave sills range in elevation from 35 m to 90 m (Holtedahl, 1984) and almost certainly reflect inter-stadial or interglacial sea-levels. Similar caves are also found in Lofoten and the Vesterålen where they are also located above the Weichsellan marine limit (Møller, 1985). The sills of these caves decrease in elevation to the south from ca. 40 m (one exceptionally at 58 m although rubble filled) down to 20 m at Røst. The close similarities of all these caves suggest their contemporaneous formation and if truly interglacial reveals ongoing and differential coastal uplift. The discovery of an Eemian transgression at an elevation of 30-45 m near Bergen (Mangerud et al., 1981) is consistent with such a rise in the land. Around the northern end of the Gulf of Bothnia, in northern Finland, an undated interglacial marine fauna, has been found at an elevation of 200 m (Tynni, 1982). From the volume of sediment eroded from the Barents Sea considerable uplift and erosion has occurred in the Quaternary, although it is impossible to isolate any tectonic effects from the impacts of deep glacial scouring and consequent isostatic uplift.

7.3 North Africa - Eurasia Plate Boundary

The pole of relative rotation between Africa and Eurasia is estimated currently to lie at 20.2°N 23.5°W (Chase, 1978). This location is within a short distance of the western end of the plate boundary, allowing the plates to pivot and thereby avoid a significant length of compressional deformation in the oceanic lithosphere. This situation prevailed for much of the Tertiary.

Because the current pole of relative rotation of Africa-Eurasia is located close, and to the south, of the Azores-Gibraltar plate boundary, there is a marked change in present-day tectonics along this boundary: from extensional volcanism and WNW-ESE rifting in the Azores (which have developed since 7Ma), through dextral transcurrent motion along the 400 km long 80° trending Gloria Fault (Laughton et al., 1972), to large-scale overthrusting and sinistral strike-slip faulting on NE-SW faults in the oceanic crust and continental margin to the west of (and passing onland in) Portugal (Grimison and Chen, 1986).

7.3.1 *Deformation in Western Europe*

From the southwest corner of Iberia, through central western Europe, the regional Plio-Quaternary tectonics have been dominated by sinistral displacement on faults oriented NE-SW to NNE-SSW. This broad zone of deformation appears to reflect a response to the difficulty of forming a simple plate boundary through the western Mediterranean. In the Eastern Pyrenees some of the most pronounced transtensional basins along the east coast of Iberia occur where the coastal zones of rifting intersect the eastern Pyrenees. The changing stress regime in southern France during the Pliocene is indicated by the rotation of the orientation of dykes from NW-SE between 8 and 6Ma to N-S for those intruded between 3.4 and 0.7Ma (Feraud and Champreton, 1983).

The NNE-SSW trending Upper Rhine Graben also currently acts as a zone of sinistral shear. The thickest Quaternary basins (with sediment thicknesses in excess of 300 m) are found in two sections of the graben, where the bounding faults trending N-S (Illies and Greiner, 1978). Most of the active faults in the graben strike ca. 170°, oblique to the general graben axis.

7.3.2 *Southern North Sea*

Following a phase of stability in Miocene times, subsidence in the southern North Sea resumed in the late Pliocene, reflecting some process of active tectonic subsidence, not explicable through the long-term influence of Mesozoic rifting. In the Dutch part of the Central North Sea Graben, the ratio of Tertiary to Quaternary sediment thicknesses is 4:5, implying Quaternary subsidence rates around twenty times the Tertiary average (Kooi et al., 1989). There is evidence to suggest that this subsidence has accompanied extensional tectonic activity. The main axes of Quaternary North Sea subsidence follow the Broad Fourteens and Dutch Central Grabens. Originally, major rifts in the Triassic and Jurassic, these became the locus of significant inversion during the late Cretaceous. These inversions now appear to be collapsing in association with a broad transtensional 'lazy S-shaped' basin. Post-Lower Miocene extensional faulting is also distributed to the west, on several NW-SE trending faults in the central Southern North Sea, associated with underlying salt-movements (Oudmeyer and de Jager, 1993),

On the basis of sediment thicknesses, and assuming an extension phase from 2 to 1Ma, Kooi et al. (1991) estimated Beta values of up to 1.10 with values in excess of 1.03 across a width of at least 40 km, implying a cumulative extension across the southern North Sea of 2-3 km. These results were found to accord with Pliocene palaeobathymetry and gravity modelling, but appear to be exaggerated by almost an order of magnitude relative to the magnitude of accompanying faulting. Onshore, in the Roer Valley Graben (Lower Rhine Graben), there is a clear association between rifting and subsidence. Tertiary extension initiated in the early Oligocene (Zijerweld et al., 1992), requires a cumulative 6-10% crustal extension, implying a total horizontal extension of ca. 3 km, (equivalent to a long-term average of around 100 m per million years). This latter figure is consistent with rates of extension calculated from the individual fault displacements observed through the Quaternary (see Ahorner, 1962). To the northwest of the Roer Graben, wells show an additional extension of around 4%. Since about 3Ma, uplift has also dramatically increased on the southern flanks of the Roer Graben (Zijerweld et al., 1992). It is only in this period that extension has also occurred across the southern North Sea.

7.3.3 Uplift on the Flanks of the North Sea

Accompanying the subsidence of the North Sea there has been uplift of the eastern and western flanks, raising Pliocene marine sediments up to 200 m above sea-level in south east England and also raising the Neogene sediments that underlie much of Denmark. From the evidence of raised interglacial shorelines in England this uplift can be seen to have been proceeding until at least 200,000 year BP.

7.4 Fennoscandian Border Zone

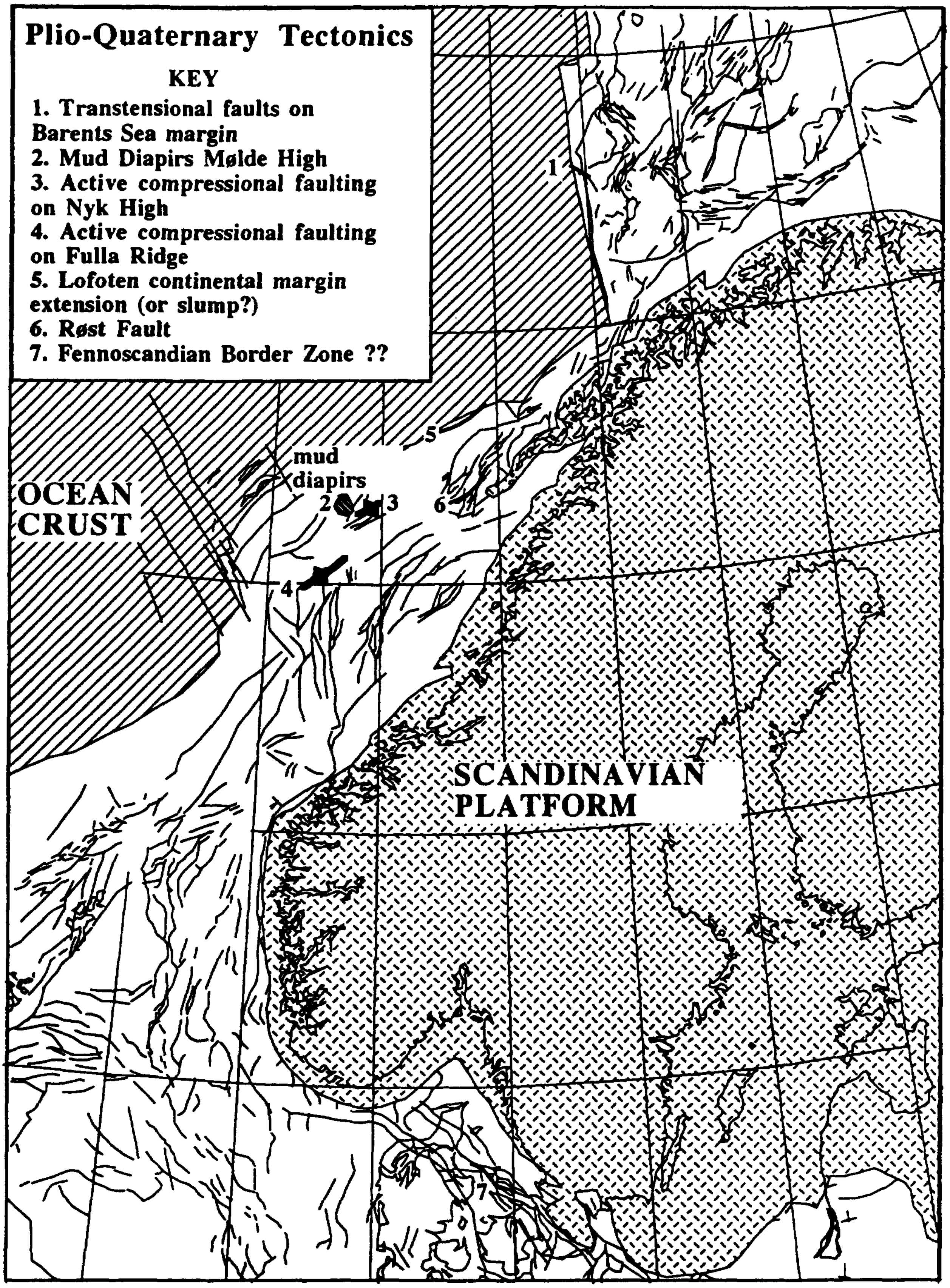
The form of the bedrock surface across parts of the Fennoscandian Border Zone both onshore and onland has suggested to a number of authors that there is evidence for some Quaternary extension along structures within part of the Fennoscandian Border Zone. This would be consistent with the style of tectonics seen across the southern North Sea. However, while there are depressions following faults of the FBZ it is not clear that any of them are incontrovertibly the product of fault displacement. Most recently Wannas and Floden (1993) have reviewed evidence from the Hanø Bay area offshore southern Sweden, in which they have claimed 20-60 m subsidence on the half graben during the Quaternary. Unfortunately the data for this area comprises old, low resolution single channel shallow seismic reflection profiles, which have a vertical to horizontal exaggeration typically of 20 x 1. In such a context trenches eroded by glacial scouring along the line of faults and fractures in the underlying bedrock and infilled during the post-glacial period, show up on seismic sections as tight synclinal folds. None of the seismic sections illustrated by Wannas and Floden is incontrovertibly demonstrative of neotectonic displacement and their estimate of 20-60 m displacement must be considered highly questionable.

7.5 Constraints on Plio-Quaternary Deformation in the Baltic Shield

The relaxation of compressional marginal deformation following the Miocene, suggests that the shield may also have undergone some reduction in horizontal NW-SE directed stresses during the Pliocene. However, as discussed in this chapter, there is evidence for some renewal of compressional tectonics along parts of the mid-Norway continental margin during the Pleistocene. This activity is however significantly less than that which prevailed during the Miocene both in rates of deformation and overall extent. Although it is clear that the fault movements themselves were triggered by deglaciation, the late-glacial fault ruptures of northern Fennoscandia may in part reflect some component of this renewed compressional strain-field accumulated and released within the shield. As at other periods, any deformation within the shield is likely to represent some small component of that concentrated in the weaker rifted basins.

In contrast to the deformation seen at a number of earlier periods, there is no evidence for ongoing compressional fault activity in seismic sections from the North Sea. In the central and southern Baltic Shield, apart from some suggestion of extension in the Fennoscandian Border Zone, there is no sign of active tectonic deformation passing into the shield.

The levels of strain internal to the shield implied by the Plio-Quaternary deformation seen on the continental margins appear significantly below those witnessed in the mid-Miocene. However rates of regional dome uplift may have been comparable and a tectonic strain-rate in the shield of 10^{-11} per year is realistic.



1) Transtensional Faults on Barents sea margin
2) Mud diapirs Mølde High
3) Active compressional Faulting on Mølde High
4) Active compressional Faulting on Fulla Ridge
5) Lofoten continental margin extension?
6) Røst Fault

Figure 7.1 Pleistocene tectonics around Fennoscandia

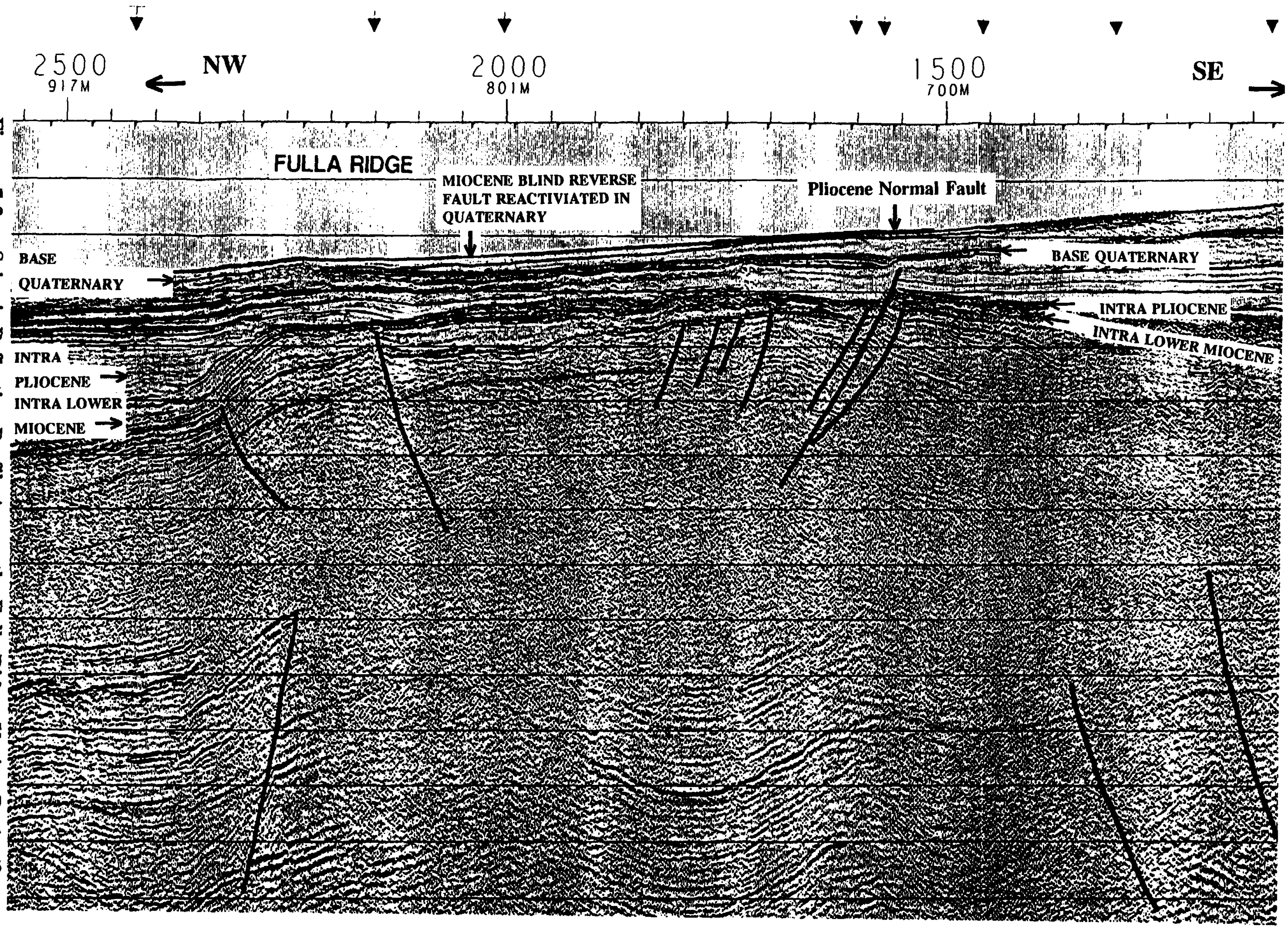


Figure 7.2 Seismic Reflection Profile Across the Fulla Ridge, Vøring Basin (from Granberg, 1992)

8.0 CONCLUSIONS

8.1 Tectonic Context

The tectonic history of Fennoscandia over the past one hundred million years has involved the transition from a location in the middle of a rifting super-continent through the creation of a new plate boundary to its present location on the continental margin of a new ocean. The first half of the period saw the build-up to the new plate boundary with as many as three premature attempts to create a continuous zone of rifting. The second half of the period saw the establishment of the new plate boundary, although with marked variations first in the orientation and later in the speed of spreading. Almost all the tectonic processes accompanying the emergence and activity of the new plate boundary have been concentrated to the west of Fennoscandia. Although the Fennoscandian Border Zone on the south-west edge of the craton, was a prominent zone of shear in the lead-up to the creation of the spreading ridge it has subsequently been relatively inert.

Throughout the past 100 million years deformation has at all periods been diverted around Fennoscandia. This can be seen from both the Fennoscandian Border Zone where deformation has consistently skirted the edge of the craton, as well as to the west where, even prior to the formation of the spreading ridge, rifting was concentrated in the present continental margin. These regions must be significantly weaker than the lithosphere of the shield and at all periods have tended to concentrate deformation. Such a difference in rheology is to be expected between the thick cold lithosphere of the craton and the thinner and warmer lithosphere of the rifted basins that lie to the south and west. The pattern of uplift of Fennoscandia during the Tertiary itself suggests that the cratonic lithosphere has controlled even mantle flow. Uplift is concentrated on the western flank of the lithospheric keel of Fennoscandia, where the mantle velocities are known to be relatively low, indicating low densities. The most plausible explanation for uplift in the late Tertiary is the movement of new lighter (hotter?) asthenosphere underneath the region in association with the re-invigoration of the Iceland hot-spot. In the absence of such a lithospheric keel, as beneath the northern North Sea, there was no equivalent permanent uplift.

8.2 Phases of Tectonics

There have been a number of distinct tectonic phases around Fennoscandia over the past 100Ma.

i) During the late Cretaceous and early Palaeocene in a series of distinct tectonic episodes dextral shear occurred along the Fennoscandian Border Zone. This shear involved left-stepping fault systems to negotiate the cratonic crust of southern Fennoscandia that created transpression. These episodes probably coincided with phases of rifting along the western margin. In the Palaeocene transtensional marginal rifting along the western continental margin of Scandinavia was again accompanied by transpressional movement along the Fennoscandian Border Zone.

ii) In the early Eocene, volcanic activity and rifting occurred along the present continental margin of Western Scandinavia to create overthicked oceanic crust, and underplated margins of sedimentary basins. These remained elevated, long after spreading had been initiated.

iii) Towards the end of the Eocene spreading directions in the North Atlantic began to rotate anticlockwise. This locked the Iceland-Faeroes fracture-zone, caused a new spreading axis to be initiated along the coast of Greenland and forced the Jan Mayen block to become rifted and rotated. Deformation was also transmitted into North-West Europe and sub-plate boundary zones of interconnection short-circuited between the North Atlantic and Africa-Eurasia plate boundaries. In the North Sea NE-SW to E-W compression was initiated, causing widespread minor inversion of some pre-existing normal faults. Further south, in the Rhine Graben E-W extension led to rifting that eventually spread through western Britain in the late Oligocene.

iv) In the mid-late Miocene, foreland compressional deformation and regional uplift accompanying the emergence of the Iceland hot-spot led to compressional deformation all around the eastern continental margin of the North Atlantic facing the hot-spot.

v) In the late Pliocene and early Pleistocene rapid subsidence accompanied by mild extension occurred across both the mid-Norway continental margin and across the southern North Sea. Mild compression has now been renewed on the Norwegian continental margin.

8.3 100Ma Stress-History of Sweden

Although the evidence of deformation comes from the margins of the shield, current tectonic stress-fields are known to be regionally consistent and to be influenced by mantle motions. It can be expected that throughout the past 100 million years the stress-field prevailing within Sweden is likely to be closely consistent with that prevailing in the surrounding region. Hence, the Swedish stress field can be reconstructed from the evidence of tectonic deformation around the Fennoscandian margins.

Reconstructing stress-fields from tectonic evidence can never be exact, principally because the exact motion vector of faults (and in particular the significance of an oblique slip component of displacement) may not be resolved. However, the style of faulting is immediately indicative of the relative magnitudes of the principal stress tensors and where there have been a range of available fault orientations, the particular structures subject to reactivation are likely to be sensitive to the orientation of the stress field.

8.3.1 Stress-history

In the late Cretaceous through to the Palaeocene, although there were separate phases of tectonic activity implying periods of higher stress, and accompanying strain concentration, the disposition of deformation implies that the major horizontal stress direction was oriented close to NNE-SSW, through the period (see Figure 8.1). From the strike of basin formation adjacent to inversion, as seen along the Fennoscandian Border Zone, the minimum stress was oriented close to WNW-ESE. To the west of Scandinavia there is less evidence of transpression through this period and a greater predominance of rifting.

A similar orientation persisted along the Fennoscandian Border Zone through the Palaeocene. Along the Norwegian continental margin the dominance of extension implies that the maximum horizontal stress magnitude was close, or even below, that of the vertical. To the south of the Faeroes-Shetland Basin dextral transpression along WSW-ENE trending faults implies a NNW-SSE orientation of the principal horizontal compressive stress direction that is itself consistent with the orientation of the Hebridean dyke-swarms. Hence, there appears to be some stress-rotation at this period perhaps chiefly caused by the impact of the Hebridean mantle plume. In effect the axis of the plume was in extension while the flanks may have been subject to some marginal compression (or stress re-orientation).

The first significant deformation to be found in the North Sea subsequent to ocean spreading indicates a stress-field that shows some rotation from almost N-S in the southern North Sea to more nearly WSW-ENE in the northern North Sea. This stress-field, which appears to fan around Western Fennoscandia appears to be a response to the interaction between the various plate boundary forces:- ridge-push spreading in the Norwegian Sea compression radiating from the blocked Iceland-Faeroes Fracture Zone, and a N-S component of shortening emerging off the African-Eurasia plate boundary through the Alps.

By the late Oligocene this radiating stress field persisted, but with compressional deformation appearing along the Norwegian continental margin accompanied by E-W extension in western Britain. During the late Mid-Miocene tectonic crisis the northern stress-field, first found along the West Scandinavian continental margin, came to dominate the whole of north west Europe, and compressional deformation was found along the eastern margin of the northern Atlantic, radiating around Iceland. Compression also passed deep into western Europe, ending rifting and subsidence in the Rhinen Graben and leading to the Western Alpine collision zone. It was at this period when a NW-SE orientation of the major horizontal stress direction prevailed across the whole of western Europe. In the late Pliocene and Quaternary renewed extension has taken over from compression both along the Rhine Graben and around the southern North Sea. However, compressional deformation has been renewed along parts of the Norwegian continental margin.

8.3.2 *Stress magnitudes*

Both the orientation and magnitudes of the tectonic stress-field have varied through time, the magnitude of horizontal stress in Scandinavia was at its highest at the time of most rapid compressional deformation on the western continental margin in the mid-Miocene. The lowest NW-SE horizontal stresses are likely to have been in the late Cretaceous and Palaeocene prior to the opening of the North Atlantic. At such time the NE-SW stress was higher than the NW-SE stress. Hence, the greatest changes in the past 100 million years have occurred in the magnitude of the NW-SE stress, and consequently it is faults and fractures lying orthogonal to this direction that will have shown the greatest changes in their properties.

8.4 The Age of the Current Tectonic Regime

For the purposes of determining evidence of tectonics of direct relevance to present-day seismotectonics, the concept of the current tectonic regime (CTR) has been proposed, as discussed in Muir Wood and Mallard, (1992). The age of the current tectonic regime becomes of great importance in determining whether past fault displacement is, or is not, a manifestation of ongoing fault activity. From the evidence of the changing pattern of Tertiary tectonics, it is possible to estimate when the CTR became established.

Across north west Europe, the period of intense mid-Tertiary tectonics ended in mid to late Miocene times with the re-emergence of the Iceland hot-spot, the acceleration of Atlantic opening, and the collision of Apulia with Europe in the western Alps. From palaeo-stress observations, it is clear that the current orientation of the maximum horizontal stress direction dates from around the end of the Oligocene culminating in the middle of the Miocene (12-10Ma), being closely related to the change in relative Africa-Eurasia motion at about this time. However, from the Norwegian continental shelf it is apparent that tectonic activity has itself become modified through the subsequent period. Although compressional structures largely ceased their activity by the early Pliocene, in the Quaternary NW-SE compressional stresses have increased again. Hence, the CTR is probably less than 2 million years old, although the prevailing stress-regime has showed some consistency for about 25Ma.

8.5 Controls on Tertiary Deformation

Deformation rates can only be refined where there is the opportunity for detailed dating throughout an individual episode. For example, the Palaeocene phase of Hebridean magmatism almost all took place within a single episode of reversed magnetism, thereby constraining the rates of crustal dilation to greater than ca.0.5mm/yr. As another example, folding on the Norwegian margin can be seen to have involved long-term growth, followed by a rapid phase of mid-Miocene activity. From seismic evidence, it appears that, throughout the major episodes of Tertiary deformation (40-25Ma), rates of displacement have typically been less than

100m per million years (10^{-1}mm/yr), and more often tens of metres per million years (10^{-2} mm/yr).

8.5.1 *Deformation within Sweden*

Although it is evident that deformation has dominantly occurred around the margins of Fennoscandia it is not possible to completely discount the possibility that some deformation also occurred within the shield during the past 100 million years. However, some important constraints can be placed on such internal tectonics. Nowhere is deformation seen to emerge from the shield, but instead can be shown to lie parallel with its borders (as for example along the Fennoscandian Border Zone). The manifestation of tectonic activity along the margin is indicative of the degree to which the warmer and thinner lithosphere of the sedimentary basins is of weaker rheology.

Two most intense periods of activity adjacent to Scandinavia were in the Middle Miocene and in the Late Cretaceous and Palaeocene. It is at these periods, if any, that some internal deformation might be predicted. Along the Fennoscandian Border Zone there is however no evidence to suggest any major zone of antithetic NE-SW strike-slip faulting passing into southern Sweden. However, some deformation may have passed into the Mesozoic rifted basins originally located along the stretched Caledonian crust of northern Norway.

8.5.2 *Speculative models of Cenozoic tectonics in Sweden*

A number of speculative models concerning the Cenozoic tectonics of Sweden and the Baltic Shield have appeared in the literature. While the absence of dated fault displacements from within the shield makes it difficult to refute one or other conceptualisation it is possible to test all such models according to the degree to which they match with the well-documented evidence of the Fennoscandian continental margins.

Mörner (1979) (and many other papers) proposed that the Fennoscandian Shield was 500m higher from about 100Ma to 22.5Ma, and some 1100-1200m higher from 22.5Ma to 0.9Ma, after which a considerable subsidence set in. The uplift and subsidence were considered to follow elliptical fields that closely resembled the geometry of the postglacial isostatic uplift. Mörner also proposed that the 'true' postglacial isostatic rebound died out 4500 years ago to be replaced by some long-term tectonic process. These ideas appear to be without any foundation: Cenozoic uplift events of the Norwegian continental margin and Western Scandinavia clearly do not have a simple spatial relationship with postglacial rebound, and occurred over extended time-periods. Modern lithosphere-asthenosphere models of postglacial rebound find good agreement with the observed uplift history of Fennoscandia without any need for additional 'tectonic' support.

Talbot and Slunga (1989) claimed that large-scale strike-slip duplexes (with local pop-up horsts and pull-apart basins) define NW trending corridors across Fennoscandia between offsets in the mid-Atlantic and Mediterranean - Himalayan plate boundaries. Strike-slip duplexes in north and south Fennoscandia are interpreted as extrapolations of movement on the Jan Mayen and Iceland transforms, passing into the continental marginal through the past 38Ma. Although nowhere stated, the implicit scale of the structures implies displacements during the Cenozoic must have been measured in kilometres. The main argument in support of this model appears to be the apparent lensoid configurations (on a scale of hundreds of kilometres) of the major fracture zones mapped in the shield. However, neither the evidence from within the shield nor beneath the continental margin lend support to this model. It is clear from the deformation all around the western margin of Fennoscandia prior to the opening of the North Atlantic that displacement skirted around the Baltic Shield and that the Fennoscandian Border Zone marks the most northerly of the NW-trending corridors of deformation interconnecting the plate boundaries. The major transform faults of the North Atlantic do not show significant post-Eocene displacement passing into the continental margin and major duplexes within the shield would anyway leave a legacy of uplifted and downwarped blocks (and sediment filled pull-apart basins) on a scale commensurate with any horizontal displacement.

8.5.3 Constraints on Cenozoic fault displacement rates in Sweden

Throughout the various sections of this report comments have been made on the likely strain-rates internal to the Baltic Shield implied by different tectonic episodes. The use of the strain-rate concept is intentional because it avoids the problem of having to define by how much individual faults within the shield may have suffered displacement in these various periods.

In terms of where regional strain has been concentrated on individual faults within the Baltic Shield one could propose two divergent models.

In the first model, within each tectonic episode all the deformation across the shield is considered to have been concentrated on a single key fault. If a tectonic phase is considered to have lasted approximately 10 million years, then a strain rate of 10^{-11}/yr would equate to a cumulative strain of 10^{-4}. Concentrating the strain distributed over a distance of 500km this fault would have moved by 50m over a ten million year period, equivalent to a slip-rate of 0.005mm/yr. Converted into seismicity this might be a magnitude 7 (2.5m displacement) earthquake every 500,000 years along each 50km length of the fault, perhaps equivalent to a return period of a magnitude 7 earthquake somewhere in Sweden of about 50,000 years.

In the second model the strain is distributed across a large number of equivalent structures. Suppose the typical separation of Baltic Shield faults most prone to reactivation in a given tectonic episode is 10km. Over ten million years a strain rate of 10^{-11}/yr is equivalent to a cumulative displacement on each of these faults of lm, and a slip-rate of 10-4mm/yr.

The return period of a magnitude 7 earthquake on these faults would be many millions of years, although the overall level of activity throughout Sweden would be much the same as in the first strain localisation model.

Models of this type can provide important constraints on what can be inferred about the long-term tectonic behaviour of any specific fault encountered in a site investigation. From knowledge of the orientation of individual faults encountered within the Baltic Shield, and their relation with contemporary active faults of comparable orientation identified on the continental margins, there is the potential to assess which faults are most likely to have been active at different periods. In the late Cenozoic the orientations most likely to have been reactivated are NNE-SSW trending dip-slip structures, consistent with those found in northern Fennoscandia to have been reactivated during deglaciation.

8.5.4 *Tectonic Analogues*

One way in which to estimate the significance of deformation within the shield is to study regions which today have a similar tectonic context to that which prevailed in Fennoscandia during the Palaeocene or Mid-Miocene. Along the southern margin of the cratonic crust of Russian shield deformation related to the present Alpine-collision zone is concentrated in Iran, where there is both transpressional and compressional deformation of styles comparable to those found along the margins of Fennoscandian in the past 100 million years, although rates and magnitudes of deformation are significantly higher in Iran. From maps of current seismicity almost no activity occurs to the north of the zone of deformation, along the Kopet Dagh mountains of north east Iran (McKenzie and Jackson, 1984), indicating that the boundary of the shield is itself a barrier to the transfer of strain.

8.6 Future Tectonics

It is, of course, impossible to be completely certain about the future tectonic state of Fennoscandia. However, one can make a number of predictions according to different time-horizons, based both on the style of changes that have occurred over the past 100 million years and also from tectonic analogues in other regions of the world.

8.6.1 *The Next Million Years*

The loading and removal of ice-sheets are likely to dominate the strain-state of Fennoscandia over the next million years. However, after isostatic recovery is completed, once a tectonic strain-field is restored, it is likely that the evidence for an increase in horizontal stress along the western margin of mid-Norway may become of greater importance in Sweden. The late-glacial faults of northern Fennoscandia may owe part of their existence to this increase in horizontal stress. There is a possibility of changes in the current activity of the continental margin even over a one million year time-period. This could theoretically expand to include the number of structures and rates of displacement observed in the mid-Miocene, although there is no evidence for any enhancement in mid-Atlantic spreading of Icelandic plume activity to cause such changes. Alternatively, the deformation could decay in significance as horizontal stresses decline.

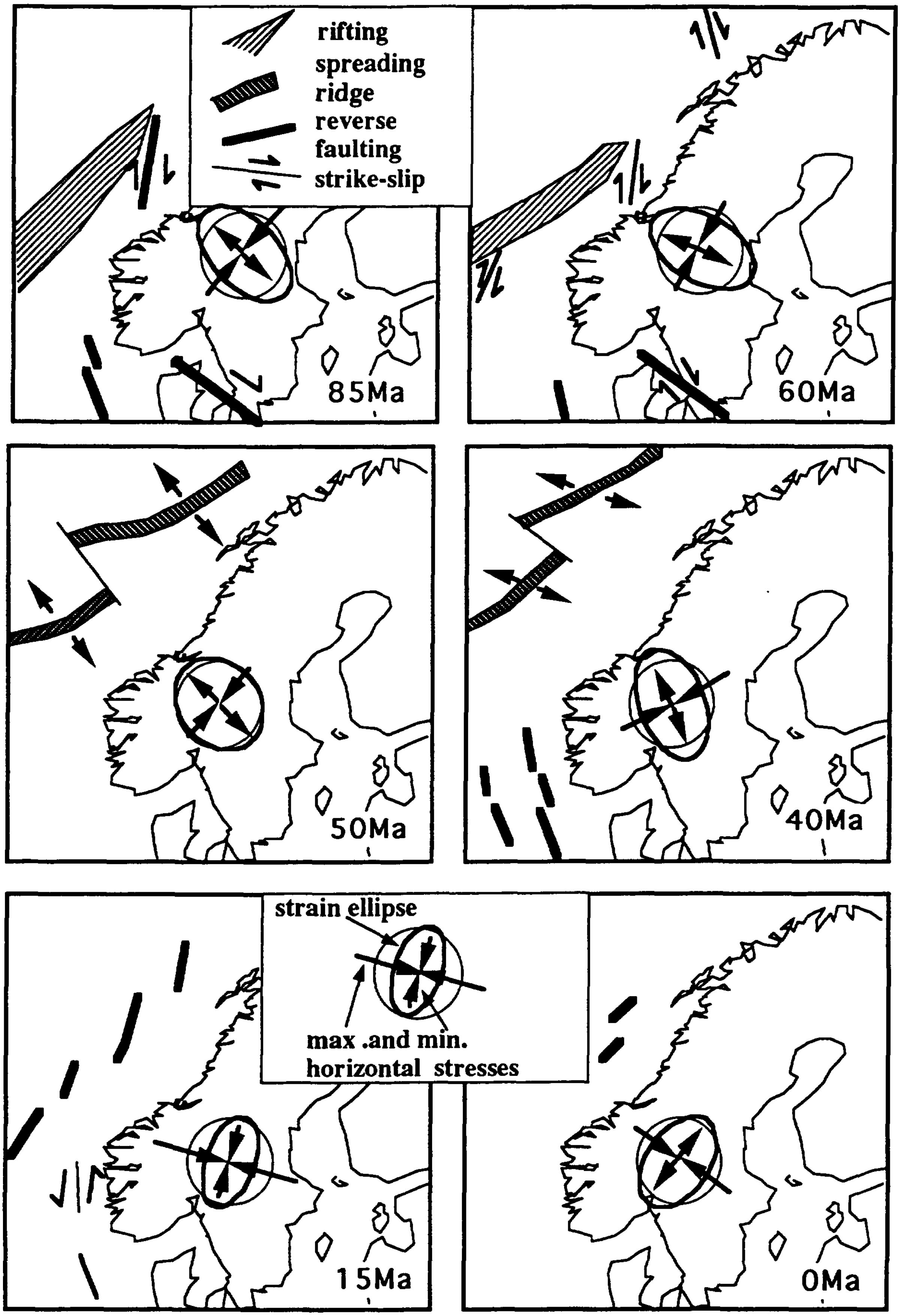
8.6.2 The Next Ten Million Years

The Africa-Eurasia plate boundary through the Mediterranean has been highly dynamic throughout the past 100 million years. At present there is some suggestion that compressional deformation is becoming transferred to the north of the western Mediterranean, as along the Ligurian continental margin. However, based on the past 50 million years it is highly unlikely that a reconfiguration of the Alpine plate boundary will impact on Sweden.

Over a ten million year time-period it is not unlikely that the Iceland plume might decline in significance, switch off, or move off the spreading ridge, to become a chain of islands like the Azores. The implications of such changes on stresses on the continental margin are difficult to predict. It is also possible that the extension seen in the southern North Sea might expand to encompass the Fennoscandian Border Zone. This could lead to low-level normal faulting (of the rates now observed in southern Holland of around 0.1mm/yr) to occur along the south-east corner of Sweden.

8.6.3 The Next One Hundred Million Years

Within the next one hundred million years it is almost inevitable that a subduction zone will develop along the continental margin of western Fennoscandia. This subduction zone is likely to be initiated further to the south, where the oceanic lithosphere is older and denser. Lying on the eastern margin of the ocean such a subduction zone is likely to be of Andean type, with volcanoes emerging through the edge of the pre-existing continent. A model for the development of this area could perhaps be the Chonos Islands in southern Chile, where a Palaeozoic orogenic belt, cut through by fjords and interspersed by a chain of volcanoes rising to more than 2500m, is now underlain by a subduction zone capable of great earthquakes. This may be the future for mid Norway 100 million years into the future. Fortunately, as volcanoes only emerge off subduction zones when the subducting (and dipping) oceanic slab has reached depths of 100-150km, (for a $<$4cm/yr plate convergence) there would typically be a window of at least 5 million years before the first volcano appeared in Trondheim.



'Arrows indicate approximate relative magnitude of principal stresses'

Figure 8.1 Changing stress regime in Sweden over past 100 Ma

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APPENDIX A

Gazetteer of Placenames and Structural Features Mentioned in the Report

ASR Aegir Spreading Ridge (see Fig 5.1)
AG Ålborg Graben (see Fig 1.2)
AF Alpine Foreland (the region to the north of the Alps
AEG Arne Elin Graben (see Fig 1.2)
AET Arne Elin Trend (see Fig 1.2)
Alpine Chain (the main chain of mountains through central Europe)
Andøya (island at 69°N off northern Norway)
Anglo-Brabant Platform (Stable region encompassing central England and Belgium)
Antrim (Palaeocene volcanic province in Northern Ireland)
Apulian promontory (The Italian microplate)
Arctic Ocean Spreading Ridge (see Fig 5.1)
Arran Volcano (Palaeocene volcanic island W of Glasgow, Scotland)

BaFi Balder Field (see Fig 1.2)
BeF Beatrice Field (see Fig 1.2)
BjB Bjornoya Basin (see Fig 1.2)
BFC Bjornoyrenna Fault Complex (see Fig 1.2)
BFZ Borglum Fault Zone (see Fig 1.2)
BrG Bresse Graben (in SW France)
BFB Broad Fourteens Basin (see Fig 1.2)
Bjornøya (island midway between Norway and Svalbard)
Black Forest Volcanic Province (Late Tertiary volcanic complex to east of Rhine Graben)
Blackstones Basin (Palaeocene volcanic complex in southern Hebrides Sea, W Scotland)
Blosseville Peninsula (E. Greenland coast around 70°N)
Bohemian Graben (Area of late-Tertiary subsidence and volcanism in the Czech Republic)
Bovey Tracey Basin (10km long NW-SE trending mid-Tertiary pull-apart basin in Devon, England)
Brendan Centre (Major Tertiary igneous centre at northern end of Porcupine Basin to the west of Ireland)

CG Central Graben (see Fig 1.2)
CF Clair Field (see Fig 1.2)
CR Clair Ridge (see Fig 1.2)
CTZ Clair Transfer Zone (see Fig 1.2)
CSF Coffee Soil Fault (see Fig 1.2)
Canna Basin (Small Mid-Tertiary basin to the SW of Skye, Hebrides)
Cardigan Bay Basin (Mid-Tertoary basin to the west of Wales)
Charlie Gibbs Fracture Zone (see Fig 5.1)
Clare Lineament (E-W former transform fault at southern end of Rockall Trough).
Cleveland Hills (E. Yorks, England)
Connemara (W. Ireland)

DCT Danish Central Trough (see Fig 1.2)
DCG Dutch Central Graben (Mesozoic Graben north of the Netherlands)
DGZ Geer Zone (N-S transcurrent fault system passing through western Svalbard)
DFZ Dowsing Fault Zone (main NW-SE fault-system through Sole Pit Basin)

ESB East Shetland Basin (see Fig 1.2)
ES Egersund Sub-basin (see Fig 1.2)
EC Erlend Centre (see Fig 1.2)
ETZ Erlend Transfer Zone (see Fig 1.2)
East Shetland Platform (see Shetland Platform)
Edda Field (see Ekofisk)
Ekofisk Field (see Fig 1.2)
Emerald Field (see Fig 1.2)

FSE Faeroes-Shetland Escarpment (see Fig 1.2)
FSB Faeroes-Shetland Basin (Trough) (see Fig 1.2)
FBZ Fennoscandian Border Zone (see Fig 1.2)
FG Fenris Graben (see Fig 1.2)
FF Fjerritslev Fault (see Fig 1.2)
FGS Fladen Ground Spur (see Fig 1.2)
F Fles Fault (see Fig 1.2)
FR Fulla Ridge (see Fig 1.2)
Faeroes Basin (see Faeroes-Shetland Basin)
Faeroes Platform (Raised non-oceanic area around Faeroes)
Faeroes- Rockall Platform (Raised non-oceanic area encompassing both Faeroes and Rockall)
Forties Field (see Fig 1.2)
Frigg Field (see Fig 1.2)

GHFZ Grena-Helsingør Fault Zone (see Fig 1.2)
Geikie Escarpment (on eastern margin of northern Rockall Trough)
Geikie Igneous Centre (on eastern margin of northern Rockall Trough)
German Molasse Basin (Continental basin filled with sediment eroded off the rising Alps in the Tertiary)
Gloria Fault (current major E-W transform fault between Portugal and the Azores)
Gorm Field (see Tyra Field)
Great Glen Fault (Major NE-SW former transcurrent through northern Scotland)
Greenland Fracture Zone (see Fig 5.1)

HT Halten Terrace (see Fig 1.2)
HB Hammerfest Basin (see Fig 1.2)
HBG Hanø Bay Graben (see Fig 1.2)
HaB Harstad Basin (see Fig 1.2)
HFZ Hornsund Fault Zone (see Fig 1.2)
Hebridean Escarpment (Continental margin to west of Hebrides)
Hebridean Volcanic Province (Palaeocene igneous prvince of western Scotland)
Helmsdale Fault (NE-SW fault parallel and to the west of Great Glen Fault)
Helvetic Shelf (shallow sea formerly on northern margin of Swiss Alps)
Hessen Depression (former northern continuation of main Rhine Graben).

IVP Inner Voring Plateau (see Fig 1.2)
Iceland-Faeroes Fracture Zone (see Fig. 5.1)
Isle of Wight Monocline (Mid-Tertiary E-W monocline on island offshore southern England)

Jameson Land (E. Greenland)
Jan Mayen Fracture Zone (see Fig 5.1)
Jan Mayen Microplate (see Fig 5.1)
Jan Mayen Plateau (Area of non-oceanic water depths, extending to the south of Jan Mayen)

Kap Dalton (E. Greenland)
Kap Brewster (E. Greenland)
Kejser Josef's Fjord (E. Greenland)
Kish Bank Basin (Small Mesozoic basin in Irish Sea to east of Dublin)
Knipovich Ridge (see Fig 5.1)
Kolbeinsey Ridge (see Fig 5.1)

LR Lindesnes Ridge (see Fig 1.2)
LCH Linerodasen-Christiansø Horst (see Fig 1.2)
LM Lofoten Margin (see Fig 1.2)
LB Lofoten Basin (see Fig 1.2)
LFZ Lofoten Fracture Zone (see Fig 1.2)
LH Loppa High (see Fig 1.2)
Labrador Sea Spreading Ridge (See Fig 5.1)
Lough Neagh Basin (Mid-Tertiary continental basin beneath Lough Neagh, northern Ireland)
Lundy (island in Bristol Channe, west of Devon, England)

MaB Maud Basin (see Fig 1.2)
MH Mølde High (see Fig 1.2)
MFB Moray Firth Basin (see Fig 1.2)
MrB Møre Basin (see Fig 1.2)
MM Møre Margin (see Fig 1.2)
MNB Mid-Norway Basin (see Fig 1.2)
Magnus Transfer Zone (NW-SE transfer fault across Faeroes-Shetland Basin)
Massif Central (Central France)
Molasse Basin (Region of Tertiary continental sedimentation to north of Alps)
Møre Platform Escarpment (see Fig 1.2)
Mull (Inner Hebridean island, west Scotland)

NF Navlinge Flexure (see Fig 1.2)
NRSZ Nordland Ridge Shear Zone (see Fig 1.2)
NS Nyk Syncline (see Fig 1.2)
Nelson Field (see Forties Field)
Neuwied Basin (Small late-Tertiary basin on SE continuation of Roer Basin in Rhenish Massif, Germany)
North Celtic Sea Basin (Mesozoic basin south of Ireland)
North Pyreneean Fault Zone (Main plate boundary fault-system parallel to Pyrenees)

OVP Outer Voring Plateau (see Fig 1.2)
Outer Isles Fault System (major NNE-SSW trending fault on inner edge of

PT Polish Trough (see Fig 1.2)
Peel Horst (Uplifted block within Roer Valley Graben, Netherlands)

Porcupine (Seabight) Basin (N-S Mesozoic basin, west of Ireland)
Porcupine Igneous Province (Region of Palaeocene igneous activity around the Porcupine Basin)
Pyreneean Chain (Mountains formed along former plate boundary separating France from Iberia)

RLFC Ringvassoy-Loppa Fault Complex (see Fig 1.2)
RP Rockall Platform (area of sunken continental crust surrounding Rockall)
RR Rona Ridge (see Fig 6.3)
RG Rønne Graben (see Fig 1.2)
RH Røst High (see Fig 1.2)
RFH Rynkobing Fyn High (see Fig 1.2)
Reykjanes Spreading Ridge (see Fig 5.1)
Rhenish Massif (Variscan age - 300Ma - deformation zone, on border between Germany and Belgium, uplifted in the late Tertiary)
Rhine Graben (Major Tertiary rift valley passing NNE-SSW along border between France and Germany)
Rhum (Hebridean island)
Roar Field (see Fig 1.2)
Rockall Bank (see Rockall Platform)
Rockall Trough (Deep-water trough, probably floored with oceanic crust, separating the Rockall Roer Valley Graben (NW-SE trending rifted basin on border between Germany and the Netherlands)

SSB Scarvister sub-basin (see Fig 1.2)
SM Senja Margin (see Fig 1.2)
SR Senja Ridge (see Fig 1.2)
SSFS Shetland Spine Fault System (see Fig 1.2)
SV Skaggerak Volcano (see Fig 1.2)
SlF Sleipner Field (see Fig 1.2)
SrB Sørvestsnaget Basin (see Fig 1.2)
SWBB South West Barents Sea Basin (see Fig 1.2)
SH Stappen High (see Fig 1.2)
StB Stord Basin (see Fig 1.2)
SD Svalis Dome (see Fig 1.2)
Scotland-Greenland Ridge (Area of shallower water connecting Scotland and Greenland via Iceland)
Scotland - Shetland Platform (Shallow water geologically stable are connecting Scotland wth the Shetlands)
Senja Fracture Zone (see Fig 5.1)
Skye (Hebridean island)
Skjold Field (see Tyra Field)
Platform from Great Britain)
Slyne Fissures (ENE-WSW trending dyke-swarm at northern end of Porcupine Basin)
Snaefellsnes (peninsula with active volcano in W. Iceland)
Sogne Basin (see Fig 1.2)
Solan Basin (southern end of Faeroes-Shetland Basin)
Solan Fault (southern end of Faeroes-Shetland Basin)
Sole Pit Basin (NW-SE Mesozoic North Sea basin offshore Eastern England)

South Harris Basin (Small mid-Tertiary Basin on Outer Isles Fault)
South West Approaches Basin (Mesozoic basin on continental margin to west of English Channel)
South West Bohemian Basement Spur (Continental promontory to north-east of the Alps)
St Kilda (island to the west of Outer Hebrides)
Storegga Slide (major underwater Holocene landslide on Møre-Romsdal continental margin)
Swabian Volcano (late Tertiary volcano in southern Germany)

TEG Tail End Graben (see Fig 1.2)
TF Traenabanken Field (see Fig 1.2)
TrF Troll Field (see Fig 1.2)
TB Tromso Basin (see Fig 1.2)
TP Trondelag Platform (see Fig 1.2)
TyF Tyra Field (see Fig 1.2)
Tor Field (see Ekofisk)

UR Utrøst Ridge (see Fig 6.3)
UH Utsira High (see Fig 1.2)

VF Valhall Field (see Fig 1.2)
VD Vema Dome (see Fig 6.3)
VVP Vestbakken Volcanic Province (see Fig 1.2)
VBV Vest Brona Volcano (see Fig 1.2)
VG Viking Graben (see Fig 1.2)
VB Vøring Basin (see Fig 1.2)
VMH Vøring Marginal High (see Fig 1.2)
VPE Vøring Plateau Escarpment (see Fig 1.2)
Vogelberg Volcano (Late Tertiary volcano in southern Rhine Graben)

WSB West Shetland Basin (see Fig 1.2)
MVGF West Viking Graben Boundary Fault (see Fig 1.2)
WGG Witch Ground Graben (see Fig 1.2)
WTR Wyville Thomson Ridge (shallow-water ridge passing to the north-west from the Outer Hebrides)
WTRC Wyville Thomson Ridge Complex (see Fig 1.2)
Weald Basin (E-W Mesozoic Basin in Sussex, and Kent, southern England)
Wessex Basin (E-W Mesozoic Basin in Dorset, Wiltshire and Hampshire, southern England)
West Lewis Basin (Mesozoic basin to west of Lewis, Outer Hebrides)
West Netherlands Basin (Mesozoic Basin passing offshore into southern North Sea)
West Shetland Margin (continental margin to west of Shetland)

Ymir Ridge (Major mid-Tertiary compressional structure adjacent to Wyville Thomson Ridge)

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Biotite and chlorite weathering at 25°C. The dependence of pH and (bi) carbonate on weathering kinetics, dissolution stoichiometry, and solubility; and the relation to redox conditions in granitic aquifers

Maria Malmström[1], Steven Banwart[1], Lara Duro[2], Paul Wersin[3], Jordi Bruno[3]
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January 1995

TR 95-02

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Lars Werme, Joachim Eriksson
Swedish Nuclear Fuel and Waste Management Co, Stockholm, Sweden
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Marie Wiborgh (ed.)
Kemakta Konsult AB, Stockholm, Sweden
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Groundwater sampling and chemical characterisation of the Laxemar deep borehole KLX02

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Bill Wallin
Geokema AB, Lidingö, Sweden
March, 1995

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Äspö Hard Rock Laboratory Annual Report 1994

SKB
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Feasibility study for siting of a deep repository within the Storuman municipality

Swedish Nuclear Fuel and Waste Management Co., Stockholm
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Ignasi Puigdomènech[1], Jordi Bruno[2]
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April 1995

TR 95-10

Investigations of subterranean microorganisms. Their importance for performance assessment of radioactive waste disposal

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Luis Moreno, Björn Gylling, Ivars Neretnieks
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June 1995

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Yvonne Ohlsson, Ivars Neretnieks
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Ove Landström[1], Eva-Lena Tullborg[2]
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Roland Pusch
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Paul R La Pointe[1], Peter Wallmann[1], Sven Follin[2]
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Temperature conditions in the SKB study sites
Kaj Ahlbom[1], Olle Olsson[1], Stefan Sehlstedt[2]
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Measurements of colloid concentrations in the fracture zone, Äspö Hard Rock Laboratory, Sweden
Anna Ledin, Anders Düker, Stefan Karlsson, Bert Allard
Department of Water and Environmental Studies,Linköping University, Linköping, Sweden
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Thermal evidence of caledonide foreland, molasse sedimentation in Fennoscandia
Eva-Lena Tullborg[1], Sven Åke Larsson[1], Lennart Björklund[1], Lennart Samuelsson[2], Jimmy Stigh[1]
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Clay Technology AB, Lund, Sweden
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Modelling of the physical behaviour of water saturated clay barriers. Laboratory tests, material models and finite element application
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Björn Lagerblad, Jan Trägårdh
Swedish Cement and Concrete Research Institute
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Kristina Skagius[1], Anders Ström[2], Marie Wiborgh[1]
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Roy Forsyth
Caledon Consult AB
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Marcus Laaksoharju[1], Claude Degueldre[2], Christina Skårman[1]
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Marcus Laaksoharju (ed.)
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Steven Banwart (ed.)
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DECOVALEX I – Bench-Mark Test 3: Thermo-hydro-mechanical modelling
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DECOVALEX I – Test Case 1: Coupled stress-flow model
Lars Rosengren[1], Mark Christianson[2]
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TR 95-32
Partitioning and transmutation (P&T) 1995. A review of the current state of the art
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Geohydrological simulation of a deep coastal repository
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