Seismic Stratigraphy and Distribution of Palaeogene Sediments West and East of Shetland

Patrick J. Condon

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DECLARATION

This thesis has been composed solely by myself. The work presented is my own unless otherwise acknowledged.

Patrick J. Condon
CONTENTS

ABSTRACT 1

ACKNOWLEDGEMENTS 3

CHAPTER 1 INTRODUCTION 5

1.1 Objectives 5
1.2 Structural setting and nomenclature 7
1.3 The database 11
1.4 Principles of seismic stratigraphy 13
1.5 Seismic stratigraphy and sea level changes 19
1.6 The control of basin forming mechanisms on post-rift subsidence 28

CHAPTER 2 THE EOCENE SUCCESSION ON THE EAST SHETLAND PLATFORM 29

2.1 Introduction 29
2.2 Location and seismic configuration of boreholes 29
2.3 ?Lower to early Middle Eocene sedimentation 37
2.4 Mid to late Middle Eocene sedimentation 42
2.5 Oligocene to Miocene 52
2.6 Depositional history of the East Shetland Platform marginal sequence 54
2.7 Reworking in BGS 80/03, 81/16 and 81/17 61
2.8 Correlation to the Hampshire Basin and Haq et al. (1987) sea level curve 64
2.9 Conclusions 68

CHAPTER 3 SEISMIC STRATIGRAPHY OF THE EAST SHETLAND PLATFORM AND BASIN 70

3.1 Introduction 70
CHAPTER 4 EOCENE SEISMIC STRATIGRAPHY WEST OF THE SHETLAND PLATFORM

4.1 Introduction 131
4.2 Correlation to the East Shetland succession 132
4.3 Comparison between log characteristics of the southern Faeroe Basin 137
4.4 Seismic stratigraphy of the southern Faeroe Basin 140
4.5 Depositional history of the Eocene succession 156
4.6 Tectonism and seismic stratigraphy in the north-eastern Faeroe Basin 160
4.7 Role of fault activity during the Eocene 167
4.8 The role of Rona Ridge in Eocene sedimentation 174
4.9 Summary and sea level curve for west of Shetland 180

CHAPTER 5 TERTIARY TECTONIC ACTIVITY AND THE EVOLUTION OF THE NE ATLANTIC

5.1 Introduction 186
5.2 Cretaceous development of the North-East Atlantic 189
5.3 Palaeocene - tectonism before crustal separation 195
5.4 Pre-rifting volcanism on the Atlantic margins 198
5.5 Lowermost Eocene tectonism and the initiation of Atlantic seafloor spreading 202
5.6 Eocene sedimentation and tectonics - passive margin evolution 206
5.7 Summary

CHAPTER 6 THEORETICAL MODELLING OF BASIN SUBSIDENCE

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.1 Introduction</td>
<td>222</td>
</tr>
<tr>
<td>6.2 Calculation of subsidence curves from well data</td>
<td>224</td>
</tr>
<tr>
<td>6.3 Subsidence mechanisms - a review</td>
<td>242</td>
</tr>
<tr>
<td>6.4 Modelling of subsidence curves</td>
<td>246</td>
</tr>
<tr>
<td>6.5 Summary of modelling philosophy</td>
<td>256</td>
</tr>
</tbody>
</table>

CHAPTER 7 SUBSIDENCE MECHANISMS AND THE ROLE OF THE SHETLAND PLATFORM DURING BASIN FORMATION

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1 Introduction</td>
<td>260</td>
</tr>
<tr>
<td>7.2 East Shetland Platform - models for subsidence</td>
<td>265</td>
</tr>
<tr>
<td>7.3 Subsidence of the East Shetland Basin</td>
<td>268</td>
</tr>
<tr>
<td>7.4 Quantification of subsidence mechanisms</td>
<td>281</td>
</tr>
<tr>
<td>7.5 Origin of East Shetland Platform subsidence</td>
<td>290</td>
</tr>
<tr>
<td>7.6 Subsidence in the Faeroe Basin</td>
<td>299</td>
</tr>
<tr>
<td>7.7 Summary and conclusions</td>
<td>303</td>
</tr>
</tbody>
</table>

CHAPTER 8 A DEPOSITIONAL AND TECTONIC HISTORY OF THE EOCENE OF THE NORTHERN UKCS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
</table>

REFERENCES

APPENDIX A Clay mineralogy                                             339
APPENDIX B Author citation for palaeontology                          344
APPENDIX C Computer program SUBSIDE                                    346
APPENDIX D Aspects of Palaeogene hydrocarbon prospectivity              360
Seismic stratigraphic principles, accompanied by well studies, have been used to establish a stratigraphy for the Eocene succession in the northern North Sea and Faeroe-Shetland Channel. The aim of this has been to compare the sedimentation in the basins to determine the role of a structural high, the Shetland Platform, and the controls on basin deposition during the post-rift, thermal subsidence phase of passive margin and basin evolution.

A six-fold seismic stratigraphy is observed in both basins. These are characterised by: Ela - a high amplitude reflection representing the tuffaceous claystones of the Balder Formation; Elb - mounded facies representing submarine fans of the Frigg Formation; Eic - downlapping reflections and onlap onto the East Shetland Platform, indicating shelf progradation and cessation of submarine fan activity; E2 - low amplitude discontinuous reflections representing a prograding muddy shelf; E3 - mounded and downlapping seismic facies, indicating the rejuvenation of fan activity, and subsequent shelf progradation; E4 - downlapping and gently mounded facies of a shelf and slope sequence.

The nature of submarine fan sedimentation changes with time. Early Eocene fans represented a large system emanating from canyons on the East Shetland Platform, and were fed by smaller fan systems developed at breaks in the basin slope. Later fans were more disparate. Where erosion is not observed these are more likely a response to shelf progradation, rather than a major sea level fall. Differences occur in the southern Faeroe Basin, where shallower facies are observed in the earlier sequences. Subsequent fans are developed in response to differential subsidence between Rona Ridge and the Faeroe Basin. Confinement of the fans occurs where compressional folds emerged at seabed to form basin relief.
A comparison between the Eocene basin margin sediments and global sea level curves, shows a sea level control on sedimentation. However, the details and distribution of the seismic facies indicates that these sea level changes are superimposed on a number of longer period tectonic episodes. These include i) an early Eocene uplift of the East Shetland Platform and tilt to the SE, ii) a mid-Eocene subsidence of the East Shetland Platform, and iii) a mid- to late Eocene inversion in the southern Faeroe Basin, involving reactivation of the Judd Fault, that may have controlled apparent sea level falls in the later sequences. These events have been related to the proximity of the region to the Iceland-Faeroe Ridge in the North-east Atlantic Ocean, in which seafloor spreading was initiated in the lowermost Eocene.

Studies of the subsidence histories of wells indicates the influences of pre-existing structure, and the basin-forming mechanisms, on the distribution of the Palaeogene succession. The broad, regional subsidence was a response to the thermal cooling of the lithosphere in the post-rift phase. The subsidence of the East Shetland Platform may have been a response to an inhomogeneous extension, although a flexural model can also be applied. Palaeocene faults movements are local, and assumed to be a response to rapid sediment loading.
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1.1 Objectives

Understanding the interaction of sedimentary processes and the vertical movements of the Earth's crust is fundamental to the problem of determining the history of a sedimentary sequence (eg. Mitchell & Reading, 1986). The evidence for faulting in the development of almost all sedimentary basins led to the understanding that these basins were formed in response to horizontal extension of continental lithosphere (McKenzie, 1978). In general, such basins exhibit syn-rift, fault-related subsidence, and a subsequent post-rift, thermal, subsidence phase. Much of the subsequent research has concentrated on the style of structural deformation developed in response to extension, and the sedimentary sequences developed during faulting. Relatively little research has been carried out on the late stage development of basins, and the sedimentary sequences formed after the cessation of faulting. Whilst many passive margins have been analysed, the determination of controls on deposition of the post-rift sequence has lacked detail (eg. Schlee & Fritsch, 1982).

To attempt a detailed sedimentary facies analysis of post-rift sedimentary sequences poses a problem. Most complete post-rift successions are preserved offshore. The construction of detailed sedimentary facies models in such areas faces the problem in distinguishing lateral changes in sedimentation from more profound temporal changes. This is especially the case for marine sequences, where sedimentary facies models have concentrated on modern analogues. In these, the lateral variations in a particular model are very evident, but the temporal variations in sedimentation are less obvious. Consequently, the influence of tectonic subsidence upon sedimentation during the post-rift evolution of a basin is still unclear.
The objective of this study has been to look at the evolution of the sedimentary patterns in the post-extensional, passive subsidence phase of two large basins; the East Shetland and Faeroe Basins. Geologically, these basins are relatively simple in that active rifting ceased over 70 Ma. ago (Ridd, 1981; Beach et al., 1987). Consequently, the role of individual faults in post-rift subsidence would be expected to be minimal, and controls on the development of sedimentary sequences would therefore be expected to be of a more regional nature. This project has concentrated on determining the importance of those regional controls.

Changes in the type of sedimentation, in a post-tectonic regime, can occur by the following mechanisms:-

i) **Fluctuations in the rate of subsidence of the basin.** Post-rift subsidence of a McKenzie (1978) -type basin consists of an exponentially decaying subsidence. Fluctuations in the post-rift subsidence, therefore, require a more detailed model of tectonism than presently exists.

ii) **Fluctuations in the rate of uplift of adjacent flanks and hinterland regions.** This controls the amount of material eroded off the hinterland and consequently the sedimentation rate in the basin. For deposition during the thermal subsidence of a simple extensional basin (McKenzie, 1978) these fluctuations would require an external mechanism. The interaction of the two mechanisms, though both apparently simple, means it is likely to prove difficult to determine which was dominant.

iii) **Changes in sea level.** Again, the response of a sediment pattern to eustatic changes in sea level depend on the significance of the two mechanisms above, relative to the change in sea level.

The response to all three mechanisms is most evident in coastal marginal sequences. What is preserved in the sedimentary succession is a series of transgressive and regressive cycles which are the result of the interaction of all three controls (Curray, 1964). Because sequences are more likely to be eroded from these areas, a more useful approach would be to analyse a whole basin, both the
marginal and deepwater facies, in order to define intra-basinal, and extra-basinal influences on the sedimentation patterns observed. The basic data for such a study are regional seismic lines and well ties.

1.2 Structural setting and nomenclature

The study area comprised the United Kingdom Continental Shelf (UKCS) north of the Orkney Islands. This area has been divided into two regions, characterised by the present day bathymetry (Fig. 1.1a). The northern North Sea, east of Shetland, is a region of shallow partially enclosed tide dominated sea, with water depths less than 200m. The Central North Sea and Moray Firth lie to the south. North of 62°N, the northern North Sea deepens rapidly and merges with the continuation of the Faeroe-Shetland Channel and the Norwegian Sea. The seafloor west of Shetland is characterised by deeper bathymetry. The continental shelf and slope west of Shetland trend NE-SW. The shelf deepens into the eastern margin of the Faeroe-Shetland Channel, a bathymetric trough between Shetland and the Faeroe Islands. Maximum water depth in the trough is 1700m. To the south-west (60°N, 7°W), the Faeroe-Shetland Channel is intersected by a NW-SE trending bathymetric high, Wyville-Thomson Ridge. This, and the Faeroe Bank, represent the south-eastern end of the Iceland-Faeroe Ridge, which consists of anomalously thick oceanic crust. South of Wyville-Thomson Ridge — and contiguous to the Faeroe-Shetland Channel — is the Rockall Trough, which reaches water depths in excess of 2500m. This region is outside the study area but is frequently referred to in discussions.

A number of structural elements, defining the sedimentary basins, (Fig. 1.1b) underlie these bathymetric features. The northern North Sea is composed of the N-S trending Viking Graben. Along the western flank of this graben a series of tilted fault blocks form the East Shetland Basin. The two basins merge south of 60°N into a single graben structure, referred to as the South Viking Graben.
Figure 1.1
a) Bathymetry map for the northern UK Continental Shelf (UKCS) and surrounding regions.
b) Map of the major structural features on the northern UKCS and surrounding regions.
The Faeroe-Shetland Channel is underlain by the Faeroe Basin which parallels the channel. Along the eastern flank of the basin is a structural high, the Rona Ridge which also trends NE-SW. The age of formation of these features is unknown, but they may have existed as structural entities as far back as the Middle Jurassic (Haszeldine et al., 1987).

Between the Faeroe Basin and East Shetland Basin is a structural high, the Shetland Platform. This is geographically sub-divided into the East Shetland Platform and West Shetland Platform. The geology of the Shetland Platform consists of crystalline rocks (Mykura, 1976), interspersed with fault-bounded Permo-Trias basins. The East Shetland Platform is overlain by an eastward thickening wedge of Tertiary to Recent sediments. Within the East Shetland Platform is a Triassic to Cretaceous depocentre, the Unst Basin (Johns & Andrews, 1985).

Because of possible confusion between geological and geographical nomenclature, the former terminology is preferred throughout the thesis. The exceptions to this are the ocean basins (e.g., Norwegian-Greenland Sea), and Rockall Trough, which is of uncertain affinity.

The Palaeogene succession in the Faeroe and East Shetland Basins, and the East Shetland Platform, forms the basis of the present study. In particular, emphasis is placed on the Eocene, which represents the most complete, post-rift (including all effects of the Palaeocene Atlantic rift episode) sequence. Within the established lithostratigraphic nomenclature (Deegan & Scull, 1977), the Eocene is represented by the Hordaland Group in the East Shetland Basin, but is otherwise classified as "Undifferentiated" throughout the Viking Graben (Fig. 1.2). A localised formation, the Frigg Formation has been defined in the centre of the Viking Graben.
Figure 1.2 Northern North Sea lithostratigraphy for the Tertiary. From Hamar et al. (1980).
1.3 The Database

The data used in this thesis were obtained from two main sources; i) boreholes and seismic obtained by the British Geological Survey and ii) wells and seismic based on commercial exploration of the area.

1.3.1 BGS Data

As part of the reconnaissance mapping programme carried out by the British Geological Survey (BGS), a number of shallow boreholes have been drilled to a maximum depth of 200m during the past 15 years. Of these, three boreholes have sampled the Palaeogene succession on the East Shetland Platform. Initial investigations simply identified these as Eocene in age without establishing the sequence within the framework of the North Sea lithostratigraphic or biostratigraphic schemes. The core recovered from these boreholes allows a more detailed lithostratigraphic and biostratigraphic correlation to be made, of sediments on the East Shetland Platform. This is discussed in Chapter 2.

Tied to the boreholes is a network (Fig. 1.3) of shallow seismic records. These comprise (in order of increasing penetration and decreasing resolution) pinger, boomer, sparker and single channel airgun systems. The sparker record is the most useful in this data set in that it has better penetration than pinger and boomer systems, and better resolution than BGS single channel airgun and commercial multichannel records. Examples of these data are shown in Chapter 2. Single channel airgun data are particularly good west of Shetland, and examples are given in Chapter 4. Vertical resolution is limited by the arrival of the first seabed multiple. Reflectors tied to the boreholes dip below the multiple, and are consequently obscured between the boreholes. The seismic configuration of the sedimentary sequences can only be analysed close to the borehole, and are not particularly useful as a correlative tool. Characteristics of the seismic stratigraphy of the Eocene succession on the East Shetland Platform are described in Chapter 2.
Figure 1.3: Location of BGS seismic lines used in the comparative study of the East Shetland Basin and Faeroes Basin. Boreholes sampling the Eocene, and tied to this network are shown on Fig. 2.1.
1.3.2 Commercially acquired data

Commercial exploration activity both east and west of Shetland has produced a network of regional seismic lines (Fig. 1.4), tied to commercial wells. Seismic data consist of multichannel records recorded to at least 6 seconds two way time (TWT). Well data consist of composite logs and wireline logs of the commercially released data set. For the Eocene, the quality of these data are somewhat variable as many wells are not logged either lithologically or by wireline methods, in the uppermost 500m - and in some cases, to the base of the Eocene. Useful well data (Fig. 1.5) are confined to ties on the seismic lines where at least the Eocene lithology was recorded, and where calibrated velocity logs (CVL) existed through the sequence. Gamma ray logs are an indicator of the 'shaliness' of a sequence and were of use in calibration. Similarly, CVL was useful because its response directly describes the acoustic properties of a formation, which are important to the seismic reflection characteristics. Because released commercial core is rare in the Eocene succession, seismic data represent the best tool for determining the basinwide stratigraphy in the East Shetland Basin and the Faeroe Basin.

In addition, palaeontological data has been acquired through the detailed analysis of the BGS core material. Furthermore, access was made possible by Robertson Group, to the results of a detailed biostratigraphy of a well in the East Shetland Basin. Biostratigraphical information otherwise consisted of the released data. A lithostratigraphical marker horizon, the Balder Formation, was taken as the base of the Eocene.

1.4 Principles of seismic stratigraphy

Stratigraphic analysis of a seismic section is based on the interpretation of reflections. These reflections result from the rebound of acoustic energy from physical interfaces. Such interfaces are produced by physical changes in the reflection coefficient in the subsurface, which is controlled by changes in the velocity and
Figure 1.4
Location of commercial multichannel seismic lines used in the comparative study of the East Shetland and Faeroe Basins.
Figure 1.5
Location of commercial wells tied to the seismic lines shown in Fig. 1.4
density of the rocks. In general, these increase with depth of burial of the sediments, although this in itself is not a factor in the formation of seismic reflections. Porosity is possibly the most important factor controlling the velocity and density of sediments (Sheriff & Geldart, 1983), as well as the nature of pore-fluids. Consequently, gas-filled sands can be distinguished by a 'bright spot'- a high amplitude reflection associated with the sharp increase in velocity at the gas-water or gas-oil contact. Similarly, reflections occur from sharp interfaces between, for instance, basaltic lavas and porous sediment (eg. Gatiliff et al., 1984).

With this physical basis to seismic reflection, it is somewhat surprising to find that the fundamental assumption of the seismic stratigraphic technique is that reflections result from chronostratigraphic horizons (Vail et al, 1977c), and not lithological changes. However, this assumption appears to be valid in many cases, and has been confidently applied to basin analysis since its inception (eg. Vail et al., 1977a; Schlee, 1981; Hubbard et al., 1985; Mitchum 1985; Colwell, 1988). It is perhaps easier to accept that unconformities will show reflections, since these are surfaces where a period of non-deposition and/or erosion has occurred. Subtle physico-chemical changes resulting from non-deposition (eg. limestone hardgrounds) and erosion (eg. weathering), may alter the acoustic properties of this surface. In this case, the reflection is chronostratigraphic, but is not the response to a single, simple geological event. The amplitude and polarity of the reflection is dependent on the reflection coefficient between the underlying and overlying rocks, and may vary along the length of the surface.

It is probable that seismic reflections related to chronostratigraphic surfaces are the result of subtle changes in acoustic parameters. The question is in determining the causes of these changes. It might be speculated that the acoustic parameters are a response to minor diagenetic and lithological changes, including the mean grain size of sediments (Horn, 1968). Small-scale lithological
changes are a response to short period, transient sedimentary processes. It is therefore possible to consider depositional hiatuses, such as tidal reactivation surfaces (Allen, 1980; Boersma & Terwindt, 1981), as similar to short-lived unconformities or disconformities. Consequently, the difference between a hiatus and an unconformity, is simply the geological time gap between the overlying and underlying sequences. Such subtle variations in physical parameters, associated with these 'unconformities', may cause reflections that are not resolvable by the seismic method. However, a reflection may result from the interference of a large number of reflections - at smaller, subtler acoustic interfaces - within a quarter-cycle of the seismic wavelength (Sheriff, 1985). In general for multichannel seismics, this is about 10-15m. Consequently, the horizons that are observed on a multichannel seismic section are a composite, resulting from sharp physical interfaces and small scale, rapid changes in the physical properties in sediments. The latter are the result of changes in the depositional environment.

Accepting this argument, seismic sections may be used for time-based geological interpretation, within the limitations imposed by the resolution of the seismic method. The basis behind the seismic stratigraphic approach of Vail et al. (1977a) is the interpretation of packages of seismic reflections, referred to as Seismic sequences. These contain concordant reflections, that are discordant with the underlying and overlying sequences. The packages are defined by Sequence Boundaries (Fig. 1.6). These surfaces reflect hiatuses, and are consequently assumed to result from short-lived gaps in sedimentation. Boundaries between packages may be defined by a surface onto which dipping reflectors terminate. This is called Baselap - or more specifically onlap (sub-horizontal reflectors onto a dipping surface) or downlap (dipping reflectors onto a sub-horizontal surface). Upper boundaries of sequences may be toplapping or the sharp truncation of reflectors. Sequences may also be concordant, especially in pelagic conditions, where sedimentation rate is low.
Figure 1.6 Concepts of seismic facies analysis. Internal configurations (Left), isopach maps, and reflection amplitude and continuity all lead to a geological interpretation of seismic sequences.
Seismic sequences comprise **seismic facies units**. These are reflection packages defined by their overall shape, reflection configuration, amplitude and wavelength. The internal configuration of a seismic sequence may therefore change laterally, dependent on the distribution of seismic facies units within the sequence boundaries. Such subtleties may be picked up more readily on shallow seismic than multichannel seismic records. Sedimentary interpretation of these seismic facies units can be made, and is referred to as seismic facies analysis. Isopachs of a sequence may be controlled by the processes of sedimentation (Hubbard et al., 1985) and may indicate seismic facies in areas where the overall shape of seismic facies units is not governed by subsequent erosion (Fig. 1.6).

### 1.5 Seismic stratigraphy and sea level changes

Vail et al. (1977b) produced a relative sea level curve from the late Triassic to the Neogene, using a modal average of three or more correlative regional cycles, derived from seismic stratigraphic studies of basins. This was produced by assuming that every seismic sequence boundary resulted from a change in the depositional conditions, in this case brought about by a shift in sea level. From this, it was concluded that sequences could be correlated, by their seismic facies, to these changes. Their model can be summarised thus (after Vail et al., 1977a; Vail et al., 1984) (Fig. 1.7):

**Sea level rise** - outstrips sedimentation and subsidence rate. Characterised by i) Coastal onlap - landward onlap of coastal and marginal facies. 
ii) Transgression in basin.

**Highstand** - Sea level rise balances uplift or subsidence. Characterised by i) Onlap lag - especially if a very rapid sea level rise occurred previously. 
ii) Toplap of coastal facies. 
iii) Downlap as delta/shelf progrades into the basin.
Figure 1.7 The depositional model of Vail et al. (1977a). Unconformities, bounding seismic sequences, are defined by the relationship between the under- and overlying sequences. Reflection configurations are shown by dashed line, with termination indicated by arrows. Important forms of reflector termination are shown.
iv) Regression in basin - formation of an onlapping sediment wedge.

Sea level fall - Sea level fall outstrips subsidence or uplift of the margin occurs at a rate greater than sea level rise or during stillstand.

Characterised by 1) Type 1 unconformity - shelf bypass margin, with canyon cutting and submarine fan deposition. These are characteristic of a rapid fall below the shelf edge of the preceding sequence

ii) Type 2 unconformity - a downward shift in coastal onlap characteristic of a gradual fall.

Lowstand - Sea level fall balances subsidence or uplift, or sea level is stationary.

Characterised by 1) Coastal toplap with basinward shift of onlap.

All of the changes are relative, and may result from the interaction of tectonic and eustatic processes. Only by correlation of a number of widespread passive margin settings is a truly global correlation achieved. The most up-to-date curve is that of Haq et al. (1987) which consists of both seismic and onshore correlations, based on magnetostratigraphic and biostratigraphic evidence.

1.5.1 Objections to the Vail et al. (1977a) sea level model

The Vail et al. (1977a) model, outlined above, has received widespread criticism, on both tectonic and sedimentary grounds. A number of arguments are based on the mechanism of formation of basins due to a particular plate fragmentation, such as Pangaea (Watts et al., 1982). Changes in flexural rigidity may result in onlap of the basin flanks that is tectonically, not eustatically induced (Watts et al., 1982). Furthermore, sea level fluctuations observed in a basin may be the result of short-term changes in intra-plate stresses (Cloetinho et al., 1987; Karner, 1987). Synchronicity of events would therefore be due to the similar ages of the basins and their response to a regional stress. Hence, unconformities would have only a regional, not necessarily a global significance.
The Vail et al. (1977a) model is simplistic in its treatment of the sedimentary consequences of sea level changes. Many of the sedimentary models of submarine fan and delta/shelf deposition have progressed since the model was first proposed, and it can no longer be taken in its strictest form. Using seismic stratigraphic techniques to determine the sedimentary processes in the Eocene of the East Shetland and Faeroe Basins, it is useful to define the limitations of a pure Vail-type approach.

Vail et al. (1977a) and Haq et al. (1987) consider the occurrence of submarine fans to be diagnostic of sea level fall. This has been expanded upon by Mitchum (1985) who argued that fans developed in a sea level fall could be distinguished by their derivation from a point source i.e. a formation of a canyon. Both are very speculative. Indeed, most contemporary submarine fans were initiated in response to the gravitational instability of the slope. This in turn is caused by a rapid influx of sediment onto the outer shelf (Coleman et al., 1983), rather than sea level fall.

Similarly, the formation of canyons may also relate to slope instabilities rather than the rapid fall in sea level below the existing shelf-edge (Miall, 1986). At the same time most canyons form by backcutting the shelf-edge (May et al., 1983; O'Connell et al., 1987) rather than aggradation of fluvial systems across the shelf (Vail et al., 1977a; Shepard, 1981). These are discussed more fully in section 3.6.1.1. O'Connell et al. (1987) argue that debris flows, formed as a result of gravitational slumping, act as a nucleus for subsequent canyon development and fan formation.

Where large delta-fed fans have evolved (e.g. the Amazon and the Rhone) (Damuth & Embley, 1981; Droz & Bellaiche, 1985), they may only be moderated by sea level fall and rise, and not dramatically switched on and off. Such systems may simply show fluctuations in the amount of coarse sediment deposited onto the basin floor. Essentially, the response of fan systems to changes in sea level is dependent on the extent of canyon systems on the shelf (cf. Droz &
Bellaiche, 1985), and the size of the submarine fan. However, Shanmugam & Moiola (1982) suggest that there are geological periods of formation of submarine fan systems, indicated by turbidites and winnowed turbidites in periods to eustatic sea level fall.

The consequence of these observations is that fan development is not restricted to sea level fall, but may occur at any time during sea level high- or lowstand. Simply, the Vail et al. (1977a) model is not a unique interpretation of sedimentary and seismic facies. Relative sea level fall must, therefore, be defined by more criteria than the presence of submarine fans and canyons. The rejuvenation of a submarine fan system, after earlier abandonment, may tentatively point to a relative sea level fall, although fans and debris flows may occur at any time (cf. Normark & Gutmacher, 1988). If a regional erosion event is synchronous with this - or the sequence is restricted in its deposition - then a relative sea level fall may be implied. Obviously, a more critical interpretation of the internal and regional configurations of seismic sequences is required, and more regard taken of lithological evidence.

This thesis contains such a regional study, with details of appropriate sedimentological models. Studies of the successions in the Faeroe Basin and the East Shetland Basin aid the distinction of tectonic events since both basins have an independent tectonic history, but are linked by the same sediment source area - the Shetland Platform. This region is comparatively small (approximately 100x260 km) and therefore sensitive to changes in sea level and tectonism. Consequently, it should prove possible to separate localised differences in sedimentary facies and distribution from a regional (eg. sea level, tectonic) influence, affecting the whole of the North-east Atlantic.

Quantification of sea level rise from coastal onlap is also a speculative excercise. The fundamental problem with Vail et al.'s (1977a) method of quantification of sea level variations from seismic data, is the lack of regard for alluvial facies within the
onlap wedge, which produce an error in the interpretation of the lateral and vertical position of the coastal margin by 10-100's of metres. A purely seismic approach does not provide the data that unravels these ambiguities. Verification of the sediments at onlap are required before a quantitative approach can be applied. Whilst errors due to alluvial sedimentation may not be significant in entirely marine clastic regimes, the erosion of sequences below the onlap wedge - especially during sea level fall - is a major problem in the quantification of sea level changes of a missing sedimentary sequence. Erosion on a shelf (Mougenot et al., 1983) has a significant effect on the seismic sequences in sea level rise, fall and stillstand. In particular, the Type 2 Unconformity may occur in a number of different circumstances (Fig. 1.8). Indeed, no attempt is made in this study to quantitatively define the amount of coastal onlap, and amplitude of relative sea level changes.

Where coastal margin sequences are not observed, the use of deeper water facies alone is discounted as Miall (1986) states:-

"It is difficult, in any case, to perceive the relevance of the concept of coastal onlap... or to understand how sea level change can be quantified using evidence from an entirely deep marine succession."

This is especially true when, as discussed above, the configuration of basin sediments is controlled by sedimentary and tectonic processes that over-ride the effect of sea level variation. An emphasis has been placed on what is observed qualitatively - rather than a questionable quantitative analysis - from the distribution of both deep marginal marine, seismic and lithological facies (including, where possible, biostratigraphic evidence).

1.5.2 Objections to the synchronicity of "eustatic" sea level curves

Much has been debated as to the real significance of the variations in sea level observed by Vail et al. (1977b). Indeed, this debate has not been allayed by the publication of the up-to-date version of the curve (Haq et al., 1987), which has included the calibration of
Figure 1.8 The relationship between transgression-regression, sedimentation rate and tectonic subsidence/uplift. Predicted unconformities and depositional types of Vail et al. (1977a) are shown.
the observed offshore seismic sequences, with marine sequences presently outcropping onshore. The problems with the curves remain the same since the offshore sequences have been analysed using the same Vail et al. (1977a) and Vail et al. (1984) model. Other oblections to the true synchronicity of sea level variations is the use of a composite chronostratigraphy, defined by numerous methods of radiometric dating and palaeontological zonation schemes. Inevitably, uncertainties and errors that creep into the correlation between different schemes, and different regions, weaken the confidence as to the true synchronicity of the events in global sequences. In defining a third order series of sea level changes (of the extent of a few millions of years) the analysis of Vail et al. (1977b) and Haq et al. (1987) begin to approach this limit.

Regional tectonism may explain some apparent sea level fluctuations. Theoretical studies of passive margins indicate that similar, third order fluctuations in tectonic subsidence can occur. Watts (1982) and Watts et al. (1982) argue that onlap may be associated with changes in flexural rigidity with time. However, their model can only explain a general onlap onto the basin flanks, not fluctuations of the type observed by Vail et al. (1977b). These might be explained by a complex model of intra-plate, in-plane stresses, causing minor and rapid changes in the base-rate (Cloetingh et al., 1987). These criticisms raise the point that analysis on basins formed contemporaneously may include coincident changes in tectonic subsidence, such as rapid changes in flexural rigidity (Kusznir & Karner, 1985). Basins formed at different times will always contain an inseparable interaction between regional tectonic uplift and subsidence and truly global sea level changes.

Although the eustatic nature of the sea level curves of Vail et al. (1977b) and Haq et al. (1987) is questionable, the curves represent an important synthesis of regional data. Analysis of seismic sections and onshore sequences in regions away from the calibration areas of the curves, is the way forward in integrating and comparing the sedimentary and subsidence histories of basins, and assessing
the importance of global changes in sea level. The north-western UKCS is not used in the production of the sea level curves, but is close enough to a region used in the calibration of the Palaeogene "global" sea level curve (the Hampshire Basin; Haq et al., 1987) to justify comparison. The Eocene in the Hampshire Basin comprises a marginal marine sequence, and the evidence for sea level changes has been substantiated by Plint (1988). For the purpose of this study, the Haq et al. (1987) sea level curve has been taken as an indicator of the regional influences on sedimentation—i.e. eustatic sea level changes, or simply regional changes in the stress regime across the North-east Atlantic region (Cloetingh et al., 1987). Whatever the causes of the apparent sea level changes in the post-rift development of the Faeroe and East Shetland Basins, they may be taken as local, as opposed to regional, influences on the UKCS, if they do not correlate to apparent sea level changes in the Hampshire Basin.

1.5.3 The approach adopted for seismic interpretation

The conclusion of this discussion is that the distribution of the seismic sequences is a function of erosional and sedimentological processes. The internal configuration of the sequences is very much controlled by fluctuations in the sea level, sedimentation rate and depositional processes, which may all be controlled by an overriding tectonism, whether it be fluctuations in subsidence or uplift. In this thesis, sedimentological models that explain the observed seismic configurations have been derived. Temporal changes, especially where basinwide abandonment of a particular facies occurs, are explained by basin subsidence or sea level changes. Rejuvenation of sandy facies, and by-pass margins, are only taken as indicative of sea level fall where underlying sequences show erosion. The simplicity of the tectonism at this stage of basin evolution means that local tectonic events may be inferred where the uncalibrated onlap curve produced from these, deviates from the Haq et al. (1987) sea level curve.
Consequently, the thesis comprises Chapter 2, on the lithological and stratigraphical evidence for the Middle Eocene marginal marine succession on the East Shetland Platform. Characteristics of the broader, basinwide stratigraphy and seismic facies - as defined by seismic interpretation and incorporating the succession outlined in Chapter 2 - are presented in Chapters 3 and 4. A comparison between the sequences in the East Shetland Basin and the Faeroe Basin is discussed at the end of Chapter 4. A history of Cretaceous-Tertiary tectonism on the northern UKCS is presented in Chapter 5, that attempts to explain some of the observations in the preceding chapters in relation to eustatic and tectonic evolution of the basins.

1.6 The control of basin forming mechanisms on post-rift subsidence

A regional analysis of the role that the earlier extensional history of the East Shetland Basin played in the post-rift subsidence history is presented in Chapters 6 & 7. This shows which subsidence mechanisms may have influenced the pattern of subsidence in the basin, and the extent to which the different basin forming mechanisms can be applied. Tectonic subsidence curves were derived for observed well sequences in the East Shetland Basin by the backstripping techniques of Watts & Steckler (1979) and Sclater & Christie (1980). These were used to test theoretical models of the consequences of lithospheric extension upon the subsidence history of a basin (eg. McKenzie, 1978; Jarvis & McKenzie, 1980; Cochran, 1983; Watts et al., 1982; Wernicke & Burchfiel, 1982; Wernicke, 1985) and Kusznir et al., 1987) - and the observed curves. The theoretical aspects and consequences of these models are described in Chapter 7.
2.1 Introduction

Three of the BGS boreholes drilled on the East Shetland Platform (Figure 2.1) penetrated a Palaeogene succession. The nearest commercial wells are as much as 50Km from the BGS drill sites. These boreholes therefore represent the only direct evidence on the nature of the Eocene sediments at their subcrop limit. The importance of these data is in detailing the Palaeogene sediments subcrop limit, and defining whether or not the subcrop limit represents the coastal margin of the Palaeogene North Sea Basin. The object of this chapter is to present a detailed sedimentary description of the Palaeogene succession in the boreholes, as well as a number of approaches to the correlation of the lithological units described. Preliminary ages were based on dinoflagellate cyst and foraminiferal analysis. The analyses were, however, not of sufficient detail to allow more than a basic correlation between the sedimentary sequences to be made. The new correlations include data provided by A C Morton (unpublished BGS report for BGS 80/03) and a detailed biostratigraphic analysis made by D W Jolley (Robertson Group - subsequently referred to as RG). Sediment analysis and clay mineralogy studies (using X-ray diffraction techniques described in Appendix A), were made on each of the cores.

The cores have been divided into a number of lithological units which are described in detail. Correlation between these units and the palaeoenvironmental and sedimentary interpretations are discussed at the end of each descriptive section.

2.2 Location and seismic configuration of the boreholes

BGS 80/03 (and 80/03A), drilled in 163m water depth, is the most southerly location of the three boreholes (60°7.0'N, 0°13.0'E) (Fig.
Figure 2.1 Location map for the three BGS shallow boreholes, marked by crosses. Platform edge and fault patterns from Hitchen & Ritchie (1987). Department of Energy Licence numbers are also indicated in the top left corner of the respective quadrant.
2.1). The core material consists of 10-12m of Pleistocene (Flandrian) shelly, gravelly clays, overlying a Palaeogene succession drilled to 162m below seabed. A gamma ray log was run within the drillstring, and the correlation between this and the sediment log is shown in Fig. 2.2. This Palaeogene succession can be divided into three distinct lithological units:

1) Upper Claystone Unit (10-22m below seabed)
2) Main Sand Unit (22-85m)
3) Lignitic Sand Unit (85-162m)

The coincident shallow seismic line shows the succession in 80/03 (Fig. 2.3). The Quaternary succession has been assumed to have a velocity of 1800ms⁻¹. An overall velocity of 2000ms⁻¹ is assumed for Tertiary sediments. The Lignitic Sand Unit is observed as a parallel bedded sequence with an undulose reflector defining the boundary between this and the Main Sand Unit. The contact between the Lignitic Sand Unit and the Main Sand Unit is poorly defined, but appears to be another undulose moderate to high amplitude reflection, as much as 110m below the top of the Lignitic Sand Unit. The Main Sand Unit is another unit of parallel reflections gently dipping to the East. Gently downlapping sequences can be observed within the upper part of the Main Sand Unit. An unconformity at the top of the Main Sand Unit is observed as an undulose eastwards dipping reflector truncating the tops of cross-stratification. The overlying unit contains wavy, and occasionally transparent reflections.

The second borehole, 81/16 (60°26.8'N, 0°30.0'E), was drilled at the Tertiary subcrop limit, 35km NE of BGS 80/03 in a water depth of 175m (Fig. 2.1). At TD, the borehole penetrated a fine grained greenish-grey sandstone with occasional red siltstone clasts. The change in lithology and the occurrence of a prominent unconformity on the seismic sections, at approximately this depth (146.40m), suggested that this represented Permo-Trias or Devonian sediments of the East Shetland Platform. The overlying Tertiary succession was
Figure 2.2 Composite lithostratigraphy, sediment log and gamma ray log for borehole 80/03 and 03a. Stipple - sands and sandstones; dashed lines - siltstones and clays.
Figure 2.3 Single channel sparker seismic section through 80/03, with line interpretation. 80/03 is ties to the seismic assuming a velocity of 2000ms⁻¹ for the Tertiary succession.
proved to be Palaeogene from 59.9 to 145m, below Plio-Pleistocene shelly, pebbly sands. A gamma ray log was run for only the upper part of the borehole (Fig. 2.4). The Palaeogene is again divisible into a threefold lithological succession:

i) Upper Claystone Unit (61-66m)
ii) Main Sand Unit (66-126m)
iii) Interbedded Silt/Sand Unit (126-145m)

The whole seismic succession in 81/16 (Fig. 2.5) is observed as a series of parallel, sub-horizontal reflectors onlapping onto a South-easterly dipping unconformity. Reflections become discontinuous in the upper part of the succession and are truncated by Neogene channels. The sequence below the unconformity is opaque, and corresponds to the sediments at the base of 81/16.

Borehole 81/17 (61°22.1'N, 0°10.1'W) is the most northerly borehole, and is situated 100 Km NW of 81/16 in 175m water depth. The borehole was drilled without gamma ray logging (Fig. 2.6). However, the core recovery through the relevant section was good and the sediment considerably less variable than the preceding two boreholes. The proven Eocene succession underlies 101.5m of Plio-Pleistocene shelly clays and gravels. Unlike the previous boreholes, only two distinct lithological units can be observed:

i) Glaucanitic Silt Unit (101.5-109.5m)
ii) Glaucanitic Sand Unit (109.5-137.6m)

The lowermost sequence lies on a crystalline basement of hornblendite. Coincident seismic shows two offlapping sequences (Fig. 2.7), which are discussed in more detail later.
Figure 2.4 Composite lithostratigraphy, sediment and (partial) gamma ray log for 81/16. Symbols as for fig 2.2.

Figure 2.5 Seismic sequence correlated to 81/16 with line interpretation. Tertiary velocity 2000ms⁻¹.
Figure 2.6 Lithology and sediment log for 81/17. No gamma ray log run.

Figure 2.7 BGS single channel sparker line through 81/17. Tertiary velocity 2000 ms$^{-1}$. 

Hornblende (1051 ± 24 Ma: Hitchen & Ritchie, 1987)
2.3 Early to early Middle Eocene sedimentation

2.3.1 BGS 80/03 - Lignitic Sand Unit (85-162m TD)

The top of the Lignitic Sand Unit is defined by a sharp contact with the overlying Main Sand Unit, at 85m below seabed. The base of the unit is undefined in this borehole.

From 138m to the 162m (TD) the Lignitic Sand Unit consists of a red and light brown colour with little carbonaceous material present. Alternating layers, up to 5mm thick, of silt and sand have a reddish tan colour that dominates in sandier layers. This suggests that post-depositional oxidation, and oxygenated groundwaters have preferentially oxidised the higher porosity sands. A thicker sand, between 150m and 162m (TD), is hard but non-calcareous, light brown with red oxidation blotches and occasional euhedral pyrites. Calcite cementation becomes important at 160m, with the sands absent of oxidative staining.

Above this the Lignitic Sand Unit is dominated by peaty silts with micaceous fine-grained sands. These sands occasionally define fining upwards sequences. At 130m a lignite of woody debris above a clean light brown sandstone is sampled. The silts below this sand are again carbonaceous but contain a coarse sub-angular sand.

Between 88-102m the Lignitic Sand Unit is comprised of carbonaceous sands and silts. A series of fining upwards sand units are present at 102m, with an intervening woody lignite at 102.5m. A coarsening upwards succession contains rootlet horizons, and is topped by 1-2cm thick lenses of lignite (98.84m). These consist of leaf fragments and pieces of well-preserved wood in sharply defined layers within the sands. Overlying the lignite is a fine-grained sand to silt with non-calcareous concretions and fragments of plant material. The sands have a milky-white to cream colour. A high gamma ray signature suggests that the top of the Lignitic Sand Unit is silty in nature.

The clay mineralogies are dominated by kaolinite, (Fig. 2.8) which constitute up to 90% of the <8 phi fraction. Smectites are rare, but
Figure 2.8 Distribution of clay minerals in 80/03 & 03A. The various biostratigraphic schemes are also displayed.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Lithostratigraphy</th>
<th>Biostratigraphy</th>
<th>Clay Mineralogy</th>
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<td>80/03 &amp; 03A</td>
<td>Deegan &amp; This Soud (1977) Study</td>
<td>King et al. &amp; Costa et al. (1983) al.</td>
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<td>Upper Claystone Unit</td>
<td>NS810</td>
<td>D14</td>
<td>1000</td>
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<tr>
<td>Middle Main Sand Unit</td>
<td>BAR-2</td>
<td>D11</td>
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<td>Lower Barren</td>
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<td>Lipshic Sand Unit</td>
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Figure 2.9 Distribution of clay minerals through the Palaeogene succession in 81/16. The various biostratigraphic schemes are also displayed.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Lithostratigraphy</th>
<th>Biostratigraphy</th>
<th>Clay Mineralogy</th>
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<td>81/16</td>
<td>Deegan &amp; This Soud (1977) Study</td>
<td>King et al. &amp; Costa et al. (1983) al.</td>
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<tr>
<td>NG</td>
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<td>D14</td>
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<tr>
<td>UCU</td>
<td>Upper Main Sand Unit</td>
<td>BAR-4</td>
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<td></td>
<td>Lower Main Sand Unit</td>
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<td>Lower Main Sand Unit</td>
<td>BAR-3</td>
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<td>HORDALAND GROUP</td>
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<tr>
<td>ISSU</td>
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UCU - Upper Claystone Unit
ISSU - Interbedded Silt/Sand Unit
NG - Nordland Group

38
vermiculite is present throughout the sequence. Gibbsite is observed as a minor mineral from 108-137m, corresponding to the siltstone dominated portion of the Lignitic Sand Unit. The association of gibbsite with kaolinite is a common one (Biscaye, 1965). The increased abundances around river systems (Biscaye, 1965; Park & Han, 1985) is attributable to tropical and sub-tropical weathering and the enrichment of alumina in laterites. The implication of this zone is that local soil profiles provide some of the clay mineralogy, which is not apparent in the mineralogy of the Main Sand Unit.

The palynological analysis of the lignites suggests an early Middle Eocene age, based on the lignite at 98.84m. This has a very rich miospore flora, dominated by Laevigatosporites haardti and Microfoveolatosporites pseudodentatus associated with Brosipollis striatobrosus. The lignite below, at 103m, has a very sparse flora, probably because the lignitic material is dominantly massive woody debris. Winnowing of this may be responsible for the removal of the flora. Prasinophysid algae (Leiospheres), acanthomorph acritarchs and marine dinoflagellate cysts are observed within the uppermost clays. All other samples show a similar age of BAR-3 zone of Bujak et al. (1980). Because samples within the lower sands are diagenetically oxidized they contain a sparse flora and highly inertinitic kerogen. The age of the base of the Lignitic Sand Unit is unproved, especially as the borehole terminates within the unit. Palynological samples in the lowermost part of the unit show no evidence of a marine influence.

2.3.2 81/16 - Interbedded Silt/Sand Unit (123-145m)
The top of the Interbedded Silt/Sand Unit can only be tentatively identified, between 110m and 125m. The poor recovery of this interval is interpreted as poorly consolidated sands, lost as washout. The top of the unit is therefore defined to be the base of a white sandstone at 123m. The base of the Interbedded Silt/Sand Unit is taken as the change from peaty silts to the coarse sands assigned to the ?Permo-Trias.
The lowest sediments in the unit are light grey claystones, occasionally micaceous at 139.8m, and free of organic material. The sandstones at 131m and 135m are light brown and fine upwards into the silt/sand laminations and finally into the peaty clays. Euhedral pyrite is present within these sands. The clays become peaty downhole, becoming soft and greenish brown. At the top the unit is dominantly silty with 1m thick sandy sequences within it. These silts and clays at the top of the unit are dark, chocolate brown and carbonaceous with a peaty smell.

The silts and sand are characterised by a smectite-poor clay mineralogy (Fig. 2.9), typically 5-15%. What smectite is present is well-crystallised (v/p=0.5-1.0; Biscaye, 1965) (Fig. 2.10d). Kaolinite and ordered illite exhibit variable proportions, kaolinite appearing to decrease, and illite consequently increase up the unit. Illite shows high crystallinity (peak width= 0.25-0.40°(2θ); Thorez, 1976). If kaolinite is taken to represent the input from acidic, swampy terrestrial environments (as inferred for the Lignitic Sand Unit in 80/03), then this trend suggests the diminishing importance of this source towards the top of the unit.

As with the Lignitic Sand Unit in 80/03, the Interbedded Silt/Sand Unit in 81/16 proves to be barren of dinoflagellate cysts (Fig 2.9). The RG palynological analysis indicates an age no older than B-3 of Bujak et al. (1980) for the single diverse assemblage, at 133.20m. This puts the age at no older than lower Middle Eocene, but correlates this unit with the Lignitic Sand Unit in 80/03 (Fig 2.17). The flora observed contains marine microplankton, increasing in proportion up the unit. Conditions would appear to have been marine throughout deposition, probably outer neritic, unlike 80/03.
Figure 2.10 Characteristic x-ray diffraction records for 81/16. Horizontal scale is angle of diffraction in degrees (2θ). Position of major peaks and their stability through a number of analytical procedures, is an indication of the clay mineralogy. Vertical scale is intensity and is arbitrary.
2.4 Middle to late Middle Eocene sedimentation
2.4.1 80/03 - Main Sand Unit (22–85m)

The top of the unit is defined by a chert lag. This is associated with a downhole decrease in the gamma ray log signature, indicating a reduction in the argillaceous content. The base of the Main Sand is similarly indicated by a downhole increase in the gamma ray signature at the top of the Lignitic Sand Unit. The unit is subdivided into three sub-units: the Upper Main Sand Unit - medium grained, coarsening upwards sandstone (22–50m); the Middle Main Sand Unit - a massive glauconitic sandstone (50–64m); and the Lower Main Sand Unit - a well-rounded quartzose sand (64–85m).

The Lower Main Sand Unit comprises a calcareous cemented fine-grained sand with occasional well-rounded coarse quartz grains. The top of the unit consists of a 10cm layer of highly abundant glauconite. Core recovery is poor at the base of the unit, and the base itself is defined by a sharp rise in the gamma ray signature. Overall, the sands are quartzose and poorly sorted with a light brown to cream coloured clay infilling the pore spaces.

The Middle Main Sand Unit is composed of an 18m thick light-brown to greenish, massive, very fine-grained sandstone with calcite cement. Grains are sub-angular to sub-rounded and glauconite is abundant (proportionally, 30–40% of the >4.0 phi fraction). Vertical surfaces show circular blotches which are probably faint worm burrows, but the unit generally shows no internal structure.

Between 22m and 50m the Upper Main Sand Unit is a coarsening upwards sequence. The base of the unit is a poorly sorted silty sand with 4–5 mm bands of carbonaceous silt. The top is a massive light brown, fine grained, moderately sorted, hard sandstone. The core recovered is calcite cemented, although regions of poor core recovery may indicate washout of poorly cemented sands. Pyrite and glauconite are pervasive throughout the sub-unit.
The clay mineralogies also show a threefold differentiation, albeit with boundaries at different depths (Morton; unpublished BGS report). From 26 to 35m (approx.) the clay mineralogy (Fig. 2.8) is dominated by sepiolite, and probably reflects invasion of the Upper Main Sand Unit by drilling fluids. This invasion indicates good porosity in the sands and minor amounts of in situ clays. The clay mineralogy from 64-85m changes to one with a higher proportion of kaolinite and corresponding decrease in the proportion of smectite. The carbonaceous silty sands, between 42m and 64m, do not show such a phenomenon and the clay mineralogy is dominated by a high proportion of smectite (80-90%). Small amounts of kaolinite and mica, with trace clinoptilolite are also present. This distinction cuts across the lithological boundary at 50m, between the coarsening upwards sandstone and the massive sandstone. It would appear that the silty sands show greater similarity to the Middle rather than the Upper Main Sand Unit. However, the clay mineralogy appears to be facies controlled, in that the clays are winnowed out of the coarser, and hence higher energy environment, sands. The similarities are more likely controlled by provenance, and suggest a common source of sediment for the Upper and Middle Main Sand Units.

The micropalaeontological dating of the Main Sand Unit in 80/03(Fig. 2.8) is based on palynological evidence as the unit is barren of foraminifera throughout. This is consistent with the implications of the above and imply continental to marginal marine environments of deposition. The dinoflagellates present indicate an age no younger than Late Eocene, between 36m and 64m, based on the presence of Areosphaeridium dictyplokus, A. pectinicformae, Glaphyrocysta semitexta, Hystrichokolpoma cinctum and Rhombodinium spp. indet. Middle Eocene ages are inferred for the sands from 68m to 88m for a poorer dinoflagellate assemblage containing A. dictyplokus, Eatonicystta ursulae and Dracodinium varielongitudum. The RG palynological analysis, however suggests that the Main Sand Unit is mostly Middle Eocene in age. The uppermost portion of the sands, between 26m and 39m maximum, contains flora identical to those in the overlying Upper Claystone Unit. This suggests that the coarser,
Winnowed sands at the top of the Main Sand Unit represent a reworking of the underlying sands. The biostratigraphic change does not, in this case, correspond to the lithological change. The position of the biostratigraphic boundary, however, is uncertain as the base of the Upper Main Sand Unit, show a flora consistent with a late Middle Eocene age. Part of the Upper Main Sand Unit therefore must have been subsequently reworked in the Oligocene.

The base of the coarsening upwards sub-unit, at 50m, relates to a region of highly condensed, or missing, palynofloral events with assemblages showing an age no younger than D11 (Costa et al., in press) in the silty sands, and ages no younger than middle D10 (Costa et al., op. cit.) at 51m - where *Diphyes ficusoides* and *Dracodinium varielongitudum* appear. It is possible, in this case to conclude that the base of the coarsening upwards sands, is represented by an unconformity, within the late to middle Middle Eocene. This is also consistent with a downhole change in kerogen types from vitrinitic to amorphous humic. The lithological change at 50m therefore reflects a change in environment as well as a biostratigraphic break.

The middle massive sand and the lower, quartzose sand show palynofloral assemblages consistent with a middle Middle Eocene age, no older than Zone B-4 of Bujak et al. (1980).

### 2.4.2 Main Sand Unit - 81/16 (66-123m)

The top of this lithological unit in 81/16 is defined by the drop in the gamma ray log from the Upper Clay Unit. The base is defined by the base of the white sandstone at 123m. The Main Sand Unit is again sub-divisible into three units: Lower, Middle and Upper Sand Units.

At 106m the sandstones are dominated by large, 2mm glauconite grains giving the sandstones a dark green appearance. These define the top of the Lower Main Sand Unit. The sands of this unit are white, fine-grained and calcite-cemented.
Between 90m and 105m, the Middle Main Sand Unit consists of a hard, calcite cemented, fine to medium grained sandstone, similar to the thin sandstone above. Glauconite is ubiquitous, and coarse rounded quartz granules occur randomly through the sandstones. The sandstones are also calcite-veined.

The Upper Main Sand Unit consists of light brown to white fine-grained quartzose sands with black silty laminations defining the top of thin fining upwards cycles. The sands themselves are well sorted and grains sub-angular to sub-rounded. These overly a layer of olive green glauconitic siltstones with calcareous bands. Mica and occasional flaky green chlorite are present, as well as pyrite. The proportion of lignitic material increases downhole.

The laminated sand and siltstones continue to 82.3m where a 34cm thick light brown calcareous sandstone is sampled. The sandstone is fine grained and dominated by white and smokey, occasionally yellow, subangular to subrounded quartz grains. Glauconite is prevalent, with pellets up to 1mm throughout the sandstone, occasionally defining glauconite-rich bands. The calcite-cemented sandstone has a sharp contact with the underlying silty sands.

The clay mineral distribution (Fig. 2.9) through the Main Sand Unit breaks down into two distinct zones:

1) Below 98m, the Lower Main Sand Unit and the lower part of the Middle Main Sand Unit show a clay mineralogy dominated by kaolinite (Fig. 2.10c). Smectite and illite, in lower proportions and less well crystallised than above (v/p=0.5-0.6; illite peak width=0.4-0.7). Illite exhibits an IV-type crystallinity (Lucas, 1963), indicating the opening of the illite structure to behave like vermiculite. Vermiculite and gibbsite occur as minor minerals. An intermediate clay content, combining the proportions of the Upper Main Sand Unit clays and the minerals of the Lower, is observed at 97.60m. This implies a reworking of the underlying sands. Overall,
the mineralogy is very similar to that observed in the Lower Main Sand Unit in 80/03.

ii) The upper part of the distribution corresponds to Upper Main Sand Unit and the upper part of the Middle Main Sand Unit. The clays consists of a high proportion of kaolinite (Fig. 2.10b), approximately 60-65% based on peak height ratio (Griffin, 1971). The rest of the mineralogy consist of equal proportions of ordered smectite and illite. Illite is characterised by a symmetric (001) peak, implying a well-crystallised form (half-height width=0.35-0.55° (2θ)) (Thorez, 1976). Smectite shows a similarly well-crystallized form (v/p=0.57-0.75). Derivation of these clay types may be from detrital or authigenic processes (Singer, 1984). Detrital sources are preferred in this case because of the present-day position and depth of the sediments.

Dinoflagellate cyst analysis of one sample in the Main Sand Unit in 81/16 contained an assemblage suggestive of the Middle Eocene coleothrypta zone of Costa & Downie (1976). This sample was well within the Main Sand Unit, and RG samples at the top of the unit indicated an assemblage consistent with the upper part of zone D11 of Costa et al. (in press) (Fig. 2.9), or at least BAR-3 zone (Bujak et al., 1980). This correlates with the Main Sand Unit in 80/03 (Fig. 2.17). No samples were taken in the uppermost 5m of the unit, so there is no direct evidence of Oligocene flora in this, as there is in 80/03. The intra-Main Sand Unit hiatus, at 50m in 80/03, is also observed. Again, Zone B-5 of Bujak et al. (1980) is missing, and suggest the hiatus occurs at 90m. An age corresponding to the lower part of D10 (Costa et al., in press), or Zone B-4 (Bujak et el. (1980), is obtained from samples at 98.50m, where Diphyes ficusoides and Dracodinium varielongitudum first occur. This correlates to the Middle and Lower Main Sand Units in 80/03. The glauconite horizon corresponding to the base of the middle Main Sand Unit in 80/03, is observed as the glauconite-rich sands at 106m in 81/16. The glauconite horizons correlate biostratigraphically,
suggesting a regional transgressive event separating the lower and middle sequences of the Main Sand Unit.

Reworked flora reported by the RG analysis occur in the Upper Main Sand Unit, to 73m. Apart from Palaeocene, Middle and Early Eocene, Middle and Late Jurassic flora are observed.

A seismic correlation ties the Main Sand Unit to a sequence of continuous cross-stratified units to the south-west of 80/03 (Fig. 2.11). The unit in this case breaks into five cycles of south-eastwards, shallow (1°-2°) dipping cross-stratification, throughout the whole of the Main Sand Unit. The southeasterly extremity of the seismic unit is not seen as the Eocene succession dips below the seabed multiple, and the internal structure is too detailed for multichannel seismics. Depositional models for this seismic sequence are discussed in Section 2.6.2

2.4.3 81/17 - Glauconitic Sand Unit (109.5-137.6m)

This interval occurs between the crystalline basement and the overlying Glauconitic Silt Unit. The sands are dominantly sub-angular to sub-rounded quartz, with few micas. Quartz grains are again clear or yellowish, and pelletal glauconite is ubiquitous, and up to 1mm diameter. The sand is uniform throughout the unit, but quartz grains are mostly yellow at the base.

The clay mineralogy of the Glauconitic Sand Unit is characterised by a high proportion of kaolinite (Fig. 2.12). Smectite dominates over illite with smectite proportion as high as 47%, and dominating the mineralogy at the base of the unit. Smectite is reasonably well-crystallised (Fig. 2.13) \((v/p=0.54-0.88)\). Gibbsite is present, as a minor mineral at the top of the unit, corresponding to the zone of reworked microfossils discussed below.

The first diagnostic foraminiferal assemblage occurs at the top of the unit (Fig. 2.12). A single specimen of Planulina palmerae defines the Middle Eocene age of the sands, restricting the
Figure 2.11 Seismic line showing the cross-bedded sequence south of 80/03. This seismic sequence correlates with the Main Sand Unit.
uppermost interval to the benthonic foraminiferal zone NSB5 (King, 1983). This specimen is specific to this zone and suggests an early to middle Middle Eocene. Planktonic foraminifera are in greater abundance in the top of the unit and are characteristic of the NSP6 to NSP7 zones of King (1983), which are themselves correlated to NSB5.

The occurrence of *Turrilina brevispira* at 120.4m indicates zone NSB3 to NSB4, equivalent to the Early Eocene. The zonation is based on basinal assemblages and it is possible that this taxon has a younger range on the shelf. The palynology is more consistent to Middle Eocene ages, especially the middle of Zone D10 of Costa et al. (in press), with *Dracodinium varielongitudum* occurring at 119.70m. This assigns the Glauconitic Sand Unit to the Lower Main Sand Unit in the other two boreholes (Fig. 2.17). The presence of reworked flora, of Callovian-Kimmeridgian, ?Toarcian and early Cretaceous ages, may explain the discrepancy, in that the early Eocene foraminifera may be reworked. This is tenable, since it places younger reworked species below older and might define the onset of reworking. Any form of reworking is certainly sparse below 120.4m.

### 2.4.4 Glauconitic Silt Unit (101.5-109.5m)

This unit consists of a greenish moderately sorted glauconitic silty sand. The grain size is uniform throughout the sequence and no variations or internal structures are observed. The silty sands consist of sub-angular to sub-rounded quartz grains with mica and occasional garnets. Pelletal glauconite is abundant.

The clay mineralogy shows a high proportion of kaolinite (Fig. 2.12), typically 40-50% with smectite dominating over illite (smectite-illite ratios as high as 6.0). Smectites are well- to moderately well-crystallised, with v/p ratios from 0.88 to 0.52. Illites are in low proportion (Fig. 2.13), but appear to be well- to moderately well-crystallised (illite(001) peak width= 0.4-0.7).
Figure 2.12 Clay mineral distribution in the Palaeogene succession of borehole 81/17. Biostratigraphical zonation is also shown.

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<tr>
<th>Lithology</th>
<th>Biostratigraphy</th>
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<tr>
<th>Glauconitic Silt Unit</th>
<th>Undiagnostic</th>
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<tr>
<td>NSB5</td>
<td>NSP6/7</td>
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<td>Gibbsite</td>
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**Figure 2.13** Characteristic X-ray diffraction records for the Eocene sedimentary units in 81/17. Scales as for Fig. 2.9.
The Glauconitic Silt Unit contains a sparse foraminiferal assemblage dominated by calcareous and occasional agglutinated foraminifera. The sparsity and low diversity preclude a good age dating by this method. The presence of calcareous, benthonic foraminifera suggests a deposition in a restricted shelf environment. The RG palynological analysis, however, proves that this unit is correlatable to the middle part of the Main Sand Unit in both 80/03 and 81/16. The occurrence of Heteraulacacysta porosa, Phthanoperidinium echinatum and Diphyes colligerum restricts the possible age to BAR-3 to BAR-1 of Bujak et al. (1980), or no older than mid-Zone D10 of Costa et al. (in press). This correlates the Glauconitic Silt Unit with the Upper Main Sand Unit in 81/16 (Fig. 2.17) and the base of the coarsening-upwards and middle, massive sands of the same unit in 80/03. The last occurrence of Homotreblium floripes/vallum group at 108m is taken to define an age no older than BAR-1 of Bujak et al. (1980) (roughly equivalent to lower D11 of Costa et al. (in press)).

Reworking is again prevalent within this equivalent of the Upper Main Sand unit. Reworked dinoflagellate cysts recorded include types diagnostic of Middle to Late Kimmeridgian, Early Cretaceous and Pliensbachian-Bajocian ages. Middle to Late Jurassic and Middle to Late Carboniferous spores have also been recognised.

2.4.5 Seismic correlation of 81/17
The single channel sparker line running through the borehole gives a good record of the two sedimentary sequences (Fig. 2.6). Two seismic sequences are observed, both onlapping onto a north-easterly dipping undulose, opaque basement. This corresponds to the metamorphic basement observed at the base of 81/17. The lowermost, Glauconitic Sand Unit corresponds to an offlapping sigmoid progradational sequence dipping towards the north-east and thickening from 0m to 90m before disappearing below the seabed multiple. This unit also fills undulations in the basement where a local onlap occurs within hollows. The subcrop limit of the Glaucocnitic Sand Unit is slightly below the point at which basement levels out to be broadly horizontal. This suggests that the onlap limit is primary, i.e. not
significantly influenced by a subsequent planation/erosion event. The Glauconitic Silt Unit, on the other hand, shows onlap onto the Glauconitic Sand Unit and subcrops against it. The sequence is again sigmoid progradational. The top of this unit is also planed by the overlying high amplitude reflections of the well-cemented Pliocene shelly gravels. A seismic sequence is present above the Glauconitic Silt Unit, but is truncated by the Pliocene sequence to the north of the drillsite.

2.5 Oligocene to Miocene

2.5.1 Upper Claystone Unit - 80/03 (10-22m)
The unit directly below the Pleistocene is a calcareous claystone dominated sequence with abundant glauconite. The top of the unit is defined by a change from greenish grey to brown clays. At the base, the claystones coarsen downwards into a fine to very fine grained poorly sorted sand with large (5mm) rounded clasts of chert. Occasionally, the claystones are highly glauconitic in bands up to 20cm thick.

The gamma ray signature associated with the Upper Claystone Unit shows a high reading, with occasional troughs, indicating the dominance of fine grained material in the sequence, accompanied by thin, 1m, sandier layers.

The clays within the unit are dominated by smectites and illite (Fig. 2.8), with minor chlorite. Heavy mineral assemblages show a relatively high proportion of amphibole (Morton, unpublished BGS report).

The age of this unit is Upper Palaeogene to Neogene, with each of the various micropaleontological techniques giving different ages within this. The calcareous foraminifera showed both low numbers and diversity. The foraminifera that were present indicated a Late Oligocene to Early Miocene age (NSB8 to NSB9; King, 1983) based on the occurrence of *Asterigina guerichi* and *Bulimina elongata*. The appearance of *Elphidium inflatum* would tend to indicate a Middle
Miocene age (NSB10 to NSB11), however. The very restricted numbers of foraminifera put some doubt into such a well defined age. Dinoflagellate cyst analysis supports a Late Oligocene age, whereas the RG palynological analysis indicates a mid-lower Oligocene age (Zone D14 of Costa et al. (in press)). The palynoflora appears to be associated with reworking of Jurassic and Carboniferous (Westphalian to Namurian) flora. The only consistent ages within all these dates appears to be Middle to Late Oligocene.

2.5.2 Upper Claystone Unit - 81/16 (61-66m)
The top of the Palaeogene succession is similar to BGS 80/03 in that it consists of a claystone unit. This is below a zone of poor core recovery which might represent the sandier facies of this unit. The low gamma ray reading certainly implies a low argillaceous content. The claystone is olive-green silty and micaceous. Occasional glauconite is present, as well as laminations of coarser quartz and lithic, mostly metamorphic, grains. Trace pyrite and limonite were also observed within the sieved sand fraction. Thin bands of calcareous claystone are also present.

The Upper Claystone Unit is correlatable with an increase in the gamma ray log between the overlying sands and the Main Sand Unit (Fig. 2.4).

The clay mineralogy is illite-rich, with kaolinite as the most abundant secondary mineral (Fig. 2.9). Crystallinity of the illite is high (0.35°(2θ), Fig. 2.10a), whereas that of smectite is low (v/p ratio is 0.38). The mineralogy is reflected by the overlying Plio-Pleistocene mineralogy, which also contains minor amounts of vermiculite, implying that the mineralogy is reworked from the underlying Palaeogene succession.

The micropalaeontology of this unit indicates a middle Lower Oligocene age (Rupelian) from the Dinoflagellate cyst analysis. The assemblage contains Chiropteridium species, Cordosphaeridium cantharellum, Wetzeliella symmetrica, W. gochtii and Homotreblium
floripes, which are also observed in the Upper Claystone Unit of BGS 80/03. The correlation of these units is confirmed by the RG palynological analysis, which assigns a mid-Lower Oligocene age (Fig. 2.9). Reworked flora observed in this analysis include Paleocene to Middle Eocene, Cretaceous, Middle-Upper Jurassic, and Namurian to Westphalian.

2.6 Depositional history of the East Shetland Platform marginal sequence

2.6.1 Lower to early Middle Eocene

The facies interpretation of the Lignitic Sand Unit breaks down into three parts. The presence of abundant carbonaceous material and the terrestrial flora imply that the whole of the unit is deltaic in origin. The uppermost part of the unit, dominated by fining upwards sands and lignites is indicative of an 'abandonment' facies on the delta plain (Elliott, 1986), especially as the base of the Main Sand Unit is marine. The mixing of a marine flora with algae, considered to be typical of high stress, salinity mixing environments (D.W Jolley, pers. comm.; Tappan, 1980) suggests that this point represents a transition between a fully continental/deltaic to fully marine facies. The evidence for salinity mixing, associated with peaty clays, suggests a marine influence on the delta plain and a tide-influenced delta model is invoked. The dominance of terrestrial flora in 80/03 indicates a fluvially dominated part of such a delta.

The facies interpretation of the Interbedded Silt/Sand Unit has to take into account the abundance of lignitic material in the upper part, combined with the marine microplankton found in this sequence. Lignites and coarse wood fragments immediately imply some low energy environment, such as a deltaic interdistributary swamp or lacustrine facies. The marine flora, however, suggests that the swamp was open to the sea, and thus the sediments may more readily be interpreted as a marine bay, with sands representing washover fans or minor river deltas. The clay mineralogy also suggests the increased importance marine-supplied clays and non-delta-plain sources,
supplied by longshore drift and currents (Park & Han, 1985) towards the top of the unit. Changes in climate or hinterland rejuvenation (Sladen & Batten, 1984), during Interbedded Silt/Sand Unit deposition, are not reflected in any changes in sedimentation.

An analogue in the Mahakam delta (Allen, Laurier & Thouvenin, 1979) suggests that the sands in 80/03 represent pseudo-point bars within a fluvial channel. The pyritic sands at the base of the unit may represent channel mouth bar deposits in either model; subsequent oxidation and exposure may being the cause for the absence of marine flora in this sand.

The Eocene delta system proposed for the northern North Sea must have been small in comparison to Mississippi or Mahakam analogies. The absence of a large hinterland area (Fig. 2.14a), outlined by the dimensions of the Shetland Platform suggest that fluvial systems draining into the delta must have been poorly developed, and less than 100 Km long.

The dominance of kaolinite in the clay mineralogy indicates that chemical conditions prevalent at the time of sediment deposition resulted in the groundwater alteration of the clay assemblage. Kaolinite is a product of intense acid leaching of soils in organic-rich marsh soils, conditions which rapidly degrade any detrital illites, smectites and mixed layer clays (Keller, 1970; Grim, 1962). The rocks that would have constituted a source for the Palaeogene sediments derived from the Shetland Platform, would have produced a far more varied clay mineralogy if chemical weathering were not important. The Palaeogene clay mineralogies observed over a wide area of the basins in North-west Europe, from Vering Plateau to the Central North Sea, indicate a more smectite-rich composition attributed to the eruption - and subsequent alteration - of volcanic products during this period (Karlsson et al., 1978; Froget, 1981).

In these cases, kaolinite is only present as a secondary or minor mineral throughout the Palaeogene. The terrestrial environment and post-depositional oxidation which has occurred is reflected in the
Figure 2.14a) Evolution of the Eocene margin on the East Shetland Platform. Lower to early Middle Eocene based on Lignitic Sand Unit and Interbedded Silt/Sand Unit, and show the inferred regions of delta progradation. b) Middle Middle Eocene based on Main Sand Unit and Glauconitic Sand and Silt Units. Basinal extent of the facies is schematic, as are the positions of channels in the deltaic sequence.
micropalaeontology (Fig. 2.8). This oxidation is likely to have taken place during the progradation of the delta and preceding subsequent abandonment.

2.6.2 Middle to late Middle Eocene

The facies interpretation for the Main Sand Unit follows the three-fold subdivision outlined above. The presence of glauconite and carbonaceous material in the sands suggests that these were deposited in a marginal marine conditions. The environment may have been too severe to support a foraminiferal assemblage, but bioturbation may have been important in destroying any sediment structures that were present in the sands. The sands may therefore represent either shelf shoals and bars or the upper shoreface of barrier beach deposits.

The occurrence of a highly glauconitic bed between the Middle and Lower Main Sand Units suggests a hiatus, which is not reflected in the micropalaeontology. The hiatus is thus short-lived and rapid. For glauconite to be so well developed water depths of over 50m must have been exceeded during the transgression, producing conditions allowing the 'verdissement' of foraminifera (if present) and faecal pellets (Odin & Matter, 1981).

The Middle Sand Unit can be interpreted to have had the same depositional conditions as the Lower Main Sand Unit, although the finer grain size, in 80/03, might suggest that the energy of the environment was lower.

The unconformity between the Upper and Middle Main Sand Units, in 81/16, is represented by a change from a moderately high energy to low energy environment with a decrease in the thickness of sandier units and the increase in the amount of silt laminations within the sands.

The presence of kaolinite in high proportions in the Main Sand Unit of 80/03, suggests that the clay mineralogy is controlled by locally
acids, rather than the gross mineralogy of a Shetland Platform provenance. Similarly, the absence of apatite from the heavy mineral assemblage of this part of the Main Sand Unit indicates the influence of meteoric waters (Morton, 1986). This implies that the Upper and Middle Main Sand Units were deposited in similar conditions in which clay mineralogy was influenced by reworking of continental weathering products.

The relative increase in abundance of planktonic foraminifera in 81/17 is taken to imply more open, probably deeper marine conditions. The species involved suggest a middle shelf environment for the upper reworked sands. This is based on the present day distribution of foraminiferal species (Murray, 1971) and so is not conclusive, since distribution patterns may not be the same. The lower part of the sands have proportionally fewer foraminifera and similar to the Glaucophanic Silt Unit in this respect. An inner shelf environment is thus concluded for these sands.

The position of 81/17 on the upper surface of the shelf cliniforms, 1-2 km from the palaeosshelfbreak (as defined on the seismic in the manner of Vanney & Stanley, 1983) is consistent with the palaeoenvironmental conclusions above. The sigmoid progradational configuration also implies a low energy shelf (Mitchum et al., 1977; Mougenot et al., 1983).

The disconformity between the units in 81/17, as observed on the seismic, corresponds to the Zone B-5 (Bujak et al., 1980) hiatus between the Upper and Middle Main Sand Units, observed within 80/03 and 81/16. This suggests that the unconformity is present here, if not directly observed, and relates to some change in sea level, producing two shelf wedges. The minor planation of the Glaucophanic Sand Unit implies that the effect of lowstand causing the unconformity was not great this far North.

The correlation of the Main Sand Unit to the cross-stratified seismic sequences supports a marginal marine interpretation.
However, the low angle of dip of the cross-stratification precludes a delta-front progradation, where idealised Gilbert-type deltas exhibit 10°-25° dipping foresets (Elliott, 1986) (Fig. 2.15a). The internal configuration of the seismic sequence is also significantly different from the deltaic Lignitic Sand Unit.

Cyclic progradation of a barrier beach sequence is a possible explanation for the observed cross-stratification (Fig. 2.15b). Dips and height of each cross-stratified unit are similar to those observed in the Galveston Island barrier, Texas (McCubbin, 1982). The seaward dipping foresets would thus represent time-lines for the prograding shoreface of a barrier succession. The problem with this model is that each of the five sequences would have to represent a barrier, rapidly prograding over the shelf and terminated by a rapid transgression, drowning the whole of the barrier system (Heward, 1981). If the sediments in 80/03 are to be equated to the seismic sequences then each unit should show a coarsening upwards from lower shoreface muds and laminated sands to upper shoreface sands and lagoonal mud facies. Each unit should then be topped by an erosional surface and a transgressive boundary, defined by a change in sedimentation to a very much deeper facies. The sediments in 80/03 are in fact consistent with extensive shoreface erosion, leaving only shoreface deposits, and sedimentary facies are less variable than such a model implies. For such a model there is a paradox in that the sediments imply poor preservation (Heward, 1981) of the barrier systems, whereas the seismic imply excellent preservation.

A more likely interpretation, encompassing the sedimentological evidence from 80/03, is of shelf sand-wave deposition (Fig. 2.15c). Features similar to those described above have been described on existing shelves (Field et al., 1981; Berne et al., 1988). Deposition would appear to be similar to that of Zones 2-3 of the model applied to the Dalradian Jura Quartzite described by Anderton (1976), in which a significant supply of coarse sediment would be required, unlike the sediment starved, or tidal-swept shelves of more recent examples (Kenyon & Stride, 1970; Kenyon, 1970). This
Figure 2.15 Models for the cross-stratified sequence in Fig. 2.11. Each model is presented with the equivalent interpretation of the seismic sequence. For discussion see text.
sediment supply may be reworked from the underlying Lignitic Sand Unit sediments. Controls on sediment distribution are, in this case, the influence of tidal and storm currents, and the sand waves deposited in the shoreline and high-energy current-dominated shelf zone of Johnson (1986). Each sequence boundary may represent major periods of storm surge and the consequent modification of the shelf sediments.

The sand wave models produced by Allen (1980) are significantly more complicated than the seismic configuration. However it is possible to equate the horizontal surface between the cross-stratified units as E1 erosion surfaces, produced by the planation of the seabed in front of the sand wave. The cross-stratification thus equates to the broad appearance of the internal structure of the sand wave produced by E2 and E3 surfaces (Allen, 1980). The dips are broadly consistent with the Class V/VI sand wave, implying a fluctuating symmetrical tidal flow.

2.7 Reworking in BGS 80/03, 81/16 and 81/17

Reworking of flora was described within the boreholes above. It is possible to define this reworking into two distinct phases which are broadly correlatable within the three boreholes. The importance of the reworking is that it allows quite specific source areas to be delimited. Sediments outside of these two phases can be assumed to be entirely derived from the rocks that compose the Shetland Platform.

2.7.1 Phase 1 Reworking

This involves late Middle Eocene reworking of:

80/03 Jurassic and Upper Jurassic flora
81/16 Upper to Lower Jurassic with Upper Cretaceous (Cenomanian to Campanian) and Mid-Eocene to Palaeocene
81/17 Callovian-Kimmeridgian, ?Toarcian, Middle to late Jurassic microspores and early Cretaceous dinoflagellates.
Numbers of reworked flora appear to reach a maximum above the base of the Upper Main Sand Unit (at 50m in 80/03; 90m in 81/16; approx. 111m in 81/17).

By the late Middle Eocene all of the East Shetland Basin, Viking Graben and almost all of the East Shetland Platform were covered by Tertiary sediments (e.g. Day et al., 1981; Lovell, 1984), precluding these areas from having been reworked at this time. The only areas that are feasible for reworking are those beyond the present subcrop limit of the pre-late Middle Eocene. Since the subcrop limit for these sediments is that of the Lower Main Sand Unit, the only source would therefore be the presently exposed areas of the platform. Consequently, the only sediments of the reworked types that are present are those in the Unst Basin, to the NNW of 80/03 and 81/15, and SE of 81/17 (Fig. 2.16). The only problem with this source is that far from there being no Maastrichtian over the basin - which was probably removed during the Palaeocene - there is no Upper Cretaceous at all mapped over the Unst Basin (Johns & Andrews, 1985). However, the Pre-Tertiary succession is difficult to trace from the East Shetland Basin to the Unst Basin, and the succession to the east of the Tertiary subcrop limit has not been sampled within the Unst Basin. It is possible that a thin Upper Cretaceous cover existed over this basin, which has subsequently been removed from areas outside the Tertiary limit. Within the Unst Basin, it is possible to delimit the probable source area to the North-western half-graben, which is the only area of Mesozoic sediment beyond the Eocene subcrop limit. This half-graben contains a sequence from Lower Jurassic, in the extreme west, to Lower Cretaceous in the east against the fault. The sediments are presently dipping to the east, and hence all exposed at the seabed (Johns & Andrews, 1985). Exposure may have been such that only the most westerly sediments were reworked. Sediment transport directions to 80/03 and 81/16 therefore represent movement of material from at least 60 Km from the north-west. Flora in 81/17 may simply be a function of the proximity (37 Km) of this borehole to the source area.
2.7.2 Phase 2 Reworking

This phase involves Middle Eocene to Lower Oligocene reworking of:

- 80/03 (?Lower) Cretaceous, Jurassic, Carboniferous and Palaeocene to Middle Eocene flora in the Main Sand Unit above the top Eocene unconformity, and the Upper Claystone Unit (Oligocene).
- 81/16 Similar flora, but only within the Upper Claystone Unit.
- 81/17 Early Cretaceous, middle to late Kimmeridgian, Lower to Middle Jurassic dinoflagellates, Middle to late Jurassic and Middle to late Carboniferous spores. These are present in the Main Sand Unit/Glauconitic Silt Unit, above the intra-Middle Eocene hiatus.

The presence of Mesozoic flora is again compatible with derivation from the Unst Basin. Middle Jurassic dinoflagellates are observed in the Unst Basin (Johns & Andrews, 1985), in association with a marine influence in carbonaceous lagoonal swamp shales and sands. Kimmeridgian deposition of dark radioactive shales occurs within the basin. This is similar to the East Shetland Basin which has already been discounted for the earlier phase of reworking. The microfossil types are therefore consistent, if not diagnostic, of an Unst Basin source.

However, the presence of Carboniferous spores poses an enigma as there are no sediments of this type outcropping or subcropping on the platform, which might be recognised as a source. There is evidence of reworking of Carboniferous microfossils during the late Cretaceous to early Palaeocene in the West Shetland Basin, in wells as far apart as quadrants 205S and 209N (Hitchen & Ritchie, 1987). Thus there is an indication of deposition of Carboniferous sediments over an area far wider than the Clair oilfield on Rona Ridge - the only region where Devonian-Carboniferous sediments have been proved this far North. Unless Rona Ridge was exposed during the late Cretaceous to early Tertiary - which is unlikely since even Clair has a Maastrichtian to Santonian succession - there is only one area which can be postulated to have been a Shetland Platform Carboniferous source. This is the Sandwick Basin, adjacent to the
Walls Boundary Fault, west of the Unst Basin. The succession within this basin has been sampled by only one BGS borehole, 78/10, and the thin soft, red sandstone sampled, has been ascribed a tentative Devonian age (Evans et al., 1981), and a Permo-Trias age (Hitchen & Ritchie, 1987). Even taking a Devonian age, it seems unlikely that Late Carboniferous sediments would still be present after having been potentially exposed during the whole of the Mesozoic and early Tertiary.

A more likely possibility is that the Carboniferous flora are the result of reworking of those Palaeocene sediments containing an already reworked Carboniferous flora, and whose equivalents are observed in the West Shetland Basin. This is supported by the presence of extensive Palaeocene sediments with reworked Carboniferous in the West Shetland Basin (D W Jolley, pers. comm.), and the Palaeocene to Middle Eocene flora observed in association with the Carboniferous in the BGS boreholes. The implication of this is that Phase 2 represents the rejuvenation of a region of Palaeocene-Eocene deposition, during a late Middle Eocene, and a subsequent Oligocene phase. It is unclear as to where this area might be on the platform, but the implication of an Unst Basin source for the Mesozoic flora and dinoflagellates, suggests the area NW of this basin, in quadrant 209SE (Fig. 2.16).

In conclusion, the reworking of Palaeocene sediments into the southerly boreholes only, and the absence of reworking in the Upper Main Sand Unit of 81/16 may suggest controls imposed by a tilt of the platform towards the south-east. Similarly, the earlier occurrence of Carboniferous flora in 81/17 may reflect a proximity to source.

2.8 A correlation to the Hampshire Basin and Haq (1987) sea level curve

Biostratigraphy (Fig. 2.17) allows a correlation to be made between the marginal succession of the Middle Eocene East Shetland Platform,
Figure 2.16 Reworking events in the BGS boreholes and the interpretation of the source areas and sediment transport directions.
and a similar succession in the Hampshire Basin (Fig. 2.18). This has significance for the processes influencing deposition. The Hampshire basin has been tied to global sea level changes by Haq et al. (1987) and Flint (1988). A comparison between the two similar settings allows an analysis of the importance of sea level changes during Middle Eocene deposition on the margin of the northern North Sea.

The Lignitic Sand and Interbedded Silt/Sand Units are equivalent to the upper Wittering Division of the Bracklesham Group in the Hampshire Basin (Fig. 2.18) (Flint, 1988). The deltaic sequence in the early Middle Eocene of the northern North Sea correlates to the transgressive lagoonal succession in the upper Vittering Division of Flint (1988). The subsequent abandonment and inundation of the margin is very similar to the marine transgression and shelf sedimentation in the lower Earnley Division.

A more equivocal correlation occurs when an attempt is made to correlate the Main Sand Unit to the Hampshire Basin stratigraphy. This is most notable at the base of the Lower Main Sand Unit. Where Flint (1988) indicates a regional transgression followed by erosion (T3-E4; Figure 2.18), the northern North Sea margin shows a condensed Zone B-3 sequence. An erosive base to the Lower Main Sand Unit can only be inferred. The Hampshire Basin event is characterised by channelling in estuarine sediments removing about 10m of sediment at Alum Bay (Flint, 1988). Reworking of the Lignitic Sand and Interbedded Silt/Sand Units, or simply little erosion of the top of Zone B-3, may account for the continuation of this zone into the Lower Main Sand Unit. Transgressive-erosive sequence T3-E4 may therefore be present at the base of the Lower Main Sand Unit.

The significant horizon at the base of the Middle Main Sand Unit, in both 81/16 and 80/03, is a glauconite zone. Micropalaeontology correlates the Middle Main Sand Unit to the Earnley Division of the Bracklesham Group in the Hampshire Basin. The glauconite horizon, however, does not correspond to either of Flint's (1988) E4 or T4
Figure 2.17 Biostratigraphic correlation for the Eocene sedimentary units defined in the three BGS boreholes. Biostratigraphic zonation is that of Bujak et al. (1989).

Hampshire Basin (Hunt, 1989)

Relative change of coastal onlap (Hug et al., 1987)

Figure 2.18 Correlation between the Hampshire Basin Eocene succession and the northern North Sea and global sea level curve. Dashed unconformities = transgressive markers (marked T-15 for Hampshire Basin). Solid unconformities = erosive events (E2-E5 in the Hampshire Basin) (after Pilt, 1986).
marker horizons. Indeed, the Lower and Middle Main Sand Units correlate with the sequence between these markers. This glauconite event may therefore have a more tectonic and basinal significance, marking a minor erosion followed by transgression, irrespective of global sea level changes. This regression may be the result of a relative, regional uplift of the East Shetland Platform at this time.

The Upper Main Sand Unit is equivalent to the Lower Barton Beds of the Hampshire Basin and is associated with the absence of a number of floral sub-zones at the base. This event is recognised in 81/17 as the seismic sequence boundary separating the two prograding wedges. Reworking is important within the Upper Main Sand Unit, and rejuvenation of the source region around the Unst Basin appears to have taken place. This suggests a hiatus through the equivalent of the Selsey Division.

The main features of the northern North Sea Middle Eocene succession correlate, with the exception of one glauconite event, with the Hampshire Basin succession. The Hampshire Basin has been used in the correlation of the Haq et al. (1987) global sea level curves. The further correlation to the northern North Sea supports a similar control, whether global or not, for the northern North Sea and the Hampshire Basin coastal margins.

2.9 Conclusions

The Eocene sediments at the western edge of the East Shetland Platform can be sub-divided into a series of lithological and biostratigraphical units (Fig. 2.17). Facies analysis indicates that the sediments represent the coastal marginal equivalent of basinal shales observed in the centre of the northern North Sea. The Lignitic Sand and Interbedded Silt/Sand Units record the progradation of a tidally dominated delta across the platform during the early Middle Eocene. Sediment transport was into the South Viking Graben, towards the south-east. Clay mineralogy was
dominantly environmentally controlled. Subsequent transgression and tidal reworking of the deltaic sediments followed during the middle to late Middle Eocene, across the central and southern parts of the platform. This reworking was associated with deposition of the Lower and Middle Main Sand Units. Sands were derived from the northwestern part of the platform in the late Middle Eocene, based on reworking events in the biostratigraphy. The succession can be biostratigraphically correlated to the Hampshire Basin, where global sea level changes have been recognised as the dominant control on sedimentation. In detail, though, one marker horizon is not correlatable, and may represent an episode of regional uplift of the East Shetland Platform.
3.1 Introduction

The previous chapter outlined the evidence for a progressive transgression and submergence of the East Shetland Platform margin during the Middle Eocene. Comparisons between the East Shetland Platform and the Hampshire Basin showed a series of inter-regional unconformities. It is possible to assess the significance of these events in the basin evolution, using a seismic stratigraphic approach. Consequently, this chapter outlines the seismic interpretation of a number of commercial multichannel seismic lines. Locations of examples presented are given in Fig. 3.1. The interpretation of the lines is accompanied by the seismic and sediment facies analysis. A relative sea level curve, based on the interpretation is also presented.

Seismic interpretation has an element of subjectivity in that no two interpretations, by different interpreters, are likely to be identical. The credibility of a seismic model is very much controlled by the quality and quantity of evidence supporting the model. This supporting evidence and the regional interpretation of the seismic stratigraphy are discussed.

The Eocene succession has been sub-divided into 6 seismic stratigraphic units. These are assigned the informal names of - from lowermost to uppermost Eocene - Ela, Elb, Elc, E2, E3 and E4 (Fig. 3.2). A summary of the description of the seismic sequences is given in Table 3.1.

3.2 Seismic sequence Ela

For the purpose of this study the base of the Eocene has been defined by the Ash Marker horizon. As the base of the Eocene is not strictly
Figure 3.1. Location map for the seismic sections and line interpretation shown in the succeeding figures.
Table 3.1
Characteristics and interpretations of seismic sequences and facies units in the Eocene of the East Shetland Basin

<table>
<thead>
<tr>
<th>SEISMIC SEQUENCE</th>
<th>EXTERNAL FORM</th>
<th>INTERNAL CONFIGURATION</th>
<th>BASEAL REFLECTOR TERMINATION</th>
<th>NATURE OF BASEAL REFLECTOR</th>
<th>REFLECTOR AMPLITUDE</th>
<th>REFLECTOR CONTINUITY</th>
<th>COMMENTS</th>
<th>INTERPRETATIONS</th>
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<tbody>
<tr>
<td>E4</td>
<td>WEDGE</td>
<td>PROGRADING</td>
<td>DOWNLP</td>
<td>PLANAR</td>
<td>MODERATE</td>
<td>CONTINUOUS</td>
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<td></td>
<td>BASE OF SLOPE SLUMPS</td>
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<td></td>
<td>SHEET</td>
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<td></td>
<td></td>
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<td>EAST SHEETLAND PLATFORM</td>
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<td>DOWNLP</td>
<td>PLANAR</td>
<td>MODERATE</td>
<td>CONTINUOUS</td>
<td></td>
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<td>DIS-CONTINUOUS</td>
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<td>DIS-CONTINUOUS</td>
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<td>DIS-CONTINUOUS</td>
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<tr>
<td></td>
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<td>DOWLP</td>
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<td>DIS-CONTINUOUS</td>
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<td></td>
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<td>1</td>
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<td>DOWLP</td>
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<td>MODERATE</td>
<td>TO HIGH continuous</td>
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<td>DIS-CONTINUOUS</td>
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<td></td>
<td>3a</td>
<td>MOUND</td>
<td>CHAOTIC WITH</td>
<td>IRREGULAR</td>
<td>LOW TO HIGH</td>
<td>DIS-CONTINUOUS/CONTINUOUS</td>
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<tr>
<td></td>
<td>3b</td>
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<td>DOWNLAPPING</td>
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<td>CONTINUOUS</td>
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<td></td>
<td>3c</td>
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<td>MODERATE</td>
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<td>IRREGULAR</td>
<td>HIGH</td>
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72
Figure 3.2 Seismic section and line interpretation showing the Eocene seismic stratigraphy described in Chapter 3. Location map at bottom is enlargement of Fig. 3.1.
defined by either palaeontological or lithostratigraphical events, there are discrepancies in the position of the Ash Marker horizon within the chronostratigraphic record. Deegan & Scull (1977) assign the tuffaceous Balder Formation as Palaeocene to early Eocene. Heritier et al. (1979) assign the tuffaceous sands in the Frigg sand to the Sparnacian to Ypresian, preferring to place the tuffs in the upper Palaeocene (Heritier et al., op. cit.; their Figure 4). On the other hand, Sarg & Skjold (1982) assign the tuffaceous sands, silts and clays in the Balder field, to the lower Ypresian. More recently, Stewart (1987) assigns his tuffaceous sands and silts of seismic sequence 9 to the early Ypresian, based on a change in palynomorph assemblages. This confusion is reflected in the released well data used in this study, the position of the biostratigraphically defined Palaeocene-Eocene boundary reflecting the preference of the individual exploration companies. Some confusion is caused by the tuff deposition being more than a single instantaneous event. Indeed phases of tuff deposition occurred throughout the Sparnacian and Ypresian (Jacque and Thouvenin, 1975; Knox & Morton, 1983). The period of major tuff deposition appears to have been 53-54.5 Ma. To remove the discrepancies produced by the well data and the uncertainties described above, the base of the Eocene was defined lithostratigraphically as the base of the tuffaceous claystones. This was on the basis that this lithostratigraphic event is recognised in the majority of wells.

The Ash Marker, corresponding to sequence Ela, is a high amplitude continuous seismic reflector observed throughout the East Shetland Basin. Reflection amplitude decreases onto the East Shetland Platform and in the South Viking Graben, where the Ash Marker lies within the submarine fan sands of the late Palaeocene and early Eocene Frigg Formation (Deegan & Scull, 1977; cf. Sarg & Skjold, 1982). This relates to reworking of the tuffs in areas of moderate to high energy conditions, either shallow marine or turbiditic. On the East Shetland Platform the extent of the Ash Marker is discernible only to the Palaeocene subcrop limit. West of this, it is lost within the high amplitude basement reflections, if indeed it onlaps any further onto
the platform. The tuffs are, in general, 10-20m thick, and so the reflector accounts for both the top and base of the tuffs. Little of the internal configuration of the Ela seismic sequence is discernible. Thin, downlapping reflectors are rarely seen.

Sediment grading in the volcanic beds, in the East Shetland Basin, suggest the aeolian fall of the pyroclastic ash (Jacque & Thouvenin, 1975). Interbedded non-tuffaceous sediments are quartzose with poorly crystallized glauconite, and associated with pyritized Coscinodiscus sp.1. Therefore, conditions were entirely marine, favouring the preservation of the tuffs. A proximal, Hebridean source has been favoured (Knox & Morton, 1983) for the volcanism.

3.3 Seismic sequence Elb

The overlying seismic sequence extends from the edge of the East Shetland Basin across the whole of the East Shetland Basin, and into the Viking Graben, reaching a maximum thickness of 150-200m. The unit can be divided into a number of recognisable seismic facies, based on the contacts with the upper and lower units, and the internal configuration of the sequence (Fig. 3.3).

3.3.1 Quadrants 2, 3, 8 and 9

**Facies 1.** The unit Elb is represented by a thin series of downlapping reflectors at the edge of the East Shetland basin. Reflectors are moderate amplitude and of variable continuity. The internal configuration of reflectors show oblique progradation.

This facies is interpreted as a prograding shelf and slope system.

**Facies 2.** (Fig. 3.4) Within the East Shetland Basin, the unit becomes dominated by low amplitude discontinuous reflectors. These are observed to onlap the Palaeocene shelf-edge (Rochow, 1981). Occasionally, poorly defined downlapping and concave upwards reflectors are observed.
Figure 3.3. Map showing the distribution of seismic facies units through the Elb seismic sequence.
Horizontal dash - onlap;
Diagonal dash - downlap;
Wavy solid - mounded facies with numbers indicating the facies type of mounds;
Horizontal solid - horizontal concordant reflectors.
Figure 3.4. Seismic line showing the onlap of Elb seismic facies 2, onto the palaeotopography produced by the Upper Palaeocene Moray Group shelf deltaics, as defined by Rochow (1981). Location shown on Fig.3.1.
Figure 3.5 Seismic line with mounded facies 3 of Elb, showing two of the different types of mound observed.
Figure 3.6 Comparison and interpretation of the Elb facies 3 mounds to observed present day and ancient submarine fan systems.
The low amplitude and discontinuity of reflections suggests that this facies is dominantly silty. The onlap therefore appears to be marine, rather than coastal (Mitchum et al. 1977), and represents the inundation of the late Palaeocene shelf-deltaics, and the progressive infilling of the basin topography that these produced.

**Facies 3.** (Fig. 3.5) Within the East Shetland Basin a number of high amplitude reflectors can be observed, mostly concentrated around quadrants 9N and 3S. These are mounded, with reflectors downlapping to the west and east. The mounded facies has a maximum thickness of 150-200m with, in 3/24, an undulose upper surface with a wavelength of 2.5Km and present relief of approximately 40m. Within the mounds the reflections are discontinuous and chaotic at the centre, becoming more continuous on the flanks and dipping away from the centre (Facies 3a). These appear to be single channel and levee systems (cf. Droz & Bellaiche, 1985) (Fig. 3.6a). Towards the east the mounds have a longer wavelength and lower relief (Facies 3b). Internal configuration shows parallel undulose reflectors, with occasional horizontal, discontinuous onlapping reflections. Reflectors are more continuous than within the mounds and signal amplitudes are higher. Facies 3b is interpreted as a series of suprafan lobes (cf. Sarg and Skjold, 1982) which overlie facies 3a (Fig. 3.6b), representing fan retrogradation. Towards the north, concentrated in 3/18 and 3/19, reflectors are again low amplitude, low angle concave upwards (Facies 3c) (Fig. 3.6c). These reflectors have continuity for only 0.5km, throughout the mounded sequence, with the appearance of overstepping channels, as might be expected for a meandering channels system (cf. Damuth et al., 1983).

This facies is interpreted as the development of a submarine fan system in E1b. The high amplitudes and the laterally varying reflection configuration support this interpretation. Similar mounded structures have been described in the early to middle Thanetian, by Sarg & Skjold (1982). Similarly, the lower Eocene submarine fans forming the Frigg field show such a mounded relief (Heritier et al, 1979).
3.3.2 Dipmeter interpretations of Facies 3

Well ties support the above interpretation. Wells 3/19-1 and 3/25-1 (for location see Fig. 1.5) contain some dipmeter data which is consistent with the lateral and vertical variations in deposition in these mounded facies of Elb. The released dipmeter logs are of poor quality, but even within the scatter there are recognisable patterns and trends in dip orientations.

Well 3/25-1 (Fig. 3.7) divides into six sequences. The upper dipmeter unit is composed of scattered dips with no recognisable patterns, the lithology of this unit being a dark grey clay and thin limestone streaks. This can be interpreted in terms of pelagic sedimentation, probably in the seismic sequence Elc, above the Elb submarine fan system. Subsequent dipmeter units are dominated by single dip increase upwards patterns indicative of erosive surfaces of channels (Selley, 1979), supported by the dominance of sands within this succession in 3/25-1. Dip directions vary from NE to E. One unit of scattered dips is interpreted as a channel fill. Absence of stratification suggests rapid deposition as a debris flow.

Lithologically, the sequence is composed of white, fine-medium grained sand with thin grey shales and limestones.

A core in dipmeter unit V of the Elb sequence corresponds to one of the dipmeter units that exhibit a single dip-increasing downwards pattern. This pattern implies a channel-fill sequence (Selley, 1979). The core is poorly preserved but consists of a fine to medium grained sandstone, occasionally calcite cemented, with intraformational shale clasts. Shales are found at the base of the core. The rapid vertical change in lithology, and the evidence for reworking is consistent with a channel infill by a density flow.

Well 3/19-1 (Fig. 3.8) similarly divides into 7 recognisable dipmeter units within the Elb sands. The lowermost sequence consists of 2-7m thick dip-increasing downwards patterns (dipmeter sequence VIII-1798m to 1823m). Dips are towards the SE and NW. That infill is both silty and sandy, rather than a sand dominated series of stacked
Figure 3.7 Dipmeter interpretation for sequence Elb in well 3/25-1.

a) Seismic configuration. Box shows position of dipmeter reading.

b) Dipmeter log with sonic and gamma ray logs.

Well No.: 3/25-1
Figure 3.8 Dipmeter interpretation for the Elb and E1c sequence in well 3/19-1, with sonic and gamma ray logs.
channels, suggests fluctuations in the energy of the submarine fan system during its evolution. Channels may have infilled by debris flows down the system, with silty backfilling of abandoned channels. This is overlain by a layer of horizontal dips (dipmeter sequence VI - 1762m to 1764m), followed by scattered dips. The uppermost unit, from 1663m to 1733m, contains scattered dips and overlies a sequence of 1-2m thick dip-increasing downwards layers (1733m to 1762m). Dip directions are scattered and orientated towards the SW and SE. These are separated by 4-5m of scattered dips, suggesting stacking of channels with axes oriented NW-SE (Selley, 1979). These patterns coincide with both sandy and silty units, and suggest that the patterns represent stacked channels in a mid-fan/lower fan setting (Selley, 1979; cf. Sarg & Skjold (1982)).

The submarine fans developed in the Elb seismic sequence are widespread. In wells tied to the seismic coverage, sands in Elb are observed both in the basin and at the edge of the East Shetland Platform. The latter appear to be isolated fan sands and may represent parts of feeder fans to the main basinal system, similar to the La Jolla Fan, California (Hand & Emery, 1963; Graham & Bachman, 1983). No other dipmeter data was available through the sequence but the predominance of a SE transport direction suggests that fans were derived from the East Shetland Platform. There is no evidence to suggest that any part of the Elb fan system - in the UK sector - was derived from the Norwegian sector (McGovney & Radovich, 1985).

3.3.3 Quadrants 210 and 211
To the north, seismic sequence Elb is restricted to the basin and does not onlap onto the East Shetland Platform. The sequence is thins over the North Shetland Platform. Internally, reflectors are mostly parallel, of moderate amplitude and continuous over the most of the region. The sequence appears to be a drape of sediment over the pre-existing structure and wells tied to the sequence suggest that basin sedimentation is dominated by claystones. At the southern edge of the Magnus Basin a wedge of moderate amplitude, continuous reflectors in Elb. These downlap onto the Ela relector (Fig. 3.9). The nearest well
Figure 3.9
Seismic line showing the downlap of Elb into the Magnus Basin, off the northern edge of the East Shetland Platform.
tie is 210/13-1, at the toe of this downlap, which records only a sequence of olive-grey mudstones coarsening to siltstones. This appears to represent a muddy shelf sequence similar to Facies 1, described in quadrants 2, 3, 8 and 9.

There is, however, evidence of some fan activity this far North (Fig. 3.10). An undulose, chaotic series of reflectors ties to sands in well 210/29-1, and mudstones in 210/29-2 (Fig. 3.10a). The internal configuration is mounded in the lower part of Elb, with concave-upwards reflectors concentrated in the upper part of the sequence and separate from the mounds. The interpretation is of mid- to lower fan sand lobes overlain by a prograded mid-fan channel system. The interpreted fan system is juxtaposed against the basin-flanking fault and restricted to the two wells above (Fig. 3.10b). Sedimentation is again claystones in 210/30 and 210/25, where thin sands are also observed. These may represent lower fan sand lobes and mark the distal part of this fan system.

On seismic evidence alone, there appears to be a broad fan system developed in the centre of the Magnus Basin. The isopach map (Fig. 3.11) indicates a lobate shaped thickening of the Elb sediments (Elc being thin across Magnus Basin). On seismic lines this appears as a wedge of concordant reflectors. This wedge consists of a series of smaller wedges stacked on top of each other (Fig. 3.12). Reflections are discontinuous although occasionally mounded and downlapping reflections are observed. The lobe thins under the present Tampen Spur feature. This initially aroused suspicion that this lobate shape was a series of multiples, but the slope of the top Elb surface is very much less than that of the recognisable multiples in the section above. From the isopach map and the dip of reflections the fan appears to derive from the southern and eastern parts of quadrant 210, where smaller fans are also observed. It is likely that these are part of the same feeder system transporting sediment from the East Shetland Platform towards the NE.
Figure 3.10
Submarine fan deposition off the northern part of the East Shetland Platform. Seismic line with well tie.
Figure 3.11 Isopach map of Elb over the Magnus Basin. Contours in metres, depth conversion based on an average velocity of 2350 ms\(^{-1}\). Positions of wells around the thick lobe are also shown.
Figure 3.12
Internal configuration of the "Tampen Fan" observed on the isopach map of Fig. 3.11.
No wells penetrate this lobate structure, the nearest, well 211/7-2, being located in the area of thin El sediments south of Magnus Basin. This records a sequence dominated by pale grey and green mudstones. Thin, well-sorted sands are observed, with the gamma ray signature suggesting coarsening upwards sequences (Selley, 1979). This may represent a lower fan facies, dominated by pelagic sedimentation with thin distal parts of sand lobes (cf. Nelson, 1976). Claystone-dominated lithologies are observed in wells 210/15-1, 210/15-2, 210/20-1 and 211/12-6. This precludes the sands in 211/7-2 from being sourced from fans to the South or due East, and restricts the sands to being derived from a fan system to the North of all of these wells.

3.4 Seismic sequence Elc

In quadrants 2, 3, 8 and 9 the submarine fan systems of sequence Elb are overlain by a less varied seismic facies (Fig. 3.13). The mounded facies are absent and the sequence is overlain by a downlapping series of low amplitude continuous reflectors which prograde across the whole of the East Shetland Basin. The seismic facies are more regionally controlled, and change laterally from the East Shetland Basin to the East Shetland Platform.

3.4.1. East Shetland Basin and Viking Graben

Low amplitude reflections dominate the Elc sequence in the basins, poorly defining a series of downlap onto the Elb submarine fans. Wells 9/08-4, 9/19-3, 3/19-1 and 3/25-1 tied within the East Shetland Basin record only a dark grey to grey-green mudstone sequence for Elc. The dipmeter log for well 3/19-1 (Fig. 3.8) records a series of dip-increasing upwards structures in the claystones. Dip orientations are orthogonal to those in Elb in 3/19-1, being dominated by SW-NE directions. This is supported by seismic evidence of a mounded facies locally around the well (Fig. 3.14). The mound appears as a series of discontinuous reflectors east of the maximum extent of the downlapping reflectors and where Elc is thinnest. These mounds may represent muddy turbidites or contourites (Stow & Lovell, 1979).
Figure 3.13: Seismic traces map for the Fig. 3 sequence.

Shading as for Fig. 3.3.
probably the former in that the mounds do not exhibit the regular internal configuration of Sangree & Widmier (1977) or Gonthier et al. (1984) (Fig. 3.14). The dipmeter configuration is also very similar to that of mud lobes in the Jurassic of the Brae Field (Stow et al., 1982). Dip directions may be related to a lateral position on the fan lobe. The absence of sandy contourites and the low relief compared to many present day examples (Gonthier et al., 1984) suggests that bottom currents were not well developed and were stable (Gonthier et al., 1984). The thinning of sequence Elc east of this mounded facies suggests low sedimentation in the centre of the basin at this time, but corresponds to the thickening of submarine fan sediments in sequence Elb. Consequently, there is no evidence of bottom current reworking of submarine fan sands into Elc.

In quadrants 210 and 211, sequence Elc is difficult to trace but appears to become thin east of the downlapping wedge seen in Fig. 3.10., compared to Elb. Again, this may be the response to low sedimentation rate, and silt deposition during sequence Elc.

3.4.2 Elc on the East Shetland Platform and the timing of canyon development

The sequence Elc extends onto the East Shetland Platform, and is traceable to the Eocene subcrop limit. On the East Shetland Platform sequence Elc consists of a series of moderate amplitude, gently downlapping reflections. Occasionally, smaller scale downlaps are observed within the overall configuration.

This sequence ties with the lower part of BGS 81/16, described in Chapter 2. This suggests that the East Shetland Platform Elc sediments are probably the continuation of the deltaic sequence, but include the shelf succession of the Lower and Middle Main Sand Unit. The moderate amplitude reflections therefore represent sandy shelf progradation across the East Shetland Platform with development of sand waves within the succession, extending as far as the edge of the East Shetland Platform.
Figure 3.14. a) Internal configuration of E1c tied to well 3/19-1. b) & c) are interpretations based on published details of contourite mounds are also presented, and highlight the dissimilarities between these and a).
Over much of the East Shetland Platform sequence Elc overlies Palaeocene sediments or basement rocks. A number of concave upwards incisions, approximately 2Km wide and 50m deep, can be observed below the Elc reflections (Fig. 3.15). These are traceable across the platform and show a NW-SE trend with little evidence of meandering. The infill to the incisions show an onlap onto the walls of these basement features.

The incisions are interpreted as canyons, related to submarine fan development. Whilst the canyons are overlain by Elc sediments, this does not necessarily signify an Elc age for the construction and utilisation of these features. It may simply signify that the canyons became inactive only at this time. Similarly, the Elc age does not imply that sediments infilling the canyons are of this age. There may be a thin infill of Elb to Palaeocene sediments that would signify an earlier age for canyon development.

There is, however, other evidence which constrains the timing of canyonisation of the platform. The canyons are located high up the platform, east of the Palaeocene subcrop limit (Fig. 3.16). There is indeed no evidence of canyons incised into the basement below the Palaeocene. The last stage of Palaeocene submarine fan development is observed in the Witch Ground Graben during the middle Thanetian (Stewart, 1987), and is overlain by the shelf-deltaic sediments corresponding to the Moray Group (Deegan & Scull, 1977; Rochow, 1981; Mudge & Bliss, 1983). For the canyons to have been cut during the mid-Thanetian, preservation is required during the Moray Group regression; the subsequent (or contemporaneous) exposure of the western margin of the East Shetland Platform, until the deposition of Elc to the present Eocene subcrop limit. Similarly, if the canyons had remained open from the mid-Thanetian, then a more complex series of fills might be expected. These would include a secondary series of erosive surfaces within the canyon. The absence of any high amplitude reflections between the base of the canyons and the top of the Elc succession suggest that the time between the lowermost uppermost
Figure 3.15. Seismic section of canyons on the East Shetland Platform. Infill consists of onlapping and occasional downlapping reflectors of Elb?-E1c sequence.
upfill was not that great, and did not involve a complicated evolution in between.

The top of the Palaeocene in blocks 8/5 and 9/3 shows a broad undulation which truncates Palaeocene reflections and thins the underlying sediments. This represents a low, broad incision into the top Palaeocene, indicating a post-Palaeocene regional erosion surface. The overlying sequence is again Elc and it is concluded that canyon development occurred during the deposition of Elb. Support for this is provided by the close correlation between the "Frigg-equivalent" fan distribution of Heritier et al. (1979) and the positions of the East Shetland Platform canyons (Fig. 3.16).

3.5 Seismic sequence E2

The transition between the Elc and E2 sequences is subtle. There are great similarities between the two sequences and the sequence boundary is low amplitude. This is especially true in the centre of the East Shetland Basin, where reflections are conformable. However, a common surface for downlap is defined by the infrequent seismic events in E2. The sequence is restricted to the East Shetland Basin and Magnus Basin, and does not extend onto the East Shetland Platform (Fig. 3.17).

The internal configuration of sequence E2 consists of low amplitude discontinuous reflections. Occasionally, more continuous downlapping reflections are observed and onlap onto the Elc sequence occurs at the basin margins. High amplitude reflections are rare and similarly discontinuous.

The Unst Basin proves an exception to this, however, and the E2 sequence is observed as a thin layer onlapping at the Eocene subcrop limit. A thin moderate amplitude unit is also correlated to E2, on the north-eastern edge of the Unst Basin (Fig. 3.18). The nearest well tie, 210/19-1 records only the siltstones at the toe of the
Figure 3.16 Two way time contour map for the reflector of the East Shetland Platform, below the Paleocene to Eocene succession. Shading indicates the position of the Lower Eocene submarine fans from Heritier et al. (1979). Sub-Cretaceous faults are marked to define the East Shetland Platform and East Shetland Basin.
Figure 3.17 Map of the facies distribution of seismic sequence E2. Shading as for Fig. 3.3.
Figure 3.18 Seismic section showing the downlap of E2 at the edge of the Unst Basin. This is interpreted as a barrier beach succession. Nearest well tie, 210/19-1, is located at the toe of the downlap sequence.
downlap. An adjacent well, 210/19-2, within the downlap facies samples a light brown and grey calcareous mudstone.

The well ties throughout the East Shetland Basin all show the same lithology. Sequence E2 consists of dark grey and brownish-green mudstones becoming olive-green and micaceous further north in quadrant 210. In addition there is a prominent increase in the amount of limestone observed, in the form of both stringers and thin, 2-3m, beds. Thin sands observed at the base in E2 in 9/19-3 may represent thin turbidites, which fine upwards according to the gamma ray log (Selley, 1979). These may, however, be at the top of E1c, rather than base of E2, and are very localised.

Despite its restricted distribution, sequence E2 is dissimilar to the E1 sequences in that it (see enclosure) does not exhibit any controls by the underlying structure.

3.6 Seismic sequence E3

The boundary between seismic sequences E2 and E3 is a moderate amplitude reflection across which the amplitude and continuity of reflections increase from E2 to E3. The sequence E3 encroaches onto the East Shetland Platform where it directly overlies E1c (Fig. 3.19). The observed seismic disconformity on the East Shetland Platform is consistent with the intra-Middle Eocene disconformity recorded in the BGS boreholes. It is possible to equate the upper Main Sand Unit with 80/03 and 81/16 to the seismic sequence E3, and the missing section to E2.

On the East Shetland Platform, E3 is composed of moderate to occasionally high amplitude reflectors which gently downlap onto the E1c sequence. These downlaps have an apparent dip towards the east. Overall the succession on the platform changes from one of continuous downlap - relating to shelf progradation - to one of localised, small scale submarine fan deposition. These do not appear to be related to any channelling, but the absence of a marginal sequence in E2 on the
Figure 3.19 Map of the facies distribution of seismic sequence E3. Shading as for Fig. 2.2.
platform implies that deposition of the bulk of the E3 sediments on the East Shetland Platform postdates an early E3 planar erosion event.

Further north, on the edge of the East Shetland Platform, sequence E3 is composed of low relief (40m) mounds over thin E2 sediment. In well tie 9/07-1 (Fig. 3.20), such a mound is composed of a greenish grey silty claystone, with occasional beds of medium to coarse grained argillaceous sandstone. Channel facies, similar to E1b facies 3, are not evident around the mound. The mounded nature and claystone-dominated lithology are interpreted as deposition of a low energy, outer fan lobe. Small downlapses adjacent to the well, and in the upper part of E3, are similar to the downlapses further west on the East Shetland Platform. This implies that the shelf sedimentation on the platform postdates the development of the E3 low energy fan systems.

Within the East Shetland Basin, in quadrants 3 and 9, seismic sequence E3 is composed of moderate to high amplitude discontinuous reflections. A number of these discontinuous reflections are concave upwards, defining channel sequences (Fig. 3.21). These channels are isolated and are generally more frequent in block 3/24 although smaller channels occur towards the SE part of quadrant 3. Low amplitude downlap occurs in thin units (approximately 60m), below the channels in block 3/24. This can be interpreted in terms of a fan system, with progradation of middle fan channels over lower fan lobes. Progradation is tentatively inferred to be from the NW, as the lobe facies is elongate along a roughly NE-SW direction, with channels concentrated to the NW.

A feature of the isopach map of E3 is a NE-SW oriented elongate ridge (Fig. 3.22). This occurs to the south-east of the fan system described above. The change in thickness across the feature is about 50m in 7Km. The orientation cuts across the trend of the underlying Mesozoic structure, in part overlying the edge of the Shetland Platform. In well 3/29-1, in the centre of this feature, the lithology of sequence E3 is dominated by pale grey-green mudstones.
Figure 3.20 Well tie 9/07-1 and a thin argillaceous sand mound in sequence E3.
Figure 3.21 Interpretation of seismic line showing the formation of channels within E3, with downlapping infills, suggesting lateral accretion and meandering of the channels.
Figure 3.22 Detail of the isopach map for E3, indicating the lithology of well ties in a NE-SW trending ridge. Depth conversion based on an average velocity of 2100 ms⁻¹.
Figure 3.23 Map of the facies distribution in seismic sequence E4. Shading as for Fig. 3.3. Arrows indicate the direction of downlap.

Figure 3.24 The isopach map for the southern part of sequence E4, with well ties indicated. Depth conversion based on a velocity of 2000ms⁻¹.
and siltstones, with interbedded limestones. These features may simply represent a ridge created by bottom flowing currents modifying the E3 seafloor topography (possibly at the end of E3), rather than the formation of a linear submarine fan system.

Seismic sequence E3 is relatively thin (100m) over the Magnus Ridge and into the Magnus Basin. Reflections are generally of moderate amplitude, parallel, horizontal and continuous. This apparent quiescent sequence contains very fine to medium-grained sandstones in well 210/29-2. This appears on the seismic record as a localised downlap sequence on the edge of the Unst Basin, but the facies of this sandstone is uncertain. It may represent a barrier beach system, but the poor sorting and the presence of lithic grains suggests considerably less reworking than might be expected for shoreface sands.

3.7 Seismic sequence E4

Sequence E4 is restricted in its distribution (Fig. 3.23) and does not extend significantly onto the platform. The distribution breaks down into two locations, far south and far north of the Viking Graben.

3.7.1 Southern part of E4

In quadrant 9 and the southern half of quadrant 3 a thin series of high amplitude reflectors downlap onto the E3 sequence. These form a wedge of oblique prograding reflections that thins towards the East Shetland Platform and the Viking Graben (Fig. 3.24). The wedge is oriented broadly N-S, parallel to the basin axis.

The dominant direction of downlap of the internal reflectors of E4 is towards the east. This pattern is common on lines to the south of quadrants 9N. Further north than this, in blocks 3/23 and 3/24, the downlap is superseded by a small scale channel facies, with low amplitude chaotic reflections at the toe of the downlap package. The configuration suggests the development of a prograding slope, with
minor slumps and debris flows concentrated at the base of the shelf-slope. Channelling may be the result of erosion at the shelfedge, but these do not appear to have been as important, or as large, as the canyons mentioned earlier.

Well ties (9/08-4, 9/17-1a, 9/17-2) indicate that the downlapping sequence is composed of claystones. Sandstones are observed in wells 9/08-4 and 9/19-3 (Fig. 3.25), and are related to the toe of the downlapping shelf. In the case of 9/08-4 (Fig. 3.26), the sands are medium to coarse grained and moderately sorted. Hard pieces of lignite are found in the sands. The seismic configuration of the sands is a sub-horizontal high amplitude reflection at the toe of the downlap, and which is itself downlapped upon by the shelf sequence. No channelling is observed in the sequence or in the underlying top of E3. The sands are thus interpreted as slump deposits. In well 9/19-3, the sands are very fine to coarse grained glauconitic and also contain lignitic debris. The seismic response to these sands is a chaotic series of discontinuous reflectors. Similarly, these reflections are at the base of a downlap sequence, where E4 thins. Hence it appears that these are similar in origin to the 9/08-4 sands.

3.7.2 Northern part of E4
A seismic package is observed above E3 in the north of quadrant 210. This package is disconnected from E4 in the south, as described above, but the position in the seismic stratigraphy and released biostratigraphy is taken to suggest a similar age for both sequences.

This northern package consists of two wedges of reflectors that onlap the underlying Eocene succession over the North Shetland Platform. The wedges are elongate to the E-W for the northerly, and NE-SW for the southerly wedge. The top of either wedge is a high amplitude reflection that terminates basinward with a recognisable steep downlap (Fig. 3.27). The wedges are thin, but internally contain reflectors that downlap the top of E3. Apparent downlap directions
Figure 3.25 Internal configurations of sequence E4 with well ties. a) 9/08-4; sequence as for Fig. 3.2. b) 9/19-3.

Figure 3.26 Lithology of 9/08-4 for Eocene seismic stratigraphic succession.
are towards the north for the more northerly E4 package, and towards the northwest for the southerly package.

No wells tie directly to this sequence, as observed in this study. However, a similar package to the southerly wedge is observed to tie with well 209/9-1 (K. Hitchen, pers. comm.). The lithology at this horizon is a 60m to 90m thick green, medium grained sandstone with subrounded to rounded quartz grains. A characteristic of the sandstone is the occurrence of polished quartz grains. This lithology is also observable in wells 209/6-1 and 209/3-1 (Fig. 3.28).

The continuity of the downlap in the packages suggests that E4 in the north is not of turbiditic origin. The basinward termination of the packages is very clear and suggests a sharp transition from the inferred sandy nature of the wedges to thin time-equivalent shales. A tidal shelf origin is inferred for the sands, similar to that described for the cross-stratified seismic sequence in Chapter 2.

3.8 Evolution of the Eocene submarine fan systems

As has been described, the presence and absence of sands in the Eocene succession (Fig. 3.29) in the East Shetland Basin is directly related to the activity of submarine fan systems. Such systems have been taken by Mitchum et al. (1977) as an indication of lowstand sea level depositional systems. This simplistic approach has been criticized by Miall (1986). Consequently, in attempting any analysis involving sea level, it is imperative that the relative controls on, and importance of, any submarine fan system be investigated before ascribing any particular significance to its occurrence. Therefore an evolutionary history is presented for the Eocene shelf and slope sediments, and the depositional controls discussed.

3.8.1 The B1b fan system

The submarine fan system in B1b can be taken as a major system emanating off the East Shetland Platform and, it is argued, linking to the Frigg Formation sands. The system was established after the
Figure 3.28 Distribution and thickness of the E4 sands based on correlation of the E4 wedge to upper Eocene- lower Oligocene polished quartz sands in wells in quadrant 209. Depth conversion velocity for the E4 isopach map is 2000ms⁻¹.

Figure 3.27 Seismic line showing the northern part of sequence E4. The line interpretation shows the prominent downlap of the sands onto the earlier Eocene sequence. Location of line is shown on Fig. 3.28.
Figure 3.29 Chronostratigraphic diagram showing the seismic, lithological and distributional variations of the seismic stratigraphic units of the Eocene in the northern North Sea.
cessation of fan activity in the late Palaeocene and during deposition of the Ela tuffaceous claystone sequence in the East Shetland Basin. This lowermost sequence is taken as indicating sea level rise. A model linking the Eocene submarine fans from a common canyon source can be invoked (Fig. 3.30). A differentiation of fans, based on position and internal structure, is proposed.

1) "Platform" Fans. These consist of the East Shetland Platform canyons and sands in quadrant 9N and 3S (eg. Bruce; Heritier et al., 1979). This region is very similar to upper fan regions of some of the major present-day fans. Whilst not all fans are fed by canyons (eg. Stow, 1985), deposition in larger systems is controlled in this manner (eg. the Rhone, Nova Scotian and the La Jolla Fans - Droz & Bellaiche, 1985; Hill, 1984; Graham & Bachman, 1983). The dimensions of the canyon and upper fan single channel are very similar to East Shetland Platform canyons. Overbank deposits and levees alongside the upper fan canyons are poorly represented, suggesting effective channelling of all turbidite flows down the canyons. Small scale lobes at the termination of the single channel/canyon system are similar to the fan model of Normark (1978), but in the East Shetland Basin represent only part of a larger system. Termination at this point is likely to be the result of an abrupt decrease in angle of slope at the termination of the canyon onto the eastern edge of the Shetland Platform.

ii) "Basinal" Fans. The middle fan facies is contained in the East Shetland Basin and is composed of numerous feeder channels. The fan systems of this type are West Frigg and S. Alwyn (Heritier et al., 1979). This is again similar to most submarine fans systems, whereby sediment is distributed away from the main feeders (Mutti, 1977). The middle fan as described for Elb is very similar in seismic configuration to the Upper Fan channels of the Rhone (Droz & Bellaiche, 1985). Abandonment and burial of channels would produce a sequence similar to that for the middle Amazon fan (Damuth, 1983). Whilst there is no evidence of channels stacking, the channels similarly do not show any great lateral migration within the sequence, and are larger in relief than the suprafan channels and
Figure 3.30 A linked fan model for the lower Eocene fans of Eib. The model suggests that all the fans were derived from the canyon system on the East Shetland Platform, and that subsequent distribution was controlled by topography of the basin floor.
lobes of Sarg & Skjold (1984). Indeed, the repetition of channel sequences on seismic sections may result from meandering of a single channel across the basin (cf. Garrison et al., 1982). The more frequent channels in 3/18 and 3/19 may represent a middle fan facies more typical of Droz & Bellaiche (1985).

The more northerly fan systems in the south of quadrant 210 and the "Tampen Fan", are of this type in that the fans are located at the base of the platform slope (210S) and within the Magnus Basin for the "Tampen Fan". The latter again shows a shift towards the north, strongly controlled by the trend of the Magnus Basin.

The position of the basinal fans is offset from the platform fans. The derivation of the Heimdal system from Bruce is recognisable but other systems are separated. The major controlling factors in fan development appear to have been the edge of the East Shetland Platform; the shelf-edge of the Palaeocene Moray Group and the basin axis. These represent changes in the basin slopes within the lower Eocene basin (Fig. 3.31), at the base of which fan systems such as South Alwyn and Frigg accumulated. Changes in the sediment transport direction from the platform to the basin are invoked, and supported by transport directions for West Frigg (McGovney & Radovich (1985).

The system described suggests that a large scale fan developed, fed by channels trending SE. If the connection between the canyons and the Frigg sands, as implied above, is correct, then the sands that make up Frigg must have changed course, from southeast to north. This is supported by the conclusions of McGovney & Radovich (1985) for the deposition of West Frigg, derived from the south. The basin controls for this change in trend are either the Coriolis Effect (Stow, 1986) or a northward dipping basin floor, similar to the change in transport direction from the La Jolla canyon to the San Diego Trough (La Jolla Fan System; Graham & Bachman, 1983). These may have resulted from compactional and fault influences (Badley et al, 1988), or differential subsidence of the north Viking Graben.
<table>
<thead>
<tr>
<th>SHELF BYPASS</th>
<th>PLATFORM FANS</th>
<th>TOPOGRAPHIC BYPASS</th>
<th>BASINAL FANS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fans directed towards south-east. Canyons grade platform.</td>
<td>Fans fed directly from canyons or from platform fans. Characterised by development of mid-fan channels and lower fan lobes. Distribution influenced by Coriolis effect and/or northward dipping basin floor. (eg. West Frigg, South Alwyn)</td>
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<tr>
<td>Fans directed towards south-east. Canyons grade platform.</td>
<td>Sands deposited at termination of canyons. Turbidite flows bypass site of Type 1. Deposited at toe of Palaeocene palaeo-shelf. (eg. Bruce)</td>
<td>Sands deposited at termination of canyons. Turbidite flows bypass site of Type 1. Deposited at toe of Palaeocene palaeo-shelf. (eg. quadrant 35 elongate fan)</td>
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**Figure 3.31 Schematic diagram of the controls on spatial distribution of E1b submarine fans, produced by the basin topography.**
The submarine fan system that flowed into the Magnus Basin is a comparatively smaller system feeding from the region of the East Shetland Platform north of the Unst Basin. Direction of sediment transport, as has been discussed appears to be towards the northeast, although some modification of the direction may result from the topographic low over the Magnus Basin structure. No channelling or canyon formation is observed, and upper fan deposits may be represented by multiple feeders and debris flows, feeding the middle and lower fan.

3.8.1.1 Significance of East Shetland canyons

The reasons for the SE trending canyons in the upper fan are important for the understanding of the controls on the development of this particular fan system. The formation of canyons is common for shelf-slope systems and is initiated by various processes.

i) Shelf loading and slope instabilities resulting from oversteepening at the shelf-break (McGregor et al., 1982; Coleman et al., 1983).

ii) Slope instabilities nucleating as result of earthquake activity and faulting (e.g. Herzer & Lewis, 1979).

iii) Downcutting of the slope by shelf-derived submarine mass-flow (Reimnitz & Gutierrez-Estrada, 1970).

iv) Reactivation of pre-existing drowned fluvial channels (e.g. Shepard, 1981; von der Borch, 1969).

v) Aggradation of an exposed shelf during lowstand (Mitchum et al., 1977).

Similarly the position and dimensions of submarine canyons are controlled by:

i) Position and dimensions of a pre-existing canyon/fluvial channel (Lindseth & Beraldo, 1985).

ii) Energy of density flows downcutting the canyon floor.

iii) Nature of canyon floor or flanks. Canyons will cut further back, and deeper into, soft, muddy sediment, than into consolidated sediments or crystalline basement. Hence the size of canyons is not necessarily a reflection of the size of downslope fan.
iv) Tectonic influences affecting the aggradation of canyon floors. Canyons, as with fluvial systems can concentrate in the hanging wall of faults, parallel to fault trends (Picha, 1979), parallel to the tectonic trends (eg. the Newport Channel; Graham & Bachman, 1983) or aggrade the slope, perpendicular to tectonic trends (eg. the Carlsbad canyon of Graham & Bachman, 1983; Surlyk, 1987).

The canyons of the East Shetland Platform are small in comparison to a number of present day examples (eg. Coleman et al, 1983, Mitchum, 1985). This may reflect the aggradation in comparatively more consolidated and coarser sediment than usually associated with slope deposition. The canyons are cut into the margin of the East Shetland Platform and would thus require a significant amount of back-cutting into this consolidated sediment, if canyon formation induced by slope failure is invoked. There is also no evidence for concentration of canyons in the hanging wall of faults or, indeed, in any pre-existing lineation on the East Shetland Platform. Whilst the seismic sequences below the canyons is obscured by canyon floor multiples, there is no evidence of a structural trend of NW-SE in the Viking Graben. Faults with this trend associated and controlling the Brae Jurassic fan-delta/ submarine fan sediments (Stow et al., 1982; Turner et al., 1987) are associated with offsetting of other N-S trending Jurassic faults, more like transfer faults (Gibbs, 1984), but do not imply the influence of a widespread Permo-Trias (or older) tectonic lineation. There is also no evidence of channelling around the Permo-Trias Fair Isle Basins on the platform. Indeed the absence of channels into the Magnus Basin submarine fan system suggests controls on canyon formation may have been more fundamental.

It is to be concluded that the dominant control on the orientation of the canyons on the East Shetland Platform is tectonic, and that the canyons have developed in a trend parallel to the direction of tectonic tilt of the East Shetland Platform during the lower Eocene. Faulting has played no part in the nucleation and evolution of the
canyons. The canyons are concluded to have aggraded platform rocks as a result of a relative sea level fall for the lower Eocene. It should be noted that this does not preclude either aggradation of the platform during uplift, or a true eustatic sea level fall. Indeed some element of tectonic control can be inferred from the observation that the canyons are aligned perpendicular to the trend of the Shetland Spine Fault system on the western edge of the West Shetland Platform.

3.8.2 Deposition of sequence E1c (Fig. 3.32a)
As described, the E1c sequence was absent of significant fan development. The presence of muddy mounds was noted, which suggested the influence of basinal (i.e. slope and shelf-edge), rather than marginal (i.e. shelf-bypass), influences on sedimentation. The switching off of submarine fans is a phenomenon rarely discussed, and the only process attributed is that of relative sea level rise. Other possible processes, such as upper/middle fan channel switching (cf. Droz & Bellaiche, 1985), are precluded as fan sediments are nowhere observed in sequence E1c. It is possible that these migrated to the Central Graben. However, it is concluded that the mudstones of Elb imply a highstand shelf and slope, with the consequent inundation of the Shetland Platform submerging the source area for the Elb fans. This is supported by the widespread deposition of this sequence over the East Shetland Platform. However, there is no evidence of reworking of the Elb submarine fans into E1c sandy contourites. This may simply occur because of the lack of significant marine water circulation within the basin. A tidal dominated shelf may have restricted sands to the inner shelf, as observed in Chapter 2.

3.8.3 Deposition of sequence E2 (Fig. 3.32b)
The similarities between seismic sequences E1c and E2 suggest that both were deposited in the same environment. The sequence boundary between the two units is essentially a Type 2 unconformity (Vail & Todd, 1981). This suggests a small relative sea level fall, on the indication of the basinwards migration of the coastal onlap. This marginal marine facies has been removed from the E2 sequence by
Figure 3.32. Depositional models for the Eocene seismic sequence.

a) E1c.  
b) E2.
Figure 3.32 (Continued)

(c) E3.

d) E4.
subsequent erosion, and so the sequence may be the result of either sea level rise or fall. On seismic evidence alone, a small relative sea level fall is tentatively concluded. The lack of coarse clastics, and the abundance of thin carbonates in the claystones indicating a low sediment accumulation rate, suggests that E2 was deposited in a highstand, probably following the subsequent sea level rise.

3.8.4 Deposition of sequence E3 (Fig. 3.32c)
The re-emergence of submarine fan sands occurs in sequence E3, but the major system in the lower Eocene is not repeated. The type and distribution of sands is also dissimilar. Fan systems are smaller, predate the downlapping shelf system and no recognisable canyons are observed. However, an erosional event at the base of E3 is responsible for the removal of the E2 highstand deposits from the East Shetland Platform. A rapid fall in sea level could cause shelf planation of the E2 sequence (cf. Mougenot et al., 1983; Mougenot, 1985); development of a shelf by-pass margin and reworking of E2 sediments into the lowermost E3 turbidites. That a major fan system, like that of Elb, does not develop suggests a subsequent rapid rise in sea level. Consequently, the distribution of sands is sporadic, but like Elb, concentrated to the southeast of the East Shetland Platform. The fan deposition was terminated by the progradation of mud-dominated shelf during what is again interpreted as a highstand.

3.8.5 Deposition of E4.
The fan sands of sequence E4 are local, chaotic mounds at the toe of a muddy downlapping shelf sequence. An interpretation based on the principles of Vail et al. (1977a) would suggest that this sequence was similar to the E1 lowstand-highstand cycle, with the sands predating shelf sedimentation - analogous to Elb - and muddy shelf analogous to Elc. However, there is no evidence for shelf bypass in E4 and the internal configuration of the sands is very different to the well-developed submarine fan system of E1. A model is proposed for these sands that takes into account Miall's (1986) sedimentological objections to a pure Vail et al (1977a) type approach.
The presence of sandy slumps at the base of a muddy shelf is a problem which must be accounted for in any model linking the sands and shelf successions in a synchronous model. The absence of evidence for the initiation of a regional submarine fan system similar to that set up in the early Eocene (Elb), or an erosive event (E3), suggests that such a synchronous depositional model is valid. The shelf may have developed in the manner illustrated by Asquith (1970), whereby shelf-deltaic sands are restricted to the coastal margin and a broad muddy shelf develops. It is possible that the slumps are related to the rapid migration of isolated shelf sands (Flemming, 1978) in strong ebb tidal conditions, or isolated deltaic systems (Coleman et al., 1983) across the top of the muddy shelf and the foundering of these over the shelf-slope. Small-scale shelf sand features might therefore accumulate at specific positions, which are related to shelf morphology. Consequently, the base of the slope would be characterised by cumulative slump deposits. Canyons are not a necessary prerequisite for this form of slumping, simply sand wave migration oblique to the trend of the shelf-break. The presence of lignitic debris certainly argues for derivation of the sands from an inner shelf environment. The oblique progradational nature of the shelf suggests a high energy shelf, controlled by deeper water and tidal currents. The shelf sources of these slumps may have been removed by a subsequent Oligocene erosion event.

The development of the E4 sequence suggests synchronous deposition of a muddy shelf with a clastic-dominated shoreline (Fig. 3.32d). The wedges of E4 sands probably represent this sandy inner shelf. Dominant controls on sedimentation appear to be a tidal current oblique to the direction of shelf progradation, which was west to east. Hence tidal currents towards the northeast or southeast are implied by this model. A dominant current direction towards the northeast is preferred, since this allows for the downlap directions for the northern E4 sequence, by a current following the edge of the platform. Note that this represents a shift in the direction of dominant tidal currents from Elc, and the tidal shelf sand waves discussed in Chapter 2. Such a shift in transport direction may
simply reflect the difference in transport direction produced by the wave dominated inner zone of the shelf, and the current-dominated outer zone (Flemming 1980), or a more fundamental change in water circulation in the East Shetland Basin.

3.8.6 A detailed well tie
Validation of the seismic model has been obtained from the details of the sediment record in the East Shetland Basin Eocene succession. In the detailed micropalaeontological analysis of one of the well ties (published by permission of the Robertson Group) a series of transgressive and regressive cycles has been identified (Fig. 3.33). These are based on influxes of terrestrially derived palynoflora, suggesting an increased sediment transport from the source region. It should be noted that regression is not necessarily a direct indicator of lowstand deposition. Relative changes in sea level are the interaction of rate of sediment input, subsidence and uplift and true rise/fall in sea level (section 1.5.1). Hence both highstand and lowstand - where sediment input rate or platform uplift outstrips the contemporaneous rise in sea level - will result in regressive phases as shelf progradation occurs.

Obviously location of the well within the Eocene sequence is important. It is sufficient to note that the well is located in the mid-fan region of the Elb submarine fan system, and is a claystone dominated succession. The hydrodynamic properties of spores and microplankton are similar to that of shales and clays, and consequently are likely to have been winnowed from fan sands etc. The well is therefore recording the background influx of terrestrial flora, rather than a facies control.

It can be seen that the interpreted lowstand for Elb is represented by a series of transgressive cycles. The Moray Group shelf-deltaics, in the Palaeocene, are indeed seen to exhibit a regressive assemblage. It appears that the influx from, for instance, fluvial systems was restricted to the margin and did not extend into the basin. Fluctuations in the cycles might be a response to shorter
Figure 3.33 Transgressive-regressive cycles for a well tie to the seismic sequence. Cycles are based on influxes of terrestrial and shelfal flora. Reproduced by kind permission of Robertson Group.
period sea level changes or a more fundamental sedimentological change in coastal marginal sedimentation.

The progressive inundation of the margin, during Elb-c and the consequent highstand of Elc are represented by a major transgressive phases, followed by more regressive cycles. A gradual change to transgression at the top of Elc may be the result of the reduction in sediment input, as a response to the highstand. This transgression is reflected in the Main Sand Unit in the BGS boreholes. Seismic sequence E2 is represented by a gradual switch from a transgressive-type assemblage to a regressive, supporting a highstand rather than lowstand deposition for this sequence. The restricted distribution of this sequence must therefore relate to a late E2/early E3 erosion event.

The cycle representing the E3 seismic sequence indicates a rapid transgression followed by more gradual transgression. The sequence is very similar to that in sequence Elb, as both erosive episodes are represented by transgressive cycles. This may simply be due to the condensing of the base of the sequences during the shelf-bypass erosive event. The conclusion is somewhat different to that of Vail et al. (1984). However, this would be consistent with the erosion of E2 on the East Shetland Platform, and with the thin fan sands. The overlying sequence E4 is represented by a series of gradual transgressive cycles with an overall regressive trend. This is consistent with the progradation of the muddy shelf sediments - each regressive termination to the cycle possibly representing an influx of marine flora, possibly basin current related.

In conclusion, the transgressive-regressive cycles in the palynology of a East Shetland Basin well are complex. However, the cycles can be interpreted consistent with the seismic stratigraphy outlined above.
3.9 A relative sea level curve for the Eocene, East Shetland Basin.

From the seismic interpretation above, and the supporting evidence it is possible to produce a relative sea level curve for the Eocene in the East Shetland Basin (Fig. 3.34). Essentially, each sequence represents a change in sea level, the direction of which is determined from the internal configuration of the seismic sequence, the lithology, distribution and changes in palynofloral assemblages associated with each sequence. Ages from the detailed well tie (and commercial biostratigraphical zonation) are consistent with the ages derived from the BGS boreholes.

A comparison can be made between this curve and the East Shetland Basin Tertiary sea level curve of Vail et al. (1977a) (subsequently referred to as the Vail curve), and the global sea level curve, produced by Haq et al. (1987) (Haq curve). The Haq curve is based on detailed land data from northwestern European basins. A correlation between these curves is aided by the ages assigned to the Main Tuff Zone of Jacque & Thouvenin (1981) by Knox & Morton (1983) of 54-54.5 Ma. The correlation of the biostratigraphy of the unnamed well tie, to the nannoplankton zones of Martini (1971) used by Haq et al. (1987) allows a reasonable comparative study of the broad changes in sea levels. This highlights a number of differences between the East Shetland Basin curve of this study, and the Haq global sea level curve.

One noticeable difference is in the significance attached to the Lower Eocene. The Haq and Vail curves attach little or no significance to the base of the Eocene, with the development of a minor sea level fall within an overall first order sea level rise. The timing of the Frigg fan deposition is important to the correlation. McGovney & Radovich (1985) infer a late Ypresian age for rapid deposition of the sands, on the basis that this fits the sea level fall at this time. Commercial biostratigraphical evidence, however, suggests that the fans are spread throughout sequence Elb, diminishing towards the top of the sequence. Heritier et al. (1979)
Figure 3.34. A northern North Sea relative onlap curve based on the Eocene seismic sequence.

a) Compared to the Vail et al. (1977) Tertiary sea level curve for the North Sea.

b) Compared to the Haq et al. (1987) global sea level curve.
place the sands in the *Wetzeliaella coleothrypta* zone, with an Ypresian shale cap. Consequently, the top Ypresian sea level fall is not observed in the Frigg fan. The submarine fan model presented above suggests that the Frigg fan, at the distal part of the sequence, is probably the youngest fan in the Viking Graben. Fan sands on the platform may point to an earlier age of deposition.

Similarly, the middle Eocene is a period of first order stillstand on the Haq curve, whereas this study suggests that the sea level in the East Shetland Basin was rising. The Vail curve indicates a relative sea level rise throughout the Middle Eocene, and does not recognise a sea level fall associated with E3 fan deposition. The margin of the basin in the Middle Eocene, as described in Chapter 2, was influenced by the recognisable changes in sea level between 44.0 and 49.5 Ma. The intra-E-4 hiatus, indicated by a glauconite horizon in the BGS boreholes is not seen as a seismic sequence. Sequence E2 is missing off the East Shetland Platform and probably correlates to the hiatus at the top of the Middle Main Sand Unit. This inference is supported by commercial biostratigraphy. The erosion event, which removed the marginal marine facies of sequence E2, is believed to be the result of the sea level fall that correlates with E3. Indeed, the Haq curve shows a better correlation in the late Middle Eocene, reflecting the E3 and E4 sea level falls.

The implication of the comparison (and that in Chapter 2) is that, the East Shetland Basin was influenced by the third order sea level changes recorded by Haq et al. (1987). These are, however, superimposed on top of a number of longer period - first and second order - sea level changes, presumed to be of local tectonic origin. These are a Lower Eocene lowstand and subsequent rapid change to highstand; and a mid-Eocene erosion event removing sequence E2 from the East Shetland Platform. The latter may be a relative effect resulting from hinterland uplift and subsequent planation followed by subsidence. During the highstand, deposition at the margin was controlled by the short period global fluctuations in sea level. Little evidence of erosion of the sequences in the Unst Basin implies
that this was undergoing differential subsidence throughout the Eocene. An event in the middle Middle Eocene resulted in the deposition of the E2 shelf wedge system, probably within a sea level rise. On the basis of the East Shetland seismic sequence, this may have been globally or tectonically induced.
Chapter 4
Eocene seismic stratigraphy west of the Shetland Platform

4.1 Introduction

The previous chapter outlined a six-fold seismic stratigraphy for the East Shetland Platform and East Shetland Basin. The tectonic influences concluded for the relative onlap curve of these units can be tested by analysing the similar succession for the region West of the Shetland Platform, in the Faeroe Basin. A comparison between the relative onlap curve for each basin, and the Haq et al. (1987) global sea level curve, should be able to distinguish between tectonic and sea level events.

Tectonic movements should be distinguished by a number of criteria:-

i) Poor correlation to the global sea level curve would indicate some tectonic event.

ii) Asymmetry of the seismic facies in the basins may represent a tectonic movement of the Shetland Platform.

iii) Preservation of some of the eroded sequences - for instance, E2 and E4 - would indicate basin subsidence events. Rapid changes from shelf to basinal seismic facies would support this interpretation.

Well control is poorer in the Faeroe Basin, since most wells have been drilled on the Rona Ridge high. Wells in the deeper part of the basin have been used to fully document the lithological succession, but the succession in the centre of the Faeroe Basin is much more dependent on the seismic interpretation.

The seismic sequence in the Faeroe Basin has been diagnosed by correlation with the East Shetland Basin. The methods of correlation are discussed in this chapter. Aspects of the seismic stratigraphy and relevant well control are presented, and the locations of sections illustrating this chapter are shown in Figure 4.1.
4.2 Correlation to the East Shetland succession

The sequences on either side of the Shetland Platform have been correlated. Since both sequences are dominantly controlled by the Shetland Platform, as indicated for the Faeroe Basin in this chapter, it is not too simplistic to assume that identical sequences will occur on either side of the platform. Seismic correlation can be made across the northern part of the Shetland Platform into the eastern flank of the northern Faeroe Basin. This suggests that identical sequences are present in both basins.

Unfortunately, seismic correlation into the deeper parts of the northern Faeroe Basin is not possible. The quality of seismic lines shot in the north of the basin, in quadrants 209 and 208 (Fig. 4.1), is degraded by multiples (Fig. 4.2, Fig. 4.3). The presence of a strong Miocene-Recent unconformity combined with the sea bottom reflection have resulted in a series of multiples that obscures the sequence boundaries, and much of the detail within the sequences. Even re-processed lines suffer from the loss of useful information in the succession between the top of the Palaeocene and the unconformity mentioned above. Hence other forms of correlation are required to document and correlate the Eocene succession east and west of the Shetland Platform.

The best correlative tool is the wireline – especially sonic – log profile and characteristics of claystone dominated wells. The seismic characteristics being independent on the sonic properties of the lithologies, the sonic log proves most useful. The sonic log response is therefore linked to the seismic stratigraphic sequences. Gamma ray logs provide lithological information on the presence of sands. Coarsening and fining upwards sequence can occasionally be recognised (Selley, 1979). A summary of the correlation is given in Figure 4.4. Distinctive log character for each of the Eocene seismic units, based on the internal character of each unit on the sonic log, is described below.
Figure 4.1 Location map for the seismic sections shown in this chapter. Numbers on the map refer to licence quadrant numbers.

Figure 4.2 Diagram showing the variations in quality of the Eocene on seismic lines. These variations limit the extent to which a purely seismic stratigraphic approach can be applied to the Faeroe-Shetland Basin.
Figure 4.3
Example of the seismic lines observed in the north-eastern Faeroe Basin. The section highlights the problem with strong seabed and unconformity multiples throughout the succession.
Figure 4.4 Correlation diagram for wells both west and east of the Shetland Platform. Common sonic log characteristics from the East Shetland Basin have been used to define the seismic stratigraphic sequences in the southern Faeroe Basin wells. Datum for the diagram is the top Eocene, as defined by released biostratigraphies.
4.2.1 Sonic log character of sequence Ela

Unit Ela is composed of the tuffaceous claystones of the Balder Formation (Deegan & Scull, 1977). The presence of volcanic grains results in a high sonic velocity and a corresponding low gamma ray signature, producing a "barrel"-like profile, as recognised by Deegan & Scull (1977). The same unit can be recognised in wells in the Faeroe Basin in the late Palaeocene to early Eocene, in the Rockall Trough (Jones & Ramsay, 1982), and on the Hebrides Shelf and Wyville-Thomson Ridge (Stoker et al., 1988). It is again presumed that the development of the thickest tuffs west and east of the platform is contemporaneous (cf. Hitchen & Ritchie, 1987), the base of these thick tuffs representing the base of the Eocene succession.

4.2.2 Sonic log character of sequence Elb

Directly above the barrel-shaped signature of sequence Ela the sonic velocity drops in sequence Elb, gradually increasing up the sequence before terminating in a log kick at the Elb-c boundary. The low velocity may be a relative effect due to the lack of tuffaceous material. Alternatively, it may be associated with a smectite-dominated clay mineralogy produced by the alteration of reworked volcanic products. The sonic profile is characterised by occasional high amplitude peaks, 3-4m thick, which represent siltier or sandier layers associated with the submarine fan sands in this sequence.

Where sands dominate the lithology, the sonic velocity of Elb is higher and much more controlled by gross lithological variations. In this instance the log is characterised by a blocky sonic profile, related to the sand-silt alternations.

4.2.3 Sonic log character of sequence Elc

Sequence Elc is characterised by a velocity increase at the base of the sequence. This is prominent in wells in quadrant 9, but becomes less distinctive further north towards Magnus Basin. Here the low velocity of unit Elb is similarly less pronounced. The sonic profile of Elc is characterised by a flat, invariant velocity, with rare higher velocity peaks and an occasional 'blocky' appearance, which
allows better distinction than the recognition of a distinct log kick. The gamma ray profile is similarly less variable than for sequence Elb, reflecting the absence of sands in the Elc claystones.

4.2.4 Sonic log character of sequence E2
Sequence E2 in the East Shetland Basin, is dominated by low amplitude discontinuous reflections, with occasional high amplitude reflections produced by limestone bands observed in well ties. This is observed in the sonic log character which has a broad 'blocky' appearance, interspersed with abundant high velocity peaks up to 8m thick.

4.2.5 Sonic log character of sequence E3
The log profile of E3 is dependent on the position of the well with respect to the shelf and fan systems outlined in Chapter 3. The sonic log profile close to the channels and fans in E3 (eg. 9/08-4) is devoid of high amplitude peaks, observed in sequence E2. The profile is much more 'blocky' and sonic velocity much more variable overall. Slope and basinal regions outside the fan system (eg. 9/19-3) produce a profile more like that of E2, but with a recognisable log kick between the two units.

Further north, the E3 sequence is again very similar to the E2 sequence (eg. 210/13-1), occasionally showing a blocky appearance.

4.2.6 Sonic log character of sequence E4
This unit again exhibits some variability in its log response, but is generally separated from E3 by a sharp velocity increase into E4 and higher velocities for the E4 sequence. Sand-dominated successions show an invariant velocity in the log response, with occasional high and low velocity kicks. The shelf claystones are characterised by a blocky appearance to the sonic log.

4.3 Comparison between log characteristics - west and east
The differentiation based on sonic log characteristics for the East Shetland Basin can be compared to a differentiation in the Faeroe
Basin. Both wells described below are located between the northwestern flank of Rona Ridge and the axis of the Faeroe Basin, and can therefore be expected to have the most complete Eocene succession.

4.3.1 206/2-1a

Location: 60°59.731'N, 2°45.552'W; KB elevation: 42'

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Depth (ft. KB)</th>
<th>Depth (m. MSL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E4</td>
<td>4097-4815'</td>
<td>1236-1455m</td>
</tr>
<tr>
<td>E3</td>
<td>4815-5680'</td>
<td>1455-1718m</td>
</tr>
<tr>
<td>E2</td>
<td>5680-5940'</td>
<td>1718-1798m</td>
</tr>
<tr>
<td>E1c</td>
<td>5940-6400'</td>
<td>1798-1938m</td>
</tr>
<tr>
<td>E1b</td>
<td>6400-6993'</td>
<td>1938-2119m</td>
</tr>
<tr>
<td>E1a</td>
<td>6993-7160'</td>
<td>2119-2170m</td>
</tr>
</tbody>
</table>

The tuffaceous claystones in 206/2-1a, between 2119m and 2170m show the characteristic barrel-shape of the E1a unit in the East Shetland Basin. The characteristic low velocity sequence E1b is observed above 2119m. Velocity gradually increases between 2010m and 1983m, but the top of the sequence appears to be at 1916m, where a log break indicates an upward increase in velocity. The log break is taken as the base of the E1c sequence. Both E1b and E1c are characterised by a peaky configuration, related to limestone bands and calcareous cementation, becoming slightly 'blocky' for E1c, implying discrete siltier beds. Overall, the log response and lithology indicates low clastic input. An increase in the peaky appearance is taken as the E2 equivalent sequence between 1717m and 1800m. This is itself terminated by a log drop, similar to that observed at the E2-E3 sequence boundary in well 9/19-3 in the East Shetland Basin. The log configuration of E3 is again 'peaky', but changes at 1455m, where the sonic log becomes less variable with fewer but higher velocity peaks. This may represent the change to E4 deposition.

Lithologies in the well are claystone-dominated and the wireline logs show a number of distinctions, which allow the sub-divisions of
the Eocene in the East Shetland Basin to be correlated with the Eocene in the middle of the Faeroe Basin. Despite the very different depositional conditions (well 206/2-1 is claystone dominated throughout the whole of the Eocene, rather than occasionally sandy as in the East Shetland Basin) the distinctions are clear.

4.3.2 204/28-1

Location: 60°9.817'N, 4°32.017'W; KB elevation: 25m

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Depth (m. KB)</th>
<th>Depth (m. MSL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E4</td>
<td>740-825m</td>
<td>715-800m</td>
</tr>
<tr>
<td>E3</td>
<td>825-905m</td>
<td>800-880m</td>
</tr>
<tr>
<td>E2</td>
<td>905-1100m</td>
<td>880-1075m</td>
</tr>
<tr>
<td>E1c</td>
<td>1100-1162m</td>
<td>1075-1127m</td>
</tr>
<tr>
<td>E1b</td>
<td>1162-1250m</td>
<td>1127-1225m</td>
</tr>
<tr>
<td>E1a</td>
<td>1250-1323m</td>
<td>1225-1298m</td>
</tr>
</tbody>
</table>

Sequence E1a of the Eocene succession in this well is taken as the tuffaceous claystone unit. This shows the barrel shaped gamma/sonic profile characteristic of the East Shetland Basin wells. Gamma ray profile increases and sonic velocity decreases uphole into the top of the 'barrel', indicating a gradual reduction in the proportion of tuffaceous material, rather than the base of unit E1b. Overlying the Balder Formation is a mudstone-sandstone sequence (1198-1225m msl) that grades into a sandstone dominated sequence between 1140 and 1198m, topped by a sharp log drop, defining the top of sequence E1b. The overlying unit is a claystone with thin sands at the top. The sonic log shows an occasionally peaky appearance in the claystones, becoming less variable in the sands. This can be correlated with the E1c sequence of the East Shetland Basin.

Above the sand (880-1075m msl) the sonic log response displays an abundantly peaky signature corresponding to limestone interbeds, correlated to that of E2 in the East Shetland Basin. There is no log kick, similar to that associated with the base of E3, in the claystones, suggesting that unlike well 9/19-3 in the East Shetland Basin, the 'peaky' claystones are probably all within this E2.
sequence. The top of the E2 sequence is taken as the base of a thick glauconitic sandstone. Above this boundary the sonic log signature is invariant, up to 785m (msl), where a velocity decrease uphole defines the base of the E4 sequence, and is within the Eocene succession as defined by the released biostratigraphical ages. This latter unit is tentatively defined, since its occurrence in the East Shetland Basin is limited. Biostratigraphy of the well indicates that this sequence is within the Mid-Late Eocene.

It is argued that the characteristics of the sonic log for 204/28-1, in the southern part of the Faeroe Basin, are very similar to those of the Eocene succession outlined for the East Shetland Basin. The lithostratigraphy is directly comparable. While there is limited biostratigraphical evidence to support this correlation, the similarity is a good working model for sub-dividing the Eocene succession in the Faeroe Basin.

Having concluded that the sequences in the East Shetland Basin and the Faeroe Basin are similar, it is possible to describe the Faeroe Basin stratigraphy in terms of units equivalent to those observed in the East Shetland Basin. Hence the nomenclature is kept the same, for this basin, as for the East Shetland Basin.

4.4 Seismic stratigraphy in the southern Faeroe Basin

The Eocene succession in the Faeroe Basin is best imaged on the seismics in quadrants 204 and 205 (60°-61°N, 3°-5°W), on the southwestern margin of the basin. Because of the particular nature of the sequence and the water depth (exceeding in parts 1000m) the BGS is very useful in the southern Faeroe Basin. Multichannel seismic data is less affected by multiples in this region and is tied to two wells. Well control is based on well ties 204/28-1, located on the southeastern flank of the basin and 206/2-1, situated in the central Faeroes Basin on the western flank of Rona Ridge.
4.4.1 Seismic sequence Ela

The tuffaceous clays of sequence Ela are present in well 204/28-1. The tuffs are observed on seismic as a high amplitude continuous reflection, traceable over the Rona Ridge into the West Shetland Basin (Fig. 4.5). The reflector is traceable across the whole of the Faeroe Basin, but merges with the high amplitude reflections of the Faeroes Plateau lava series along the western flank of the Faeroe Basin. Below the lava reflector the details of the pre-Tertiary succession is obscured. The earliest Faeroes basalts are the Lower Series extruded in the Upper Palaeocene (Landenian - Lund, 1983) prior to the deposition of the Ela tuffs (Smythe et al., 1983), and it is this surface with which the Ela reflectors merge.

Within the basin the top of sequence Ela shows evidence of folding along a NE-SW and E-W axes, parallel to the axis of the Faeroe Basin. The folds observed have an amplitude of approximately 200-300m, with a wavelength of 37 Km. A monoclinal fold structure is observed in quadrant 204SW, with the same trend, but exhibiting an en echelon fold structure. Differential compaction, along the edge of fault blocks and the plateau basalts, is not a likely cause of these structures which are truncated at seabed and not buried to any great depth.

4.4.2 Seismic sequence Eib

Over most of the Faeroe Basin in quadrants 204 and 205, sequences Eib and Elc consist of horizontal, moderate to low amplitude reflections. Along the western flank of Rona Ridge, however, sequence Eib exhibits a series of downlapping reflections of moderate amplitude, thinning towards the north-east (Fig. 4.6a), and delineating a prograding wedge of shelf sediments along the axis of Rona Ridge, derived from the Shetland Platform. Some onlap of the top of Ela is observed at the toe of this downlap wedge, implying marine, rather than coastal onlap (cf. Facies 2; Section 3.3.1). West of well 204/28-1, the Eib sequence rises to seabed, and shows a similar, smaller scale downlap unit approximately 100m thick, in the upper part of the sequence (Fig. 4.7). The isopach map for Eib (Fig.
Figure 4.5 Isochron map of the top of sequence Ela in the southern Faeroe Basin. Contours are in milliseconds, at 100 ms intervals. "V" shading shows the position of the Faeroe-Shetland Flood Basalts.
Figure 4.6 a) Seismic facies map for the southern Faeroe Basin. Horizontal lines - parallel, horizontal reflectors; diagonal dashes - downlap, arrows indicating dip direction; horizontal dashes - onlap; wavy lines - mounded.

b) Isopach map for the Elb sequence, based on a sonic velocity of 2400 m/s. Contours in 50 m intervals.
Figure 4.7 a) Seismic section showing the difference in direction of downlap between the E1b deltaic wedge and the E4 shelf wedge. The difference is produced by the removal of the Faeroe-Shetland Flood Basalts as a source for sediment in the later Eocene succession.
Figure 4.7 b) Sequence Elb as imaged on the BGS single channel airgun records.
4.6b) shows this to be localised in a lobate form thinning into the basin. Direction of downlap and the lobe shape indicate derivation from the southwest. Smaller lobes extend from this main lobe, directed more towards the north-east. BGS airgun data shows the sequence to be composed of a series of horizontal reflectors, with a 100m thick cross-bedded unit within the centre of the sequence.

This lobate shape is interpreted as a deltaic system prograding into the basin from the Faeroes Plateau. The presence of a number of thick and thin lobes, and the development of downlap in intervals throughout the sequence, suggests distributary switching of the delta system. The delta lobes become more restricted through time, prograding less into the basin. This suggests gradual depletion of the Faeroes Plateau source area. Derivation from the Faeroes Plateau implies the emergence of this region as an important sediment source during the early part of Elb. Deposition of the delta system is more likely to be have developed during, or after, extrusion of the Middle and Upper Faeroes Basalt series, which are much less widespread than the Lower Series (Smythe et al., 1983). At the same time, a less variable shelf-deltaic wedge was prograding from the Shetland Platform.

In well 204/28-1, the Elb sequence is represented by well sorted grey, medium grained quartz sands, interbedded with light brown mudstones. Occasional gravels are recorded. The well is situated at the south-western end of the Shetland Platform shelf wedge, and within the shelf wedge, and the sands probably represent shelf sands rather than slumps or fan sands. A mounded series of reflectors north-west of the well, similar to facies 3b (Fig. 3.5), probably represents the slump equivalents of the 204/28-1 shelf sands. The mound is covered by a drape that onlaps the E1a reflection, and suggests a short-lived, localised submarine fan system.

4.4.3 Seismic sequence E1c

Sequence E1c shows less variation in reflection type than sequence Elb (Fig. 4.8a), and is developed as a series of low to moderate
Figure 4.8 a) Seismic facies map for the Elc sequence in the southern Faeroe Basin. Shading as for Figure 4.6

b) Isopach map for sequence Elc, based on a velocity of 2300m/s. Contours at 50m intervals. Diagonal shading indicates where the sequence is absent.
amplitude sub-horizontal reflectors over most of the southern Faeroe Basin. The base of Elc is poorly defined, but is taken as the top of the uppermost downlapping sequence observed in sequence Elb. A small, restricted series of downlapping reflectors, similar to the shelf wedge of Elb is observed in block 204/28, but does not extend very far into the Faeroe Basin. Occasional low amplitude downlapping reflectors are observed in Elc, but these are localised and not as well-defined as the shelf wedges in Elb. Similarly, the isopach map (Fig. 4.8b) also indicates the absence of major shelf-type sedimentation in the southern Faeroe Basin.

Over the en echelon fold structure observed at the top of Ela, sequence Elc is missing, and sequence Elb subcrops below thin Neogene sediments, at seabed. This erosion does not influence the trend of isopachytes, and suggests that erosion postdates the deposition of sequence Elc. Around this structure Elc reflectors are sub-horizontal and low to moderate amplitude. Within the basin, localised patches of thicker Elc are associated with downlapping, low amplitude reflectors.

In well 204/28-1, Elc is observed as a thin grey-green mudstone overlain by a carbonaceous light brown mudstone. A 15m thick sand occurs at the top of the sequence. Analysis of the gamma ray log indicates a gradual coarsening upwards of the sand from the mudstones. The well is located within the small downlapping shelf wedge, and the sands are probably shelf sand bodies, or small fans, similar to those observed in Elb.

Apart from this, sedimentation appears to have been in a low energy environment, with little clastic input into the basin during this time. The absence of sediments sourced from the Faeroes Plateau suggests the inundation of this area as a result of either sea level change or an increase in the rate of subsidence in the basin, from Elb to Elc time.
4.4.4 Seismic sequence E2

The base of sequence E2 is defined by a moderate amplitude reflection that dips into the basin along the western margin of Rona Ridge, and reflections are concordant with those of Elc in the deeper parts of the Faeroe Basin. At the flanks of the folds observed in the top of Ela to Elc, the E2 reflectors onlap (Fig. 4.9a).

Along the western margin of Rona Ridge a wedge of moderate amplitude downlapping reflectors is observed. This thins into the basin, and downlaps onto a lower E2 series of horizontal, moderate to low amplitude reflectors. The age of this sequence is unknown and is discussed below. A distinct sub-horizontal reflector between the two sequences is developed at the distal end of the downlapping package, where a mounded facies is also observed. On BGS airgun data the E2 shelf wedge is clearly displayed with a low amplitude downlapping internal configuration (Fig. 4.10). This wedge thins at its toe, becoming concordant with the lower series of E2 reflectors, with occasional areas of downlap, and it is this lower series that onlaps the top of Elc in the en echelon fold structure.

The isopach map of E2 (Fig. 4.9b), shows the development of the shelf wedge derived from the Shetland Platform. The thickness of the sequence increases into the basin as a result of the progressive marine onlap of the Elc upper surface. The isopachytes also show the influence of an ENE-WSW trending ridge associated with the en echelon fold in quadrant 204S.

In well 204/28-1, the E2 succession is dominantly a light to medium grey mudstone, with traces of lignite, becoming sandier and more glauconitic towards the top of the sequence. The well penetrates the downlap wedge, and the coarsening upwards nature of E2 in the well supports the interpretation of shelf progradation. The shelf is represented by a a variable, sonic log character between 875-975m (msl) at the top of the interpreted E2 sequence in well tie 204/28-1 (Fig. 4.4). The shelf wedge is believed to represent the top of E2.
Figure 4.9 a) Seismic facies map for the E2 sequence in the southern Faeroe Basin. Shading as for Figure 4.6.
Figure 4.10 BGS single channel airgun record showing the E2 shelf wedge. Sequence E3 is represented by the higher amplitude wedge of discontinuous reflections onlapping the E2 shelf wedge.
The seismic tie to 206/2-1 is better fitted to this interpretation than to the shelf being equivalent to E3.

The development of two discrete sequences within E2, and the controls on thickness produced by the emergence of the anticlinal structures, suggests that the compression associated with these folds initiated at the beginning of E2, with possibly one more important compressional phase within E2. This subsequent phase is followed by the rejuvenation of a Shetland Platform derived shelf wedge system.

4.4.5 Seismic sequence E3
The E3 sequence is developed as an infill to the E2 sequence, reaching a maximum of 300m thickness in the north of quadrant 204. The base of E3 is recognised as a change from the prograding downlap of the shelf wedge, to a sequence of low amplitude discontinuous reflectors. The E3 sequence is thin over the areas of folding, and infills the base of the E2 slope, as observed on BGS seismic lines.

The shelf wedge in E2 is replaced in the north-east of 205 by a small, 100m thick lobe of downlapping reflectors (Fig. 4.11a). This lobe is thinner than the shelf wedges observed in the earlier sequence, which were up to 200m thick, and the downlap is less well-defined. The interpretation of this region as an off-shelf area is consistent with the observed onlap of E3 onto the E2 slope, and the absence of any erosion of the E2 shelf wedge to the south. Lack of an erosional event suggest in-place drowning of the E2 sequence similar to that involved in the preservation of the Moray Group palaeo-shelf edge in the East Shetland Basin.

In the south of quadrant 204, the E3 sequence becomes confined between the en echelon fold uplift, and the slope of the E2 shelf. The isopach map (Fig. 4.11b), based on BGS airgun data, indicates the topographic confinement of a submarine lobe between these two features. Internally this lobe shows evidence of onlap onto the fold region (Fig. 4.11a) and the E2 slope, with reflectors being
Figure 4.11 a) Seismic facies map for E3 sequence. Shading as for Figure 4.6.

b) Isopach map for the E3 sequence, based on a sonic velocity of 2000 ms\(^{-1}\). Contours are at 50m intervals.
discontinuous and slightly concave upwards. The reflectors also exhibit a discontinuous downlap towards the south-west, parallel to the axis of the E3 lobe. This feature therefore appears to have been derived from the north-east and fed laterally into the topographic low developed between the E2 palaeoshelf sequence and the emerging region of broad folding. Compressions associated with the development of these folds appear to have been active immediately before, or during, the deposition of the E3 sequence.

The lithology of E3, observed in well 204/28-1, is a sand-dominated sequence of coarse grained sands with interbedded brown siltstones, at the edge of the fan lobe. This suggests that the slightly concave upwards reflectors are a series of broad, probably ephemeral, mid-fan type sand-filled channels within this lobe structure. Derivation appears to be from shelf sediments, probably on Rona Ridge and in the West Shetland Basin.

4.4.6 Seismic sequence E4

On the basis of the well log correlation E4 has a broader distribution than the upper Eocene sequence observed in parts of the North Sea. However, the sequence does show the development of an elongate, NE-SW trending downlap sequence to the north-west of Rona Ridge (Fig. 4.12a). The base of E4 is defined by this downlap surface. Downlap is towards the NE. The interpreted shelf wedge is derived from the central eastern part of quadrant 205, and progrades away from Rona Ridge. To the northwest of this downlap facies, E4 is represented by low amplitude discontinuous reflectors.

The shelf facies described above is limited in distribution by the emergence of a NW-SE trending topographic high associated with the en echelon folding. The topographic high separates E4 into a shelf-dominated northern region and the topographic low developed between the ridge, trending ENE-WSW and Rona Ridge. Within this low, sequence E4 has identical internal configuration to sequence E3. The downlap in the north-western end of the lobe is part of the shelf.
Figure 4.12a) Seismic facies map for the E4 sequence in the southern Faeroe Basin. Shading as for Figure 4.6.

b) Isopach map for the E4 sequence in the southern Faeroe Basin, based on a sonic velocity of 1950ms⁻¹. Contours at 50m intervals.
system, which changes to a gentle marine onlap in the south-eastern portion.

In well 204/28-1, the E4 sequence is observed as a series of medium to coarse grained, shelly sands interspersed with siltstones. The sands are located at the south-west extremity of the E4 shelf wedge, suggesting that this is a sand-dominated slump system. The seismic ties confirm this interpretation since the sequence does not contain the seismic facies variations that characterised the major submarine fans observed in earlier sequences in the East Shetland Basin and the southern Faeroe Basin. The presence of glauconite in the sands implies marine deposition and shelly material is probably derived from the shelf, presumably the downlap wedge observed to the north-west.

4.5 Depositional history of the Eocene succession

The succession in the Faeroe Basin can be sub-divided into a similar number of units as in the East Shetland Basin. Despite the lack of supporting evidence showing the synchronicity of the units, the correlation of the units in the west to the units in the east has been discussed. It is possible, with this interpretation, to compare the sedimentation and distribution either side of the Shetland Platform. An illustration of the depositional history for the southern Faeroe Basin is given, with the broader evolutionary history of the whole of the basin, in Figure 4.22.

E1a. The occurrence of the tuffaceous bands within claystones suggest that deposition in the Faeroe Basin was very similar to that in the East Shetland Basin. This was shown to be transgressive, which is the same as the conclusion of Mudge & Rashid (1987) for the Faeroe Basin. Consequently, a relative sea level rise resulted in the inundation of the Faeroe Basin, after deposition of Palaeocene deltaic and shelf sands on the basin margin (Mudge & Rashid, op. cit.).
Eib. The most noticeable difference between the sequence west and east of the Shetland Platform is the absence of major submarine fan systems in the Faeroe Basin. The Faeroe Basin is characterised by deltaic sedimentation. This sequence shows derivation from the Faeroes Plateau, with a shelf system established along the margin of Rona Ridge. The Faeroes deltaic system appears as an elongate form, which indicates a high constructive, probably fluviolally dominated, system. The shelf system emerging from the Shetland Platform is much more widespread. This indicates a greater wave or tidal influence on the margin. Input from the Shetland Platform was from a broad multiple source, possibly several small fluvial systems, rather than a large well-established system, as probably existed on the Faeroes Plateau. That such a system did not develop on the Shetland Platform is a reflection of the limited extent of the source region that this provided. The East Shetland and Faeroe Basins show a common regressive nature to sedimentation. However, the southern Faeroe Basin was much shallower during deposition of Sequence E1b.

The significance of this shallowing is the implication that the type of deposition and distribution was more controlled by regional tectonic conditions. The absence of a recognisable submarine fan system in the southern Faeroe Basin is taken to imply that the basin floor was both flat and shallow. If the rate of sediment input into both the basins was the same, then the difference could be explained by regional uplift of the Faeroe Basin, compared to the East Shetland Basin. Consequent tilting of hinterland and platform areas would have resulted from this tectonism.

Elc. The removal of the Faeroes Plateau as a major source of sediment for the sequences after Elb deposition is consistent with the increasing influence of both differential subsidence and relative sea level rises on basin deposition. Sequence Elc is characterised by low sediment input and the development of a small restricted shelf, in the Faeroe Basin. Whereas the north North Sea Elc sequence is characterised by progradation of a highstand shelf, the same sequence in the Faeroe Basin does not exhibit such
highstand sedimentation. Shelf deposits may be confined to the top of Rona Ridge and the West Shetland Basin.

During Elc deposition the lack of a prominent highstand shelf sequence implies that water depths over the whole of the Faeroe Basin were either too deep to allow a shelf to establish, or that sediment rate was very low. In this case, water depths in the basin were deeper than those in the East Shetland Basin, a change in conditions from those implied for deposition of the Eb sequence. Hence, differential subsidence within the Faeroe Basin - during Elc deposition - overtook subsidence in the East Shetland Basin, although both basins show the effect of relative sea level rise.

E2. The progressive infilling of the Faeroe Basin by marine onlap suggests that for the early part of the sequence sedimentation was low energy and deep marine. Unlike the East Shetland Basin where a highstand muddy shelf was established, the sequence on the west again does not show shelf progradation in this early part of E2. Water depths would appear to have been too deep to establish such a system, resulting either from renewed subsidence in early E2, low sedimentation rate or the residual effects of rapid subsidence during Elc deposition.

The establishment of a shelf system in the south of the Faeroe Basin, that downlaps onto the marine onlapping sequence suggests a major change in deposition on this side of the Shetland Platform. If the shelf wedge represents upper E2 sediments, then there is no similar sequence observed in the East Shetland Basin. This does not rule out an upper E2 age for the sequence, as it has been postulated that an erosive event removed all E2 sediments from the East Shetland Platform. The shelf wedge may then represent a highstand shelf at the top of Sequence E2 which is not observed in the East Shetland Basin. Alternatively, it may be an event local to the Faeroe Basin. That shelf development is observed at the same time as fold development suggests a link, and it may be that compression
results in the formation of a shallow sea in the southern Faeroe Basin, allowing shelf deposition to resume.

E3. This sequence consists of a smaller shelf sequence with the development of a fan lobe channelled between the topographic expression of the E2 shelf and the fold ridge. The absence of major channel systems is very different to the situation described in the East Shetland Basin. A fan system similar to that described by Heller & Dickinson (1985), is proposed for this sequence. Sediment slumping off the slope has been funnelled into the topographic low and a series of ephemeral sandy channels established, as indicated by the hummocky internal configuration of the middle of the submarine fan lobe. A mounded facies to the south-east represents the lower fan or distal ramp facies and represent sandy sheet flows. The deltaic environment required for the sand-rich input to the model of Heller & Dickinson (op. cit.) is not seen in the basin, although part of the downlapping wedge is observed in 204/28-1. Progressive funneling of the slumps would build up a thick, localised sequence.

Other small scale lobes are observed north of this major feature, and may represent unfunnelled similar structures. These are probably fed by multiple feeders on the slope.

The development of sand-rich turbidites is consistent with the reactivation of fan sedimentation in the East Shetland Basin during E3. The absence of regressive shelf sediments over the fan sands may again be a reflection of the deeper water depths, or a lower sediment input, in the Faeroe Basin than were prevalent in the East Shetland Basin at the time.

E4. The appearance of a large widespread shelf wedge system is similar to the deposition of E4 in the East Shetland Basin. The age of this unit is Upper Eocene in 204/28-1, so it seems unlikely that this downlap is correlatable to the Oligocene, rather than E4. Overall, the sequence is much sandier on the west than the East, and
suggests preferential drainage towards the west, or the proximity of the source area to this particular margin.

4.6 Tectonism and seismic stratigraphy in the north-eastern Faeroe Basin

As mentioned earlier, the details of the seismic stratigraphy in the northern part of the Faeroe Basin are obscured by recurrent multiples. However, the well correlation outlined above, and seismic profiles on the margin of the basin can be used to give a broad geological history.

The seismic sequence Ela is recognisable in the area. In well 209/6-1, the main tuff zone is observed below a 4m thick coal. Tuffaceous horizons are also interbedded with the late stage basaltic lavas from the Erlend Igneous Complex. These basalts are associated with sills K-Ar minimum age dates of uppermost Palaeocene to lowermost Eocene (Harland et al., 1982; Hitchen & Ritchie, 1987). Elsewhere tuffs are predominantly observed above sands and coals. This phase of terrestrial shelf-deltaic deposition in the basin is taken to be late Palaeocene (Mudge & Rashid, 1987), and is contemporaneous with the deposition of the early phase of pyroclastics (Knox & Norton, 1983). Late stage tuffs, deposited in grey-brown siltstones in 209/3-1, 209/9-1 and 208/15-1 and olive-grey siltstones in 208/27-1, are assigned to the Ela sequence (Fig. 4.13). As observed for the East Shetland Basin succession, the Ela sequence corresponds to transgression and inundation of the late Palaeocene shelf-deltaics.

The rest of the seismic sequences in E1 are only detailed close to the edge of the North Shetland Platform - the name assigned to the northernmost part of the Shetland Platform (Fig. 4.14). Along this margin, undulose and chaotic reflectors are observed dipping into the basin (Fig. 4.15). The internal configuration of this wedge is similar to submarine slides described by Prior et al. (1984) and Normark & Gutmacher (1988) (Fig. 4.15b). Sequence Elc, as for the northern East Shetland Basin, is probably very thin in the centre of
Figure 4.13 Lithostratigraphic correlation for wells over and around the Erlend Igneous Complex, in the northern Faeroe Basin. Wells amount to a ENE-WSW section across the complex. Datum in this case is sea level.
<table>
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<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
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<th>12</th>
<th>13</th>
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<tr>
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<td>0.5</td>
<td></td>
<td>1.0</td>
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<td>1.5</td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

**Figure 4.14** Seismic line and interpretation along the western flank of the North Shetland Platform. Although some parts of the section are obscured by multiples, parts of each sequence show the succession to be identical to that on the eastern flank of the North Shetland Platform.
Figure 4.15 a) Seismic section along the western flank of the North Shetland Platform.
Northern Faeroe Basin

<table>
<thead>
<tr>
<th>Onlap/downlap</th>
<th>Discontinuous, low amplitude</th>
<th>Downlap</th>
<th>Undulose, downlapping</th>
<th>Gentle downlap, occ. chaotic</th>
</tr>
</thead>
<tbody>
<tr>
<td>occ. high amplitude</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Truncation of Ash Marker

8.25 km

Scarps & gullies
Debris flow
No continuous reflectors
Pressure ridges
Spillover lobes
Hummocky & Smooth
Outrunner blocks

Kitimat Slide analogue
(Prior et al., 1984)

Erosion of seafloor

Figure 4.15 b) An analysis of the internal configuration of the mounds indicates a similarity to the submarine slide at Kitimat (Prior et al., 1984).
the basin, forming a restricted shelf wedge system close to the edge of the North Shetland Platform.

There appears to be little influence exerted by the Erlend Igneous Complex on the Elb/Elc sequences. Tentative correlations suggest that the Elb sequence in 209/9-1 is a sandstone (Fig. 4.16). An open marine, middle energy environment is indicated in the released well report.

A wedge of downlapping reflectors on the complex has been assigned an Eocene age (Gatiffany et al., 1984: their figure 4, Line F). If the deltaics are taken as Eocene then it implies the emergence of the western flank of the igneous complex, whilst the eastern flank was submerged. The absence of eastward directed wedges is taken as evidence that this was not the case. The age is unproved and it is possible that this series of reflections represents a delta lobe - synchronous with basalt extrusion - in the late Palaeocene. Indeed, the limit of the wedge is consistent with the recognisable shelf edge limit of the late Palaeocene observed in the region (cf. Rochow, 1981) (Fig. 4.16). Therefore, the Erlend Igneous Complex did not act as a source for Eocene sediments, unlike the basalts in the southern Faeroe Basin. Rapid planation of the complex must therefore have taken place during the late Palaeocene, followed by subsidence during the early Eocene.

The sequences E2 and E3 are poorly recognised in quadrants 208 and 209. The sequence in 209/9-1 between the ?Elb/Elc sandstones and the polished quartz sands of E4, probably contains both of these sequences. This interval consists of argillaceous siltstone, implying low energy deposition. To the west - in wells 209/3-1 and 209/6-1 - sandy clays and limestones are observed throughout the sequences.

Wells in the central Faeroe Basin (eg. 208/15-1) show the same lithological sub-divisions as the East Shetland Basin. Sandy siltstones over the Eocene tuffaceous siltstones are consistent with
Figure 4.16 Map showing the nature of the northern Faeroe Basin at the end of the Palaeocene. The map is based on seismic interpretation and well correlation of the late Palaeocene. Shape of the volcanic escarpments of the Erlend Igneous Complex and the Faeroe-Shetland Flood Basalt are taken from Hitchen & Ritchie (1987). A deltaic wedge on the Erlend basalts from Gatliiff et al. (1984) is interpreted as being of this age.
the E1b/E1c sequences. Thin limestones in the overlying grey brown clays are consistent with those observed in the E2 sequences throughout the East Shetland Basin. Overlying thin, glauconitic sands may represent the E3 fans.

The distribution of sequence E4, in the northern Faeroe Basin, is described in detail in Chapter 3. The distribution of the E4 sediments appears to be controlled by the northwestern flank of the North Shetland Platform.

Therefore, it can be concluded that the northern part of the Faeroe Basin was influenced by the same controls on sedimentation, as the East Shetland Basin. No evidence of folding is observed in the northern Faeroe Basin.

4.7 Role of fault activity during the Eocene

A number of folds have been described, that disturb and influence the Eocene seismic sequences in the southern Faeroe Basin. The timing and duration of the deformation is discussed, with particular reference to similar features on both Wyville-Thomson and northern Rockall Trough.

The shape of the en echelon folding in the top of E1a, is consistent with a NW-SE dextral strike-slip motion. This region is in the vicinity of the position of the NW-SE trending Judd Fault (Kirton & Hitchen, 1987) (Fig. 4.17). The geometry of the Judd Fault is one of low-angle fault with an apparent dip of 45° to the north-east. The fault detaches at about 6 Km with no evidence of roll-over in the Cretaceous hanging wall sediments. The structure is interpreted as a transfer fault along which sinistral strike-slip occurred during Mesozoic extension of the Faeroe Basin. Recognisable movement of this fault occurred through the Upper Jurassic and Cretaceous and ceased during the Palaeocene. However, it would appear that fault movements occurred in the post-Palaeocene crumpling the soft sediment cover over the fault (Fig. 4.18). Whereas, pre-Eocene
Figure 4.17 Structural analysis of the folds observed in the top of Ela. Two fold sets are defined. One, trending approximately E-W, are associated with NW-SE dextral movements in the vicinity of the deeper Judd Fault. The other set, trending NE-SW is compatible with a compression perpendicular to the axis of the Faeroe Basin.
Figure 4.18
Isometric diagram showing the response of soft sediment to subtle movement on underlying pre-existing fault structure. The model is related to the deformation in the Ela sequence.
movements on the fault, associated with extension were sinistral, the movement is reversed in the Eocene, and implies minor inversion tectonics. That the fault does not break the sediments suggests that fault movements were slow and gradual and with small displacements involved, creating a palaeotopographic feature only.

Differential subsidence of the basin has caused eastwards tilting of the Eocene sediments in post-Eocene time. The tilting has resulted in the present monoclinal appearance of a number of the folds.

Other folds to the north-west of this structure are more consistent with NW-SE oriented compression, normal to the trend of faults associated with Rona Ridge. This direction is parallel to the orientation of the Judd transfer fault and suggests pure compression of the Faeroe Basin in the post-Palaeocene. The folds are similar to those in the vicinity of the Judd Fault, in that no movement can be observed on the underlying structure. The position of the folds may be controlled by reverse rotational movement of tilted fault blocks, in the manner of Jackson (1980), although these are obscured on the seismics.

From the isopach maps the major Eocene phases of fault movement and consequent fold development, occurred in the Middle Eocene (Fig. 4.19). The influence of the seismic sequence is initiated in E2, and influences the sedimentary style into E4 (Upper Eocene). Later, Oligocene movements, as evoked by Earle et al. (in press) may have occurred, but the Oligocene succession is removed from the top of the en echelon fold structures, and is thin over much of the rest of the region. It appears to be a post-folding sequence on Fig. 4.19, and the rapid differential subsidence of the basin to the present depth may have occurred at this time.

More regionally, there is evidence of fault movements consistent with those described here. Mapping of the sediments (Stoker & Hitchen, 1988) on Wyville-Thomson Ridge has indicated the presence of faults in the late Palaeocene-early Eocene tuffaceous sandstones.
Figure 4.19 Seismic section showing the onlap of the Eocene units onto the NE-SW trending folds. The section emphasises that formation of the folds is contemporaneous with Eocene deposition, and constrains the timing of major movement of the faults to the upper Eocene.
overlying the basalts (Stoker et al., 1988). A series of NW-SE trending faults, up to 29 km in length, downthrow to the south-west (Fig. 4.20a). Between faults are en echelon anticlines and synclines, trending NW-SE to NE-SW, concentrated around the faults, but distributed over the whole of the south-east part of the ridge.

These structures represent the effect of NW-SE dextral shearing of the Wyville-Thomson Ridge (Fig. 4.20b). The concentration of faults in the south-eastern flank of the ridge is associated with the formation of a releasing bend (Crowell, 1974) in the deeper structural stepover in the fault pattern (Stoker & Hitchen, 1988). The orientations and positions of faults suggests the deformation was produced during an attempted stepover of the the major northerly NE-SW trending deep fault after the late Palaeocene-early Eocene. Normal movements on the faults is also compatible with extensional stresses within the releasing bend. Folds on the northern flank of the ridge also suggest a pervasive dextral shear across the ridge, with the same trend as strike-slip movements within the Faeroe Basin. The formation of a Palaeogene depocentre adjacent to the releasing bend in the northern Rockall Trough (Stoker & Hitchen, op. cit.) is compatible with this interpretation. Minor downfaulting of the edge of Wyville-Thomson Ridge is consistent with partial formation of an extensional duplex, associated with the bend, in the manner of Woodcock and Fischer (1986).

Folds within the structural stepover are, however, not compatible with a NW-SE dextral shear (cf. Sanderson & Marchini, 1984). The compression orientation from the anticline-syncline pair suggests a subsequent, and post-Palaeocene, sinistral shear (Fig. 4.20b). The absence of a pervasive deformation, with this sense of shear, on Wyville-Thomson Ridge implies a less influential phase of compression in northern Rockall Trough.
Figure 4.20


b) Interpretation of the Wyville-Thomson structure in terms of two phases of strike-slip movement along the bounding faults of the northern Rockall Trough. Strain ellipses for the direction of strike-slip movement show orientation of compression and extension for the two phases.
4.8 The role of Rona Ridge in Eocene sedimentation

The onlap of the seismic sequences onto Rona Ridge, and into the West Shetland Basin is complicated by the presence of seabed multiples on the seismic lines. The Eocene succession becomes thinner and in places, more sand-prone, precluding any wireline correlation. However a lithostratigraphy can be defined that encompasses a number of the seismic sequences (Fig. 4.21).

The thickest sand succession is observed in well 205/20-1, and consists of a medium to coarse-grained sand with traces of gravel and lignite interbeds. The downlapping internal configuration of corresponding seismic sequences adjacent to Rona Ridge implies progradation of the shelf-deltaic sequence during Elb (Fig. 4.22a). The lignites may represent the top of the Elb sequence. Corresponding sands are traceable to the north overlying a claystone which changes from a red-brown to dark brown and blue-grey. This colour change is possibly associated with the chemical alteration of the top of the Eocene tuffs. The sands are less continuous and become interspersed with grey-black siltstones, and are taken as indicating the edge of the shelf-deltaic sequence in the immediate post-tuff deposition. This is supported by the lack of similar sands in wells on the northern part of Rona Ridge, where grey to light grey siltstones and limestones were deposited. It is probable that the abandonment of the shelf, implied by the lignites in 205/20-1 is contemporaneous with the fluctuations in deposition at the shelf edge. An Elc timing is probable for this event. The shelf edge appears not to have moved significantly to the north throughout the Eocene, remaining on the southern extent of Rona Ridge, south of quadrant 206, until late Eocene-early Oligocene. During this time shelf sands are observed as far north as 207/2-1, and may be contemporaneous to the sands in E4 in the far north of the basin.

The continuity of the shelf-type sands throughout 205/20-1 suggests that the southern part of Rona Ridge was established as a structural
Figure 4.21 Correlation of the lithologies in the Eocene to Oligocene along Rona Ridge. This diagram shows the formation of the Eocene shelf wedge on the southern part of Rona Ridge, which does not prograde into the northern Faeroe Basin until late Eocene - early Oligocene.
Figure 4.22. A series of interpretations of the development of the Faeroe Basin during the Eocene, highlighting the tectonic activity and changes in sedimentation.

a) Sequence E1b.

b) Sequence E1c
Figure 4.22

c) Sequence E2.

d) Sequence E3.
Figure 4.22

e) Sequence E4.
high and was broadly stable throughout the Eocene. There is, however, evidence of movements of the ridge relative to the basin throughout the Eocene. These are based on the differences between sedimentation in the Faeroe Basin and on Rona Ridge, as well as lateral variations along the ridge structure.

The E2 shelf sub-sequence is not observed in the northern North Sea and has probably been removed by the E3 relative sea level fall. Consequently, the presence of the well-preserved E2 shelf sequence in the southern Faeroe Basin indicates a tectonic event. The subsequent restriction of the shelf to Rona Ridge implies an increased palaeo-relief. Both this and the complete preservation of the E2 shelf sub-sequence, must be caused by rapid subsidence of the basin or uplift of the ridge (Fig. 4.22d). Either process would create a marked palaeo-relief between the tectonic elements favouring slope and basin floor deposition in the southern Faeroe Basin. In-plane compressional stress would indeed result in both effects (Karner, 1986; Cloetingh et al., 1987).

The thinning of the Palaeogene succession in quadrant 205SW (around well 205/23-1, and shown by Evans & Mykura (1985)) suggests a subsequent inversion in the West Shetland Basin. The sediments at seabed are Maastrichtian to Danian in age and suggest exposure of the sediments adjacent to the Shetland Spine Fault during the Tertiary. No definite age can be placed on the timing of this exposure, but it is tempting to correlate this event with the mild Eocene inversion in the Faeroe Basin.

The succession in the southern extension of Rona Ridge, is similar to that observed on the structural high. Well 202/3-1 contains a 400m thick sequence of coarse sands and lignites. Tuffaceous claystones below the sands have a sonic log character consistent with Ela, below Elb (showing the characteristic velocity decrease) and probably Elc. Similar sequences are observed in 202/3-2 and 202/2-1. The succession in 202/8-1 is clay-dominated with thinner interbedded sands. This represents the edge of the shelf. Some
evidence of mild inversion of the underlying fault structure can be observed (Evans & Kirton, 1986; their section 2). This inversion is not reflected by the inferred direction of transport for the shelf sediments, and is concluded to be the effect of a later Oligocene inversion. It can be concluded that the focus of shelf sedimentation was in the southern part of Rona Ridge. This region was influenced by tectonic controls on the basin and margin.

In conclusion, Rona Ridge controlled the deposition of Eocene shelfsediments in the Faeroe Basin. Rona Ridge acted as a structural block, separating the deformation and observed compression in the Faeroe Basin from the marginal marine sediments of the West Shetland Basin controlled by global sea level changes. Consequently, global changes in sea level and tectonic events - in the southern Faeroe Basin - have been shown to be independent and superimposed upon each other. The northern Faeroe Basin appears to have undergone quiescent subsidence throughout the Eocene disturbed only by the regression and sea level fall associated with sequence Elb, and shelf progradation during sequence E4.

4.9 Summary and sea level curve for the west of Shetland

The construction of the Eocene sea level curve for the Faeroe Basin (Fig. 4.23) is based upon the succession in the southern part of the basin. The sequences are difficult to trace into the West Shetland Basin. A relative onlap curve is therefore impossible to define, although it is possible to construct an uncalibrated qualitative curve for changes in sea level. The Eocene observed in wells in the West Shetland Basin implies invariant shelf conditions throughout the Eocene.

The details of the sea level curve are therefore based on the sequence in the basin itself. This is contaminated by phases of tectonic subsidence, as discussed. The changes in sea level are taken as being defined by the sedimentation type. In general, widespread rejuvenation of submarine fans is taken to represent a
Figure 4.23 A comparison of the sea level curves for the Eocene successions east and west of the Shetland Platform. Curves are comparable, and highlight tectonic phases within the Faeroe basin, superimposed on the sea level curve.

<table>
<thead>
<tr>
<th>QUIESCENT PHASE</th>
<th>ACTIVE PHASE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Subsidence of</td>
<td>Uplift of</td>
</tr>
<tr>
<td>Eriand Igneous</td>
<td>Faeroes Plateau and</td>
</tr>
<tr>
<td>Faeroes Complex</td>
<td>Southern</td>
</tr>
<tr>
<td>Basin</td>
<td>Increase</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>RELATIVE COASTAL ONLAP CURVE</th>
</tr>
</thead>
<tbody>
<tr>
<td>E4 - E2 - E1 - E0</td>
</tr>
<tr>
<td>Lower - Middle - Upper</td>
</tr>
</tbody>
</table>

Delta progradation

Submarine fan activity

Judd Fault activity

Fault activity

Influencing palaeogeography

Basin tectonic events

Delta progradation

Submarine fan activity

Judd Fault activity

Influencing palaeogeography

Basin tectonic events
Table 4.1 Comparison of events in the northern North Sea and Faeroe-Shetland Basin, based on the seismic stratigraphy of both basins, with relative sea level changes indicated.

<table>
<thead>
<tr>
<th>Sequence</th>
<th>Faeroe-Shetland</th>
<th>East Shetland</th>
</tr>
</thead>
<tbody>
<tr>
<td>E4</td>
<td>Sea level rise</td>
<td>Sea level rise</td>
</tr>
<tr>
<td></td>
<td>1) Prograding sandy shelf</td>
<td>1) Prograding sandy shelf</td>
</tr>
<tr>
<td></td>
<td>2) Sea level highstand system</td>
<td>2) Sea level highstand system</td>
</tr>
<tr>
<td></td>
<td>3) Probable continued fault movement &amp; inversion</td>
<td></td>
</tr>
<tr>
<td>E3</td>
<td>Sea level fall</td>
<td>Sea level fall</td>
</tr>
<tr>
<td></td>
<td>1) Shelf-deltaic system on Rona Ridge</td>
<td>1) Erosion of sequence E2</td>
</tr>
<tr>
<td></td>
<td>2) Deepening of southern Faeroe Basin - ?uplift of southern Rona Ridge?</td>
<td>2) No apparent change in subsidence</td>
</tr>
<tr>
<td></td>
<td>3) Turbidite fans in southern Faeroe Basin</td>
<td>3) Turbidite fans in basin</td>
</tr>
<tr>
<td></td>
<td>4) Fault movements and folding near basin axis confine submarine fans.</td>
<td>4) Fans concentrated in South Viking Graben</td>
</tr>
<tr>
<td>E2</td>
<td>Sea level rise</td>
<td>Sea level rise</td>
</tr>
<tr>
<td></td>
<td>1) Small-scale submarine fans</td>
<td>1) No submarine fans</td>
</tr>
<tr>
<td></td>
<td>2) Uplift of Rona Ridge and shelf sediments on Rona Ridge</td>
<td>2) No shelf sedimentation</td>
</tr>
<tr>
<td></td>
<td>3) Fault movements cause folds in basin</td>
<td>3) No apparent tectonism</td>
</tr>
<tr>
<td>E1c</td>
<td>Sea level rise</td>
<td>Sea level rise</td>
</tr>
<tr>
<td></td>
<td>1) Shelf sediments on Rona Ridge</td>
<td>1) Shelf sediments on East Shetland Platform</td>
</tr>
<tr>
<td></td>
<td>2) Claystones in basin</td>
<td>2) Claystones in basin</td>
</tr>
<tr>
<td>E1b</td>
<td>Sea level fall</td>
<td>Sea level fall</td>
</tr>
<tr>
<td></td>
<td>1) Shelf and deltaic sedimentation in basin</td>
<td>1) Development of major submarine fan system in basin</td>
</tr>
<tr>
<td></td>
<td>2) Sediment derived from Faeroe Plateau</td>
<td>2) Incision of SE-directed canyons on platform</td>
</tr>
<tr>
<td></td>
<td>3) Tectonism - regional, probably thermal uplift of basin</td>
<td>3) Tectonism - Tilting of East Shetland Platform</td>
</tr>
<tr>
<td>E1a</td>
<td>Sea level rise</td>
<td>Sea level rise</td>
</tr>
<tr>
<td></td>
<td>1) Tuffaceous claystone deposition</td>
<td>1) Tuffaceous claystone deposition</td>
</tr>
</tbody>
</table>
higher sedimentation rate on the shelf and into the basin - resulting from relative sea level fall. Sequences dominated by parallel reflections may represent transgression of the shelf, and consequent fall in sedimentation rate - indicating sea level rise.

The tuffaceous claystones of sequence Ela are again indicative of the inundation of the late Palaeocene shelf. At this time the basin was undergoing progressive landlocking by lava flows (Nudge & Rashid, 1987), after which it might be supposed, sea level effects would be independent of global changes. However, the basin did not become entirely landlocked and rapid transgression was both subsidence and sea level controlled. Since the Erlend Igneous Complex is not an influence on post-Palaeocene sedimentation, the formation of a seaway connecting the Faeroe and Møre Basins by subsidence is the probable cause for transgression in the northern Faeroe Basin. The plateau basalts in the southern Faeroe Basin are a source for the Elb sediments. Hence, transgression in the southernmost part of the Faeroe Basin is induced by sea level rise, and not solely by the formation of a northern seaway to the Møre Basin and subsidence of the northern Faeroe Basin.

The Elb regression - indicated by delta progradation in the south - implies a relative sea level fall. The comparison between the northern North Sea relative onlap curve and the Haq et al. (1987) global sea level curve, imply a tectonic event throughout Elb. This supports the concept of a basinwide tectonic event in the Faeroe Basin, rather than global sea level changes. The direction of sediment transport in the southern Faeroe Basin, and the role of the plateau basalts implies a northern tilt to the basin floor during Elb deposition, associated with quiescent subsidence of the Erlend Igneous Complex and northern flood basalts. Debris flows in the northern Faeroe Basin may have been caused by this rapid subsidence, and oversteepening of the Lower Eocene shelf edge. Alternatively, they may have been caused by fault movements (cf. Normark & Gutmacher, 1988). Either way, a tectonic mechanism is concluded.
The quiescent nature of the sequence Elc sedimentation is consistent with the submergence of the Elb deltaic sequences, in a sea level rise. This is similar to the nature of the sequence in the northern North Sea. Loss of the Faeroes Plateau as a sediment source may reflect rapid subsidence of the basalts. The delay between emplacement of the basalts and the observed subsidence of the southern Faeroes Basin suggests a tectonic, possibly volcanic related mechanism.

The two sub-sequences of E2, are consistent with highstand deposition. The lowermost sub-sequence represents the onset of a sea level rise, and the basin infill by marine onlap. The uppermost sequence consequently represents the highstand shelf system of this sea level rise. The initiation of fault movements at this time, and their influence on deposition within the Faeroes Basin, does not appear to rejuvenate the Shetland Platform as a sediment source.

The formation of submarine fans in sequence E3 is consistent with the sea level fall invoked for this sequence in the northern North Sea. Basin topography, resulting from folding, causes confinement of the fans. It is possible that, far from being eustatic, the rejuvenation of the Shetland Platform may be related to this compression (cf. Cloetingh et al., 1987). Indeed, downwarping of the basins and uplift of flanks is a consequence of compression. However, the correlation of the sequence to a coincident global sea level fall in the late Middle Eocene implies that compressional uplift of the platform was masked by the more important eustatic/regional influences.

The highstand shelf wedge system in E4 is similar to the northern North Sea. Both sequences E3 and E4 are consistent with a tectonically inactive Shetland Platform, with sedimentation influenced only by global changes in sea level.

In conclusion, it can be seen that the same seismic sequences are recognisable in both the northern North Sea and the Faeroe Basin.
The Shetland Platform, between the Eocene subcrop limit east and west of Shetland, represents an area of only 100 km wide by 260 km long. Events influencing the margin of one basin were, consequently, likely to affect the margin on the opposite side of the platform. Although tectonic subsidence events are more important in the Faeroe Basin, these sequences show a consistent pattern in the changes in sediment type and distribution. Tectonic pulses can be isolated in the middle and late Eocene that affect sediment distribution without over-riding the controls imposed by global sea level changes. Rona Ridge is the structural element that plays the most part in the tectonic evolution of the Faeroe Basin through the Eocene. Although a topographical high throughout the Tertiary, Wyville-Thomson did not apparently play a part in the sedimentation of the Faeroe Basin. This may have been because the feature was submerged shortly after formation (McKenna, 1983).
Chapter 5

Tertiary tectonic activity and the evolution of the NE Atlantic

5.1 Introduction

A comparative study of the seismic stratigraphies of the East Shetland and the Faeroe Basins has shown a number of subtle, but recognisable, tectonic phases. These occur within the apparently quiescent subsidence of these basins (Ziegler, 1987) during the Eocene. The effects of the tectonism is small, having little observable influence on the relative sea level curve across the Shetland Platform. However, the deformation indicates a marked asymmetry in the development of the basins west and east of the Shetland Platform. This asymmetry points towards the influence of the North-east Atlantic in Palaeogene sedimentation (Vann, 1978). The mild deformation requires some explanation within the tectonic framework of the North-east Atlantic region. Similarly, it is important to the controls on platform subsidence to know if the deformation is part of a regional tectonism, or simply a local or marginal phenomenon. Hence a qualitative measure of the stresses imposed on the Shetland Platform can be made, that can equally be applied to similar platform and massif structures.

The tectonic episodes that are most important to this study - and have been defined in the previous two chapters - are as follows:-

i) An early (but significantly, not lowermost) Eocene uplift and eastward tilting of the Shetland Platform. A shelf-deltaic sequence in the Faeroe Basin indicates a broad uplift of the basin, simulating a sea level fall that is not observed on the global sea level curve of Haq et al. (1987). Similarly, rejuvenation of submarine fans in the East Shetland Basin suggest an eastward tilting of the East Shetland Platform.

ii) A mild (middle Middle Eocene to ?Lower Oligocene) inversion of the centre of the southern Faeroe Basin indicating a pervasive NW-SE
oriented compression across the basin. This is not observed in the
East Shetland Basin

The East Shetland Basin and Faeroe Basin border the NE Atlantic
Ocean (Fig. 5.1). This active tectonic province might be expected to
have had the greatest effect on the adjacent intra-continental
basins, rather than invoking the Alpine influence observed in the
southern part of NW Europe (Ziegler, 1981, 1982, 1987; Glennie &
Boegner, 1981; Beach, 1987). Ziegler (1987) argues that periods of
relaxation in the compressive stresses induced by Alpine collision
are responsible for the sea floor spreading in the northern North
Atlantic and Norwegian-Greenland Sea. It can be argued that the
Alpine compression has an important effect on the whole of NW
Europe, not simply on the inverted basins in Central Europe
(Ziegler, 1982). The deformation directly associated with the
compression however, decreases towards the north. The absence of
post-Palaeocene compression structures in the East Shetland Basin
might be taken as evidence that the Eocene tectonism observed in the
southern Faeroe Basin is not as widespread as expected for
compressive stresses derived purely from Alpine tectonics, though
this is discussed in more detail within this chapter. It is tempting
to link the minor inversion of the Faeroe Basin to the interaction
of the Alpine compression and compressive stresses resulting from
the changes in spreading rate and shifts in spreading centres
throughout the Tertiary. The effect of the latter on the Atlantic
marginal basins has not been discussed in any systematic form.

This chapter therefore documents the tectonic evolution of the NE
Atlantic during rifting - from Jurassic-Cretaceous, with the final
phase in the Palaeocene - into the post-rift phase during the
Eocene. The aim is to produce a history of the Atlantic that is
applicable to the borderland basins, and to assess the role of
Atlantic tectonism and opening in deformation of the basins adjacent
to the rift zone during these times. Consequently, the evolution
includes documentary evidence of the Palaeogene succession in the NE
Atlantic Province north of the Charlie Gibbs Fracture Zone, from the
Figure 5.1 Location map for the NE Atlantic and marginal basins.
northern Norwegian margin to southern Rockall Trough. Cretaceous rift subsidence, the phase preceding Atlantic break-up is discussed. This represents the crustal regime prior to the final Atlantic rifting, and contrasts sharply with Palaeocene to Eocene sedimentation. Aspects of the Alpine influences on the North Sea are also discussed.

5.2 Cretaceous development of the North-east Atlantic.

The Triassic to Jurassic development (Fig. 5.2) of basins on the continental shelves bordering the NE Atlantic is well-documented. However, little detail is known of the areas adjacent to the Atlantic ocean basins (eg. Møre Basin, Faeroe Basin and Rockall Trough). Middle Jurassic rifting and subsidence almost certainly occurred in the Faeroe Basin (Haszeldine et al., 1987; Kitchen & Ritchie, 1987). Subsidence in the East Shetland Basin occurred during the Upper Jurassic (eg. Hay, 1978; Gray & Barnes, 1981; Ziegler, 1981, 1982; Beach et al., 1987; Badley et al., 1988). Extension during the Jurassic consequently defined the structural extent of the Shetland Platform. In the Cretaceous, for the most part, there was continued evolution of the extensional system, but also renewed subsidence in a number of the Atlantic marginal basins. Within this, however, there are spatial variations in the relative amount and style of subsidence. These are discussed below, basin by basin.

The last phases of faulting ceased in the lower Cretaceous in the East Shetland Basin (Ziegler, 1981; Beach et al., 1987), although Badley et al. (1988) argue that the lower Cretaceous (Ryazanian onwards) merely represents the isostatically controlled thermal subsidence postdating extensional fault movements. The rest of the Cretaceous is represented by a basin infill of shales in the Viking Graben and East Shetland Basin (Wheatley et al, 1987), the onlap of which encroaches the edge of the East Shetland Platform during the Campanian to Maastrichtian (from released biostratigraphy of UK well 9/16-1).
Figure 5.2 Correlation of tectonic events in the marginal basins, preceding the final phase of Atlantic rifting and seafloor spreading.
On the western side of the Shetland Platform, however, the tectonic regime was still one of active extension, with emergence of Rona Ridge and the deposition of coarse sands and gravels in the Albian (Ridd, 1981; Hitchen & Ritchie, 1987) into the Faeroe Basin. Movement of the Shetland Spine Fault, that ceased at the end of the Albian, preserved a sequence of continental sediments within the West Shetland Basin. In the post-Albian the depocentre shifted into the centre of the Faeroe Basin (Hitchen & Ritchie, op. cit.). The West Shetland Basin and Rona Ridge appear to have been maintained as structural high areas throughout the Cenomanian to Coniacian, and were transgressed by the Santonian to Maastrichtian mudstones and shales. Tectonic subsidence may be the cause of onlap across Rona Ridge, slightly earlier than on the East Shetland Platform. Reactivation of faults in a moderate extension event (20-30%), between the Turonian and Maastrichtian (Duindam & van Hoorn, 1987) resulted in occasional emergence and erosion of Rona Ridge, but was otherwise masked by the global sea level rises occurring at this time. Igneous activity at the end of the Cretaceous emplaced a number of igneous centres, notably Erlend and Brendan's Dome.

Price & Rattey (1984) (Fig. 5.3a) established a regional model for the mid-Cretaceous events from the Rockall Trough to the Voring Basin. In this, it was postulated that the development of ocean basins occurred in Rockall Trough, Faeroe, Møre and Voring Basins in an NNE-SSW trend. The spreading centre in the basins was offset by a series of ENE-WSW transform faults, about a pole of rotation north of the Voring Basin. The degree to which seafloor spreading occurred therefore decreased from southern Rockall Trough, where magnetic anomalies similar to oceanic crust patterns are observed (Roberts et al., 1981), to sill intrusion in the Møre Basin (Price & Rattey, 1984) which might now include the volcanic centres of Nelson & Lamy (1987). Bukovics et al. (1984) dispute any evidence for formation of oceanic crust in the Møre and Voring Basins, and the formation of oceanic crust in the centre of the Faeroe Basin and Rockall Trough is equivocal. Indeed there is some equivocation as to the presence of oceanic crust in the southern Rockall Trough (Megson, 1987).
Despite the presence of the Axial Opaque Zone (Ridd, 1983) close to the axis of the Faeroe Basin, there is little evidence of full scale seafloor spreading throughout the Mesozoic history of the Faeroe Basin (cf. Smythe, 1983). A number of seismic interpretations of the deep structure of the Faeroe Basin show a series of westward dipping reflectors on the eastern flank of the basin, and eastward dipping reflectors on the western flank, adjacent to the Faeroe Plateau. These have been interpreted in the style of continental extensional tectonics, as a series of tilted fault blocks (Nudge & Rashid, 1987) above a detachment fault (Gibbs, 1987; Earle et al., in press). The crustal thickness of 8 km, in the centre of the Faeroe Basin (Smythe, 1983; Bott, 1984), is similar to the thicknesses of layers 2 and 3 of oceanic crust, but may similarly represent highly extended continental crust. The crustal thickness is less than that of the inferred continental crust in the northern Rockall Trough (Roberts et al., 1988), and suggests greater extension in the Faeroe Basin. The latter is a narrower basin, and may have concentrated extension within a narrower zone. With these constraints there is little space left in which to place a spreading centre. Indeed, the crustal structure of the contiguous Rockall Trough is suggestive of greatly extended continental crust (Roberts et al., 1988) and there is geochemical evidence for continental crustal contamination of Palaeocene lavas in northern Rockall Trough (Morton et al., 1988). Consequently if the Faeroe Islands represent igneous extrusion over continental crust (Casten 1973; Casten & Nielsen, 1975), there is little reason to believe that a fully-fledged spreading centre formed. Any Mesozoic igneous activity along the axis of the Faeroe Basin may have been of the form of thin dyke intrusion into greatly extended continental crust, in the manner of Megson (1987). It is therefore possible that the deformation prior to, and during the Cretaceous was wholly intra-continental, involving only minor intrusion of oceanic-type intrusives along the axes of basins undergoing major extension (Fig. 5.3b).

Along the Norwegian margin, to the north of the Shetland Platform, the Møre Basin were already well established by the start of the
Figure 5.3 Comparison of the different models for basin formation in the mid-Cretaceous. a) after Price & Rattey (1984); b) based on (Roberts et al., 1988), (Negson, 1987) and (Morton et al., 1988).
Cretaceous. Fault movements were synchronous with the Viking Graben and had evidently ceased by the early Cretaceous (Nelson & Lamy, 1987). Broad, thermal subsidence was dominant for most of the Cretaceous, although early Cretaceous lavas (Hamar & Hjelle, 1984) and Upper Cretaceous volcanic centres have been interpreted (Nelson & Lamy, 1987). The latter have been inferred as being part of a regional elevation of the basin as a result of early melting prior to ocean crust formation. The Haltenbank segment, to the north-east of the Møre Basin shows signs of fault reactivation and formation in the early Cretaceous, decreasing in activity into the Upper Cretaceous (Gabrielsen & Robinson, 1984). Subsidence in the Møre Basin, however, was most rapid in the mid-Cretaceous and probably induced by thermal contraction and gravity loading of extended continental crust (Bukovics et al., 1984). Non-uniform extension and the recession of an underlying mantle plume (if present, although there is no real evidence for this) might have played a part in the rapidity of Møre Basin subsidence.

In conclusion, the regional tectonic setting during the Upper Cretaceous is one of precursor tectonic movements and volcanism prior to the more dramatic activity in the Palaeocene to lower Eocene.

The significance of this is that the structural configuration on the borders of the Palaeogene Atlantic rift was comprised of a heterogeneous continental crust. This was composed of regions of extended and unextended continental crust, bounded by deep lineaments of Caledonian and younger ages. Although major crustal or rheological boundaries were formed by earlier rifting, such as Rockall Trough or the Viking Graben, these did not become the subsequent locus of seafloor spreading. Consequently, the Palaeocene rifting occurred to the NW of Rockall Trough and the Faeroe Basin, in a NE-SW orientation. It is argued below that what was exploited and reactivated in these borderland basins was merely the structural configuration of continental crust that was in a preferential orientation to the principle stress directions during the
Palaeocene. This stress was induced both by rifting and the presence within the rift zone of the Iceland Hotspot.

5.3 Palaeocene - tectonism before crustal separation

The final phase of Atlantic rifting was one of intense igneous activity and associated deformation of the pre-existing basins outlined above. This deformation was variable throughout the Atlantic Province, and in all cases reactivated pre-existing structure rather than generating a new series of folds and faults. A regional map of the deformation is shown in Fig. 5.4.

The Palaeocene in the North Sea was associated with the formation of submarine fan sands, indicating a rejuvenation of the stable source areas (Rochow, 1981). The disruption of the quiescent subsidence characterising the Upper Cretaceous did not occur until the Thanetian, although there is evidence of volcanic ash as far back as the Danian (Fitch et al., 1978). An unconformity between Maastrichtian and Danian chalks probably represents a change in sea level in the early Danian (cf. Haq et al., 1987). Subsequent Thanetian erosion and incision of the Danian occurred at the basin margins in the Central North Sea (Stewart, 1987), and reworked chalks (Johnson, 1987) and Danian and Cretaceous foraminifera occur in the Late Danian to early Thanetian (Stewart, 1987). Knox et al. (1981) recognised two distinct tectonic phases within the central North Sea for the Thanetian of i) Central Graben subsidence, uplift of the Shetland Platform and widespread regression and ii) subsidence of the Moray Firth Basin, accompanied by rejuvenation of the Scottish Highlands, volcanism and widespread transgression. Morton (1982) found similar phases in the Viking Graben.

In the East Shetland Basin, the isopachs for the early Palaeocene are shown to be thickest in the hanging wall of the bounding faults (Mudge & Bliss, 1983), and therefore indicative of contemporaneous fault movements rather than pure sedimentary facies controls.
Figure 5.4 Regional picture of Palaeocene tectonism in the period prior to seafloor spreading and ridge development.

References: Steel et al. (1984); Caselli (1987); Upton (1988); Roberts et al. (1984b); Lake & Karner (1987); Glennie & Boegner (1981); Walker & Cooper (1987).
Structurally, the reactivation of the bounding faults of the East Shetland Basin during the Palaeocene has been interpreted in terms of a new rift phase (Beach et al., 1987). Badley et al. (1988) dispute the evidence for a true rift, arguing that fault movements are associated with the accommodation of a rejuvenation of sedimentation rate into - and consequently increased loading of - the East Shetland Basin and Viking Graben. The absence of widespread extension in the Viking Graben supports a loading mechanism. However, this structural style is very similar to fault movements in the Faeroe Basin. The fault movements are discussed in a later chapter.

In the Faeroe Basin the same base Danian unconformity or hiatus (Hitchen & Ritchie, 1987) is observed. A subsequent tectonic pulse in the early Palaeocene led to reactivation of the Shetland Spine Fault in the middle and southern parts of the basin, and the formation of sub-basins adjacent to the fault. The Bryozoan Sands of Hitchen & Ritchie (op. cit.) are considered an equivalent to fault-controlled sequences of Mudge & Bliss (1983) in the East Shetland Basin (Middle Palaeocene to Danian). Hence both basins exhibit the same rejuvenation of the Shetland Platform, and indeed other intra-basinal highs such as Rona Ridge and the Halibut Horst, enhancing the interpretation of a tectonic uplift in the late Danian-early Thanetian. Global sea level falls in the Danian may have contributed to the rejuvenation of the platform as a sediment source, but it is difficult to explain how this alone produced the intra-basinal relief on highs such as Halibut Horst. Subsequently, the Faeroe Basin became progressively restricted by the formation of the sub-aerial lavas of the Faeroe Plateau (Mudge & Rashid, 1987).

The Møre Basin, north of the Viking Graben, shows evidence for unaffected subsidence and deposition throughout the Palaeocene. The Voring Basin, however, appears to have been undergoing uplift, with deposition of a comparatively thin lower Tertiary sequence. Timing of the uplift is unclear, and this phase of dextral transpression (Bøen et al., 1984) of both the Voring Basin and Halten Terrace
(Caselli, 1987) may extend into the Eocene. Palaeocene "fans" (Mutter, 1984) near the foot of the Vøring Plateau Escarpment indicate marine deep water conditions at this time.

There is a distinct change in the style of Palaeocene tectonism away from the Atlantic rift zone (Figure 5.4). The basins in southern England and central Europe were at this time undergoing inversion and uplift as a result of Alpine compression. This is significantly different to the Atlantic margin, where uplift was limited to structural highs, not basins. Therefore, the direct consequences of Alpine compression may not be observed in the northern NW European basins, and the direct influence of the Atlantic rifting is more apparent in the tectonism at this time.

5.4 Pre-rifting volcanism on the Atlantic margins

The volcanism on the Atlantic margin is extensive (Fig. 5.4). Tertiary igneous rocks are observed as far apart as Greenland and the British Isles, a pre-Atlantic extent of 1200-1300 Km, along the 2000Km length of the North-east Atlantic Rift. This represents a total volume of greater than 1-2x10$^{27}$Km$^3$ (White, 1988). The volcanism is observed in the form of oceanward dipping reflectors (representing MORB-type lava flows, along the margin of the present ocean basins) newly formed eruption centres, basaltic lavas and dykes on the stable basement of Greenland and the British Isles and flood basalts and sills in the Mesozoic basins (ie. Faeroe, Vøring and More Basins) (Upton, 1988). Tertiary eruptions associated with igneous complexes within the Mesozoic basins (eg. Erlend and Brendan Igneous Complexes in the Faeroe Basin and Anton Dohrn and Rosemary Bank Seamounts in the Rockall Trough)) simply represent the rejuvenation of pre-existing Upper Cretaceous igneous complexes (Duindam & van Hoorn, 1987), rather than the formation of new centres.

Dating of Palaeocene basalts, discussed by Been et al. (1984), indicates extrusion during the late Palaeocene on the Vøring
Plateau. This is contemporaneous with the extrusion of the Faeroes Lower Series and prior to the formation of the first Cenozoic oceanic crust in the north-east Atlantic (Magnetic Anomaly 24 (Nunns, 1983) in the Lower Eocene (Harland et al., 1982; Berggren et al., 1985). Overall, volcanism in the Atlantic Province was continuous throughout the Middle and Upper Palaeocene. The main phase of Hebridean volcanism is dated as Middle Palaeocene (ca. 60-57 Ma.; Macintyre, 1977), whereas the Upper and Middle Series of basalts on the Faeroes Plateau (Lund, 1983) were still being extruded into the Lower Eocene.

The igneous activity of the North-east Atlantic is associated with, and concentrated around, the Iceland Hotspot. Rift tectonics alone are not responsible for this activity since margins such as the Biscay Margin show only simple faulting and subsidence, unassociated with any intrusive or extrusive volcanism (Montadert et al., 1979; Le Pichon & Barbier, 1987; White, 1988). The topographic effect of the hotspot is, at present, Iceland, representing an uplift of 2.5 Km (White, 1988). Bathymetrically, Iceland forms part of a transverse ridge, trending NW-SE, from Greenland to the Faeroes. This ridge is composed of 30 km thick oceanic crust (Bott, 1983), formed between 55 and 35 Ma. Vink (1984) has attempted to explain the formation of Iceland-Faeroe Ridge in terms of the migration of the ridge zone over the hotspot and channelling of hotspot asthenosphere into the spreading ridge.

The Iceland Hotspot is the obvious source for local tectonism along the borderlands of the Palaeocene Atlantic rift zone. The extent of the volcanism is consistent with the dimensions of hotspot influences on unruptured crust (White, 1988; cf. Courtney & White, 1986) (Fig. 5.5). The model invoked is one in which the mantle plume associated with the Iceland Hotspot spreads out under the crust. In the case where the crust is homogeneous, no volcanism is observed outside the rift zone, similar to the Lower to Middle Palaeocene evolution of the Scottish Massif. Uplift of the stable continental regions, such as the Shetland Platform and Scottish Massif (Watson,
Figure 5.5 A model for uplift and volcanism in the Faeroe-Shetland Basin during the Palaeocene. Note how the model explains tilting of the Shetland Platform and shallowing of the Faeroe-Shetland Basin. (Cf. Courtney & White, 1986).
would have occurred as a result of the thermal and buoyancy effects of an underlying mantle plume. However, inherent crustal weaknesses are likely to have been exploited by partial melts generated in the mantle plume as it spreads out under the lithosphere.

Vann (1978) proposes such a model for the formation of the late Palaeocene dyke swarms of the Hebridean Province. Igneous centres occur at the intersection of dyke swarms and major NE-SW faults. This occurs in a region of relatively homogeneous, crystalline basement. The proximity of regions of extended and fractured crust, such as the Mesozoic basins, probably resulted in the intrusion of far greater volumes of igneous material into these regions. At the same time, reactivation of basin faults and minor amounts of extension within the Faeroe Basin may have contributed small amounts of partial melting below the basin. Hence, much of the volcanism observed in the borderlands is concentrated in the Mesozoic basins, and rejuvenated Cretaceous volcanic conduits. The high number of sills in the basins is a response to the thickness of sediment and are probably developed in the widespread form of Sheridan (1981), rather than directly above a spreading centre (Gibb & Kanaris-Sotiriou, 1988). The Faeroe flood basalts probably represent the intense intrusion of basaltic magma in and around a highly faulted continental crustal block. Similarly, the linear nature of Wyville-Thomson Ridge suggests that the 10 Km of basalts that form the ridge (Roberts et al., 1983) were intruded along a pre-existing crustal lineament transverse to the Rockall and Faeroe Basins. The massive thickness of the basalts in the ridge may be the result of the proximity of the northeastern edge of the lineament to the rift zone and hotspot centre.

Basaltic sills are found in the West Shetland Basin, close to the edge of the West Shetland Platform (eg. well 206/13-1; Hitchen & Ritchie, 1987). The lack of Palaeocene volcanism in the East Shetland Basin suggests that the influences of the mantle plume were reduced or non-existent in the Viking Graben. Consequently, the
influences of the hotspot must have decreased across the Shetland Platform. The likely response to this would have been to tilt the platform towards the east throughout the Middle to Upper Palaeocene.

The Palaeocene volcanism and tectonism can be linked to rifting associated with stresses around the Iceland Hotspot. The stresses reactivated Mesozoic structures along the Norwegian margin and in the Faeroes-Shetland Basin whereas the buoyancy effects of the underlying mantle plume were responsible for the domal uplift of the borderlands in the vicinity of the hotspot. The maximum extent of the uplift may have reached the middle of the East Shetland Platform, all of the Scottish Massif and the Inner Moray Firth Basin. This tectonic uplift occurs over a far wider extent that the observed volcanism. Mesozoic structures were exploited by partial melts from the hotspot, and were intruded along faults and crustal lineaments, extruded via pre-existing volcanic centres, or intruded through thinned continental crust. In places, crustal material was assimilated into the melts (Gariepy et al., 1983; Morton et al., 1988). Sills are common in the thicker sedimentary sequences, especially in the Faeroe Basin and Voring Basin. There is no evidence to suggest widespread reactivation of the intra-basinal spreading centres that may have existed in the Cretaceous.

5.5 Lowermost Eocene tectonism and the initiation of Atlantic seafloor spreading

Throughout this study, the base of the Eocene has been taken as the base of a tuffaceous claystone sequence, the Balder Formation. The deposition of the tuffs is widespread, and correlatable between the East Shetland Basin, Viking Graben and Faeroe Basin. Other events are also correlatable to the Upper Palaeocene-Lower Eocene, at therefore approximately the same time as deposition of the Balder Formation (Fig. 5.6). Sills in the Faeroe Basin have been K/Ar dated as lowermost Eocene (Hitchen & Ritchie, 1987), considered unreliable by Gibb & Kanaris-Sotiriou (1988), on the basis of heavy alteration. A more reliable Ar/Ar date of 50 Ma for a tholeiitic sill was
### Figure 5.6

Tentative correlation of the volcanism associated with the Palaeocene-Eocene boundary, and the ensuing sedimentary sequences. Data from Roberts et al. (1984b) and Stoker et al. (1988).

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<th>STANDARD AGES</th>
<th>ROCKALL PLATEAU</th>
<th>NORTH HEBRIDES SHELF</th>
<th>WYVILLE-THOMSON RIDGE</th>
<th>FAEROES PLATEAU</th>
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- **Basalts**
- **Pyroclastic**
- **Epiclastic**
- **Coal-bearing Sequences**
- **Glauconic Horizon**
obtained by Fitch et al. (1988). Upper and Middle Series basalts were extruded onto the Faeroes Plateau at this time (Fitch et al., 1978). Gibb et al. (1986) used chemical similarities between sills and the Upper Series Faeroes basalts to conclude an Eocene age, but also a genetic link between the two styles. This idea has been expanded to the inference that the locus of igneous activity shifted from the Faeroe Basin to the Faeroes Plateau, during the Upper Palaeocene. The tectonic implication of this is that periods of Faeroe Basalt extrusion may be related to periods of increased basin subsidence resulting from the removal of asthenospheric buoyancy and decay of thermally induced uplifts. This might be represented in the sediment record as a series of transgressions, and indeed is similar to the transgressive-volcanic phases of Knox et al. (1981). The onlap curve of Stewart (1987) shows a number of transgressions which are at a similar time to the Lower and Middle phases of Faeroes Basalt extrusion (Fig. 5.7). This is best observed for the Lower Series. The subsequent regression - that might be expected during the deposition of the Intra-basaltic Series - is masked by a global sea level rise. The sea level fall at the start of deposition of the Balder Formation is not obvious in the East Shetland Basin although erosion in the Central Graben is seen at the top of Stewart's sequence 8. It is tempting to correlate the subsequent transgression, observed in the East Shetland Basin and Faeroe Basin, during deposition of the Balder Formation to the extrusion of the Faeroes Middle and Upper Series basalts.

More importantly, in that they represent a vast volume of igneous material, oceanwards dipping reflectors were extruded along the flanks of the rift zone. Magnetostratigraphic analysis suggests that the ocean-dipping reflectors were extruded immediately preceding magnetic anomaly 24B (early Eocene - Harland et al., 1982). The extrusion of these basalts appears to have been sub-aerial, suggesting a broad uplift of the rift zone. Vann (1978) argues that the magnitude of uplift in the vicinity of the Faeroes was of the order of 1-2 km.
Figure 5.7 A tentative correlation between the transgressive phases of Stewart (1987) and extrusion of the Faeroe Basalt Series. The correlation is obscured by a global sea level rise during sequences 7 & 8.
Tuffs in the northern Rockall Trough are of lower Eocene age (Jones & Ramsay, 1982), somewhat later than the age of the Balder Formation (Roberts et al., 1984b), and are interbedded with sandstones of shallow marine origin (Stoker et al., 1988).

On the western margin of Rockall Plateau, early Eocene sedimentation is characterised by the cessation of sub-aerially extruded (Harrison et al., 1984) basaltic lava flows (Roberts et al., 1984a). Coarse pyroclastic tuffs were deposited contemporaneous to the late Palaeocene-early Eocene tuffaceous sediments in the East Shetland Basin (Fig. 5.6). The subsequent alteration of the pyroclastic material to glauconite is taken to represent a shallow marine hiatus related to a major transgression immediately preceding Anomaly 24B (Morton, 1984; Backman et al., 1984). This is contemporaneous with the sea level rise and transgression indicated for the East Shetland Basin, during the deposition of the Balder Formation. It also appears to be contemporaneous to the extrusion of the basalts represented by the ocean-dipping reflectors, although the basalts on Rockall Plateau are below the glauconitised sediments. The transgression observed on Rockall Plateau is therefore immediately after the major basalt flows and tuff deposition of the early Eocene. An Atlantic influence can be invoked for the timing of this transgression, and it may represent the moment of crustal separation.

5.6 Eocene sedimentation & tectonics - Passive margin evolution

There is little published detail of the development of the Eocene succession after deposition of the lowermost Eocene tuffs. The following discussion is based on the seismic stratigraphic work presented in Chapters 3 and 4, for the East Shetland Basin and Faeroe Basin.

The subsequent development of the Atlantic margin was one of overall quiescence. The initiation of seafloor spreading resulted in the migration of the passive margins away from the influence of the
Iceland Hotspot and the cessation of all but minor volcanism in the Atlantic marginal basins. The overriding control on the sedimentation within the basins was probably eustatic sea level changes. However, adjacent basins were subsiding at different rates. Thus, the East Shetland Basin shows a regressive phase in the early Eocene, associated with fan development. At the same time, the regression in the southern Faeroe Basin is dominated by shelf and deltaic sedimentation. The northern parts of both basins, however, indicate a rapid subsidence that submerged the igneous centres and the northern extension of the Shetland Platform. This northern region is also the southernmost part of the Møre Basin and Norwegian Sea oceanic basin and was probably directly influenced by rapid subsidence of those basins at this time. The consequence of this would be to have opened up a seaway between the Møre, northern Faeroe and East Shetland Basins, that was similarly linked to the ocean basins in the Norwegian Sea and Greenland Sea (Talwani & Eldholm, 1977). The removal of the barrier between the Atlantic marginal basins and the ocean basins may be the cause of the influx of planktonic foraminifera that is misinterpreted by Mudge & Bliss (1983) as representing a transgression *sensu stricto* at the top of their unit 4. The seismic stratigraphy presented here has indicated that this transgression *sensu lato* may be defined by a micropalaeontological influx, but it is certainly not related to a rise in sea level in the northern UKCS, Atlantic marginal basins.

5.6.1 Causes of the early to middle Eocene regression

This still leaves the problem of the early to middle Eocene regression that developed a major fan system in the East Shetland Basin. The regression in the Faeroe and East Shetland Basins are not related to any change in sea level in the Haq et al. (1987) global sea level curves, nor is the regression observed on the Rockall Plateau (Fig. 5.6). The relative sea level fall therefore appears to be local to the northwestern margin of the UKCS. Canyonisation of the East Shetland Platform, shallow marine sedimentation in the Faeroe Basin and the importance of the southern part of the Faeroes Plateau during the deposition of unit Elb, all suggest a tectonic
uplift of this area with respect to the thermally subsiding basins to the north. This tectonic uplift appears to have been centred to the west, where the Faeroe Basin is uplifted regionally. The tilt of the Shetland Platform, inferred by the change in sediment type across the structural high, represents the effects at the edge of uplifted region.

Two models are proposed; a thermal model and a flexural model, of which the former is the preferred. The basis for the latter are that flexural uplift would be the result of the piling up of up to 10 Km of dense, basaltic lavas on the western flank of the Faeroe Basin, and on the Wyville-Thomson Ridge (Roberts et al., 1983) (Fig. 5.8). The consequences of this would be to develop a flexural moat (Ten Brink & Watts, 1985) along the flanks of the Wyville-Thomson Ridge and around the Faeroes Islands. An analogy to this is shown in Figure 5.8b. The effective elastic thickness quoted is a measure of the rigidity of the lithosphere at the time of loading, and is taken as the same as that for the oceanic crust around the Hawaiian Seamount (effective age 44Ma - Ten Brink & Watts, 1985). The uplift ridge is east of the moat developed in the centre of the Faeroe Basin. Indeed, the uplift coincides with the edge of the Shetland Platform, and for this model can produce a relief between the moat and uplift bulge of 600-700m. These two features might be observed in the depocentre adjacent to the Wyville-Thomson Ridge during the Palaeocene-Eocene (Stoker & Hitchen, 1988), and the shallowing or delayed subsidence of the southern Faeroe Basin and southern Faeroe Plateau. However, the model assumes a very high effective elastic thickness (Watts et al., 1982) for continental crust that has undergone a complex and prolonged thermal history, including proximity to recently formed oceanic crust. Lower elastic thickness would merely result in a less regional effect, including less widespread, and lower amplitude flexural uplift. A more regional uplift of unextended continental crustal areas, such as the Shetland Platform must be invoked.
Figure 5.8 Model of a flexural loading analogue for the post-rift uplift and tilting of the Shetland Platform. a) The geophysical interpretation of the thick basalts associated with the Wyville-Thomson Ridge and the Faeroes Ridge. b) The consequences of loading of a lithosphere with rigidity of 44 Ma oceanic crust (in this case emplaced in the Albian), with an effective elastic thickness of 25 Km.
A thermal effect is therefore the preferred explanation for this phase of uplift. The effect of an underlying thermal anomaly such as a mantle plume, or simply the heating and thermal expansion of the mantle, may produce the observed uplift of the platform areas. Such phenomena underneath extended continental crust would result in the slowing down of subsidence or minor uplift, dependent on the amount of heating and the crustal thickness (McKenzie, 1978; Hellinger & Sclater, 1983). As for the flexural model, the focus of heating would need to be to the west to explain the eastward tilting of the Shetland Platform. However, the thermal source could be under the whole of the Faeroes Plateau and Wyville-Thomson Ridge, and affecting the whole of the Faeroe Basin and West Shetland Platform, diminishing rapidly to the north into the Norwegian Sea. This description adequately describes the residual effect of the Iceland Hotspot under the Greenland-Scotland Ridge, a source which is supported by the continuing extrusion of the Upper Series Faeroe Basalts after deposition of the Balder Formation (Roberts et al., 1984b). That regression occurs - rather than the transgressions observed to be synchronous with the earlier Faeroe basalt extrusions - suggests that the final extrusions of the Upper Series do not significantly alter the extent of the residual thermal anomaly. The subsequent transgression, during deposition of the Eic seismic sequence, relates to the migration of the region away from the vicinity of the Iceland Hotspot.

This migration away from Iceland during Eic deposition may have initially been fast, causing a rapid subsidence of the East Shetland Platform. Subsequent gradual subsidence resulted in the imposition of shorter period sea level fluctuations on a gradually subsiding, onlapped coastal margin on the East Shetland Platform. In the west, the migration was represented by the final subsidence of the southern part of the Faeroe Basin, and the Faeroes Plateau. At this time the north-south division of the basin subsidence disappeared and subsequent subsidence of basins was uniform.
5.6.2 Eocene compressive features

During the Middle to Upper Eocene the basins became tectonically quiescent with the major control on sedimentation being global changes in sea level. Modifying, but not inherently controlling, sedimentation was a phase of local compressional deformation in the southern Faeroe Basin. This deformation is probably synchronous to activity on Wyville-Thomson Ridge. Similar deformation is not seen in the East Shetland Basin, although small-scale, Eocene compression structures are observed in the Stord Basin (Biddle & Rudolph, 1988) (Fig. 5.9). The concentration of compressional structures around Wyville-Thomson Ridge nonetheless suggests a link between this structure and inversion.

Structural inversion of basins on the NW European margin has been previously observed. The major inversions, involving the reversal of basin faults and the removal of basinal sequences has been observed in most of the basins in southern England. These have been divided into two distinct tectonic phases - the Laramide (end Cretaceous-early Tertiary, in general the Palaeocene) and the Pyrenean (late Eocene to late Oligocene). A stricter dating of these phases is precluded by the lack of good dating of deformation elsewhere in NW Europe (Biddle & Rudolph, 1988). The tectonism associated with these two phases is directly related to resurgent phases of Alpine orogenic compression (Ziegler, 1982), as well as the Iberian collision.

The Laramide phase of inversion is associated with the initiation of the final phase of Atlantic and a period of Alpine collision. The whole of NW Europe was therefore in compression throughout this period, and the style of deformation related to the proximity of basins and highs to these two active regions. Hence, basins in southern England - close to the Alpine compressional front - were inverted by reversal of normal faults. Basins bordering the Atlantic were not so deformed, and highs were uplifted by the thermal and locally transpressive regime (Caselli, 1987) of the Atlantic rift
EOCENE POST-RIFT TECTONISM

Volcanism
- Lava flows, sills and dykes
- Oceanic spreading centres

Crustal response
- Uplift
- Subsidence
- Sediment transport direction

Figure 5.9 Regional map of the Eocene post-rift tectonism along the western NE Atlantic margin. Data from Steel et al. (1984), Mutter (1984), Srivastava & Tapscott (1986), Biddle & Rudolph (1988), Holloway & Chadwick (1986), Masson & Parson (1983).
zone. Volcanism was also characteristic of the Atlantic margin. Consequently inversion structures are more characteristic of the southern UK basins, and the North Sea south of thick Zechstein evaporites (Biddle & Rudolph, 1988). However, individual inversion structures are much more widespread. It is likely that in these cases, inversion only took place in regions where compressive stresses were concentrated, for instance, along the structural continuation of the Tournquist Line in the North Sea.

The idea that the two active regions affected the whole of the N European region is consistent with Zieglers (1987) explanation of the Tertiary Atlantic seafloor spreading, as a result of the relaxation of the Alpine compression at the end of the Palaeocene. This may have allowed basins close to the rift zone to undergo regional subsidence, whilst still continuing less severe inversion. Biddle & Rudolph (1988) conclude that the regional compression set up by the interaction of these two active areas was trending north-west.

The difference therefore between the southern Faeroe Basin (and Wyville-Thomson Ridge) phase of deformation, and the Laramide events is that the former is somewhat later than the latter, bridging the broad timespan between Laramide and Pyrenean events. The direction of compression is the same as that for the Laramide inversions. The region lies along a continuation of the Tournquist Line, and it has been postulated that this affected the position of the Faeroes Palaeocene volcanic centre (Pegrum, 1984). Movements on the Tournquist Line were sinistral transtension in the Jurassic and early Cretaceous. This was replaced in the late Cretaceous by dextral transpression (Pegrum, op. cit.), similar to that observed into the Eocene on the Judd Fault. The movement during the Laramide event is ascribed to compression in the Carpathian mountains, although this had evidently ceased by the Eocene (Ziegler, 1982). Deformation is traceable in a broad line along the Tournquist Line to the Central Graben, up to block 15/9 demonstrably during the Eocene (Pegrum & Ljones, 1984), and possibly to the Wyville-Thomson
Ridge. This later stage of movement has been ascribed to rifting in the Alpine belt. The effects of this lineament in the southern Viking Graben and outer Moray Firth Basin, are at present unknown.

The events in the southern Faeroe Basin are also correlatable to changes in the seafloor spreading patterns in the NE Atlantic. During the late Lower Eocene to middle Middle Eocene, the North Atlantic saw the emplacement of crust of magnetic anomalies 20 and 21. However, it has been suggested that this was the period of formation of a large part of the Iceland-Faeroe Ridge (Voppel et al., 1979; Vink, 1984), and the initiation of a ridge jump in the Norwegian Sea (Nunns, 1983). An interpretation of the magnetic anomaly map of Nunns et al. (1983) (Fig. 5.10) can show that spreading during anomalies 20-21 is traceable across the Iceland-Faeroe Ridge. The northwestern margin of the Faeroes Plateau consists of a complex zone of anomalies, and anomalies 22-24 are not easily traced across the ridge. These anomalies apparently curve around the complex zone of anomalies. Anomalies 21 and 20 are more easily traced across the ridge from the south, appear less curved on the southern flank of the ridge, and also appear to overlap the earlier anomalies. Anomalies 21 and 20 show no change in orientation or trend in the Norwegian Sea, although spreading clearly slowed after this chron 20.

This event does not show up in the poles of rotation for the Norwegian Sea, although spreading across the Iceland-Faeroe Ridge may have been too small to show up in these data. However, Srivastava & Tapscott (1986) indicate a slight eastward change in the pole of rotation between anomalies 21 and 24 in the Norwegian Sea.

With regards to the Faeroe Block, of probably continental origin (Casten, 1973,1975; Bott, 1984), this represents a shift of the axis of seafloor spreading - for the ocean basin south of the ridge - from the southern to northern edge of the block. It is proposed that this produced an anticlockwise rotation of the Faeroe Block, with
Figure 5.10 Model for the deformation observed during the middle Middle Eocene to ?Lower Oligocene in the southern Faeroe Basin.

a) Magnetic anomaly over the Iceland-Faeroe Ridge. b) Interpretation of events on the Ridge from chron 20 to chron 24. Anomalies divided into normal and reversed pair for each chron (prefixed C).
dextral shear on the structural feature underlying Wyville-Thomson Ridge and Faeroe Bank. Such movement is not restricted to a continental regime and shear along oceanic transforms has been documented (Parson & Searle, 1986). The timing of this event is probably related to the relaxation of the NW European ambient stress field as a result of the Alpine rifting. The presence of strike-slip movements in the southern Faeroe Basin are not therefore directly controlled by the Alpine rifting, but arise from the consequential shift in locus of the NE Atlantic spreading ridge system.

The subsequent change to sinistral transpressional motion inferred for the southern margin of Wyville-Thomson Ridge is undated. It probably represents a phase of middle Oligocene tectonism in the region of the Orkney-Faeroe Alignment (Earle et al., in press). The locus of compression appears to have changed from the Faeroe Basin to the northern Rockall Trough. As with the timing the cause for this is speculative. It may relate to:-

1) Oligocene inversion events associated with the Pyrenean orogeny or Alpine compression which have been invoked for compressional structures in the North Sea (Pegrum & Ljones; Biddle & Rudolph, 1988). Late Eocene to Oligocene folds are also observed on the Biscay margin, related to Pyrenean compression (Masson & Parson, 1983).

2) Transpression in the Voring Basin, and less so in the Møre Basin have been assigned an Oligocene age (Bukovics et al., 1984). Sinistral movements on the Nordland Ridge occurred in the Middle Oligocene to Miocene (Bøen et al., 1984). These appear to be related to the rearrangement of the sea-floor spreading axes in the Norwegian-Greenland Sea, which resulted in simultaneous spreading east and west of the Jan Mayen Ridge. Termination of the spreading in the Norwegian Sea and on the Iceland-Faeroe Ridge occurred in the late Oligocene (Taiwani & Eldholm, 1977; Nunnis, 1983; Vink, 1984). This phase appears to diminish towards the south and is considered to be a local deformation on the Norwegian margin rather than a regional Atlantic event.
Tertiary Sinistral movements along Wyville-Thomson Ridge are therefore likely to be related to a regional orogenic event in the Oligocene not - as for the first and more important movement - a more local compression.

5.6.3 Oceanic influences on sequence E4
As indicated in section 3.6.5, the model for E4 sedimentation showed very different current directions than preceding sequences. Sequence E4 was dated as late Eocene to early Oligocene. This correlates to the cessation of strike-slip along the Hornsund Fault Zone in Svalbard (Spencer et al., 1984), and the separation of Greenland from Norway (Taiwani & Eldholm, 1977) during Chron 13 (Harland et al., 1982). The effect of this would have been to initiate circulation of Arctic cold waters into the Norwegian-Greenland Sea, and temperatures are observed to drop sharply (Buchardt, 1978). This may have stimulated southward ocean currents entering via the recently formed seaway. Since the Iceland-Faeroe Ridge was acting a high (Berggren & Schnitker, 1983), and the East Shetland Basin probably represented a bathymetric step from the Norwegian Sea, the ocean currents would have been diverted towards the west. The current directions in the East Shetland Basin may result from eddy currents produced by this change in ocean current direction (Fig. 5.11). Bottom waters may have flowed though the Faeroe Basin at this time.

5.7 Summary

The discussion above has highlighted a number of the tectonic influences on the post-extensional phase of the Atlantic marginal basins. The magnitude of these events are related to the type of activity occurring with the NE Atlantic rift zone and subsequent ridge. These can be summarised as follows:

1) Middle to Upper Cretaceous. The cessation of active faulting in the basins was followed by rapid thermal subsidence, and possible minor seafloor spreading in the areas that had undergone most extension (Rockall Trough and possibly Faeroe Basin). Basins were
Fig. 5.11 Seismic sequence E4 deposition and changes in the ocean water circulation in the Norwegian–Greenland Sea at the end of the Eocene.
filled by erosion of unextended platform areas such as the Shetland Platform.

(i) Upper Cretaceous to uppermost Palaeocene-lowermost Eocene. Igneous centres formed in a number of basins that extruded basalts and rhyolites (Ridd, 1983), in response to rifting and uplift to the north-west of the Atlantic marginal basins. The volcanic activity increased into the Upper Palaeocene and resulted in the formation of areas of thick flood basalts that permeated the pre-existing crustal weaknesses. Stresses set up in the basins reactivated major crustal structures, and phases of uplift of the Shetland Platform may relate to fluctuations in the influence of the mantle plume under the Iceland Hotspot. The extrusive activity ended abruptly, culminating in the extrusion of the basalts of the ocean-dipping reflector series and a burst of tuffaceous activity correlatable across the basins. Seafloor spreading was initiated as the compressive stresses in the Alpine Foreland momentarily decreased.

(ii) Eocene-Oligocene. A transgression contemporaneous to the last phase of pyroclastic activity may have resulted from the migration of mantle material towards the Reykjanes and Norwegian Sea Ridges. Seafloor spreading was established soon after, and the northern segments of the Viking Graben and Faeroe Basins subsided with the Møre Basin. An uplift of the Shetland Platform and the southern Faeroe Basin resulted from the residual influences of the Iceland Hotspot, and the last outpouring of basaltic lavas on the Faeroes Plateau. Subsequent controls on sedimentation were dominated by global sea level changes. Movements along the structure underlying Wyville-Thomson Ridge resulted in minor inversion of the southern Faeroe Basin. This movement was the result of shifting rates of spreading and small changes in the poles of rotation of individual continental fragments along the margin of the Atlantic Ocean. These caused changes in the ambient stress field across the continental margin, set up by Alpine orogenic and Atlantic spreading effects. Hence, Eocene tectonics was controlled by Atlantic induced stresses. These may have influenced the deposition of the seismic sequences.
especially E2 and E3 (Middle Eocene). Later Oligocene movements may have been the result of the more widespread Pyrenean orogenic event influencing this ambient stress field.

The overall conclusion is that the post-rift development of the western margin of the Atlantic was controlled by the state of stress of the whole of NW Europe. This was not controlled purely by Alpine compressional phases, but was the result of the interaction of both Alpine and Atlantic induced stresses. Deformation was dependent on the proximity of the basins to the causes of those stresses (Fig. 5.12). Hence the thermal subsidence phase of the evolution of the Atlantic marginal basins was influenced by fluctuations in the ambient stress field resulting from Upper Cretaceous to lowermost Eocene rifting, Alpine tectonics and and subtle changes in the spreading history of the Atlantic. These had an influence on individual elements of the pre-existing fragmented and heterogeneous continental crust adjacent to the ocean basins and ridges.
Figure 5.12 Correlation between compressive stresses induced by the Alpine orogenic events and the shifts in Atlantic spreading centres. The diagram shows that the Middle to Late Eocene deformation in the southern Faeroes-Shetland Basin does not easily fit into the recognised tectonic events.
Chapter 6
Theoretical modelling of basin subsidence

6.1 Introduction

This chapter represents a change from the geological approach to post-rift development of the Faeroe Basin, East Shetland Basin and the East Shetland Platform. A more theoretical approach, albeit encompassing geological data, is applied to the basins in an attempt to show how much of the post-rift subsidence pattern was inherited from the earlier rift phase of the basins. The aim of this chapter is to present a number of techniques used to analyse the influence that pre-existing basin structure and the configuration of extended lithosphere had upon the subsidence of the East Shetland Platform.

Most of the modern theory of extensional basin formation stems from the two-phase evolutionary model of McKenzie (1978) (Fig. 6.1). This essentially consisted of a extension-related subsidence, that thinned the lithosphere. A subsequent unfaulted subsidence phase occurs as the extended lithosphere cooled to its equilibrium thickness. Because of lateral heat flow in the extended lithosphere, the flanks of the basin are progressively uplifted, resulting in the erosion - and consequent crustal thinning - of unextended crust. The subsidence of the eroded flanks produces a classic "Steer's Head" basin (Dewey, 1982). The flexural rigidity of the crust also plays a major part in the development of this type of basin (Watts et al., 1983), by accommodating the load of the basin over a wider, down-flexed area than by Airy isostasy. A two-dimensional model can be applied to the East Shetland Basin and East Shetland Platform, which appear to be of this form. The East Shetland Platform is intrinsically involved in the evolution of the East Shetland Basin, and consequently, an analysis of the mechanism of basin formation provides an insight into the role of the platform.
Figure 6.1 McKenzie (1978) type horizontal stretching and its effect on a lithospheric section:
(a) isostatically balanced lithosphere pre-stretching,
(b) immediately after stretching by a factor B showing initial fault-controlled subsidence,
(c) lithosphere has almost re-equilibrated and thermal subsidence is added to initial subsidence.
(d) Predicted subsidence curves for water-filled basin formed by extension, (e) sedimentary basin formed by stretching. Basin flank onlap of Dewey's (1982) classic steer's head basin may be caused by lithospheric flexure, lateral heat flow or heterogeneous stretching (see text for details).
Diagram based on Chadwick (1985a).
(f) Cross-section of the northern North Sea, showing the onlap of Tertiary sediments onto the basin flanks.
This steer's head shape of the basin and platform has been shown in Chapters 2 & 3, to have controlled the position of the marginal facies in the Middle Eocene. Additionally, this region of broad subsidence separated the marginal deposition from the basinal sedimentation, and equivalent deltaic sediments are not observed in the basin north of 59°N. The flank uplift and erosion model has been applied to the East Shetland Platform by Leeder (1983). However, little is known as to why some basins, such as the Faeroe Basin do not show this shape. Sediments in the Faeroe Basin do not onlap the West Shetland Platform (Fig. 6.2), and whilst erosion of the onlap may have occurred as a result of inversion (Chapter 4), the West Shetland Platform does not appear to have undergone any significant subsidence.

It is therefore considered important to make a more detailed analysis of some of the factors that influence subsidence during the Palaeogene. This analysis includes reference to the earlier history of the basins, based on documentary evidence and the sequences observed within the basins. The fundamental data for this chapter is subsidence curves obtained from the released well data in both the East Shetland Basin (Fig. 6.3) and the Faeroe Basin (Fig. 6.4). The curves have been modelled using some of the theory that has been applied to extensional basin development.

6.2 Calculation of subsidence curves from well data

The basic technique for obtaining a subsidence curve from the stratigraphy of a well follows the approach of Steckler & Watts (1978) and Sclater & Christie (1980). The sediments overlying basement are progressively removed, until basement reaches present datum, and a depth to basement versus age relationship can be established. This is a curve showing the subsidence history of the well. The basic steps involve:

1) Decompaction - the progressive removal of the effect of sediment compaction from the well.
Figure 6.2 The Shetland Platform, showing the subcrop limit of the Tertiary. Note how the Tertiary is deposited over a wide area of the East Shetland Platform, but is restricted to a narrow zone on the West Shetland Platform. Diagonal shading - Tertiary absent; criss-cross - Permo-Triassic basins exposed on the Shetland Platform.
Figure 6.3 Location and identification of wells in the East Shetland Basin for which the sediments were backstripped and the tectonic subsidence curves were computed.
Figure 6.4 Location of wells in the Faeroe Basin, for which backstripping was applied.
ii) Sediment unloading - removal of the loading effect of the sediments overlying basement.

iii) Water depth correction - increases the observed subsidence by accounting for the fact that sedimentation does not take place at a fixed datum, sea level.

iv) Correction for sea level changes - The fluctuations in water depth may result from rapid tectonic subsidence, outstripping the rate of sediment accumulation. Alternatively, they may also result from temporal variations in sea level datum in response to global, eustatic changes, and can be corrected for the latter.

These steps, and the parameters used in this study are discussed below.

6.2.1 Sediment decompression

As sediments are buried pore fluids are expelled and the volume, and consequently thickness, of a particular stratigraphic unit decreases. Because of compaction, the present thickness of a sediment sequence in a well is not the true thickness that was deposited. Hence, it is not a direct measure of tectonic subsidence (cf. Clarke, 1973). This initial thickness can be found using a porosity-depth relationship and progressively bringing the succession up to surface.

The porosity-depth relationship of Sclater & Christie (1980) is an exponential, of the form:

\[ \phi = \phi_0 e^{-cz} \]

\( \phi \) - porosity at depth \( z \)  
\( \phi_0 \) - initial porosity at surface  
\( c \) - porosity constant

The parameters for the porosity-depth relationship for the North Sea have been determined by Sclater & Christie (1980), based on sonic log porosity data. Different relationships were similarly obtained for the different lithologies; sand, shale, chalk and shaley sand. Subsequent analysis of neutron and sonic log data, in the northern North Sea, was carried out by Zervos (1986). This later work is
therefore more applicable to the northern North Sea lithologies. The use of two logs is a more rigorous approach in that sonic logs alone have an optimum porosity range of 5-20% (Merkel, 1981), and do not measure fracture or vugular porosity. This dual approach was also based on a lot more well data (28) than for Sclater & Christie (1980) (based on 8 wells). The values used for this work have been taken from Zervos (1986), although these only are only slightly different to those of Sclater & Christie (1980) (Table 6.1).

Based on this approach the thickness of a unit varies during decompaction in the form of:

\[ Z_2' = (Z_2 - Z_1) - \left( e^{-z_2} - e^{-z_2'} \right) \phi_0 / c + \left( 1 - e^{-z_2} \right) \phi_0 / c \]
\[ Z_3' = Z_2' + Z_3 - \left( e^{-z_2} - e^{-z_2'} \right) \phi_0 / c + \left( e^{-z_2} - e^{-z_3} \right) \phi_0 / c \]

\( Z_1, Z_2, Z_3 \) - Original depth to tops of units
\( Z_2', Z_3' \) - Decompacted depth to tops \( (Z_1' = 0) \)

(After Sclater & Christie, 1980)

**TABLE 6.1 Porosity-depth parameters**

<table>
<thead>
<tr>
<th></th>
<th>( \phi_0 )</th>
<th>( c \left( \times 10^{-5} \text{ cm}^{-1} \right) )</th>
<th>( \rho_w' )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shale</td>
<td>0.63</td>
<td>0.51 (0.55)</td>
<td>2.721</td>
</tr>
<tr>
<td>Sand</td>
<td>0.49</td>
<td>0.27 (0.24)</td>
<td>2.651</td>
</tr>
<tr>
<td>Shaley sand</td>
<td>0.56</td>
<td>0.39 (0.42)</td>
<td>2.681</td>
</tr>
<tr>
<td>Chalk</td>
<td>0.70</td>
<td>0.71 (0.71)</td>
<td>2.711</td>
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</tbody>
</table>

Most of the lithologies, taken from well descriptions, in the northern North Sea fall into these broad categories. Composite units, comprising alternating sands and shales were defined as a shaley sand unit.

229
To produce a curve, the sediment column overlying basement in the well is progressively stripped. Stratigraphic time markers divide the well into a number of units (Fig. 6.5).

6.2.2 Triassic sediments and basement

Compacted sediments are presumed to lie on a completely compacted basement. Although a number of marginal or platform wells penetrate crystalline basement or Devonian sediments, most of the basin wells in the northern North Sea terminate in Permo-Triassic sediments. Some of these, such as in the Magnus Basin, have a major unconformity between the Trias and the overlying Cretaceous sediments. It is not possible to ignore the effect of the Triassic sediments underlying a particular well. If the Triassic has compacted during the subsequent history of the basin, there is an inherent overestimate of the tectonic input into the curve (Fig. 6.6). Indeed, the Triassic succession produces probably the greatest uncertainty into any subsidence work in the northern North Sea, including decompaction of seismic successions (Beach et al., 1987; Giltner, 1987; Shorey & Sclater, in press). Variations in thickness in the East Shetland Basin are poorly documented, varying from 1000 to 3000 m across the East Shetland Basin into the Viking Graben, and controlled by major faults (Ziegler, 1982). Dating of Triassic events is also poorly defined, due to the lack of any useful biostratigraphic markers (Fisher, 1984; Morton et al., 1987). It has, however, been widely suggested that the sediments are mostly late Triassic (Ziegler, 1982; Beach et al., 1987). In the light of the uncertainties, any considerations based on the Triassic succession in the northern North Sea are somewhat conjectural.

An estimate of the thickness of Triassic sediments below the well are obtained from an isopach map (Fig. 6.7), (Ziegler, 1982), or from seismic lines across the well. The lithology of the sequence is presumed to be a mixture of shales and sands, and are thus assigned the porosity-depth characteristics of shaley sand. The Triassic sequence is then included in the decompaction procedure.
Figure 6.5 Backstripping - an example based on well 2/15-1. Units are defined by chronostratigraphical markers. Note how the sediment thicknesses increase as the basement is brought to surface. This is the result of progressive decompaction, in the manner of Sclater & Christie (1980).
Figure 6.6 Comparison between the subsidence curves obtained assuming full compaction of the Triassic sequence before deposition of the Jurassic, and taking Triassic compaction into account. Note how the shape of the curve remains the same, but the overall amount of subsidence, ascribed to tectonic processes, increase markedly.
Figure 6.7 Isopach map for the Triassic sequence, and from which the thickness of the compacted Triassic at the base of each well was computed. Locations of wells are indicated, and correspond to locations shown in Fig. 6.3.
6.2.3 Progressive sediment unloading

The wells are backstripped by removing the top layer and allowing the underlying units to decompact along the specified porosity-depth relationship. The depth to basement is then recomputed. The mean sediment density of the remaining sediment column is also re-evaluated by the following:

\[ p_{\text{m}} = \frac{1}{S} \left( \frac{p_{\text{i}}}{\delta_{\text{i}}} + (1 - \delta_{\text{i}})p_{\text{g,i}} \right) S' / S \]

- \( \delta_{\text{i}} \) - mean porosity of \( \text{ith} \) layer
- \( p_{\text{g,i}} \) - sediment grain density of \( \text{ith} \) layer
- \( S' \) - recomputed depth to top of \( \text{ith} \) unit
- \( S \) - recomputed total sediment thickness

The re-evaluated thickness for the sediment column (less the backstripped units) is then unloaded assuming Airy isostasy:

\[ Y = \frac{S (\rho_{\text{m}} - \rho_{\text{w}})}{(\rho_{\text{m}} - \rho_{\text{w}})} \]

- \( Y \) - water loaded depth to basement
- \( \rho_{\text{m}} \) - mantle density 3.33 g cm\(^{-3}\)
- \( \rho_{\text{w}} \) - water density 1.03 g cm\(^{-3}\)

6.2.4 Water depth correction

Palaeobathymetry is an important correction to the depth of unloaded basement derived above. Two indicators to palaeobathymetry are used. The sedimentary facies indicators (e.g., red beds, glauconite, cross-bedding) help to make coarse distinctions between, for instance, alluvial, coastal plain and deeper marine palaeoenvironments. Facies maps therefore enable some detailing of the position of, for instance, coastal plains in the Jurassic (Eynon, 1981) or the late Palaeocene shelf break (defined by seismic interpretation in the manner of Rochow, 1981). By far the most useful indicator is micropalaeontological assemblage data. Such an analysis has been carried out in the central North Sea (Barton & Wood, 1984). It should be noted that the use of present day faunal distributions as a guide to the past is tenuous. Temporal changes in oceanographic and climatic conditions mean that distributions of particular fauna were probably not the same in the past as they are at present.
combination of these approaches, however, enables a broad palaeobathymetric scheme to be produced.

Uncertainties abound in the strict determination of the water depth, especially in abyssal environments. By far the greatest uncertainty in the northern North Sea occurs in the Upper Jurassic and Cretaceous. A significant basin relief may have occurred in the Upper Jurassic, with the deposition of the Magnus submarine fan sandstones (De'ath & Schuylenberg, 1981). At the same time, marginal sands occur in the Kimmeridgian observed along the edge of the East Shetland Platform (Wheatley et al., 1987). The Upper Jurassic Kimmeridge clays have been assigned a broad depth range (2-500m) throughout the East Shetland Basin. Similarly, there is evidence of a Lower Cretaceous 'marginal' limestone facies at the edge of the East Shetland Platform (Amiri-Garroussi, 1988). This probably indicates the margin of the basin (Wheatley et al., 1987). A shelf paleobathymetry has been assigned to all the wells located on the Shetland Platform. The only estimate of the depth in the centre of the graben is 300m (Wood, 1982), although this may be in error. A depth range of 200-400m has been assigned for the Cretaceous palaeobathymetry in the East Shetland Basin (Table 6.2).

Similar uncertainties occur in the Palaeogene, where agglutinated foraminifera are the basis for biostratigraphic sub-division (Berggren & Gradstein, 1981). These species have a very broad depth range, existing from 800-1500m (Wood, 1982), although the domain of agglutinated foraminifera in a restricted basin may well have been from 200m to an unknown depth (Berggren & Gradstein, 1981).

Some limitations on the maximum water depth for specific horizons can be made in the Palaeogene, using the backstripping approach (Fig. 6.8). The basin relief can be reconstructed if a number of wells are decompacted back to a common horizon - for instance, the top Palaeocene - and one well is known to be of shelf origin. If the difference in depth to a recognisable marker horizon (in this case basement or the Top Triassic) between wells is considered to be
Figure 6.8 Example of the determination of the maximum water depth estimate for a well in the eastern edge of the East Shetland Basin. Use of a control well and the decompacted depths to and thickness at a particular common horizon enable an estimate to be made.
unchanged since the last phase of rifting, the maximum water depth for an unknown well can be determined. The assumption for basement ignores the effect of post-Palaeocene differential subsidence between wells. This effect is likely to make the water depth estimate a maximum, since the control well in this case is located on the East Shetland Platform, and shows a shallow palaeobathymetry. The estimates fit with the palaeoenvironmental interpretation for the top Palaeocene, and most wells to the east of the palaeo-shelf edge (Rochow, 1981) show a maximum water depth of 0-300m (Table 6.2). Higher stratigraphic horizons are therefore also likely to be in accordance (Table 6.2).

This approach is very specific to the North Sea, and there only to this post-faulting sequence. Were it to be applied to other basins, the assumptions would have to be critically analysed beforehand.

Most of the wells in the Faeroe Basin are located on the Rona Ridge, and within the West Shetland Basin. Palaeobathymetry of these wells has been taken from published material. On this evidence, the Rona Ridge has acted as a structural high throughout most of the basin evolution - most importantly during the Jurassic (Ridd, 1981; Hitchen & Ritchie, 1987). Subsequent Cretaceous transgression events can be ascribed to similar events in the North Sea (Ridd, 1981), although Lower Cretaceous conglomerates suggest that the Rona Ridge still maintained some influence on sedimentation (Hitchen & Ritchie, 1987). Submarine fan sedimentation in the West Shetland Basin, during the Palaeocene was derived from Rona Ridge and the Shetland Platform (Hitchen & Ritchie, 1987). Shallowing of the basin during the Palaeocene, is recorded by Mudge & Rashid (1987). These results are summarised in Table 6.3, and for specific wells in Table 6.2.
**Table 6.2** List of values for the water depth correction given to individual subsidence curves based on biostratigraphic and geologic data for the basins east and west of the Shetland Platform

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<tr>
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<th>Cretaceous</th>
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<td>204/28-1</td>
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<td>2-400</td>
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<td>30-200</td>
<td>2-400</td>
<td>2-400</td>
<td>0-200</td>
</tr>
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U: Stratigraphical marker undefined
- Sequence absent

238
Table 6.3 Summary of the palaeoenvironments of the main structural features west of the Shetland Platform.

<table>
<thead>
<tr>
<th>Period</th>
<th>Faeroe-Shetland</th>
<th>Rona Ridge</th>
<th>West Shetland</th>
</tr>
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<tbody>
<tr>
<td>Present</td>
<td>200-1600m</td>
<td>150-200m</td>
<td>50-200m</td>
</tr>
<tr>
<td>Recent-Eocene</td>
<td>Bathyal</td>
<td>Inner-Outer</td>
<td>Inner Shelf</td>
</tr>
<tr>
<td>Palaeocene</td>
<td>Probably</td>
<td>Inner-Outer</td>
<td>Inner Shelf</td>
</tr>
<tr>
<td>Upper Cretaceous</td>
<td>Bathyal</td>
<td>Outer Shelf</td>
<td>Outer Shelf</td>
</tr>
<tr>
<td>Lower Cretaceous</td>
<td>Bathyal</td>
<td>North-exposed</td>
<td>Inner-Outer</td>
</tr>
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<td>Upper Jurassic</td>
<td>Bathyal</td>
<td>Outer Shelf</td>
<td>Inner-Outer</td>
</tr>
<tr>
<td>Middle Jurassic</td>
<td>Bathyal</td>
<td>Outer Shelf</td>
<td>Inner-Outer</td>
</tr>
<tr>
<td>Permo-Triassic</td>
<td>CONTINENTAL</td>
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</table>

Water depths (van Hinte & Deighton, 1987)

<table>
<thead>
<tr>
<th></th>
<th>Continental</th>
<th>Sub-littoral</th>
<th>Shelf</th>
<th>Bathyal</th>
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<tr>
<td></td>
<td>Inner</td>
<td>Outer</td>
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<td></td>
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<tr>
<td>&gt;0m</td>
<td>0-30m</td>
<td>30-100m</td>
<td>100-200m</td>
<td>200-400m</td>
</tr>
</tbody>
</table>

6.2.5 Sea Level Correction

Changes in water depth can result from phases of rapid subsidence or uplift, changes in the sedimentation rate or simply from changes in the sea level datum. The latter were initially recognised by Pitman (1978) as being related to volume changes in mid-ocean ridge systems. A correction must be applied to account for these eustatic changes in sea level throughout the burial history, in order to obtain a true tectonic subsidence curve.

239
Current ideas on eustatic sea level changes are based on the Vail et al. (1977b) curve. This was derived from inter-regional stratigraphic studies. The first and second order curves are generally accepted as of global extent (Falvey & Deighton, 1982; van Hinte & Deighton, 1987), albeit with some modification (Hallam, 1984). Higher frequency changes in this curve may be contaminated by regional tectonic events, and so are discounted in this analysis (Section 1.5.2). Calibration is based on the amplitude of the late Cretaceous sea level high (Pitman, 1978), for which there are varied estimates. Estimates range from 220m (Falvey & Deighton, 1982) and 250m (Haq et al., 1987) to 375m (Sleep, 1976) and 400m (Hallam, 1984) (Fig. 6.9). A lower value of the highstand amplitude appeared to fit the subsidence better in that it did not produce a net uplift in the curves during the late Cretaceous - an event for which there is no lithological evidence. A first and second order Vail curve was therefore applied, with an end Campanian highstand of 200m above present day sea level (van Hinte & Deighton, 1987).

Wood (1982) presents reservations about the use of the sea level correction on curves where the water depth uncertainty outweighs the sea level change. However, it is still believed to be good practice to remove some of the effect of sea level for the parts of the curve where water depth is less than 200m. The shape of the whole curve is also important for the subsequent modelling, and is dependent on the shape of the sea level curve as much as the amplitude of the sea level fluctuations.

Application of the sea level correction, and the positioning of stratigraphical markers on the subsidence curve, has been made using the chronostratigraphic timescale of Harland et al. (1982).

6.2.6 A fully corrected subsidence curve

The fully corrected tectonic subsidence, decompacted to a specific horizon, is therefore:

\[ Y = S(\rho_m - \rho_w)/(\rho_m - \rho_w) - WD - \Delta SL(\rho_m/\rho_m - \rho_w) \]
Figure 6.9 Comparison of the two most up to date sea level curves. Both are basically a first and second order Vail et al. (1977) curve with minor modifications. Calibration, however, is quite different.
The values obtained are plotted against time in Ma. Stratigraphic markers have been calibrated to time using the Harland et al. (1982) time scale. Whilst there may be errors and uncertainties in the biostratigraphic dating and the radiometric ages used for calibration, these are not the largest uncertainty in the curve and have therefore been ignored. Unconformities have been plotted as periods of no sediment deposition on curves uncorrected for water depth and sea level changes. On fully corrected curves the missing sections only fix the subsidence curve at the end of the section, since erosion may have occurred during the unconformity. Unless otherwise shown by biostratigraphical evidence, the age at the beginning and end of the unconformity is taken as the earliest and latest ages of the underlying and overlying chronostratigraphic units, respectively. The resultant shape of the fully corrected curve with time is a reflection of tectonic subsidence of the basement (Fig. 6.10).

6.3 Subsidence mechanisms - a review

Basin subsidence in a horizontal extensional regime, can result from a number of different mechanisms. For the purpose of this chapter, these are defined as thermal mechanisms, in that subsidence is a response to the isostatic and thermal consequences of lithospheric extension, both of which can be readily modelled (Fig. 6.11):-

1) Lithospheric extension by pure shear (McKenzie, 1978). This is characterised by two interdependent phases of subsidence. An initial phase of subsidence is caused by the thinning of the lithosphere by extension, and the isostatic equilibration. The second phase of subsequent subsidence is related to the cooling of the thinned lithosphere, and follows an exponential trend.

2) Depth-dependent pure shear of the lithosphere (Rowley & Sahagian, 1986; White & McKenzie, 1988). The rheological properties of the lithosphere (Royden & Keen, 1980), and the aseismicity of the lower crust (Chen & Molnar, 1983) suggest a change from brittle deformation in the upper crust to ductile deformation in the lower crust and upper mantle. Since brittle deformation is localised by
Figure 6.10 An example of subsidence curves produced for the well referred to in Fig. 6.5. Each curve shows the changes produced by the different corrections applied. Note how the Hallam (1984) sea level curve overcompensates for the late Cretaceous highstand.
Figure 6.11 The different mechanism by which the lithosphere may undergo extensional deformation. Differences in style of deformation are associated with different considerations of the rheology of the lithosphere.
discrete faults and lineations, the lateral extent of these two types of deformation may vary. The ratio of crustal to sub-crustal extension, at any specific location, critically controls the amount of subsidence in both the extensional and post-extensional phases of basin formation.

iii) Lithospheric extension by simple shear (Wernicke, 1985). The two phases described in i) are spatially offset by a lithospheric shear plane. Fault-bounded subsidence occurs away from the area of sub-crustal thinning. In this setting, extensional and post-extensional subsidence may be independent at the same location.

All of these mechanisms are passive (Bally, 1980) because the lithosphere deforms without any change in the physical and chemical composition of the mantle. Active subsidence mechanisms result in physico-chemical changes in the mantle rheology. In particular, subsidence is a response to crustal creep (Bott, 1980), sub-crustal phase changes (Falvey & Middleton, 1981) and the influence of asthenospheric convection (Steckler, 1986; Moretti & Chenet, 1987). Most of these mechanisms, however, are only applicable to regions of anomalous mantle conditions and where seafloor spreading has, or almost, occurred.

6.3.1 Simple Shear basin formation - application to the North Sea

Extension resulting from simple shear is caused when the lithosphere extends along a dipping fault plane or planes. Consequently, the deformation at the base of the crust or lithosphere may be offset from the surface expression of extension. Theoretical modelling of this offset has assumed that subsidence is controlled by separate crustal and sub-crustal extension factors (Royden & Keen, 1980), defined here as $\beta_c$ and $\beta_{sc}$ respectively. Similarly, the crustal extension may consist of both a brittle and ductile extension factor (Hellinger & Sclater, 1983). This approach has been used to show the isostatic response to independent crustal or sub-crustal thinning (Coward, 1986). However, Kusznir et al. (1987) indicate that this approach is merely a pure shear approximation to simple shear, and propose a model encompassing upper crust deforming by simple shear,
and the lower by ductile pure shear (Fig. 6.11). The two regions are separated by a low-angle detachment. This new approach predicts an asymmetric basin, but where the second phase, thermal decay basin is not laterally offset. Depth of detachment to a lower crustal region of pure shear is an important control on the width and asymmetry of the basin. In this case, detachment at the base of the crust results in a symmetric basin. Where extension is based on a dipping fault, the importance of isostatic compensation must also be considered. Unloading of the lithosphere by lateral movement of the crust above the controlling fault, results in uplift of the hanging wall.

The positioning of the thermal basin directly over the rift basin, in the northern North Sea, implies that whole lithosphere simple shear sensu stricto (Wernicke, 1985) does not apply. However, a simple shear model does not need to be rejected. The East Shetland Basin, and Viking Graben, represent a much more complex situation than any of the theoretical models. Interference of a number of different detachment horizons may have led to the development of a basin that regionally conforms to a pure shear system.

6.4 Modelling of the subsidence curves

Using the mathematical basis for extensional tectonics, the resultant subsidence curve can be modelled. A modelling program has been written that uses the following subsidence mechanisms to produce theoretical subsidence curves:

1) Instantaneous homogeneous rifting (McKenzie, 1978)
2) Finite rifting (Jarvis & McKenzie, 1980)
3) Inhomogeneous extension based on detachment of the lithosphere at the Moho (Hellinger & Sclater, 1983)

Lateral heat flow has also been included in the calculated theoretical subsidence curves. This has greatest effect on regions flanking the theoretical basin, in that heat flow to the flanks causes uplift (Cochran, 1983; Watts et al., 1982).
8.4.1 Basin subsidence - a finite difference approach
Subsidence can be modelled in terms of the isostatic response to changes in crustal thickness and the geothermal gradient within continental lithosphere, resulting from lithospheric extension. Thus, analysis of heat flow from the lithosphere can be used to construct a theoretical subsidence history. To allow for all of these parameters a finite difference model was created (Smith, 1978). This followed the principles of Long & Lowell (1982), in solving the heat flow equation, although to accommodate the lateral heat flow, a two-dimensional equation was solved:

\[ \frac{\delta T}{\delta t} = K \left[ \frac{\delta^2 T}{\delta x^2} + \frac{\delta^2 T}{\delta y^2} \right] \]

K - thermal diffusivity

On a grid this is rewritten:

\[ \frac{\delta T}{\delta t} = K \left\{ \left( T_{i-1,j+1} + T_{i,j-1} + T_{i+1,j+1} + T_{i,j-1} \right) / h_x^2 - \frac{2(h_x^2 + h_y^2)T_{i,j}}{h_x^2 h_y^2} \right\} \]

Ti,j - temperature at point i,j

hx,hz - vertical and horizontal distance between points on grid

\[ T_{i-1,j-1}, \quad T_{i+1,j-1}, \quad T_{i,j-1}, \quad T_{i+1,j+1}, \quad T_{i-1,j+1}, \quad T_{i,j+1}, \quad T_{i,j} \]

This is determined in finite difference form using the Peaceman-Racheford stepwise computation (Smith, 1978; p. 39). The first step (n+1) solves the equation in the horizontal plane from temperature values at time step n, the second (n+2) in the vertical plane from temperatures from the preceding time (n+1). This is purely to cut down the processing time in the computation. As long as the number of alternate steps is kept equal and the time step remains constant, no instabilities develop in the computation.

Step 1 \( \left( 1/K \delta t + 2/\delta x^2 \right) T_{i,j,n+1} - (1/\delta x^2) (T_{i-1,j,n+1} + T_{i+1,j,n+1}) \)
\[ = (1/K \delta t - 2/\delta x^2) T_{i,j,n} + (1/\delta x^2) (T_{i-1,j,n} + T_{i+1,j,n}) \]

Step 2 \( (1/K \delta t + 2/\delta y^2) T_{i,j,n+2} - (1/\delta y^2) (T_{i,j+1,n+2} + T_{i,j-1,n+2}) \)
\[ = (1/K \delta t - 2/\delta y^2) T_{i,j,n+1} + (1/\delta y^2) (T_{i-1,j,n+1} + T_{i+1,j,n+1}) \]
These steps can be computed by means of a Gaussian elimination of a tridiagonal matrix (Appendix C).

6.4.2 Initial parameters

The model consists of a rectangular lithosphere defined by a grid of points, each of which has the properties of position, temperature and a thermal conductivity (Fig. 6.12). These are primed to give a simple initial model for the crust and sub-crustal lithosphere. Geothermal gradient is assumed to be linear, from 0°-1333°C (Parsons & Sclater, 1977) from surface to base of lithosphere. Heat flow is not allowed out of the sides of the model, or downwards. Hence, the left and right edges of the model contain 'buffer zones', where the temperature structure is kept constant. The lower face, equivalent to the top of the asthenosphere is defined by another buffer zone of constant temperature, 1333°C. This allows heat to flow upwards out of the model only. Note that this still allows heat to flow vertically and laterally within the defined model. Initial crustal and lithospheric thicknesses may be changed within the program but are generally taken as 30 km and 125 km respectively (Parsons & Sclater, 1977). Diffusivity is defined by:

\[ K = \frac{k}{\rho c} = 19.044 \text{ km}^2\text{Ma}^{-1} \]

\( k \) - thermal conductivity of the crust 2.5 Wm\(^{-1}\)K\(^{-1}\)
\( \rho \) - density of lithosphere as a whole 3330 kgm\(^{-3}\)
\( c \) - specific heat of upper mantle 1.250 Kjkg\(^{-1}\)K\(^{-1}\)

Bott (1982)

Diffusivity contains continental and mantle thermal properties and has been chosen so that the time constant of the lithosphere (\( \tau \)) is approximately 62 Ma (\( \tau = a^2/\pi^2K \); McKenzie, 1978).

Geothermal gradient is taken as linear, of the form:-

\[ T_z = T_m z/a \]

where \( z \) is depth taken from datum, rather than the base of the lithosphere as in the case of McKenzie (1978) and Hellinger & Sclater (1983).
INITIAL CONDITIONS OF FINITE DIFFERENCE MODEL

$T_0 = 1333°C$

Temperature $T_{x_i}z_j = T_0 Z_j B_{x_i}$

Initial Subsidence $S_i = 3.39 (1 - 1/B_{x_i})$

Elevation $E(t) = F(p_m,p_w)\sum(T_{i,j} - T_0 Z_j/a)$

Subsequent Subsidence $S(t) = S_f + \sum \triangle E(t)$

Figure 6.12 Features of a finite difference model for simulating the effects of extension on a homogeneous ideal continental crust.
For the purpose of this study the lithosphere is assumed to have mantle properties, the subcrustal lithosphere representing ~75% of the lithosphere. This diffusivity is therefore used throughout, and is applied uniformly to the lithosphere.

6.4.3 Lithospheric extension

Extension of the lithosphere is applied by defining the extension factor $\beta$. From this the subsidence, assuming instantaneous stretching, is computed by:

$$S = a \left( \rho_m - \rho_c \right) tc/a (1 - \alpha T_m tc/2a) - q T_m \rho_m/2) (1 - 1/6)$$

$$\rho_m \left( 1 - \alpha T_m \right) - \rho_c$$

(McKenzie, 1978)

From the values in Table 6.4 this simplifies to:

$$S = 3.39(1-1/6) \quad t_c = 30 \text{ km}$$

$$= 4.66(1-1/6) \quad t_c = 35 \text{ km}$$

Inhomogeneous extension occurs when crustal and sub-crustal lithosphere deform by different amounts. This may be a response to different rheologies in the crust and sub-crustal lithosphere in which brittle extension is localised by the position of faults and bounding lineaments, and ductile extension is controlled more by creep processes. Such a process requires a detachment between the brittle and ductile layers, which may not necessarily be the Moho.

Table 6.4 Values assigned to parameters in the subsidence formulae and the finite difference model (Parsons & Sclater, 1977)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Definition</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a$</td>
<td>thickness of lithosphere</td>
<td>125 km</td>
</tr>
<tr>
<td>$t_c$</td>
<td>thickness of crust</td>
<td>30 km</td>
</tr>
<tr>
<td>$\rho_m$</td>
<td>density of mantle material at 0°C</td>
<td>3.33 g cm$^{-1}$</td>
</tr>
<tr>
<td>$\rho_c$</td>
<td>density of crustal material at 0°C</td>
<td>2.78 g cm$^{-1}$</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>water density</td>
<td>1.03 g cm$^{-1}$</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>thermal expansion coefficient</td>
<td>$3.28 \times 10^{-6}$ C$^{-1}$</td>
</tr>
<tr>
<td>$T_m$</td>
<td>temperature at base of lithosphere</td>
<td>1333°C</td>
</tr>
<tr>
<td>$\tau$</td>
<td>thermal decay constant of lithosphere</td>
<td>62.8 Ma</td>
</tr>
</tbody>
</table>
The lower crust may itself behave in a ductile manner. However, the amount that the crust subsides in response to inhomogeneous extension is isostatically dependent on the ratio of crust to subcrustal lithosphere, and not necessarily to the position of a detachment between brittle and ductile processes. In this model initial fault subsidence, during inhomogeneous extension, is computed as:

\[ S_c = \left[ (\rho_m - \rho_e) t_c (1-\alpha T_m t_e / 2a) - \alpha \rho_m T_m t_e / 2 \right] \left( 1 - 1 / \beta_c \right) / \left( \rho_m (1-\alpha T_m) - \rho_w \right) - \left[ \alpha \rho_m T_m (a-t_e) / 2 \right] \left( 1 - 1 / \beta_c \right) / \left( \rho_m (1-\alpha T_m) - \rho_w \right) \]

\[ a / \beta_c = t_c / \beta_e + (a-t_e) / \beta_w \]

Substituting for \( \beta_e \), water-loaded subsidence which applies for a comparison to the subsidence curves, this simplifies to:

\[ S_c = 5.557 \left( 1 - 1 / \beta_c \right) - 2.440 \left( 1 - 1 / \beta_w \right) \]

Modelling of the various basin concepts can be attempted by varying the values of \( \beta_c \) and \( \beta_w \). The thermal structure of the lithosphere is redefined, such that the initial linear geothermal gradient becomes:

\[ T_e = \beta_l T_m z / a \quad \text{Homogeneous} \quad 0 < z < a / \beta_l \]

\[ T_x = \beta_c T_m z / a \quad \text{Inhomogeneous} \quad 0 < z < t_c / \beta_c \]

\[ T_x = T_m (1 + \beta_w (z / a + 1 - 1 / \beta_w)) \quad \text{Inhomogeneous} \quad t_c / \beta_c < z < a / \beta_l \]

Deformation of the grid pattern is ignored during extension and points on the grid remain fixed with respect to the top of the lithosphere. Where uplift occurs, as a response to \( (1 - 1 / \beta_w) > 2.283 (1 - 1 / \beta_c) \), erosion is taken as instantaneous. The removal of crust by this method is accounted for by lowering the reference datum at this point. The subsequent cooling of the extended subcrustal lithosphere therefore results in subsidence of the erosionally thinned crust, in the manner of Sleep (1971) and Hellinger & Sclater (1983).

Because a two-dimensional model is used for simulating the effect of lateral heat flow, different crustal and subcrustal \( \beta \) values can be ascribed along the model. Hence the model can be used to simulate the evolution of small basins or passive margins.
6.4.4 Thermal subsidence.
The consequent thermal subsidence phase of basin evolution is modeled using the finite difference heat flow formula. Time steps of 0.5 Ma are used, and the temperature structure recalculated. This is directly related to \( e(t) \), the elevation above the final depth to which the upper surface of the lithosphere sinks (McKenzie, 1978). Hence, the amount of subsidence that has occurred during the time step is computed from the difference in the value of \( e(t) \) at the start and end of the time step. Thus for each step:

\[
\Delta e = e(t)-e(t-1)
\]

At the end of the step, the boundary conditions, as for the initial model are re-established. The model therefore remains an open system, with heat lost through the upper, crustal surface, and input into the lithosphere at its base. Asthenosphere temperature remains constant at 1333°C throughout. Thermal contraction of the grid (Royden & Keen, 1980) throughout this stage is negligible.

Where crust is elevated above datum, instantaneous erosion occurs, producing a subsequent subsidence of platform areas. The consequences of the model are shown in Fig. 6.13.

6.4.5 Finite extension rates
Subsidence associated with lithospheric extension - as outlined in section 6.4.3 - has thus far been considered to be instantaneous. Where extension occurs over a finite interval, heat flow throughout the period of rifting must be taken into consideration (Jarvis & McKenzie, 1980; Cochran, 1983). In general, the approach to finite extension is to consider it as a series of discrete, instantaneous events (Fig. 6.14a). Extensional subsidence is computed by the formulae above, based on the crustal and sub-crustal lithosphere thicknesses resulting from earlier phases of the finite extensional event. Total crustal and subcrustal \( \beta \) values are defined for the event, which are divided into a number of smaller sub-events such that:
Figure 6.13 The consequences of extension of the finite difference model of the lithosphere. Lateral and vertical heat flow occur, and the flanks of basins, and platform regions undergo uplift and erosion.
\[
N_{ext} = \frac{\text{Length of finite extension}}{\text{Time-step}}
\]
\[
\beta_n = (\beta_{\text{initial}})^{1/N}
\]

For time steps of 0.5 Ma, the thermal relaxation of the base of the lithosphere is considered negligible. Consequently, at the end of the rifting period, the lithospheric thicknesses are:
\[
t_e = t_{\text{pre-rifting}}/\beta_n
\]
\[
a = a_{\text{pre-rifting}}/\beta_n
\]

These extensional phases are separated by short-lived thermal phases of subsidence. The mathematical basis behind this approach becomes complicated, especially when lateral heat flow effects are introduced (Cochran, 1983). Since the pre-rift geothermal structure is no longer linear, for extensional events succeeding the initial event, the simple formulae for thermal subsidence no longer apply.

The effect of a subsequent extensional phase is to elevate the lithosphere geotherms, bringing them closer together and closer to the upper crustal surface. This can be simulated in a finite difference model by simply multiplying the temperature value at any one point by the relevant crustal or sub-crustal lithosphere \( \beta \) value. The base of the lithosphere is similarly raised by this factor.

Subsequent thermal subsidence follows the same principles as defined for the simple instantaneous extension.

6.4.6 Subsequent extensional phases

Extension in a phase much later in the evolution of a basin can be modelled. The initial, fault-related, subsidence is again dependent on the ratio of crustal and lithosphere thicknesses prior to the extensional event (Dewey, 1982). Crustal thickness remains the same as at the end of the last phase of faulting (ie. \( t_e/\beta_n \)). Sub-crustal lithospheric thickness is altered to include both the earlier extensional events and the post-extensional thermal relaxation of the lithosphere isotherms. This latter effect is compensated for by the following (Fig. 6.14b):-
Figure 6.14 Consideration of
a) the effects of finite extension, and
b) the thickness of the lithosphere at time $t$ after the last extensional event.
\[ a_{t2} = a - (a_a/\beta_a) (1 - e^{-t2/\tau}) \] (Modified from Le Pichon & Sibuet, 1981)

- \( a_{t2} \): lithosphere thickness at time of subsequent extensional phase
- \( t2 \): time elapsed since previous extensional phase
- \( \tau \): decay time constant (taken as 62.8 Ma)

The thermal effects of this later extensional event are simulated in the same way as for finite extension. The subsidence of the crust at \( t2 \) is a function of the crustal extension only, independent of whether extension was instantaneous, finite extension rate or multiple rifting (Fig. 6.15).

6.5 Summary of modelling philosophy

A theoretical model for the subsidence of a basin and adjacent margins has been developed, that uses a finite difference approach. The model simulates instantaneous and finite rifting, with both lateral and vertical heat flow. Subsequent rift phases can also be incorporated.

Inhomogeneous stretching can be used in the modelling. Detachment of the crust and sub-crustal lithosphere is assumed to occur at the Moho, without any specific reference to mechanisms. Hence the model is valid for both extensional regimes, and thermal attenuation of the base of the lithosphere (Sleep, 1971).

A subsidence curve can be produced within the model by defining a specific point at which subsidence is recorded. This may be within the basin or on the flanks, where lateral heat flow and inhomogeneous extension are important. Modelling of a subsidence curve, derived from a well, can be achieved. The curves, therefore, have two fixed points: the initiation of subsidence, and the present day water-loaded depth to basement (Fig. 6.15). The latter is a measure of the crustal extension, assuming the lithosphere is presently cooled. The philosophy behind the modelling (Fig. 6.16) has been to begin with the oldest event and obtain a reasonable fit between the theoretical and well subsidence (referred to as the
Curves obtained from finite difference models

A - Inhomogeneous finite extension (60 Ma.) $\beta_c = 1.10$, $\beta_{sc} = 1.30$

B - Homogeneous finite extension (60 Ma.) $\beta_c = 1.10$

C - Homogeneous finite extension (10 Ma.) $\beta_c = 1.10$

Figure 6.15 Examples of subsidence computed from the finite difference model, compared to a subsidence curve obtained from the mathematical approach of McKenzie (1978). Minor differences in total subsidence are related to errors and approximations in finite difference analysis. Nevertheless, the final subsidence is observed to be dependent on crustal thinning only, not the mechanism by which this occurred.
Figure 6.16 Flowchart for the principles behind the interpretation of the subsidence curves. Homogeneous extension is assumed until the curves can no longer fit the geological evidence.
geo-history curve on Fig. 6.16) curves. Homogeneous extension is assumed until the curves cannot be modelled with this assumption. Subsequent modelling assumes inhomogeneous extension, and whilst there are many non-unique solutions (in that there are two variables $\beta_\infty$ and $\beta_{vo}$), the curves must ultimately coincide at the present day water-loaded depth to basement.
Chapter 7
Subsidence mechanisms and the role of the Shetland Platform during basin evolution.

7.1 Introduction

The evolution of the East Shetland Platform is intimately related to the formation of the adjacent basins. Consequently, subsidence of the East Shetland Platform, and its role throughout basin development, is integral to the mechanisms of basin evolution in the northern North Sea. The object of this chapter is to review the subsidence history of the East Shetland Basin and the Faeroe Basin basins (Fig. 7.1), and attempt to explain the origin of the consequent onlap of Upper Cretaceous and Tertiary sediments onto the East Shetland Platform. The ideas presented in this chapter are based on subsidence curves obtained from wells in the basins, and models following the theoretical studies of basin history. The details of these have been explained in Chapter 6.

The chapter outlines the different mechanisms which may account for basin subsidence, with particular reference to subsidence of the basin flanks. Evidence for the different subsidence mechanisms is discussed in the light of the theoretical modelling of the subsidence curves. Conclusions are based on a regional study of subsidence curves, and related to the geological history of the northern North Sea.

7.1.1 Seismic reflection and refraction and constraints on subsidence models

Seismic reflection and refraction have determined the crustal thickness in the northern North Sea (Solli, 1976; Beach et al. 1987; Klemperer, 1988) (Fig. 7.2). Estimates of crustal thickness vary from 30-35 km under the East Shetland Platform, 20-25 km under the northern extension of the East Shetland Platform (the 'North
Figure 7.1 Regional map of the northern North Sea. Shaded area indicates the position of wells from which subsidence curves were derived. Note this area is along the western flank of the Viking Graben, and that almost all the wells are located in the East Shetland Basin.
Figure 7.2 Crustal configuration of the northern North Sea based on a) seismic refraction, and b) & c) deep seismic reflection. Seismic reflection models also indicate the two approaches to the northern North Sea, simple shear (Beach et al., 1987) and an approximate pure shear model (Klemperer, 1988).
Shetland Spur' of Beach et al., 1987) to 13-16 Km (excluding the Mesozoic-Cenozoic sediments) under the Viking Graben. It has been found that this is much greater than that predicted by the modelling of extension factors from subsidence. In the centre of the Viking Graben this discrepancy in the $\beta$ values is as much as 3.3 (subsidence curve) to 2.3 (seismic reflection) (Beach et al., 1987). Similarly, Kleinperer (1988) obtains a $\beta$ value of 2.0 for the axis of the Viking Graben. This may simply relate to uncertainties in the seismic velocities, or may indicate an element of crustal underplating in the centre of the Viking Graben. Active processes may have played a part in the evolution of the Viking Graben, although the seismic velocity evidence is equivocal on this point (cf. White et al., 1987). Furthermore, the absence of volcanism in the Viking Graben argues against such a mechanism. A major problem in the use of sediment thicknesses and subsidence curves for quantifying crustal extension factors is the assumption of the initial crustal thickness. Consequently, this discrepancy is reduced when an initial crustal thickness of 35 km is used. The extension factor based on seismic reflection increases, and the position of the 26 km crustal isopachytye of Klemperer (1988) is very close to that predicted from the backstripping approach (ie. where $\beta=1.34$) (Fig. 7.3; Table 7.1).

This difference is independent of the discrepancy between extension estimates from subsidence curves and estimates from simple summing of the heave of faults in the basin (Ziegler, 1983). Whilst the latter difference may prove reconcilable (White et al., 1986) in theoretical terms, it is very difficult to prove in practice.

Crustal thickness values obtained by seismic reflection, therefore, - whilst giving an indication of the overall amount of extension - give limited information (Klemperer, 1988) on the mechanisms that existed during the evolution of the northern North Sea. Indeed, whole crustal $\beta$-values derived from seismic, or backstripping, methods are independent of the mechanism of lithospheric extension (as shown in Fig. 6.15). Temporal and spatial changes in subsidence
Figure 7.3 Whole crustal extension factors based on the magnitude of subsidence obtained by backstripping of wells.
mechanism can be inferred by a regional study of subsidence curves, bearing in mind that the absolute values obtained may not be entirely in accordance with the crustal thickness values derived seismically.

7.2 East Shetland Platform - models for subsidence
Because of its location, on the western flank of the East Shetland Basin and Viking Graben, subsidence of the East Shetland Platform must have been inextricably related to subsidence in the basins. The difficulty is in determining the mechanism for subsidence because, in a structural context, the platform shows relatively little evidence for extension. The platform may have undergone one (or none) of the following evolutionary processes:

i) Homogeneous pure shear. Extension is confined to within the bounding faults of the Viking Graben and the East Shetland Basin. In this case platform, subsidence can only be related to flexural downwarping. Otherwise, subsidence can only result from lateral heat flow and consequent uplift and erosion of the basin flanks. Without erosion of the flanks, the thermally induced uplift subsides back to sea level (Sleep, 1971). The amount of uplift is dependent on the magnitude of extension in the basin.

ii) Inhomogeneous pure shear. Depth-dependent stretching results in a broad extension of the sub-crustal lithosphere under the unextended flanks of a basin. In this case, it would require upper mantle extension in the Viking Graben to occur over a wider area than the bounding faults. As for i), subsidence of the flanks requires erosion of uplifted crustal material. White & McKenzie (1988) have suggested that platform subsidence may result, if the region of crustal ductile extension is broader than the region of crustal brittle extension. Again flexural downwarping may play a part in flank subsidence (Watts et al., 1982).

iii) Simple Shear. Lateral movement of the crust along a low angle detachment may induce isostatic uplift of unextended footwall. At the same time, a broader zone of ductile extension and flexure may aid the formation of a broad post-rift basin.
7.2.1 Lithospheric flexure
Subsequent basin development may be modified by the mechanical effects of extension and sediment loading. Most important of these are footwall uplift, and the flexure of the continental crust. Footwall uplift involves the geometrical consequences of the rotation of fault blocks. Flexure involves the accommodation of sediment loading by a rigid crust. The deflection caused by loading may, therefore, be over a wider area than occupied by the load itself. Many of the early studies of crustal flexure focussed on oceanic crust (Fig. 7.4a). Watts (1975) defined flexure as the response to loading of the elastic thickness ($T_e$) of the crust. The amount of deflection of the crust as a result of loading was observed to be related to the age of the oceanic crust at the time of loading. Consequently, it was established that effective elastic thickness of oceanic crust was coincident with the change in depth with time of the 300-600°C oceanic isotherms. On this basis, Watts et al. (1982) derived a model for flexure within a sedimentary basin, based on the elastic thickness being equivalent to the depth of the 450°C isotherm (Fig. 7.4b). This produced a steer's head basin (Dewey, 1982), with downwarping of the basin flanks. In general, the older the thermal age of the lithosphere (and consequently, the larger the effective elastic thickness), the broader the area of deflection, and lower its amplitude. Consequent changes in the elastic thickness resulted from thermal decay of the extended lithosphere. This gradual change resulted in sediment sequences onlapping the steer's head basin (Fig. 7.4c).

The problem with this model is the fundamental assumption that continental and oceanic lithosphere have the same rheological properties. Whilst the continental crust may follow the same flexural principles, the effective elastic thickness of continental crust may not simply follow the same isotherms as for oceanic. Few values for effective elastic thicknesses for continental regions have been documented, most of which are based on very short geological time scales of loading and lithosphere age (Watts et al., 1982). Studies of intra-continental basins suggests that the
Figure 7.4 Illustration of some of the theory behind the flexural rigidity of continental crust. a) The relationship between the age of oceanic lithosphere and the effective elastic thickness ($T_e$). b) A similar relationship for the continental lithosphere. Both from Watts et al. (1982). c) The theoretical response of a basin to a gradual increase in $T_e$ related to cooling of the basin.
effective elastic thickness is of the order of 5 km during the post-rift subsidence phase (Barton & Wood, 1984; Nunn & Sleep, 1984; Warner 1987). Generally, this includes the assumption that syn-rift subsidence is accommodated by Airy isostasy (Barton & Wood, 1984). This might only be the case where sub-crustal extension provides a greater heat flow and a consequent resetting of the thermal age of the lithosphere. White & McKenzie (1988) present a theoretical basis for a low elastic thickness of continental crust. There may, however, be situations when it is very much higher than this, although no value for upper limit has been established (Ahern & Mrkvicsa, 1984). Kusznir & Karner (1985) argue that flexural strength changes, during cooling, from being controlled by the quartz rheology to the olivine rheology of the lithosphere. This represents an increase in the flexural strength of the lithosphere with time. Similarly, much of the modelling of flexure has assumed a homogeneous crust, and has not taken into account the accommodation of differential loading by small scale movements of faults and crustal blocks. This in itself, may simulate the effect of Airy isostasy and low effective elastic thickness (cf. Badley et al., 1988).

The role of flexure is therefore a complex one. To simplify the modelling discussed here, the flexural rigidity has, initially, been ignored. The gravity data in the northern North Sea indicates a very low flexural strength and most of the subsidence can be explained assuming Airy isostasy. Where subsidence curves deviate significantly from the models then flexural rigidity of the crust may be invoked as a possible additional factor. Use of wells across the region allow definition of areas where such deviations from the thermal model occur.

7.3 Subsidence of the East Shetland Basin
7.3.1 Recognisable subsidence phases
The magnitude and shape of the subsidence history, using curves derived from well data, gives an indication of the type of subsidence mechanism involved in basin formation. Similar studies in
the North Sea (Barton & Wood, 1984; Zervos, 1986; Beach et al., 1987; White, 1988) have concentrated on a two phase subsidence history - equivalent to the extensional and post-extensional subsidence phases of McKenzie (1978). Triassic extension have briefly been considered by Giltner (1987) and Zervos (1986). Every model based on horizontal lithospheric extension (depth-dependent, inhomogeneous, finite (Jarvis & McKenzie, 1980; Cochran, 1983))- is therefore directly applicable.

Curves show common features and have been divided into groups (Table 7.1), and for economy of space, only representative examples are presented in this chapter. These groups are:

S1 - Jurassic and much of the Cretaceous is present in the well (Fig. 7.5).

S2 - Wells in which all the early Cretaceous is absent, and late Cretaceous sediments directly overlie Triassic and Jurassic (Fig. 7.11 & 7.12)

S3 - Wells on the East Shetland Platform that contain a thin Mesozoic sequence, and for which the subsidence curve indicates rapid subsidence during the Palaeocene (Fig. 7.13).

S4 - Wells in the East Shetland Basin for which the subsidence curve displays a rapid subsidence phase during the Palaeocene (Fig. 7.7).

Four distinct phases of basin subsidence have been recognised on the subsidence curves for the northern North Sea. These are:-

a) Triassic-Upper Jurassic. Rapid subsidence across the East Shetland Basin.

b) Upper Jurassic. A second phase of rapid subsidence across the basin.

c) Lower Cretaceous. Gentle subsidence and a thin succession across the basin.

d) Palaeocene. Rapid subsidence that is most distinct along the margin of the East Shetland Basin.

The mechanisms invoked, and the values obtained for these events are outlined below.
Table 7.1 List of 8-values obtained by modelling of the well subsidence curves. Values include the Triassic and Upper Jurassic determined directly from modelling. Whole crustal values are computed either by direct modelling or by computation from: S. = C. / (1-1/8)

Le Pichon & Sibuet (1985)

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<th>Upper Jurassic</th>
<th>Whole crustal</th>
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<td>Maximum B</td>
<td>B-value</td>
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</tr>
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</table>

1 - Computed from modelling of subsidence curves
2 - C. = 7.136 for crustal thickness of 30 km at top Triassic
3 - C. = 8.899 for crustal thickness of 35 km at base Triassic (hence minor discrepancies occur between Jurassic maximum and whole crustal values)
Figure 7.5 A group S1 subsidence curve of 3/02-1, with an almost complete Mesozoic and Cenozoic sequence. Note how curve can not be modelled in terms of homogeneous extension.

All curves in this chapter show subsidence since Triassic only.
7.3.2 Triassic-Lower Jurassic subsidence

The timing of extension during the Triassic is conjectural. Badley et al. (1988) argue that the rifting occurred in the Permian to early Triassic on the eastern margin of the Viking Graben. The late Triassic to middle Jurassic represents, in this case, a thermal subsidence sequence. However, this proves difficult to trace across the graben and it is possible that extension began at a later date along the western flank of the Viking Graben, since Beach et al. (1987) define their Rift Phase 1 as a homogeneous extension from Middle Triassic to early Jurassic.

β-values for the Triassic have been computed from the estimated thickness of Triassic for each well (Table 7.2). The subsidence during the Triassic has been taken as the change in depth to basement from the base to the top of the Triassic. Because the timing and extent of Triassic rifting is unknown, subsidence has been assumed to be entirely extensional. A β-value was calculated using the equation of Le Pichon & Sibuet (1981). Since the present day thickness of the East Shetland Platform is 30-35 km, and the platform underwent some extension through the Triassic, the upper value has been used for the pre-Triassic crustal thickness.

Because the Triassic subsidence may include some post- and syn-rift thermal subsidence the β-values are a maximum. From Jarvis & McKenzie (1980), the criterion for assuming instantaneous stretching is that the duration of the rifting be less than 60/β^2. This computes as 33.4-41.7 Ma for β=1.20-1.34, (Table 7.2) which represents almost all of the Triassic. Consequently, syn-rift thermal subsidence does not appear to have been important during the Triassic.

In some wells extension has been modelled into the Lower Jurassic, but this is more likely the result of an underestimate of Triassic subsidence and extension. In these wells the β-values in the Lower Jurassic are small (1.01), and have little effect on the overall value of Triassic extension.
Table 7.2 Computation of the lithospheric extension from the decompacted and unloaded Triassic subsidence.

<table>
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<th>Well Number</th>
<th>Triassic decompacted sediment thickness (m)</th>
<th>Triassic subsidence (m)</th>
<th>$t_c$ (km)</th>
<th>$a$ (km)</th>
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<td>1.14</td>
<td>30.6</td>
</tr>
<tr>
<td>211/12-1</td>
<td>1400</td>
<td>1117</td>
<td>1.32</td>
<td>26.6</td>
</tr>
<tr>
<td>211/16-1</td>
<td>1500</td>
<td>1134</td>
<td>1.32</td>
<td>26.5</td>
</tr>
<tr>
<td>211/18-1</td>
<td>1600</td>
<td>1194</td>
<td>1.34</td>
<td>26.0</td>
</tr>
<tr>
<td>211/21-1a</td>
<td>1800</td>
<td>1296</td>
<td>1.39</td>
<td>25.3</td>
</tr>
<tr>
<td>211/28-2</td>
<td>1560</td>
<td>1183</td>
<td>1.34</td>
<td>26.0</td>
</tr>
<tr>
<td>211/28-6</td>
<td>1600</td>
<td>1233</td>
<td>1.36</td>
<td>25.7</td>
</tr>
</tbody>
</table>

tc - crustal thickness at the end of the Triassic, assuming the subsidence is entirely due to extension

a - lithospheric thickness at the end of the Triassic, assuming extension occurs instantaneously at the end of the Triassic.
The residual effects of this major rift phase are particularly important for the post-Triassic subsidence of the northern North Sea. Hence, if rifting took place into the Lower Jurassic, as also suggested by Beach et al. (1987), the thermal subsidence from the late Permian-Triassic extension would be occurring during the Middle Jurassic and would have a profound effect on the subsidence during the Jurassic, as well as the estimates of post-Triassic crustal extension (Giltner, 1987; Hellinger et al., in press).

Triassic subsidence has therefore been represented by two different regimes in the modelling of the northern North Sea subsidence curves. Firstly, by modelling the post-Triassic subsidence with an initial Lower Jurassic lithospheric thickness of 125 km, and crustal thickness of 30 km. This presumes that all subsidence from the Triassic has ceased. Hence the basin is not already undergoing thermal subsidence related to an earlier event. All post-Triassic subsidence is thus dependent only on extension in the Jurassic. Consequently, this model maximises the amount of extension in the post-Triassic. In the second model, the lithosphere and crustal thickness have been assigned values based on the amount of Triassic extension. This model therefore presumes that the Triassic event has just taken place, and that no thermal subsidence has yet occurred after this event. Consequently, this model minimises the amount of post-Triassic extension, because the basin is already subsiding by thermal decay through the Jurassic. In reality, the Triassic may contain both fault-related and thermal subsidence (Giltner, 1987).

7.3.3 Upper Jurassic

A second phase of extension occurred in the Upper Jurassic (Beach et al., 1987; Badley et al., 1988). Throughout the Middle Jurassic, shoreface and lagoonal sediments of the Brent Group were deposited across the East Shetland Basin as far north as block 210/25 (61° 20'N) (Brown et al., 1987). Subsequent deposition was of marine shales and limestones across the basin and related to basin foundering during the Oxfordian. The development of the Magnus submarine fan, during the Kimmeridgian (De'ath & Schuyleman, 1981),
implies the formation of structural topography produced by tilting of the Magnus Ridge, and generation of a sediment source.

As discussed in the preceding section, β-values for the Upper Jurassic extension are critically dependent on the lithosphere regime at the end of the Triassic. The mean extension value for the Upper Jurassic event was 1.12 ± 0.07 (Table 7.1). Where the lithosphere is assumed to have cooled to its pre-rift thickness of 125 km by the end of the Triassic, the Jurassic subsidence model can be made to fit the subsidence curve. However, this places a substantial amount of extension into the Lower Jurassic. Where Triassic extension is assumed to have just taken place, the models can be made to fit during the Upper Jurassic extension, but do not match during the preceding Middle Jurassic quiescence. This is especially true for Group S1 wells north of 61°N on the Magnus Ridge. The magnitude of the mismatch on Magnus Ridge is as much as 400m.

This mismatch on Magnus Ridge may be explained in terms of erosion of the top of the Bathonian-Callovian sequence by Upper Jurassic footwall uplift. The response of the East Shetland Basin to extension in the Viking Graben would appear to have been the foundering of fault blocks parallel to the Viking Graben. Localised, small scale (150-300m) mismatches south of 61°N are probably the effect of tilting of fault blocks. Uplift of the Magnus Ridge does not follow the Viking Graben but the same process probably prevailed there (Fig. 7.6).

Footwall uplift will result in a delay in subsidence, into the later stage of the extensional event, or indeed into the thermal subsidence phase. In the most severe cases, no extensional subsidence phase is recorded (Sawyer, 1986), and is a fundamental problem of the location of exploration wells on the crests of tilted fault block structure (Sawyer, op. cit.; Giltner, 1987). Such a model is indeed more likely than a depth-dependent extension, in
that the effects of the latter would not be so localised. The magnitude of the footwall uplift is discussed later.

Beach et al. (1987) have postulated that the Upper Jurassic extension in the north Viking Graban was influenced by NE-SW sinistral transtension. Most of the observed Jurassic subsidence can be explained by homogeneous extension south of 61°N, but the position of footwall uplift is not inconsistent with a sinistral strike-slip origin to the More Basin (Beach, 1985). No wells penetrate the deep part of this basin, so the origin remains inconclusive, in terms of pre-Cretaceous subsidence history.

Wells on the East Shetland Platform (9/16-1 and 3/28-1; Table 7.1) show the presence of thin Upper Jurassic shales. There is no evidence that this area was undergoing erosion induced by either inhomogeneous extension or footwall uplift. Moreover, the lack of convincing evidence for a basin margin sequence along the edge of the East Shetland Platform (Turner et al., 1987; Harris & Fowler, 1987) argues against active uplift of the edge of platform at this time.

Consequently, it can be concluded that the Upper Jurassic extension was broadly homogeneous. The absence of significant uplift along the flanks of the basins argues against simple shear deformation across the East Shetland Basin. Uplifts, as evidenced by the deviation of the subsidence curve from a pure shear thermal model, can be explained by tilting of fault blocks and footwall uplift. Magnus Basin may have resulted from localised sinistral shear during the Upper Jurassic.

### 7.3.4 Lower Cretaceous

The Lower Cretaceous is thin over the East Shetland Basin, and consists of shales and occasional marls. The margin of the Shetland Platform shows the development of limestones in a marginal setting (Wheatley et al., 1987; Amiri-Garoussi, 1987). The rest of the basin contains fault-controlled shales, which in places onlap the Jurassic.
to Triassic tilted fault blocks. Consequently, the Lower Cretaceous has been considered both a syn-rift (Beach et al., 1987) and post-rift sequence (Badley et al., 1988). The equivocal nature of the sequence requires both alternatives to be considered in the modelling of the subsidence curves.

In the East Shetland Basin the Lower Cretaceous is dominated by a lengthy unconformity, with the Neocomian absent. Rawson & Riley (1982) suggest that this hiatus is composed of a series of eustatically controlled transgressions and regressions. From their interpretations the 'Barremian' transgression indicated in a number of wells may represent condensed — but nevertheless complete — Lower Cretaceous sequences. However, there is a relationship between wells in which Lower Cretaceous sediments are entirely absent and their location within the East Shetland Basin. Wells located on Magnus Ridge, the margin of the Magnus Basin and south of 60°30′N show this absence of Lower Cretaceous sediments. A very thin sequence of fine-grained sandstone on the Shetland Platform (well 9/16-1 and Harris & Fowler, 1987) suggests that the coastal margin was developing on the flanks of the basin, west of the edge of the graben. This implies that the wells with little or no Lower Cretaceous represented palaeotopographic highs during this period. These highs coincide with areas where early Jurassic uplift was indicated. It is probable active faulting, and/or fault relief, continued from the Upper Jurassic into the Lower Cretaceous, which is preserved in the topographic lows of the fault blocks.

7.3.5 Upper Cretaceous

Within the East Shetland Basin the Upper Cretaceous is represented by shales, that encroach onto the East Shetland Platform. The distribution of the Upper Cretaceous is characterised by the lack of faulting in the northern North Sea. All fault movements had evidently ceased by the Aptian (Badley et al., 1988).
On subsidence curves, the Upper Cretaceous is observed as a phase of exponential subsidence that begins in the Cenomanian, and continues to the present.

In terms of extensional basins, this period appears to be the thermal subsidence phase related to the thermal decay of extended lithosphere (McKenzie, 1978).

7.3.6 Palaeocene
During the Palaeocene, the rejuvenation of the East Shetland Platform (Knox et al., 1981) as a sediment source area caused submarine fan sands to be deposited in the East Shetland Basin. At the same time the bounding faults to the basin were reactivated, and a palaeoslope formed along the edge of the platform. Submarine fans were developed along the scarp (Heritier et al., 1981; Stewart, 1987). Only later, during the late Palaeocene, was the platform onlapped (Wheatley et al., 1987). Subsidence associated with fault movement was not observed on the Magnus Ridge, although thinning of the Palaeocene occurs over the structure.

A number of the subsidence curves show an increased rate of subsidence during the Palaeocene (Fig. 7.7), with a concomitant increase in the sediment accumulation rate. These curves are located along the western flank of the East Shetland Basin, and on the East Shetland Platform. Apparent $\beta$-values for this event—assuming all Palaeocene subsidence was related to extensional fault subsidence—are of the order of 1.0–1.1, and locally 1.10–1.20 (Fig. 7.8).

Caution must be applied to an interpretation of this event as a true 'rift phase' (Beach et al., 1987). It has been considered by Badley et al. (1988) that this event represents the accommodation, by planar fault movement, of the load applied by a rapidly deposited thick wedge of sediment. This model implies that the movement on the fault is contemporaneous to, or slightly post-dates, sediment deposition. It similarly implies that the syn-depositional movements deform the Palaeocene and cause it to drape the fault structure.
Figure 7.7 A group S2 and S4 subsidence curve of 3/23-1 with a subsidence increase in the Palaeocene.
Figure 7.8 A map of the maximum apparent extension of the northern North Sea during the Palaeocene. Values are based on the magnitude of the subsidence increase observed in wells during this time.
However, the Thanetian sand sequences downlap onto the earlier Palaeocene, and show onlap onto the fault surfaces. This suggests the formation of a sediment wedge along a scarp that already existed (Fig. 7.9).

Despite this seismic evidence it is probable that the fault movements are as Badley et al. (1988) claim. As described in Chapter 5, the NW European region was undergoing compression from the late Cretaceous to the Palaeocene (Biddle & Rudolph, 1988), and most of the observed deformation is compressive (Ziegler, 1987) or transpressive (Caselli, 1987). It is unlikely therefore that basins outside the Atlantic rift zone of active volcanism were undergoing active extension. It is more likely that doming and uplift of the East Shetland Platform resulted in an influx of sediment into the basin to which the faults responded. The lack of intra-basinal fault block rotation and footwall uplift also argue against an extensional origin to the fault movements.

It is concluded therefore that the Palaeocene fault movements did not play an important part in the evolution of either the East Shetland Basin or the East Shetland Platform.

7.4 Quantification of subsidence mechanisms
7.4.1 Upper Jurassic to Lower Cretaceous footwall uplift
The response of crustal blocks to movements on high-angle planar faults over a detachment horizon is to rotate in a 'domino'-style (Wernicke & Burchfiel, 1982). Undefined mechanisms at the detachment provide spatial conservation. Consequently, the footwall to each individual fault is uplifted, as the hanging wall subsides. The mean subsidence of the block is the theoretical value of basin subsidence, derived from $\beta$-values (Barr, 1987) (Fig. 7.10). From geometrical considerations, the magnitude of uplift is a function of the angle of the faults, the fault spacing and the basin $\beta$-value. Values for footwall uplift are derived from Barr (1987; his figure 7), and are specified for $\beta$ and the initial fault spacing, a'.
Figure 7.9 Interpretation of a seismic line, in the vicinity of quadrant 9N, through the area of fault reactivation during the Palaeocene.
Figure 7.10
Quantification of the magnitude of footwall uplift, derived from Barr (1987). Relationship between extension, fault spacing and amount of uplift above sea level.

Open squares - estimates for Magnus Ridge Triassic and upper Jurassic (A), and Upper Jurassic only (B). Solid square - estimate for quadrant 3S.

(After Barr, 1987)
For the Magnus Ridge, the spacing between the major faults at the top Jurassic is 24 km (Chesher & Turner, 1987). If the crustal extension for both the Triassic and Jurassic events is taken into account, then $\beta=1.25$. Fault spacing, prior to extension and rotation, was consequently $(24/1.25)=19$ km. Ignoring effects of Upper Jurassic and Lower Cretaceous erosion (Bowen, 1975; Barr, 1987), the footwall uplift, above sea level, is about 1.7 km. The width of the exposed fault block is 7 km, encompassing all the wells which show the maximum deviation in the subsidence curves (Fig. 7.6). The amount of erosion may be less than this uplift, if a finite value of erosion rate is taken into consideration. Finite extension will also reduce the amount of uplift, as syn-rift thermal subsidence counteracts the rotational uplift of the fault block crests. The apparent delay in subsidence, from the Upper Jurassic to the mid-Cretaceous, after extension, is simply explained by the time taken for the fault crest to return to sea level.

Even assuming a Jurassic origin for Magnus Ridge, the magnitude of footwall uplift is sufficient for the observed discrepancies and delays. Correcting for the Triassic sequence in well 210/15-1, provides an estimate of $\beta=1.10$ for the Upper Jurassic event (Fig. 7.10 and Fig. 7.11). This implies a footwall uplift of 1 km.

The time after rifting, at which the footwall crest sinks below sea level is dependent on the erosion rate of the crest, and the subsidence rate of the basin. The rate of subsidence is:

$$-\frac{dU}{dt} = (1-\rho_c/\rho_m)R + dS(t)/dt$$

the solution of which is:

$$U(t)=U_0-S(t)-(1-\rho_c/\rho_m)Rt$$

$U_0$ - Initial value of uplift
$S(t)$ - Subsidence since time of uplift
$R$ - Erosion rate

The amount of erosion observed - $(1-\rho_c/\rho_m)Rt$ - is taken to be the same as for the Brent block, an estimate of 700-1000m (Barr, 1987). $S(t)$ is related to the crustal extension of the basin as a whole. Consequently, the footwall crest subsides to sea level when
Figure 7.11 Example of a Group S2 subsidence curve (210/15-1) for the crest of Magnus Ridge, showing both an inhomogeneous extension and footwall uplift interpretation to the missing Cretaceous section. Curves are plotted as post-Triassic subsidence decompacted to base Triassic.
Taking a footwall uplift of 1700m, the subsequent thermal subsidence is insufficient to reduce this back to sea level. The uplifts associated with the footwall are computed assuming instantaneous extension. The probable effect of finite extension is to reduce the total uplift, by superimposing thermal subsidence onto the isostatic and rotational uplift. Furthermore, the rise in sea level from the Upper Jurassic to mid-Cretaceous was of the order of 100-150m (Haq et al., 1987). A smaller value of initial uplift (1000m) subsides to sea level in 60 Ma. This is very close to the observed time of onlap of the footwall crest.

The other region of absent Lower Cretaceous is quadrant 3S. In this area, fault spacing is presently 13.25 km. Total extension of $\beta=1.37$ is concluded from the subsidence curves, giving $a'=9.67$ km. Hence the amount of footwall uplift is only 100-250m above sea level. The width of blocks above sea level is only 1.12 km. This should have subsided below sea level within a few 10's Ma, unless continual extension propped up the footwall crests.

The biggest problem with this model is that the subsidence — after the delay of up to 50 Ma due to uplift — during the Upper Cretaceous is very much more rapid than implied by subsidence following a simple, footwall uplift model. From the shape of the modelled curves, some mechanism is required that prevents the dissipation of the thermal effects of extension for a significant length of time. Giltner (1987), indeed, argues that away from the crest of fault block, the subsidence curves derived from seismic data do not easily model as homogeneous extension. This is supported by well 3/02-1 (Fig. 7.5), in which the entire succession is present, but the shape of the curve is not that of homogeneous extension. However, the footwall uplift approach gives the best approximation to the observed tectonism in the Upper Jurassic to Lower Cretaceous, if not the Upper Cretaceous.
7.4.2 The role of depth-dependent stretching during the Lower Cretaceous

On the Group S2 subsidence curves the Lower Cretaceous is observed as a period of little or no subsidence. The crust, during this period has the appearance of being held or buoyed up, against the predicted thermal subsidence. Some of this may be apparent in that palaeo-bathymetric data is scarce for the whole of the period. However, the evidence for erosion of fault block crests suggests that these East Shetland Basin was close to, or above sea level. Thus, this period indicates a large deviation from modelled pure shear extension on both Jurassic maximum and minimum extension values. The maximum magnitude of the deviation, between modelled and observed subsidence, is of the order of 300-800m.

If the duration of the Upper Jurassic fault phase is increased to include the Lower Cretaceous, then a depth-dependent model for extension, in the East Shetland Basin, can explain the lack of apparent subsidence. A close match to the subsidence curve is obtained for sub-crustal lithosphere $\beta_c$ values of $\approx 1.60-2.00$ (Fig. 7.12). A major constraint on the extension is that regional extension must be the same for upper crustal and sub-crustal extension factors (Rowley & Sahagian, 1986). Inhomogeneous extension of the East Shetland Basin is only possible if the $\beta_c$-value for the Viking Graben is as high as 2.0 (Beach et al., 1987). The corresponding $\beta_c$-value for the Viking Graben would also have to be 1.60, implying a broad, regional pure shear of the sub-crustal lithosphere. Areas of Lower Cretaceous non-deposition - or simply, reduced sedimentation - may represent where active faulting rotated fault block crests, and caused footwall uplift to near sea level.

Where the full thermal effect of Triassic rifting is considered, it becomes very difficult to model the Lower Cretaceous in terms of depth-dependent stretching. The rapid heat loss and subsidence resulting from the early stages of post-rift subsidence have first
Figure 7.12 Example of an S2 curve (210/29-1) which shows a good fit between predicted and observed subsidence for an inhomogeneous extension event during the Lower Cretaceous. Curves as for Fig. 7.11.
to be counteracted. This increases the $\beta_{m}$ for the East Shetland Basin to at least 1.8, and consequently introduces space problems for the areas of Lower Cretaceous non-deposition.

This depth-dependent approach also requires a major assumption in the continuation of the Upper Jurassic rifting as late as the Albian. Badley et al. (1984) argue against fault activity in the Oseberg region occurring later than Berriasian, although active faulting may have continued for longer in other areas. Even so, extension appears to have been sporadic across the basin, rather than a widespread event similar to the Upper Jurassic activity. This puts a severe constraint on the depth-dependent model, in that rapid subsidence occurs immediately after the cessation of depth-dependent extension.

7.4.3 Summary of basin mechanisms during evolution of the East Shetland Basin

The following conclusions can be drawn as to the controlling mechanisms on the development of the East Shetland Basin:

i) Triassic subsidence, as far as can be inferred, occurred by a homogeneous extension in a north-trending basin.

ii) Lower and Middle Jurassic was characterised by minor extension and thermal subsidence.

iii) Upper Jurassic extension appears to have been homogeneous - or mildly inhomogeneous - in form. Footwall uplift was responsible for the uplift and erosion of structural highs. Such a mechanism also explains the lack of Lower Cretaceous sediments over much of the East Shetland Basin.

iv) Lower Cretaceous sedimentation was low, and dominated by the structural regime of the Upper Jurassic. Footwall crests began subsiding, but may have also been maintained by thermal effects of a minor inhomogeneous extension event centred on the Viking Graben.

v) Upper Cretaceous to present subsidence was rapid, and exponentially decaying. The rapidity may have been aided by an input of heat into the basin during the Lower Cretaceous, but none of the models is entirely satisfactory in explaining this rapid subsidence.
Fault reactivation took place in the Palaeocene as a result of uplift of the East Shetland Basin, and rapid loading of the basin.

The processes that took place in the East Shetland Basin can be related to the origin and development of the East Shetland Platform.

7.5 Origin of East Shetland Platform subsidence

Similar subsidence phases to the East Shetland Basin are observed on the East Shetland Platform. On the platform itself, most wells terminate in the Devonian, below Cretaceous to Palaeocene sediments. Wells 3/28-1 and 9/16-1 differ from the rest in that they penetrate a Triassic sequence of sands. Basins such as the Unst, Fetlar and the Fair Isle Basins, have also proved to be infilled with Permo-Triassic sediments. Thus, there is evidence that extension during this time was not isolated to the very deep Viking Graben, and that mild extension occurred on the East Shetland Platform. The value of Triassic-Lower Jurassic extension is small ($\beta=1.04$) (Fig. 7.13), and does not relate to the total crustal extension inferred from the present unloaded depth to basement (1.11 - well 3/28-1; Table 7.1).

Most of the wells on the East Shetland Platform only record subsidence from the Palaeocene, but the total extension may be composed of a series of events that both extended the crust, and hindered the immediate subsidence of the platform. Since the Triassic, an extra $\beta=1.16$ must have been provided by some subsequent mechanism. The Upper Jurassic is the only extensional event which can have caused this extra apparent extension.

7.5.1 Flank uplift and erosion models - Magnitude problems

The influence of flank uplift and subsequent erosion have been postulated as mechanism for later flank subsidence below sea level (Sleep, 1971; Leeder, 1983; Hellinger & Sclater, 1983). The mechanism requires that attenuation (by extension or heating) of the sub-crustal lithosphere produces an uplift of unextended or less extended crust, at the flank of a basin. The amount of uplift is

$$U_0 = 1.40(1-1/\beta_{\infty})$$

(Hellinger & Sclater, 1983)
Figure 7.13 A well (9/16-1) on the East Shetland Platform in which a fairly complete section is observed, within a small fault system. Note how little evidence there is of uplift or exposure of crustal basement above sea level.
for crust of thickness 30 km - approximate crustal thickness at the start of Upper Jurassic extension in the northern North Sea. The maximum amount of uplift that is required to produce the observed crustal thinning by erosion only, is equated by

\[ D_{\text{w}} = 1.4U_0 \]

(Hellinger & Sclater, 1983)

\( D_{\text{w}} \) - Water loaded subsidence at time \( t = 0 \)
\( U_0 \) - Initial uplift due to inhomogeneous extension

The maximum unloaded subsidence in quadrant 3 on the East Shetland Platform is 980m, which to be produced by erosion alone required 700m of uplift. Such uplift, by sub-crustal lithospheric extension only, requires a \( \beta_{\text{w}} = 1.9 \), very much greater than estimates for the margins of the basin as a whole. Similarly, this thermal uplift-erosion model alone can not adequately explain why the platform does not rapidly subside through the Lower Cretaceous. When a finite value is specified for the erosion rate of uplifted crust, a delay occurs between the instantaneous uplift of the crust, and the subsequent subsidence to sea level. A lower value of erosion rate increases this delay time. However, this means that less of the uplifted crust is eroded. Consequently, application of a finite erosion rate, requires a larger value of \( \beta_{\text{w}} \) to produce an observed 980m of subsidence.

The Upper Jurassic is the only extensional event which can have caused this extra apparent extension. As has been discussed, the platform and crests of tilted fault blocks may have been close to sea level or exposed, but there is no geological evidence for active erosion of the edge of the East Shetland Platform. It is concluded that uplift-erosion models, on their own, can not explain the flank subsidence in the northern North Sea. On the same basis footwall uplift and isostatic rebound of unloaded crust - in requiring erosion of the crust - are also discounted. Indeed, where the platform becomes broad, in quadrant 3S, the footwall uplifts in the basin are very low.
7.5.2 Inhomogeneous extension of the East Shetland Platform

Subsidence on the East Shetland Platform may be related to an inhomogeneous extension (White & McKenzie, 1988). This may be particularly important around quadrant 9W, where the platform is composed of a granitic pluton (Donato, Tully, 1982). The density contrast between the granite and the basement rocks of the East Shetland Platform may have aided the delay in subsidence of this region (Scrutton et al., 1987). The granite plays an important role in controlling the fault locations, and shaping the basin. It is highly possible that extension occurred in the sub-crustal lithosphere, and ductile lower crust below the granite. The platform is broadest to the south, and this may represent a shadow zone behind the rigid granitic block. Consequently, ductile extension would have occurred over a broader area than brittle extension. (Fig. 7.14) The relationships of Hellinger & Sclater (1983) imply that there is a point where the ratio of $\beta_c$ and $\beta_{mc}$ produce no initial subsidence. This ratio is:

$$\lambda_{nc}/\lambda_c = 2.976$$

where $S_i = C_1\lambda_c - C_2\lambda_{nc} = 0$ \hspace{1cm} Hellinger & Sclater

and $\lambda_c = (1-1/\beta_c)$ \hspace{1cm} (1983)

$\lambda_{nc} = (1-1/\beta_{nc})$

$$C_1/C_2=\{(a(p_m-p_c)tc(1-\alpha T_m T_c/2a)-\alpha p_m T_m T_c/2)\}/$$

$$\{(a-t_c)^2\alpha p_m T_m\}$$

$$= 2.976$$

(Symbols and values from Table 6.4)

For this to occur in one event on the East Shetland Platform, where well 3/28-1 shows a $\beta_c = 1.11$, $\beta_{mc}$ must have been 1.42. The crustal thinning has no fault expression, and may be related to ductile, lower crustal extension only. The thickness of this is constrained, since:

$$t_c/\beta_c = t_m/\beta_m + t_d/\beta_d$$ \hspace{1cm} Hellinger & Sclater

$$= (t_m + t_d)/\beta_c$$

(1983)
Figure 7.14 Map of the inhomogeneous extent of the East Shetland Platform. Note how the position of the maximum extent of unfaulted subsidence occurs close to the granite. Depth dependent stretching may have been generated by the resistance of the homogeneous granite to fracturing during extension. The rest of the platform may lie in a 'shadow' zone, where extension is entirely lower crustal.
Hence,

\[ t_d = \{ (1/\beta_b - 1/\beta_d) / (1/\beta_b - 1/\beta_c) \} t_b \]

- \( t_b \) - Thickness of brittle crust  
- \( \beta_b \) - Extension factor of brittle crust  
- \( t_d \) - Thickness of ductile crust  
- \( \beta_d \) - Extension factor of ductile crust  
- \( t_c \) - Total thickness of crust  
- \( \beta_c \) - Whole crust extension factor

Since ductile extension in the crust can be considered to have the same value as in the sub-crustal lithosphere (in this case 1.42) and the whole crust extension factor is 1.11, then:

\[ t_d = 0.504 t_b \]

and since \( t_d = t_c - t_b \)

\[ t_b = t_c / 1.504 \]

This constrains the ductile extension to the lower 10-13km of the continental crust of the East Shetland Platform.

Again, however, the values of extension are very much greater than those observed for any single event in the East Shetland Basin and Viking Graben. The platform subsidence would therefore have to be formed by the superimposition of Triassic and Upper Jurassic extension. The extension on the platform was small, during the Triassic. It is therefore possible that the contrast between crustal and sub-crustal lithosphere extension was too small to be observed on the coarse approximations for basin development during this period.

In summary, thermal and mechanical models relating to extension provide only an approximate solution to the problem of platform subsidence. The best approximation is that of depth dependent extension, over both the Triassic and Upper Jurassic extension events, in that it produces subsidence without resorting to erosion of the crust - a mechanism for which there is no geological evidence.
7.5.3 Flexure?

As has been mentioned, there is no geophysical basis for a large effective elastic thickness of the lithosphere in the northern North Sea. However, the geological effects observed may be related to flexure of a plate of significant elastic thickness. The calculation of elastic thickness assumes a negligible syn-rift flexural rigidity (Barton & Wood, 1984) and gravity modelling of the North Sea has disregarded problem areas such as the western flank of the Viking Graben (Barton & Wood, op. cit.). Whilst no attempt has been made to use flexural rigidity in this study, a prediction of the effects of such a model can be made (Fig. 7.15), and constraints placed on the applicability of such a mechanism.

The consequences of flexural loading are the development of regions of flexural downwarping (Moat) and upwarping (Bulge). During the Jurassic, when the Viking Graben was the major depocentre (Badley et al., 1988), elastic thickness would have been at a minimum. This would have resulted in a very small amount of flexural downwarp, but a characteristically high amplitude flexural uplift of the flanks of the Viking Graben. This would also lead to a relative uplift akin to, and accompanying, the footwall uplift in the East Shetland Basin, and may have continued into the Lower Cretaceous. Subsequent cooling of the lithosphere would increase the elastic thickness of the crust. Consequently, the load imposed by early and mid-Cretaceous sediments would produce a region of flexural downwarping adjacent to the Viking Graben, and development of a flexural bulge further away from the depocentre. This migration would be further aided by the flexural loading of the East Shetland Basin, as a result of renewed sedimentation into this basin. Consequently, the flexural bulge migrates onto the East Shetland Platform.

A minor change in elastic thickness is all that is required for the migration of the flexural bulge across the East Shetland Basin, and onto the East Shetland Platform. Fig. 7.16 shows a simplistic relationship for the elastic thickness and the lateral extent of deflection of the crust. It can be seen from this relationship that
Figure 7.15 A possible mechanism by which flexure could have controlled the development of the East Shetland Basin and Platform, with a theoretical subsidence curve shown below.
Figure 7.16
A graph showing the migration of flexural uplift, as $T_e$ changes. Although simplistic in dealing with a point load, the relationship shows that minor changes in flexural rigidity could have influenced the East Shetland Basin and East Shetland Platform. Equations from Turcotte & Schubert (1982).

Relationship between elastic thickness and lateral extent of crustal deflection for a point source

Equations:

- $X_0 = 0.75\pi\alpha$
- $X_b = \pi\alpha$
- $\alpha = \frac{4D}{(\rho_m - \rho_s)g\rho_p}$
- $D = \frac{E\tau_s^3}{12(1-\nu^2)}$

- $E$ - Young's Modulus (70 GPa)
- $\nu$ - Poisson's Ratio (0.25)
- $T_e$ - Elastic thickness

$X_0$ - Width of flexural downwarp

$X_b$ - Distance to point of maximum uplift

Comparable distance from centre of Viking Graben

East Shetland Basin
Shetland Platform

$T_e$ (km)
a change in elastic thickness from 5 km to 10 km is sufficient to cause a migration of the flexural bulge across the East Shetland Basin. It should be noted that the conclusion is simplistic, and would require the East Shetland Basin to be independent of the Viking Graben, in terms of its flexural response to loading.

The analogy requires a rapid shift in the style of sediment load accommodation, during the mid-Cretaceous, from the graben to the broader basin. This may be generated by a change in the control of flexure by a quartz dominated to an olivine dominated rheology (Kusznir & Karner, 1985), occurring approximately 100Ma after the resetting of the thermal age of the lithosphere.

7.6 Subsidence in the Faeroe Basin

The Faeroe Basin (Fig. 7.17) is easier to model in terms of homogeneous extension, without recourse to mechanical consequences of extension (Fig. 7.18). A number of fault-related events are observed - Upper Jurassic, mid-Cretaceous and Upper Cretaceous. The absence of Lower Cretaceous sediments, in a number of wells on Rona Ridge, can readily modelled by inhomogeneous extension (Fig. 7.19) of the structural high. Footwall uplift, computed in the same manner as for the East Shetland Platform is 1300m ($\beta = 1.15$; $a' = 26$ km), and may therefore also contribute to the absence of Lower Cretaceous sediments. Little is known of the thickness of Triassic sediments in the West Shetland Basin, although only one well, 205/25-1, does not terminate in crystalline basement. Extension factors for the basin are therefore a maximum for the post-Triassic sequence. Rapid subsidence is again observed in the Upper Cretaceous, which has been attributed to the initiation of seafloor spreading in the Faeroe Basin (Scrutton, 1987) but equally may be related to fault movements in the West Shetland Basin (Duindam & van Hoorn, 1987), and is modelled as the latter. Otherwise, it is difficult to model this rapid subsidence in terms of lithospheric extension and thermal subsidence.
Figure 7.17 Map showing the distribution of crustal extension across the wells in the Faeroe Basin. Dotted line - extent of Erlend Igenous Complex; Dashed line - approximate extent of Rona Ridge.
Figure 7.18 Examples of subsidence for the Faeroe Basin (205/22-1a). Note how the modelled curves are a better fit to the observed subsidence.
Figure 7.19 Examples of subsidence for the Faeroe Basin (205/20-1). Includes the influence of footwall uplift of Rona Ridge.
The effects of late Cretaceous-Palaeocene volcanism are not observed on most of the curves. This is, in part, a function of the short duration of this tectonic episode compared to the time gap between stratigraphic marker horizons on the subsidence curves. Even so, well 209/9-1, situated on the Erlend Igneous Complex, shows a pause in subsidence at this time (Fig. 7.20). Minor uplift is observed on the corrected subsidence curve, a function of the unloading of the Campanian to Palaeocene volcanic sequence. This sequence was assumed to have undergone no compaction since emplacement, and to impose a load with the maximum density of cooled basalt (3.30 g/cm$^3$). An unconformity prior to this apparent uplift, provides further evidence of uplift at this time. Consequently, the subsidence curve can be modelled by a thermal event from the late Cretaceous to the Eocene (Fig. 7.20). Rapid subsidence during the Eocene is consistent with the Faeroe Basin history outlined in Chapter 5.

The fundamental difference between the West and East Shetland Basins is the absence of significant onlap of Tertiary sediments onto the West Shetland Platform, south of 61°N (Fig. 6.2) for Tertiary subcrop limits). This precludes flexural, and significant lithospheric inhomogeneous extension, being important in the evolution of the West Shetland Platform, although these mechanisms might have been important in the evolution of Rona Ridge. The apparent younger thermal age of the basin means that the rapid switch in subsidence, due to flexural changes as discussed for the East Shetland Basin, may not have been initiated.

Further detailed analysis is precluded by the localised distribution of the exploration wells west of Shetland. Good quality seismic data and deep wells penetrating to crystalline basement would be required before a more detailed study could be undertaken for this basin.

### 7.7 Summary and conclusions

A number of thermal and mechanical models for the formation of the East Shetland Basin and the East Shetland Platform, have been
Figure 7.20 Subsidence curve from a well on the Erland Igneous Complex (209/9-1). Volcanics have been assigned a low porosity constant and a high density. Consequent decompaction shows uplift, which is modelled as a thermal anomaly during the late Cretaceous and Palaeocene.
proposed. None of the models is entirely satisfactory in explaining all the observed magnitudes and timing of subsidence, but some approximate better than others. Arguments on the basis of simplistic lithospheric extension cannot account for all of the observations, and a simplistic flexural model has been proposed.

Homogeneous or mildly inhomogeneous extension of the East Shetland basin provides the best-fit model. All extension values are consistent with this interpretation, although intra-basinal simple shear mechanisms may simply have superimposed upon each other to produce an apparent pure shear extension.

It can be concluded that, until the end of the Lower Cretaceous, subsidence in the East Shetland Basin was countered by footwall uplift generated during the Upper Jurassic extension. Where extension continued into the Lower Cretaceous, footwall uplift similarly controlled sedimentation on the crests of fault blocks. Subsequent subsidence of the fault blocks occurred, and is characterised by rapid subsidence in the Upper Cretaceous. The lack of erosive geological environments along the margin of the platform suggests that erosion played no major part in the subsidence mechanism of this particular structure. The most likely mechanism for the evolution of platform subsidence was inhomogeneous extension, related to extension under a granite pluton. Subsidence of the platform is only observed from late Palaeocene onwards and may have been delayed by flexure, or the buoyancy of the granite (Scrutton et al., 1987). This may have been aided by flexural bulging followed by flexural downwarping, and the effects of flexural changes in the basin are consistent with the observed subsidence patterns.
Chapter 8

A depositional and tectonic history of the Eocene of the northern UKCS

The Eocene is one of the few stratigraphical units in the northern UKCS about which little had been previously determined of the controls on deposition, and the tectonic regime during deposition. Over the region, the Eocene has widely been regarded as 'undifferentiated'. Interpretation of regional seismic data in this study has enabled the sub-division of the Eocene succession in both the East Shetland and the Faeroe Basins with biostratigraphical and lithostratigraphical data providing supportive palaeoenvironmental information. The results have been incorporated into a relative sea level curve and compared against the Haq et al. (1987) global sea level curve. Tectonic events have been inferred, where the two curves deviate. This chapter presents a summary of the depositional and tectonic history (Fig. 8.1) for the East Shetland Basin, Shetland Platform and the Faeroe Basin.

The Eocene was deposited during the post-rift phase of Atlantic margin development. Prior to Atlantic separation, the basins bordering the present Atlantic Ocean formed a single continental plate that underwent lithospheric extension from the Permo-Triassic onwards. Extension ceased in the East Shetland Basin during the early Cretaceous. Footwall uplift, and possibly flexural upwarping, were responsible for the delayed subsidence in the East Shetland Basin, which commenced in the mid-Cretaceous (Chapter 7). Rapid subsidence in the Faeroe Basin, during the mid-Cretaceous, may have been the result of minor seafloor spreading, or a westward shift in the locus of extension.

The final rift phase in the North-east Atlantic took place during the late Cretaceous and Palaeocene. Igneous activity was widespread, affecting east and west Greenland, and the Norwegian and British basins and platforms. The emplacement of volcanic centres preceded
Figure 8.1 Summary diagram of the features of the seismic stratigraphy of the Faeroe Basin and the East Shetland Basin. The diagrams essentially display a WNW-ESE cross-section.
Basin infill followed by "highstand/lowstand" shelf progradation.

Muddy shelf progradation

Initiation of fold development
Southern Faeroe Basin in compression

Confinement of submarine fan in synclinal low
Bottom water currents in deeper part of basin

Middle Eocene (Sequence E2)
Confinement of submarine fan in synclinal low
Rapid subsidence of basin preserves the E2 shelf sequence
Erosion of Sequence E2 from the ESP
Shelf bypass followed by rapid shelf progradation and snap of the East Shetland Platform

Late Middle Eocene (Sequence E3)
"Lowstand/Highstand" cycle

Upper Eocene (Sequence E4)
Continued fold development

Figure 8.1 (Continued)
voluminous sill and dyke intrusion throughout the north-western UKCS. Volcanism exploited the pre-existing crustal weaknesses, and so concentrated basaltic sills in the faulted basins. Volcanism was local to the rift zone, in the form of oceanward dipping reflector sequences, along Rockall Bank, Voring Plateau and the Greenland continental margin except in the vicinity of the Iceland Hotspot. In the proximity of the Iceland Hotspot, volcanism was dominated by dyke and sill intrusion over a far wider areas. Analogous present day mantle plumes at present do not show such widespread volcanism (Courtney & White, 1986) and the sill intrusion in the Faeroe Basin may indicate a minor amount of extension at this time. The absence of igneous material in the East Shetland Basin and the lack of significant faulting suggests that this basin did not undergo extension during the Palaeocene. Due to its proximity to the hotspot, the Shetland Platform was uplifted, and shed clastic material eastwards into the East Shetland Basin and the Central and Viking Grabens. Periods of fan development and shelf bypass in the Central Graben correlate loosely to periods of non-volcanism on the Faeroe Plateau, which was undergoing subsidence at this time (Hitchen & Ritchie, 1987; Mudge & Rashid, 1987).

To the south in North-west Europe, the Palaeocene was a time of Alpine-driven compression and basin inversion. The style of Palaeocene tectonism was controlled by the proximity of the pre-existing basin and platform structures in North-west Europe to the Atlantic or Alpine margins. The northernmost part of the continental plate, between Svalbard and Greenland was undergoing transtension.

After a prolonged history of rift tectonics in both the East Shetland Basin and the Faeroe Basin, pre-existing regions of extension were abandoned, at the end of the Palaeocene. Fully-fledged seafloor spreading took place to the north-west of the northern UKCS during the Lower Eocene. The initiation of seafloor spreading may have been caused by a relaxation of the ambient stress field over the whole of North-west Europe, at the cessation of a major phase of Alpine compression.
Subsequent Eocene sedimentation on the northern UKCS can be subdivided on the basis of the observed major seismic sequences observed therein. After the progradation of shelf-deltaic facies in both the East Shetland and Faeroe Basins, in the late Palaeocene. A relative sea level rise resulted in the submergence of the delta plains. Contemporaneous volcanicity resulted in the deposition of a tuffaceous claystone (Seismic sequence Eta) (Fig. 8.1). Tuff deposition in the uppermost Palaeocene-lowermost Eocene is a correlateable event across the western UKCS that has been recognised in drill sites from Rockall Plateau to the Norwegian Shelf. The event immediately pre-dates sea floor spreading and the rapid submergence of the Atlantic rift zone.

Whilst the Atlantic margin was subsiding, during the onset of seafloor spreading in the Lower Eocene, the Shetland Platform, southern Faeroe Basin and Faeroes Plateau were undergoing local uplift. This resulted in the formation of a submarine fan complex in the East Shetland Basin (Seismic sequence Elb). Canyons cut back into the surface of the platform, and a series of smaller fans fed the main Frigg fan system. Shallow marine sedimentation in the Faeroe Basin was derived from both the Shetland Platform and the Faeroes Plateau basalts. The northern Faeroe Basin, and the northern part of the East Shetland Basin, were undergoing rapid subsidence in response to thermal subsidence of the Norwegian Sea oceanic basin. The increased relief between the East Shetland Basin and the Shetland Platform resulted in slope instability and debris flows on the flanks of the basin. The cause of the uplift was probably residual thermal effects related to the Iceland Hotspot. By this mechanism platform regions were preferentially uplifted with respect to basins.

Subsequent deposition was controlled by the subsidence of the platform regions during the early Middle Eocene. Onlap of the East Shetland Platform occurred at this time (Seismic sequence Elc) (Fig. 8.1), and the submarine fan systems became inactive. Disconformities and hiatuses in the marginal succession show that eustatic sea
level changes were superimposed on the slow tectonic subsidence that occurred throughout the Middle Eocene. Rapid subsidence of the southern Faeroe Basin, along the north-western flank of Rona Ridge, occurred at this time. This rapid subsidence preserves an additional shelf wedge sequence in the Faeroe Basin that was probably eroded from the East Shetland Platform.

During the later stages of this relative sea level rise, faults in the Faeroe Basin were reactivated and a mild inversion occurred. The slow movement on the faults caused the overlying sedimentary layer to crumple, and form monoclinal, anticlinal and synclinal structures. The event appears to have affected the southern Faeroe Basin, Wyville Thomson Ridge and northern Rockall Trough. The southern part of Rona Ridge became the focus for shelf sedimentation. Whilst Eocene inversion structures are observed throughout North-west Europe, the folds in the southern Faeroe Basin can be related to changes in Atlantic sea floor spreading, rather than Alpine collision tectonics.

The late Middle Eocene (Seismic sequence E3) is characterised by the resumption of fan activity and rapid shelf progradation in the East Shetland Basin. In the southern Faeroe Basin the fold structures emerged to form palaeobathymetric ridges. These acted to confine a submarine complex at the base of Rona Ridge. The increased relief between Rona Ridge and the basin floor produced a fan complex fed by a delta ramp, rather than a canyon system. Shelf sediments were confined to Rona Ridge. This is believed to have been caused by a sea level fall, correlated to a similar event in the Haq et al. (1987) sea level curve. There is little evidence to suggest that the compression, associated with the folding, was responsible for this event - as might be expected for an intra-plate stress mechanism for sea level changes.

The Upper Eocene to Lower Oligocene is characterised by the progradation of shelf sands in both the East Shetland and Faeroe Basins (Sequence E4) (Fig. 8.1). Shelf currents oblique to the
direction of shelf progadation caused sand waves to slump, over the
shelf edge. Deposition of the slumped sands in the East Shetland
Basin was controlled by basinal currents, which developed in
response to renewed abyssal circulation in the Norwegian-Greenland
Sea. The circulation pattern was associated with the separation of
Greenland and Svalbard and the influx of Arctic waters into the
previously landlocked ocean basin. The continuation of the Iceland-
Faeroe Ridge as a bathymetric entity routed circulation into the
northern North Sea. Arctic bottom waters may have flowed in the
deeper parts of the Faeroe Basin.

The history summarised above indicates that sedimentation during the
Eocene was initially controlled by regional tectonism associated
with the involvement of the Iceland Hotspot in the Atlantic rift
zone. Subsequent deposition was controlled by more regional
fluctuations in sea level. Mild inversion did not influence platform
uplift, but served to confine basin sedimentation. Middle and Upper
Eocene sedimentation and tectonism can be correlated with changes in
sea floor spreading.

The results have an importance for passive margins and the Atlantic
marginal basins, in particular. Post-rift subsidence and
sedimentation is no longer perceived to be a quiescent phase, in
detail, but influenced by more fundamental and regionally important
tectonism, including the ambient stress field produced by the
processes acting at the plate margins. Such tectonism might be
observed in similar detailed studies on the Norwegian and Greenland
continental shelves. The Norwegian shelf lies close to the Voring
Plateau, a possible site of hotspot migration (Vink, 1984), and
intra-basinal highs (eg. the Trondelag Shelf; Caseili, 1987) may
prove analogous to the Faeroe Plateau or the Shetland Platform.
Eocene sedimentation in these areas may prove to be similarly
controlled by post-rift uplift and the initiation of rapid oceanic
circulation in the uppermost Eocene.
direction of shelf progradation caused sand waves to slump, over the shelf edge. Deposition of the slumped sands in the East Shetland Basin was controlled by basinal currents, which developed in response to renewed abyssal circulation in the Norwegian-Greenland Sea. The circulation pattern was associated with the separation of Greenland and Svalbard and the influx of Arctic waters into the previously landlocked ocean basin. The continuation of the Iceland-Faeroe Ridge as a bathymetric entity routed circulation into the northern North Sea. Arctic bottom waters may have flowed in the deeper parts of the Faeroe Basin.

The history summarised above indicates that sedimentation during the Eocene was initially controlled by regional tectonism associated with the involvement of the Iceland Hotspot in the the Atlantic rift zone. Subsequent deposition was controlled by more regional fluctuations in sea level. Mild inversion did not influence platform uplift, but served to confine basin sedimentation. Middle and Upper Eocene sedimentation and tectonism can be correlated with changes in sea floor spreading.

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319


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327


333


335


Appendix A

Clay mineralogy studies

A1 Controls on clay mineralogy

The use of clay mineralogy determination was decided upon to assess its usefulness of this aspect as a stratigraphic tool. A number of factors central to the composition and proportion of clays within a sedimentary sequence is controlled by:

1) Provenance. The types of rock in a sediment source region can be expected to exert some control on the types of clays. For instance, a higher content of kaolinite in recent sediments at high latitudes, is explained in terms of marine reworking of kaolinite-rich Mesozoic sediments (Darby, 1975; Bjorlykke & Elverhoi, 1975). Some clay mineralogies are dominated by reworking compositions (e.g. Bent, 1986), others only have some input (Karisson, 1978). Reworking and secondary reworking will increasingly complicate the clay mineralogy with time. However, provenances commonly are heterogeneous in composition and can not be the cause of compositions of one or two clay minerals.

2) Climate. One of the most important studies on clay mineralogy distribution (Biscaye, 1965) showed that different minerals dominate in different latitudes. Notably, kaolinite was found to be associated with low latitudes, and chlorite with high latitudes. This can be viewed in terms of climatic controls on weathering, where high latitudes are dominated by physical weathering and low latitudes by chemical processes. Thus, kaolinite, smectite, palygorskite and iron-oxides are associated with tropical to temperate climates (Singer, 1984). This approach was applied to the North Sea Tertiary by Karisson et al., 1978). The clay mineralogy showed a change from smectite-dominated clays (interpreted as the result of halmyrolitic weathering of volcanic glass) to chlorite. This was taken to represent a gross climatic change with time. Assumptions within a climatic interpretation of
clay mineralogy abound, and caution should be applied to any more
detailed interpretation in this form (Singer, 1984).

iii) Diagenesis. Clay minerals are sensitive to chemical
conditions both at surface, and at depth. A number of studies have
shown the changes in clay mineralogy associated with burial
diagenesis (eg. Weir et al., 1975; Dypvik, 1983). The most
important change, that of smectite to illite occurs over a
temperature range of 65-100°C (Dypvik, 1983), with burial depths
of 2-3 Km. This particular reaction will not be considered in the
BGS boreholes due to the shallow depths involved. However,
diagenesis is a process that occurs almost immediately after
deposition (Huang et al., 1975; Singer & Stoffers, 1980). This
early diagenesis are partly associated climatic conditions in
that these affect the chemical conditions of soil horizons and
groundwaters.

A2 Methods and criteria for recognition of clay minerals
Clay mineralogies were determined for boreholes 81/16 and 81/17,
for comparison with the already determined mineralogies in 80/03.
Smear-on-glass slides (Theisen & Harward, 1962) were prepared from
the <63μm obtained from particle size and pipette analyses
(Galehouse, 1971). Slides were subjected to solvation with
ethylene glycol at 60°C, and heated to 180°C. This allowed a semi-
quantitative analysis of the kaolinite, smectite and illite
proportions to be made (Table A1). Illite was determined from the
presence of the 10Å peak. Since this corresponds to all minerals
with a mica structure, including illite, there is some component
of micaceous material within the estimated proportions of illite.
This is especially relevant as mica flakes were observed in the
sediments. Kaolinite was recognised by the characteristic 7Å and
3.59Å peaks. The presence of chlorite was determined by slow
scanning (1 minute) of the 3.54 and 3.59Å peaks,
differentiating chlorite from kaolinite. However, most samples
showed only 2-3% of chlorite determined by this method, and
chlorite is discounted as a major mineral in the following
descriptions. A number of samples heated to 500°C, showed the loss of both the kaolinite and chlorite peaks, indicating that the low amounts of chlorite present were a poorly formed soil-derived type (Thorez, 1976).

Smectite was recognised, and semi-quantitatively measured from the expansion of the 12-15Å peak to 17Å on glycolation. A number of samples from 81/16 were K+-saturated, and the type of smectites diagnosed, following the interpretative procedure of Thorez (1976, p.21). Essentially, this is a measurement of the position of the glycolated, K+-saturated smectite (001) peak. Expansion to 17Å is taken to signify low-charge, effectively authigenic, smectite (Jonas, 1975). This implies a derivation from the weathering of framework silicates (Thorez, 1976). An expansion to 14Å observed is similarly concluded, as high-charge illite transformed to smectite produces a lower expansion to 12Å. Lack of expansion, signifying interlayer mixing (Thorez, 1976), was not observed in this analysis.

Glaucnite, although abundant in a number of the samples produces a series of diffraction peaks similar to illite. It is very difficult to distinguish the two minerals and so the illite proportions may include some glauconite. This is very low in all cases, because the 10Å peak is low magnitude. Furthermore Odin & Matter (1981) observe that glauconite, as a replacement mineral, is only statistically important in the 100-1000μm size range, outside the <63μm fraction discussed here.

A number of characteristics of the clay minerals can be determined from the shape of the diffraction peaks obtained, which give further indications of the importance of the three controls outlined above.

1) Illite. The width of illite is a measure of the “open character of illite”, or the extent to which K+ has been stripped from the illite lattice. Such characteristics have been defined in terms of
the similarities to the behaviour of other minerals (hence an illite classified as Iv describes an illite that behaves similarly to a vermiculite (Thorez, 1976)). The crystallinity of illite can be determined by analysing the width of the 10\AA{} peak. In this case the width was measured at half-height (Moriarty, 1977), on glycolated samples. This width, and the ratio of the illite (001) and (002) diffraction peaks (the 10\AA{}/5\AA{} intensity ratio) have been shown to be related to the intensity of weathering (Thorez, 1973).

ii) Smectite. A measure of the crystallinity of smectite is given by the valley/peak ratio of the 17\AA{} peak on glycolated samples (Biscaye, 1965) This parameter is stratigraphically and diagenetically controlled (Thorez, 1976). Similarly, the smectite/illite ratio for a sample is the ratio of the 17\AA{} peak (corresponding to the smectite (001) peak), to the 10\AA{} peak (illite (001)). The value of the ratio, coupled with the semi-quantitative proportion of smectite is taken to imply the extent to which interlayer mixing is significant. Such mixing has been shown to be provenance controlled (Sladen & Batten, 1984; Hurst, 1985).

The details of the clay mineralogies is discussed with each borehole. The results can be summarised as indicating that the dominant control on the clay mineralogy was the effect of climate and palaeoenvironment. Reworking of mineralogies is suggested for part of 81/16, where a later sequence, the Lower Main Sand Unit, has a clay mineralogy identical to the earlier Lignitic Sand Unit of 80/03.
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<td>2.00</td>
<td>5.110</td>
<td>0.52</td>
<td>14.5</td>
</tr>
<tr>
<td></td>
<td>111.90</td>
<td>4</td>
<td>38.5</td>
<td>15</td>
<td>39.5</td>
<td>1.60</td>
<td>6.000</td>
<td>0.54</td>
<td>13.4</td>
</tr>
<tr>
<td></td>
<td>119.98</td>
<td>4</td>
<td>24</td>
<td>15</td>
<td>22</td>
<td>2.50</td>
<td>3.400</td>
<td>0.88</td>
<td>15.4</td>
</tr>
<tr>
<td></td>
<td>128.93</td>
<td>3.5</td>
<td>23</td>
<td>10</td>
<td>21</td>
<td>-</td>
<td>2.500</td>
<td>0.69</td>
<td>17.2</td>
</tr>
<tr>
<td></td>
<td>135.70</td>
<td>0</td>
<td>13</td>
<td>21</td>
<td>16</td>
<td>-</td>
<td>-</td>
<td>0.72</td>
<td>0.0</td>
</tr>
</tbody>
</table>

TABLE A1. Semi-quantitative analysis of clay minerals, based on Griffin (1971)
Appendix B
Author citation appendix
Systematic palaeontology of taxa discussed in the text.

DINOFLAGELLATE CYSTS

Division PYRROPHYTA Pascher, 1914.
Class DINOPHYCEAE Fritsch, 1929.
Order PERIDIINIALES Haeckel, 1894.


Cordosphaeridium cantharellum (Brosius) Gocht, 1969.
Diphyes colligerum (Deflandre & Cookson) Cookson, 1965a.
D. ficusoides Islam, 1983b.
Dracodinium varielongitudum (Williams & Downie) Costa & Downie, 1979.

Etonicosysta ursulae (Morgenroth) Stover & Evitt, 1978
Glaiphyrocysta semitecta (Bujak in Bujak et al., 1980) Lentin &

Miospores

Pteridophyte spores:
Microfoveolatosporites pseudodentatus Krutzsch, 1959.
Angiosperm pollen:

*Brosipollis striatobrosus* (Krutzsch) Krutzsch, 1968.

**BENTHONIC FORAMINIFERA** (from King, 1983)

**DISCORBACEA:**
*Asterigerina guerichi* Franke, 1912

**BULIMINACEA:**
*Bulimina elongata* D'Orbigny, 1846
*Turrilina brevispira* Dam, 1944

**ROTALIACEA:**
*Elphidium inflatum* Reuss, 1861

**ORBITOIDAEA:**
*Planulina palmerae* Bellen, 1941
Appendix C

Computer program "SUBSIDE"

This Fortran program was developed to model the subsidence of the continental crust, founded on changes in the thermal structure of the lithosphere during extension. The model is based on a finite difference approach, in which points are defined in space, and assigned particular properties. An interval of time occurs before the properties of the grid points are recomputed, based on mathematical formulae. In this case the lithosphere is defined by a grid of 60x30 points all assigned an initial temperature value, and a constant value for conductivity.

Output consists of a water-loaded subsidence curve, at a defined point. Subsidence occurs by two mechanisms:-

i) Lithosphere extension - Necking of the crust produces a subsidence. The magnitude of this is controlled by the initial crustal and lithospheric parameters, and the magnitude and duration of the extension phase. The subsidence during extension is computed using standard formulae of Hellinger & Sclater (1983).

ii) Thermal - This is defined by changes in elevation above equilibrium, controlled by the temperature structure at any one time. This temperature structure is computed by the finite difference approach. For finite extension, heat flow is allowed to occur between the 0.5 Ma fault steps.

C1. Input parameters externally defined

Input is via a sequential file defined on Channel 8

<table>
<thead>
<tr>
<th>Definition</th>
<th>Variable</th>
<th>Format</th>
</tr>
</thead>
<tbody>
<tr>
<td>Title</td>
<td></td>
<td>16 Characters</td>
</tr>
<tr>
<td>No. of crustal blocks</td>
<td>N</td>
<td>I2</td>
</tr>
<tr>
<td>Point at which subsidence curve</td>
<td>POINT</td>
<td>F5.1</td>
</tr>
<tr>
<td>is to be computed in model</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Duration of first extensional phase</td>
<td>TRIFT</td>
<td>F5.1</td>
</tr>
</tbody>
</table>
Duration of modelling TEND F5.1
No. of subsequent extensional episodes NRIFT I2

C1.1 Two-dimensional extension
For each crustal block (I = 1 to N - max. 40) the following parameters are required:

<table>
<thead>
<tr>
<th>Definition</th>
<th>Variable</th>
<th>Format</th>
</tr>
</thead>
<tbody>
<tr>
<td>Position of left-hand margin of block</td>
<td>X(I)</td>
<td>F5.1</td>
</tr>
<tr>
<td>Crustal β</td>
<td>BC(I)</td>
<td>F5.2</td>
</tr>
<tr>
<td>Sub-crustal β</td>
<td>BSC(I)</td>
<td>F5.2</td>
</tr>
</tbody>
</table>

The lateral extent of the crustal model is defined by the difference between the first and last values of X(I) (Variable ΔX, below) - consequently, the last values of BC(I) and BSC(I) are redundant.

C1.2 Multiple extension
For each subsequent extensional phase (I = 1 to NRIFT - max 10) the following input is required:

<table>
<thead>
<tr>
<th>Definition</th>
<th>Variable</th>
<th>Format</th>
</tr>
</thead>
<tbody>
<tr>
<td>Start time of extensional phase from beginning of model</td>
<td>RTIME</td>
<td>F5.1</td>
</tr>
<tr>
<td>Duration of extensional phase</td>
<td>TNRIFT</td>
<td>F5.1</td>
</tr>
<tr>
<td>Number of crustal blocks</td>
<td>NRPTS</td>
<td>I2</td>
</tr>
</tbody>
</table>

Parameters for the two-dimensional model (J = 1 to NRPTS) are subsequently required as for the first extensional phase. These are defined as XRIFT(I,J), BCR(I,J) and BSCR(I,J).

C2 Parameters of two-dimensional model defined in program

C2.1 Initial parameters for lithosphere modelling

<table>
<thead>
<tr>
<th>Definition</th>
<th>Variable</th>
<th>Format</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crustal thickness</td>
<td>TC(I)</td>
<td>30-35 km</td>
</tr>
<tr>
<td>Lithosphere thickness</td>
<td>AL(J)</td>
<td>125 km</td>
</tr>
<tr>
<td>Horizontal no. of grid points</td>
<td></td>
<td>80 (1-80; L to R)</td>
</tr>
<tr>
<td>Vertical no. of grid points</td>
<td></td>
<td>45 (1-45; Top to base)</td>
</tr>
</tbody>
</table>
Spacing of horizontal points  $\Delta X/60$  
Spacing of vertical points  $125/30 = 4.1666 \text{ km}$
Initial temperature at each point  $T_{\text{T0}(x,z)} = 1333\text{HZ}/125$
Diffusivity at each point  $C_K = 19.04 \text{ km}^2/\text{Ma}^{-1.2}$
Number of finite difference steps  $NT = \text{TEND multiplied by 2}$

C2.2 Boundary conditions for lithosphere model

These values define the edge of the model and control the amount and direction of heat loss from the model.

<table>
<thead>
<tr>
<th>Left &amp; Right edges</th>
<th>Horizontal grid points 1-10 and 70-80</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Temperature is assigned to that at equivalent depth at horizontal grid point 11 or 69.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Bottom edge</th>
<th>Vertical grid point 31-45</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Temperature is fixed at 1333°C</td>
</tr>
<tr>
<td></td>
<td>Short-lived instantaneous increase in temperature above 1333°C is allowed.</td>
</tr>
</tbody>
</table>

| Upper edge          | Temperature is defined as 0°C, and heat flow is allowed across this boundary. |

C2.3 Finite difference computation

The finite difference part of the program consists of a matrix solution of the heat flow equation in two-dimensions. The Peaceman-Racheford solution upon which this is based performs an alternating horizontal and vertical Gaussian elimination. Consequently the finite difference part of the program begins with an alternation in the declaration of the shapes of the matrices. The solution is obtained by:

$$\text{NPTSxNPTIS tridiagonal matrix multiplied by } NPTSx1 = \text{NPTSx1 matrix performed } \text{MIN times.}$$

Values of the tridiagonal matrix are a function of spacing and diffusivity only. The diagonals are defined in three arrays: DL - Lower; D - Middle; DU - Upper. The known $NPTSx1$ matrix is a function of the spacing, temperature of adjacent points and diffusivity, and is defined by array $DXZ$. 

348
The solution to the above is performed by the NAG Library routine F04EAF. The solution is returned in the array DXZ.

Subsidence is obtained from the newly computed temperature structure using SUBROUTINE ELEV. This performs a summation (approximating an integration) of the difference between computed and equilibrium temperature structure. This becomes subsidence in the main body of the program through the array PSUB(ICOUNT) - where ICOUNT is 1 to TEND*2.

C2.4 Parameter during single-phase finite extension
Fault subsidence is quantified using the equations of Hellinger & Sclater (1983), relating extension, crustal and lithosphere thicknesses to subsidence. For finite rates of extension the input β-value is modified into smaller steps at 0.5 Ma interval such that the resultant crustal thinning is equal to that required. Consequently:

\[ EP = \frac{1}{(\text{Length of extensional phase})} \]

\[ BTC(I) \text{ or } BTSC(I) = (BC(I) \text{ or } BTC(I))^{EP} \]

After each fault phase the temperature structure is increased by the relevant β-value. Prior to each fault-thermal subsidence phase the lithosphere and crustal thicknesses are reset such that:

\[ TC(I) = \frac{TC}{BTC(I)} \]

and whole lithospheric extension is:

\[ BLT(I) = \frac{AL(I)}{TC(I)} = \frac{AL(I)(AL(I) - TC(I))}{BTSC(I) + TC(I)/BTC(I)} \]

\[ AL(I) = \frac{AL(I)}{BLT(I)} \]

No allowance is made - in the computation of fault subsidence - for the decay of heat flow, and consequent increasing \( AL(I) \) throughout the extension phase. The amount of subsidence computed enters the array PSUB(ICOUNT).

C2.5 Initial parameters during multiple phase of extension
Between phases of extension the finite difference model is allowed to run. At the start of a subsequent phase of extension, the thickness of the lithosphere is computed, assuming exponential decay of the 1333°C isotherm back to 125 km. This appears in the program as:\n
349
**Definition**

Time since last extension \( T_{THEM} \)

Proportion of time constant \( DECAY = -\frac{T_{THEM}}{62.8} \)

Lithosphere thickness \( AL(I) = AL(I) + (125 - AL(I))(1-e^{DECAY}) \)

The temperature structure is again altered in the manner of finite extension. The amount of subsidence computed enters the array \( PSUB(\text{ICOUNT}) \).

**C2.5 Output**

The program outputs a theoretical subsidence curve. Plotting routines are derived from ERCC GRAPHPACK, an Edinburgh-based graphics package. A glossary of the routines used is provided, in approximate order of usage.

**OPENGR (22)** - opens channel 22 for the reception of the plotting file.

**GRAREA** - defines the total plotting area.

**SCALGR** - defines the point of origin, X and Y scales and orientation. May be called several times to define different windows.

**AXISGR** - draws axis from specified point in specified direction with no. of ticks spaced as required.

**ANNOGR** - Sends the pen to a specific point in readiness for plotting, and defines size of subsequent graphics.

**DRNUMG, DRSTRG, DRSYM G** - draws number, text string of special symbol respectively.

**LINESG** - draws line between points defined by an array in X- and Y-axes.

**PLOTGR** - elementary plotting routine. parameters are pen up/down and dashed indicator, followed by destination and size of dash.

and space where not redundant.

**CHPNGR** - changes colour of pen.

**CLOSGR** - closes channel 22.
PROGRAM SUBSIDE

*** DATA INPUT FROM CHANNEL 8 ***

*** External routine: NAG subroutine F04EAF ***

IMPLICIT DOUBLE PRECISION (A-H,O-Z)

CHARACTER TITLE*16

DIMENSION T(1010),X(40),BT(90),XP(40),YP(45),XTEM(45),TXZ(90,45)
DIMENSION ST(90),BC(90),BSC(90),TT0(90,45)
DIMENSION BTC(90),BSC(90),TC(90),BLT(90),AL(90)
DIMENSION DXZ(90),DU(90),DL(90),D(90),MCH(5),Y(90)
DIMENSION TR(5),XR(5,2),BR(5),ITLE(4),SL(20)
DIMENSION PSTUB(1010),BEND(11),PEND(11),TREND(11),BSEND(11)

C Dimensions of arrays containing the values for the various rift phases.
DIMENSION RTIME(10),TNRIFT(20)
DIMENSION XRIFT(10,10),BCR(10,10),BSCR(10,10)
DIMENSION ICH1(3),ICH2(3),JCH(2)
DATA ICH1 /'Dept'/,ICH1(2)/'h (K'/,ICH1(3)/'m)'/
DATA ICH2 /'Time'/,ICH2(2)/' (Ma'/,ICH2(3)'/')'/

DATA JCH(1)/'TRIF'/,JCH(2)/'T= '/
DATA MCH(1)/'GEOL'/,MCH(2)/'OGY'/,MGR/' 1 Km'/

*** INPUT MODEL PARAMETERS ***

READ(8,18)TITLE

CALL ISTRNG(TITLE,ITLE,16)
WRITE(6,13)
WRITE(6,18)TITLE

READ(8,10)N,POINT,TRIFT,TEND,NPTFT
WRITE(6,19)POINT

DO 100 I=1,N
READ(8,12)X(I),BC(I),BSC(I)

IF (NRTIE .NE. 0) THEN
DO 110 I=1,NRTIE
READ(8,14)RTIME(I),TNRIFT(I),NRPTS
DO 110 J=1,NRPTS
READ(8,12)XRIFT(I,J),BCR(I,J),BSCR(I,J)
ENDIF

WRITE(6,16)

110 CONTINUE

ELSE
ENDIF

10 FORMAT(A16)

CALL ISTRNG(TITLE,ITLE,16)
WRITE(6,13)
WRITE(6,18)TITLE

13 FORMAT( ' *** LATSUB MODELLING PROGRAM *** ')
19 FORMAT(F5.1)

DO 100 I=1,N
READ(8,12)X(I),BC(I),BSC(I)

100 CONTINUE

11 FORMAT(I2,3(F5.1,1X),I2)
12 FORMAT(F5.1,1X,F5.3)
14 FORMAT(2(F5.1,1X),I2)
16 FORMAT(' READING IN COMPLETED')

EP=1/(TRIFT*2.0)
C DEFINE GRID FOR 125KM LITHOSPHERE
HZ=4.166666
C GIVES A COLUMN OF 30 POINTS IN 125KM LITHOSPHERE
HX=(X(N)-X(1))/60.
HZ2=HZ**2
HZ2=HZ**2

351
DO 200 J=10,70
AL(J)=103.7
TC(J)=29.0
XP(J)=HX*(J-10)
PMIN=XP(J)-HX/2
PMA=XP(J)+HX/2
IF(POINT.GT.PMIN.AND.POINT.LE.PMA) IP=J
200 CONTINUE
DO 300 K=1,45
300 YP(K)=(K-1)*HZ
DO 400 J=10,70
DO 400 M1,N
IF(XP(J).GE.X(M)) THEN
BTC(J)=BC(M)**EP
BTSC(J)=BSC(M)**EP
BLT(J)=AL(J)/((AL(J)-TC(J))/BTSC(J)+TC(J)/BTC(J))
CALL RIFT(BTC(J),BTSC(J),TC(J),AL(J),YP,XTEM)
IF(J.EQ.IP) THEN
C Computation of the initial instantaneous subsidence
C associated with non-uniform lithospheric extension.
CALL ELEV(XTEM,YP,HZ,ET1)
BEND(1)=BC(M)
BSEND(1)=BSC(M)
TREND(1)=TRIFT
IB=1
BL=BLT(IP)
GC=1-1/BTC(IP)
GSC=1-1/BTSC(IP)
CF=0.335*GSC
IF(GC.GE.CF) THEN
DENS=2.1544
ELSE
DENS=3.1844
ENDIF
PSUB(1)=((0.53*TC(IP)*(1-0.022*TC(IP)/AL(IP))
+0.0728*TC(IP))*(1-1/BTC(IP))-0.0728*(AL(IP)-TC(IP)))
/((1-1/BLT(IP)))/DENS
IF(PSUB(1).GT.0) PSUB(1)=0.0
ELSE
ENDIF
ELSE
ENDIF
DO 600 K=1,45
600 TTO(J,K)=XTEM(K)
400 CONTINUE
DO 575 J=10,70
AL(J)=AL(J)/BLT(J)
TC(J)=TC(J)/BTC(J)
575 CONTINUE
WRITE(6,15)
15 FORMAT( ' INITIAL MODEL COMPUTED ')
NT=TEND*2
IPN=0
ICOUNT=1
T(ICOUNT)=0.0
IF(T(ICOUNT).EQ.TRIFT)THEN
  PEND(IB)=PSUB(1)
  TREND(IB)=T(ICOUNT)
ELSE
ENDIF
NARR=1
CK=25.12
WRITE(6,87)T(1),PSUB(1),SI
C==========================================================================
C FINITE DIFFERENCE MODELLING AT EACH TIME INTERVAL
C==========================================================================
DO 700 M=1,NT
  ICOUNT=ICOUNT+1
  IFAILO
  IPLOTO
  TP=0.5
  T(ICOUNT)=TP*M
C COMPUTATION OF EQN MATRIX
IF(NARR.EQ.1) THEN
  N1=2
  N2=1
  NIN=44
  NPTS=80
  D1=HX2
ELSE
  N1=2
  N2=1
  NIN=79
  NPTS=44
  D1=HZ2
ENDIF
C BOUNDARY CONDITIONS
DO 750 J=1,80
  DO 750 K=1,45
    TTO(J,1)=0.0
    IF(K.GE.31)TTO(J,K)=1333.0
    IF(J.LE.10)TTO(J,K)=TTO(11,K)
    IF(J.GE.70)TTO(J,K)=TTO(69,K)
  CONTINUE
750 C COMPUTATION OF VALUES IN HORIZ AT NEXT TIME FROM VERT NOW
IF(NARR.EQ.1) THEN
  IND=1
  J1=J-1
  J2=J+1
  TRAN=TTO(K,J1)+TTO(K,J2)
  DXZ(K)=TTO(K,J)*(CD-2./HZ2)+(TRAN)/HZ2
ELSE
  C ALTERNATIVE STEP
  IND=2
C==========================================================================
C==========================================================================
363
J2 = J + 1  
J1 = J - 1  
TRAN = T0(J, K) * (CD - 2 / HX2)  
DXZ(K) = TRAN + (T0(J1, K) + T0(J2, K)) / HX2  
ENDIF

900 CONTINUE
CALL F04EAF(NPTS, D, DU, DL, DXZ, IFAIL)
DO 800 K = 1, NPTS
IF (IND.EQ.2) TXZ(J, K) = DXZ(K)
IF (IND.EQ.1) TXZ(K, J) = DXZ(K)
800 CONTINUE
NTR = NARR
IF (NTR.EQ.1) NARR = 2
IF (NTR.EQ.2) NARR = 1

C==============================================================================
C End of finite difference solution
C==============================================================================
C Computation of subsidence at the point specified, for each
time the FD equation is computed.
TXZ(IP, I) = 0.000
DO 1100 K = 2, 31
1100 XTEM(K) = TXZ(IP, K)
CALL ELEV(XTEM, YP, HZ, ET)
DS = ET - ET1
PSUB(ICOUNT) = PSUB(ICOUNT - 1) + 1.448 * DS
ET1 = ET
DO 1200 J = 10, 70
DO 1200 K = 2, 31
1200 TTO(J, K) = TXZ(J, K)
C
C Option for finite rifting
C
IF (T(ICOUNT).LE.TRIFT) THEN
C Compute fault subsidence at the specified point.
C
C Reset all the crustal thickness values for the effect of the last rift
DO 725 J = 10, 70
725 BLT(J) = AL(J) / (TC(J) / BTC(J) + (AL(J) - TC(J)) / BTSC(J))
BL = BLT(IP)
SI = ((0.53 * TC(IP) * (1 - 0.022 * TC(IP) / AL(IP)) - 0.0728 * TC(IP))
* (1 - 1 / BTC(IP)) - 0.0728 * (AL(IP) - TC(IP)) * (1 - 1 / BL)) / 2.1544
ICOUNT = ICOUNT + 1
PSUB(ICOUNT) = PSUB(ICOUNT - 1) - SI
T(ICOUNT) = TP * M
IF (T(ICOUNT).EQ.TRIFT) THEN
PEND(IB) = PSUB(ICOUNT)
TREND(IB) = T(ICOUNT)
IB = IB + 1
ELSE
ENDIF
WRITE(6, 87) T(ICOUNT), PSUB(ICOUNT), SI
87 FORMAT (F5.1, 3(4X, F10.6))
C Reset temperature values due to extension event.
C Geotherms are raised by a factor B (crustal and sub-crustal
354
C where appropriate). Hence all temperatures are
C raised by the same factor at a fixed point.

DO 825 J=10,70
  AL(J)=AL(J)/BLT(J)
  TC(J)=TC(J)/BTC(J)
END DO 825

DO 825 K1,31
  IF(YP(K) .LE. TC(J)) THEN
    TTO(J,K) = TTO(J,K) * BTC(J)
  ELSE
    TTO(J,K) = 1333.0 + TTO(J,K) * BTSC(J) - 1333.0 * BTSC(J) / BLT(J)
  ENDIF
  IF(TTO(J,K) .GT. 1333) TTO(J,K) = 1333.0
END DO 825

DO 1300 NR=1,NRIFT
  IF(T(ICOtJNT).GE.RTIME(NR).AND.T(ICOtJNT).LE.(RTIME(NR)+TNRIFT(NR))) THEN
    IF(T(ICOtJNT).EQ.RTIME(NR)) THEN
      IF(NR.GT.1) TTHERM = T(ICOtJNT) - RTIME(NR-1)
      IF(NR.EQ.1) TTHERM = T(ICOtJNT) - TRIFT
      DECAY = -1.0 * TTHERM / 62.8
    ELSE
      DECAY = -0.5 / 62.8
    ENDIF
  DO 1325 J=10,70
    AL(J) = AL(J) + (125.0 - AL(J)) * (1 - EXP(DECAY)) / (2.0 * TNRIFT(NR))
    EP = 1 / (2.0 * TNRIFT(NR))
  DO 1400 J=10,70
  DO 1400 MR=1,NRPTS
    IF(XP(J) .GE. XRIFT(NR,MR)) THEN
      IF(J.EQ.IP) THEN
        BEND(IB) = BCR(NR, MR)
        BSEND(IB) = BSCR(NR, MR)
      ELSE
        BEND(IB) = BCR(NR, MR)
        BSEND(IB) = BSCR(NR, MR)
      ENDIF
      BTC(J) = BCR(NR, MR) ** EP
      BTSC(J) = BSCR(NR, MR) ** EP
      BLT(J) = AL(J) / ((AL(J) - TC(J)) / BTSC(J) + TC(J) / BTC(J))
      TC(J) = TC(J) / BTC(J)
  DO 1450 K1,31
    IF(YP(K) .LE. TC(J)) THEN
      TTO(J,K) = TTO(J,K) * BTC(J)
    ELSE
      TTO(J,K) = 1333.0 + TTO(J,K) * BTSC(J) - 1333.0 * BTSC(J) / BLT(J)
    ENDIF
    IF(TTO(J,K) .GT. 1333) TTO(J,K) = 1333.0
  END DO 1450
END DO 1325

DO 1300 NR=1,NRIFT
  IF(T(ICOtJNT).EQ.RTIME(NR)) THEN
    TTHERM = T(ICOtJNT) - RTIME(NR-1)
  ELSE
    TTHERM = T(ICOtJNT) - TRIFT
  DECAY = -1.0 * TTHERM / 62.8
  ELSE
    DECAY = -0.5 / 62.8
  ENDIF
END DO 1300

DO 1325 J=10,70
  AL(J) = AL(J) + (125.0 - AL(J)) * (1 - EXP(DECAY)) / (2.0 * TNRIFT(NR))
  EP = 1 / (2.0 * TNRIFT(NR))
END DO 1325

CALL ELEV(XTEM, YP, HZ, ET1)
ELSE
  CALL ELEV(XTEM, YP, HZ, ET1)
END IF

If the time is past the rifting period then look for
a subsequent rift phase.

DO 1300 NR=1,NRIFT
  IF(T(ICOtJNT).GE.RTIME(NR).AND.T(ICOtJNT).LE.(RTIME(NR)+TNRIFT(NR))) THEN
    IF(T(ICOtJNT).EQ.RTIME(NR)) THEN
      IF(NR.GT.1) TTHERM = T(ICOtJNT) - RTIME(NR-1)
      IF(NR.EQ.1) TTHERM = T(ICOtJNT) - TRIFT
      DECAY = -1.0 * TTHERM / 62.8
    ELSE
      DECAY = -0.5 / 62.8
    ENDIF
  DO 1325 J=10,70
    AL(J) = AL(J) + (125.0 - AL(J)) * (1 - EXP(DECAY)) / (2.0 * TNRIFT(NR))
    EP = 1 / (2.0 * TNRIFT(NR))
  DO 1400 J=10,70
  DO 1400 MR=1,NRPTS
    IF(XP(J) .GE. XRIFT(NR,MR)) THEN
      IF(J.EQ.IP) THEN
        BEND(IB) = BCR(NR, MR)
        BSEND(IB) = BSCR(NR, MR)
      ELSE
        BEND(IB) = BCR(NR, MR)
        BSEND(IB) = BSCR(NR, MR)
      ENDIF
      BTC(J) = BCR(NR, MR) ** EP
      BTSC(J) = BSCR(NR, MR) ** EP
      BLT(J) = AL(J) / ((AL(J) - TC(J)) / BTSC(J) + TC(J) / BTC(J))
      TC(J) = TC(J) / BTC(J)
  DO 1450 K1,31
    IF(YP(K).LE.TC(J)) THEN
      TTO(J,K) = TTO(J,K) * BTC(J)
    ELSE
      TTO(J,K) = 1333.0 + TTO(J,K) * BTSC(J) - 1333.0 * BTSC(J) / BLT(J)
    ENDIF
    IF(TTO(J,K) .GT. 1333) TTO(J,K) = 1333.0
  END DO 1450
END DO 1325

DO 1300 NR=1,NRIFT
  IF(T(ICOtJNT).EQ.RTIME(NR)) THEN
    TTHERM = T(ICOtJNT) - RTIME(NR-1)
  ELSE
    TTHERM = T(ICOtJNT) - TRIFT
  DECAY = -1.0 * TTHERM / 62.8
  ELSE
    DECAY = -0.5 / 62.8
  ENDIF
END DO 1300

DO 1325 J=10,70
  AL(J) = AL(J) + (125.0 - AL(J)) * (1 - EXP(DECAY)) / (2.0 * TNRIFT(NR))
  EP = 1 / (2.0 * TNRIFT(NR))
END DO 1325

CALL ELEV(XTEM, YP, HZ, ET1)
ELSE
  CALL ELEV(XTEM, YP, HZ, ET1)
END IF
CONTINUE
ELSE
ENDIF
CONTINUE
CALL ELEV(XTEM,YP,HZ,ET1)
IF(T(ICOUNT).EQ.(RTIME(NR)+TNRIFT(NR)))THEN
PEND(IB)=PSUB(ICOUNT)
TREND(IB)=T(ICOUNT)
IB=IB+1
ELSE
ENDIF
SI=((0.53*TC(IP)*(1-0.022*TC(IP)/AL(IP))-_0.0728*TC(IP))*(1-1/BTC(IP))_0.0728*(AL(IP)-
TC(IP))*(1-1/BLT(IP)))/2.1544
DO 1425 J=10,70
AL(J)=AL(J)/BLT(J)
TC(J)=TC(J)/BTC(J)
CONTINUE
ICOUNT=ICOUNT+1
PSUB(ICOUNT)=PSUB(ICOUNT-1)-SI
T(ICOUNT)=TP*M
WRITE(6,87)T(ICOUNT),PSUB(ICOUNT),SI
ELSE
ENDIF
CONTINUE
C If the time does not correspond to a rift period then do not reset.
C
ENDIF
CONTINUE
C *** PLOTTING ROUTINES *** ERCC GRAPHPACK ***
C
CALL OPENGR(22)
CALL DRSTRG(IDENT,7)
CALL GRAREA(0.00,0.00,20.00,10.0,1)
CALL SCALGR(0.00,0.00,1.00,1.00,0.00)
CALL BOX(0.00,0.00,15.50,10.00)
C Plot the model parameters above the curve with pointer to position
CALL SCALGR(2.00,8.50,0.04,0.50,0.00)
CALL BOX(0.00,1.00,X(N),2.00)
DO 1250 I=1,N
IL=I+1
IF (I.LT.N)THEN
XMEAN=(X(I)+X(I1))/2-5.0
CALL ANNOGR(XMEAN,1.26D0,2.0D0,0.0D0)
BETA=BC(I)
IF(NRIFT.NE.0)THEN
DO 1275 IN=1,NRIFT
WRITE(6,42)IN,I,BCR(IN,I),N
42 FORMAT(10X,I2,2X,I2,F5.2,1X,I2)
BETA=BETA*BCR(IN,I)
ELSE
ENDIF
1275 CONTINUE
ELSE
ENDIF
CALL DRNUMG(BETA,1,2)
ELSE
ENDIF
CALL ANNOGR(X(I)-8.D0,2.1D0,2.D0,0.D0)
CALL DRNUMG(X(I),3,1)
CALL PLOTGR(1,X(I),2.D0,1.D0,1.D0)
CALL PLOTGR(2,X(I),1.D0,1.D0,1.D0)
1250 CONTINUE
CALL ANNOGR(XP(IP)-3.D0,2.1D0,3.D0,0.D0)
CALL DRNUMG(11)
CALL SCALGR(11.D0,5.05D0,0.04D0,1.6D0,0.D0)
CALL AXISGR(-250.D0,-3.D0,2,1.D0,3)
CALL AXISGR(-250.D0,0.D0,1,10.D0,25)
CALL AXISGR(0.D0,0.D0,4,1.D0,3)
C
DO 500 I=1,4
DP=(I-1)*1.0
DP1=-1.*(DP+0.05)
CALL ANNOGR(-265.D0,DP1,2.D0,0.D0)
CALL DRNUMG(DP,1,1)
500 CONTINUE
C
CALL ANNOGR(-266.D0,-1.6D0,2.D0,0.D0)
CALL DRSTRG(ICH1,10)
DO 650 I=1,14
AXP=(I-2)*20.+10
IF(I.EQ.1)AXP=0.
AXP1=-1.*(AXP+8.)
CALL ANNOGR(AXP1,0.1D0,2.D0,0.D0)
CALL DRNUMG(AXP,3,0)
650 CONTINUE
C Plot time-subsidence curve
DO 1225 I=1,ICOUNT
TEMP=T(I)
T(I)=TEMP-TEND
1225 CONTINUE
CALL CHPNGR(4)
CALL PLOTGR(1,T(I),0.D0,1.D0,1.D0)
CALL PLOTGR(2,T(I),PSUB(1),1.D0,1.D0)
CALL LINESG(T,PSUB,1,ICOUNT,1.D0,0.D0,0,1.D0)
CALL ANNOGR(T(I)-4.D0,0.4D0,10.D0,0.D0)
CALL DRSYM(73)
DO 1500 I=1,IB-1
TREND(I)=TREND(I)-TEND
IF(I.EQ.1) THEN
TINC=TREND(I)-0.6*TRIFT
ELSE
TINC=TREND(I)-0.6*TNRIFT(I-1)
ENDIF
CALL ANNOGR(TREND(I)-4.D0,0.4D0,10.D0,0.D0)
CALL DRSYM(73)
CALL ANNOGR(TINC,1.2D0,2.D0,0.D0)
CALL DRSYM(164)
CALL DRSYM(61)
CALL DRNUMG(BEND(I),1,2)
CALL ANNOGR(TINC,1.05D0,2.D0,0.D0)
CALL DRSYM(162)
CALL DRSYM(61)
CALL DRNUMG(BSEND(I),1,2)
CONTINUE
1500 CONTINUE
DO 1530 I=1,IB-2
RTIME(I)=RTIME(I)-TEND
CALL ANN0GR(RTIME(I)-4.D0,0.4D0,10.D0,0.D0)
CALL DRSYM(73)
1530 CONTINUE
CALL CLOSGR
END
C
C
C
C
SUBROUTINE RIFT(B,BS,TCP,A,YPR,XTER)
C Computes the temperature structure at the first extension for
C inhomogeneous extension of the crust and sub-crustal lithosphere.
C Input: B-values for crustal and sub-crustal lithosphere
C Crustal and lithosphere thickness
C Depth to grid points
C Output: Temperature for specific depth
IMPLICIT DOUBLE PRECISION (A-H,O-Z)
DIMENSION YPR(46),XTER(46)
T0=1333.0
BL=A/(A-TCP/BS+TCP/B)
ATC=TCP/B
ATH=A/BL
DO 350 K=1,45
IF(YPR(K).LE.ATC) XTER(K)=T0*B*YPR(K)/A
IF(YPR(K).GT.ATC.AND.YPR(K).LE.ATH)XTER(K)=
$$T0*(1+BS*(YPR(K)/A_1/BL))$$
IF(YPR(K).GT.ATH)XTER(K)=T0
350 CONTINUE
RETURN
END
C
C
C
C
SUBROUTINE ELEV(TEMP,ZP,DZ,EL)
C Compute the elevation of the upper surface of the lithosphere
C from the value at infinity
C Input: Array of temperature values at depth: x is constant
C Vertical distance between grid points
C Output: Elevation at point x
IMPLICIT DOUBLE PRECISION (A-H,O-Z)
DIMENSION TEMP(40),ZP(40)
DATA A/125.0/,T0/1333.0/
SUMTZ=0.0
DO 150 J=1,31
TZ=TEMP(J)-ZP(J)*T0/A
T21=TZ*DZ
SUMTZ=SUMTZ+T21
150 CONTINUE
EL=SUMTZ*3.28E-05
RETURN
END
C
C
C
SUBROUTINE BOX(XMIN, YMIN, XMAX, YMAX)
C Plotting routines for a box around the defined window.
IMPLICIT DOUBLE PRECISION (A-H,O-Z)
CALL PLOTGR(1, XMIN, YMIN, 1.D0, 1.D0)
CALL PLOTGR(2, XMAX, YMIN, 1.D0, 1.D0)
CALL PLOTGR(2, XMAX, YMAX, 1.D0, 1.D0)
CALL PLOTGR(2, XMIN, YMAX, 1.D0, 1.D0)
CALL PLOTGR(2, XMIN, YMIN, 1.D0, 1.D0)
RETURN
END
Appendix D

Aspects of Palaeogene hydrocarbon prospectivity

The Tertiary has been an important hydrocarbon play in the northern North Sea, since the earliest exploration in the area. The discovery of the Forties oilfield, in October 1970, focussed exploration on the Palaeocene sands of submarine fan origin. Subsequent Palaeocene discoveries occurred in the Maureen, Montrose, Andrew, Cod, Balmoral and Cyrus fields in the central North Sea (Fig. D1). Heavy oil has been discovered in the Moray Group shelf deltaics, including the Bressay field. The Eocene has proved of commercial interest in the Gannet field in the Central Graben and the Frigg gas field in the south Viking Graben. Interest has recently been revived in the Eocene succession, with the discovery of the Alba and Forth oilfields. Within the East Shetland Basin, the Palaeocene-Eocene has been shown little interest and exploration activity has concentrated on the Jurassic.

D1 Source rock for Palaeogene hydrocarbon occurrences

Because of their immaturity and lack of organic vitrinite, the late Palaeocene shales do not represent a potential source rock for Palaeocene-Eocene reservoirs (Barnard & Cooper, 1981). However, oil and gas shows in the Palaeogene succession are related to areas of good Jurassic source rock maturation. Consequently, areas such as the Beryl Embayment are potential areas for migration of oil and gas into the Palaeogene succession. Throughout the northern North Sea Jurassic reservoirs are overpressured (Harris & Fowler, 1987), and release of hydrocarbons into the Palaeogene only occurs where the overpressure seal of mid-Cretaceous mudstones is breached (Fig. D2). Consequently, mechanisms for migration into the Palaeogene appear to be related to underlying fault movements, resulting from differential compaction, localised stresses (Pegrum & Ljones, 1984) and halokinesis (Armstrong et al., 1987). The latter mechanism is only important south of 59° 30' N. The areas of breaching are likely to include the crests of
Hydrocarbon occurrences in the Palaeogene

- Graben structure
- Oil field
- Gas field
- Hydrocarbon shows
- Platform areas

Figure D1. Map of Palaeogene hydrocarbon discoveries and shows in the North Sea.
Clastic reservoir types

Hydrocarbon migration paths

East Shetland Platform

Lateral migration along channels and canyons into shelf sands

E1b-c shelf-deltaics

E4 muddy shelf seal

Rare generally sealed by E1c/E2 claystones

E4 slope sands

E3 channel ?sands

Fracture of seal by subsequent fault movement

Mid-Cretaceous seal

Fracture of seal by subsequent fault movement into Lwr. Eocene

Present sea level

Oligocene-Recent

Frigg/Balder-type play

Bressay-type play

Bruce-type play

Brent/Heather-type Jurassic play

Figure D2. Prospective migration paths and reservoirs in the Eocene of the northern North Sea.
tilted fault blocks, the flanks of the East Shetland Basin and South Viking Graben, which show signs of reactivation during the Palaeogene. The flanks and Jurassic-rich areas of the North Sea grabens therefore represent a region of potential Palaeogene oil finds.

The West Shetland Basin is much more problematical, in that the area has to date shown little in the way of recoverable reserves, let alone Palaeocene-Eocene oil finds. The highest organic content occurs in the source rocks of the Lower Cretaceous (Ryazanian and Volgian), supplemented by the Kimmeridgian. Within the basin these sequences are thin and immature, with significant hydrocarbon generation localised against the Shetland Spine Fault (Bailey et al., 1987). However, potential source areas still exist in the deeper parts of the Faeroe Basin. Gas-blanking is indeed observed on single channel seismic records in the southern Faeroe Basin. This suggests either biogenic gas formation or migration of mature hydrocarbons close to the seabed. The top and bottom of the Faeroe Basin oil window is approximately 3660-4115m and 4420-4575m. According to Bailey et al. (1987), oil generation from the Volgian would have occurred from 50Ma. to 8Ma. ago, and was replaced by dry gas production 27 Ma. ago. Consequently, the formation of traps and reservoirs in the Eocene would be synchronous to the major episode of hydrocarbon maturation and migration, and does not preclude this play on the flanks of the Faeroe Basin.

D2 Prospective Palaeogene reservoirs

Good reservoir qualities are high porosity and permeability. In general, porosity is controlled by sorting and permeability by grain size. Consequently, well sorted sands represent the most important clastic reservoir rocks. Subsequent cementation has an important bearing on the reservoir potential of clastics. In the Palaeogene, the optimum reservoir properties are associated with the shelf-deltaic sequences. Such regions may include beach ridge and dune complexes, behind the delta front, such as the Bressay field. During the mid to
late Palaeocene, submarine fan deposition was widespread, and the best reservoirs were developed in the mid to lower fan lobe and channel setting (Sarg & Skjold, 1984). This is best developed in the Central Graben. In the northern North Sea, the distribution of this facies is more widespread, defining a broad sheet (Rochow, 1981; Morton, 1982; Mudge & Bliss, 1983). Consequently, there is no absence of sand reservoir in the Palaeocene northern North Sea.

The Eocene consists of discrete horizons and locations of sands. The Frigg fan complex is located in the centre of the basin, but was fed by a composite series of sub-fans located on the flank of the graben and the edge of the East Shetland Platform. Distribution of these, as discussed in Chapter 3, was controlled by basin topography (eg. edge of platform; Moray Group shelf edge; end of canyon). These sub-fans have oil shows, such as Bruce. As discussed in Chapter 3 canyons on the East Shetland Platform indicate a transport direction towards the south-east, suggesting an anticlockwise rotation of the transport direction towards the north as the sands enter the graben axis. Further north, the seismic stratigraphy shows the development of debris flow deposits along the western flank of the graben. The rest of the basin appears barren of fan sands, unless smaller, localised lobes existed.

Subsequent shelf progradation - during the deposition of the Middle Eocene sequences E1c-E2 - was silt-dominated, and it is considered that the shelf-slope sequences are unlikely to contain significant reservoir types. The marginal sands, discussed in Chapter 2, represent the best reservoir rocks for this sequence. The Lower Sand Unit is occasionally silty, and is kaolinite infiltrated, providing poor reservoir qualities at the subcrop limit. However, it is possible that the delta-front deposits predicted to lie in the middle of quadrant 8 represent the best facies for hydrocarbon migration. Porosity is also enhanced by the flow of meteoric groundwaters - and consequent heavy
mineral dissolution - through the unit. This also has an important
bearing on the biodegradation of hydrocarbons. Deposition of the Main
Sand Unit suggests the possibility of localised sandy sandwaves
amongst the sand-silt laminated sand waves predicted for the cross-
stratified seismic unit.

Sands observed in sequences E3 and E4 are associated with localised
submarine fans, and base of slope deposits. These have a reasonable
potential for reservoir rocks.

In the Faeroe Basin, the only sands observed in the Palaeocene are the
Bryozoan Sands of Hitchen & Ritchie (1987) and the upper Palaeocene
shelf deltaics of Mudge & Rashid (1987). Eocene deltaics are inferred
in the southern Faeroe Basin. It is predicted that these are a mixture
of sands dominated by reworked basaltic lithics, and interbedded
lignites and clays. This sequence overlies the region of dry gas and
condensate generation in the Faeroe Basin. Eocene shelf sands are
observed on Rona Ridge and the West Shetland Basin, although the
reservoir potential of the downslope facies is limited to the southern
Faeroe Basin. Here, periodic formation of large scale submarine fans
are observed on the seismics, and it is probable that sands are
concentrated in the confinement structure discussed in Chapter 4.

D3 Entrapment
It has been shown that oil migration has occurred into the Palaeogene
succession, and that there are sediments with reasonable reservoir
potential. However, the biggest problem with the Palaeogene succession
is the formation of trap structures. Only one field shows evidence of
structural deformation (15/9 Gamma; Pegrum & Ljones, 1984), although
halokinesis has dominated the deposition and hydrocarbon accumulation
in Gannet (Armstrong et al., 1987). With the absence of major
faulting, most of the traps are likely to be stratigraphic pinch-outs
of reservoirs with overlying shales acting as the seal (Fig. D2). For
example, the Balder and Frigg fields are trapped by the closure of the mounded upper surface of submarine fan lobes (Sarg & Skjold, 1982; Heritier et al., 1979). Smaller scale traps result from pinch-out of individual channel sands (e.g., Cod field; Kessler et al., 1980). Stratigraphic traps in the Moray Group shelf-deltaics are formed by the pinch-out of beach and barrier ridges.

Prospective Eocene sands are likely to be similarly trapped, and the mid-fan setting of the East Shetland Basin is the best region for this. Migration paths are likely to fill pinch-outs close to the point of breakage of the mid-Cretaceous seal. After these are filled to spill-point, migration is likely into progressively higher stratigraphic levels, and laterally into basin margin settings. Vertical migration is dependent on the fracture of the claystone seal (sequences E1c and E2). Middle Eocene reservoirs may therefore contain oil, and the channelled facies of sequence E3 (Chapter 3) the best region for trapping. Similarly, the Middle Eocene shelf sediments may have been charged by hydrocarbons. However, the absent of faults extending into the sand probably precludes major hydrocarbon accumulation in the E3 and E4 sequences. Small, very subtle stratigraphic traps may be located in the shelf sands on the East Shetland Platform, and charged by lateral migration of hydrocarbons. This is dependent on there existing a linked series of sands on the edge of the platform. Shallow buried reservoirs are susceptible to bacterial attack and oxidation. Consequently, the biodegradation of the hydrocarbons and the lack of hydrostatic pressure would hinder the development of any such find.

Stratigraphic and structural traps exist in the Faeroe Basin, related to the Middle to late Eocene inversion. Anticlines may be prospective, although the structures along the axis of the basin are unlikely to contain good reservoir rocks. Folds in the vicinity of the Judd Fault (Chapter 4) may represent a good, structural trap for the Elb deltaic
sands. Later confinement by this structure may have allowed stratigraphic trapping in Middle Eocene submarine fans.

D4 Conclusion
The best prospectivity in the Palaeogene succession remains the submarine fan sands, concentrated in the Palaeocene and Lower Eocene, but interspersed throughout the Eocene (Fig. D2). Hydrocarbons charged these reservoirs by breaking of the mid-Cretaceous seal to underlying, overpressured Jurassic reservoirs. Lateral migration of hydrocarbons along channel fills may have allowed oil to reach the East Shetland Platform. Subsequent lack of a trap mechanism, and biodegradation of the hydrocarbons will have removed much of the commercial reserves in this area. It is possible that some subtle traps still exist on the East Shetland Platform.
Addendum

Figure 7.6 A map showing areas of missing Lower Cretaceous sediments, and their relation to major structures. Map taken from Chesher & Turner (1987) and Chesher & Stoker (1986).
Contours in metres
Velocity = 2100 m/s
O Commercial well tied to seismics

British Geological Survey
Mineral Earth Sciences
Map projection OSGB 36, International Spheroid European Datum
Scale 1:200 000
Date: 02/03/92

E3 ISOPACH
E3 Subcrop Limit

O Commercial well tied to seismics