Timing and controls of structural inversion in the NE Atlantic Margin

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To my mother
Declaration

This thesis has been written by myself and the work reported within has been conducted by myself, unless otherwise stated. The work has not been submitted in any previous application for a degree.
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Abstract

Passive continental margins are generally thought to be characterised by tectonic quiescence as they experienced gentle thermal subsidence following the extensional events that originally formed them. Analysis of newly acquired and existing 2-D seismic data from the Rockall Plateau to the Faroe Shelf, however, has demonstrated that the NE Atlantic Margin is the site of significant active deformation. Seismic data have revealed the presence of numerous compression-related Cenozoic folds, such as the Alpin, Ymir Ridge and Wyville-Thomson Ridge anticlines. The presence, timing and nature of these structures have provided new insights into the controls and effects of contractional deformation in the region. Gravity models support the presence of low-density sediments in the core of folds. This is consistent with folds developed due to the inversion of sedimentary basins. The spatial extent of the deformation could thus reflect the differing underlying basin morphologies. The mapping and dating of angular unconformities that define the folding suggest that the growth of these compressional features occurred in five main phases – Thanetian, late Ypresian, late Lutetian, Late Eocene (C30) and Early Oligocene, each of which appear to have been driven by regional events affecting the NE Atlantic Margin. The late Ypresian, Late Eocene (C30) and Early Oligocene events correlate with the timings of hot-spot influenced ridge-push based on the ages of V-shaped ridges. The late Lutetian event is tentatively also ascribed to hot-spot influenced ridge-push. Mohr-Coulomb circle plots, however, reveal that the forces exerted by hotspot-influenced ridge push appear to be insufficient to result in the reactivation of faults in the underlying Mesozoic basins. Alpine and Pyrenean Orogenies may have, thus, produced the additional force needed for the inversion of these basins. The Thanetian phase, just prior to Atlantic Ocean spreading, would suggest compression due to depth-dependent stretching and associated asthenospheric upwelling. The regional studies make clear that compression can have a profound effect on seabed bathymetry and consequent bottom-water current activity. Bottom-water currents have directly formed the early Late Oligocene, late Early Miocene (C20), Late Miocene – Early Pliocene, and early Late Pliocene (C10) unconformities. These unconformities have been interpreted to result from both global and regional
changes in deep water circulation. The entry of bottom-water currents from the Faroe-Shetland Channel into the Rockall Trough is restricted by the Wyville-Ymir Ridge Complex, and takes place via the syncline (Auðhumla Basin) between the two ridges. The growth of this compressional feature is now thought to have controlled the distribution of bottom-water currents in the Rockall Trough, and the resulting unconformity formation and sedimentation therein.
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Symbols and Abbreviations

Map Features
- anticline axis
- syncline axis
- reverse fault
- normal fault
- continent-ocean boundary
- compressional structure
- bathymetric high
- bathymetric low
- bathymetric contour

Seismic Features

Stratigraphic Surfaces
- C10 (late Early Pliocene)
- Late Miocene - Early Pliocene
- C20 (late Early Miocene)
- early Late Oligocene
- Early Oligocene
- C30 (Late Eocene)
- late Lutetian
- late Ypresian
- top basalt
- Thanetian

Abbreviation
- TWT Two-way time
1.0 Introduction

1.1 Rationale
Passive continental margins have traditionally been thought to be relatively tectonically quiescent during post-rift thermal subsidence following syn-rift extension which formed them. However, many passive continental margins around the world appear not to simply undergo passive subsidence (McKenzie, 1978), but show evidence of structural inversion whereby former sedimentary basins are uplifted and the normal faults that originally defined and transected them are reactivated under compression. Indeed, continental margins off Angola (Hudec and Jackson, 2002), East USA (Whitjack et al., 1998), Brazil (Cobbold et al., 2001; Meisling et al., 2001), NW Australia (Karner and Driscoll, 1999), and NW Europe (Boldreel and Andersen, 1993), have all undergone structural inversion. Inversion, it seems, has played a major role in unconformity formation, which was once attributed to sea level changes and bottom-water current activity in continental margins.

Inversion has long been identified in areas throughout the NE Atlantic Margin. These areas include the Rockall-Faroe area (Boldreel and Andersen, 1993, 1994, 1998), Hatton Bank (Hitchen, 2004; Johnson et al., 2005), the Faroe-Shetland Basin (Ritchie et al., 2003) and along the Norwegian Margin in the Vøring Basin (Blystad et al., 1995; Lundin and Doré, 2002). Inversion in the NE Atlantic Margin has been variably attributed by these authors to normal ridge push (Boldreel and Andersen, 1993; Doré and Lundin, 1996), hotspot-influenced ridge push (Lundin and Doré, 2002), Alpine compression (Doré and Lundin, 1996; Brekke, 2000; Roberts, 1989; Vågnes et al., 1998) and Pyrenean compression (Knott et al., 1993).

The main purpose of this study is to identify compressional structures and to better constrain the timing of their formation using an extensive seismic dataset coupled with well data in a strategic area of the NE Atlantic. The study area comprises the northern part of the Rockall Plateau to the south western part of Faroe Shelf (Figs. 1.1 and 1.2) and for the purposes of this study has been named the Rockall-Faroe area. Boldreel and Andersen (1993) were the first to attribute domal features in this area to compression. However, these authors lacked the extensive seismic dataset and well data which is available in this study. Thus, the timing of formation of these
Fig. 1.1. Regional map of the NE Atlantic showing the location of the Rockall-Faroe study area. The position of the continent-ocean boundary of the NE Atlantic Margin and the Charlie Gibbs Fracture Zone is based on Kimbell et al. (2005). The position of ocean ridges and transform faults, north of latitude 60°N, is based on Lundin and Doré (2002). Abbreviations: EJMFZ = East Jan Mayen Fracture Zone, FFZ = Faroe Fracture Zone, CGFZ = Charlie Gibbs Fracture Zone, FSB = Faroe-Shetland Basin, MB = Møre Basin. Bathymetric map (contours in metres) is courtesy of GEBCO Digital Atlas published by the British Oceanographic Data Centre on behalf of IOC and IHO (2003).

KEY

- Rockall-Faroe study area
- continent-ocean boundary
- active ocean ridge
- extinct ocean ridge
- major transform fault
Fig. 1.2. Location of seismic, gravity and well/borehole data used in the Rockall-Faroe study area. Seismic data was provided by BGS, CGG Veritas, Fugro Multi Client Services, WesternGeco and GEUS. Well and borehole data were used to constrain the ages of horizons which were identified and mapped on seismic lines. Gravity data determined the potential nature of underlying structures.
structures was not well constrained. In addition, the underlying basement structures were not discerned, as in this study, through gravity models. Previous authors were also unable to test the effectiveness of driving mechanisms using models as was done in this study.

The Rockall-Faroe area (Fig. 1.1) is an ideal area for such a study due to the availability of 2-D seismic, well and gravity data (Fig. 1.2). Inversion features (anticlines, reverse faults and associated unconformities) were mapped and interpreted from seismic data to determine the orientation and distribution of these features. The use of wells enabled unconformities to be dated and the timing of compressional events to be established.

Indeed, studying the timing, nature and orientation of inversion features in the NE Atlantic Margin has shed some light on the inversion mechanisms responsible for their formation.

1.2 Continental Margins

The characteristics of continental margins play a vital role in understanding the tectonic and non-tectonic post-rift development of continental margins. Indeed, the characteristics of the Rockall-Faroe area, in the NE Atlantic Margin, can influence the development of compressional structures and unconformities present within the Rockall-Faroe area. Thus, an understanding of the development and the structure of continental margins can facilitate a better understanding of the nature and origin of compressional structures.

1.2.1 Development

Continental margins are formed as a result of continental break-up, which is a consequence of continued rifting of the crust and lithosphere as a result of extension. Different extensional mechanisms have been proposed for rifting which results in the thinning of the crust and lithosphere:
**Pure Shear**

Pure shear extension (McKenzie, 1978) involves the uniform and symmetrical extension in the lithosphere (Fig. 1.3). The brittle upper crust deforms by normal faulting while the ductile lower crust and lithosphere thins by ductile stretching.

![Fig. 1.3. Schematic diagram of continental crust undergoing pure shear extension (McKenzie, 1978). Note the normal faulting in the upper crust and the ductile stretching occurring in the lower crust.](image)

**Simple Shear**

Simple shear extension (Wernicke, 1985) is associated with detachment faulting (Fig. 1.4). The fault cuts the upper and lower crust and lithosphere, stopping at the asthenosphere.

![Fig. 1.4. Schematic diagram of continental crust undergoing simple shear extension (Wernicke, 1985). Note the detachment fault displacing the crust and lithosphere.](image)

**Pure Simple Shear**

Bott (1992) proposed a model which incorporates both mechanisms (Fig. 1.5). Asymmetrical faulting of the upper crust is separated from a region of symmetrical ductile thinning of the lower crust by a detachment surface. The lithosphere also deforms by ductile stretching.
Fig. 1.5. Schematic diagram of continental crust undergoing both pure shear and simple shear extension (Bott, 1992). Note the normal faults occurring in the upper crust which taper to a detachment surface between the upper and lower crust.

During rifting, hot asthenosphere wells up to replace thinned lithosphere (McKenzie, 1978). The asthenosphere, at a lower pressure, undergoes decompression melting to form magma. If rifting continues until the crust and lithosphere breaks (continental break-up), the decompressional melts cool to form oceanic crust in the space created by continental break-up (McKenzie and Bickle, 1988).

During extension, pre-rift sediments are cut by normal faults and accommodation space is created for the deposition of syn-rift basin sediments (Fig. 1.6). After break-up, post-rift sediments are deposited and the continental margin no longer undergoes extension as all the extension is taken up by ocean spreading. The Bay of Biscay continental margin shows the general structure of a passive continental margin (Fig. 1.7).
Cessation of Extension
Extension ends and normal faults become inactive. Syn-rift sediment is no longer accommodated in the hanging wall of the normal fault. Post-rift sediments are deposited at this time.

Extension
As extension continues pre-rift sediment and upper crustal rock are rotated and accommodation space created for the deposition of relatively thick syn-rift sediment in the hanging wall of the normal fault.

Initiation of Extension
Extension produces normal faults which cut pre-rift sediments (sediments deposited prior to rifting).

McKenzie (1978) proposed a model for the subsidence of non-volcanic continental margins during and after rifting (Fig. 1.8). He assumed rapid stretching of the continental lithosphere which produces an initial subsidence. Non-volcanic margins initially subside by at least 2 km when stretching occurs (White, 1988). As the lithosphere thins, there is passive upwelling of hot asthenosphere (McKenzie, 1978). The asthenosphere cools and this produces subsidence. The subsequent thermal subsidence is exponentially decreasing as the lithosphere regains its original thickness by the cooling of the asthenosphere (McKenzie, 1978).
Volcanic margins, however, undergo uplift at the time of rifting (Fig. 1.8; White, 1988; Skogseid and Eldholm, 1988). Uplift in volcanic margins at the time of rifting is thought to be the result of:

1. Dynamic uplift due to the presence of a mantle hotspot convecting upwards (Courtney and White, 1986; White et al., 1987; White and McKenzie, 1989). This convective flow produces broad swells observed around hotspots (White and McKenzie, 1989);

2. New igneous material added to the crust and subsequently the crust is elevated to maintain isostatic equilibrium (White et al., 1987; White, 1988; White and McKenzie, 1989);

3. The reduction in density of the residual asthenosphere as partial melt rises and is removed (White and McKenzie, 1989).

Skogseid and Eldholm (1988) have modelled subsidence in a volcanic margin. The model shows uplift during rifting and prior to sea-floor spreading (Fig. 1.8). The subsidence of a normal margin is also presented based on McKenzie (1978). The extrusion and intrusion of igneous rocks and the elevation of the North Atlantic volcanic margin during rifting, was followed by subsidence at normal rates as the margin thermally equilibrated (Hyndman and Roberts, 1987; cited in White, 1988).
1.2.2 Continental Margin Sediment

Sediment type and distribution differs through time and space in continental margins. The sediment is related to the different stages of continental margin development. Boillot (1981) has defined four stages of continental margin development and their subsequent sediment deposition:

**Continental Rift Stage**

This stage takes place during continental rifting prior to sea floor spreading. Sediments deposited in the continental rift stage include continental sediment, such as fluvial and lacustrine deposits, and lagoonal sediment, such as evaporites. Intercalated lava flows, a product of rifting, may also be extruded.
Red Sea Stage
Sea-floor spreading produces a young ocean basin with continental shelves on opposing margins of the basin. Coral reefs may grow on the shelf edges. Sapropels and black shales are typically formed due to limited circulating ocean waters resulting in deep waters (1 km depth) remaining anoxic. Hydrothermal activity, driven by heat of the young oceanic crust, may result in the precipitation of sulphides within the sediment.

Narrow Ocean Stage
Submergence of the rift flanks results in detrital sediments transported into the marine basin. Turbidites, intercalated with hemipelagic sediments, start depositing in the deeper basin.

Atlantic Stage
The ocean basin is deep and wide enough to facilitate oceanic circulation. This results in the transportation of finely laminated sediment (contourites) by deep contour currents, in addition to the deposition of turbidites. At this stage, thermal subsidence wanes and the continental margin is prone to the effects of marine transgressions and regressions due to eustatic changes in sea level.

1.2.3 Volcanic Margins
'The only difference between volcanic and non-volcanic margins is that the potential temperature of the asthenosphere is up to 100-150°C higher beneath the former' (White et al., 1987a). This higher-than-normal asthenospheric temperature results in anomalously high volumes of magmatism in volcanic margins.

Asthenospheric material upwells during the thinning of the lithosphere. If stretching continues above values for $\beta$ of above five, new oceanic crust is formed (White, 1988; White and McKenzie, 1989). As the asthenosphere upwells during the thinning of the lithosphere, it decompresses, generating partial melt. The hotter the asthenosphere the more partial melt is generated (White, 1988). The melt generated rises rapidly, until it is either extruded as basaltic flows or intruded into or beneath the crust (White and McKenzie, 1989).
Volcanic margins are characterized by:

1. Uplift during rifting resulting in relatively small syn-rift basins. This also results in less overall subsidence of volcanic margins compared to non-volcanic continental margins (Skogseid and Eldholm, 1988). In the North Atlantic, for example, there is anomalously shallow bathymetry throughout the region (White, 1988);

2. Relatively rapid volcanism. The bulk of magmas in flood basalt provinces of Karoo and Deccan, for example, were ejected within 2 - 3 Ma and this was followed by smaller volumes of magma over a further period of 5 - 10 Ma (Campbell and Griffiths, 1990);

3. Wedges of seaward-dipping reflectors mainly constituting basalt lava flows, which are located at the transition between thinned continental and oceanic crusts (Larsen et al., 1994);

4. Mainly basaltic lava flows, pyroclastic deposits and igneous complexes (Skogseid, 2001);

5. Underplated igneous material at the base of the lower crust (White et al., 1987a; 1987b; White, 1988);

6. Relatively thick oceanic crust (10 - 12 km) (Skogseid, 2001).

Three zones can be represented on volcanic continental margins (Fig. 1.9). An inner zone (zone 1) where the igneous rocks are represented by dykes, sills and plateau basalts. A central zone (zone 2) contains seaward-dipping reflector surfaces and an outer zone (zone 3) of normal oceanic crust (Larsen et al., 1994). Such a structure is evident on the Namibia volcanic margin (Fig. 1.10).
1.2.4 Volcanism in the NE Atlantic Margin

In the North Atlantic region there was an initial phase of rift magmatism at 62 - 58 Ma followed by more intense magmatism during the initiation of sea-floor spreading at 56 - 52.5 Ma (Saunders et al., 1997: cited in Fitton et al., 1997). During the initial stages of rifting in the North Atlantic, there was massive extrusion of volcanic rocks and underplating of igneous rock to the lower crust in just 2 - 3 My (White et al.,
According to Skogseid and Eldholm (1988) extensive subaerial volcanism in the North Atlantic during continental break-up lasted for about 3 Ma. Magmatism in the North Atlantic occurs over an area with a diameter of 2000 km (White, 1988) and takes the form of:

**Underplated Material**

New igneous material, up to 15 km thick, was added to the lower crust of the NE Atlantic Margin. This underplating is evident for the Vøring Plateau (White et al., 1987, White, 1988) and the Hatton Bank Margin (Fig. 1.11; White et al., 1987a; 1987b; White, 1988). Under the lower crust of the Hatton Bank the new igneous material has a well-defined velocity of 7.3 – 7.4 km/s (White et al., 1987b). These magmatic bodies underlying the lower crust are assumed to be gabbroic in composition and thermal models suggest that the underplated material gradually accreted from several smaller bodies (Clift and Turner, 1998).

![Velocity model over the North Atlantic](image)

**Basalt lava flows**

Much of the NE Atlantic Margin is covered with a veneer of basalt lava flows. The basalt lava flows were extruded near sea level (Maresh and White, 2005). In the NE Rockall Trough 1.2 km of Paleocene extrusive basalts were recorded (Maresh and White, 2005) while the maximum proven composite thickness of basalt lava flows drilled in the Faroe Island is 6.6 km (Ellis et al., 2002; Passey, 2004). Each basalt flow is separated with a tuff or breccia layers (Maresh and White, 2005). The distribution of basalt lava flows is shown in Fig. 1.12.
Seaward-dipping reflectors

In the NE Atlantic Margin seaward-dipping reflectors (Fig. 1.9) were first recognized as lava flows by Hinz (1981), later confirmed by direct sampling (Roberts et al., 1984a; ODP Leg 104 Scientific Party 1986; Parson et al., 1986; cited in White, 1988). The extrusive basalt which comprises the seaward-dipping reflectors is 3 - 6 km thick occurs as a zone 50 - 100 km wide along the North Atlantic Margins (White, 1988). The extent of seaward-dipping reflectors located on the edges of the continental margin is shown in Fig. 1.12.

Fig. 1.12. North Atlantic Igneous Province showing early Tertiary onshore and offshore igneous rocks. Modified from Saunders et al. (1997), Fig. 1.
Igneous Centres

Igneous centres take the form of large scale features of the sea bed called seamounts or are buried by Cenozoic sediment. Seamounts include the Sigmundur, Rosemary Bank, Anton Dohrn and Hebrides Terrace in the Rockall Trough and Mammal in the Hatton-Rockall Basin. Buried igneous centres include the Darwin Igneous Centre in the NE Rockall Trough.

According to O’ Connor et al. (2000), the intrusion of igneous centres such as the Rosemary Bank, Anton Dohrn and Hebrides seamounts began in the late Cretaceous, and continued from early Paleocene to Mid Eocene in at least four discrete phases – 62, 52, 47 and 42 Ma. Late Upper Cretaceous ages of the Anton Dohrn and Rosemary Bank seamounts was also derived by Tate et al. (1999). The Darwin Igneous Centre is dated as Paleocene in age (Tate et al., 1999). Basalt and phonotephrite samples have been drilled on the Rosemary Bank suggesting sourcing from both asthenospheric mantle and subcontinental lithospheric mantle (Hitchen et al., 1997). Twin circular positive gravity anomalies over George Bligh Bank are interpreted as igneous intrusions (Edwards, 2002).

Skogseid and Eldholm (1988) presented a model for the formation of a volcanic-type passive margin based on the outer Vøring margin (Fig. 1.13). In the late stages of rifting there is dyke injection and subaerial extrusion of andesite flows. The thinned lithosphere breaks and there is the emplacement of basalt lava flows and seaward-dipping reflectors. The intrusive centre subsides below sea level, 3 My after break-up as ocean spreading ensues (Skogseid and Eldholm, 1988). In the Vøring margin sills and dykes are present in Mesozoic sediments and a lower crustal velocity body (7.2 - 7.5 km/s) has been mapped beneath the outer margin at the base of the crust (Skogseid, 1994). Flood basalts and seaward-dipping reflectors are also present.
Fig. 1.13. Schematic model for the development of a volcanic-type passive margin based on the Vøring Margin (redrawn from Skogseid and Eldholm, 1987). Abbreviation: SDR = seaward dipping reflectors.

1.3 Iceland Plume

The massive volcanism that produced the North Atlantic Igneous Province has been attributed to anomalously hot asthenosphere caused by the presence of a mantle plume (White, 1988, White and McKenzie, 1989, Cox, 1989; Griffiths and Campbell, 1990). Rockall Trough volcanism of the Rosemary Bank, Anton Dohrn and the Hebrides Terrace seamounts has been attributed to the pulsing of hot Iceland plume material (O’Connor et al., 2000).
Ascending convective thermal plumes in the mantle have long been believed to form uplifted ridges, swells and volcanic islands (Morgan, 1971). The Cape Verde Rise represents a site of uplift and volcanism over at least the last 20 My (Courtney and White, 1986). The uplift and volcanism has been attributed to the presence of a mantle plume. Courtney and White (1986) used heat flow, geoid and bathymetric relief values to model the hotspot mechanism over the Cape Verde Rise. An axisymmetric convection model (Fig. 1.14) offered a good fit to the observed data and accounts for the broad heat flow anomaly and the uplift and volcanism observed on the Cape Verde Rise. The model suggests that the Cape Verde Rise swell has been generated by an ascending thermal plume in the underlying mantle. The swell can be attributed to dynamic uplift caused by the ascending thermal plume (Courtney and White, 1986).

Characteristics of hotspots based on the Cape Verde model (Fig. 1.14; Courtney and White, 1986; cited in White and McKenzie, 1989):

1. Narrow (150 - 200 km wide) central plume of rising asthenosphere with abnormally high temperatures
2. A mushroom-shaped head forms on top of the rising plume as the convected material flattens and laterally spreads under the plate away from the centre of the plume. The mushroom-shaped head of asthenosphere has temperatures 100 – 200 °C higher than the ambient asthenosphere.
3. The mushroom-shaped head of the plume has a diameter of 1000 - 2000 km.
4. The dynamic uplift as a result of the convecting mantle reaches a maximum of 1000 - 2000 m.

White (1988) assumes that a similar mushroom-shaped hotspot was developed beneath the NW European-Greenland plate prior to rifting. This hotspot provides a good explanation for the extensive volcanism present in the North Atlantic margins. Voluminous partial melts would have been produced where the rift ran across the 2000 km diameter head of the mushroom hotspot, where the temperature of the asthenosphere is higher than normal (White, 1988; White and McKenzie, 1989). In areas distal to the ‘mushroom’, such as on the Biscay margin, normal asthenosphere temperature still existed and thus no volcanism occurred here during continental rifting.
Fig. 1.14. Axisymmetric convection model showing temperature variations across a mantle plume beneath the Cape Verde swell (White, 1988 based on Courtney and White, 1986). Temperatures are in degrees Celsius and are in relation to the mean asthenosphere value. The plume has a relatively narrow conduit and a broader mushroom-shaped head of asthenosphere material which spreads laterally away from the plume centre beneath the overlying plate.

The assumption that a plume consists of a narrow vertical pipe which supplies hot mantle to the asthenosphere, feeding a radial flow in the asthenosphere away from the plume, beneath a moving plate is held by Sleep (1990).

Griffiths and Campbell (1990) stated that an ascending plume consists of a large buoyant head followed by a narrow vertical conduit. They have used experimental data and calculations to show that a plume from the core-mantle boundary should have a diameter of the order of 1000 km and would enlarge to ~ 2000 km as the plume head flattens beneath the lithosphere. This theoretical prediction fits well with the extent of continental flood volcanism (2000 - 2500 km diameter) in the Deccan and Karoo flood basalt provinces (Campbell and Griffiths, 1990). The ~ 2000 km diameter of the plume head also matches the extent of volcanism in the North Atlantic (White, 1989).

The 10 - 15 km thick igneous material which underplated the lower crust on the NE Atlantic margin requires a 100 - 150 °C rise of the temperature in the asthenosphere at the time of rifting (White, 1988). This was derived from curves of melt thickness as a function of stretching factor calculated from an empirical relation between melt fraction, pressure and temperature (McKenzie and Bickle, 1988; cited by White, 1988). The 100 - 150 °C increase in asthenospheric temperature is in good
agreement with the asthenospheric temperature increase postulated in the Cape Verde hotspot convection model (Fig. 1.14).

According to White and McKenzie (1989) the anomalously thick Iceland-Faroes Ridge was formed directly above the central plume. This view is also supported by Larsen et al. (1994). The thickness of the Iceland-Faroes Ridge reaches 35 km (Bott and Gunnarson, 1980; cited in White and McKenzie, 1989). However, Lawver and Müller, 1994, have constructed Iceland hotspot tracks to show that the initiation of the mantle plume predates the opening of the North Atlantic. Thus, the massive early Tertiary volcanism along the North Atlantic margins is the result of rifting in existing thinned crust independent of the arrival of the Iceland plume (Lawver and Müller, 1994).

Smallwood and White (2002) proposed that the Iceland plume took the form of two hot sheets of asthenospheric melt at the time of continental break-up. A sheet of mantle was located along the zone of rifting between Greenland and NW Europe at the time of break-up. The plume then developed into an axisymmetric shape with a hot sheet of mantle extending from the Davis Strait on the western margin of Greenland to the west coast of Scotland and the east coast of Northern Ireland. It is believed that the central core of the Iceland plume generated the anomalously thick crust (30 - 40 km) of the Greenland-Iceland-Faroes Ridge (Smallwood and White, 2002) where the two hot sheets intersected. Like the Cape Verde model, the surrounding 2000 km diameter region received the lateral outflow from the plume which caused regional elevation (Smallwood and White, 2002).

Whilst it is recognized that plumes originate from the mantle, the specific source of plumes has been debated (Fig. 1.15). Plumes may originate from:

1. The 670 km discontinuity between the upper and lower mantle (White and McKenzie, 1989)
2. The D” layer at the core-mantle boundary (Griffiths and Campbell, 1990)
3. Convective instability at the 670 discontinuity as plume from the D'' layer stalls at this thermal boundary (Tackley et al., 1993: cited in Fitton et al., 1997).

The presence of a plume offers a good explanation for the voluminous magma produced at volcanic margins. It also accounts for the isotopic and trace-element signatures of plume in basalts from the Reykjanes Ridge as far south as 62° N (Hart et al., 1973; Schilling, 1973; cited in Larsen et al., 1994). However, depleted MORB-like basalts present on the Hatton Bank and the southeast Greenland margin show no evidence of a plume isotopic signature (Larsen et al., 1994). The MORB asthenosphere could have been heated at the periphery of the plume head to give magma of this affinity (Larsen et al., 1994).

![Schematic diagrams of three plume models](image)

Fig. 1.15. Schematic diagrams of three plume models (Fitton et al., 1997). A plume sourced from an instability in the boundary layer between the upper and lower mantle (Model 1) will be composed of upper mantle with entrained lower-mantle material in its centre. A plume originating at the core-mantle boundary (Model 2) will be composed mostly of lower mantle. A lower mantle plume which stops at the 670 km discontinuity (Model 3) will produce a hot lower mantle that may instigate instabilities in the upper mantle to produce a plume as in Model 1.

Mutter et al. (1988) (cited in Larsen et al., 1994) presents a model which attributes enhanced melt generation to locally convecting asthenosphere driven by the strong thermal contrast between the normal-temperature asthenosphere and the cooler
continental lithosphere. However, this model fails to account for the plume signatures present in basalts in the NE Atlantic margin. The Icelandic mantle plume was situated under East Greenland in the Early Tertiary during continental break-up (Morgan, 1971; White, 1988; White and McKenzie, 1989).

1.3.1 V-shaped Ridges

V-shaped ridges along the Reykjanes Ridge, south of Iceland, have been attributed to the lateral flow of pulses of hot asthenosphere from the Iceland plume (Vogt, 1971). V-shaped ridges on the Reykjanes Ridge record the passage of hotter asthenosphere from the Iceland mantle plume (O’Connor et al., 2000). These ridges are delineated by free-air gravity anomalies across the Reykjanes Ridge (Fig. 1.16) with the crest of each ridge corresponding to gravity highs (Jones et al., 2002). The V-shaped ridges are symmetrical about the spreading axis and they progressively cross younger isochrons as they converge towards the south (Vogt, 1971) propagating up to 1000 km from Iceland (Jones et al., 2002). There is crustal thickness variation of ~2 km between the ridges and intervening troughs (Jones et al., 2002). The ridges are probably formed as a result of pulses of anomalously hot asthenosphere flowing down the spreading axis (White and Lovell, 1997; O’Connor et al., 2000). Excess volumes of melt, which form the ridges, are produced when this anomalously hot asthenosphere decompresses beneath the spreading centre (White et al., 1995). The V-shaped ridges can be explained by an upward convecting plume under Iceland and lateral flow of hotter-than-normal mantle away from Iceland channelled beneath the Reykjanes Ridge (White et al., 1995). However, modelling has shown that the observed topography south of Iceland along the Reykjanes Ridge is the result of radially symmetric lateral flow away from the plume centre (Iceland) rather than being channelled preferentially along the ridge axis (Ito et al., 1996; Ito, 2001).
Fig. 1.16. Free-air gravity anomalies over the Reykjanes Ridge. V-shaped ridges are delineated verging southward on the Reykjanes Ridge. Diagram from Ito et al. (2001).
1.4 Aims and Objectives

Syn-rift basins are formed by the process of extension during the development of continental margins. These basins are bounded by normal faults which have the potential of reversing their direction of slip upon inversion (Fig. 1.17). The aim of the study is to examine the compressional features in the area by integrating seismic, well/borehole and gravity data in order to understand the Cenozoic structural evolution of the Rockall-Faroe area. The study also considers the effects of potential inversion mechanisms. The presence of an active Iceland plume in the Cenozoic offers the opportunity to study the potential effects of hotspot-influenced ridge push, which is unique as many continental margins are limited to the effects of only normal ridge push. In addition to hotspot-influenced ridge push, the effects of the Iceland Plateau body force and the more distal Alpine and Pyrenean orogenies can also be studied.

![Diagram of two stages of fault activity](image)

**a.** Normal faults become inactive as extension ceases.

**b.** Compression results in the inversion of pre-existing normal faults.

Fig. 1.17. Schematic diagram of the inversion of a normal fault in a half-graben. The normal fault is formed as a result of extension resulting in the deposition of syn-rift sediments (a), re-activate with a reversal in slip direction, during compression, to result in the uplift and folding of pre-rift, syn-rift and post-rift sediments (b). Diagram based on Williams et al. (1989).
The project aims to:

1. Date unconformities marked by onlap or erosional truncation using well data. This is crucial in determining the timings of both tectonic (compressional) and non-tectonic events.

2. Use seismic data to map unconformities in the Rockall-Faroe area. The mapped unconformities can be used to construct isochrons (time maps of horizons) and isochores (time difference maps). The isochron maps highlight compressional features while the isochore maps reveal sediment distribution patterns.

3. Determine the origin of unconformities by their nature and extent. Unconformities in continental margins can be the result of compression, sea-level changes or bottom-water current activity. The study examines the nature of the unconformities in the Rockall-Faroe area and aims to distinguish compressional phases from phases of tectonic inactivity.

4. Understand the role of compression in influencing sediment distribution.

5. Use gravity data to analyse the potential sub-basalt structures in the Rockall-Faroe area to understand the nature of compressional structures and the potential controls on inversion.

6. Study the nature and timing of the various driving mechanisms in order to link them to the observed compression in the Rockall-Faroe area. The driving mechanisms explored are ridge push (normal and hotspot-influenced), Alpine and Pyrenean compression, depth-dependent stretching and the Iceland Plateau body force.

7. Determine quantitatively the forces required to invert normal faults using Mohr-Coulomb diagrams and finite-element analysis. This was done to better appreciate the relative contributions and effectiveness of certain driving mechanisms.

8. Establish whether there is any link between compression in the Rockall-Faroe area and regional events based on the timing and effectiveness of compressional mechanisms.
1.5 Dataset and Methodology

Seismic, well/borehole and gravity data, in addition to finite-element models have been used to understand the structure and nature of structures in the Rockall-Faroe area and the driving mechanisms which brought them about. The following is an appraisal of each of the tools used to understand the timings and controls of structural inversion in the Rockall-Faroe area.

1.5.1 Seismic Data

Use has been made of 40,000 km of 2-D seismic data (line-length) acquired over the past twenty years extending from the Rockall Plateau to the Faroe Shelf (Fig. 1.2). Seismic data were provided by BGS, Fugro Multi Client Services, CGG Veritas and GEUS. The seismic data differs in quality and depth of penetration (Table 1.1). Landmark software was used to image seismic data.

Table 1.1. Seismic data used in this study.

<table>
<thead>
<tr>
<th>Operator</th>
<th>Prefix</th>
<th>Year</th>
<th>Type</th>
<th>Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>BGS</td>
<td>92/01</td>
<td>1992</td>
<td>High-resolution data. Small airgun array, short cable</td>
<td>Central and north Rockall Trough</td>
</tr>
<tr>
<td>BGS</td>
<td>93/02</td>
<td>1993</td>
<td>Large airgun array, long cable</td>
<td>West Rockall Basin</td>
</tr>
<tr>
<td>BGS</td>
<td>98/01</td>
<td>1998</td>
<td>High-resolution data. Small airgun array, short cable</td>
<td>Hatton Bank</td>
</tr>
<tr>
<td>BGS</td>
<td>00/01</td>
<td>2000</td>
<td>High-resolution data. Small airgun array, short cable</td>
<td>Hatton Bank, Hatton-Rockall Basin, Rockall Bank</td>
</tr>
<tr>
<td>BGS</td>
<td>02/02</td>
<td>2002</td>
<td>High-resolution data. Small airgun array, short cable</td>
<td>North Hatton Bank and north Rockall Trough</td>
</tr>
<tr>
<td>BGS</td>
<td>03/01</td>
<td>2003</td>
<td>High-resolution data. Small airgun array, short cable</td>
<td>Area between Lousy and Hatton Banks, Hatton Bank and north Rockall Trough</td>
</tr>
</tbody>
</table>
### 1.5.2 Well Data

Eleven wells and boreholes were used to date unconformities in the Rockall-Faroe area (Table 1.2). The location of these wells and boreholes are shown in Fig. 1.2. Check shot data were available for five wells/boreholes - 164/07-1, 164/25-2, 164/25-1, 154/01-1 and 982. Check shot data consists of one-way times measured at known depths in the wells during drilling. The check shot data facilitates the calculation of a time-depth curve in Landmark (software used to image seismic data) for the conversion of the depths in wells to two-way times represented on seismic images. This conversion, depth to two-way time, makes the well data compatible with the seismic data, allowing the calibration of horizons by the well data. The check shot data for borehole 982 was used to calculate the time-depth curve for borehole 116 (6.5 km to the west of 982), which lacked check shot data but was located at a site with better quality seismic data.

The check shot data for well 163/06-1 and four boreholes – 117, 94/1, 94/4, and 610 - were not available. Calibration of stratigraphic surfaces at the sites of well 163/06-1 and boreholes 117, 94/1 and 94/4 was done by extrapolating the sediment log, of the well and the boreholes, between the known sea bed and top-basalt horizons on the seismic data. The seismic interpretation in the vicinity of borehole 610, located outwith the study area, is based on previous work (Dolan, 1986).
<table>
<thead>
<tr>
<th>Well/borehole</th>
<th>Company</th>
<th>Year drilled</th>
<th>Total depth (m)</th>
<th>Check shot data</th>
<th>Area drilled</th>
</tr>
</thead>
<tbody>
<tr>
<td>163/06-1</td>
<td>British National Oil Corporation</td>
<td>1980</td>
<td>3686</td>
<td>absent</td>
<td>NE Rockall Trough</td>
</tr>
<tr>
<td>164/07-1</td>
<td>Conoco (UK) Limited</td>
<td>1998</td>
<td>5129</td>
<td>present</td>
<td>NE Rockall Trough</td>
</tr>
<tr>
<td>164/25-1</td>
<td>British Petroleum</td>
<td>1988</td>
<td>4050</td>
<td>present</td>
<td>NE Rockall Trough</td>
</tr>
<tr>
<td>164/25-2</td>
<td>British Petroleum</td>
<td>1992</td>
<td>2728</td>
<td>present</td>
<td>NE Rockall Trough</td>
</tr>
<tr>
<td>154/01-1</td>
<td>Enterprise Oil</td>
<td>2000</td>
<td>3050</td>
<td>present</td>
<td>NE Rockall Trough</td>
</tr>
<tr>
<td>982 (6.5 km east of borehole 116)</td>
<td>Ocean Drilling Project</td>
<td>1995</td>
<td>615</td>
<td>present</td>
<td>Hatton-Rockall Basin</td>
</tr>
<tr>
<td>117</td>
<td>Deep Sea Drilling Project</td>
<td>1970</td>
<td>313</td>
<td>absent</td>
<td>Hatton-Rockall Basin</td>
</tr>
<tr>
<td>94/1</td>
<td>British Geological Survey</td>
<td>1994</td>
<td>65</td>
<td>absent</td>
<td>Central Rockall Trough</td>
</tr>
<tr>
<td>94/4</td>
<td>British Geological Survey</td>
<td>1994</td>
<td>59</td>
<td>absent</td>
<td>Central Rockall Trough</td>
</tr>
<tr>
<td>610</td>
<td>Deep Sea Drilling Project</td>
<td>1983</td>
<td>723</td>
<td>absent</td>
<td>South Rockall Trough</td>
</tr>
</tbody>
</table>

Table. 1.2. Wells and boreholes used to calibrate stratigraphic surfaces in the Rockall-Faroe study area. For the location of wells (italics) and boreholes see Fig. 1.2. Total depths for commercial wells in metres below rotary table (mbrt). Total depths for boreholes in metres below sea floor (mbsf).

1.5.3 Gravity Data

Gravity data acquired during BGS seismic surveys have been used to construct gravity models of potential underlying structures within the study area. GravMag software (Pedley et al., 1993) was used to construct the models from five gravity profiles (for location, see Fig. 1.2). GravMag calculated gravity anomalies for constructed polygons with specific densities. The shape, extent and densities of
polygons were manipulated to achieve calculated gravity anomalies that best matched the observed Bouguer gravity anomalies of the five gravity profiles. The top surface of constructed models was constrained by sea level and the sea-bed bathymetry. Seismic data was used to constrain the top-basalt stratigraphic surface and the Cenozoic sediment thicknesses within the models.

The gravity anomalies calculated by GravMag for the models is relative to a background density. This background density was derived from a reference model (Fig. 1.18). The reference model was constructed to represent an ideal structure (0 - 50 km) of a continental margin. The thicknesses and densities of upper and lower crust, and mantle are based on previous work by Klingelhofer et al. (2005). A background density of approximately 3 Mg/m$^3$ was calculated from the weighted average of densities of the reference model. This background density was incorporated in the models to achieve the best fit between the calculated and the observed Bouguer gravity anomalies.

![Fig. 1.18. Reference model used to calculate the background density of gravity models. The thicknesses and densities of rocks (Mg/m$^3$) are based on Klingelhofer et al. (2005).](image)

1.5.4 Modelling rock behaviour in response to compression

SAVFEM$^\text{TM}$ (Structural Analysis Via Finite-Element Methodology) (Applied Mechanics Inc.) was used to build a model to determine the response of rock materials to forces. The software allowed properties to be assigned to elements or grids and simulated their behaviour when these elements are compressed. Material consisting of these grids and representing sediment, basement (upper crust), lower crust and mantle, were constructed and assigned properties for elastic moduli,
Poisson’s ratios and densities (Table 1.3). The Elastic or Young’s modulus gives the ratio of the stress applied to a material to the strain (% shortening) and is a measure of the stiffness of the rock (Davis and Reynolds, 1996). Poisson’s ratio is the ratio of lateral strain to longitudinal strain and describes the degree to which a core of rock bulges as it shortens (Davis and Reynolds, 1996).

<table>
<thead>
<tr>
<th>Material</th>
<th>Elastic Modulus (MPa)</th>
<th>Poisson’s ratio unitless</th>
<th>Density (kg/m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sediment (sandstone)</td>
<td>15000</td>
<td>0.3</td>
<td>2300</td>
</tr>
<tr>
<td>Crust</td>
<td>60000</td>
<td>0.26</td>
<td>2600</td>
</tr>
<tr>
<td>Lithosphere</td>
<td>100000</td>
<td>0.3</td>
<td>3400</td>
</tr>
</tbody>
</table>

Table 1.3. Parameters used for SAVFEM model. The elastic modulus for the crust based on properties of granite (Davis and Reynolds, 1996) and the poisson’s ratio of the crust is from Zandt and Ammon (1995). The poisson’s ratio of sediment is based on Rieke and Chilingarian (1973) and the elastic modulus of sediment is based on (Plona and Cook, 1995). The elastic modulus and the poisson’s ratio of the lithosphere are based on Ruiz et al. (2006). The density of sediment is based on Keary and Brooks (1991) and the densities of the crust and the lithosphere are based on Fowler (2001).

The software also simulates deformation based on strain-stress properties of the material. The sediments were simulated to follow a generalized power law isotropic plastic material properties data set for the Berea sandstone (Couples et al., 2007). The materials representing the crust and the lithosphere were simulated to deform as isotropic elastic materials.

SAVFEM™ was also used to simulate fault movements. Friction-gap elements were constructed, using SAVFEM™, and allowed adjacent parts of the model to slide past one another according to the assigned angle of friction, thereby simulating fault movement. Faults in the model were assigned an angle of friction of 30°. Boundary forces were exerted on the sides of the model and the vertical displacements, due to fault movement, were recorded. The principles of stress distributions and faulting utilized by SAVFEM™ are based on Hafner (1951).
1.6 Thesis Layout

The thesis consists of six main chapters, each of which represents different areas of study and research.

Chapter Two

Chapter two summarizes the history of the NE Atlantic Margin which has led to the development of structures such as basins and lineaments at certain locations along the margin. The structural setting of the margin plays a key role in determining the sites of inversion in the Rockall-Faroe area. This chapter also highlights the previous work on inversion along the NE Atlantic Margin, within and outwith the Rockall-Faroe area. The advantages of this study over previous studies of inversion in the Rockall-Faroe area are also highlighted.

Chapter Three

Chapter three discusses the types of unconformities that can be present on seismic data and the characteristics of these unconformities which facilitate their identification. Unconformities and the top-basalt surface in the Rockall-Faroe area are described and displayed on seismic data. The well data used to constrain the ages of these stratigraphic horizons are presented. The level of uncertainty in the interpretation of these unconformities in the Rockall-Faroe area, based on the availability of well and seismic data, is also highlighted.

Chapter Four

This chapter describes the nature and orientation of the various compressional structures present in the Rockall-Faroe area. Seismic images of compressional structures and their associated unconformities are shown. Isochron maps are present to delineate the orientation of structures. Gravity models are also present in this chapter to portray possible underlying structures beyond the scope of the seismic data.
Chapter Five
Chapter five presents isochore maps to understand the sediment distribution across the Rockall-Faroe area through time and space. These sediment patterns may reflect the development and evolution of compressional structures. This chapter also highlights the role of bottom-water current activity in the distribution of sediment and the formation of unconformities. The potential influence of compression on bottom-water current activity in the Rockall-Faroe area is addressed. This chapter also assesses the role of eustatic sea level changes in controlling the distribution of sediment in the Rockall-Faroe area.

Chapter Six
Chapter six describes the potential driving mechanisms of inversion in the Rockall-Faroe area. These are ridge-push (normal and hotspot-influenced), Alpine and Pyrenean compression, depth-dependent stretching and Iceland Insular Margin body force. The potential of each mechanism is assessed based on the direction, extent and timing of their forces. The effectiveness of the magnitude of ridge push is also assessed using Mohr-Coulomb diagrams and finite-element models. The potential causes of changes in bottom-water currents are also explored in this chapter. The timing of compressional features and the potential mechanisms forming unconformities have been used to outline the tectonostratigraphic evolution of the NE Atlantic Margin from the Late Paleocene to the early Late Pliocene.
2.0 The NE Atlantic Passive Margin

2.1 History of the NE Atlantic Passive Margin

In Chapter 1 the mechanics of rifting and the formation of volcanics, in continental margins, were explored. In this chapter, the history of these and other events from the Late Paleozoic to the Cenozoic are presented. It is the timing of these events that has established the structural framework of the NE Atlantic margin in the Cenozoic. This structural framework has influenced and controlled the development of inversion features in the Rockall-Faroe study area. It is, thus, vital to examine the history of the NE Atlantic continental margin in order to shed light on the development of these features present in the study area.

**Pre-Cambrian - Silurian**

The basement rock of the NE Atlantic Margin was created in the Pre-Cambrian (Woodcock and Strachan, 2000; Coward et al., 2003) and comprises crystalline rocks which originated from at least three tectonic blocks – Laurentia, Baltica and Eastern Avalonia. These terranes accreted during the Caledonian collision which took place in the Silurian – Early Devonian (Woodcock and Strachan, 2000; Coward et al., 2003). The Rockall-Faroe area occurs within the Laurentian terrane which comprises Archaean gneiss (~ 2900 Ma) which was reworked during the Mid-Proterozoic (1800-1600 Ma) (Coward et al., 2003). Basement rocks outcrop the Rockall-Faroe area in the Hebrides Shelf (Hitchen et al., 2002) and the Rockall Bank (Morton and Taylor, 1991; Dickin, 1992; Hitchen, 2004). In the West Shetland Basin the Rona Ridge consists of Lewisian basement which is covered by Devonian-Carboniferous sediments (Stoker et al., 1993).

**Devonian**

Caledonian continental collision, between Laurentia and Baltica, ceased in the late Silurian to Early Devonian (Coward et al., 2003). Basin development took place onshore and offshore southern Norway and in the northern North Sea (Coward et al., 2003). In the Hebrides Shelf and the West Orkney region, the North Lewis Basin and the West Orkney Basin are inferred to have developed in Devonian time as a
Fig. 2.1. Devonian structure of NW Europe. (a) Palinspastic map for the Mid Devonian showing the distribution of active faults and sediment facies. Note the lack of basin development in the Rockall-Faroe study area at this time. (b) Mid Devonian global tectonics at around 380 Ma. (c) Devonian faults of north-west Europe. Diagram from Coward et al. (2003).
result of post-Caledonian Orogenic collapse (Stoker et al., 1993). Post-Caledonian Orogenic collapse was first proposed by McClay et al. (1986) for the Orcadian Basin and the Midland Valley Graben of Scotland where Old Red Sandstone (> 9 km thick) was deposited in the Devonian. The Clair Basin, to the north of the Orcadian Basin, was also formed in the Devonian (Nichols, 2005). Active basins within the Devonian are shown in Fig. 2.1. There is no evidence to suggest Devonian rifting activity within the Rockall-Faroe area.

**Carboniferous**
Acadian-Variscan compression in the Early Carboniferous resulted in the formation of the massive Variscan mountains in NW Europe (Fig. 2.2; Coward et al., 2003). The Variscan mountain belt was well established by the Late Carboniferous and preceded the onset of rifting in the Early Permian. Carboniferous basin development was evident in the area of the Erris Trough, north-west of Ireland on the eastern flank of the Rockall Trough (Chapman et al., 1999) and in the Clare Basin located west of Ireland (Croker, 1995).

**Permian - Triassic**
Gravitational collapse of the Appalachian and European Variscan mountains, formed in the Carboniferous, resulted in the onset of continental rifting in the NE Atlantic Margin (Coward et al., 2003). Permo-Triassic rifting and deposition of continental red bed successions took place in the West Orkney Basin, North Lewis Basin, and Flannan Trough and West Shetland Basin (Stoker et al., 1993). The Erris Trough, north-west of Ireland, also underwent Permo-Triassic rifting (Chapman et al., 1999). Triassic rifting affected or formed basins such as the West Shetland Basin, the North Celtic Sea Basin, and the Wessex Basin (Fig. 2.3; Coward et al., 2003). Rifting in the Permo-Triassic within the Rockall-Faroe study area is not recorded in previous work.
Fig. 2.2. Late Carboniferous structure of NW Europe. (a) Palinspastic map for the Late Carboniferous showing the distribution of active faults and sediment facies. Note the location of the Clare Basin (Croker, 1995) and Erris Trough (Chapman et al., 1999) which developed in Carboniferous time (Chapman et al., 1999). (b) Late Carboniferous global tectonics at around 300 Ma. (c) Late Carboniferous faults of NW Europe. Diagram from Coward et al. (2003).
Fig. 2.3. Mid-Triassic structure of NW Europe. (a) Palinspastic map for the Mid Triassic showing the distribution of active faults and sediment facies. Note rifting in the West Shetland Basin, North Celtic Basin, Wessex Basin and Western Approaches Basin at this time. (b) Mid-Triassic global tectonics. (c) Permian and Triassic salt basins in north-west Europe. Diagram from Coward et al. (2003).
**Jurassic - Cretaceous**

Rifting continued in the Jurassic to Cretaceous periods in NW Europe. In the Jurassic there was rifting in the Faroe Shetland Basin and the North Rona Basin (Stoker et al., 1993). The Porcupine Basin, Celtic Sea Basin and Western Approaches Basin, which underwent extension in the Mid-Triassic, continued to rift in Early Jurassic times (Coward et al., 1993).

In the Early Cretaceous there was rifting in the Labrador Sea between Greenland and North America (Fig. 2.4; Coward et al., 2003). According to Musgrove and Mitchener (1996), the main phase of rifting of the Rockall Basin (Rockall Trough), within the study area, occurred in the Early Cretaceous. In the Møre Basin, rifting also took place in the Early Cretaceous (Knott et al., 1993). Late Jurassic and Early Cretaceous rifting was observed in the Vøring Basin (Knott et al., 1993).

In the Late Cretaceous, rifting continued in the Rockall Basin, Faroe-Shetland Basin and in the Vøring Basin (Coward et al., 2003).

**Paleocene - Oligocene**

Two periods of volcanism occurred in the NE Atlantic from Paleocene to earliest Eocene times:


According to Coward et al. (2003), the volcanic events probably coincided with phases of extension. Volcanism in the NE Atlantic Margin took the form of underplated material, basalt lava flows, seaward-dipping reflectors and seamounts (see section 1.2.4). High degree of volcanism in the NE Atlantic has been attributed to the presence of a mantle plume (White, 1988; White and McKenzie, 1989; Cox, 1989; Griffiths and Campbell, 1990). The structure of NW Europe in the Paleocene is shown in Fig. 2.5.

Sea-floor spreading between Greenland and Europe, to form the NE Atlantic, was initiated in the Early Eocene – 53 Ma (Doré et al., 1999); 54 Ma (Smallwood et al.,
Fig. 2.4. Early Cretaceous structure of NW Europe. (a) Palinspastic map for the Early Cretaceous showing the distribution of active faults and sediment facies. Note rifting of the Rockall Trough within the Rockall-Faroe study area. (b) Early Cretaceous global tectonics at around 120 Ma. (c) Early Cretaceous structures in north-west Europe. Diagram from Coward et al. (2003).
Fig. 2.5. Paleocene structure of NW Europe. (a) Palinspastic map for the Paleocene showing the distribution of active faults, sediment facies and volcanic rocks associated with the North Atlantic mantle plume. The position of the plume in the Late Paleocene time is shown (Torsvik et al., 2001). (b) Paleocene global tectonics at around 60-55 Ma. (c) Paleocene structures in northwest Europe. Diagram from Coward et al. (2003).
Fig. 2.6. Early Oligocene structure of NW Europe. (a) Palinspastic map for the Oligocene showing the distribution of active faults and sediment facies. (b) Oligocene global tectonics at around 30 Ma. (c) Oligocene structures in north-west Europe. Diagram from Coward et al. (2003).
2002) resulting in the formation of the Reykjanes Ridge, Aegir Ridge and Mohns Ridge. Oceanic crust, formed at these ridges, separated the NE Atlantic Margin from the East Greenland Margin in the Oligocene (Fig. 2.6). Sea-floor spreading in the Labrador Sea, which began in the Paleocene (Chalmers, 1997), ceased in the late Eocene (Nunns, 1983; Hinz et al., 1993).

**Miocene**

In the Early Miocene there was the initiation of the Kolbeinsey Ridge (Fig. 2.7; Coward et al., 2003). This resulted from a ridge-jump across the Jan-Mayen microcontinent from the now extinct Aegir Ridge (Fig. 2.8). This has been attributed to the presence of the Iceland plume (Lundin and Doré, 2002). The higher gravitational potential and weaker lithosphere found above the plume core would have made rifting more favourable to the NW of the Aegir Ridge to form the Kolbeinsey Ridge (Smallwood and White, 2002). Such plate reorganization could have resulted in increased stress to result in compression in the Early Miocene (Lundin and Doré, 2002). Fault reactivation, resulting in basin inversion and the uplift of half-graben fill, associated with strike-slip transfer structures, occurred along the NE Atlantic Margin in Miocene times (Coward et al., 2003). In the Miocene there was also compression along the Alpine foreland (Coward et al., 2003).
Fig. 2.7. Miocene structure of NW Europe. (a) Palinspastic map for the early Miocene showing the distribution of active faults and sediment facies. (b) Miocene to Pliocene global tectonics at around 10-5 Ma. (c) Miocene structures in north-west Europe. Diagram from Coward et al. (2003).
Fig. 2.8. Plate tectonic evolution of the Norwegian-Greenland Sea. Grey and yellow dots indicated the position of the Iceland Plume centre in the Cenozoic; grey dots mark the previous positions and yellow dots mark the approximate position at each reconstruction. The position of the Iceland Plume centre is from Torsvik et al. (2001). AR = Aegir Ridge, JM = Jan Mayen micro-continent, KR = Kolbeinsey Ridge, KnR = Knipovitch Ridge, MR = Mohns Ridge, RR = Reykjanes Ridge. Active ridges are marked by red lines, and inactive ridges are green. Diagram from Lundin and Doré (2002).
2.2 Structure of the NE Atlantic Margin

The NE Atlantic Margin extends from the southwest of Ireland to northern Norway (Shannon et al., 2005a). The margin consists of a number of basins and structural highs (Figs. 2.9 - 2.11) transected by NW-trending lineaments (Figs. 2.10 - 2.11).

2.2.1 Basins

The basins within the NE Atlantic Margin, from north to south, include the Vøring, Møre, Faroe-Shetland, NE Rockall, North Rockall, Hatton-Rockall and South Rockall basins. These basins contain thicker sediment compared to surrounding structural highs, based on 3D gravity modelling (Fig. 2.11). These basins were formed as a result of extension during continental rifting, revealed by the presence of normal faults as exemplified by a profile across the Flett Sub-basin to the south of the Faroe Shetland Basin (Fig. 2.12). An attempt has been made to link the Cenozoic sediment stratigraphy between these basins across the NE Atlantic Margin (Fig. 2.13). Stratigraphic boundaries or unconformities of similar ages have been mapped across the margin. These include the C30 (Late Eocene), Late Oligocene to Mid Miocene, and Early Pliocene megasequence boundaries (Fig. 2.13). Basins within the NE Atlantic Margin are underlain by continental crust (Shannon et al., 2005a). According to Joppen and White (1990), however, the Rockall Basin is underlain by either thinned continental crust heavily intruded by syn-rift igneous rocks or oceanic crust based on seismic velocity gradients.

2.2.2 Structural Highs

Structural highs within the margin surround basins and include the Rockall High, Hatton Bank High, Ymir Ridge, Wyville-Thomson Ridge, Faroe Bank High, and the Faroe Platform (Figs. 2.9 - 2.10). Some of these structural highs have been described as horst blocks (Vanneste et al., 1995), seamounts (Ritchie et al., 1999) and compressional structures (Boidreel and Andersen, 1998). In this study, the nature of the structural highs present in the Rockall-Faroe area is examined.
2.2.3 Lineaments

3D gravity modelling has revealed the presence of lineaments in the NE Atlantic Margin (Kimbell et al., 2005). The lineaments are generally NW-trending, segmenting basins along the margin (Fig. 2.10 and Fig. 2.11). Some of these lineaments have been interpreted as transfer zones which could have been formed during Mesozoic rifting to accommodate displacement between rift segments (Morley et al., 1990; cited in Doré et al., 1997). Transfer zones may develop over pre-existing lineaments or tectonic grain (Cartwright, 1992; cited in Doré et al., 1997). The Anton Dohrn lineament (Fig. 2.10), for example, represents a major terrane boundary separating Archaean crust (Lewisian Terrane), in the north, from Proterozoic crust to the south (Dickin, 1992). Some structural highs exist on lineaments. The Wyville-Thomson Ridge (Fig. 2.10), for example, is part of a complex of lineaments that includes the Bill Bailey’s Bank and Ymir Ridge, extending from the continent-ocean boundary (Kimbell et al., 2005).
Fig. 2.9. Structural highs and basins across the NE Atlantic Margin. Abbreviations: AD, Anton Dohrn igneous centre; CR, Corona Ridge; EFH, East Faroes High; EP, Erlend Platform; F, zone of fractured oceanic crust (outlined with a white dashed line); FBC, Faroe Bank Channel; FBH, Faroe Bank High; FLB, Faroe Bank High; FLB, Flett Sub-basin; FZ, fracture zone; HH, Helland Hansen Arch; HT, Hebrides Terrace igneous centre; JH, Judd High; KR, Kolbeinsey Ridge; M, Mammal igneous centre; MB, Magnus Basin; MR, Munkagrunnur Ridge; NERB, North-east Rockall Basin; NRB, North Rona Basin; NSP, North Shetland Platform; NVG, North Viking Graben; OL, Ormen Lange Dome; R, Rockall igneous centre; RR, Rona Ridge; S, Swithin igneous centre; SG, Sogn Graben; WOB, West Orkney Basin; WSB, West Shetland Basin; WTR, Wyville-Thomson Ridge; YR, Ymir Ridge. Age of ocean crust after Muller et al. (1997). Diagram from Kimbell et al. (2005).
Fig. 2.10. Gravity variations across the NE Atlantic Margin. Free-air anomalies offshore; Bouguer anomalies onshore (Kimbell et al., 2005). Abbreviations for structural elements as in Fig. 2.9. Lineaments: ADL, Anton Dohrn Lineament Complex; CL, Clair Lineament; JL, Judd Lineament; JML, Jan Mayen Lineament; MFL, Marflo Lineament, ML, Magnus Lineament; MTL, Møre-Trøndelag Lineaments; SHL, South Hatton Lineament; WTL, Wyville-Thomson Lineament Complex.
Fig. 2.11. Sediment thickness across the NE Atlantic Margin based on regional 3D gravity modelling (Kimbell et al., 2004). Abbreviations for structural elements and lineaments as in Fig. 2.9 and Fig. 2.10 respectively. Diagram from Kimbell et al. (2005).
Fig. 2.12. NW-SE regional cross-section over the Flett sub-basin and the Rona Ridge south of the Faroe-Shetland Basin. For the position of these features refer to Fig. 2.11. Diagram from Lamers and Carmichael (1999).
Fig. 2.13. Cenozoic stratigraphy of the NE Atlantic Margin from the Porcupine to the Vøring Basins (STRATAGEM Partners, 2002). For location of areas refer to Fig. 2.11.
2.3 Inversion Structures

According to Coward et al. (2003), inversion features within the NE Atlantic Margin formed in the Oligocene (Fig. 2.6) and in the Miocene (Fig. 2.7). Inversion features have been previously recorded in the Norwegian Margin, Faroe-Shetland Basin and in the Rockall-Faroe area (Fig. 2.14).

2.3.1 Rockall-Faroe area

Structural highs in the continental margin were first ascribed to compression by Boldreel and Andersen (1993). Structures such as the Hatton Bank, the Wyville-Thomson Ridge, the Ymir Ridge and the Munkagrunnur Ridge were studied within the Rockall-Faroe area (Boldreel and Andersen, 1993; 1994; 1998). Late Paleocene - Early Eocene, Oligocene and Middle Miocene ages were proposed for the compressional events. In later years, there were more studies over the Wyville-Thomson Ridge (Johnson et al., 2005; Ziska and Varming, 2008; Ritchie et al., 2008), Ymir Ridge (Ziska and Varming, 2008; Ritchie et al., 2008) and the Hatton Bank (Hitchen, 2004; Johnson et al., 2005). Johnson et al. (2005) proposed Paleocene, Late Eocene, Top Palaeogene and Intra-Miocene ages for the growth of the Wyville-Thomson Ridge (Fig. 2.15). These ages were also supported by Ritchie et al. (2008). The Ymir Ridge was attributed to compression in the Late Eocene (C30), Top Palaeogene and Intra-Miocene (Ritchie et al., 2008). Ziska and Varming (2008) proposed a non-compressional origin - a transient rifting episode in the early Paleocene - for the Wyville-Thomson Ridge and the central segment of the Ymir Ridge. The south segment of the Ymir Ridge, however, is viewed as being the result of compression (Ziska and Varming, 2008).

The Munkagrunnur Ridge may have been formed due to compression in latest Paleocene or later times - possibly during the Eocene, Oligocene and the Miocene (Ritchie et al., 2008). In addition to these ridges, the Alpin Dome was also mapped and dated (Stoker et al., 2005a; Johnson et al., 2005; Ritchie et al., 2008). Phases of compression, for this anticline, occurred during the Late Eocene (C30), the Mid Miocene and possibly the Late Oligocene (Top Palaeogene) (Ritchie et al., 2008).
Fig. 2.14. Compressional structures previously mapped in the NE Atlantic Margin. Basemap from Kimbell et al. (2005). Fold abbreviations: AA = Anticline A, ADo = Alpin Dome, BBB = Bill Bailey's Bank, FR = Fugløy Ridge, HA = Helland Hansen Arch, HB = Hatton Bank, LB = Lousy Bank, MA = Modgunn Arch, OL = Ormen Lange Dome, MR = Munkagrunnur Ridge, WTR = Wyville Thomson Ridge, YR = Ymir Ridge. Lineament abbreviations as in Fig. 2.10.
Late Eocene (C30) growth was proposed by Johnson et al. (2005) for the Hatton Bank (Fig. 2.16).

Fig. 2.15. Interpreted SW-NE seismic profile across the Ymir and Wyville-Thomson Ridges (Johnson et al., 2005). For location of anticlines see Fig. 2.14. Seismic horizons: 1 = sea bed, 2 = Intra-Neogene (late Early Pliocene), 3 = C10 (late Early Pliocene), 6 = Intra-Miocene, 7 = Top Palaeogene, 9 = Intra-Eocene c, 11 = Intra-Eocene a, 15 = Top Paleocene lavas. Faults are shown by yellow lines.

Fig. 2.16. Interpreted N-S seismic profile across the Hatton High (Bank) (Johnson et al., 2005). For location of the anticline see Fig. 2.14. Seismic horizons: 1 = sea bed, 3 = C10 (late Early Pliocene), 8 = C30 (Late Eocene), 12 = Intra-Eocene a, 15 = Top Paleocene lavas. Faults are shown by yellow lines.
2.3.2 Faroe-Shetland Basin

Boldreel and Andersen (1993) identified Mid- or Late Miocene compression in the Faroe-Shetland Basin. NE- and NNE-trending folds have been mapped within the Faroe-Shetland Basin by Ritchie et al. (2003) and Ritchie et al. (2008), such as the Fugløy Ridge (Fig. 2.17) and Anticline A (Fig. 2.18). The folds were formed mainly during early to mid-Miocene times (Ritchie et al., 2003). Fold growth may have continued into Early Pliocene times and there is also evidence of older phases of compression (Ritchie et al., 2003). Previous work by Nicholson (2005) revealed a compressional age of Late Paleocene for an inverted NE-trending graben within the Faroe-Shetland Basin. Inversion of the Judd High anticline (for location, see Fig. 2.9) occurred in the late Ypresian, the late Lutetian, and the Oligocene (Smallwood, 2004).
Fig. 2.17. Interpreted NW-SE seismic profile across the Fugløy Ridge (Fugløy Anticline or Anticline C) (Ritchie et al., 2003). For the location of Fugløy Anticline see Fig. 2.14. Seismic horizons: 1 = sea bed, 2 = Intra Neogene Unconformity, 3 = Pliocene Unconformity, 4 = Mid-Miocene Unconformity, 5 = Intra-Miocene Unconformity, 6 = Top Palaeogene Unconformity, 7 = Opal A-Opal C/T transition event, 8 = Top Palaeogene lavas.

Fig. 2.18. Interpreted NW-SE seismic profile across anticline A (Ritchie et al., 2003). Seismic horizons as in Fig. 2.17. For the location of Anticline A see Fig. 2.14.
2.3.3 Norwegian Margin

Compressional structures have been identified within the Norwegian margin (Lundin and Doré, 2002; Brekke, 2000; Løseth and Henriksen, 2005). The Helland Hansen Arch (Fig. 2.19), in the Vøring Basin, is marked by a Mid-Miocene unconformity which defines the time of compression (Brekke, 2000; Løseth and Henriksen, 2005). According to Blystad et al. (1995), the Helland Hansen Arch developed during Eocene-Oligocene and Late Miocene times. Domes, such as the Helland Hansen Arch and the Ormen Lange Dome, in the Norwegian margin, formed mainly in the Middle Eocene to Early Oligocene, and in the Early Miocene (Lundin and Doré, 2002).

![Interpreted NNW-SSE seismic profile across the southern Helland Hansen Arch](image)

Fig. 2.19. Interpreted NNW-SSE seismic profile across the southern Helland Hansen Arch (Doré et al., 2008). For the location of Helland Hansen Arch refer to Fig. 2.14.
The NE Atlantic margin shows strong evidence for compression within the Cenozoic. A summary of compressional ages recorded in the NE Atlantic Margin is shown in Table 2.1.

<table>
<thead>
<tr>
<th>Author</th>
<th>Area</th>
<th>Age of compression</th>
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<tbody>
<tr>
<td>Boldreel and Andersen (1993)</td>
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<td></td>
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<td>Oligocene</td>
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<td>Late Paleocene – Early Eocene</td>
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<td>Hatton Bank</td>
<td>Late Eocene</td>
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<td>Johnson et al. (2005)</td>
<td>Wyville-Thomson Ridge</td>
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<td>Top Palaeogene</td>
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<td>Hatton Bank</td>
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<td>Ritchie et al. (2008)</td>
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<td></td>
<td></td>
<td>Top Palaeogene</td>
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<td></td>
<td>Alpin Dome</td>
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<td>Mid Miocene</td>
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<td>Late Oligocene (Top Palaeogene)</td>
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<td>Late Eocene</td>
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<tr>
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<td>Mid or Late Miocene</td>
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<td>Ritchie et al. (2003)</td>
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<td>Mid-Miocene</td>
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<td>Eocene-Oligocene</td>
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<td>Late Ypresian</td>
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<td>Nicholson (2005)</td>
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<tr>
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<td>Helland Hansen Arch (Vøring Basin)</td>
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<td>Mid Eocene – Early Oligocene</td>
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Table 2.1. Summary of the ages of compressional events in the NE Atlantic Margin from previous work. For the location of previously studied areas refer to Fig. 2.14.
2.4 Mechanisms and Controls

Previous work has attempted to link the ages of compressional structures in the NE Atlantic Margin to regional events. Compressional features within the NE Atlantic Margin have been attributed to:

- Ridge push (Bott, 1991; Boldreel and Andersen, 1993; Doré and Lundin, 1996; Lundin and Doré, 2002).
- Alpine Compression (Doré and Lundin, 1996; Brekke, 2000; Roberts, 1989; Vågnes et al., 1998).
- Pyrenean Compression (Knott et al., 1993).
- Plate reorganization (Boldreel and Andersen, 1993; Lundin and Doré, 2002).
- Iceland Plateau body force (Doré et al., 2008).

It has been proposed that many of the compressional structures are inverted basins. These include the Helland Hansen Arch (Grunnaleite and Gabrielsen, 1995; Vågnes et al., 1998; cited in Mosar et al., 2002), the Ormen Lange Dome (Doré and Lundin, 1996), the Naglfar Dome (Lundin and Doré, 2002), the Wyville-Thomson Ridge (Klingelhofer et al., 2005), and the Ymir Ridge (Ziska and Varming, 2008).

The growth of compressional structures has also been linked to lineaments. Several domes in the Norwegian Margin, for example, are located en echelon along the Jan Mayen Lineament extending into the margin from the Jan Mayen Fracture Zone (Fig. 2.14). These domes include the Nora, the Edwarda, the Modgunn Arch, and the Ormen Lange. The Wyville-Thomson Ridge and the Ymir Ridge also lie on lineaments (Fig. 2.14). These lineaments were probably reactivated as part of a ramp anticline complex (Kimbell et al., 2005).
2.5 Summary

Mesozoic rifting in the NE Atlantic has created pre-Cenozoic basins along the continental margin. In Chapter 6 the role of these and other basins in the formation of anticlines in the Rockall-Faroe study area is examined. Some of these anticlines have been previously mapped and dated in the Rockall-Faroe area by previous authors such as Boldreel and Andersen (1993) and Johnson et al. (2005). However, hitherto, their ages of formation have remained uncertain. In this study, there has been critical use of the well data available to date horizons. These dated horizons were mapped throughout the study area using extensive seismic datasets. Previous studies, such as Boldreel and Andersen (1993), lacked well data to date horizons with certainty and the relatively limited seismic data available also hindered the extensive mapping of compressional features. In the Johnson et al. (2005) study, there was the lack of mapping of horizons from the site of wells. Most horizons, on the south Ymir Ridge, where inferred although well data was available in the NE Rockall Trough. The mapping of horizons from the NE Rockall Trough to other areas such as the south Ymir Ridge, Mordor, North Rockall Trough, and Alpin Dome resulted in a greater certainty in the ages of these horizons mapped in compressional structures. In addition, the seismic data available in this study facilitated not only the refining of the extent of previously mapped folds, such as the Alpin Dome, but the mapping of new folds such as the Dawn, Bridge, Onika and Viera anticlines. In this study, the Ymir Ridge has also been mapped as three distinct anticlines.

Gravity data, not used or available in the previous studies, allowed the nature of compressional structures to be examined. This was crucial in understanding the origin of these structures and the potential controls on inversion.

Compressional structures have been previously attributed to certain driving mechanisms. The lack of well constrained ages for these features in the Rockall-Faroe area, however, has made such a link between compression and mechanisms uncertain. In addition, the effectiveness of any mechanism was never tested qualitatively. This study incorporates finite-element modelling and Mohr-Coulomb plots to determine the efficiency of ridge push in producing the compressional features present in the Rockall-Faroe area.
3.0 Unconformities and stratigraphic surfaces

3.1 Introduction
Unconformities are stratigraphic surfaces which represent an interruption in the
deposition of sediment (McQuillin et al., 1984) or a break in a sequence of strata
representing a period of non-deposition of sedimentary material (Lapidus and
Winstanley, 1990). Unconformities define sediment packages in both seismic and
sequence stratigraphy (Mitchum et al., 1977a; 1977b; Mitchum, 1977). In the
Rockall-Faroe area, unconformities were identified on seismic profiles and have been
dated using well and borehole data available in this study. The ages of these
unconformities are crucial in determining the timing of compressional and non-
compressional events affecting the Rockall-Faroe area.

3.1.1 Types of Unconformities

Disconformity
This unconformity is an erosional surface which separates two series of sediment
strata which are parallel (Lapidus and Winstanley, 1990). The erosional surface
represents the removal of sediment by subaerial or submarine erosion (Nichols,
1999). Unconformities marked by erosional truncation are present within the
Rockall-Faroe study area.

Angular unconformity
An angular unconformity separates two series of sediment strata which are not
parallel (Lapidus and Winstanley, 1990). Tilting of sediment strata followed by
subsequent deposition of sediment can form an angular unconformity. Angular
unconformities have been formed with the Rockall-Faroe area as a result of the
deposition and onlap of sediment on folded strata.
Non-conformity
A non-conformity is the surface situated between overlying stratified sediments and underlying unbedded metamorphic or plutonic rocks (Lapidus and Winstanley, 1990). In the Rockall-Faroe area sediments onlap igneous centres to form this unconformity.

Paraconformity
A paraconformity is an unconformity situated between two series of parallel sediment strata representing prolonged periods of non-deposition with little evidence of erosion (Lapidus and Winstanley, 1990). Hiatuses of non-deposition, not marked by erosional truncation, are evident in the Rockall-Faroe area. On the east margin of the Rockall Bank, for example, there is evidence on seismic data of Pliocene and Miocene sediment onlapping onto Late Eocene sediment.

![Fig. 3.1. Types of unconformities (redrawn from Lapidus and Winstanley, 1990).](image)

Each unconformity type (Fig. 3.1) is formed by an event which promotes an interruption in the deposition of sediment on a surface. Unconformities are thus important markers for events, such as folding, uplift and bottom-water current changes, affecting the Rockall-Faroe area.
Fig. 3.2. Seismic profile across a locality in the South Rockall Trough. Note the onlap of reflectors, reflector terminations and high amplitude reflectors, which are used to identify unconformities on seismic data.

(Seismic data courtesy of Fugro Multi Client Services)
3.1.2 Unconformities and reflector relationships on seismic data

The nature of reflector surfaces on seismic data can reveal the presence of unconformities. Unconformities can be identified on seismic by reflector terminations, onlapping reflectors or in some cases reflectors with high amplitudes. Reflector terminations can indicate erosional truncation whilst onlapping reflectors can represent an angular unconformity. Reflectors of high amplitude are the result of relatively large differences in the acoustic impedance (product of density and sonic velocity) of the two materials bounding a stratigraphic surface (Nichols, 1999). The high amplitude reflector can represent a change in lithology resulting from changes in depositional environment (Nichols, 1999) which may indicate a paraconformity. The different characteristics of reflectors used to identify unconformities on seismic profiles are shown in Fig. 3.2.

In addition to the onlap and erosional truncation of reflector surfaces, which may define unconformities, reflectors may also show clinoforms, downlap and offlap (Fig. 3.3). Clinoforms are gently sloping depositional surfaces formed through progradation of sediment into deeper water (Mitchum et al., 1977a; Nichols, 1999). Downlap describes inclined strata which terminate downwards onto a horizontal surface (Mitchum, 1977; Nichols, 1999) whilst offlap refers to reflection patterns generated from strata prograding into deepwater (Mitchum, 1977).

![Diagram showing reflector patterns and reflector relationships on seismic data]

Fig. 3.3. Reflector patterns and reflector relationships on seismic data (redrawn from Nichols, 1999).

One of the principal aims of this study is to recognize, map and date the various types of unconformities within the Rockall-Faroe area, by using the different
reflector patterns. The ages and characteristics of unconformities can elucidate the timings and nature of the mechanisms with formed them.

3.2 Methodology

Eleven wells/boreholes have been used to date or constrain the ages of unconformities within the study area. The locations of these wells and boreholes are shown in Figs. 3.4 - 3.7. These wells/boreholes are as follows:

Commercial wells:
- 154/01-1
- 163/06-1
- 164/07-1
- 164/25-1
- 164/25-2

BGS boreholes:
- 94/01
- 94/04

Deep Sea Drilling Project (DSDP) boreholes:
- 116
- 117

Ocean Drilling Project (ODP) boreholes:
- 610
- 982

For information on operator, drill date and total depth of penetration for each well and borehole refer to section 1.5.2.

Unconformities were first mapped in Landmark and their depths given in two-way time. In order to determine the ages of unconformities using well data, the depths (m) in wells were converted to two-way times. This was done using check shot data where one-way times were measured at known depths during drilling of the wells. The check shot data for wells 164/25-1, 164/25-2, 164/07-1, 154/01-1, and borehole 116/982 were used to calculate time-depth curves in Landmark in order to convert
Fig. 3.4. Seismic and well/borehole data in the Rockall-Faroe area of the NE Atlantic Margin. Wells (purple circles) and boreholes (red circles) were used to constrain the ages of stratigraphic surfaces and seismic data were used map these surfaces within the study area. Bathymetric contours (in metres) are represented by blue lines.
Fig. 3.5. Compressional structures and seismic data in the Rockall-Faroe study area.
Fig. 3.6. Seismic dataset in the NE Rockall Trough - Faroe Shelf area. Wells used to constrain the ages of stratigraphic surfaces and the location of seismic illustrations are shown by solid black lines.
Fig. 3.7. Seismic dataset across the Rockall Bank and Hatton-Rockall Basin. Boreholes used to constrain the ages of stratigraphic surfaces and the location of seismic illustrations are shown by solid black lines.
depths in wells into two-way times. This proved useful as all the major unconformities were dated using these wells.

For well 163/06-1 and boreholes 117 and 94/01, the depth in the wells were converted to two-way times by simply superimposing the well depth profile onto the two-way time (TWT) section from the seabed to the top-basalt horizon. The sea bed and top-basalt surfaces were selected as the horizons are easily identified in both the well/borehole data and the seismic TWT profiles. This was done under the assumption that the velocities within the post-basalt Cenozoic sequence of sediments did not vary significantly along the profile. This assumption was valid for the following reasons:

1. The C30 (Late Eocene) stratigraphic surface derived from well 163/06-1 is in good agreement with the position of the C30 surface of the proximal Onika Anticline.
2. The Upper Miocene-Upper Eocene unconformity in borehole 94/1 core matches the position of an unconformity, marked by onlap, in the seismic section.
3. The Oligocene-Early Eocene unconformity in borehole 117 correlates with a strong amplitude reflector marked by onlap and erosional truncation in the seismic profile.

Although no unconformities were penetrated by BGS borehole 94/4, the borehole revealed Upper Oligocene mass flow deposits of a lowstand fan (Stoker, 1995; Stoker, 1997). The age of the fan was used to constrain the position of the Late Eocene (C30) unconformity on the east margin of the Rockall Bank and in the Rockall Trough.

Site 610 is located in the South Rockall Trough, outwith the study area (Fig. 3.4). This borehole was instrumental in the dating of the C20 (late Early Miocene) unconformity in the Central-South Rockall Trough (Ruddiman et al., 1994). The C20 unconformity is marked by a layer rich in smectite that was possibly derived from the diagenesis of volcanic ash (Dolan, 1986). There is also a marked increase in sonic velocity across this layer, attributed to the dissolution of biogenic silica (Baldauf, 1986).
Fig. 3.8. NW-SE seismic line across the Feni Ridge in the South Rockall Trough (for location, see Fig. 3.4). Interpretation of C30 (Late Eocene), C20 (late Early Miocene) and C10 (late Early Pliocene) unconformities is based on Stoker et al. (2001). Note the onlap onto the C30 unconformity. The C20 unconformity is defined by the onlap of a relatively thick smectite layer (Dolan, 1986).

(Seismic data courtesy of Fugro Multi Client Services)
3.3 Unconformities and the top-basalt surface in the Rockall-Faroe study area

Unconformities have been previously mapped and dated in the Central to South Rockall Trough (Stoker et al., 2001). These are the C30 (Late Eocene), C20 (late Early Miocene) and C10 (late Early Pliocene) unconformities (Fig. 3.8). In this study, these and other unconformities have been mapped and dated across the Rockall-Faroe area. Unconformities have been identified by onlap, downlap and erosional truncation. Some unconformities help to constrain the ages of folding, while others may be due to the action of bottom-water currents. The ages of unconformities mapped in the Rockall-Faroe study area are shown in Fig. 3.9.

![Unconformities Diagram](image)

Fig. 3.9. Age of unconformities dated and mapped in the Rockall-Faroe study area. Time scale based on Gradstein et al. (2004).

The following section seeks to further describe the extent, nature and expression of unconformities and other stratigraphic horizons. The wells used to date stratigraphic surfaces and the depths and seismic character of horizons are shown in Tables 3.1 – 3.9.
**Thanetian**

An unconformity, recognized in the West Lewis Basin, has been dated as Early Thanetian to Early Eocene in age by reference to well 164/25-1 (Fig. 3.10). The unconformity is marked by onlap of the Early Eocene Balder Formation onto Early Thanetian basalts. This hiatus in Late Thanetian sediment may be due to erosion as a result of uplift from the inversion of the West Lewis Basin. Consequently the unconformity, in this study, has been interpreted as Thanetian in age based on a Late Paleocene unconformity marking inversion in the Faroe-Shetland Basin (Nicholson, 2005). Late Paleocene - Early Eocene compression has also been previously recorded in the Rockall-Faroe area (Boldreel and Andersen, 1993; Johnson et al., 2005). The Base Balder unconformity [Late Paleocene - Early Eocene (Jolley et al., 2002)] has been dated and mapped in the southern Faroe-Shetland Basin (Smallwood and Gill, 2002). The unconformity in the south West Lewis Basin underlies the Balder Formation and could, therefore, represent the Base Balder unconformity present in the southern Faroe-Shetland Basin.

**Late Ypresian**

The late Ypresian, or the late Early Eocene horizon, has been dated by wells 164/25-1, 164/25-2, 164/07-1, 154/01-1 and 163/06-1 (Figs. 3.10 – 3.15). This horizon has also been dated, in previous work, across the Judd anticline, in the south Faroe-Shetland Channel (Smallwood, 2004). This previous interpretation was used to map the unconformity across the Judd anticline in this study. Here, the late Ypresian unconformity is marked by onlap and defines folding.

Onlap onto the late Ypresian unconformity is best expressed on the eastern limb of the Bridge Anticline, on the north limb of the Wyville-Thomson Ridge Anticline and within the Mordor Anticline (structures described in section 4.4). On the south limb of the Hatton Bank Anticline (structure described in section 4.4) there are two unconformities pre-dating the C30 unconformity, one of which could represent a late Ypresian surface. Outwith the NE section of the Rockall-Faroe area and the south limb of the Hatton Bank Anticline, the late Ypresian unconformity was not identified due to poor seismic penetration.
Late Lutetian

The late Lutetian horizon has been dated by wells 164/25-1, 164/25-2, 164/07-1, 154/01-1 and 163/06-1 (Figs. 3.10 – 3.15). This unconformity has also been dated, in previous work on the Judd Anticline (Smallwood, 2004), located in the south Faroe-Shetland Channel, where it is marked by onlap. The unconformity is also characterized by onlap in the Mordor, Ymir Ridge (North, Central and South), Wyville-Thomson Ridge and the Bridge anticlines (structures described in section 4.4). One of the post-C30 unconformities on the southern limb of the Hatton Bank Anticline may represent the late Lutetian unconformity. Poor seismic penetration in other areas, such as the North Rockall Trough, precludes the mapping and dating of this unconformity.

C30 (Late Eocene)

The C30 unconformity has been previously dated and mapped in the south Rockall Trough by Stoker et al. (2001). In the south Rockall Trough the C30 unconformity is marked by onlap (Fig. 3.8). The C30 unconformity has been dated by wells 164/25-1, 164/25-2, 164/07-1, 154/01-1 and 163/06-1 in the NE Rockall Trough (Figs. 3.10 – 3.15). The age of the unconformity has also been constrained by boreholes 117 (Fig. 3.16) and 94/01 (Fig. 3.17) in the Hatton-Rockall Basin and on the east margin of the Rockall Bank, respectively. These wells and boreholes and the presence of good quality seismic data, allowed the C30 stratigraphic surface to be mapped in areas outwith the South Rockall Trough, in the Rockall-Faroe study area. This unconformity has also been dated in previous work in the Faroe-Shetland Basin (Davies and Cartwright, 2002). The C30 horizon is a major angular unconformity marked by onlap on the Hatton Bank, Alpin, Mordor, Ymir Ridge and Wyville-Thomson Ridge, Onika, and Viera anticlines (structures described in section 4.4). The C30 unconformity also defines the age of reverse faulting on the south limb of the Hatton Bank and the north West Lewis Basin (structures described in section 4.4).
Early Oligocene

The Early Oligocene horizon has been dated using wells 164/25-1 and 164/25-2 in the NE Rockall Trough (Figs. 3.11 – 3.12). Borehole 116 (Fig. 3.16 and Fig. 3.21) has revealed an unconformity between Upper and Lower Oligocene (top Lower Oligocene) within the Hatton-Rockall Basin. This unconformity, in the Hatton-Rockall Basin, has been inferred as a composite surface, comprising the Early Oligocene unconformity and a later early Late Oligocene unconformity (Fig. 3.21). The Early Oligocene unconformity is marked by onlap on the south limb of the Central Ymir Anticline and in the Alpin and Mordor anticlines (structures are described in section 4.4). On the north limb of the Wyville-Thomson Ridge, an inferred Early Oligocene and early Late Oligocene composite unconformity is marked by strong onlap (section 4.4).

Early Late Oligocene

This unconformity has been dated using well 164/25-2 in the NE Rockall Trough. Here, the stratigraphic surface is marked by onlap of Late Oligocene shale onto Early Oligocene sand (Figs. 3.12 and 3.20). The early Late Oligocene unconformity is also marked by onlap in the North Rockall Trough (described in section 5.4). Previous work has revealed a Top Palaeogene erosional unconformity within the Faroe-Shetland Channel (Smallwood, 2004). In this study, the Top Palaeogene unconformity is interpreted as early Late Oligocene in age. This interpretation was used in the mapping of an Early Oligocene/early Late Oligocene composite unconformity on the north limb of the Wyville-Thomson Ridge.

C20 (late Early Miocene)

The C20 unconformity was dated using borehole 610 (Fig. 3.19) in the southern Rockall Trough outwith the study area (Ruddiman et al., 1994). Here, the unconformity is marked by the onlap of a smectite layer (Dolan, 1986) which extends into the South Rockall Trough (Fig. 3.8). The C20 smectite layer appears to be absent in the NE and North Rockall Trough and the Hatton-Rockall Basin. In the NE Rockall Trough, the C20 stratigraphic surface has been constrained using wells 164/25-1 and 164/25-2 (Figs. 3.11 – 3.12). In the North Rockall Trough, the C20
unconformity is marked by onlap and erosional truncation (described in section 5.4). In the Hatton-Rockall Basin the C20 horizon, marked by onlap and erosional truncation on the south limb of the Hatton Bank Anticline (described in section 4.4), has been dated using borehole 116 (Fig. 3.16).

**Late Miocene – Early Pliocene**

An unconformity is mapped between the C20 (late Early Miocene) and C10 (late Early Pliocene) unconformities in the NE Rockall Trough. The age of this unconformity has been inferred as Late Miocene – Early Pliocene based on previous work in the Faroe-Shetland Channel (Davies and Cartwright, 2002) and its position between the C20 and C10 unconformities. This unconformity is characterized by erosional truncation along the eastern margin of the Rockall Bank (Fig. 3.8) and the northern margin of the Alpin Anticline (described in section 5.4). In the NE Rockall Trough, the C20 unconformity has been truncated by the Late Miocene- Early Pliocene unconformity (Fig. 3.15).

**C10 (late Early Pliocene)**

The C10 unconformity, in this study, has been dated using well 164/25-2 in the NE Rockall Trough, borehole 116 in the Hatton-Rockall Basin and borehole 94/01 in the Central Rockall Trough. In the Faroe-Shetland Channel, an Intra Neogene unconformity (INU) have been mapped previously (STRATAGEM partners, 2002; Ritchie et al., 2003) and may correlate with the C10 unconformity in the Rockall Trough. This unconformity is marked by erosional truncation and onlap within the study area. Erosional truncation is evident in the Hatton-Rockall Basin (described in section 5.4) and in the NE Rockall Trough (Fig. 3.13).

**Top Basalt**

The top-basalt horizon is a very strong reflector on seismic data throughout the Rockall-Faroe study area. In this study, the basalt surface has been dated by wells 154/01-1, 163/06-1, 164/07-1, 164/25-1 and 164/25-2 in the NE Rockall Trough (Fig. 3.20) and borehole 117 in the Hatton-Rockall Basin (Fig. 3.21). The wells and the borehole reveal a top-basalt surface age of Late Paleocene – Early Eocene.
Table 3.1. Characteristics of stratigraphic surfaces dated by well 164/25-1 in the NE Rockall Trough. Well data courtesy of British Petroleum (BP).

<table>
<thead>
<tr>
<th>Well borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Depth of Reflector from drilling platform (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>164/25-1</td>
<td>Pliocene-Early Miocene</td>
<td>815</td>
<td>1258</td>
<td>Moderately strong seismic reflector marked by onlap and erosion.</td>
<td>Unconformity marks the deposition of Pliocene sediments directly on Early Miocene sediments (see Fig. 3.20). The Mid to Late Miocene hiatus may be the result of erosion due to uplift or bottom-water current activity. Mapping suggests a Late Miocene-Early Pliocene/C10 (early Late Pliocene) composite unconformity at the well site.</td>
</tr>
<tr>
<td>Early Miocene-Early Oligocene</td>
<td>823</td>
<td>1266</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
<td>Unconformity marks the deposition of Early Miocene sediments directly on Early Oligocene sediments (see Fig. 3.20). Mapping has revealed this stratigraphic surface as an early Late Oligocene/C20 (late Early Miocene) composite unconformity.</td>
<td></td>
</tr>
<tr>
<td>Early Oligocene</td>
<td>877</td>
<td>1320</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
<td>Unconformity defines fold growth of the Alpin and Ymir Ridge anticlines.</td>
<td></td>
</tr>
<tr>
<td>Late Eocene (C30)</td>
<td>923</td>
<td>1366</td>
<td>Strong seismic reflector marked by onlap.</td>
<td>Unconformity defines fold growth of Alpin, Mordor, Ymir Ridge and Viera anticlines.</td>
<td></td>
</tr>
<tr>
<td>Late Lutetian</td>
<td>982</td>
<td>1425</td>
<td>Forms a relatively weak seismic reflector.</td>
<td>Unconformity marks folding of the Ymir Ridge, Bridge and Mordor anticlines.</td>
<td></td>
</tr>
<tr>
<td>164/25-1</td>
<td>Late Ypresian</td>
<td>1178</td>
<td>1621</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
<td>This unconformity marks folding in the Dawn and North Ymir Ridge anticlines.</td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>Early Thanetian – Early Ypresian (Top basalt)</td>
<td>1388</td>
<td>1831</td>
<td>Strong seismic reflector marked by onlap.</td>
<td>This angular unconformity represents a hiatus between Early Thanetian basalt and Ypresian sediment (see Fig. 3.20). The unconformity may be equivalent to the Base Balder Formation unconformity in the south Faroe-Shetland Basin (Smallwood and Gill, 2002) and correlate with an unconformity observable within the Wyville-Thomson Ridge basalt succession.</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 3.10. Interpreted seismic reflection profile in the NE Rockall Trough. Stratigraphic surfaces constrained using well 164/25-1. For location of seismic line and well see Fig. 3.6. Seismic data courtesy of CGG Veritas.
Fig. 3.11. Interpreted seismic reflection profile in the NE Rockall Trough. Stratigraphic surfaces constrained using well 164/25-1. For location of seismic line and well see Fig.3.6. Seismic data courtesy of CGG Veritas.
<table>
<thead>
<tr>
<th>Well/ borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Depth of Reflector from drilling platform (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>164/25-2</td>
<td>late Early Pliocene (C10)</td>
<td>539</td>
<td>1236</td>
<td>Forms a strong seismic reflection characterised by downlap at the well and erosional truncation and onlap away from the well.</td>
<td>Stratigraphic surface characterized by the downlap of the Sula Sgeir fan.</td>
</tr>
<tr>
<td></td>
<td>early Late Oligocene</td>
<td>793</td>
<td>1490</td>
<td>Forms a strong seismic reflection showing onlap. Gamma ray spike, at this horizon, due to an abrupt change from sand to shale.</td>
<td>This unconformity marks the deposition of Late Oligocene shale onto Early Oligocene sand (see Fig. 3.20).</td>
</tr>
<tr>
<td></td>
<td>Early Oligocene</td>
<td>833</td>
<td>1530</td>
<td>Forms a moderately strong seismic reflection marked by onlap.</td>
<td>Unconformity defines fold growth of the Alpin, Mordor and Ymir Ridge anticlines.</td>
</tr>
<tr>
<td></td>
<td>Late Eocene (C30)</td>
<td>1153</td>
<td>1850</td>
<td>Forms a very strong amplitude reflector marked by onlap.</td>
<td>Unconformity defines fold growth of the Alpin, Mordor, Ymir Ridge, Onika and Viera anticlines.</td>
</tr>
<tr>
<td></td>
<td>Late Lutetian</td>
<td>1233</td>
<td>1930</td>
<td>Forms a relatively weak seismic reflector.</td>
<td>Unconformity defines fold growth of the Ymir Ridge, Bridge and Mordor anticlines.</td>
</tr>
<tr>
<td></td>
<td>Late Ypresian</td>
<td>1359</td>
<td>2056</td>
<td>Forms a moderately strong seismic reflector characterised by onlap.</td>
<td>This unconformity defines fold growth of the Dawn and North Ymir Ridge anticlines.</td>
</tr>
<tr>
<td></td>
<td>Top basalt</td>
<td>1493</td>
<td>2190</td>
<td>Very strong seismic reflector.</td>
<td>Parallel bedded layers evident beneath top basalt surface. Surface can be mapped throughout the Rockall-Faroe area.</td>
</tr>
</tbody>
</table>

Table 3.2. Characteristics of stratigraphic surfaces dated by well 164/25-2 in the NE Rockall Trough. Well data courtesy of BP.
Fig. 3.12. Interpreted seismic reflection profile in the NE Rockall Trough. Stratigraphic surfaces constrained using well 164/25-2. For location of seismic line and well see Fig. 3.6. Seismic data courtesy of CGG Veritas.
<table>
<thead>
<tr>
<th>Well/ borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>below sea bed (m)</td>
<td>from drilling platform (m)</td>
<td></td>
</tr>
<tr>
<td>164/7-1</td>
<td>Late Eocene (C30)</td>
<td>796</td>
<td>1590</td>
<td>Strong seismic reflector marked by onlap.</td>
</tr>
<tr>
<td></td>
<td>late Middle Eocene</td>
<td>1042</td>
<td>1836</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
</tr>
<tr>
<td></td>
<td>late Early Eocene</td>
<td>1158</td>
<td>1952</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
</tr>
<tr>
<td></td>
<td>Top basalt</td>
<td>1241</td>
<td>2035</td>
<td>Very strong seismic reflector.</td>
</tr>
</tbody>
</table>

This unconformity can be mapped to wells 164/25-1 and 164/25-2. Unconformity defining folding of the Alpin, Mordor, Ymir Ridge and Viera anticlines.

This unconformity can be mapped to well 164/25-2 where it is dated as late Lutetian.

This unconformity can be mapped to well 164/25-2 where it is dated as late Ypresian.

Parallel bedding within basalt lava flows evident beneath top-basalt surface.

Table 3.3. Characteristics of stratigraphic surfaces dated by well 164/7-1 in the NE Rockall Trough. Well data courtesy of Conoco (UK) Limited.
Fig. 3.13. Interpreted seismic reflection profile in the NE Rockall Trough. Stratigraphic surfaces constrained using well 164/07-1. For location of seismic line and well see Fig. 3.6. Seismic data courtesy of CGG Veritas.
<table>
<thead>
<tr>
<th>Well/borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>154/01-1</td>
<td>Early Oligocene to Late Tertiary</td>
<td>69, 1115</td>
<td>Strong seismic reflector.</td>
<td>This reflector is the C10 unconformity of Stoker et al. (2001) and can be mapped to wells 164/25-1 and 164/25-2.</td>
</tr>
<tr>
<td></td>
<td>Late Eocene (C30)</td>
<td>181, 1227</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
<td>This unconformity can be mapped to wells 164/25-1 and 164/25-2.</td>
</tr>
<tr>
<td></td>
<td>Late Middle Eocene</td>
<td>327, 1373</td>
<td>Strong seismic reflector marked by onlap.</td>
<td>This unconformity can be mapped to well 164/25-2 where it is dated as late Lutetian.</td>
</tr>
<tr>
<td></td>
<td>Late Early Eocene</td>
<td>684, 1730</td>
<td>Moderately strong seismic reflector marked by onlap.</td>
<td>This unconformity can be mapped to well 164/25-2 where it is dated as late Ypresian.</td>
</tr>
<tr>
<td></td>
<td>Top basalt</td>
<td>1009, 2055</td>
<td>Moderately strong reflector.</td>
<td>Top-basalt surface can be mapped throughout the Rockall-Faroe area.</td>
</tr>
</tbody>
</table>

Table 3.4. Characteristics of stratigraphic surfaces dated by well 154/01-1 in the NE Rockall Trough. Well data courtesy of Enterprise Oil.
Fig. 3.14. Interpreted seismic reflection profile in the NE Rockall Trough. Stratigraphic surfaces constrained using well 154/01-1. For location of seismic line and well see Fig. 3.6. Seismic data courtesy of CGG Veritas.
<table>
<thead>
<tr>
<th>Well/ borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Depth of Reflector from drilling platform (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>163/6-1</td>
<td>late Early Pliocene (C10)</td>
<td>140</td>
<td>1529</td>
<td>Strong seismic reflector characterised by erosional truncation</td>
<td>The age of this unconformity was constrained by wells 164/25-1 and 164/25-2. Erosional truncation would suggest the unconformity was formed as a result of bottom-water current activity.</td>
</tr>
<tr>
<td></td>
<td>Late Miocene-Early Pliocene</td>
<td>180</td>
<td>1569</td>
<td>Strong seismic reflector characterised by erosional truncation</td>
<td>Erosional truncation would suggest bottom-water current activity.</td>
</tr>
<tr>
<td></td>
<td>Early Oligocene</td>
<td>260</td>
<td>1649</td>
<td>Strong seismic reflector marked by onlap</td>
<td>The exact age of this unconformity was constrained by wells 164/25-1 and 164/25-2.</td>
</tr>
<tr>
<td></td>
<td>Late Eocene (C30)</td>
<td>460</td>
<td>1849</td>
<td>Strong seismic reflector marked by onlap</td>
<td>Unconformity defines folding of the Alpin, Mordor and the Ymir Ridge anticlines.</td>
</tr>
<tr>
<td></td>
<td>Late Mid Eocene</td>
<td>766</td>
<td>2155</td>
<td>Moderately strong seismic reflector marked by onlap</td>
<td>Unconformity defines folding of the Alpin, Mordor and the Ymir Ridge anticlines.</td>
</tr>
<tr>
<td></td>
<td>Late Early Eocene</td>
<td>1059</td>
<td>2448</td>
<td>Moderately strong seismic reflector marked by onlap</td>
<td>Unconformity defines folding of the Alpin, Mordor and the North Ymir Ridge anticlines.</td>
</tr>
<tr>
<td></td>
<td>Top basalt</td>
<td>1241</td>
<td>2630</td>
<td>Very strong seismic reflector</td>
<td>Parallel bedding within basalt lava flow evident beneath top-basalt surface. Surface can be mapped throughout the Rockall-Faroe area.</td>
</tr>
</tbody>
</table>

Table 3.5. Characteristics of stratigraphic surfaces dated by well 163/6-1 in the NE Rockall Trough. Well data courtesy of British National Oil Corporation.
Fig. 3.15. Interpreted seismic reflection profile in the NE Rockall Trough. Stratigraphic surfaces constrained using well 163/061-1. For location of seismic line and well see Fig. 3.6. Seismic data courtesy of CGG Veritas.
<table>
<thead>
<tr>
<th>Well/ borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>116</td>
<td>Early Pliocene – Late Pliocene (C10)</td>
<td>60</td>
<td>Forms a low amplitude conformable reflection at the well location. Horizon defined by erosional truncation between the Hatton-Rockall Basin and the North Rockall Trough (see Fig. 6.22).</td>
<td>Erosional truncation away from well site can be attributed to bottom-water current activity.</td>
</tr>
<tr>
<td>Late Miocene–Early Pliocene</td>
<td>160</td>
<td>High amplitude reflector.</td>
<td>Unconformity inferred as forming a composite unconformity with the C10 (late Early Pliocene) surface in the north Hatton-Rockall Basin.</td>
<td></td>
</tr>
<tr>
<td>Early Miocene – Middle Miocene (C20)</td>
<td>530</td>
<td>Forms a moderate amplitude conformable reflection at the well location. Erosional truncation and channels define the unconformity on the south limb of the Hatton Bank.</td>
<td>Erosional truncation and channels can be attributed to bottom-water current activity.</td>
<td></td>
</tr>
<tr>
<td>Early Oligocene – Late Oligocene</td>
<td>750</td>
<td>Strong amplitude reflector at the well site.</td>
<td>Unconformity interpreted as an Early Oligocene/early Late Oligocene composite unconformity (see Fig. 3.21). The Early Oligocene unconformity defines fold growth of the Hatton Bank. The early Late Oligocene unconformity is characterised by erosional truncation on the south limb of the Hatton Bank attributed to bottom-water current activity.</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Well/ borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>117</td>
<td>Early Eocene – Oligocene</td>
<td>140</td>
<td>Very strong amplitude reflector.</td>
<td>This unconformity is inferred as the C30 (Late Eocene)/Early Oligocene composite unconformity in the Hatton-Rockall Basin (see Fig. 3.21).</td>
</tr>
<tr>
<td></td>
<td>Top basalt</td>
<td>311</td>
<td>Strong amplitude reflector.</td>
<td>Top-basalt surface defines the Rockall Bank</td>
</tr>
</tbody>
</table>

Fig. 3.16. Interpreted seismic reflection profile in the Hatton-Rockall Basin. Stratigraphic surfaces constrained using boreholes 116 and 117. For location of seismic line and boreholes see Fig. 3.7.
<table>
<thead>
<tr>
<th>Well/borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>94/01</td>
<td>late Early Pliocene (C10)</td>
<td>17</td>
<td>Strong seismic reflector marked by erosional truncation.</td>
<td>This unconformity separates Lower Pliocene sediments from Pliocene/Holocene sediments (Fig. 3.17). The erosional truncation is most likely due to bottom-current activity.</td>
</tr>
<tr>
<td></td>
<td>Late Eocene – Late Miocene</td>
<td>33</td>
<td>Strong seismic reflector marked by onlap.</td>
<td>This unconformity was mapped as C30 (Late Eocene) by Stoker et al. (2001) in the Central Rockall Trough where it defines the dramatic onlap of Oligocene sediments onto Eocene sediment (Fig. 3.17)</td>
</tr>
</tbody>
</table>

Fig. 3.17. Interpreted seismic reflection profile in the east margin of the Rockall Bank. Stratigraphic surfaces constrained using borehole 94/01. For location of seismic line and borehole see Fig. 3.7.
Fig. 3.18. Interpreted seismic reflection profile in the northeast margin of the Rockall Bank. Stratigraphic surfaces constrained using boreholes 94/04. For location of seismic line and borehole see Fig. 3.7.
<table>
<thead>
<tr>
<th>Well/borehole</th>
<th>Stratigraphic Surface</th>
<th>Depth of Reflector below sea bed (m)</th>
<th>Seismic Character</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>610</td>
<td>late Early Miocene</td>
<td>675</td>
<td>Very strong seismic reflector.</td>
<td>Reflector marked by a rich smectite layer (Dolan, 1986). This characteristic has been used to map the C20 unconformity in the South Rockall Trough.</td>
</tr>
</tbody>
</table>

Table 3.9. Characteristics of the late Early Miocene (C20) unconformity dated by borehole 610 in the South Rockall Trough. Borehole data courtesy of the Deep Sea Drilling Project.
Fig. 3.19. Interpreted seismic reflection profile in the South Rockall Trough. Stratigraphic surface constrained using borehole 610. For location of seismic line and boreholes see Fig. 3.4. Seismic data courtesy of WesternGeco.
Fig. 3.20. Lithologies present in wells in the NE Rockall Trough. The positions of stratigraphic surfaces are also shown. The depth represents the distance below the sea surface. For the location of these wells see Fig. 3.5.
Fig. 3.21. Lithologies present in boreholes (a) 116 and (b) 117 in the Hatton-Rockall Basin. The depth marks the distance below the sea bed. Note the composite unconformities in both boreholes. For the location of these boreholes see Fig. 3.6.
3.4 Folding and the formation of unconformities

Compression can result in the tilting of sediment strata (Fig. 3.22). Subsequent younger sediments onlap the tilted sediment resulting in the formation of an angular unconformity. Angular unconformities continue to develop recording each compressional event.

Fig. 3.22. Development of angular unconformities as a result of folding. (a) unfolded sediment strata (b) onlap of younger sediments onto folded sediment strata defines an angular unconformity (c) a second folding event results in the formation of another angular unconformity.

3.5 Uncertainties in the dating and mapping of unconformities

Wells and boreholes which were used to date the stratigraphic surfaces are located in the NE Rockall Trough and the Hatton Rockall Basin (Fig. 3.4). Further constraints on the mapping of stratigraphic surfaces were made using boreholes located on the east margin of the Rockall Bank (Fig. 3.4). The certainty in the dating and mapping of stratigraphic surfaces throughout the Rockall-Faroe area depends on the location of these wells/boreholes, the presence and quality of seismic data and the fidelity of well ties between both. The quality and quantity of well and seismic data available in the study area is generalized in Fig. 3.23. The level of confidence in mapping stratigraphic horizons varies across the Rockall-Faroe study area.
Fig. 3.23. Seismic and well data in the Rockall-Faroe area. Generalized levels of uncertainty in the dating and mapping of stratigraphic surfaces in sediment are superimposed onto the map. Structural highs are blank as there is little sediment over these areas for much seismic interpretation of stratigraphic surfaces. Abbreviation: AB = Auðhumla Basin

KEY
- good seismic data quality and well control
- good seismic quality and poor well control
- poor seismic quality and lack of well data
- lack of seismic and well data
**NE Rockall Trough**

The NE Rockall Trough has the highest concentration of wells in the study area. Wells have yielded ages for stratigraphic surfaces from the Thanetian to the Pliocene. High quality and deep penetrating seismic data in the area have facilitated the mapping of the top-basalt surface and the late Ypresian, late Lutetian, C30 (Late Eocene), Early Oligocene, early Late Oligocene, C20 (late Early Miocene), Late Miocene-Early Pliocene and C10 (late Early Pliocene) unconformities throughout the NE Rockall Trough. These surfaces can also be mapped to proximal areas such as across the Mordor Anticline and the south limb of the Ymir Ridge Anticlines due to good quality seismic data between the NE Rockall Trough and these areas.

**North Rockall Trough**

The North Rockall Trough contains an extensive seismic data set with relatively poor imaging of pre-C30 sediment compared to the NE Rockall Trough. However, there are areas in the North Rockall Trough containing high resolution seismic data – south of the Lousy Bank and on the eastern margin of the Rockall Bank (Fig. 3.23). Although the C30 (Late Eocene) and C10 (late Early Pliocene) unconformities are constrained by borehole 94/01 on the eastern margin of the Rockall Bank, the North Rockall Trough lacks well data and the seismic quality to constrain stratigraphic surfaces identified in the NE Rockall Trough. An attempt has been made to map unconformities - C30 (Late Eocene), early late Oligocene, C20 (late Early Miocene), Late Miocene-Early Pliocene, and C10 (late Early Pliocene) - from the NE Rockall Trough to the North Rockall Trough using the available seismic data, of relatively poor quality.

**North of the Alpin Anticline, and the channels between the Lousy, Bill Bailey’s and Faroe Banks**

The C30 (Late Eocene) unconformity in these areas has been identified confidently based on the surface’s characteristically strong onlap within the study area. However, the seismic data available for these areas are scarce and of moderate quality and hence the confidence of other seismic picks is, at best, moderate. Interpretation of other stratigraphic surfaces, north of the Alpin Anticline – Early
Oligocene, early Late Oligocene, C20 (late Early Miocene), Late Miocene – Early Pliocene and C10 unconformities – is based on the characteristics and stratigraphic levels of these horizons in other areas such as the North and NE Rockall Trough. These criteria were also used to map unconformities in the channels between Lousy, Bill Bailey’s and Faroe Banks.

**Auðhumla Basin**

The Auðhumla Basin is separated from the NE Rockall Trough by the Bridge Anticline. Although there is extensive good quality seismic data in the area, there is onlap of the stratigraphic surfaces of the NE Rockall Trough onto the Bridge Anticline. Stratigraphic surfaces were thus mapped within the Auðhumla Basin by jump correlation from the NE Rockall Trough.

**Faroe Bank Channel**

Good quality seismic data is present across the Faroe Bank Channel. However, the lack of well data within the channel meant that the stratigraphic surfaces had to be inferred from the nature of unconformities present in the NE Rockall Trough. Interpretation of stratigraphic surfaces in the Faroe Bank Channel was also based on previous mapping in the south Faroe-Shetland Channel (Smallwood, 2004).

**Judd Anticline**

The area around the Judd Anticline lacks both well and seismic data. Interpretation of unconformities in this area is independent of unconformities mapped in the NE Rockall Trough, but is based on previous work by Smallwood et al. (2004).

**Hatton-Rockall Basin**

Unconformities were identified on the south limb of the Hatton Bank Anticline. These unconformities were then dated by correlating along seismic data to distal boreholes 116 and 117. Good quality seismic data facilitated the confident mapping of horizons at the south limb of the Hatton Bank Anticline to the borehole sites. The pre-C30 unconformities were not dated directly but inferred based on their positions between the top-basalt and C30 (Late Eocene) stratigraphic surfaces.
C20 (late Early Miocene) unconformity in the North Rockall Trough

The C20 unconformity has been dated in the South Rockall Trough, outwith the study area, at site 610 (Fig. 3.4). This unconformity was dated as late Early Miocene and characterised by a thick layer of overlying smectite. In this study, the late Early Miocene stratigraphic surface was dated and mapped in the NE Rockall Trough and its position on seismic data is in agreement with the interpretation by Tate et al. (1999). This stratigraphic surface, however, when mapped from the NE Rockall Trough into the North Rockall Trough, is stratigraphically higher than the previously assigned C20 layer in the North Rockall Trough (STRATEGEM partners, 2002). The C20 layer in the North Rockall Trough was previously mapped based on a thick layer which looked similar to the smectite layer present and dated in the South Rockall Trough. The smectite (in the South Rockall Trough), is believed to be originated from volcanic ash derived from the Norwegian margin (Dolan, 1986) by the Norwegian Sea Overflow (Stoker et al, 2001). If this ash layer is the result of Norwegian Sea overflow into the Rockall Trough, it is conceivable that in the North Rockall Trough conditions were more erosional than in the South Rockall Trough. Hence, deposition of the ash layer was more likely in the South Rockall Trough. There is a lack of good quality seismic data in the North and Central Rockall Trough, in this study, to confidently link the newly mapped C20 stratigraphic surface in the North Rockall Trough with the C20 ash layer in the South Rockall Trough.

3.6 Summary

Well data, coupled with an existing seismic data set, have been used to date and map stratigraphic surfaces in the Rockall-Faroe area. The presence of good quality seismic data has facilitated the mapping of well dated unconformities, at the well sites, to more distal areas. Where there is a lack of connecting seismic data to the well site, an attempt has been made to infer the horizons based on the distribution and nature unconformities present in the NE Rockall Trough. The stratigraphic surfaces dated and mapped in the Rockall-Faroe area are essential in establishing the ages of Cenozoic compressional and non-compressional events.
4.0 Inversion Features

4.1 Introduction
Over the past two decades there has been an increasing realisation of the general role that structural inversion may have played in the tectonic development of the NW European Shelf in general and the NE Atlantic passive continental margin in particular (Boldreel and Andersen, 1993; Doré and Lundin, 1996; Ritchie et al., 2003; Johnson et al., 2005).

With the quality of seismic data and the quantity of wells available, it is now possible to calibrate and define the specific effects of compression. The major aim of this chapter is to highlight the role of structural inversion in the NE Atlantic Margin as a major mechanism in the formation of unconformities and folds.

In the NE Atlantic Margin horizontal compressional stresses have affected Cenozoic basalt lava flows and younger sediments. Within the study area, these stresses resulted in the formation of compressional structures which vary in shape, size and in orientation (Figs. 4.1 – 4.2). Most of these structures influence the bathymetry of the sea bed and are also delineated on isochron maps of horizons.

Compressional structures identified in this study mainly comprise anticlines and synclines, although a number of reverse faults have also been recognized. Elongated folded compressional structures are termed anticlines (positive relief) and synclines (negative relief). These structures have defined fold axes which can be mapped using seismic data.

4.2 Dataset
An extensive 2-D seismic dataset was used to visualize compressional structures in the study area. Use has been made of 40,000 km of 2-D seismic data (line-length) acquired over the past twenty years extending from the Rockall Plateau to the Faroe Shelf (see Fig. 3.4). These seismic lines were provided by BGS, Fugro Multi Client Services, CGG Veritas, WesternGeco and GEUS. For more details on the seismic data available in this study, refer to section 1.5.1.
Fig. 4.1. Location of compressional features in the Rockall-Faroe area. Note the varying lengths and orientations of folds. Faults are defined at the Palaeogene top-basalt surface. Locations of seismic illustrations shown by dotted black lines. Bathymetric contours (in metres) are represented by blue lines.
Fig. 4.2. Compressional structures from the Rosemary Bank to the Faroe Shelf. Locations of seismic illustrations are shown by black lines.
4.3 Features of Folds

A fold can be described by its amplitude and width. A fold's amplitude is the height of the fold from its crest to the trough of the adjacent syncline measured from a particular stratigraphic surface which best defines the fold. The amplitude of a fold, which defines the sea bed, is derived from the addition of the depths of water and sediment. The depths are calculated from the two-way time of water (TWT\textsubscript{w}) and sediments (TWT\textsubscript{s}) in the adjacent syncline. Where the fold is completely covered by sediment a simple depth conversion from the two-way time of sediment (TWT\textsubscript{s}) is required.

![Illustration of the methodology used in the calculation of fold amplitude. The two-way time intervals of water (TWT\textsubscript{w}) and sediment (TWT\textsubscript{s}) are depth-converted. The addition of these depth-converted thicknesses gives the amplitude of the fold.](image)

In this study, shortening across some compressional structures have yielded values of \(\leq 1.0\%\). The amount of shortening is derived for individual structures using the formula:

\[
\left[ \frac{(L_0 - L_f)}{L_0} \right] \times 100
\]

where \(L_f\) = length after shortening and \(L_0\) = original length

It should be noted that this shortening may have been increased by the effects of sediment loading, especially where adjacent synclines to anticlines are filled with thick sediments. This small degree of shortening has produced open-gentle folds of large widths in relation to their amplitudes resulting in long-wavelength structures with interlimb angles > 90\(^\circ\). The width of a fold is measured as the distance between the trough centres on either side of the fold. Folds in the study area are symmetrical and asymmetrical. Symmetrical folds have limbs of similar dip angles while dip angles of the limbs of asymmetrical folds are different. In the study area
asymmetrical folds have developed on the hanging walls of reverse faults and are termed fault-propagation folds.

Gravity modelling has been used to distinguish folds from seamounts and horst blocks. Seamounts and horst blocks consist of rocks of higher density and thus give higher gravity anomalies than folds which generally consist of lower density sediment. Reverse faults also suggest folding as a result of compression. It should be noted that sediment loading and differential subsidence could also produce apparent folded strata. These mechanisms are explained in detail in Section 4.7.

4.4 Compressional Structures
A number of compressional structures have been mapped within the study area (Fig. 4.1). The names of some compressional structures are new, whilst others have been derived from previous work. Table 4.1 shows the names of structures used in this study and the previous names of these structures. Structures located on the western and northern limits of the study area described from west to east are:

- Hatton Bank
- Lousy Bank
- Bill Bailey’s Bank
- Faroe Bank – North and South
- Faroe Bank Channel Syncline
- Vine
- Judd

The descriptions of these structures are followed by the descriptions of more inner compressional structures from west to east:

- Alpin
- Dawn
- Mordor
- Ymir Ridge – North, Central, South
- Wyville-Thomson Ridge
- Auðhumla Basin
- Bridge
- Onika
- West Lewis reverse faults
- Viera

<table>
<thead>
<tr>
<th>This Study</th>
<th>Previous variations</th>
</tr>
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<tbody>
<tr>
<td>Hatton Bank Anticline</td>
<td>Hatton Bank (Boldreel and Andersen, 1994; Stoker, 2002; Doré et al., 1997; Kimbell et al., 2005, McInroy et al., 2006)</td>
</tr>
<tr>
<td>Lousy Bank Anticline</td>
<td>Lousy Bank (Boldreel and Andersen, 1994, 1998; Doré et al., 1997; Stoker, 2002)</td>
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<tr>
<td>North Faroe Bank Anticline</td>
<td>Unnamed (Boldreel and Andersen, 1998).</td>
</tr>
<tr>
<td>South Faroe Bank Anticline</td>
<td>Faroe Bank (Boldreel and Andersen, 1994, 1998; Doré et al., 1997; Stoker, 2002)</td>
</tr>
<tr>
<td>Faroe Bank Syncline</td>
<td>Faroe Conduit (Stoker et al., 2005b)</td>
</tr>
<tr>
<td>Vine Anticlines</td>
<td>Unnamed (Johnson et al., 2005)</td>
</tr>
<tr>
<td>Judd Anticline</td>
<td>Judd Inversion Anticline (Smallwood et al., 2004), Judd High (Sørensen, 2003), mapped but not named (Boldreel and Andersen, 1998)</td>
</tr>
<tr>
<td>Alpin Anticline</td>
<td>Alpin Dome (Johnson et al., 2005, Stoker et al., 2005a; Stoker et al., 2005c)</td>
</tr>
<tr>
<td>Mordor Anticline</td>
<td>Unnamed (Johnson et al., 2005)</td>
</tr>
<tr>
<td>Ymir Ridge Anticlines – North, Central and South.</td>
<td>Ymir Ridge (Tate et al., 1999, Boldreel and Andersen, 1998; Johnson et al., 2005; Stoker et al., 2005a)</td>
</tr>
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Isochron maps and present-day bathymetry, in addition to seismic data, have been used to define the orientation and extent of compressional features in the study area. The characteristics of compressional structures are herein described based on observations and interpretations from seismic data.

**Hatton Bank Anticline**

The Hatton Bank Anticline (Figs. 4.4 – 4.7) represents a NE-trending anticline with a width of ~ 40 km and an amplitude of 1.1 km (measured from the top basalt surface). This anticline runs along ~ 210 km of the western edge of the Rockall Plateau continental margin parallel to the continent-ocean boundary (Fig. 4.1), and represents ~ 0.25 % shortening. It is characterised by the folding of basalt lava flows and overlying Eocene-Early Oligocene sediment (Fig. 4.4). To the south of the Hatton Bank Anticline lies the Hatton-Rockall Basin, containing Cenozoic sediments overlying basalt lava flows. Although there is a lack of internal reflectors in much of the succession directly overlying the top-basalt surface, there is some evidence of onlap onto the top-basalt surface (Fig. 4.5). This suggests that a structural high existed prior to the deposition of Ypresian sediment. Eocene sediment on Hatton Bank’s southern limb is cut by reverse faults (Fig. 4.5). The reverse faults have an E-W orientation, are 15-25 km in length and verge to the north and south. Reverse
Fig. 4.4. NNW-trending seismic profile across the Hatton Bank Anticline (see Fig. 4.1 for location).
Fig. 4.5. NNW-trending seismic profile of unconformities and reverse faults on the south limb of the Hatton Bank Anticline (for position, see Fig. 4.4). Note onlap of sediments onto the Intra-Eocene 1, Intra-Eocene 2, C30 (Late Eocene) and Early Oligocene unconformities. The early Late Oligocene and the C20 (late Early Miocene) unconformities are defined by onlap and erosional truncation. The early Late Oligocene unconformity is also marked by downlap. North and south verging reverse faults cut basalt lava flows and Eocene unconformities.
Fig. 4.6. NNW-trending seismic profile across the Hatton Bank Anticline (see Fig. 4.1 for location).
Fig. 4.7. NNW-trending seismic profile of unconformities and reverse faults on the south limb of the Hatton Bank Anticline (for position, see Fig. 4.6). Note the presence of a fault-propagation fold, underlain by a north verging reverse fault which cut basalt lava flows and Eocene unconformities. The Intra-Eocene 1, C30 (Late Eocene) and the Early Oligocene unconformities are marked by onlap. Erosional truncation defines the C10 (late Early Pliocene) unconformity.
faulting is inferred as Late Eocene (C30) in age based on the onlap of Early Oligocene sediments onto the C30 unconformity occurring on the hanging wall of one of the reverse faults (Fig. 4.5). Some reverse faults are also associated with fault propagation folds (Figs. 4.6 - 4.7). Unconformities defining folding in the Hatton Bank Anticline are the Intra-Eocene 1, the Intra-Eocene 2, the C30 (late Eocene) and the Early Oligocene unconformities (Figs. 4.5 and 4.7). The Intra-Eocene unconformities have been inferred based on the well-dated C30 (late Eocene) and Late Paleocene basalt stratigraphic surfaces. The Intra-Eocene unconformities may correspond to the late Ypresian and late Lutetian unconformities dated in the NE Rockall Trough. The early Late Oligocene, C20 (late Early Miocene) and C10 (early Late Pliocene) unconformities are inferred to be the result of bottom-water current activity because of erosional truncation associated with them (Fig. 4.5).

On the northern margin of the Hatton Bank Anticline there are inferred reverse faults, 1-8 km in length, with a NE-SW and an E-W orientation, imaged on the sea bed (Fig. 4.8). These faults are arranged en-echelon with apparent relay ramps bridged in some places (Fig. 4.8). Fault-propagation folds associated with these reverse faults are exposed on the sea bed as NE- and E-W trending ridges due to relatively little sediment cover on the western margin of the Hatton Bank Anticline (Fig. 4.9 and Fig. 4.10). Reverse faulting may have occurred at the time of the C30 (Late Eocene) faulting event on the southern limb of the Hatton Bank Anticline.

Most of the Hatton Bank area is covered with Paleocene – Early Eocene basalt lava. The high acoustic impedance of the basalt lava results in poor imaging of pre-basalt sediments (Maresh and White, 2005). However, in some areas, where basalt lava is absent (basalt windows) the underlying sediments can be imaged. To the southwest of the Hatton Bank Anticline, for example, there is an ENE-trending reverse fault, 20 km in length imaged beneath a basalt window (Fig. 4.11). This fault appears inverted as it is associated with a pop-up structure on its hanging wall. The inversion structure has a maximum width of ~ 8 km and the uplift of such a structure may have exposed the overlying basalt to erosion. There is a syncline, 50 km south of the reverse fault also located within a basalt window (Fig. 4.12). The syncline, like the reverse fault is ENE- trending. It is 15 km in length and has an amplitude of ~ 200
Fig. 4.8. Swath bathymetric map of an area on the north limb of the Hatton Bank Anticline (for location, see Fig. 4.1). Orientation of inferred reverse faults delineated by shallower bathymetry. Location of illustrated seismic lines shown by solid black lines.
Fig. 4.9. SW-NE seismic profile of sea bed ridges on the north limb of the Hatton Bank Anticline (for location, see Fig. 4.8). The ridges are inferred to be fault-propagation folds underlain by SE-trending reverse faults which cut the top basalt surface. Seismic data courtesy of GEUS.
Fig. 4.10. NW-SE seismic profile of sea bed ridges on the north limb of the Hatton Bank Anticline (for location, see Fig. 4.8). The ridges are inferred to be carbonate mounds which overlie fault-propagation folds underlain by SE trending reverse faults which cut the top basalt surface.
Fig. 4.11. E-W seismic profile across a basalt window on the Hatton Bank (for location, see Fig. 4.1). Note the contrast between the well layered sediment strata within the basalt window and the lack of imaging beneath the top basalt surface. The pop-up structure at the basalt window is inferred to be underlain by a SE verging reverse fault. Sediments within this structure are inferred as Mesozoic in age based on previous work by Hitchen (2004).
Fig. 4.12. SSW-NNE seismic profile of a syncline within a basalt window on the Hatton Bank (for location, see Fig 4.1). Note the erosional unconformity between the inferred Mesozoic dipping strata and the overlying sediment with horizontal layers. An unconformity is also present within the Mesozoic sediment and is marked by onlap. The Mesozoic age of sediment in this basalt window is based on previous work by Hitchen (2004).
m. An unconformity, marking onlap within the syncline, may mark a phase of pre-Eocene folding.

Lousy Bank Anticline
The Lousy Bank anticline (Fig. 4.13) is a domal structure lying between the Bill Bailey’s Bank and Hatton Bank anticlines (Fig. 4.1). Like the Hatton Bank Anticline, it is parallel and juxtaposed to the continent-ocean boundary, trending in a NE direction. The anticline is 125 km long, has a maximum width of 95 km, an amplitude of up to 1.9 km (measured from the top-basalt surface) and represents 0.5% shortening. The anticline is marked by folding of basalt lava flows and the onlap of sediments onto the C30 (Late Eocene) unconformity (Fig. 4.14). On the southeast limb of the anticline, in the Lousy Bailey Channel (Fig. 4.15), the C30 (Late Eocene), early Late Oligocene, C20 (late Early Miocene) and Late Miocene-Early Pliocene/C10 unconformities have been mapped. Here, the C30 (late Eocene), early Late Oligocene and C20 (late Early Miocene) unconformities are marked by onlap while the Late Miocene-Early Pliocene/C10 composite unconformity is marked by erosional truncation (Fig. 4.16). The C30 unconformity is inferred to be the result of compression, while the early Late Oligocene and the C20 unconformities, and the Late Miocene-Early Pliocene/C10 composite unconformity are the result of bottom-water current activity.

Bill Bailey’s Bank Anticline
The Bill Bailey’s Bank Anticline (Fig. 4.17) is a domal fold lying between the Faroe Bank and Lousy Bank anticlines (Fig. 4.1). It is NW-trending, and its axis is almost co-linear to the Dawn, the North Ymir Ridge and the Wyville-Thomson Ridge anticlines. The Bill Bailey’s Bank Anticline has a length of 100 km, a maximum width of 85 km, an amplitude of up to 1.7 km (measured from the top-basalt surface) and constitutes ~ 0.9% shortening. On the north limb of the anticline within the Bill Faroe Channel the early Late Oligocene, C20 (late Early Miocene), Late Miocene-Early Pliocene and C10 (late Early Pliocene) unconformities are marked by onlap (Fig. 4.18). The early Late Oligocene unconformity encloses a feature interpreted as a possible submarine fan.
Fig. 4.13. N-S seismic profile of the Lousy Bank Anticline (for location, see Fig. 4.1). Note the onlap of Oligocene and Miocene sediments onto the C30 (Late Eocene) unconformity in the North Rockall Trough. The C30 unconformity is eroded at the top of Lousy Bank.
Fig. 4.14. N-S seismic profile across the southeast limb of the Lousy Bank Anticline (for position, see Fig. 4.13). Note the onlap of sediments onto the C30 (Late Eocene) unconformity. The C10 (late Early Pliocene) unconformity truncates the Late Miocene - Early Pliocene unconformity.
Fig. 4.15. E-W seismic profile of the Lousy Bailey Channel (for location, see Fig. 4.1). Note the downlap of an Early Oligocene fan onto the C30 (late Eocene) unconformity on the western margin of the Lousy Bank Anticline. The early Late Oligocene and C20 (late Early Miocene) unconformities are marked by onlap. The interpretation of a pre-C30 diapir underlain by a normal fault which cuts the top-basalt surface is based on previous work by Vanneste et al. (1995).
Fig. 4.16. E-W seismic profile across the Lousy Bailey Channel (for position, see Fig. 4.15). Note the onlap of sediments onto the C30 (Late Eocene) unconformity on Bill Bailey’s Bank, the early Late Oligocene and the C20 (late Early Miocene) unconformities. The Late Miocene-Early Pliocene/C10 (late Early Pliocene) composite unconformity truncates the C20 unconformity. The interpretation of a diapir underlain by a normal fault which cuts the top-basalt surface is based on Vanneste et al. (1995).
Fig. 4.17. SW-NE seismic profile of the Bill Bailey’s Bank Anticline (for location, see Fig. 4.2). On the southwestern margin, the C30 (Late Eocene) unconformity is truncated at the top of the anticline.
Bill Bailey's Bank

Bill Faroe Channel

Faroe Bank

Fig. 4.18. E-W seismic profile of the Bill Faroe Channel (for location, see Fig. 4.2). Note the onlap of sediment on the early Late Oligocene, C20 (late Early Miocene), Late Miocene-Early Pliocene and C10 (late Early Pliocene) unconformities. The feature located between the early Late Oligocene unconformity and the top basalt is interpreted as a possible Early Oligocene (Rupelian) submarine fan.
Faroe Bank Anticlines

The Faroe Bank is interpreted as comprising two anticlines with different trends (Fig. 4.2). These anticlines are characterized by the folding of basalt lava flows and the onlapping of Eocene and younger sediments. The North Faroe Bank Anticline (Fig. 4.19) is 65 km in length and is NW-trending. This trend and position suggests it may be the complementary anticline to the adjacent Faroe Bank Channel Syncline (Fig. 4.2). On the northern limb of the North Faroe Anticline, in the Faroe Bank Channel, the inferred late Lutetian and C30 (Late Eocene) unconformities are marked by onlap, while the C10 unconformity is marked by erosional truncation (Fig. 4.20). Apparent reverse faults affecting basalt in the northeast of this anticline further suggest a compressional origin (Fig. 4.20).

The South Faroe Bank Anticline (Fig. 4.21) is a 60 km long NE-trending fold with an amplitude of up to 1.78 km (measured from the top-basalt surface). The late Lutetian, C30 (late Eocene), Early Oligocene, and C20 (late Early Miocene) unconformities inferred on the eastern margin of the South Faroe Bank Anticline are marked by onlap. There is also the presence of an inferred Thanetian unconformity marking onlap within basalt lava flows. The Late Miocene-Early Pliocene/C10 composite unconformity is defined by erosional truncation, cutting the Early Oligocene and C20 unconformities.

Faroe Bank Channel Syncline and Vine Anticlines

The Faroe Bank Channel Syncline (Fig. 4.22) is NW-trending and lies between the north-eastern Faroe Bank and the Faroe Shelf. It is characterized by the folding of basalt lava flows and Eocene sediment. The syncline has an amplitude of up to 2.2 km (measured from the top-basalt surface) and is 50 km in length. On the north limb of the syncline, there is an inferred Thanetian unconformity defined by the thinning of basalt lava flows towards the Faroe Shelf. This unconformity may mark the initial growth of the syncline. The inferred late Ypresian, late Lutetian and C30 (Late Eocene) unconformities are marked by onlap. The Faroe Bank Channel Syncline is not present throughout the Faroe Bank Channel as the top-basalt surface becomes flat-lying to the southeast between the Wyville-Thomson Ridge and Faroe Shelf.
Fig. 4.19. SW-NE seismic profile of the northwestern edge of the North Faroe Bank Anticline (for location, see Fig. 4.2). Note the interpreted reverse faults which folds the top-basalt surface on the NE margin of the bank. The inferred late Lutetian and C30 (Late Eocene) unconformities are marked by onlap.
Fig. 4.20. SW-NE seismic profile of the northeast limb of the North Faroe Bank Anticline (for position, see Fig. 4.19). Note the onlap marking the late Lutetian and C30 (Late Eocene) unconformities. Erosional truncation defines the early Late Oligocene/C10 (late Early Pliocene) composite unconformity. Reverse faults affecting the top-basalt surface are inferred on the northwestern limb of the North Faroe Bank Anticline.
Fig. 4.21. E-W seismic profile of the east limb of the South Faroe Bank anticline and the north limb of the Wyville-Thomson Ridge Anticline (for location, see Fig. 4.2). An inferred Thanetian unconformity marks onlap within basalt lava flows. The C10 (late Early Pliocene) unconformity truncates the Early Oligocene and C20 (late Early Miocene) unconformities. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.22. SW-NE seismic profile of the Faroe Bank Channel Syncline (for location, see Fig. 4.2). Note the onlap of sediment marking the late Ypresian, late Lutetian, and C30 (Late Eocene) unconformities and the erosional truncation marking the C10 (late Early Pliocene) unconformity. There is thinning of basalt lava flows towards the Faroe Shelf defining an inferred Thanetian unconformity. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.23. SW-NE seismic line profile of the Vine Anticlines (for location, see Fig 4.2). Note the inferred north-verging reverse faults underlying the folds. The late Lutetian and C30 unconformities are marked by onlap. The early Late Oligocene/C10 (late Early Pliocene) composite unconformity is defined by erosional truncation. Seismic data courtesy of Fugro Multi Client Services.
The Vine Anticlines (Fig. 4.23) are located on the western edge of the Faroe Shelf to the southeast of the Faroe Syncline. They are interpreted as folds which propagate from a series of NW-trending en echelon reverse faults which range in length from 8 to 75 km (Fig. 4.2). The reverse faults dip towards the southwest and run parallel to the Faroe Shelf which defines the northern edge of the Faroe Bank Channel.

**Judd Anticline**

The Judd Anticline (Fig. 4.24) is an east-trending fold located 100 km northeast from the eastern end of the Wyville-Thomson Ridge. The fold has a width of ~ 40 km and an amplitude of up to 800 m (measured from the late Ypresian unconformity). On the southern limb of the Judd Anticline the late Ypresian, late Lutetian, C30 (late Eocene) unconformities are marked by onlap and have been defined by Smallwood (2004). The TPU (Top Palaeogene unconformity) and the C10 (early Late Pliocene) unconformity are characterised by erosional truncation. The thickening of sediment from the Kettla Tuff to Base Balder stratigraphic surfaces towards the crest of the Judd Anticline suggests that the sediment was deposited within a basin.

**Alpin Anticline**

The Alpin Anticline is a slightly arcuate east to west trending symmetrical anticline in the North Rockall Trough between Bill Bailey’s Bank and Rosemary Bank (Fig. 4.2). The fold plunges towards the east resulting in the crest of the Alpin Anticline exposed as sea-bed bathymetry in the west (Fig. 4.25), but covered by sediment towards the east (see Fig. 3.15). The Alpin Anticline has an amplitude of up to 700 m [measured from the C30 (Late Eocene) unconformity] (Fig. 4.25), a length of ~ 150 km and a maximum width of up to 30 km. This anticline represents folding of Paleocene basalts and Eocene-Oligocene sediment. On the western section of the Alpin Anticline (Fig. 4.25) the Eocene sediment has a thickness of up to 1500 m from the top-basalt surface to the crest of the fold. The C30 (Late Eocene) and the Early Oligocene unconformities, characterized by onlap, record the phases of compressional growth of the Alpin Anticline (Fig. 4.26). Poor seismic imaging between the top-basalt surface and the C30 (Late Eocene) unconformity, on the western section of the fold (Fig. 4.25), precludes the mapping of any possible Early
Fig. 4.24. SW-NE seismic profile of the south limb of the Judd Anticline (for location, see Fig. 4.2). The Thanetian, late Ypresian, late Lutetian and C30 (Late Eocene) unconformities are marked by onlap while the early Late Oligocene and the C10 (late Early Pliocene) unconformities are marked by erosional truncation. Note the thickening and onlap of Ketta-Tuff to Base Balder sediments towards the crest of the Judd Anticline. The interpretation of stratigraphic surfaces is based on Smallwood and Gill (2002), Smallwood (2004) and Smallwood et al. (2004). It is inferred that the early Late Oligocene unconformity, in this study, represents the Top Paleogene unconformity (TPU) of Smallwood et al. (2004). Seismic data courtesy of CGG Veritas.
Fig. 4.25. N-S seismic profile of the Alpin Anticline (for location, see Fig. 4.2). Note the thicker pre-C30 (Late Eocene) sediments at the crest of the Alpin Anticline. The north limb of the anticline is flanked by thicker sediment compared to the south limb.
Fig. 4.26. N-S seismic profile across the northern limb of the Alpin Anticline (for position, see Fig. 4.25). Note the onlap of sediments onto the Early Oligocene, early Late Oligocene, C20 (late Early Miocene), and the Late Miocene - Early Pliocene unconformities. The C20 and Late Miocene - Early Pliocene unconformities are also marked by erosional truncation.
Eocene unconformities within the Alpin Anticline. However, the late Ypresian and late Lutetian stratigraphic surfaces have been correlated within the eastern section of the Alpin Anticline (see Fig. 3.15). These surfaces show that there is a lack of thinning of Eocene sediment towards the crest of the Alpin Anticline suggesting that the fold formed subsequent to late Ypresian and late Lutetian times. The distinct onlap of Oligocene sediment onto the C30 (Late Eocene) unconformity along the length of the Alpin Anticline suggests that the fold was formed in the Late Eocene. Post-lava sediment thickness is greater north of the Alpin Anticline than south of the fold. The thickness of onlapping sediment onto the Early Oligocene unconformity is ~ 550 m and ~ 400 m respectively on the southern and the northern limb of the fold, measured from the base of the limb, at the Early Oligocene unconformity, to the sea bed. On the northern limb of the Alpin Anticline, four unconformities overlie the Early Oligocene unconformity – the early Late Oligocene, C20 (late Early Miocene), Late Miocene – Early Pliocene, and C10 (early Late Pliocene) unconformities (Fig. 4.26). Whilst these unconformities are all characterized by onlap, the C20 and the Late Miocene - Early Pliocene unconformities are also marked by erosional truncation. These unconformities are inferred to be the result of bottom-water current activity. The characteristics of these unconformities, related to bottom-water current activity, are explained in section 5.4.3.

Dawn Anticline
The Dawn Anticline (Figs. 4.27 - 4.29) is a NW-trending linear anticline located on the southern end of the Faroe Bank and is parallel to the Wyville-Thomson Ridge Anticline and the North Ymir Ridge Anticline. A bathymetric gorge separates the Dawn Anticline from the North Ymir Ridge Anticline (Fig. 4.2). The fold is 30 km in length and has a width of 18 km. The amplitude of the Dawn Anticline (measured from the top-basalt surface) ranges from 2.1 km in the south to 0.34 km in the north of the fold as it merges with the South Faroe Bank Anticline (Fig. 4.27). The late Ypresian, late Lutetian and C30 (Late Eocene) unconformities are marked by onlap and are interpreted as representing compressional phases of growth of the Dawn Anticline (Figs. 4.28 and 4.29). The Dawn Anticline shows clear evidence of uplift
Fig. 4.27. N-S seismic profile of the Dawn Anticline (for location, see Fig. 4.2). A bathymetric gorge separates the Dawn Anticline from the Sigmundur seamount. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.28. N-S seismic profile across the south limb of the Dawn Anticline (for position, see Fig. 4.27). Note the erosional nature of the inferred C20 (late Early Miocene) unconformity. The inferred composite Late Miocene-Early Pliocene/C10 (late Early Pliocene) unconformity truncates the C30 (Late Eocene) and C20 unconformities. The late Ypresian, late Lutetian and C30 unconformities are marked by onlap. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.29. E-W seismic profile of the Dawn Anticline (for location, see Fig. 4.2). Note the uplift of the basalt escarpment. The late Ypresian, late Lutetian and C30 (Late Eocene) unconformities are marked by onlap. Seismic data courtesy of Fugro Multi Client Services.
of a basalt escarpment, interpreted as representing a paleoshoreline, located on the fold’s northern limb (Fig. 4.29).

**Mordor Anticline**
The Mordor Anticline (Figs. 4.30 – 4.31) is a symmetrical NW-trending fold with an amplitude of 500 m (measured from the late Lutetian unconformity) and is ~ 25 km in length. The fold is surrounded by the Sigmundur seamount and the North and Central Ymir Ridge anticlines, and lies partly within a basin bounded by basalt escarpments (Fig. 4.30). The Mordor Anticline is characterized by both the folding of basalt lava flows and overlying Cenozoic sediments. Growth of the Mordor Anticline is defined by the late Ypresian, late Lutetian, C30 (Late Eocene) and Early Oligocene unconformities as represented by onlap on these stratigraphic surfaces (Fig. 4.31). Sediments onlapping the Early Oligocene unconformity on the eastern and western limbs, flanking the fold, are up to 500 m and 350 m thick respectively. The Late Miocene – Early Pliocene unconformity has truncated the Early Oligocene and early Late Oligocene unconformities at the crest of the fold and the C20 (late Early Miocene) unconformity located more distally above the South Ymir Ridge Anticline (Fig. 4.30). The crest of the Mordor Anticline is covered in a veneer of Late Pliocene – Recent sediment, 100 m thick, resulting in the fold showing no expression on the sea bed.

**Ymir Ridge Anticlines**
The Ymir Ridge is a bathymetric high comprising a series of three anticlines – the North, Central and South Ymir Ridge anticlines. The anticlines are all NW-trending but differ in size and shape.

**North Ymir Ridge Anticline**
The North Ymir Ridge Anticline is NW-trending, has a width of ~ 20 km in the west (Fig. 4.32) broadening to ~ 28 km towards the east (Fig. 4.33 – 4.35) and is ~ 25 km in length. In the west the amplitude of the fold (measured from the top-basalt surface) is 0.75 km and 1.8 km measured from the adjacent northern and southern troughs respectively (Fig. 4.32). The fold has an amplitude of ~ 1.4 km (measured
Fig. 4.30. W-E seismic profile of the Mordor Anticline (for location, see Fig. 4.2). Note that the anticline lies within a basin bounded by basalt escarpments. The late Ypresian, late Lutetian, C30 (Late Eocene), Early Oligocene, and early Late Oligocene unconformities are marked by onlap. The Early Oligocene, early Late Oligocene, and the C20 (late Early Miocene) unconformities are truncated by the Late Miocene - Early Pliocene unconformity. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.31. E-W seismic profile of the Mordor Anticline (for position, see Fig. 4.30). The Mordor Anticline lies within a basin bounded by basalt escarpments. The late Ypresian, late Lutetian, C30 (Late Eocene), Early Oligocene, and early Late Oligocene unconformities are marked by onlap. The Early Oligocene and the early Late Oligocene unconformities are truncated by the Late Miocene - Early Pliocene unconformity. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.32. N-S seismic profile across the North Ymir Ridge Anticline (for location, see Fig. 4.2). Note the onlap defining the Thanetian, late Ypresian, late Lutetian and the C30 (Late Eocene) unconformities. The Thanetian unconformity marks onlap within basalt lava flows. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.33. Uninterpreted N-S seismic profile of the Mordor, North Ymir Ridge and Wyville-Thomson Ridge anticlines (for location, see Fig. 4.2). Note the Auðhumla Basin Syncline between the Ymir and Wyville Thomson Ridge anticlines. A basalt escarpment is located in the syncline between the Mordor and Ymir Ridge anticlines. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.34. N-S seismic profile of the Mordor, North Ymir Ridge and Wyville-Thomson Ridge anticlines (for location, see Fig 4.2). Note the Auðhumla Basin Syncline between the Ymir and Wyville Thomson Ridge anticlines. A basalt escarpment is located in the syncline between the Mordor and Ymir Ridge anticlines enclosing Ypresian sediments of the Mordor Anticline. Seismic data courtesy Fugro Multi Client Services.
Fig. 4.35. N-S seismic profile across the North Ymir Ridge Anticline (for position, see Fig. 4.34). Note the onlap defining the Thanetian, late Ypresian, late Lutetian, C30 (Late Eocene), and Early Oligocene unconformities. The Thanetian unconformity marks onlap within basalt lava flows. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.36. N-S seismic profile across the north limb of the Wyville-Thomson Ridge Anticline (for position, see Fig. 4.34). Note the onlap defining the Thanetian, late Ypresian, late Lutetian, C30 (Late Eocene), Early Oligocene and the C20 (late Early Miocene) unconformities. On the north limb of the Wyville Thomson Ridge the C10 (late Early Pliocene) unconformity truncates the late Ypresian, late Lutetian, C30, and the Early Oligocene/early Late Oligocene unconformities. Seismic data courtesy of Fugro Multi Client Services.
from the top-basalt surface) in the east (Fig. 4.35). A Thanetian unconformity, defined by onlap, is inferred within the basalt lava flows of the North Ymir Ridge Anticline suggesting folding at this time. The late Ypresian, late Lutetian and C30 (Late Eocene) unconformities on the south limb of the anticline are marked by onlap (Fig. 4.32).

Central Ymir Ridge Anticline
The Central Ymir Ridge Anticline is NW-trending, has an average width of ~ 30 km and is ~ 23 km long. The fold is asymmetric with the southern limb having a steeper gradient than the northern limb. The amplitude of the fold decreases from 1.24 km in the west (Fig. 4.39) to 1.15 km in the east (Fig. 4.41). This difference in amplitude is attributed to post-late Ypresian reverse faulting which affected the crest at the northwest end of the fold to increase the fold’s amplitude (Fig. 4.39). The limbs of the northwest section of the fold are also cut by north dipping reverse faults. Towards the southeast it is evident that the late Ypresian unconformity is folded over the crest of the fold (Fig. 4.41). The late Lutetian, C30 (Late Eocene) and the Early Oligocene unconformities are marked by onlap on both limbs of the fold and represent the ages of compression for the Central Ymir Ridge Anticline.

South Ymir Ridge Anticline
A plateau (Fig. 4.43) separates the Central Ymir Ridge Anticline from the South Ymir Ridge Anticline. The South Ymir Ridge Anticline is NW-trending, and is ~ 7 km long. There is no seismic data showing the full width of the fold, but this is estimated to be approximately 26 km based on the half-width of the fold (Fig. 4.44). The late Lutetian, C30 (Late Eocene) and the Early Oligocene unconformities area marked by onlap on the south limb of the fold and are inferred to represent the ages of compression of the South Ymir Ridge Anticline.

Wyville-Thomson Ridge Anticline
The Wyville-Thomson Ridge Anticline (Figs. 4.36 and 4.40) is a NW-trending anticline, 200 km in length and represents ~ 1% shortening. The ridge links the Faroe Bank and the Hebridean/West Shetland Shelf, separating the Rockall Trough
Fig. 4.37. Uninterpreted N-S seismic profile of the Central Ymir Ridge and Wyville Thomson Ridge anticlines (for location, see Fig. 4.2). Note the north-dipping reverse faults which cut the top basalt surface within the Auðhumla Basin and on the south limb of the Central Ymir Ridge Anticline. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.38. N-S seismic profile of the Central Ymir Ridge and Wyville-Thomson Ridge anticlines (for location, see Fig. 4.2). Note the Auðhumla Basin syncline between the two anticlines and the lack of Ypresian sediments within the basin. There are north-dipping reverse faults which cut the top basalt surface within the basin and on the south limb of the Ymir Ridge Anticline. An inferred Thanetian unconformity marked by onlap within the basalt lava flow is located within the Wyville Thomson Ridge Anticline. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.39. N-S seismic profile across the Central Ymir Ridge Anticline (for position, see Fig. 4.38). Note the onlap defining the late Ypresian, late Lutetian, C30 (Late Eocene), Early Oligocene and the early late Oligocene unconformities. The Central Ymir Ridge Anticline is cut by a south dipping reverse faults, while north dipping reverse faults are present on the north and south limbs of the fold. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.40. N-S seismic profile across the north limb of the Wyville-Thomson Ridge Anticline (for position, see Fig. 4.38). Note the onlap defining the Thanetian, late Ypresian, late Lutetian, C30 (Late Eocene), Early Oligocene/early Late Oligocene and the C20 (late Early Miocene) unconformities. The composite Early Oligocene/early Late Oligocene unconformity and the C10 (late Early Pliocene) unconformity are marked by erosional truncation.
Fig. 4.41. N-S seismic profile across the Central Ymir Ridge Anticline (for location, see Fig. 4.2). Note the onlap defining the C30 (Late Eocene), Early Oligocene and the early late Oligocene unconformities. Ypresian sediment onlap over the crest of the anticline. The Thanetian unconformity is absent within the Central Ymir Ridge Anticline. Seismic data courtesy Fugro Multi Client Services.
Fig. 4.42. E-W seismic profile across the Central Ymir Ridge Anticline (for location, see Fig. 4.2). Note the folding of the Late Ypresian unconformity over the anticline and the presence of a thick Lutetian wedge of sediment on the west limb of the anticline. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.43. SW-NE seismic profile across the Ymir Ridge Plateau (for location, see Fig. 4.2). Note the onlap defining the Early Oligocene unconformity. Seismic data courtesy of CGG Veritas.
Fig. 4.44. E-W seismic profile across the South Ymir Ridge (for location, see Fig. 4.2). Note the onlap defining the late Lutetian, the C30 (Late Eocene) and the Early Oligocene unconformity. Seismic data courtesy of Fugro Multi Client Services.
from the Faroe Shetland Channel. The Wyville-Thomson Ridge Anticline has a width of ~ 28 km and an amplitude of up to 1.7 km (measured from the top-basalt surface). The anticline is characterized by extensive folding of basalt lava flows and the tilting of post-Paleocene sediments in the limbs of the fold (Figs. 4.36 and 4.40). On the north limb of the Wyville-Thomson Ridge Anticline the unconformities inferred as being the result of folding – Thanetian, late Ypresian, late Lutetian, C30 (Late Eocene) and the Early Oligocene – are marked by onlap. The Thanetian unconformity defines onlap within basalt lava flows. The Early Oligocene stratigraphic surface is inferred as forming a composite unconformity with the early Late Oligocene unconformity on the north limb of the Wyville-Thomson Ridge. The early Late Oligocene unconformity was previously interpreted as the Top Palaeogene unconformity (TPU) (Smallwood, 2004). Onlap also defines the C20 (late Early Miocene) unconformity. The C10 (early Late Pliocene) unconformity truncates the late Ypresian, late Lutetian, C30, Early Oligocene/early Late Oligocene and C20 unconformities on the north limb of the Wyville-Thomson Ridge Anticline (Figs. 4.36 and 4.40).

**Auðhumla Basin Syncline**

A syncline, the Auðhumla Basin, lies between the Ymir Ridge and Wyville-Thomson Ridge Anticlines and is filled with Cenozoic sediments (Figs. 4.34 and 4.38). Within the Auðhumla Basin Syncline there are 6-12 km long fault propagation folds which run in a northwest direction and their associated reverse faults dip to the northeast (Fig. 4.2). The development of the Auðhumla Basin Syncline is governed by the growth of both the Ymir Ridge and Wyville-Thomson Ridge anticlines. The postulated Thanetian unconformity on the northern limb of the North Ymir Ridge Anticline is considered to mark the initial growth of the Auðhumla Basin Syncline adjacent to this anticline (Fig. 4.32). The Auðhumla Basin Syncline adjacent to the Central Ymir Ridge Anticline, however, formed later with the growth of this anticline. Here, the initial growth of the syncline is defined by the late Lutetian unconformity marked by the onlap of Late Eocene sediment (Fig. 4.41).
Bridge Anticline

The Bridge Anticline (Fig. 4.45) is a NE-trending symmetrical fold between the Wyville-Thomson Ridge Anticline and the southern end of the Central Ymir Ridge Anticline. The fold separates the Auðhumla Basin from the NE Rockall Trough (Fig. 4.2). The Bridge Anticline has an amplitude of up to 1.2 km (measured from the top-basalt surface) and a width of 40 km. The fold is characterized on seismic profile by the folding of basalt lava flows and the overlying Cenozoic sediments. Middle Eocene sediment (up to 800 m thick) onlap the top-basalt surface at the crest of the fold and downlap the late Ypresian unconformity on the eastern limb of the fold (Fig. 4.46). These sediments are themselves onlapped by a wedge of Late Eocene sediment (~ 500 m thick). The late Lutetian unconformity, separating the Middle and Late Eocene sediment packages, and the C30 (Late Eocene) unconformities on the eastern limb of the fold are marked by strong onlap. Ypresian sediment onlap onto the top-basalt surface on the eastern limb of the Bridge Anticline suggesting a structural high existed at this time. The compressional growth phases of the Bridge Anticline are discussed in Section 5.3.

Onika Anticline

The Onika Anticline (Fig. 4.47 and Figs. 4.49 - 4.50) is a NNE-trending fold located to the southeast of the South Ymir Ridge Anticline. The fold has an amplitude of up to 800 m [measured from the C30 (Late Eocene) unconformity (Fig. 4.49)] and represents folding of mainly Eocene sediment. The Onika Anticline lies within a basin bounded by basalt escarpments and its development from basin sediments is supported by the thickness of Early – Middle Eocene sediment increasing towards the crest of the fold (Fig. 4.47). The late Lutetian and the C30 (Late Eocene) unconformities are marked by onlap (Figs. 4.49 - 4.50). This fold is not expressed at the sea bed as its crest is covered by approximately 500 m of sediment. The western limb of the Onika Anticline is flanked by sediments up to 1100 m thick. On the eastern margin of the Onika Anticline an additional sediment contribution from the Sula Sgeir fan results in thicker sediments (up to 1400 m thick). Differential sediment loading, as a possible mechanism of formation for the Onika Anticline, is discussed in section 4.7.1.
Fig. 4.45. E-W seismic profile across the Bridge Anticline (for location, see Fig. 4.2). Note the downlap of Middle Eocene sediments onto the late Ypresian unconformity on the eastern limb of the Bridge Anticline. All other unconformities are marked by onlap. Seismic data courtesy of Fugro Multi Client Services.
Fig. 4.46. E-W seismic profile across the Bridge Anticline (for position, see Fig. 4.45). Note the onlap and downlap of Lutetian sediment, respectively, on the crest and the east limb of the Bridge Anticline. There is strong onlap onto the top basalt surface and the late Lutetian, C30 (Late Eocene), Early Oligocene and C10 (early Late Pliocene) unconformities on the east limb.
Fig. 4.47. NW-SE seismic profile across the Alpin Anticline, Onika Anticline and the inverted West Lewis Basin (for location, see Fig. 4.2). Eocene sediment of the Onika Anticline is bounded by basalt escarpments. Note the thicker sediment from the top-basalt surface to the late Lutetian unconformity, towards the crest of the Onika Anticline. The late Ypresian and late Lutetian unconformities are cut by a reverse fault to the west of the West Lewis Basin. Seismic data courtesy of CGG Veritas.
Fig. 4.48. NW-SE seismic profile of the inverted West Lewis Basin (for position, see Fig. 4.47). Note the reverse fault which cuts the late Ypresian and the late Lutetian unconformities and stops at the C30 (Late Eocene) unconformity. The fault does not affect the overlying Early Oligocene unconformity. Seismic data courtesy of CGG Veritas.
Fig. 4.49. NW-SE seismic profile across the Onika Anticline (for position, see Fig. 4.47). Note the distinctive onlap of Oligocene sediment onto the C30 (Late Eocene) unconformity. The C20 (late Early Miocene) and younger unconformities appear unperturbed by the underlying Onika Anticline. Within the Onika Anticline sediment thickness, from the top-basalt surface to the late Lutetian surface, increase towards the crest of the fold. The Onika Anticline lies within a basin bounded by basalt escarpments. Seismic data courtesy of CGG Veritas.
Fig. 4.50. SW-NE seismic profile of the Onika and Ness anticlines (for location, see Fig. 4.2). Note the onlap onto the late Lutetian and C30 unconformities. Seismic data courtesy of CGG Veritas.
West Lewis reverse faults

The West Lewis reverse faults (Fig. 4.48 and Fig. 4.51) partially bound the margins of the NE-trending West Lewis Ridge located on the east margin of the NE Rockall Trough (Fig. 4.2). These reverse faults are interpreted as normal faults which have undergone inversion. Along the eastern side of the ridge the reverse fault is 40 km in length, with the West Lewis Basin as the hanging wall of the fault. This reverse fault cuts the late Ypresian and late Lutetian unconformities in the north (Fig. 4.48), but towards the south, it forms a fault-propagation fold (Fig. 4.51). A Thanetian unconformity, defined by onlap within basalt lava flows, is located in the south West Lewis Basin (Fig. 4.51) and may define the initiation of reverse faulting in the south West Lewis Basin. A second 10 km long reverse fault bounds the West Lewis Ridge, to the southwest, where it also forms a fault-propagation fold (Fig. 4.51). These fault propagation folds are not expressed at the sea bed as they are covered by Sula Sgeir fan sediments ~ 500 m and ~ 700 m thick on the western and eastern sides of the West Lewis Ridge respectively. Fold development associated with the West Lewis reverse faults have not perturbed the C10 (early Late Pliocene) unconformity. Towards the northeast there is no inversion of the normal fault bounding the western side of the West Lewis Ridge (Fig. 4.48).

Viera Anticline

The Viera Anticline (Fig. 4.52 - Fig. 4.54) is an approximately 30 km long NW-trending fold that lies to the southwest of the West Lewis Ridge in the NE Rockall Trough (Fig. 4.2). The amplitude of the anticline (measured from the late Ypresian unconformity) is 360 m in the west (Fig. 4.53) and up to ~ 700 m in the east (not illustrated). The C30 (Late Eocene) unconformity, which is interpreted as marking the initial age of folding, is defined by onlap of Early Oligocene sediment (Figs. 4.52 - 4.54). Sediment was deposited preferentially on the north margin of the fold, up to 800 m, compared to the southern margin (~ 600 m). This could have resulted in differential sediment loading. Differential sediment loading as a mechanism for the formation of the Viera Anticline is discussed in section 4.5.1.
Fig. 4.51. SW-NE seismic profile of inverted faults that bound the West Lewis Ridge (for location, see Fig. 4.2). Note the fault propagation folds above the inverted faults on either side of the West Lewis Ridge. A Thanetian unconformity was dated within the West Lewis Basin. Seismic data courtesy of CGG Veritas.
Fig. 4.52. SW-NE seismic profile across the north limb of the Viera Anticline and the south limb of the Wyville-Thomson Ridge Anticline (for location, see Fig. 4.2). Onlap is evident on the late Ypresian, Late Lutetian, C30 (Late Eocene), and Early Oligocene unconformities. Seismic data courtesy of CGG Veritas.
Fig. 4.53. SW-NE seismic profile across the west end of the Viera Anticline (for location, see Fig. 4.2). Note the distinct onlap of Early Oligocene sediments on the C30 (late Eocene) unconformity. Seismic data courtesy of CGG Veritas.
Fig. 4.54. NW-SE seismic profile across the east end of the Viera Anticline (for location, see Fig. 4.2). Note the distinct onlap of Oligocene sediment onto the C30 (Late Eocene) unconformity. Seismic data courtesy of CGG Veritas.
4.5 Isochron Maps

The isochron maps, in this study, are two-way time surface maps of stratigraphic horizons interpolated between 2-D seismic lines. Some of these maps delineate the approximate extent and orientation of folds in the Rockall-Faroe area.

Top Basalt

The top-basalt surface has been calibrated as Late Paleocene – Early Eocene by wells 163/06-1, 164/07-1, 164/25-1 and 164/25-2 in the NE Rockall Trough. This surface forms an easily identifiable horizon that can be traced throughout the Rockall-Faroe area. Many anticlines are defined by the top-basalt lava surface (Figs. 4.55 – 4.56) indicating compression of basalt lava flows. Basalt-covered anticlines which reflect the effects of compression are Hatton Bank, Lousy Bank, Bill Bailey’s Bank, Faroe Bank, Dawn, Wyville-Thomson Ridge, North Ymir Ridge, Central Ymir Ridge, South Ymir Ridge and Bridge anticlines. The North Ymir Ridge, Central Ymir Ridge and South Ymir Ridge anticlines appear as distinctly separate folds (Fig. 4.56). Structures such as the Alpin and Mordor anticlines are less defined by the top-basalt surface.

In addition to compressional structures, igneous centres, such as the Faroe Channel Knoll, Sigmundur, Darwin, Rosemary Bank, and Mammal, are well delineated by the top-basalt surface. Bill Bailey’s Bank and Lousy Bank, which are also well defined by the top-basalt surface, were originally thought by Ritchie et al. (1999) to also be igneous centres. However, results of this study reveal that these structures are unlikely to be igneous centres.

Seismic data have revealed a number of basalt escarpments. These escarpments are interpreted as representing paleo-shorelines formed when lava flows encountered a body of water (Naylor et al., 1999). The location of basalt escarpments, mapped in this study, is shown in Fig. 4.56. Some compressional structures, such as the Mordor Anticline (Fig. 4.30), and the Onika Anticline (Fig. 4.47) are enclosed by these basalt escarpments. In the case of the Onika Anticline (Fig. 4.49), the thickening of Early – Middle Eocene sediment towards the crest of the fold suggests the presence of a former basin bounded by basalt escarpments.
Fig 4.55. Top-basalt horizon isochron of the Rockall-Faroe area. Note the compressional structures (italics) delineated by the top basalt surface. Igneous centres are also well expressed by the top basalt surface (refer to text). The top-basalt surface was calibrated by the wells shown on the map. The top-basalt surface was also calibrated within the Hatton-Rockall Basin using borehole 117 (for location, see Fig. 3.4). The lack of the top-basalt isochron in some areas, such as in the North Rockall Trough, is due to poor seismics. Basalt escarpments are shown as white dotted lines. Abbreviation: FCK=Faroe Channel Knoll.
Fig. 4.56. Top-basalt horizon isochron from the Rosemary Bank to the Faroe Bank Channel. The position of basalt escarpments (inferred as paleoshorelines) are indicated by white dotted lines. Note that the Mordor and the Onika anticlines are enclosed by these basalt escarpments.
The lack of the mapped top-basalt surface in the North Rockall Trough is due to poor seismic resolution of the horizon in this area. The absence of the top-basalt isochron surface in the north-eastern Lousy Bank and the west Bill Bailey's Bank area is due to the lack of seismic data. The top-basalt surface ranges from 200 ms two-way time depth at the crest of structures, such as Faroe Bank, to > 4000 ms in the North Rockall Trough.

Late Ypresian
The late Ypresian surface (Figs. 4.57 – 4.58) is limited to the NE Rockall Trough, the Ymir and the Wyville-Thomson Ridges, and the Faroe Bank Channel areas. In these areas the late Ypresian unconformity can be dated and mapped with confidence. Wells used to calibrate the late Ypresian stratigraphic surface, in this study, are 154/01-1, 163/06-1, 164/07-1, 164/25-1 and 164/25-2 in the NE Rockall Trough. The late Ypresian surface (Fig. 4.58) has defined the Bridge, Onika, Mordor and the Viera anticlines. The northern and southern limbs of the Wyville-Thomson Ridge Anticline, the northern limb of the North Ymir Ridge Anticline, the Central Ymir Ridge and the South Ymir Ridge anticlines are also expressed by the Late Ypresian surface. The late Ypresian surface ranges from 800 ms two-way time depth on the Hebrides Shelf to 3000 ms in the NE Rockall Trough.

Late Lutetian
The late Lutetian surface (Figs. 4.59 – 4.60), like the late Ypresian surface, is limited in extent to the NE Rockall Trough, the Ymir and Wyville-Thomson Ridges and the Faroe Bank Channel areas. The late Lutetian unconformity was mapped with confidence in these areas due to good well control and seismic coverage. Wells used to calibrate the late Lutetian stratigraphic surface, in this study, are 154/01-1, 163/06-1, 164/07-1, 164/25-1 and 164/25-2 in the NE Rockall Trough. Structures expressed by the late Ypresian surface are also delineated by the late Lutetian surface such as the Bridge, Onika, Mordor and the Viera anticlines. The late Lutetian surface ranges in two-way time depth from 1000 ms on the Faroe Shelf to 2600 ms in the NE Rockall Trough.
Fig. 4.57. Late Ypresian horizon isochron of the Rockall-Faroe area. Note the compressional structures (italics) delineated by the late Ypresian stratigraphic surface. The late Ypresian unconformity was calibrated using the wells shown on the map. The interpretation of the late Ypresian unconformity in the Judd Anticline area is based on Smallwood et al. (2004). The lack of the late Ypresian isochron, in most areas, is due to poor seismic data and a lack of well control.
Fig. 4.58. Late Ypresian isochron from the Rosemary Bank to the Faroe Bank Channel. The Mordor, Bridge and Viera anticlines are clearly defined by this isochron.
Fig 4.59. Late Lutetian horizon isochron of the Rockall-Faroe area. Note the compressional structures (italics) delineated by the late Lutetian stratigraphic surface. Wells used to calibrate the late Lutetian unconformity are shown on the map. The lack of the late Lutetian isochron, in most areas, is due to poor seismics and a lack of well control.
Fig. 4.60. Late Lutetian horizon isochron from the Rosemary Bank to the Faroe Bank Channel. The Mordor, Bridge, Onika and the Viera anticlines are clearly defined by this isochron.
**C30 (Late Eocene)**

The C30 surface (Figs. 4.61 – 4.62) extends over the entire study area ranging in from two-way time depth 300 ms on the Hebrides Shelf to > 3000 ms in the North Rockall Trough. The surface has been mapped confidently in the NE Rockall Trough and the Central Rockall Trough due to the availability of well data and good quality seismic data. The C30 unconformity has been calibrated using wells 154/01-1, 163/06-1, 164/07-1, 164/25-1 and 164/25-2 in the NE Rockall Trough, and borehole 94/1 in the Central Rockall Trough. The unconformity has also been constrained in the Hatton-Rockall Basin using borehole 117. Anticlines well defined by the C30 surface are the Alpin, Mordor, Onika, Viera and Bridge anticlines (Fig. 4.62). The limbs of anticlines such as the Hatton Bank, Lousy Bank, Bill Bailey’s Bank, North Ymir Ridge, Central Ymir Ridge and the Wyville-Thomson Ridge anticlines are also defined by this surface.

The C30 isochron reveals that the Alpin and Mordor Anticlines are separate folds (Fig. 4.62). The Onika and Viera anticlines are also distinctly separated. The C30 surface is not mapped on the Mammal, Rosemary Bank, Darwin, Sigumundur and Faroe Channel Knoll igneous centres. The crests of the Wyville-Thomson Ridge, Ymir Ridge, Dawn, Faroe Bank, Bill Bailey’s Bank and Lousy Bank Anticlines lack the C30 surface due to erosion. The absence of the C30 surface in other areas, such as the western margin of the Lousy Bank, is due to a lack of well-calibrated seismic data.

**Early Oligocene**

The Early Oligocene horizon has been calibrated in this study using wells 163/06-1, 164/07-1, 164/25-2 and 154/01-1 in the NE Rockall Trough and borehole 116 in the Hatton-Rockall Basin. At borehole 116 the Early Oligocene surface forms a composite unconformity with the early Late Oligocene horizon (see Fig. 3.21). The Early Oligocene surface (Figs. 4.63 – 4.64) ranges in two-way time depth from 300 ms, on the Faroe Shelf, to > 3000 ms in the North Rockall Trough. The Alpin and Bridge anticlines are well defined by the Early Oligocene surface. The Early Oligocene surface also defines the northern and southern limbs of the Wyville-Thomson Ridge and the northern limbs of North and Central Ymir Ridge anticlines.
Fig. 4.61. C30 (late Eocene) horizon isochron of the Rockall-Faroe area. Note the compressional structures (italics) delineated by the C30 (Late Eocene) stratigraphic surface. The C30 stratigraphic surface has been calibrated using the wells shown on the map and by boreholes 94/1 and 117 (for the location of boreholes, see Fig. 3.4). Seamounts, shown on the top-basalt isochron (see text), lack the C30 isochron as the C30 unconformity onlaps onto the margins of these structures.
Fig. 4.62. C30 (Late Eocene) isochron from the Rosemary Bank to the Faroe Bank Channel. The Alpin, Mordor, Bridge, Onika and Viera anticlines are clearly defined by this isochron.
Fig. 4.63. Early Oligocene horizon isochron of the Rockall-Faroe area. Note the compressional structures (italics) delineated by the Early Oligocene isochron. Early Oligocene stratigraphic surface was calibrated by the wells shown and by borehole 116 in the Hatton-Rockall Basin outwith this area (for location, see Fig. 3.4). Structural highs lacking the Early Oligocene isochron due to the onlap of the Early Oligocene unconformity onto the margins of these structures.
Fig. 4.64. Early Oligocene horizon isochron from the Rosemary Bank to the Faroe Bank Channel. The Alpin, Mordor and Bridge anticlines are clearly defined by this isochron.
The Mordor, Onika and Viera anticlines are less defined by the Early Oligocene surface.
The Early Oligocene unconformity is not mapped on structural highs, such as Sigmundur and Rosemary Bank seamounts. The absence of the Early Oligocene surface on the crests of compressional structures such as, the Wyville-Thomson Ridge, Ymir Ridge and Faroe Bank anticlines is due to the erosion.

*Early Late Oligocene*

The early Late Oligocene surface (Figs. 4.65 – 4.66) has been calibrated in this study by well 164/25-2 in the NE Rockall Trough and borehole 116 in the Hatton-Rockall Basin. At borehole 116 the early Late Oligocene surface forms a composite unconformity with the Early Oligocene horizon. The surface defines the Bridge Anticline and the northern limb of the Wyville-Thomson Ridge Anticline (Fig. 4.66). The early Late Oligocene stratigraphic surface is relatively shallow in the Faroe Bank Channel, at a two-way time of 1900 ms, shallowing to 1000 ms as the unconformity onlaps the Faroe Shelf and the Wyville-Thomson Ridge. The early Late Oligocene surface descends markedly in the North Rockall Trough to two-way time depths > 2100 ms. The early Oligocene surface can be broadly divided into two zones:

- A NE-trending high that extends from Hatton-Rockall Basin (south of Hatton Bank Anticline) to the Faroe Bank Channel.
- A low, south of the aforementioned high, within the Rockall Trough.

The overlying sea-bed bathymetry mimics this broad demarcation of the early Late Oligocene stratigraphic surface.

Across the Alpin and Mordor anticlines the early Late Oligocene surface is absent as it is truncated by the C20 (late Early Miocene) and Late Miocene-Early Pliocene unconformities respectively (Figs. 4.26 and 4.31).
Fig. 4.65. Early Late Oligocene horizon isochron of the Rockall-Faroe area. The early Late Oligocene stratigraphic surface was calibrated using well 164/25-2 shown on the map and borehole 116 in the Hatton Rockall Basin outwith this area (for location, see Fig. 3.4). The lack of the Early Oligocene isochron in some areas is due to onlap of the early Late Oligocene horizon onto the margins of structural highs, erosional truncation of the horizon by younger unconformities (see text), and a lack of seismic data.
Fig. 4.66. Early Late Oligocene horizon isochron from the Rosemary Bank to the Faroe Bank Channel. The Bridge Anticline is defined by this isochron.
C20 (late Early Miocene)
The C20 (late Early Miocene) surface (Figs. 4.67 – 4.68) was constrained using well 164/25-2 in the NE Rockall Trough. The surface does not delineate any compressional features with the exception of the Bridge Anticline. The C20 surface, like the early Late Oligocene surface, can be broadly divided into two zones:

- NE-trending high, north of the Rosemary Bank, of almost constant two-way time depth of ~ 1900 ms from the West Lewis Ridge to the Hatton-Rockall Basin, with the exception of the Wyville-Thomson Ridge, Ymir Ridge and Bridge anticlines. These features exhibit a shallower C20 surface of ~ 1600 ms two-way time depth.
- A low of two-way time depth > 2400 ms in the North Rockall Trough

These zones of the C20 surface correlate with the overlying bathymetry.

On the Mordor and the Alpin anticlines the C20 surface is absent due to erosional truncation of the C20 unconformity by the Late Miocene-Early Pliocene unconformity (Figs. 4.26, 4.30 and 4.47).

Late Miocene – Early Pliocene
The Late Miocene – Early Pliocene surface (Figs. 4.69 – 4.70) was constrained using well 164/25-2 in the NE Rockall Trough. The surface does not delineate any compressional features within the Rockall-Faroe area maintaining a two-way time depth of ~ 1800 ms from the West Lewis Ridge to the east margin of the Lousy Bank Anticline and within the Hatton-Rockall Basin. Two-way time depths > 2200 ms are reached in the North Rockall Trough.

C10 (late Early Pliocene)
The C10 (late Early Pliocene) surface (Figs. 4.71 – 4.72) was calibrated by wells 154/01-1, 162/25-1 and 164/25-2 in the NE Rockall Trough and borehole 116 in the Hatton-Rockall Basin. The Bridge Anticline is defined by this surface. Apart from the Bridge Anticline, the C10 surface is generally expressed as a NE-trending high of a constant depth of ~ 1700 ms from the Faroe Bank Channel to the Hatton-Rockall Basin. Within the North Rockall Trough the C10 horizon descends to two-way time depths > 2100 ms.
Fig. 4.67. C20 (late Early Miocene) horizon isochron of the Rockall-Faroe area. The C20 stratigraphic surface has been constrained using well 164/25-2 shown on the map and well 610 outwith the study area (for location, see Fig. 3.4). The lack of the C20 isochron in areas is due to onlap of the C20 horizon onto the margins of structural highs, erosional truncation of the horizon by younger unconformities (see text), and the lack of seismic data.
Fig. 4.68. C20 (late Early Miocene) isochron from the Rosemary Bank to the Faroe Bank Channel. The Bridge Anticline is defined by this isochron.
Fig. 4.69. Late Miocene-Early Pliocene horizon isochron of the Rockall-Faroe area. The Late Miocene-Early Pliocene stratigraphic surface was constrained using well 164/25-2. The lack of this isochron, in some areas, is due to onlap of the Late Miocene-Early Pliocene horizon onto the margins of structural highs and the lack of seismic data.
Fig. 4.70. Late Miocene - Early Pliocene isochron from the Rosemary Bank to the Faroe Bank Channel. There is no distinct definition of any anticline by this isochron.
Fig. 4.71. C10 (late Early Pliocene) horizon isochron of the Rockall-Faroe area. Wells used to calibrate the C10 stratigraphic surface are shown on the map. In addition, borehole 116 in the Hatton-Rockall Basin, outwith this area (for location, see Fig. 3.4), has also dated and constrained the C10 stratigraphic surface. The lack of the C10 isochron in some areas is due to onlap of the C10 unconformity onto the margins of structural highs.
Fig. 4.72. C10 (late Early Pliocene) horizon isochron from the Rosemary Bank to the Faroe Bank Channel.
Fig. 4.73. Bathymetric map of the Rockall-Faroe area. Note that structures such as Hatton Bank, Lousy Bank, Bill Bailey’s Bank, Ymir Ridge and Wyville-Thomson Ridge anticlines are well imaged on the sea bed.
Many structural anticlines affect the present-day sea bed (Fig. 4.73). These include the Hatton Bank, Lousy Bank, Bill Bailey's Bank, Faroe Bank, Wyville-Thomson Ridge and Ymir Ridge anticlines. This is evidence for compressional structures shaping the bathymetry of the sea bed. In section 5.4.5 the study explores the influence of this change in bathymetry, as a result of compression, on bottom-water current activity.

4.6 Gravity Modelling

4.6.1 Introduction

The NE Atlantic Margin study area is covered in a veneer of Paleocene – Early Eocene basalt lava flows. The flows mainly comprise basalt separated by either tuff or breccia layers. This lithological variation results in differing velocities and densities in the succession creating a high impedance contrast and resulting in high reflection of seismic energy (Maresh and White, 2005). Consequently, structures below the basalt lava flows remain very poorly imaged on the seismic data.

Gravity data (Bouguer gravity anomalies) acquired by BGS during seismic acquisition across the Rockall-Faroe area, have been used to model the underlying structures in the Rockall-Faroe study area that are not imaged on seismic data. The location of gravity profiles used in this study is shown in Fig. 4.74. GravMag software (Pedley et al., 1993) was used to build gravity models in an attempt to reproduce gravity anomalies (based on the density of rocks) which best match the observed gravity anomalies. For details on the construction of gravity models using GravMag refer to Section 1.5.3.
Fig. 4.74. Map of compressional features in the Rockall-Faroe Area. The locations of illustrated gravity profiles are shown by dotted black lines.
<table>
<thead>
<tr>
<th>Rock</th>
<th>Density Mg/m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reduction density used for sea water</td>
<td>2.2</td>
</tr>
<tr>
<td>Cenozoic sediments</td>
<td>2.2</td>
</tr>
<tr>
<td>Basalt lava flows</td>
<td>2.5</td>
</tr>
<tr>
<td>Pre-Cenozoic sediments</td>
<td>2.25 and 2.4</td>
</tr>
<tr>
<td>Igneous intrusions</td>
<td>2.7-3.0</td>
</tr>
<tr>
<td>Basement (Upper Crust)</td>
<td>2.7</td>
</tr>
<tr>
<td>Lower Crust</td>
<td>2.9</td>
</tr>
<tr>
<td>Mantle</td>
<td>3.32</td>
</tr>
</tbody>
</table>

Table 4.2 Densities of units used in Gravity Models.

The rock densities used in the gravity models (Table 4.2) are based on approximate densities of common rock types (Keary and Brooks, 1991) and previous work on gravity modelling in the Rockall Trough (Archer et al., 2005; Klingelhofer et al., 2005). The densities of pre-Cenozoic sediment was taken as 2.25 Mg/m³ and 2.4 Mg/m³ in order to respectively represent the lower and upper end of the range of possible densities for these rocks (Klingelhofer et al., 2005). The age of this sediment, which underlies the basalt lava flow, is based on the late Paleocene – Early Eocene age of the top-basalt surface (see section 3.3).

In the models illustrated here, the thickness of basalt and crust are based on previous work by White et al. (1987), Klingelhofer et al. (2005) and Funck et al. (2008). The thickness of Cenozoic sediment was constrained using seismic data, while the extent and depth of igneous intrusions were guided by the observed gravity anomalies. The proportion of lower crust, upper crust (basement) and pre-Cenozoic sediment was also manipulated to fit the observed gravity anomaly.

4.6.2 Gravity Model Results

Gravity anomalies are dependent on the density of rocks. High gravity anomalies are indicative of high density rock, while low gravity anomalies represent lower density rock. It is this distinction which has allowed various sub-surface features to be observed across the Rockall-Faroe study area.
Lousy Bank and Rosemary Bank

The observed gravity data reveals an abrupt change in the gradient of the gravity anomaly of the north-western edge of the Lousy Bank. It is inferred that such an abrupt change in the gravity anomaly has been produced by sharp changes the thickness of underlying pre-Cenozoic sediment.

To test such a hypothesis models were constructed incorporating the following:

- Varying thicknesses of basalt lava flow and densities of pre-Cenozoic sediment (Figs. 4.75 – 4.78),
- An abrupt change in the gradient of the Moho (Fig. 4.79)
- Seaward-dipping reflectors (Fig. 4.80)

All models across the Lousy Bank and Rosemary Bank (Figs. 4.75 – 4.80) reveal pre-Cenozoic sediment underlying the Lousy Bank. Models represented in Figs. 4.75 – 4.78 all show a relatively sharp boundary between pre-Cenozoic sediment and basement beneath the western limb of the Lousy Bank. Such a boundary was consistent with the observed gravity anomaly irrespective of the density of the underlying pre-Cenozoic sediment or the thickness of basalt lava flows. However, the maximum thicknesses of the modelled pre-Cenozoic sediment coring the Lousy Bank are dependent on this density of sediment and thickness of basalt (Table 4.3).

The minimum thickness of pre-Cenozoic sediment (3.33 km) was achieved using a density of 2.25 Mg/m$^3$ and a basalt thickness of 2 km (Fig. 4.75). A density of 2.4 Mg/m$^3$ for pre-Cenozoic sediment and a basalt lava flow thickness of 2 km (Fig. 4.76) produced the greatest thickness of sediment at 6.39 km.

<table>
<thead>
<tr>
<th>Thickness of basalt lava flow (km)</th>
<th>Density of pre-Cenozoic sediment (Mg/m$^3$)</th>
<th>Maximum thickness of pre-Cenozoic sediment (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.0</td>
<td>2.25</td>
<td>3.33</td>
</tr>
<tr>
<td>2.0</td>
<td>2.4</td>
<td>6.39</td>
</tr>
<tr>
<td>1.0</td>
<td>2.25</td>
<td>3.79</td>
</tr>
<tr>
<td>1.0</td>
<td>2.4</td>
<td>5.22</td>
</tr>
</tbody>
</table>

Table 4.3. Maximum thickness of pre-Cenozoic sediment within the Lousy Bank.
Fig. 4.75. Gravity profile across the Rosemary Bank and the Lousy Bank. The basalt thickness overlying Lousy Bank is 2 km (Funck et al., 2008). The pre-Cenozoic sediment has a density of 2.25 Mg/m³. Note the match between the calculated and observed gravity anomalies of an igneous intrusion beneath Rosemary Bank and sediment coring the Lousy Bank. Crustal thickness based on Klingelhofer et al. (2005). Densities of rocks are shown in Mg/m³. The background density used in this model is 3.0 Mg/m³.
Fig. 4.76. Gravity profile across the Rosemary Bank and the Lousy Bank. The basalt thickness overlying Lousy Bank is 2 km (Funck et al., 2008). The pre-Cenozoic sediment has a density of 2.4 Mg/m$^3$. Note the match between the calculated and observed gravity anomalies of an igneous intrusion beneath Rosemary Bank and sediment coring the Lousy Bank. Crustal thickness based on Klingelhofer et al. (2005). Densities of rocks are shown in Mg/m$^3$. The background density used in this model is 3.0 Mg/m$^3$. 

![Gravity profile diagram](image-url)
Fig. 4.77. Gravity profile across the Rosemary Bank and the Lousy Bank. The basalt thickness overlying Lousy Bank is 1 km. The pre-Cenozoic sediment has a density of 2.25 Mg/m$^3$. Note the match between the calculated and observed gravity anomalies of an igneous intrusion beneath Rosemary Bank and sediment coring the Lousy Bank. Crustal thickness based on Klingelhöfer et al. (2005). Densities of rocks are shown in Mg/m$^3$. The background density used in this model is 3.0 Mg/m$^3$. 
Fig. 4.78. Gravity profile across the Rosemary Bank and the Lousy Bank. The basalt thickness overlying Lousy Bank is 1 km. The pre-Cenozoic sediment has a density of 2.4 Mg/m³. Note the match between the calculated and observed gravity anomalies of an igneous intrusion beneath Rosemary Bank and sediment coring the Lousy Bank. Crustal thickness based on Klingelhöfer et al. (2005). Densities of rocks shown in Mg/m³. The background density used in this model is 3.0 Mg/m³.
Fig. 4.79. Gravity profile across the Rosemary Bank and the Lousy Bank with an abrupt change in the gradient of the Moho underlying the Lousy Bank. The basalt thickness overlying Lousy Bank is 1 km. The pre-Cenozoic sediment has a density of 2.40 Mg/m$^3$. Crustal thickness based on Klingelhöfer et al. (2005). Note the relatively low calculated gravity anomaly, west of the Lousy Bank, from an abrupt change in the gradient of the Moho. Densities of rocks shown in Mg/m$^3$. The background density used in this model is 3.0 Mg/m$^3$. 
Fig. 4.80. Gravity profile across the Rosemary Bank and the Lousy Bank with seaward-dipping reflectors (SDR) just off the northwest limb of the Lousy Bank. The basalt thickness overlying Lousy Bank is 1 km. Crustal thickness based on Klingelhöfer et al. (2005). The pre-Cenozoic sediment has a density of 2.40 Mg/m³. Note the relatively low calculated gravity anomaly, west of the Lousy Bank, from seaward-dipping reflectors with the same density as the basalt lava flows. Densities of rocks shown in Mg/m³. The background density used in this model is 3.0 Mg/m³.
A model incorporating a change in Moho depth was also constructed in an attempt to best fit the observed gravity data without any major changes in the thickness of pre-Cenozoic sediment (Fig. 4.79). However, a change in the depth of the Moho did not produce the observed gravity anomaly.

Seaward-dipping reflectors incorporated in modelling (Fig. 4.80) also did not produce the observed gravity anomaly across the west limb of the Lousy Bank. According to White (1988), the extrusive basalt comprising seaward-dipping reflectors is 3-6 km thick along the North Atlantic Margins. This was used to help constrain the thickness of the seaward-dipping reflectors on the western limb of the Lousy Bank.

The models represented in Figs. 4.75 – 4.78 best match the observed gravity data. However, a pre-Cenozoic sediment density of 2.4 Mg/m$^3$ is most likely more realistic for sediment bounded by or just overlying basement rock. In addition, a basalt lava flow thickness of 2 km on Lousy Bank has been previously modelled by Funck et al. (2008). The preferred model across the Lousy Bank is thus Fig. 4.76 with a maximum modelled thickness of pre-Cenozoic sediment of 6.39 km.

Modelling has revealed that the pre-Cenozoic sediment unit underlying the basalt lava flow is bounded by basement highs and is approximately 90 km wide (Fig. 4.76). This width coincides with the axis and limbs of the Lousy Bank Anticline. Unlike the Lousy Bank, the Rosemary Bank is modelled as high density rocks consistent with the presence of an igneous intrusion. The observed gravity anomaly over the Rosemary Bank also suggests the presence of an inner core of denser rock. Pre-Cenozoic sediment up to 3.8 km thick is intruded by the Rosemary Bank seamount and is separated from sediment of a similar age in the Lousy Bank by a basement high (Fig. 4.76).

**Lousy, Bill Bailey’s and Faroe Banks**

Crustal thickness of the order of 25 km has been modelled beneath Lousy, Bill Bailey’s and Faroe Banks (Figs. 4.81 – 4.83). The modelling of thinned crust (channels) between the banks is based on previous work by Funck et al. (2008). In the channels between the banks 17 km thick crust is modelled and fits the observed
Fig. 4.81. Gravity profile across the Lousy, Bill Bailey’s and Faroe Banks. The underlying structure of the Moho is based on Funck et al. (2008). The basalt thickness overlying Lousy Bank is ~ 2 km (Funck et al., 2008). The pre-Cenozoic sediment has a density of 2.25 Mg/m³. Note the match between the calculated and observed gravity anomalies with the banks cored by pre-Cenozoic sediment. Densities of rocks shown in Mg/m³. The background density used in this model is 2.985 Mg/m³.
Fig. 4.82. Gravity profile across the Lousy, Bill Bailey’s and Faroe Banks. The underlying structure of the Moho is based on Funck et al. (2008). The basalt thickness overlying Lousy Bank is ~ 2 km (Funck et al., 2008). The pre-Cenozoic sediment has a density of 2.4 Mg/m³. Note the match between the calculated and observed gravity anomalies with the banks cored by pre-Cenozoic sediment. Densities of rocks are shown in Mg/m³. The background density used in this model is 2.99 Mg/m³.
Fig. 4.83. Gravity profile across the Lousy, Bill Bailey’s and Faroe Banks. The underlying structure of the Moho is based on Funck et al. (2008). The basalt thickness overlying Lousy Bank is ~ 1 km. The pre-Cenozoic sediment has a density of 2.4 Mg/m$^3$. Note the match between the calculated and observed gravity anomalies with the banks cored by pre-Cenozoic sediment. Densities of rocks are shown in Mg/m$^3$. The background density used in this model is 2.99 Mg/m$^3$. 
gravity anomalies. This crustal thickness, in the channels, is higher than the 8 km thick crust modelled by Funck et al. (2008). The banks are all cored by pre-Cenozoic sediment. The thickest depocentres of pre-Cenozoic sediment are not located directly beneath the axis of the anticlines. Table 4.4 shows the maximum thickness of pre-Cenozoic sediment beneath the banks.

<table>
<thead>
<tr>
<th>Bank</th>
<th>Density of pre-Cenozoic sediment (Mg/m³)</th>
<th>Maximum thickness of pre-Cenozoic sediment (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lousy</td>
<td>2.25</td>
<td>4.61</td>
</tr>
<tr>
<td></td>
<td>2.40</td>
<td>5.99</td>
</tr>
<tr>
<td>Bill Bailey’s</td>
<td>2.25</td>
<td>6.49</td>
</tr>
<tr>
<td></td>
<td>2.40</td>
<td>9.99</td>
</tr>
<tr>
<td>Faroe</td>
<td>2.25</td>
<td>5.65</td>
</tr>
<tr>
<td></td>
<td>2.40</td>
<td>7.99</td>
</tr>
</tbody>
</table>

Table 4.4. Maximum thickness of pre-Cenozoic sediment within the Lousy, Bill Bailey’s and Faroe Banks. The basalt thickness overlying Lousy Bank is ~ 2 km.

Where the basalt thickness overlying Lousy Bank is ~ 1 km (Fig. 4.83), the maximum thickness of pre-Cenozoic sediment, with a density of 2.4 Mg/m³, is 6.25 km. It should be noted that the maximum thickness of pre-Cenozoic sediment for all the banks does not correlate with the crest of the banks. The crest of the Faroe Bank, for example, appears to overlie a basement high.

**Bill Bailey’s Bank and Alpin Anticline**

Pre-Cenozoic sediment has been modelled with a thickness of 3.6 km and 5.3 km, for densities 2.25 and 2.4 Mg/m³ respectively, beneath the axis of the Bill Bailey’s Bank Anticline (Figs. 4.84 - 4.85). The pre-Cenozoic sediment decreases to ~ 1 km thickness beneath the Alpin Anticline where it is enclosed by two basement highs (Fig. 4.84 - 4.85). Basement highs, up to 3 km high, are also located beneath the limbs of the Bill Bailey’s Bank Anticline.

**Wyville-Thomson and North Ymir Ridges**

Gravity modelling suggests that the Wyville-Thomson Ridge and North Ymir Ridge Anticlines are cored by pre-Cenozoic sediment (Figs. 4.86 - 4.89). The thickest pre-Cenozoic sediment is located between the two ridges and is 5 km and 8.1 km thick.
The pre-Cenozoic sediment has a density of 2.25 Mg/m$^3$. Note the calculated gravity anomaly of the sediments coring the Bill Bailey's Bank matches the observed gravity anomaly. Densities of rocks are shown in Mg/m$^3$. The background density used in this model is 2.987 Mg/m$^3$. 

Fig. 4.84. Gravity profile across the Alpin Anticline and the Bill Bailey's Bank. Crustal thickness based on Funck et al. (2008).
Fig. 4.85. Gravity profile across the Alpin Anticline and the Bill Bailey’s Bank. Crustal thickness based on Funck et al. (2008). The pre-Cenozoic sediment has a density of 2.4 Mg/m³. Note the calculated gravity anomaly of the sediments coring the Bill Bailey’s Bank matches the observed gravity anomaly. Densities of rocks are shown in Mg/m³. The background density used in this model is 2.99 Mg/m³.
Fig. 4.86. Gravity profile across the Wyville-Thomson and North Ymir Ridges. The crustal thickness is based on Tate et al. (1999) and the basalt thickness (~1 km) is based on Klingelhöfer et al. (2005) and Archer et al. (2005). The density of the pre-Cenozoic sediment is 2.25 Mg/m³. Note the match between the calculated and observed gravity anomalies of the sediment coring the Wyville Thomson and North Ymir Ridges. Densities of rocks are shown in Mg/m³. The background density used in this model is 3.0 Mg/m³.
Fig. 4.87. Gravity profile across the Wyville-Thomson and North Ymir Ridges. The crustal thickness is based on Tate et al. (1999) and the basalt thickness (~ 1 km) is based on Klingelhöfer et al. (2005) and Archer et al. (2005). The density of the pre-Cenozoic sediment is 2.4 Mg/m³. Note the match between the calculated and observed gravity anomalies of the sediment coring the Wyville Thomson and North Ymir Ridges. Densities of rocks are shown in Mg/m³. The background density used in this model is 2.97 Mg/m³.
Fig. 4.88. Gravity profile across the Wyville-Thomson and North Ymir Ridges. The crustal thickness is based on Tate et al. (1999). The basalt thickness is ~ 2 km. The density of the pre-Cenozoic sediment is 2.25 Mg/m³. Note the match between the calculated and observed gravity anomalies of the sediment coring the Wyville-Thomson and North Ymir Ridges. Densities of rocks are shown in Mg/m³. The background density used in this model is 3.0 Mg/m³.
Fig. 4.89. Gravity profile across the Wyville-Thomson and North Ymir Ridges. The crustal thickness is based on Tate et al. (1999). The basalt thickness is ~2 km. The density of the pre-Cenozoic sediment is 2.4 Mg/m³. Note the match between the calculated and observed gravity anomalies of the sediment coring the Wyville-Thomson and North Ymir Ridges. Densities of rocks are shown in Mg/m³. The background density used in this model is 2.97 Mg/m³.
Fig. 4.90. Gravity profile across the Rockall and Hatton Banks. The crustal thickness is based on White et al. (1987), while the basalt thickness is based on Funck et al. (2008). The density of the pre-Cenozoic sediment is 2.25 Mg/m$^3$. Note the match between the calculated and observed gravity anomalies of the igneous intrusions of the Swithin and Mammal seamounts and of the relatively thick sediment overlying basement in the Hatton Bank. Densities of rocks are shown in Mg/m$^3$. The background density used in this model is 3.0 Mg/m$^3$. 
Fig. 4.91. Gravity profile across the Rockall and Hatton Banks. The crustal thickness is based on White et al. (1987), while the basalt thickness is based on Funck et al. (2008). The density of the pre-Cenozoic sediment is 2.4 Mg/m³. Note the match between the calculated and observed gravity anomalies of the igneous intrusions of the Swithin and Mammal seamounts and of the relatively thick sediment overlying basement in the Hatton Bank. Densities of rocks are shown in Mg/m³. The background density used in this model is 2.997 Mg/m³.
for densities of 2.25 Mg/m³ and 2.4 Mg/m³ respectively. The pre-Cenozoic sediment, with a density of 2.4 Mg/m³, beneath the ridges is separated into four distinct sediment depocentres separated by basement highs (Fig. 4.87). The crest of the Ymir Ridge overlies a basement high while the crest of the Wyville-Thomson Ridge overlies a sediment depocentre, ~15 km wide and up to 7.3 km in thickness, surrounded by two distinct basement highs (Fig. 4.87).

**Rockall and Hatton Banks**

Pre-Cenozoic sediment has been modelled within Hatton Bank. Here, these sediments have maximum thicknesses of 3.5 km and 5.9 km, for densities 2.25 Mg/m³ and 2.4 Mg/m³ respectively (Figs. 4.90 – 4.91). The pre-Cenozoic sediment, of density of 2.4 Mg/m³, continues into the Hatton-Rockall Basin where it appears to be 4.2 km thick in the vicinity of the Mammal seamount. The Mammal seamount is modelled as an igneous intrusion cored with higher density igneous rock. The basement (upper crust) becomes shallower upon approaching the Swithin seamount, which is also modelled as an igneous intrusion. The relative basement high at the Swithin igneous centre separates pre-Cenozoic sediments in the Hatton-Rockall Basin from sediment within the Rockall Bank. The Rockall Bank, pre-Cenozoic sediment, of density 2.4 Mg/m³, attains a maximum thickness of 4.6 km (Fig. 4.91).

### 4.6.3 Limitations in gravity models

The gravity models are crude representations of the possible underlying structures in the Rockall-Faroe area. Any interpretation of these structures should take into account the limitations in the gravity modelling:

- Magnetic anomalies were not calculated to refine the model. The magnitude of magnetic anomalies is dependent on basalt lava flows and seamounts. The magnetic anomalies could have, thus, been used to fine tune the thicknesses of basalt lava flows.
- Modifications to the background density were made to adjust the datum shift between the observed and calculated gravity anomalies. Thus, the modelling is relative rather than absolute.
• More than one structural possibility exists for these models. For example, the thickening of sediment can have similar effects on the calculated gravity anomalies in the model as thickening the lower crust. The gravity models should, thus, be seen as possible structures whose calculated gravity anomalies fits the observed data.
• The modelling does not allow for variations in densities within polygonal units which are used to build the models.

4.6.4 Interpretation
Although limitations in the gravity modelling exist, there are certain features which are evident in the underlying structures of the Rockall-Faroe area.
The abrupt change in the observed gravity anomaly, on Lousy Bank, can only be simulated by implementing a sharp boundary between pre-Cenozoic sediment and underlying basement. Gravity modelling has also shown that this sharp gravity gradient cannot be ascribed to abrupt changes between the lower crust and the mantle (Fig. 4.79) or by the presence of seaward-dipping reflectors of basalt (Fig. 4.80). Indeed, the sharp gravity gradients are most easily explained by rapid variations in pre-Cenozoic sediment and underlying basement rock (Figs. 4.75 – 4.78).
The observed gravity anomalies over the Hatton, Lousy, Bill Bailey’s and Faroe Banks, and the North Ymir and Wyville-Thomson Ridges cannot be replicated with high density igneous rocks in the core of these structures. This suggests that these structures are not volcanic seamounts.
The Bill Bailey’s Bank, Faroe Bank, Wyville-Thomson Ridge and Ymir Ridge, contain basement highs. These basement highs could represent horst blocks or may represent uplifted basement as a result of compression. The nature of these basement highs as compressional structures or horst blocks is examined in Section 4.7.2.
4.7 Assessment of the relative importance of Sediment Loading, Differential Subsidence and Compression in forming anticlines.

4.7.1 Sediment Loading

It has been recognised that sediment loads have the potential to produce antiforms in the NE Atlantic Margin (Fejerskov and Lindholm, 2000; Stuevold et al., 1992; Kjeldstad et al., 2003). The east limb of the Helland-Hansen Arch in the Vøring Basin, for example, has been attributed to differential loading (Stuevold et al., 1992). This sediment loading was pronounced, on one side of the fold, due to high sedimentation rates of a dense terrigenous Pliocene wedge on thick Cretaceous clays comprising the Helland-Hansen Arch (Stuevold et al., 1992). The west limb of the Helland-Hansen Arch was formed due to thermal subsidence during the Eocene and the Oligocene (Stuevold et al., 1992).

Criteria have been developed in this study to differentiate between anticlines formed by sediment loading and those that are a result of far field compression. These principles are based on previous work on the Helland Hansen Arch (Kjeldstad et al., 2003), but have also been refined by considering the potential effects of a sediment load, such as the Sula Sgeir fan (Fig. 4.47), on the underlying sediments (Fig. 4.92). Sediment loading could result in the following characteristics seen in Fig. 4.92c:

1. Compaction of underlying sediment directly beneath the sediment load
2. Formation of a dipping slope with overlying parallel sediment. There is no onlap of overlying sediment onto the slope of the antiform.
3. Thicker sediment at the crest of the antiform.
4. If there is mobilization of sediment from beneath the sediment load there could be a lack of distinct sediment layers within the antiform. This diapiric pillowing of material to form the antiform should bend the existing parallel overlying sediment layers.

It is also accepted in this study, that antiforms whose amplitudes are greater than the thickness of any juxtaposing sediment cannot be solely the result of sediment loading. These characteristics have been used to differentiate between antiforms, formed as a result of sediment loading, from those that are the result of compression.
In this study the Onika and Viera anticlines have the potential of being formed as a result of differential sediment loading. This is due to the distinct difference between the amounts of sediment deposited on opposite sides of the folds. However, the nature of the anticlines suggests that they are the result of compression.

Fig. 4.92. Schematic diagram showing the potential effects of a sediment load on underlying sediment. The diagram is based on the deposition of the Sula Sgeir Fan in the NE Rockall Trough (Fig. 4.47). A sediment load would compact underlying sediment to result in the formation of an apparent antiform away from the sediment load.

Onika Anticline

The Onika Anticline may be the result of differential sediment loading by the Sula Sgeir Fan (Fig. 4.47). The additional sediment load by the Sula Sgeir fan could have formed the east limb of the fold. If the Onika Anticline was the result of sediment loading it should have formed from Late Pliocene to Recent times based on the age of the Sula Sgeir fan (see Fig. 3.11). However, the distinct onlap of Oligocene sediment onto the C30 (Late Eocene) unconformity strongly suggests that the fold
was present in the Late Eocene. A Late Pliocene formation of the fold would also have potentially resulted in the folding of the C20 (late Early Miocene), the Late Miocene – Early Pliocene, and C10 (late Early Pliocene) unconformities. These unconformities, however, appear unfolded over the crest of the Onika Anticline (Fig. 4.49). In addition, the Sula Sgeir fan sediment load cannot account for the west limb of the fold.

The Ness anticline (Fig. 4.50), adjacent to the Onika Anticline, juxtaposes, and is parallel to, a basalt escarpment (Fig. 4.56). Furthermore, folding of the sediment layers within the Ness Anticline matches the morphology of the top basalt surface of the escarpment (Fig. 4.50). The formation of the Ness Anticline could, therefore, be attributed to buttressing of sediment against the basalt escarpment. If this buttressing is not the result of sediment loading, compression offers the most viable explanation for the formation of the Ness Anticline and by extension the proximal Onika Anticline.

**Viera Anticline**

The Viera Anticline is flanked by thicker sediment on its northern margin compared to its southern margin (Fig. 4.53). However, onlap of Oligocene sediment onto the C30 (Late Eocene) unconformity (Figs. 4.53 – 4.54) would suggest that the Viera Anticline was already formed at C30 time. An initial structural high would have thus existed in the position of the Viera Anticline in order to facilitate onlap by Early Oligocene sediment.
4.7.2 Differential Subsidence

Structures such as the Bill Bailey's Bank and the Lousy Bank have previously been attributed to differential subsidence. Banks or horst blocks which have undergone less extension than surrounding areas, subside to a lesser extent (Vanneste et al., 1995). This differential subsidence could result in the draping of overlying sediment strata over the banks and result in fold structures (Fig. 4.93).

Gravity models, in this study, have demonstrated that the Bill Bailey's Bank, Faroe Bank, Wyville-Thomson Ridge and the North Ymir Ridge, are likely to contain basement highs (Figs. 4.81 – 4.89) surrounded by relatively thick sediments. For structures, such as the Faroe Bank (Fig. 4.82) and the North Ymir Ridge (Fig. 4.89) the crest of the fold correlates with the crests of the basement high. In the light of this, the study assesses whether these structures are the result of compression (Fig. 4.94) or differential subsidence (Fig. 4.95).
Erosion of uplifted sediment and the emplacement of Paleocene basalt lava flows. Post-Paleocene compressional phases can result in the tightening of the fold and the formation of angular unconformities.

Compression results in the formation of a fold. There is uplift of sediments.

Sediment deposited within basin

Fig. 4.94. Schematic diagram of the formation of a basement high as a result of compression. Compression results in the uplift of basement rock and subsequent onlap of younger basalt lava flows.

Differential subsidence results in the draping of Paleocene basalt lava flows over the horst block.

Emplacement of basalt lava flows.

Sediments deposited in basins juxtaposed to an existing horst block.

Fig. 4.95. Schematic diagram of the formation of a basement high as a result of differential subsidence. Lateral varying extension results in the formation of a horst block (a). This is then followed by the emplacement of basalt lava flows (b), which drapes over the horst block during differential subsidence (c).
4.7.3 Interpretation

Whilst differential sediment loading and differential subsidence have the capacity to produce antiforms, these mechanisms cannot account for the timing of formation of folds in the study area. Differential sediment loading can enhance the apparent amplitude of folds, but cannot account for the distinct onlap defining the fold prior to the sediment load. Differential subsidence cannot account for the post-breakup Cenozoic unconformities which define structures such as the Wyville-Thomson Ridge and the North Ymir Ridge anticlines. Atlantic Ocean spreading in the early Early Eocene (Doré et al., 1999) precludes rifting on the margin after this time. In addition, post-breakup thermal subsidence was gradual (Skogseid and Eldholm, 1988). The timing of rifting and gradual nature of thermal subsidence suggests that any post-breakup differential subsidence was not episodic and could not have resulted in the post-breakup unconformities associated with these compressional structures. Compression can account for the onlap which defines unconformities present within folds. In addition, gravity modelling suggests the Lousy Bank does not contain any basement high and is cored by low density sediment (Fig. 4.78). It seems more likely, then, that this style of structure can be attributed to the inversion of sedimentary basin further supporting compression as the mechanism of formation. In Chapter 6, the study examines the potential mechanisms causing compression and assesses whether these mechanisms can account for the timing of compression observed in the Rockall-Faroe area.

4.8 Summary

Compressional structures, in the North Rockall-Faroe area, include folds (anticlines and synclines), and reverse faults. These structures differ in size, orientation and shape. The folds in the study area have long wavelengths and relatively small amplitudes which represent < 1 % shortening. The amplitudes of folds range from 0.36 km (west end of the Viera Anticline) to 2.2 km (Faroe Bank Channel Syncline). Many folds, such as the Hatton Bank, Lousy Bank and Faroe Bank Anticlines, have defined the overlying sea-bed bathymetry. There are three dominant orientations of folds and reverse faults - NE, NW and E-W trends. The axes of the Hatton Bank, Lousy Bank and South Faroe Bank Anticlines
are parallel to the adjacent NE-trending continent-ocean boundary. However, the NW-trending Bill Bailey's Bank and North Faroe Bank Anticlines also lie adjacent to the continent-ocean boundary. Other NW-trending structures include the Faroe Syncline and the Vine, Dawn, Wyville-Thomson Ridge, Ymir Ridge (North, Central and South), and Viera anticlines. The West Lewis reverse faults and the Bridge Anticline have NE strikes. E-W trends are exhibited by the Hatton Bank reverse faults, the Alpin and the Judd anticlines.

The limbs of the Onika Anticline and the Viera Anticline underlie different thicknesses of sediment on each side of the anticlines. The folds are thus prone to the effects of differential sediment loading. The onlap of sediment onto the folds however, suggests that these structures where present before sediment loading took place. Whilst differential sediment loading could have enhanced the growth of the Onika and Viera anticlines, these folds were initiated and formed as a result of compression.

Folds are less defined on isochrons from the early Late Oligocene. This may represent a waning of compressional activity in the Rockall-Faroe area from this time.

The Lousy, Bill Bailey's and Faroe Banks were previously ascribed to volcanic intrusions (Ritchie et al., 1999). However, gravity models suggest that these structures are cored by sediment and are not directly underlain by igneous intrusions. Gravity models also reveal that the Bill Bailey's Bank, Faroe Bank, Wyville-Thomson Ridge and the North Ymir Ridge, contain basement highs. These basement highs could be horst blocks or may be compressional in origin. However, differential subsidence, associated with the formation of horst blocks, fails to account for the post-Paleocene angular unconformities defining structures such as the Wyville-Thomson Ridge Anticline. If igneous intrusion and differential subsidence did not produce these structures, compression offers the most viable mechanism for their formation.
5.0 Sediment Distribution

5.1 Introduction
Sediment distribution in the Rockall-Faroe area varies temporally and spatially. The study has established that this area has undergone active deformation throughout the Cenozoic. Changes in sediment distribution, thus, are not simply the result of eustatic changes (Shannon et al., 2005b), but reflect the changes in the structural architecture of the margin as a result of compression. Basins can act as major sediment depocentres provided that sediment supply is sufficient. In contrast, sites of uplift impede the accumulation of sediments. Furthermore, uplifted areas can act as major sediment sources and barriers to prograding sediments. In this study, isochore maps were constructed to assess the variation in the thickness of sediment in the Rockall-Faroe area. The distribution of sediment allows further understanding of the evolution of compressional structures in the study area.

In addition to sediment distribution, sediment type also changes through time. As the margin evolves and becomes deeper there is a change from continental facies deposits (fluvial, lacustrine and lagoonal) to hemipelagic sediments, turbidites and contourites (Boillot, 1981). These changes in sediment type have been previously recorded in the Rockall-Faroe area. There is evidence in the Hatton-Rockall Basin, for example, of a Late Paleocene shallow or near-shore environment prior to ocean spreading of the Atlantic (Laughton et al., 1972). The Late Eocene (C30) unconformity marks a rapid change in depositional style in the Rockall Basin from fluvial/near-shore clastic sedimentation to deep-water mud and ooze deposition (McInroy et al., 2006). The lower Neogene succession (Miocene-early Pliocene) in the Rockall Trough and the Faroe Bank Channel is interpreted to comprise contourite sediments (Stoker et al., 2005c).

5.2 Isochores
Isochore maps represent the two-way time differences between known horizons. Assuming that there are no large variations in sonic velocities (~ 2000 m/s) in sediments, the isochore maps also depict the thicknesses of sediments (m) deposited
during a certain time interval. These isochores, thus, can reveal the changes in sediment distribution through time and space.

**Top Basalt – late Ypresian**
The top basalt to late Ypresian isochore (Figs. 5.1 – 5.2) reveals the absence of thick Early Eocene sediments between the Ymir Ridge (North, Central and South) and the Wyville-Thomson Ridge anticlines in the Auðhumla Basin Syncline. The Mordor, Onika and Viera anticlines, however, are marked by relatively thick Early Eocene sediments averaging ~ 600 m, ~ 500 m, and ~ 650 m respectively. The Faroe Syncline contains relatively thin Early Eocene sediment (averaging ~ 220 m in thickness). A small patch of relatively thick sediments (average thickness of ~ 400 m), 15 km in length is located close to the centre of the syncline.

**Late Ypresian – late Lutetian**
The late Ypresian – late Lutetian isochore (Figs. 5.3 – 5.4) reveals the presence of thick Middle Eocene sediment (~ 500 m thick) juxtaposed to the south of, and parallel to, the Wyville-Thomson Ridge Anticline. The sediment lies on the axes of the Bridge and Central Ymir Ridge anticlines. Unlike Early Eocene sediment, Middle Eocene sediments of the Mordor and Viera anticlines are relatively thin, < 200 m thick. The Onika Anticline, however, is marked by relatively thick sediment (~ 400 m thick). The sediment thickness within the Faroe Bank Channel Syncline decreases from ~ 500 m at the southern end to ~ 300 m at the northern end of the syncline.

**Late Lutetian – C30 (Late Eocene)**
Thick Late Eocene sediment, up to 700 m thick, is present within the NE Rockall Trough and is surrounded to the west by the Onika and South Ymir Ridge anticlines and to the north by the Bridge and Wyville-Thomson Ridge anticlines (Figs. 5.5 – 5.6). The sediment juxtaposes the late Ypresian – late Lutetian thick sediment package, which overlies the axes of the Bridge and Central Ymir Ridge anticlines (Fig. 5.4). However, thin Late Eocene sediment, < 100 m thick, is present on the
Fig. 5.1. Top basalt - late Ypresian isochore of the Rockall-Faroe area. Note the relatively thick Ypresian sediments in the Mordor, Onika and the Viera anticlines. Bathymetric contours (in metres) are represented by blue lines.
Fig. 5.2. Top basalt - late Ypresian isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown by dotted white lines. Note the thick Ypresian sediments within the Viera, Onika and Mordor anticlines. The thick sediments of the Onika and Mordor anticlines are partially bounded by basalt escarpments. There is a lack of Ypresian sediment in the Auðhumla Basin.
Fig. 5.3. Late Ypresian - late Lutetian isochore of the Rockall-Faroe area. The Mordor and Viera anticlines are devoid of sediment.
Fig. 5.4. Late Ypresian - late Lutetian isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. Relatively thick sediment is present in the positions of the Bridge, Central Ymir Ridge and Onika anticlines. The North Ymir Ridge, Mordor, South Ymir Ridge and Viera anticlines are devoid of sediment.
Fig. 5.5. Late Lutetian - C30 (Late Eocene) isochore of the Rockall-Faroe area. Note the relatively thick sediment to the SE of the Bridge Anticline.
Fig. 5.6. Late Lutetian - C30 (Late Eocene) isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. Note the relatively thick Mid-Eocene sediment between the Onika, Bridge and Wyville Thomson Ridge anticlines. The Mordor, Bridge, South Ymir Ridge, Onika and the Viera anticlines are devoid of Mid-Eocene sediment. There is no sediment thickness variation across the reverse fault from the West Lewis Ridge to the West Lewis Basin.
axes of the Bridge and Central Ymir Ridge anticlines (Fig. 5.6). Late Eocene sediment, ~ 350 m thick, is present within the Auðhumla Basin (Fig. 5.6).

C30 (Late Eocene) – Early Oligocene
Thick Early Oligocene sediment, with an average thickness of ~ 550 m, is present within the NE Rockall Trough (Figs. 5.7 – 5.8). These sediments are surrounded by the thick Late Eocene sediments (southeast of the Bridge Anticline), the Wyville-Thomson Ridge Anticline, the Hebridean Shelf, the West Lewis Inverted Basin, the Viera Anticline and the Onika Anticline. The Auðhumla Basin appears devoid of Early Oligocene sediment. In contrast, sediment up to 800 m thick is located in the Faroe Bank Channel at the southern end of the Faroe Bank Channel Syncline (Fig. 5.7). The Alpin Anticline, Mordor Anticline and the fault-propagation folds south of the Central Ymir Ridge Anticline all appear devoid of Early Oligocene sediments.

Early Oligocene – early Late Oligocene
There appears to be relatively thinner sediment between the Early Oligocene and early Late Oligocene stratigraphic surfaces in the Rockall-Faroe area (Figs. 5.9 – 5.10). Sediments, with an average thickness of ~ 450 m, are restricted to the SE of the Bridge Anticline in the NE Rockall Trough, west of the North Ymir Ridge Anticline and south of the Sigmundur seamount (Fig. 5.10).

Early Late Oligocene – C20 (late Early Miocene)
Early Late Oligocene to Early Miocene sediment, ~ 450 m thick, occurs in the North Rockall Trough, SE of the Alpin Anticline (Fig. 5.11). Sediment of a similar thickness is also located in the north-western section of the NE Rockall Trough (Fig. 5.12). Other areas appear devoid of sediments of this time interval.

C20 (late Early Miocene) – Late Miocene-Early Pliocene
There is a lack of Middle to Late Miocene sediment in the North and NE Rockall Trough (Figs. 5.13 – 5.14). This may be due to the erosional nature of the Late Miocene-Early Pliocene unconformity evident by the truncation of the C20 unconformity (see Fig. 4.49). This erosion by the Late Miocene-Early Pliocene
Fig. 5.7. C30 (Late Eocene) - Early Oligocene isochore of the Rockall-Faroe area. Note the relatively thick sediment to the north of the Viera Anticline and the lack of sediment south of the fold.
Fig. 5.8. C30 (Late Eocene) - Early Oligocene isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. Note the lack of Rupelian (Early Oligocene) sediment in the Onika and Viera anticlines. Sediment decreases relatively abruptly from the north section of the West Lewis Ridge to the juxtaposed West Lewis Basin across the reverse fault.
Fig. 5.9. Early Oligocene - early Late Oligocene isochore of the Rockall-Faroe area. Note the lack of sediment at this time interval.
Fig. 5.10. Early Oligocene-early Late Oligocene isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. The Bridge, Onika and Viera anticlines are devoid of sediment.
Fig. 5.11. Early Late Oligocene - C20 (late Early Miocene) isochore of the Rockall-Faroe area. Note the relatively thick sediment in the North Rockall Trough southwest of the Alpin Anticline. These sediments have been inferred as contourite deposits.
Fig. 5.12. Early Late Oligocene - C20 (late Early Miocene) isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. The Bride, Onika and Viera anticlines are devoid of sediment. Relatively thick sediment is present to the west and south west of the Onika Anticline.
Fig. 5.13. C20 (late Early Miocene) - Late Miocene-Early Pliocene Isochore of the Rockall-Faroe area. The thickest sediment in this time interval is located in the Hatton-Rockall Basin.
Fig. 5.14. C20 (late Early Miocene) - Late Miocene-Early Pliocene isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. Note the lack of Mid- to Late Miocene sediment.
unconformity could have removed Mid-Late Miocene sediments in the North and NE Rockall Trough.

**Late Miocene-Early Pliocene – C10 (late Early Pliocene)**

Relatively little Late Miocene-Early Pliocene to late Early Pliocene (C10) sediment is found throughout the study area (Fig. 5.15). This may be due to the short time interval coupled with the erosion of sediment by the C10 unconformity. However, relatively thick sediment, ~400 m thick, is located southwest of Bill Bailey’s Bank (Fig. 5.15) and within the NE Rockall Trough (Fig. 5.16).

**Top basalt – C30 (Late Eocene)**

The top basalt – C30 (Late Eocene) isochore reveals relatively thick Eocene sediment, at least 1 km in thickness, located in the position of the Alpin Anticline (Figs. 5.17 – 5.18). The thick sediment is bounded by the Sigmundur, Darwin and Rosemary Bank seamounts suggesting deposition within a basin. The NE Rockall Trough also contains relatively thick sediment having an average thickness of ~900 m. This thickness of Early to Late Eocene sediment is evident across the Onika Anticline, where it is bounded by a basalt escarpment (Fig. 5.18) suggesting the Onika Anticline was formed within a basin.

**C30 (Late Eocene) – C10 (late Early Pliocene)**

The C30 (Late Eocene) – C10 (late Early Pliocene) sediment is relatively thin over the position of anticlines (Figs. 5.19 – 5.20). The crest of the Alpin Anticline, for example, is devoid of Oligocene to Early Pliocene sediment. The lack of Oligocene – Early Pliocene sediment is also apparent across the Mordor, Bridge, Onika and Viera anticlines and the Auðhumla Basin Syncline (Fig. 5.20). Relatively thick Oligocene to Early Pliocene sediment, ~1000 m thick, is present in the NE Rockall Trough, to the north and west of the West Lewis Ridge. The North Rockall Trough contains relatively thick sediment, >1000 m thick (Fig. 5.19). In the Faroe Bank Channel, northwest of, and adjacent to, the Faroe Channel Knoll, there is an accumulation of relatively thick sediment, up to 1000 m thick (Fig. 5.20).
Fig. 5.15. Late Miocene-Early Pliocene - C10 (late Early Pliocene) isochore of the Rockall-Faroe area. Note this interval contains little sediment with the exception of the area northeast of the Onika Anticline and southwest of the Bill Bailey's Bank Anticline.
Fig. 5.16. Late Miocene-Early Pliocene - C10 (late Early Pliocene) isochore from Rosemary Bank to Faroe Bank Channel. Basalt escarpments shown in dotted white lines. Note the lack of Early Pliocene sediment present in the north study area. Area of relatively thick sediment is located northeast of the Onika Anticline.
Fig. 5.17. Top basalt - C30 (Late Eocene) isochore of the Rockall-Faroe area. Note the relatively thick Eocene sediments in the Alpin, Mordor, Onika, and Viera anticlines and the Faroe Bank Channel syncline.
Fig. 5.18. Top basalt - C30 (Late Eocene) from the Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. Note the relatively thick Eocene sediment within the Alpin, Onika, and Viera anticlines.
Fig. 5.19. C30 (Late Eocene) - C10 (late Early Pliocene) isochore of the Rockall-Faroe area. Note the lack of sediment on the Alpin, Mordor, Onika and Viera anticlines.
Fig. 5.20. C30 (Late Eocene) - C10 (late Early Pliocene) isochore from Rosemary Bank to the Faroe Bank Channel. Basalt escarpments shown in dotted white lines. The Alpin, Mordor, Bridge, Onika and the Viera anticlines are devoid of Oligocene - Early Pliocene sediment.
5.2.1 Fan Systems

Tate et al. (1999) suggested the presence of a major fan system, Early Oligocene in age, located in the NE Rockall Trough to the east of the Onika anticline. In this study, the C30 (Late Eocene) – Early Oligocene isochore reveals that this interpreted Early Oligocene fan forms an anomalously thick package of sediment in the NE Rockall Trough (Fig. 5.8). On seismic data these sediments are characterized by relatively chaotic and discontinuous reflectors (see Fig. 4.47). Older anomalously thick sediment wedges, also characterized seismically by chaotic and discontinuous reflectors, such as the Late Eocene sediment wedge on the east limb of the Bridge Anticline (see Fig. 4.45), have been interpreted, in this study, as fan systems. The following is a description of each interpreted fan system in the Wyville-Thomson Ridge and Ymir Ridge area and the NE Rockall Trough:

Middle Eocene

The Middle Eocene postulated fan system is delineated by the late Ypresian – late Lutetian isochore (Fig. 5.4) and is located south of the Wyville-Thomson Ridge Anticline. It is represented by sediments with thicknesses of at least 300m and lies above the axes of the Bridge and Central Ymir Ridge anticlines (Fig. 5.21). The Middle Eocene sediment comprising the fan, downlaps and onlaps onto the late Ypresian unconformity on the northern limb of the Central Ymir Ridge Anticline (Fig. 5.22) and onlaps onto the top basalt of the southern limb of the Wyville-Thomson Ridge Anticline (Figs. 5.22 – 5.23). The Middle Eocene fan system does not overlie the axis of the North Ymir Ridge Anticline.

Late Eocene

The Late Eocene postulated fan system is well imaged by the late Lutetian – C30 (Late Eocene) isochore (Fig. 5.6) and is located to the southeast of the Bridge Anticline between the Middle Eocene and Early Oligocene fan systems (Figs. 5.21 and 5.24). Sediments in the distal part of the fan consist of mainly siltstone [sampled by well 164/07-1 (see Fig. 3.20)]. The fan thins towards the Onika and Ness anticlines and onlaps onto these folds (Fig.5.25). On the east limb on the Bridge
Fig. 5.21. Map of sediment packages, greater than 300 m thickness, from the Wyville-Thomson Ridge to the NE Rockall Trough. Sediment packages are interpreted as a series of submarine fans. The locations of wells 164/07-1 and 164/25-2, which sampled rocks in the distal parts of the Late Eocene and Early Oligocene fans respectively, are shown. The location of seismic illustrations are indicated by solid black lines.

KEY Fan systems

- Early Oligocene
- Late Eocene
- Middle Eocene
Fig. 5.22. N-S seismic profile of the Central Ymir Ridge and Wyville-Thomson Ridge anticlines (for location, see Fig. 5.21). Note the downlap and onlap of Mid Eocene sediment (pink shading) onto the late Ypresian unconformity on the Central Ymir Ridge Anticline and the onlap of these sediments onto the Wyville-Thomson Ridge Anticline. Late Eocene sediments onlap the late Lutetian unconformity on the north limb of the Central Ymir Ridge Anticline. Seismic data courtesy of Fugro Multi Client Services.
Fig. 5.23. N-S seismic profile of the Ymir Ridge plateau and the Bridge and Wyville-Thomson Ridge anticlines (for location see Fig. 5.21). Note the wedge of Mid Eocene sediment (pink shading) downlapping and onlapping the late Ypresian unconformity on the Bridge Anticline and onlapping onto the Wyville Thomson Ridge Anticline. Seismic data courtesy of Fugro Multi Client Services.
Fig. 5.24. NW-SE seismic profile across the Bridge Anticline and the NE Rockall Trough (for location see Fig. 5.21). Note the series of inferred sediment wedges - Mid Eocene (pink shading), Late Eocene (blue shading) and Early Oligocene (yellow shading) fans - on the Bridge Anticline and in the NE Rockall Trough. Seismic data courtesy of Fugro Multi Client Services and CGG Veritas.
Fig. 5.25. SW-NE seismic profile of the Onika and Ness anticlines (for location, see Fig. 5.21). Note the onlap of the inferred Late Eocene fan (blue shading) onto the late Lutetian unconformity of the Onika and the Ness anticlines. Seismic data courtesy of CGG Veritas.
Fig. 5.26. SW-NE seismic line across the north limb of the Viera Anticline and the south limb of the Wyville-Thomson Ridge Anticline (for location, see Fig. 5.21). The Early Oligocene sediment, in yellow shading, has been interpreted as a fan system by Tate et al. (1999). Seismic data courtesy of CGG Veritas.
Anticline, onlap of Late Eocene sediment onto the late Lutetian unconformity is evident (see Fig. 4.46).

**Early Oligocene**

The Early Oligocene postulated fan system is well imaged on the C30 (Late Eocene) – Early Oligocene isochore (Fig. 5.8). The fan is NE-trending, located within the NE Rockall Trough and following the orientation of this depocentre (Fig. 5.21). The distal part of the fan consists of mainly sandstone and siltstone [sampled by well 164/25-2 (see Fig. 3.20)]. The fan onlaps the Onika Anticline (see Fig. 4.49) and the Viera Anticline (Fig. 5.26). Onlap also occurs onto the Late Eocene fan (Fig. 5.24).

**5.3 Development of compressional structures and sediment distribution**

The distribution of sediment, in time and space in the study area, has been influenced by the development and growth of compressional features. The distribution of such sediment records and reflects the structural evolution of the area. Isochores, together with seismic and gravity profiles have been used to determine the evolution of folds through time in the study area.

**Wyville-Thomson Ridge Anticline**

The onlap of Middle Eocene sediment on the south limb of the Wyville-Thomson Ridge (Figs. 5.22 – 5.23) suggests that a structural high existed prior to the Mid-Eocene. The Wyville-Thomson Ridge Anticline contains an inferred Thanetian unconformity marked by onlap within basalt lava flows (see Figs. 4.36 and 4.40). It has been established, with the aid of gravity data, that the Wyville-Thomson Ridge is not underlain by a major igneous intrusion (see section 4.6.2). The Thanetian unconformity, thus, does not represent onlap onto a seamount. It is proposed in this study, that this unconformity represents Late Paleocene compressional growth of the Wyville-Thomson Ridge Anticline. Whilst the study establishes the origin of some post-Paleocene unconformities as compressional (see section 4.7.3), it is not possible to disprove the existence of this unconformity within the Wyville-Thomson Ridge as a result of differential subsidence (Vanneste et al., 1995) and uplift due to asthenospheric upwelling (Smallwood and Gill, 2002). The presence of a Thanetian
unconformity, which defines early inversion of the West Lewis Basin (see Fig. 4.51) and compression in the Faroe Shetland Basin (Nicholson, 2005), however, suggests a compressional origin for the Thanetian unconformity within the Wyville-Thomson Ridge Anticline. Previous work has been used to explain the development of the Wyville-Thomson Ridge. According to Tate et al. (1999) the Wyville-Thomson Ridge and Ymir Ridge anticlines developed about a composite, crustal-scale ramp-flat detachment surface during compression. This interpretation has been used in this study to explain unconformity formation and sediment distribution as a result of compression across the Wyville-Thomson Ridge and Central Ymir Ridge anticlines (Fig. 5.27).

North Ymir Ridge Anticline
The North Ymir Ridge Anticline, like the Wyville-Thomson Ridge Anticline, contains an inferred Thanetian unconformity. As with the Wyville-Thomson Ridge, this unconformity is proposed as compressional in origin. The presence of the fold in the Late Paleocene is further supported by the late Ypresian – late Lutetian isochore (Fig. 5.4). The lack of Middle Eocene sediment (inferred as fan deposits in section 5.2.1) on the fold’s axis further supports the presence of the North Ymir Ridge Anticline by the Mid-Eocene.

Central Ymir Ridge Anticline
The Central Ymir Ridge Anticline does not contain a Thanetian unconformity within basalt lava flows (Figs. 4.39 and 4.41). This suggests that the Central Ymir Ridge Anticline developed after Late Paleocene time. The late Ypresian – late Lutetian isochore supports this (Fig. 5.4), as relatively thick Middle Eocene sediment (interpreted as a fan in section 5.2.1) lie on the axis of the fold, suggesting that the fold was not present in the Mid Eocene. Seismic data reveal that the late Ypresian unconformity is folded above the crest of the Central Ymir Ridge Anticline representing folding after late Ypresian time (see Fig. 4.41). The onlap of Middle to Late Eocene sediment onto the late Ypresian unconformity on the southern limb of the fold (see Fig. 4.41) can be attributed to the late Ypresian growth of the Wyville-Thomson Ridge Anticline in the north (Fig. 5.27b).
In the late Lutetian the Central Ymir Ridge begins to form uplifting the overlying Mid Eocene fan.

Further late Ypresian growth results in subaerial erosion of the Wyville Thomson Ridge. Consequently a Mid Eocene fan is shed off the ridge towards the south.

Wyville-Thomson Ridge precursor develops on a north dipping reverse fault from a flat detachment surface. Detachment surface beneath the Wyville-Thomson Ridge Anticline was inferred by Tate et al. (1999). Onlap of basalt onto Thanetian precursor.

Fig. 5.27. Schematic diagram of the possible Early Cenozoic evolution of the Wyville-Thomson Ridge and the Central Ymir Ridge anticlines.
South Ymir Ridge Anticline

The late Lutetian – C30 (Late Eocene) isochore (Fig. 5.6) reveals that a relatively thick wedge, of Late Eocene sediment, is partially bounded to the west by the South Ymir Ridge Anticline. Although no seismic images of this sediment against the fold is available, onlap is assumed based on the onlap of the Late Eocene sediment onto the proximal Ness and Onika anticlines which also bound the thick sediment wedge (Fig. 5.25). The bounding of the Late Eocene sediment by the South Ymir Ridge Anticline suggest the fold existed by late Lutetian time.

Bridge Anticline

The Bridge Anticline bounds the north-western edge of relatively thick Late Eocene sediment (Fig. 5.6) suggesting that the fold existed by the late Lutetian. There is an accumulation of relatively thick Middle Eocene sediment overlying the crest of the Bridge Anticline (Fig. 5.4). There is onlap of Middle Eocene sediment on the top-basalt surface at the crest of the Bridge Anticline. There is also downlap onto the late Ypresian unconformity on the north-western limb of a small fold, which sits on the south-eastern limb of the Bridge Anticline (see Fig. 4.46). This style of onlap and downlap, at the base of the Middle Eocene sediment, can be formed as a result of sediments being sourced from the north or the south. The juxtaposition of the relative thick Middle Eocene sediment to the south of the Wyville-Thomson Ridge (Fig. 5.4) suggests that these sediments were sourced from the Wyville-Thomson Ridge to the north. The Late Ypresian growth phase of the Wyville-Thomson Ridge Anticline could have resulted in the erosion to form the Middle Eocene sediment wedge (Fig. 5.27b), inferred as a Middle Eocene fan system (section 5.2.1). The accumulation of the relatively thick Middle Eocene sediment on the axis of the Bridge anticline would suggest that the fold was not present at this time. The onlap of Ypresian sediment onto the top-basalt surface on the southeast limb of the Bridge Anticline (see Fig. 4.46) could be due to a slope formed from the Thanetian growth of the proximal Wyville-Thomson Ridge. The presence of thick Middle Eocene sediment and thin Late Eocene sediment over the crest of the Bridge Anticline (Figs 5.4 and 5.6) is interpreted as representing growth of the fold in the Late Lutetian (late Mid-Eocene). The formation of the Bridge Anticline in the late Lutetian would have
resulted in the uplift of the Middle Eocene sediment wedge overlying the axis of the fold. The uplifted sediment could have been eroded and shed away from the axis of the Bridge Anticline contributing to the Late Eocene fan system, which juxtaposes the Middle Eocene fan (Fig. 5.21) and sits just off the south-east limb of the Bridge Anticline (see Fig. 4.46).

**Mordor Anticline**

Thick Early Eocene sediment (Fig. 5.2) reveals a NNW-trending sediment basin at the position of the Mordor Anticline. This is interpreted from seismic data which reveals a basin bounded by basalt escarpments (see Fig. 4.31). In contrast, there is thinning of Middle Eocene sediments onto the Mordor Anticline (Fig. 5.4). Seismic data reveals that the late Ypresian unconformity of the Mordor Anticline is marked by onlap (see Fig. 5.31). The seismic data and the isochores, suggest that the initial growth of the Mordor Anticline occurred in the late Ypresian.

**Viera Anticline**

Relatively thick Early Eocene sediment (up to 800 m thick) at the position of the Viera Anticline delineates a NW-trending basin (Fig. 5.2). The late Ypresian – late Lutetian isochore (Fig. 5.4), however, reveals relatively thin Middle Eocene sediments, less than 200 m thick, at the position of the Viera Anticline. This might suggest that the growth of the Viera Anticline initiated in the late Ypresian. However, onlap of Mid Eocene sediment onto the late Ypresian unconformity cannot be confirmed due to the chaotic nature and discontinuity of reflectors within the sediment (see Fig. 4.53). Seismic data has revealed that at the west end of the fold there is no thinning of late Ypresian - late Lutetian and late Lutetian – C30 (Late Eocene) sediments towards the fold’s crest (see Fig. 4.53), suggesting that growth of the fold did not take place at these times.

It is proposed in this study that folding of the Viera Anticline initiated at C30 (Late Eocene) time. Seismic data reveals that there is marked onlap of Early Oligocene sediment, inferred as a fan (see section 5.2.1), onto the C30 unconformity (Fig. 5.26). The C30 (late Eocene) - Early Oligocene isochore (Fig. 5.8) shows that the Early Oligocene fan is bounded to the south by the Viera Anticline. This establishes the
presence of the fold at the Early Oligocene. On the north limb of the fold the Early Oligocene sediments are thick (up to 500 m), while the south limb lacks Early Oligocene sediments (Fig. 5.8). The disparity of the Early Oligocene sediments on the two limbs of the fold may be the result of the position of the fan’s major source. Tate et al. (1999) proposed that the Early Oligocene fan was sourced by canyons from the eastern margin of the NE Rockall Basin which extends along the Hebrides Shelf from the east end of the Wyville-Thomson Ridge to the West Lewis Basin. The thicker Early Oligocene sediment on the northern limb of the Viera Anticline is consistent with this interpretation.

West Lewis Basin
Late Eocene sediment has a constant thickness across the West Lewis Ridge and the adjacent West Lewis Basin across the reverse fault (Fig. 5.6). Early Oligocene sediment thickness, however, decreases abruptly across the reverse fault from the West Lewis Ridge to the West Lewis Basin (Fig. 5.8). This could have occurred as a result of a change in heights of these structures in relation to each other due to the inversion of the northern section of the West Lewis Basin in Late Eocene (C30) time. This is in agreement with the interpreted Late Eocene (C30) age for the reverse fault separating the West Lewis Ridge from the northern West Lewis Basin. Here, the reverse fault cuts the late Ypresian and late Lutetian unconformities and stops at the C30 unconformity (see Fig. 4.48).

5.4 Bottom-water current activity

5.4.1 Introduction
Bottom-water currents have the ability to erode and deposit sediments affecting the distribution of sediment and unconformities in time and space. The present-day circulation of bottom-water currents in the Rockall-Faroe area is shown in Fig. 5.28. In this chapter, the study assesses the possible role of compression in influencing, directly or indirectly, bottom-water current activity.
Fig. 5.28. Study area showing the present-day circulation of bottom-water currents (green arrows). Bottom-water currents based on Stow and Holbrook (1984), Stoker (1997) and Stoker et al. (2001). Note the bathymetric gorge between the Dawn and North Ymir Ridge Anticlines, within the Auðuhumla Basin (AB), through which the Norwegian Sea Overflow (NSO) passes. Locations of seismic illustrations are shown by solid black lines.
Fig. 5.29. N-S seismic profile north of the Alpin Anticline (for location, see Fig. 5.28). Note the upslope migrating Early Oligocene contourite onlapping the C30 unconformity. Sediments onlapping the early Late Oligocene unconformity are upward-convex and lenticular characteristic of confined contourite drifts.
Fig. 5.30. SW-NE seismic profile across the area between the North Rockall and NE Rockall Troughs (for location, see Fig. 5.28). Note the onlap onto the C30 (Late Eocene) unconformity of an inferred upslope migrating contourite drift. The unconformities present have been constrained using well data in the NE Rockall Trough. Seismic data courtesy of CGG Veritas.
According to Rebesco and Stow (2001), contourites are sediments deposited or substantially reworked in deep water by the action of bottom currents. Contourite deposition takes place in the last stages of passive margin development where the seabed is sufficiently deep enough to facilitate bottom-water current activity (Boillot, 1981). These bottom-water currents can result in the formation of unconformities (Andersen and Boldreel, 1995).

Two tectonic events have affected Cenozoic bottom-water current activity within the Rockall-Faroe area. These are the C30 sagging event and the submergence of the Greenland-Scotland Ridge.

### 5.4.2 C30 (Late Eocene) sagging

Sediment, with geometries characteristic of contourite drifts migrating upslope, onlap the C30 (Late Eocene) unconformity within the Rockall Trough. One such drift was previously studied within the Rockall Trough (Stoker, 1997; see Fig. 3.17). In this study seismic data also reveal an interpreted upslope migrating contourite drift onlapping the C30 unconformity north of the Alpin Anticline (Fig. 5.29) and between the North and NE Rockall Trough (Fig. 5.30). A subsidence curve devised from well data in the NE Rockall Trough shows an abrupt deepening of the Rockall Trough after C30 time (Fig. 5.31). This deepening was previously observed by other authors (Stoker et al., 2001; McInroy et al., 2006) and has been attributed to a reduction of dynamic support from the underlying mantle (Praeg et al., 2005). The abrupt post-C30 deepening of the Rockall Trough would have enhanced bottom-water current activity and facilitated the deposition of contourite drifts. These bottom-water currents flowed into the Rockall Trough from the south (Stoker, 1997 and Stoker et al., 2001).

![Subsidence curve for the NE Rockall Trough derived from well 164/25-2 (for location, see Fig. 3.6). The amount of subsidence for each time interval was calculated using a de-compaction method from Watts (2001). Porosities of rocks used in the calculations were based on Bond and Kominz (1984). For more details on the construction of this subsidence curve see Appendix B.](image)
5.4.3 Submergence of the Greenland-Scotland Ridge

The Norwegian Sea Overflow (NSO) flows across the Wyville-Thomson Ridge from the Faroe Bank Channel to the Rockall Trough (Stoker et al., 2005b). This flow has been attributed to the submergence of the Greenland-Scotland Ridge (Vogt, 1972; Schnitker, 1980; Wright and Miller, 1996). Various authors have dated the southward flow of bottom-water currents across the Greenland-Scotland Ridge from the Norwegian Sea as:

1. Late Eocene-early Oligocene (Miller and Tucholke, 1983; Davies et al., 2001)
2. Early-Mid Miocene (Vogt, 1972; Stoker et al., 2001)
3. Mid-Miocene (Blanc et al., 1980; Bohrmann et al., 1990; Eldholm, 1990; Ramsay et al., 1998; Stoker et al., 2005b)

Compelling $^{13}$C and $^{18}$O data from benthic foraminifera in borehole 116 in the Hatton-Rockall Basin reveal that the production of oxygenated deep water in the North Atlantic Ocean started in the late Middle Miocene (Blanc et al., 1980). The convergence of $^{13}$C and $^{18}$O values of benthic foraminifera at ~ 13 Ma from boreholes 555, 563 and 608 in the North Atlantic suggest that these sites were located within a single deep water mass in the late Middle Miocene (Ramsay et al., 1998). This is inferred as the initiation of the south flowing North Atlantic Deep Water (NADW) across the Greenland-Scotland Ridge (Ramsay et al., 1998). The flow over the Greenland-Scotland Ridge may even predate late Middle Miocene based on Wright and Miller (1996), who recorded high fluxes of Northern Component Water (NCW) in the Early Miocene. The NCW is inferred as being sourced from the Greenland and Norwegian Seas (Wright and Miller, 1996). However, the NCW could have originated in shelf areas to the south of the Greenland-Scotland Ridge (Wold, 1994; cited in Ramsay et al., 1998) and may thus not be representative of overflow across the ridge.
5.4.4 Bottom-water current unconformities

The onlap and erosional nature of the early Late Oligocene, late Early Miocene (C20), Late Miocene-Early Pliocene and the late Early Pliocene (C10) unconformities have been attributed, in this study, to bottom-water current activity. Herein are descriptions of the characteristics of the unconformities which have suggested their formation by bottom-water currents.

Early Late Oligocene

North of the Alpin Anticline this unconformity is marked by onlap of lenticular, upward-convex sediment, with peripheral channels. Whilst it remains difficult to distinguish between contourites, turbidites, hemipelagites and debrites in seismic units (Rebesco and Stow, 2001), these characteristics are typical of confined contourite drifts (Rebesco and Stow, 2001). This confined contourite drift onlaps onto an earlier upslope migrating contourite drift (Fig. 5.29). Onlap which defines the early Late Oligocene unconformity is also present within the NE Rockall Trough (see Fig. 4.47). Here the unconformity marks the onlap of Late Oligocene shale onto Early Oligocene sand (see Fig. 3.20). The Early Oligocene sand is interpreted as fan deposits (see section 5.2.1). Relatively thick Late Oligocene – Early Miocene sediments are present in the North Rockall Trough southwest of the Alpin Anticline (Fig. 5.12). These sediments onlap onto the early Late Oligocene unconformity (Fig. 5.32) and could represent contourite deposits. The early Late Oligocene unconformity is also characterized by erosional truncation in the Hatton-Rockall Basin (see Fig. 4.5).

C20 (late Early Miocene)

The C20 unconformity, which is marked by the onlap of smectite in the South Rockall Trough (Dolan, 1986), is defined by erosional truncation in the North Rockall Trough (Fig. 5.29) and in the Hatton-Rockall Basin (see Fig. 4.5). Onlap of sediment onto this unconformity on the northern limb of the Wyville-Thomson Ridge Anticline (see Figs. 4.36 and 4.40) and in the Lousy Bank Channel (see Fig. 4.16) has been inferred as being the result of changes in bottom-water current activity.
Fig. 5.32. NW-SE seismic profile within the North Rockall Trough (for location, see Fig. 5.28). Note the onlap marking the early Late Oligocene, C20 (late Early Miocene), Late Miocene-Early Pliocene, and the C10 (late Early Pliocene) unconformities. Seismic data courtesy Fugro Multi Client Services.
Fig. 5.33. NW-SE seismic profile of the north Rockall Bank margin (for location, see Fig. 5.28). Note the relatively thick Early Pliocene sediments deposited in an erosional scar carved by the Late Miocene-Early Pliocene unconformity.
Fig. 5.34. E-W seismic profile in the Hatton-Rockall Basin between the George Bligh and Rockall Banks (for location, see Fig. 5.28). Note the defined erosional truncation by the C10 (late Early Pliocene) unconformity.
Late Miocene – Early Pliocene

The Late Miocene-Early Pliocene unconformity is marked by erosional truncation on the northeast margin of the Rockall Bank where it truncates Late Oligocene – Late Miocene sediment (Fig. 5.33). The Late Miocene-Early Pliocene unconformity also truncates the C20 unconformity in the NE Rockall Trough (see Fig. 4.49) and the Early Oligocene and the late Early Oligocene unconformities on the eastern limb of the Mordor Anticline (see Fig. 4.31). An erosional channel defining the Late Miocene-Early Pliocene unconformity is also present within the North Rockall Trough (Fig. 5.29).

C10 (late Early Pliocene)

Erosional truncation by the C10 unconformity is strongly evident in the South Rockall Trough (see Fig. 3.8), the NE Rockall Trough (see Fig. 3.13) and between the George Bligh and Rockall Banks (Fig. 5.34). On the north limb of the Wyville-Thomson Ridge (see Figs. 4.36 and 4.40), the unconformity truncates the late Ypresian, late Lutetian, C30 (Late Eocene) unconformities and the Early Oligocene/early Late Oligocene/C20 (late Early Miocene) composite unconformity.

5.4.5 Influence of compression on bottom-water current activity

It has been suggested that compression could alter sea-bed bathymetry to result in changes in bottom-water current activity (Laberg et al., 2005). In this study there is evidence of compressional structures directing the flow of bottom-water currents. Bottom-water currents move into the North Rockall Trough from the Faroe-Shetland Channel via a bathymetric gorge (Fig. 5.28). The gorge lies at the west end of the Auðhumla Basin Syncline (Fig. 5.35). Previous studies have shown that this gorge has been a major pathway for both present (Dooley and Meincke, 1981) and past (Andersen and Boldreel, 1995) bottom-currents into the North Rockall Trough from the Faroe Bank Channel. The position of the Auðhumla Basin Syncline has controlled the flow of bottom-water currents into the North Rockall Trough and the subsequent distribution of unconformities and contourites within the Rockall Trough no later than Mid Miocene times (Stoker et al., 2005b). Structures, such as the
Fig. 5.35. SW-NE seismic profile of the Ymir Ridge and Wyville-Thomson Ridge anticlines in the west end of the Auðhumla Basin Syncline (for location, see Fig. 5.28). Note the erosional truncation within the Auðhumla Basin by the inferred composite unconformity. Seismic data courtesy of Fugro Multi Client Services.
Wyville-Thomson Ridge, could also have acted as major barriers to the southward flow of bottom-water currents from the Norwegian Sea prior to the Miocene.

5.5 Sediment distribution and eustasy

Eustasy (changes in sea level) has long been thought to control sediment distribution in quiescent basins. Changes in sea level, in these basins, have given rise to sediments which are bounded by unconformities and correlative conformities. These sequences develop as systems tracts (Fig. 5.36; Vail, 1987):

- Lowstand Systems Tracts develop due to sea-level fall. Sea level fall results in a lowering of base level to cause erosion. Erosional products may develop as submarine fan deposits on the basin floor.
- Transgressive Systems Tracts develop as sea level rises resulting in the landward onlap of sediments.
- Highstand Systems Tracts are due to high sea level resulting in the downlap of sediments towards the basin.

Fig. 5.36. Sequence Stratigraphy Systems Tracts formed in a eustatically controlled sedimentary basin. Abbreviations: LST = Lowstand Systems Tract, TST = Transgressive Systems Tract, HST = Highstand Systems Tract, MFS = maximum flooding surface, SB = sequence boundary. Diagram based on Vail (1987).
Sequence stratigraphy systems tracts are controlled by eustatic sea-level changes. However, it has been previously recognised in the Rockall Trough, that progradational features, such as fans and upslope migrating sediment drifts, are not produced by sea-level falls (Shannon et al., 2005b). Upslope migrating sediment drifts inferred as contourites have been attributed in previous studies to bottom-water current activity (Stoker, 1997; Stoker et al., 2001). In this study an attempt has been made to assess the possible role of sea-level changes (Fig. 5.37) in producing fan systems or sediment wedges in the Rockall-Faroe study area. Middle Eocene, Late Eocene, and Early Oligocene potential fan systems in the study area were previously discussed in section 5.2.1. In addition to these fan systems, an Early Miocene sandy wedge of sediment [see Fig. 3.20 (well 164/25-2)] and the Late Pliocene – Recent
Sula Sgeir Fan (see Fig. 3.12) are present in the NE Rockall Trough. The reflectors within both deposits are characterized by downlap. Fan or sediment wedge development in the late Mid Eocene – Late Eocene, Early Miocene and Late Pliocene – Recent correlate with falls in sea level (Fig. 5.37). The late Mid Eocene – Late Eocene progradational wedge, however, has been attributed to compressional uplift (see section 5.3). The Early Miocene and Late Pliocene-Recent fan systems are not contemporaneous with the growth phases of compressional structures in the Rockall-Faroe area and thus could represent lowstand fans deposited as a result of sea-level fall.

Late Oligocene mudstone, in the NE Rockall Trough, is characterised by parallel continuous reflectors which onlap Early Oligocene sandstone [see Fig. 3.20 (well 164/25-2) for sediment type and Fig. 4.49 for seismic image of sediment]. This onlap in the Late Oligocene is coupled with a rise in sea level and is inferred as being the result of a transgressive systems tract (Fig. 5.38). A rise in sea level would have resulted in deeper water and could have facilitated the deposition of mudstone.

In the Rockall-Faroe study area, it appears that eustasy plays a greater role in controlling the distribution of sediments in post-Oligocene times when compressional activity appeared to wane. Mid- and Late Eocene fan deposition in the Wyville-Thomson Ridge area, for example, has been interpreted to be the erosional products due to uplift of the Wyville-Thomson Ridge and the Bridge anticlines at times of compression. However, inferred Early Miocene and Late Pliocene – Recent lowstand fans in the NE Rockall Trough, occur during periods of apparent tectonic quiescence. It should be noted, however, that the sediment distribution due to sea-level changes can be interrupted by the erosive action of bottom-water current activity (Fig. 5.38). However, sea-level change can also complement deposition by bottom-water currents. A rise in sea level in the Late Oligocene, for example, is inferred to result in both a transgressive systems tract (Fig. 5.38) and contourite deposition in the North Rockall Trough (Fig. 5.29).
Fig. 5.38 (a) Seismic profile showing possible systems tracts in the NE Rockall Trough (for location see Fig. 5.28). Note the transgressive systems tract in Late Oligocene and the lowstand systems tract in Early Miocene (b) Global sea level change (Haq et al., 1987). Ages of Late Cenozoic unconformities in the NE Rockall Trough shown by dashed coloured lines while inferred sediment type represented by coloured rectangles (see key).
5.6 Summary

Sediment distribution in the Rockall-Faroe area has been influenced by both tectonic and bottom-water circulation events which have affected the margin. These events have also affected unconformity formation within the margin. The sediment distribution and unconformity formation are thus vital keys in understanding the regional tectonic phases and changes in bottom-water circulation in the Rockall-Faroe area.

The Early Eocene to Early Oligocene isochores, reveal a sequential change in the distribution of fan deposits in the Wyville-Thomson Ridge area. These changes can be related to the formation and growth of the Wyville-Thomson Ridge, North and Central Ymir Ridge and Bridge anticlines.

Anticlines, such as Mordor, Onika and Viera contain relatively thick Early Eocene deposits suggesting that the sediments were deposited in basins at this time. Thick Eocene sediment of the Alpin Anticline suggests that it too formed from the compression of basin sediment. Compression has affected the distribution of sediments in the Rockall-Faroe area. The growth of the Viera Anticline, for example, in C30 (Late Eocene) time has impeded the southward progradation of Oligocene fan deposits. The inversion of the north West Lewis Basin in C30 (Late Eocene) time has resulted in a decrease in sedimentation over the inverted basin in the Early Oligocene.

In addition, to compression, bottom-water current activity has influenced the timing and distribution of sediments. Contourites and unconformities relating to bottom-water current activity are evident in the North Rockall Trough and the Hatton-Rockall Basin. In the Rockall-Faroe area there is evidence, in the form of erosional truncation, to suggest that the early Late Oligocene, C20 (late Early Miocene), Late Miocene — Early Pliocene and C10 (late Early Pliocene) unconformities are the result of bottom-water current activity.
The influx of bottom-water currents into the North Rockall Trough from the Faroe Bank Channel takes place via a bathymetric gorge. The gorge lies between the Dawn and Ymir Ridge Anticlines in the west end of the Auðumla Basin Syncline. The position of this compressional feature has affected the distribution of contourites and bottom-water current unconformities within the North Rockall Trough.

The sediment distribution within the Rockall-Faroe area can primarily be attributed to tectonics and bottom-water current activity. The relatively abrupt subsidence of the Rockall Trough after the Late Eocene and the waning of compressional activity from the Early Oligocene have resulted in bottom-water currents being more dominant in controlling the distribution of sediment. In addition, eustatic changes in sea level could have formed lowstand and transgressive systems tracts in the NE Rockall Trough. However, the presence of submarine erosional unconformities and contourites throughout post-Eocene sediment suggest bottom-water current activity had a greater control on the distribution of sediment. Where sediment distribution can be linked in part to eustatic changes in sea level, as in the NE Rockall Trough, the expected systems tract sequence appear to be disrupted by the effects of bottom-water currents.
6.0 Driving Mechanisms

6.1 Introduction
Unconformities in the Rockall-Faroe area have been attributed, in this study, to compression and bottom-water current activity. The mechanisms for compression and the reasons for changes in bottom-water currents have been debated over the last twenty years. Hitherto, however, the ages of the stratigraphic surfaces that calibrate the timings of these events have not been known with any certainty throughout the Rockall-Faroe area. The well and extensive seismic data available in this study have facilitated the dating and mapping of these surfaces in order to assess their timing and nature. It is this timing and nature of horizons that has allowed the crucial link between unconformities and regional events.

6.2 Potential Compressional Mechanisms
Mechanisms which exert far-field horizontal stresses to the Rockall-Faroe area (Fig. 6.1) have the potential to form compressional features. These mechanisms are:

1. Ridge push – normal and hotspot-influenced ridge push
2. Alpine Compression
3. Pyrenean Compression
4. Depth-dependent stretching
5. Iceland Insular Margin body force

An attempt has been made, in this study, to evaluate the suitability of these mechanisms in forming compressional structures based on the timing of formation and the location of compressional structures.

6.2.1 Ridge Push
The term ridge push was first used by Forsyth and Uyeda (1975) to describe the postulated push exerted by mid-ocean ridges as a mechanism for the driving of the motion of the lithospheric plates. Orowan (1964) was the first to identify ridge push and attributed it to the force which drives the lithosphere away from the ridge due to the excess pressure beneath the ridge as a result of its height. The idea of plates
Fig. 6.1. Map of regional forces that have the potential to affect the Rockall-Faroe study area. The direction of forces are represented by arrows (for references see text). Bathymetric contours (in metres) are represented by blue lines. Abbreviations: CGFZ = Charlie Gibbs Fracture Zone, CSB = Celtic Sea Basin, DCG = Danish Central Graben, EJMFZ = East Jan Mayen Fracture Zone, FFZ = Faroe Fracture Zone, FSB = Faroe Shetland Basin, ISB = Irish Sea Basin, MB = Møre Basin, sNS = southern North Sea, WAB = Western Approaches Basin, WB = Wessex Basin.
diverging as a result of excess pressure at the ridge due to the elevation of the ridge crest was also suggested by Lliboutry (1969) and Parsons and Richter (1980). Similarly, Bott (1991, 1993) attributed ridge push to the horizontal pressure gradient as a result of the topographic high produced by the upwelling of asthenosphere. The horizontal pressure gradient could also be the result of cooling and thickening of oceanic lithosphere with time (Richardson, 1992). Ridge push has also been defined as the force produced as a result gravitational sliding where the oceanic lithosphere glides downwards over the lubricated underlying asthenosphere away from the ridge and towards the continental margin (Hales, 1969). Jacoby (1970) also attributed ridge push to the sliding of plates from a ridge elevated as a result of diapirism of the asthenosphere beneath the ridges.

Ridge push has been modelled numerically using elastic/viscoelastic finite element analysis (Bott, 1991; 1993). Model elements (grids) were given specific properties of density, volume expansion, viscosity and temperature to represent the parameters for oceanic crust, lithosphere and asthenosphere. During modelling the elements interacted with each other along boundaries simulating the plate boundary force produced by ocean ridges and stress distributions in adjacent plates.

Modelling by Bott (1993) was done for the following conditions:

a) Normal ridge with a spreading rate of 15 mm/yr (Fig. 6.2)
b) Normal ridge with a spreading rate of 60 mm/yr
c) Hotpot ridge with a spreading rate of 15mm/yr (Fig. 6.3)

The magnitudes of the ridge push forces for the three models were calculated using:

1. The integration of the principal stress: \( \int_0^T (\sigma_{xx} - \sigma_{zz}) \), where \( \sigma_{xx} \) and \( \sigma_{zz} \) are principal stresses in the x and z directions respectively. \( T \) is the depth to the base of the lithosphere.

2. The density moment function: \( \Delta h = -2\pi G/g \int_0^T g\Delta \rho \, dz = -2\pi G/g \, F_{rp} \),
   where z is depth, \( \Delta \rho \) is anomalous density and \( T \) is the depth of compensation.

The ridge push force \( (F_{rp}) \) is approximately proportional to the change in geoid height \( (\Delta h) \).
Fig. 6.2. Vertical displacement profile and deviatoric stresses for a model of a normal ridge spreading at 15 mm/yr (Bott, 1993).

Fig. 6.3. Vertical displacement profile and deviatoric stress for a model of a hotspot-influenced ridge spreading at 15 mm/yr (Bott, 1993).
Table 6.1. Computed values of plate boundary force developed at the edge of the models, in oceanic and continental lithosphere, as determined by: (1) finite element modelling, and (2) computation of the density-moment function. Results from Bott (1991).

The model results show that ridge push produced at a hotspot-influenced ridge imposed on old oceanic lithosphere is 2.5 times higher than that of a normal ridge (Table 6.1). According to Bott (1993) normal ridge push exerts a force of 40 MPa against the continental margin whilst hotspot-influenced ridge push exerts 100 MPa (Fig. 6.4; Bott, 1993). Increased ridge push for a hotspot-influenced ridge is due to increased shear drag acting on the base of the lithosphere, resulting from asthenospheric flow (Bott, 1991; 1993). There is also an increased pressure in the asthenosphere at the ridge crest contributing to the ridge push force (Bott, 1991; 1993). Bott’s model results also show that ridge push appears to be independent of spreading rate.
Normal ridge push, in the NE Atlantic Margin, initiated at the time of continental breakup at ~ 53 Ma (Doré et al., 1999) and would have continued until the present time. Hotspot-influenced ridge push, however, has not been continuous but occurred during distinct periods within the Cenozoic. The ages of hotspot-influenced ridge push can be determined using the ages of V-shaped ridges along the Reykjanes Ridge (Vogt, 1971; White and Lovell, 1997). The formation of V-shaped ridges as a result of the presence of a mantle plume is discussed in section 1.3.1.

6.2.2 Alpine Orogeny

The Alps formed during Early Cretaceous-Tertiary as a result of the collision between the Adriatic promontory of the African plate and the Eurasian plate (Ziegler, 1988). Ziegler (1988) defined three phases of Alpine compression:

- Late Alpine (Mid-Late Miocene)
- Main Alpine (Late Eocene – early Early Miocene)
- Early Alpine (Paleocene)

These phases were defined using the ages of nappe emplacement, molasses and flysch deposition and deformation in foreland basins in the Alps region. The direction of the Alpine force (Fig. 6.1) is based on previous work by Klein and Barr
(1986) where use had been made of in-situ measurements, earthquake focal-mechanism studies and breakout analysis.

6.2.3 Pyrenean Orogeny

The Pyrenean Orogeny resulted from the collision of Iberian microcontinent with Europe (Ziegler, 1988). This orogeny has been dated as Early Oligocene based on plate reconstructions, the age of emplacement of nappes and the transport of piggy-back basins towards the hinterlands (Williams, 1985; cited in Knott et al., 1993). Sinclair et al. (2005), however, recorded four phases of orogenic growth of the Pyrenees from Late Cretaceous to early Miocene times:

1. Inversion of extensional faults and the initiation of pro-wedge (75-58 Ma)
2. Growth of pro-wedge (58-47 Ma)
3. Growth of retro-wedge and cessation of pro-wedge growth (47-40 Ma)
4. Pro-wedge growth reactivation (Oligocene – Early Miocene)

According to Knott (1993) the Pyrenean Orogeny would have transmitted a NNE-directed compressive stress to the NE Atlantic Margin (Fig. 6.1).

6.2.4 Depth-Dependent Stretching

During the initiation of seafloor spreading, after continental rifting, lithospheric extension becomes greater than crustal extension (Kusznir et al., 2005). This was deduced from Tertiary subsidence patterns of the Lofoten, Vøring and Møre basins in the Norwegian continental margin. During continental rifting (pure shear) extension by brittle faulting in the upper crust is equal to extension by ductile deformation in the lower crust and lithosphere (McKenzie, 1978). However, pure shear stretching of the lithosphere leads to the upwelling of low-density asthenosphere. At breakup, there is greater lithospheric displacement close to the upwelling resulting in lithospheric extension exceeding continental extension in a 75-150 km marginal zone from the continent-ocean boundary towards the hinterland (Fig. 6.5: Kusznir et al., 2005). In the southern Lofoten and northern Vøring margins, for example, lithospheric and crustal extension during rifting was small ($\beta < 1.1$). Upon breakup at 54 Ma, lithosphere extension was larger ($\beta > 2.5$) while the extension in the crust remained small (Kusznir et al., 2005). Evidence for depth-dependent stretching has
been shown in other continental margins, such as the Goban Spur (Fig. 6.5: Kusznir et al., 2005), Vulcan Sub-basin, NW Australia (Baxter et al., 1999; cited in Kusznir et al., 2005), the Exmouth Plateau, NW Australia (Driscoll and Karner, 1998; cited in Kusznir et al., 2005) and the Galicia-Flemish conjugate margins (Sibuet, 1992; cited in Reston, 2007).

In late rifting or during incipient seafloor spreading, the displacement of the lithosphere close to the upwelling exceeds the lithospheric displacement far from the upwelling resulting in shortening within the intervening crust (Fig. 6.6: Withjack et al., 1998). This intervening crust may represent the 75-150 km marginal zone on the continental margin where lithospheric extension exceeds crustal extension (Kusznir et al., 2005).

Fig. 6.5. Depth-dependent stretching for the Goban Spur non-volcanic margin (Kusznir et al., 2005).

Fig. 6.6. Tectonic model for the inversion of basins in the continental margin of East USA (Withjack et al., 1998). Lithospheric displacements near the site of active asthenospheric upwelling are greater than those further away from the upwelling at the time of continental breakup. This results in crustal shortening in the intervening zone.

Reston (2007), however, has cast doubt on depth-dependent stretching affecting rifted margins. In non-volcanic margins, such as the south New Foundland Basin
and the south Iberia abyssal plain, the velocity structure has been used to show that the upper and lower crust appear to thin equally towards the margin (Reston, 2007). Reston (2007) attributes the discrepancies in extension recorded by Kusznir et al. (2005) as being the result of unrecognized polyphase and top basement faulting which resulted in the underestimation of upper crustal extension.

6.2.5 Iceland Insular Margin body force
Speculatively, the age of the Iceland bathymetric-topographic high formation has been placed within the Miocene (Doré et al., 2008). This is in agreement with the oldest rocks (~16 Ma old) which outcrop Iceland (Fougler, 2006). Compression at this time could be attributed to body forces exerted from the Iceland topographic high (Doré et al., 2008). The extra height developed on the ridge could have resulted in greater gravity wedging against the continental margin to result in increased ridge push (Fig. 6.7).

Fig. 6.7. Schematic illustration of relatively larger ridge push generated by the extra elevation of the Iceland topographic high. Diagram is based on the concept of gravitational sliding of oceanic crust and lithosphere from an elevated ridge (Jacoby, 1970).
6.3 Discussion – Effectiveness of mechanisms

6.3.1 Introduction
The timing of compressional unconformities in the Rockall-Faroe area must correlate with the timing of the compressional event(s) which formed them. The ages of all unconformities mapped and dated in the Rockall-Faroe area and regional events which have affected the NE Atlantic Margin are shown in Fig. 6.8. The following is an assessment of the potential of each mechanism in causing compression in the Rockall-Faroe area.

6.3.2 Ridge Push
Ridge push force is exerted onto the continental margins and thus offers a mechanism for compression along these margins. Normal ridge push, however, is not a viable mechanism for compression as it is continuous whereas the compressional phases recorded in the Rockall-Faroe study area are episodic. Lundin and Doré (2002) proposed that the main driving force for Cenozoic compression in the NE Atlantic is plume-enhanced or hotspot-influenced ridge push. Hotspot-influenced ridge push represents a good mechanism for compression based on the following:

1. The force is episodic as is the compression affecting the Rockall-Faroe study area.
2. Hotspot-influenced ridge push (100 MPa based on Bott, 1993) is 2.5 times the magnitude of normal ridge push. This magnitude should be sufficient to cause compressional deformation in the adjacent continent (Kusznir and Park, 1984; Molnar et al., 1993; cited in Lundin and Doré, 2002; Cloetingh and Van Wees, 2005).
3. The force extends along the entire continental margin. Compressional structures also occur along the margin, in the study area, from Hatton Bank anticline in the southwest to the Faroe Bank anticline in the northeast.
4. The timing of hotspot-influenced ridge push, based on the age of V-shaped ridges (Vogt, 1971; White and Lovell, 1997), correlate well with the timings
Fig. 6.8. Cenozoic unconformities in the Rockall-Faroe area and regional events affecting the NE Atlantic Margin.

KEY

<table>
<thead>
<tr>
<th>Unconformities</th>
<th>Compressional</th>
<th>Bottom-water current activity</th>
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1. Doré et al. (1999)
4. Sinclair et al. (2005)
7. Based on Kusznir et al. (2005)
8. Closure of the Panama isthmus (Keigwin 1982; Coates et al. 1992; Droxler et al. 1998)
9. Cessation of Messinian Crisis (Blanc & Duplessy 1982; Butler et al. 1999)
10. Iceland high formation (based on Fouger 2006) and consequent submergence of the proximal Auðnuma Basin Syncline due to Iceland topographic loading
11. Sea level rise (Haq et al. 1987)
of the late Ypresian, C30 (Late Eocene) and the Early Oligocene compressional unconformities (Fig. 6.8).

5. The regional stress directions in western Norway and the northern North Sea, determined from focal mechanisms, correlate well with ridge push directions (Fejerskov and Lindholm, 2000)

Hotspot-influenced ridge push, however, cannot account for the Thanetian unconformity (Fig. 6.8) which occurred prior to the initiation of ocean spreading. In addition, many of the compressional structures in the Rockall-Faroe area are not perpendicular to the ridge push force exerted. Other mechanisms or controls must, thus, be invoked to account for such structures.

The Iceland hotspot may have been located to the west of the Reykjanes Ridge prior to the late Early Miocene (Lawver and Müller, 1994, Torsvik et al., 2001). Reconstructions of Cenozoic plume movement (which place the Iceland plume under Greenland in the Early Cenozoic) assume that the plume is fixed in relation to the overlying plates (Lawver and Müller, 1994). In addition, the Iceland plume should have been situated in West Greenland in the Early Paleocene, in order for northwestards drift of the lithosphere to place the hotspot under present-day Iceland (Lundin and Doré, 2005). Consequently, any hotspot asthenospheric drag produced under the oceanic lithosphere would be towards the east. This drag should amplify ridge push affecting the NE Atlantic margin, but should, conversely, reduce the ridge push affecting the East Greenland margin (Fig. 6.9).
Fig. 6.9. Schematic diagram of the asthenospheric upwelling (mantle plume) affecting the North Atlantic in the Eocene. The position of the hotspot beneath Greenland is based on Clift et al. (1998) and Torsvik et al. (2001). The asthenospheric drag (black arrows) at the base of the lithosphere is directed towards the European Shelf [based on the Cape Verde Model (Courtney and White, 1986)]. Normal ridge push (blue arrow) towards the continental margins is affected by this basal drag. The magnitudes of ridge push and basal drag (units in MPa) are based on Bott (1993). The resultant forces (green arrows), produced by the addition of ridge push and basal drag, both move towards the southeast. Whilst this increases the ridge push towards the NE Atlantic Margin, the ridge push towards the Greenland continental margin should be negligible, if not non-existent.

However, some authors believe that the plume was ridge-centred in the Cenozoic, based on the presence of the Greenland-Scotland Ridge which could have formed due to spreading over a hot region of mantle (White and McKenzie, 1989; Fowler, 1990; Larsen et al., 1994). V-shaped ridges along the Reykjanes ridge also suggest that the plume was ridge-centred in the Cenozoic (Vogt, 1971; White et al., 1995; Ito et al., 1996; Ito, 2001). Although the presence of an Iceland plume is established, it still remains uncertain whether the plume was fixed in relation to the overriding plate and hence the position of the plume in the Cenozoic is still up for debate (Godfrey Fitton, 2008, personal communication). Whether the plume was ridge-centred or under Greenland in the Early Cenozoic, the presence of the Iceland plume would have enhanced the magnitude of ridge-push towards the NE Atlantic Margin.
6.3.3 Alpine Compression
Alpine compression occurred in three phases in the Cenozoic (Ziegler, 1988). The Thanetian, C30 (Late Eocene) and Early Oligocene compression events, in this study, occur at times of Alpine compression (Fig. 6.8). However, Alpine compression cannot account for compression in the late Ypresian and the late Lutetian. In addition, unlike hotspot-influenced ridge push, Alpine compression has taken place over relatively long periods of time. The main Alpine event, for example, extends continuously from Late Eocene and throughout the Oligocene (Fig. 6.8). Two distinct periods of inversion in the Late Eocene and Early Oligocene, in the Rockall-Faroe area, are more easily explained by separate compressional events. Thus, other mechanisms may be controlling compression in the Rockall-Faroe area.

6.3.4 Pyrenean Compression
Pyrenean compression extends from the Paleocene to Early Miocene (Sinclair et al., 2005). All observed compression in the Rockall-Faroe area occurs during this time (Fig. 6.8). However, like Alpine compression, Pyrenean compression is unlikely to produce the observed distinct episodes of compression over shorter periods of time in the Rockall-Faroe area.

6.3.5 Depth-dependent stretching
The push of the lithosphere towards the continental crust, which is associated with depth-dependent stretching (Kusznir et al., 2005), has been proposed for the inversion of basins on the continental margins of East USA and SE Canada (Whitjack et al., 1998). Depth-dependent stretching also has the potential of resulting in compression in the Rockall-Faroe area. The Thanetian unconformity, mapped and dated in the Rockall-Faroe area, may be the result of depth-dependent stretching, just prior to Atlantic opening. The Thanetian unconformity is dated in the south West Lewis Basin (see Fig. 3.10), marking the initial inversion of the basin. In the South Faroe Bank Anticline (see Fig. 4.21), Faroe Bank Channel Syncline (see Fig. 4.22) and the Wyville-Thomson Ridge Anticline (see Figs. 4.36 and 4.40) there is an unconformity, marked by the onlap of basalt lava flows that is inferred to be of Thanetian age. Poor imaging of
sub-basalt strata in the Hatton Bank, Lousy Bank and Bill Bailey's Bank precludes any mapping of the Thanetian unconformity which may define folding of these anticlines. In the Hatton Bank Anticline, however, there is onlap of Early Ypresian sediment onto the top-basalt surface (see Fig. 4.5) suggesting a structural high existed prior to the Early Ypresian. This structural high could have been formed as a result of compression.

Hatton Bank, Lousy Bank, Bill Bailey’s Bank and Faroe Bank anticlines are located within the marginal zone (75 – 150 km from the continent-ocean boundary) where compression by depth-dependent stretching has been predicted (section 6.2.4). The Wyville-Thomson Ridge Anticline and the inverted West Lewis Basin, however, are located outwith this marginal zone. The formation of these structures therefore requires another mechanism.

Asthenospheric upwelling is associated with depth-dependent stretching (Withjack et al., 1998; Kusznir et al., 2005). The Base Balder unconformity (Late Paleocene – Early Eocene) in the southern Faroe-Shetland Channel is attributed to uplift as a result of 'the introduction of asthenospheric mantle with an anomalously high potential temperature' (Smallwood and Gill, 2002). The Thanetian unconformity of the Rockall-Faroe area may be contemporaneous with the Base Balder unconformity (see section 3.3). Whilst Smallwood and Gill (2002) dismiss uplift as a result of compression, compression could result from the asthenospheric upwelling.

A hotspot, present prior to rifting, will cause considerable uplift – 1 km or more over a region with a diameter of 1500 - 2000 km (White and McKenzie, 1989). This produces significant gravitational potential which will assist rifting in the region (White and McKenzie, 1989). It is the view, in this study, that this gravitational potential could also facilitate compression on the periphery of the uplifted region. It has been recognized that uplifted areas (swells) releases gravitational potential energy away from the site of uplift. Oligocene-Holocene compression of the Kwanza Basin, located in the continental margin off Angola, for example, has been attributed to mantle-driven uplift of the African superswell (Hudec and Jackson, 2002). The Kwanza Basin lies in the northwest rim of the superswell resulting in a horizontal stress exerted on the basin. In volcanic margins, modelling has shown that there is uplift at the site of rifting prior to seafloor spreading (Skogseid and Eldholm,
Compressional forces could have been directed towards distal areas at the periphery of the uplifted area at the site of rifting. Depth-dependent stretching could account for the inferred Thanetian compression of folds in the marginal zone flanking the continent-ocean boundary – Hatton Bank, Lousy Bank, Bill Bailey's Bank and Faroe Bank anticlines. The associated active asthenosphere upwelling, which facilitated depth-dependent stretching (Kuznir et al., 2005) could have resulted in uplift of proximal areas which generated gravitational forces resulting in compression in peripheral areas. The forces generated by active asthenospheric upwelling could have been coupled with the effects of Alpine and Pyrenean compression also taking place at this time (Fig. 6.8).

6.3.6 Iceland Insular Margin body force and Mid Miocene compression

The formation of the Iceland topographic high and the resulting body force has been proposed for the formation of compressional structures in the Norwegian Margin in the Mid Miocene (Doré et al., 2008). Mid Miocene compression has been recorded in the Norwegian Margin (Lundin and Doré, 2002) and the Faroe-Shetland Basin (Davies and Cartwright, 2002; Ritchie et al., 2003). Boldreel and Andersen (1993) did infer a Miocene age for a stratigraphic surface thought to represent a phase of compression of the Hatton Bank and Wyville-Thomson Ridge anticlines in the Rockall-Faroe area. Johnson et al. (2005) also mapped an Intra-Miocene compressional unconformity for the Wyville-Thomson Ridge Anticline. Intra-Miocene compressional growth has also been recorded for the Ymir Ridge and Alpin Dome anticlines (Ritchie et al., 2008). This observed compressional growth in the Rockall-Faroe area together with contemporaneous growth in the Faroe-Shetland Basin and in the Vøring Basin produces an arcuate distribution of Mid Miocene compressional structures around Iceland. This suggests that compression was the result of a body force generated from the Iceland Insular Margin (Doré et al., 2008). However, the Intra-Miocene unconformity previously mapped in the Rockall-Faroe area (Boldreel and Anderson, 1993; Johnson et al., 2005; Ritchie et al., 2008) was poorly constrained. In this study no obvious Miocene unconformities associated with compression were recognized in the Rockall-Faroe study area. The C20 (late Early Miocene) is inferred as being the result of bottom-water current activity.
According to Doré et al. (2008) the acme of the development of compressional structures in the NE Atlantic Margin took place in Miocene times. However, evidence in this study suggests that any Miocene compression in the Rockall-Faroe area was not as significant as Mid Miocene compression in other areas of the NE Atlantic Margin. Such evidence includes the age of reverse faults and the nature of the C20 (late Early Miocene) unconformity in the Rockall-Faroe area.

Reverse faults
Well defined reverse faults in the study area do not appear to cut Early Miocene sediment. Reverse faults in the south limb of the Hatton Bank (see Fig. 4.5) and the north West Lewis Basin (see Fig. 4.48) are Late Eocene (C30) in age. The lack of reverse faulting in Miocene strata suggests that Mid Miocene compression was not strong in the Rockall-Faroe area.

C20 (late Early Miocene) unconformity
The C20 unconformity is unfolded above the Onika and Viera anticlines (see Figs. 4.49 and 4.53). Strong Mid Miocene compression of these anticlines should have resulted in the folding of this stratigraphic surface. In this study, the onlap marking the C20 unconformity on the eastern limb of the Lousy Bank Anticline, south limb of the Alpin Anticline, north limb of the Bill Bailey’s Bank Anticline and the north limb of the Wyville-Thomson Ridge has been attributed to a change in bottom-water current activity rather than compression. A major change in bottom-water circulation, affecting the North Atlantic, at this time is compelling (Wright and Miller, 1996; Stoker et al., 2005b). In addition, bottom-water current activity is the most likely explanation for the erosional nature of the C20 unconformity in some areas of the Rockall-Faroe area (see section 5.4.4). According to Stoker et al. (2005b), however, this change in bottom-water current circulation was the result of compression. Intra-Miocene compression resulted in the opening of the Faroe Conduit (Faroe-Shetland Channel and the Faroe Bank Channel). This allowed the passage of Norwegian Sea Deep Water (NSDW) and the subsequent overflow of bottom-water currents into the Rockall Trough. The formation of the Faroe Conduit involved the compressional growth of the Wyville-
Thomson Ridge (Stoker et al., 2005b). However, in this study another mechanism has been proposed for this change in bottom-water circulation in the Miocene (section 6.5).

It is believed in this study that Mid Miocene horizontal stress affecting the Rockall-Faroe area was not as strong as that affecting the Faroe-Shetland Basin where Mid Miocene compression is well-constrained (Ritchie et al., 2003). It is proposed that this discrepancy in Mid Miocene compression between the two areas is as a result of differences in the magnitude of ridge push against these areas at this time.

*Differences in Ridge Push*

The magnitude of ridge push can be a function of the gravity wedging effect due to the height of the ridge (Orowan, 1964; Lliboutry, 1969; Jacoby, 1970; Parsons and Richter, 1980 and Bott, 1991). The study proposes that the continental margin directly adjacent to the topographic high, at Iceland, would experience a relatively high magnitude of ridge push (Fig. 6.10). In addition, if a mantle plume centre was located at the position of Iceland, the plume head could have been moving radially away from Iceland based on the Cape Verde Model (see Fig. 1.14). The direction of asthenospheric drag, from the plume head, at the base of the Iceland-Faroe Rise is postulated to have been parallel to the direction of the force generated by gravity wedging (Fig. 6.10). Late Mid Miocene ages of V-shaped ridges (Fig. 6.8) would suggest asthenospheric drag at this time further amplifying ridge push against the Faroe Shelf and Faroe-Shetland Basin. South of the Iceland-Faroe Rise, oblique asthenospheric drag at the base of the oceanic lithosphere coupled with less gravity wedging from a lower Reykjanes Ridge could have resulted in a smaller ridge push against the Rockall-Faroe area in the Mid Miocene.
The increase in ridge height at Iceland ($\Delta h$) could have resulted in an increase in gravity wedging at the site of the Iceland topographic high. This coupled with parallel underlying asthenospheric drag at the base of the lithosphere (from an Iceland-centred plume) could have resulted in a greater ridge push against the Faroe Shelf compared to the ridge push affecting the Rockall-Faroe area.

Further north, the Vøring Basin, however, does not lie adjacent to the Iceland-Faroe Rise, but has undergone Miocene compression (Blystad et al., 1995; Lundin and Doré, 2002). The lack of Miocene compressional structures in the adjacent Møre Basin has been attributed to the Aegir Ridge absorbing the compressional force from the Iceland body force (Doré et al., 2008). Such an argument would also mean that the Vøring Basin should also be devoid of Miocene compressional structures. Doré et al. (2008) proposed that compression in the Vøring Basin was possible as a result of strain partitioning by lateral movement along the East Jan Mayen Fracture Zone. However, it still remains unclear in this study how the gravity wedging in the Iceland area could have affected the Vøring Basin via the East Jan Mayen Fracture Zone.

**6.3.7 Compression in the Late Lutetian**

The late Lutetian unconformity is the only compressional unconformity in the Rockall-Faroe area, since the opening of the Atlantic, which cannot be correlated to hotspot-influenced ridge push (Fig. 6.8). Although the Lutetian is a time of retro-wedge growth of the Pyrenees (Sinclair et al., 2005) the relatively long period of this compressional event is not likely to account for the distinct phase of compression present in the late Lutetian. In this study, it is proposed that the late Lutetian event
was triggered by hotspot-influenced ridge push, not recorded by the formation of V-shaped ridges. This is conceivable as a late Lutetian (42 Myr) phase of volcanism in the Rockall Trough has been attributed to the pulsing of hot Iceland plume material (O’Connor et al., 2000).

6.4 Controls on Basin Inversion

6.4.1 Introduction
Basin inversion has been inferred in other areas of the NE Atlantic Margin. In the Norwegian Margin, for example, the Helland-Hansen Arch and other large domes are thought to be due to compressional reactivation of normal faults developed during rifting in the Cretaceous (Grunnaleite and Gabrielsen, 1995; Vågnes et al., 1998; cited in Mosar et al., 2002). The Naglfar Dome, between the Vøring and Møre Basins (for the location of these basins, see Fig. 6.1), lies within the Hel Graben (Lundin and Doré, 2002). Lundin and Doré (2002) proposed that Mid- to Late Cenozoic compression resulted in the basin inversion within deep Cretaceous depocentres on the Norwegian-Greenland Sea margins. In the Faroe-Shetland Basin there is seismic evidence for the inversion of basins (Nicholson, 2005). In this study, the inversion of the West Lewis Basin is also evident.

6.4.2 Inversion of Normal Faults
In this study, gravity models across the Rockall-Faroe area suggests that structures such as Lousy, Bill Bailey and Faroe Banks are cored by low-density pre-Cenozoic sediments (Figs. 4.75 – 4.85). It is proposed, in this study, that the pre-Cenozoic sediments were deposited in sedimentary basins bounded by normal faults and it is the inversion about these faults that resulted in the inversion of the basins. Mohr-Coulomb diagrams (Fig. 6.11) were used to calculate the stresses that are required to invert pre-existing faults. The results are summarized in Table 6.2.
Fig. 6.11. Mohr-Coulomb diagram to determine horizontal forces needed to reactivate faults at particular dips, depths and angles of internal friction. Horizontal forces ($s_1$) are shown in italics. Dip of fault = $\theta$. Angles on the Mohr-Coulomb circles represent the intersection of a line at $2\theta$ from the centre of the circle to the circumference.
<table>
<thead>
<tr>
<th>Depth of fault (km)</th>
<th>Vertical Stress (MPa)</th>
<th>Internal angle of friction (°)</th>
<th>Dip of normal faults (°)</th>
<th>Horizontal Stress required for reactivation of fault (MPa)</th>
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<tr>
<td>2</td>
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<td>25</td>
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<td>120</td>
<td>30</td>
<td>45</td>
<td>444</td>
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</tbody>
</table>

Table 6.2. Horizontal stresses required for the reactivation of normal faults with different parameters of vertical stress, internal angle of friction and dip angles. Horizontal stresses were determined from Mohr-Coulomb diagrams (Fig. 6.11).

Hotspot-influenced ridge push exerts a force of 100 MPa onto the continental margin (Bott, 1993). This order of magnitude is sufficient to reactivate faults at shallow depth of 2 km (vertical stress = 40 MPa) at dips of 33°- 48° (Table 6.2). Listric normal faults, which bound sedimentary basins, can develop at these dips away from the ideal Andersonian dip of 60°.

Based on gravity models, however, sedimentary basins beneath structures, such as Lousy Bank, are ~ 6 km deep. At a low internal angle of 25°, 444 MPa is required to invert a normal fault at 54° (Table 6.2). Faults of a lower dip of 45° can be reactivated at a higher internal angle of 30° with the same stress magnitude. These results show that a higher force than that of hotspot-influenced ridge push is required to reactivate faults at 6 km depth.

The Helland-Hansen Arch, in the Vøring Basin, is believed to be the result of the inversion of normal faults developed during rifting (Grunnaleite and Gabrielsen, 1995; Vågnes et al., 1998; cited in Mosar et al., 2002). Kjeldstad et al. (2003) used Mohr-Coulomb failure criteria to show that reactivation of faults at 5 km depth (basement level) to form the Helland-Hansen Arch was not possible by the forces exerted by ridge push. According to Kjeldstad et al. (2003) this supports a non-compressional origin for the Helland-Hansen Arch.
However, it has been established that the Rockall-Faroe area does contain compressional structures. Thus, the inadequacy of hotspot-influenced ridge push to cause inversion of deep basin faults may simply mean that other mechanisms may be contributing stresses to result in compression.

An attempt has been made to evaluate the magnitudes of the other mechanisms affecting the study area. The magnitude of Alpine compression is difficult to constrain (Carrapa and Garcia-Castellanos, 2005). According to England et al. (1986), however, the force required for mountain building by crustal shortening is calculated at 150 – 250 MPa. If 250 MPa is transmitted due to the combined effect of both the Alpine and Pyrenean Orogenies, the Rockall-Faroe study area has the potential of having an intraplate compression stress of 350 MPa. This force is around the magnitude needed to result in the inversion of a normal fault (dip = 49°) at 6 km depth provided the internal angle of friction is low at 25°.

Hotspot-influenced ridge push is contemporaneous with Alpine and Pyrenean compression during C30 (Late Eocene) and Early Oligocene times (Fig. 6.8). The late Ypresian compressional event, however, only has a contributing force from the Pyrenean Orogeny (Fig. 6.8). However, normal faults with a dip of 45° at 4 km depth, requires a lower force of 219 MPa to be reactivated. It is conceivable that shallower faults could have been reactivated in the late Ypresian.

6.4.3 Basin Modelling

A basin model (Fig. 6.12) was constructed using SAVFEM in an attempt to simulate the forces needed to invert a basin. The model consists of sediment, basement (upper crust), lower crust and mantle rock. For more details of SAVFEM refer to section 1.5.4.

It has been previously recognized that basins within the NE Atlantic Margin follow the Caledonide (NE-SW) and Lewisian (NW-SE, N-S and E-W) structural trends which affect basement rock (Tate et al., 1999). Normal faults, with a 60° dip angle, are thus modelled to a depth of 12 km through syn-rift sediment and basement rock. The normal faults were modelled with a coefficient friction angle of 30° based on the average values for crustal rocks (Davis and Reynolds, 1996). The sediment only
reaches a maximum depth of 6.5 km based on gravity models for compressional structures, such as Lousy Bank (see Fig. 4.78), across the Rockall-Faroe area. The boundaries of the basin were subjected to forces to try to simulate ridge push (100 MPa) and proposed opposing forces of Alpine and Pyrenean compression (100 – 500 MPa). The results of uplift for given forces on the basin model are shown in Fig. 6.13.

Fig. 6.12. Schematic diagram of basin model constructed using SAVFEM. Red lines show the location of normal faults.

Fig. 6.13. Y-axis displacement or uplift (in metres) after the SAVFEM Basin Model was subjected to various forces. The red lines show the location of normal faults. Please note the different scales of y-axis displacement.

(a) 100 MPa against left boundary
(b) 100 MPa against both the left and right boundaries

(c) 100 MPa against the left boundary and 200 MPa against the right boundary.

(d) 100 MPa against the left boundary and 300 MPa against the right boundary.

(e) 100 MPa against the left boundary and 400 MPa against the right boundary.
100 MPa against the left boundary and 500 MPa against the right boundary.

Limitations in basin model:

1. The graben was constructed using normal faults which dip at 60° based on Andersonian theory (Anderson, 1951). However, the normal faults in the study area may be listric, dipping at less than 60°. The normal faults tapering abruptly at the boundary between the upper and lower crust may be unrealistic. In reality there may be a more gradual lowering of the gradient of the faults as they approach the base of the basement. In addition, the faults may be shallower, not reaching the base of the basement.

2. The base of the model is fixed. In some studies, however, it has been demonstrated that the lithosphere is also undergoing compression (Cloetingh et al., 1999) and should thus be allowed to move freely throughout its depth. This fixing was necessary to pin the model when forces were added to boundaries. Such a fix to the model’s base, however, should not significantly affect compression in the interested crustal rocks.

3. The model assumes homogeneous sediment in the basin. Heterogeneities in sediment type could facilitate flexural folding either by flexural-slip or by flexural flow. In flexural-slip folding, bucking is accommodated by layer parallel slip along contacts between layers. Flexural-flow folding is accommodated by shear within mechanically soft units between harder units (Davis and Reynolds, 1996).
Results and Interpretation

The basin models show relatively little uplift when the ridge push force (100 MPa) and additional assumed forces of Alpine and Pyrenean compression are added. In the study area, the amplitude of Lousy Bank, for example, is ~ 1.9 km. In the model, however, only 50-80 m of uplift is achieved at the centre of the basin even for large Alpine and Pyrenean forces of 500 MPa (Fig. 6.13f). This discrepancy could be the result of:

1. Flexural folding of sediment playing a vital role in producing the relatively large amplitudes of the compressional folds.
2. The coefficient of friction of the faults being lower to give greater sliding of the faults to produce greater uplifts.
3. The normal faults having a lower gradient than the proposed Andersonian 60°. The normal faults tapering abruptly at the boundary between the upper and lower crust may be unrealistic. In reality there may be a more gradual lowering of the gradient of the faults as they approach the base of the basement.

6.4.4 Sediment Folding

Basin modelling (section 6.4.3) has alluded to the significance of the folding of sediment in producing the observed amplitude of folds in the Rockall-Faroe area. In this study, isochore maps suggest that several folds—Mordor, Onika and Alpin anticlines—were formed in sediment basins. The Onika Anticline (see Fig. 4.47) and Mordor Anticline (see Fig. 4.30) are located within basins surrounded by basalt escarpments. The Alpin Anticline has relatively thick Eocene sediment bounded by the Sigmundur, Darwin and Rosemary Bank igneous centres (see Fig. 5.18). In addition to these folds there is also evidence to suggest that the Judd Anticline also formed within a basin. Thickening of Late Paleocene – Ypresian sediment towards the crest of the Judd Anticline (see Fig. 4.24) suggest that the fold developed over a former basin. The horizontal forces exerted on the Rockall-Faroe area were sufficient to result in the compression of sediments within these Cenozoic basins. Gravity modelling has suggested that the Hatton Bank, Lousy Bank, Bill Bailey’s Bank, Faroe Banks, Wyville-Thomson Ridge and North Ymir Ridge anticlines are
cored or underlain by pre-Cenozoic sediment (see section 4.6.2). Cenozoic sediments of the Alpin, Mordor, Onika, Viera and Judd anticlines have demonstrated that they are able to fold under the influence of external forces exerted on the Rockall Faroe area. It is believed in this study that pre-Cenozoic sediments in basins are also capable of folding with these external forces. The folding or shortening of pre-Cenozoic sediments can be assessed from a stress-strain experiment for limestone at a confining pressure of 103 MPa representing ~ 5 km depth. The experiment shows the strain produced in the limestone with increasing stress (Fig. 6.14). Stresses which are less than 300 MPa result in elastic deformation. Elastic deformation is recoverable and the rock returns to its original form when the stress is removed (Davis and Reynolds, 1996). Stresses greater than 300 MPa result in plastic deformation, which is non-recoverable. The compressional phases of folds in the study area demonstrate that the compressional forces were episodic and not continuous. Thus, in order for compressional structures to be permanent, sediments should have undergone plastic deformation. Hence, a force of at least 300 MPa would have been needed to ensure the permanence of these anticlines. This force, incidently, is of the same magnitude of the force which was postulated to result in the inversion of normal faults at similar depths (see section 6.4.2). This is further support for the significance of the contributions of Alpine and Pyrenean compression forces.
6.4.5 Basin Trends

In this study, there is clear evidence to show that the orientation of the Mordor, Onika and Alpin anticlines match the orientation of their underlying sediment basins (see Figs. 5.2 and 5.18). Likewise, it is proposed that the orientation of the Hatton Bank, Lousy Bank, Bill Bailey's Bank, Faroe Bank, Wyville-Thomson Ridge and Ymir Ridge anticlines follow the trends of their underlying pre-Cenozoic basins. The underlying basin morphology would have controlled the orientation of the anticlines during the inversion of the basins.

Fig. 6.14. Stress-strain diagram for limestone subjected to deformation under a confining pressure of 103 MPa (Davis and Reynolds, 1996).
It is proposed, in this study, that the triggering force for compression in the Rockall-Faroe area, in the late Ypresian, late Lutetian, C30 (Late Eocene) and Early Oligocene was hotspot-influenced ridge push from the Reykjanes Ridge. The SE-directing ridge push, coupled with Alpine and Pyrenean compression, could easily explain the inversion of NE-trending structures (Fig. 6.15) such as the Hatton Bank and Bridge anticlines. However, SE-directing ridge push may be insufficient to result in the inversion of NW-trending structures (Fig. 6.15), such as the Wyville-Thomson Ridge and Central Ymir Ridge anticlines. Lineaments which underlie NW-trending structures, in the Rockall-Faroe area, may be playing a role in their inversion.

Inversion along lineaments

It has been demonstrated in previous work that lineaments may play an important role in controlling the spatial distribution of folds. Several domes, for example, lay en echelon along the Jan Mayen Lineament extending into the margin from the Jan Mayen Fracture Zone (section 2.4). There is evidence from previous studies, through modelling, to suggest that the Wyville-Thomson Ridge, Ymir Ridge and Bill Bailey’s Bank are underlain by a complex of NW-trending lineaments (Kimbell et al., 2004; 2005). These lineaments could have been reactivated as part of a ramp
Fig. 6.16. Compressional structures in the north study area. Superimposed are the approximate positions of lineaments shown by black dashed lines (from Kimbell et al., 2005). Dextral movement of the lineament underlying the Wyville-Thomson Ridge results from a $\sigma_1$ horizontal force 30° from the lineament (Anderson, 1951). The $\sigma_1$ force is perpendicular to the Bridge Anticline.
anticline complex (Kimbell et al., 2005). Dextral movement of the NW-trending Wyville-Thomson Ridge lineament, consistent with the dextral offset between the Faroe-Shetland Channel and the NE Rockall Trough, can account for the orientation of the Bridge Anticline (Fig. 6.16).

The basins underlying the Wyville-Thomson Ridge and Ymir Ridge anticlines could have been formed as a result of strike-slip movement (Fig. 6.17). Pull-apart basins, formed by strike-slip movement, could become inverted by a change in the strike-slip direction. Thus, the Central Ymir Ridge and South Ymir Ridge anticlines could represent pop-up structures or inverted pull-apart basins along NW-trending strike-slip faults (Fig. 6.17). The position of the strike-slip faults in the NE Atlantic Margin could have been determined by tectonic grain in basement rock (Cartwright, 1992; cited in Doré et al., 1997a) and could have formed during rifting to accommodate displacement between adjacent rift segments (Morley et al., 1990; cited in Doré et al., 1997a). According to Doré et al. (1997a) these fractures (strike-slip faults) within basement were conveniently oriented to accommodate extension which led to NE Atlantic breakup.

![Diagram](image)

Fig. 6.17. Inversion of a pull-apart basin along a strike-slip system. Diagram based on Davis and Reynolds (1996).

The orientation of the horizontal stress (ridge push) required to form the Bridge Anticline, however, is oblique to the Wyville-Thomson Ridge and Ymir Ridge anticlines (Fig. 6.16). This oblique ridge push, together with Alpine and Pyrenean forces, could have been sufficient to result in the inversion of NW-trending grabens, not associated with strike-slip movement.
6.4.6 C30 (Late Eocene) sagging

The Rockall Trough (Fig. 6.1) is a basin which formed by rifting in the Early Cretaceous (Musgrove and Mitchener, 1996). Sharp subsidence of the Rockall Trough in C30 (Late Eocene) time has been described by previous authors (Stoker, 1997; Stoker et al., 2001; Praeg et al., 2005; McNally et al., 2006), and has been reinforced in this study by analysis of well data in the NE Rockall Trough (see Fig. 5.31). Praeg et al. (2005) proposed that this subsidence was the result of a loss of dynamic support by convection in the underlying mantle. However, the C30 (Late Eocene) unconformity also marks the timing of a major compression across the Rockall-Faroe study area. The onset of this subsidence at the time of compression suggests that these events may be related.

Kooi et al. (1991) has postulated flexural downwarping as the cause of accelerated subsidence in the Late Pliocene in the southern North Sea basin. Flexural downwarping of the lithosphere can result from compression (Fig. 6.18). Lithosphere of low flexural rigidity can promote the effectiveness of compressional stresses in causing substantial downwarp (Kooi et al., 1991). The Rockall Trough is characterized by relatively weak lithosphere with an elastic thickness of approximately 6 km on the eastern margin near the Porcupine Bank (Daly et al., 2004). This is smaller than the 25 km elastic thickness of continental lithosphere (Fowler, 1990). Compression-induced downwarping occurs as a result of lithospheric folding due to compression coupled to the loading effects by sediment (Kooi et al., 1991). The Rockall Trough having a relatively low elastic thickness could have been prone to compression occurring at C30 (Late Eocene). The subsidence that occurred as a result of compression was significant to increase sedimentation by:

1. Deepening to allow the influx of bottom-water currents to facilitate the deposition of contourites.
2. Increasing the accommodation space for the accumulation of sediment.

This increase in sedimentation could have further increased the subsidence instigated by compression.
6.4.7 Lithospheric control on folding and basin inversion

The C30 sagging event has alluded to the possible role of the structure of the lithosphere in controlling compression. Previous studies have shown, through modelling, that lithospheric folding is a primary response to compression (Cloetingh et al., 1999). In addition, there is evidence to suggest that the age of the lithosphere controls the wavelength of folding (Cloetingh et al., 1999). The thickness of the lithosphere may also affect the location of compressional features. In the Norwegian Margin, for example, compressional features are present within the Vøring Basin, but are absent in the adjacent Møre Basin (Fernandez et al., 2005; Doré et al., 2008). The thickness of the lithosphere beneath the Vøring and Møre Basins is 115 km and 125 km respectively, resulting in folding in the less competent Vøring Basin (Fernandez et al., 2005). The structure of the lithosphere could also play a role in controlling the distribution of folds in the Rockall-Faroe area. However, the structure of the lithosphere in the study area has not been examined in this study.

6.5 Compression in areas proximal to the NE Atlantic Margin

The forces required for the inversion of deep normal faults and the folding of pre-Cenozoic sediment suggest that hotspot-influenced ridge push force would have most likely acted together with Alpine and Pyrenean forces to result in the formation of compressional structures present in the Rockall Faroe area. According to Lundin and
Doré (2000), Alpine stresses could not have been transmitted across the spreading axis to affect the NE Greenland margin. Whilst compressional deformation is well documented in the NE Atlantic Margin, comparatively little is known regarding compression in the East Greenland margin (Lundin and Doré, 2002). Compression also appears less evident in the East Greenland Margin compared with compressional deformation in the NE Atlantic Margin. "It appears that the NE Greenland margin exhibits smaller-scale compressional deformation in comparison with the compressional domes on the Norwegian margin, which are greater in length, width, and amplitude" (Tsikalas et al., 2005). The lack of stresses from Alpine and Pyrenean compression and the postulated negligible ridge push (Fig. 6.9) is consistent with the lack of observed compressional structures.

The Traill Ø region on the East Greenland margin (Fig. 6.1) experienced compression during the Tertiary (post-54 Ma) based on the folding of sills c. 54 Ma in age (Price et al., 1997). Price et al. (1997) inferred a Late Miocene age for this compression based on compressional structures of this age in the Vøring Basin (Blystad et al., 1995) and in the Rockall-Faroe area (Boldreel and Andersen, 1993). The Late Miocene compression in the Rockall-Faroe area was attributed to Alpine compression (Boldreel and Andersen, 1993). However, if Alpine stresses cannot be transmitted to the East Greenland Margin, then the compression may not be Late Miocene in age. If hotspot-influenced ridge push and Alpine/Pyrenean compression cannot account for the compression on the East Greenland margin, compression may be the result of depth-dependent stretching. An Early Oligocene unconformity drilled on the Jan Mayen microcontinent is generally accepted as representing the onset of seafloor spreading along the Kolbeinsey Ridge (Taiwani and Udintsev, 1976; cited in Lundin and Doré, 2002). Depth-dependent stretching associated with this opening, which post-dates 54 Ma, could have resulted in compression in the Traill Ø region, located in the required 75 - 150 km marginal zone from the continent-ocean boundary (see section 6.2.4). In addition to the Traill Ø region, other areas relatively proximal to the NE Atlantic Margin have also undergone inversion (Table 6.3).
Table 6.3. Inversion proximal to the NE Atlantic Margin. For location of areas see Fig. 6.1.

Holford et al. (in press) also recorded, using apatite-fission track analysis, three Cenozoic exhumation episodes in the Irish Sea Basin (Fig. 6.1) – Early Cenozoic (65–55 Ma), Mid Cenozoic (40–25 Ma) and Late Cenozoic (20–15 Ma). These events correlate with major tectonic unconformities present in the sedimentary succession of the NE Atlantic Margin (Holford et al., in press). Indeed, the timings of the Early Cenozoic and Mid Cenozoic events correlate with the Thanetian, the C30 (Late Eocene), and the Early Oligocene timings of compressional events in the Rockall-Faroe study area. According to Holford et al. (in press) this suggests the same compressional mechanisms for the two areas despite the large distances separating them. However, the late Ypresian and late Lutetian compressional events in the Rockall-Faroe area are not represented by the uplift in the Irish Sea Basin. In addition, the Early and Mid Cenozoic uplift episodes of the Irish Sea Basin correlate well with the timings of Alpine compression (Ziegler, 1988). Furthermore, the northern North Sea is devoid of compressive deformation (Pascal & Gabrielsen, 2001; cited in Doré et al., 2008). This area lies more distally from the Alpine front compared to the Irish Sea Basin. Such observations suggest that the uplift in the Irish Basin and Cenozoic inversion of proximal areas to the Alpine front (Table 6.3), are the result of Alpine/Pyrenean compression. Compression in the Rockall-Faroe study area, however, is believed to be the result of hotspot-influenced ridge push in addition to Alpine/Pyrenean compression.
6.6 Bottom-water current activity
C30 sagging resulted in an increase in bottom-water current activity in the Rockall-Faroe area (see section 5.4.2). Whilst the C30 unconformity has been attributed to compression, the early Late Oligocene, C20 (late Early Miocene), Late Miocene-Early Pliocene and C10 (late Early Pliocene) unconformities, due to their onlap and erosional nature (see section 5.4.4), have been attributed to local and regional changes in bottom-water current activity.

**Early Late Oligocene**
The Late Oligocene was marked by a major increase in global sea level (Haq et al., 1987). Onlap of marine shale onto Early Oligocene sand in the NE Rockall Trough is interpreted, in this study, as representing this period of transgression (section 5.5). A rise in sea level would have increased the depth of the water column to result in an intensification of bottom-current water current flow.

**C20 (late Early Miocene)**
The C20 (late Early Miocene) unconformity is interpreted to coincide with the formation of the Iceland topographic high, based on the oldest rocks recorded on Iceland (Fougler, 2006). The Iceland high could have acted as a topographic load resulting in flexing of the crust (Fig. 6.19). This could have resulted in the submergence of proximal areas facilitating bottom-water current flow over the Wyville-Thomson Ridge and through the Auðuhuöl Basin Syncline (section 5.4.5). This concept is consistent with the subsidence predicted for the Greenland-Scotland Ridge in the Early Miocene and Mid Miocene by Vogt (1972) and Schnitker (1980) respectively, and the Miocene initiation of the Norwegian Sea Overflow (Stoker, 2005b).

It has been previously proposed that the rich layer of smectite defining the C20 unconformity, in the South Rockall Trough, was derived from the diagenesis of volcanic ash (Dolan, 1986). The development of the Iceland topographic high could have also been the source of volcanic material. This volcanic material could have been transported by bottom-water currents and deposited in the South Rockall Trough.
Late Miocene – Early Pliocene

The Late Miocene - Early Pliocene marks the end of the Messinian Crisis. The Messinian Crisis occurred at 6.2 Ma (Blanc and Duplessy, 1982) and ended after a duration of less than 2 Ma (Butler et al., 1999). During the Messinian Crisis the Mediterranean Sea evaporated. This reduced the salinity of the intermediate and surface waters of the North Atlantic and prevented the formation of dense waters in the Norwegian Greenland Sea (Blanc and Duplessy, 1982). The end of the Messinian Crisis would have reintroduced saline water into the North Atlantic increasing the strength of the North Atlantic Deep Water and consequently an intensification of the Norwegian Sea Overflow. This could have resulted in Late Miocene-Early Pliocene unconformity formation in the Rockall Trough.

C10 (early Late Pliocene)

The C10 (early Late Pliocene) unconformity coincided with the closure of the Panama Isthmus at 4.2 - 3.5 Ma (Keigwin, 1982; Coates et al., 1992; Droxler et al., 1998). This resulted in the intensification of the Gulf Stream, strengthening deep water formation in the Labrador Sea (Haug and Tiedemann, 1998). This could have resulted in an increase in bottom-water current activity resulting in unconformity formation within the Rockall Trough at this time.

Fig. 6.19. Elastic plate model based on the Vening Meinesz isostatic model. A topographic load bends the elastic crust downward into the fluid mantle, which is pushed aside, resulting in subsidence. Diagram based on Lowrie (1997) and Watts (2001).
6.7 Summary: Evolution of the Rockall-Faroe Area

Unconformities within the Rockall-Faroe area can be linked to regional events affecting the NE Atlantic Margin (Fig. 6.8). These events have resulted in the formation of compressional structures and have consequently influenced the action of bottom-water currents within the Rockall-Faroe area.

In the Thanetian (Fig. 6.20) it is proposed that depth-dependent stretching and asthenospheric upwelling prior to the seafloor spreading resulted in the formation of compressional structures at this time. Within the marginal zone (75 - 150 km from the continent-ocean boundary), depth-dependent stretching resulted in the initial formation of the Hatton Bank, Lousy Bank, Bill Bailey’s Bank, South Faroe Bank, and North Faroe Bank anticlines, in addition to the Faroe Bank Channel Syncline. The more distal areas from the continent-ocean boundary experienced compression due to uplift induced by asthenospheric upwelling. This resulted in the formation of the Wyville-Thomson Ridge, the North Ymir Ridge and the Judd anticlines, and the inverted south West Lewis Basin. The effect of these compressional forces was enhanced by the Alpine and Pyrenean forces which also took place at this time (Fig. 6.8).

In the late Ypresian (Fig. 6.21), hotspot-influenced ridge push, coupled with Pyrenean compression resulted in the further growth of folds formed during the Thanetian. In addition, new anticlines – Dawn and Mordor anticlines – formed at this time. Sediments within a basin bounded by basalt escarpments were compressed to form the Mordor Anticline.

Although no V-shaped ridges along the Reykjanes Ridge are late Lutetian in age, late Lutetian compression, in this study, is postulated as also being the result of hotspot-influence ridge push. This force in addition to Pyrenean compression resulted in the growth of existing folds and the formation of new folds – the Central Ymir Ridge, South Ymir Ridge, Bridge, Ness and Onika anticlines (Fig. 6.22). The Onika Anticline was formed by the compression of sediments within a basin bounded by basalt escarpments.
Fig. 6.20. Thanetian development of folds in the Rockall-Faroe area. The formation of anticlines flanking the edge of the continental margin - Lousy Bank, Bill Bailey’s Bank and Faroe Bank anticlines - are inferred to be the result of depth-dependent stretching. Inversion structures located more distally from the continent-ocean boundary, such as the Wyville-Thomson Ridge and Judd anticlines and the inverted south West Lewis basin, are inferred to be the result of compression on the fringes of upwelling asthenosphere.
Fig. 6.21. late Ypresian development of folds in the Rockall-Faroe area. The Dawn and Mordor anticlines were formed at this time. Late Ypresian compression has been ascribed to the combined forces of hotspot-influenced ridge push, Alpine and Pyrenean compression.
Fig. 6.22. late Lutetian development of folds in the Rockall-Faroe area. The Central Ymir Ridge, South Ymir Ridge, Bridge and Onika anticlines initially formed at this time. It is proposed, in this study, that late Lutetian compression was the result of hotspot-influenced ridge push and Pyrenean compression.
Fig. 6.23. C30 (Late Eocene) development of folds in the Rockall-Faroe area. The Alpin and Viera anticlines were formed at this time. The north West Lewis Basin fault was also reactivated at this time. The Rockall Trough experienced abrupt subsidence as a result of compression-induced downwarping (see text). The C30 compression coincides with the timing of hotspot-influenced ridge push, Alpine and Pyrenean compression.
Fig. 6.24. Early Oligocene development of folds in the Rockall-Faroe area. Compressional growth of existing structures (except the West Lewis Basin) occurred in the Early Oligocene. The Early Oligocene compression is attributed to hotspot-influenced ridge push, Alpine and Pyrenean compression. C30 (Late Eocene) subsidence of the Rockall Trough facilitated Early Oligocene bottom-water current activity (green arrows) which resulted in the deposition of upslope migrating contourites in the Early Oligocene (see Figs. 5.29 and 5.30). Black colour of structures denotes compressional inactivity.
Fig. 6.25. Early Late Oligocene inferred bottom-water current activity (green arrows) in the Rockall-Faroe area. Onlap of Late Oligocene shale in the NE Rockall Trough and onlap of an inferred contourite drift north of the Alpin Anticline (see section 5.4.4) support bottom-water current flow in the area shown in the Rockall Trough. The distribution of bottom-water currents in the Iceland Basin is based on Stoker et al. (2005b). Black colour of structures denotes compressional inactivity.
Fig. 6.26. Miocene - present-day bottom-water current activity (green arrows) in the Rockall-Faroe area (Stoker et al., 2005b). The Auðhumla Basin Syncline has controlled the direction of the Norwegian Sea Overflow (NSO). In the Rockall-Faroe area, the C20 (late Early Miocene) unconformity marks the initiation of the NSO due to the submergence of the Greenland-Scotland Ridge. Intensification of the NSO by regional changes in bottom-water current circulation formed the Late Miocene - Early Pliocene and C10 (late Early Pliocene) unconformities in the study area. Inactive structures are shown in black.
The C30 (Late Eocene) compressional event is inferred in this study as the most significant compressional event occurring within the Cenozoic in the Rockall-Faroe area. This is based on the extensive C30 sagging (subsidence) observed within the Rockall Trough inferred as being the result of flexural downwarping due to compression. The C30 event was the result of hotspot-influenced ridge push, and Alpine and Pyrenean compression. New compressional features – the Alpin and the Viera anticlines, and the reverse faults on the south limb of Hatton Bank and in the north West Lewis Basin – developed at this time (Fig. 6.23). The Alpin Anticline was formed by the compression of Eocene sediment within a basin bounded by igneous centres. All older folds were active during the C30 episode of compression and no new folds were formed after this time.

In the Early Oligocene, compression also resulted from hotspot-influenced ridge push, and Alpine and Pyrenean forces. The Early Oligocene compression resulted in the further growth of older compressional structures present in the Rockall-Faroe area with the exception of the West Lewis Basin (Fig. 6.24).

The Hatton Bank, Lousy Bank, South Faroe Bank, North Faroe Bank, Wyville-Thomson Ridge and Ymir Ridge anticlines were probably formed by the compression of pre-Cenozoic basins. These basins exploited the existing Caledonide and Lewisian trends upon their formation resulting in the NE-SW and NW-SE orientation of these structures. The basins underlying the Wyville-Thomson Ridge and Ymir Ridge anticlines could have been formed by strike-slip movement of the lineaments which they overlie. The orientation of pre-Cenozoic and Cenozoic basins underlying folds would have controlled the orientation of compressional structures in the Rockall-Faroe area.

Late Eocene subsidence of the Rockall Trough resulted in an increase in bottom-water activity. In the early Late Oligocene a rise in sea level resulted in a further change in bottom-water currents to result in the formation of an unconformity. The unconformity could have been due to an increase in the circulation of bottom-water currents in the North and NE Rockall Trough (Fig. 6.25). In the late Early Miocene
subsidence of the Wyville-Thomson Ridge Anticline and the Auðhumla Basin Syncline resulted in the initiation of the Norwegian Sea Overflow (NSO) (Fig. 6.26) and the formation of the C20 unconformity. Intensification of the NSO in the Late Miocene – Early Pliocene and C10 (late Early Pliocene) to produce erosional unconformities was the result of regional changes in the intensity of bottom-water current circulation.
7.0 Conclusions

Continental margins are formed from the extension of continental crust and undergo thermal subsidence after rifting has taken place (McKenzie, 1978). Many continental margins, however, deviate from the once held idea of simple tectonic quiescence during this post-rift thermal subsidence. Indeed continental margins around the world show evidence of post-rift inversion (Whitjack et al., 1998; Karner and Driscoll, 1999; Cobbold et al., 2001; Hudec and Jackson, 2002). The NE Atlantic Margin represents a classic example of inversion within continental margins. Compressional structures have been previously observed on the Norwegian Margin (Blystad et al., 1995; Lundin and Doré, 2002), the Faroe-Shetland Basin (Boldreel and Andersen, 1993; Ritchie et al., 2003) and the Rockall-Faroe area (Boldreel and Andersen, 1993; Johnson et al., 2005; Ritchie et al., 2008).

Previous studies within the Rockall-Faroe area, however, were limited in their understanding of compressional features as a result of a lack of well and borehole data, high resolution and deep penetrating seismic data, and gravity data which were available in this study. The well and borehole data allowed the dating of critical stratigraphic surfaces crucial in linking both tectonic and non-tectonic events to regional events which have affected the NE Atlantic Margin. A more extensive and better quality seismic data set not only allowed more compressional structures to be mapped but was useful in refining the axial traces of previously studied folds and reverse faults. In addition, stratigraphic surfaces were mapped over larger areas because of the availability of extensive seismic data. Gravity data was crucial in determining the underlying nature of some compressional structures, not revealed by seismic data.

Unconformities within the Rockall-Faroe area, defined by onlap and reflector terminations (erosional truncation), served as important markers for compressional events and changes in bottom-water current current activity. Well and borehole data in the NE Rockall Trough and the Hatton-Rockall Basin allowed unconformities within these areas to be dated with a relatively high degree of certainty. In the North
Rockall Trough unconformities were traced from the NE Rockall Trough and constrained using borehole data on the eastern margin of the Rockall Bank. Unconformities within the Faroe Bank Channel were constrained using information from previous work (Smallwood, 2004) and unconformities mapped within the Auðhumla Basin from the NE Rockall Trough. The presence of high quality seismic data within the Faroe Bank Channel was an asset in the mapping of these unconformities.

The compressional structures within the Rockall-Faroe area vary in size and in orientation. Two main orientations of folds are apparent – NE and NW trends. The folds have amplitudes of 0.4 – 2.2 km and are of relatively long wavelengths. Unconformities associated with folding are defined by onlap on the limbs of the fold. These are the Thanetian, late Ypresian, late Lutetian, C30 (Late Eocene) and Early Oligocene unconformities inferred to represent compressional events which took place within the Rockall-Faroe area. Isochrons of mapped unconformities and the top-basalt surface were used to construct isochores which allowed the distribution of sediments to be determined. This gave more insight into the structural evolution of the area. The ages of sediment wedges, for example, were used as additional constraints for the timing of growth of the Wyville-Thomson Ridge, North Ymir Ridge, Central Ymir Ridge and Bridge anticlines.

There is evidence to suggest that the orientation of anticlines was controlled by underlying pre-existing structures. Gravity modelling has revealed the presence of relatively low-density pre-Cenozoic sediment in the Lousy Bank, Bill Bailey’s Bank, Faroe Bank, Wyville-Thomson Ridge and North Ymir Ridge anticlines that could represent sedimentary basin sediments. Seismic data and isochore maps suggest that the Alpin, Mordor, Onika and Judd anticlines where formed by the compression of Early Cenozoic basin sediment. The Alpin, Mordor and Onika anticlines follow the trends of these basins suggesting that their orientation was controlled by the underlying basin morphology. The trends of anticlines, such as the Lousy Bank, could be representative of the orientation of underlying pre-Cenozoic sedimentary
basins. The growth of the Wyville-Thomson Ridge and Ymir Ridge anticlines may have also been controlled by underlying NW-trending lineaments.

Previous studies have attributed compressional structures within the NE Atlantic Margin to hotspot-influenced ridge push, Alpine and Pyrenean compression and the Iceland Insular Margin body force. The late Ypresian, C30 (Late Eocene) and Early Oligocene compressional events in the Rockall-Faroe area correlate with the timings of hotspot-influenced ridge push. However, this study demonstrates using Mohr-Coulomb diagrams that hotspot-influenced ridge push is insufficient to result in the inversion of any deep faults bounding the proposed sedimentary basin of the Lousy Bank. This coupled with less inversion on the East Greenland Margin suggests that the Alpine and the Pyrenean compression events occurring at these times were also contributing forces for inversion to occur. Hotspot-influenced ridge push, however, cannot account for the Thanetian compressional event occurring prior to seafloor spreading. The study proposes that this event was the result of depth-dependent stretching and associated asthenospheric upwelling. The Iceland Insular Margin body force can account for the previously studied Mid-Miocene compression in the Faroe-Shetland Basin areas (Ritchie et al., 2003). The apparent lack of Mid-Miocene compression in the Rockall-Faroe area would suggest that the gravity wedging effect due to the extra height of the Iceland Insular Margin was more pronounced against the adjacent Faroe Shelf than the more distal Rockall-Faroe area.

The erosional nature of the early Late Oligocene, C20 (late Early Miocene), Late Miocene – Early Pliocene and C10 (late Early Pliocene) unconformities is attributed to bottom-water current activity. C30 (Late Eocene) subsidence of the Rockall Trough resulted in the deposition of upslope migrating contourites in the Early Oligocene. Sea-level rise in the early Late Oligocene resulted in the extension of bottom-water current activity into the NE Rockall Trough and north of the Alpin Anticline. This resulted in onlap which defines the early Late Oligocene unconformity. Early-Mid Miocene formation of the Iceland Insular Margin may have resulted in the loading and the consequent submergence of the Greenland-Scotland Ridge (Iceland-Faroe Rise). This resulted in the initiation of Norwegian
Sea Overflow (NSO) from the Faroe Bank Channel into the North Rockall Trough and is marked by the C20 unconformity in the Rockall-Faroe area. An intensification of the NSO by the cessation of the Messinian Crisis could have resulted in the formation of the Late Miocene – Early Miocene unconformity defined by erosional truncation in the Rockall Trough. Closure of the Panama Isthmus in the late Early Pliocene would have intensified the Gulf Stream which further amplified North Atlantic Deep Water formation. This could have resulted in an increase in activity of the NSO to form the C10 unconformity in the Rockall-Faroe area.

The distribution of sediment and unconformities within the Rockall-Faroe area is mainly the result of compression and bottom-water current activity. Eustatic sea-level changes, however, can account for the presence of the early Late Oligocene unconformity and Early Miocene and Late Pliocene – Recent fan systems in the NE Rockall Trough. The preservation of transgressive and lowstand systems tracts in periods of tectonic quiescence could have been hindered by the erosive actions of bottom-water currents.

Continental margins have provided an ideal setting for petroleum plays. Shale deposited in anoxic conditions during young ocean basin development has the potential to become an ideal source for hydrocarbons (Boillot, 1981). As the ocean basin deepens, turbidite sands intercalated with hemipelagic sediments are deposited and can form reservoirs and seals (Boillot, 1981). Normal faults formed as a result of rifting can act as conduits for migrating hydrocarbons. The presence of anticlines within continental margins completes the play setting by forming ideal structural traps of four-way dip closure (Doré and Lundin, 1996; Doré et al., 1997b). Gas accumulation in the Ormen Lange Dome in the Vøring Basin, is located within such a structural trap (Lundin and Doré, 2002; Smith and Møller, 2003). The timing of growth of compressional structures is critical in determining their presence at the time of hydrocarbon migration.

The Rockall-Faroe area has demonstrated that the NE Atlantic Margin has undergone deformation to form compressional structures and is thus anything but passive. The
study area located proximal to the Iceland plume and between ridge push and Alpine and Pyrenean orogenies, has provided an ideal setting for the study of potential compressional mechanisms. Seismic interpretation, calibration of stratigraphic horizons and gravity modelling have been used to determine the timing and nature of compressional structures. This allowed the link between the formation of these structures and regional mechanisms affecting the margin to be established. The Rockall-Faroe area could thus shed more light in the understanding of inversion tectonics in the NE Atlantic Margin as a whole. The deformation mechanisms proposed for the Rockall-Faroe area, such as ridge-push, hinterland orogenic forces and depth-dependent stretching may also play a role in the inversion of basins in other continental margins. These include continental margins of East USA, Brazil, Angola and NW Australia.
Appendix

A. Seismic Acquisition and Display

In addition to interpreting existing seismic data, the study involved the acquisition of new seismic data (see Fig. 1.2). A seismic survey was conducted on May 20\textsuperscript{th} – June 9\textsuperscript{th}, 2006 aboard the RRS Charles Darwin (Fig. A1). This was done in order to acquire seismic data over areas which lacked seismic images, such as the Lousy, Bill Bailey’s and Faroe Banks (see Fig. 1.2).

![Fig. A1. RRS Charles Darwin. Seismic acquisition took place aboard this ship.](image)

The British Geological Survey funded the seismic acquisition cruise and provided, in addition to trained personnel, the following equipment:

- Airgun
- Sparker
- Hydrophones
- Magnetometer
- Gravity Meter
- SWATH

The scientific team (Fig. A2) was headed by David Smith (Senior Engineer) of the British Geological Survey (BGS). The team included geologists Heather Stewart (BGS) and Adrian Tuitt, and engineers Michael Wilson, Dave Wallis, Dave Baxter
and Iain Pheasant of the British Geological Survey. In addition, Miriam Sayago-Gil was an invited PhD student from Instituto Español de Oceanografía who studied geomorphologic features on the Hatton Bank.

Fig. A2. Scientific crew on the RRS Charles Darwin (St. Kilda in the background). From left to right: Dave Baxter, Adrian Tuitt, Dave Wallis, Iain Pheasant, Heather Stewart, Dave Smith, Michael Wilson and Miriam Sayago-Gil.

Seismic acquisition involves the use of a source, receiver and recorder (Fig. A3). Airgun (Fig. A4) and sparker (Fig. A5) sources were used to emit acoustic waves. These acoustic waves, after being emitted from the sources, were reflected by underlying sediment layers returning to the hydrophone receivers (Fig. A6) at the sea surface. The characteristics of the sources and receivers are shown in Table A1.

Fig. A3. Schematic diagram of seismic survey. Sound waves emitted from sparker and airgun sources are reflected off rock boundaries and then received by hydrophones. The recording system on the ship records the two-way times of sound waves. This is the time of emission from the source to the hydrophones after reflection off rock boundaries.
Fig. A4. Airgun used for seismic acquisition as a source for acoustic signals.

Fig. A5. Sparkers used for seismic acquisition as a source for acoustic signals.

Fig. A6. Hydrophones used for seismic acquisition as receivers for acoustic signals. The acoustic signals sourced from the airgun and the sparker sources were received by different hydrophones.
Table A1. Characteristics of sources and their respective hydrophones.

<table>
<thead>
<tr>
<th>Source</th>
<th>Frequency (Hz)</th>
<th>Hydrophone Length (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Airgun</td>
<td>30-250</td>
<td>30</td>
</tr>
<tr>
<td>Sparker</td>
<td>2000</td>
<td>10</td>
</tr>
</tbody>
</table>

The airgun and sparker were towed ~10 m behind the ship at ~10 m apart and at a depth of ¼ wavelength of their respective produced acoustic signals. The airgun was shot and its hydrophones recorded for 4 seconds after the shot. When the airgun hydrophones stopped recording at 4 seconds, the sparker was shot and its hydrophones recorded for 2 seconds, after which time the airgun was shot again. This ensured that the sparker hydrophones did not receive original airgun signals and vice versa. However, the sparker hydrophones infrequently recorded airgun multiples. The sparker with its higher frequency resolved sediment layers at shallower depths compared to the airgun whose lower frequency was able to penetrate deeper sediment layers. The integration of both sparker and airgun signals facilitated a better resolution of both shallow and deeper sediments.

The seismic recording system used was BGS CODA DA200 (Fig. A7). The quality of the seismic data was enhanced using Time Varying Grain (TVG) and Time Varying Filter (TVF). TVG compensates for the loss of energy (attenuation of signal) with time/depth. The signals from deeper layers are essentially normalized to give the same strength and clarity as the seabed. TVF sets the limits of the frequency recorded to reduce the amount of noise and for the reception of relevant signals. Seismic images were produced on sheets for interpretation whilst on the ship. In addition, the recordings were saved as files by CODA and this was loaded into Landmark at the British Geological Survey.
Landmark is the software which was used for the digital display of seismic images. This tool allowed horizons and structures to be mapped. In addition, the input of well data allowed unconformities to be dated. Isochron maps were also built using Landmark.

Arc GIS (Global Information System) was used to display isochron maps and to view structures (such as the axis of folds) imported from Landmark. Isochore (time thickness) maps have also been produced using GIS. Maps from GIS were imported to CorelDraw for editing and final display. Seismic images were also imported directly to Corel Draw from Landmark for editing and final display.
B. Burial Curve Construction

Burial curve construction from well data 164/25-2 in the NE Rockall Trough is based on a de-compaction method from Watts (2001). The method calculates the thickness of de-compacted and compacted sediment at the time of deposition of a sediment interval. It is this thickness sediment which is used as a proxy for tectonic subsidence assuming subsidence results in greater accommodation space for the accumulation of sediment.

Method for de-compaction of sediment

\[ S^* = S \frac{(1-\bar{\varnothing})}{(1-\bar{\varnothing}^*)} \]

- \( S^* \) = thickness of de-compacted layer of sediment
- \( S \) = thickness of compacted layer of sediment at a particular time
- \( \bar{\varnothing}^* \) = porosity of de-compacted layer of sediment
- \( \bar{\varnothing} \) = porosity of compacted layer of sediment at a particular time

The equation was used to calculate the thickness of compacted and de-compacted sediment for different time intervals (Fig. B1). The porosity of sediment was derived from Bond and Kominz (1984) and is based on the rock type and the depth (Table B1). The rock type at each time interval is from well data while the depth of the sediment is based on the thickness of the overlying sediment after de-compaction and compaction calculations are made on the overlying sediment layers.
Fig. B1. Thickness of compacted (C) and de-compacted (D) sediment layers from well 164/25-2. Underlying sediment layers were compacted for every time interval. The Ypresian sediment layer, for example, was de-compacted in the late Ypresian and compacted six times (C<sub>1</sub>-C<sub>6</sub>) during the later six time intervals. The overall subsidence is shown by the red line.
<table>
<thead>
<tr>
<th>Age of sediment</th>
<th>Sediment Type</th>
<th>Present -day</th>
<th>late Ypresian</th>
<th>late Lutetian</th>
<th>late Late Eocene</th>
<th>late Early Oligocene</th>
<th>late Late Oligocene</th>
<th>late Early Miocene</th>
<th>late Early Pliocene</th>
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<td>shale (7%)</td>
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<tr>
<td>Ypresian</td>
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<tr>
<td></td>
<td>tuff (23%),</td>
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<td>sand (12%)</td>
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</table>

Table B1. Porosities used in calculating the thickness of compacted and de-compacted sediment in well 164/25-2 derived from Bond and Kominz (1984). The depths of sediment layers were measured as the thickness of overlying layers.
References


Roberts, D.G., Thomson, M., Mitchener, B., Hossack, J., Carmichael, S., and Bjrnseth, 1999. Palaeozoic to tertiary rift and basin dynamics: mid-Norway to the Bay of Biscay a new context for hydrocarbon prospectivity in the deep water


