

Sub-basalt hydrocarbon prospectivity in the Faroes – Shetlands and Møre basins, Northeast Atlantic

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Abstract

The seismic imaging below the basalts in the NE Atlantic Basins is generally poor and the thickness of the basalts is difficult to predict. Two recent wells (William, 6005/13-1 and Brugdan, 6104/21-1) drilled in Faroes waters were suspended before reaching the main sub-basalt reservoir targets, because the basalts were thicker than expected. New deep-tow (18m) seismic reflection data has now allowed more detailed imaging of the sub-basalt Mesozoic strata. The base of the basalt is not always reflective and is probably a transitional contact, with low acoustic impedance. The underlying Mesozoic strata produce coherent reflectivity. Large structural closures produced by Cretaceous rifting, and Cenozoic folding have been imaged in the Rockall and Faroes-Shetland basins and these may contain Cretaceous and Palaeocene clastic reservoirs at drillable depths of about 4 km.

Introduction

Basalts of Palaeogene age extend over much of the Faroes-Shetlands and Møre basins and reach up to approximately 6 km thickness around the Faroe Islands (White *et al.*, 2005). The existing seismic data quality is generally very poor below the basalt. Promising oil and gas field discoveries are present in the shallower eastern portions of the Faroes – Shetlands, but no wells drilled through the thick basalts have encountered Cretaceous reservoir targets. The new deep-tow seismic data presented here have imaged the deeper Mesozoic strata, which enhances new play possibilities for oil exploration across this area.

Seismic Data Acquisition and Processing

Various seismic acquisition techniques have been used to try to image below the volcanic rocks in the Faroes-Shetlands area: over-under technique; ocean Bottom Seismometer surveys; wide-angle reflection and seismic refraction (e.g. Christie and White, 2008; England *et al.* 2005; Gallagher and Dromgoole, 2007 and 2008; Hobbs and Jakubowicz, 2000; White *et al.*, 1999, 2003, and 2005). However, most of these surveys have been limited to a few lines due to the high cost of acquisition.

New seismic data presented in this paper was

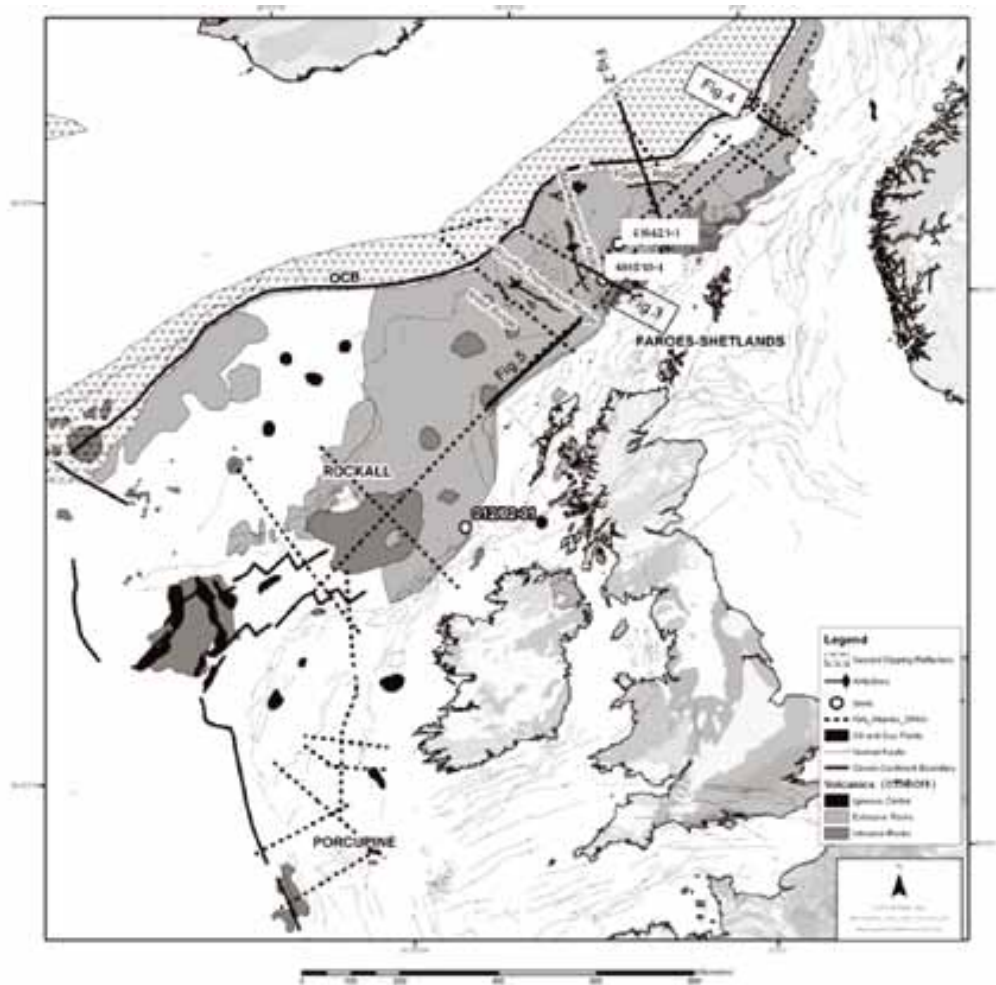


Fig. 1.

Map of Faeroes and Rockall Basins showing location of seismic lines used in this study as well as the distribution of the Paleogene volcanics. The locations of key wells mentioned in the text are also shown.

shot and processed by ION/GXT in 2008 (see Fig. 1 for line locations). The seismic line acquisition and Pre-Stack Depth Migration (PSDM) processing are all designed to provide optimum seismic imaging below thick (>1 km) Paleogene basalts, which cover much of the NE Atlantic margin (Fig. 1). This was achieved with:

- a) long receiver offset (10.2 km);
- b) large source size (7440 cu. in.) with 173 barm peak output measure through system filters;
- c) deep-towed airgun (17.5m);mer (18m); and e) minimum bubble interference.

The air gun and streamer deep tow provided a maximum energy input at 20-30 Hz frequency.

This low frequency energy penetrates through the basalts more effectively.

Another significant data quality benefit of deep tow is the reduced noise caused by near surface current and wave action.

This acquisition produced more coherent reflections below the basalts which have a high seismic attenuation factor. Data was recorded to 18 seconds TWT, to image the complete crustal structure (Fig. 2). All the PSDM seismic data was migrated using velocities derived from iterative tomographic velocity modelling. The velocity-modelled depths are within 5% of the depths from several calibration wells in the Faroes-Shetland Basin.

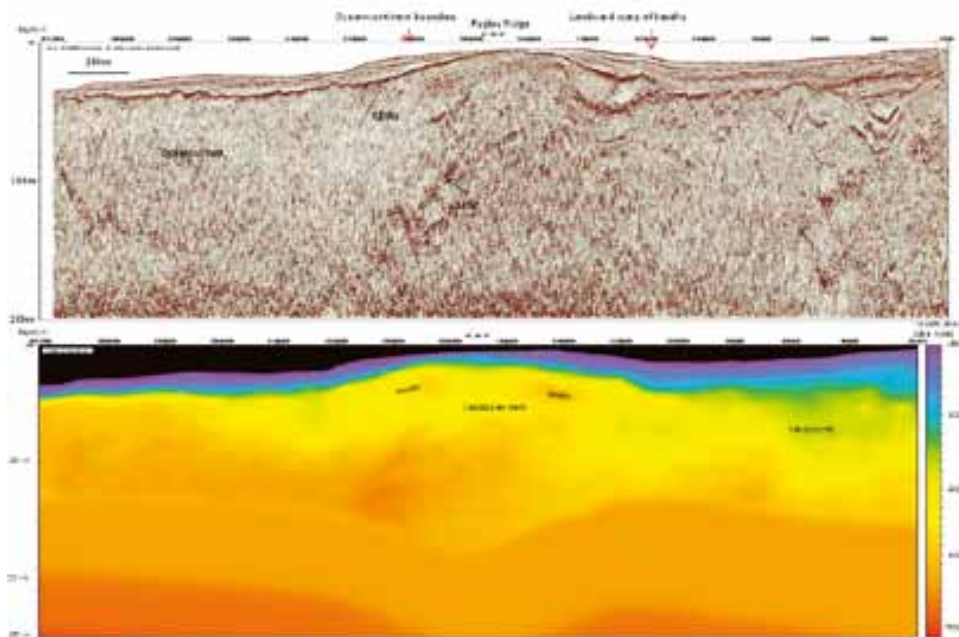


Fig. 2.

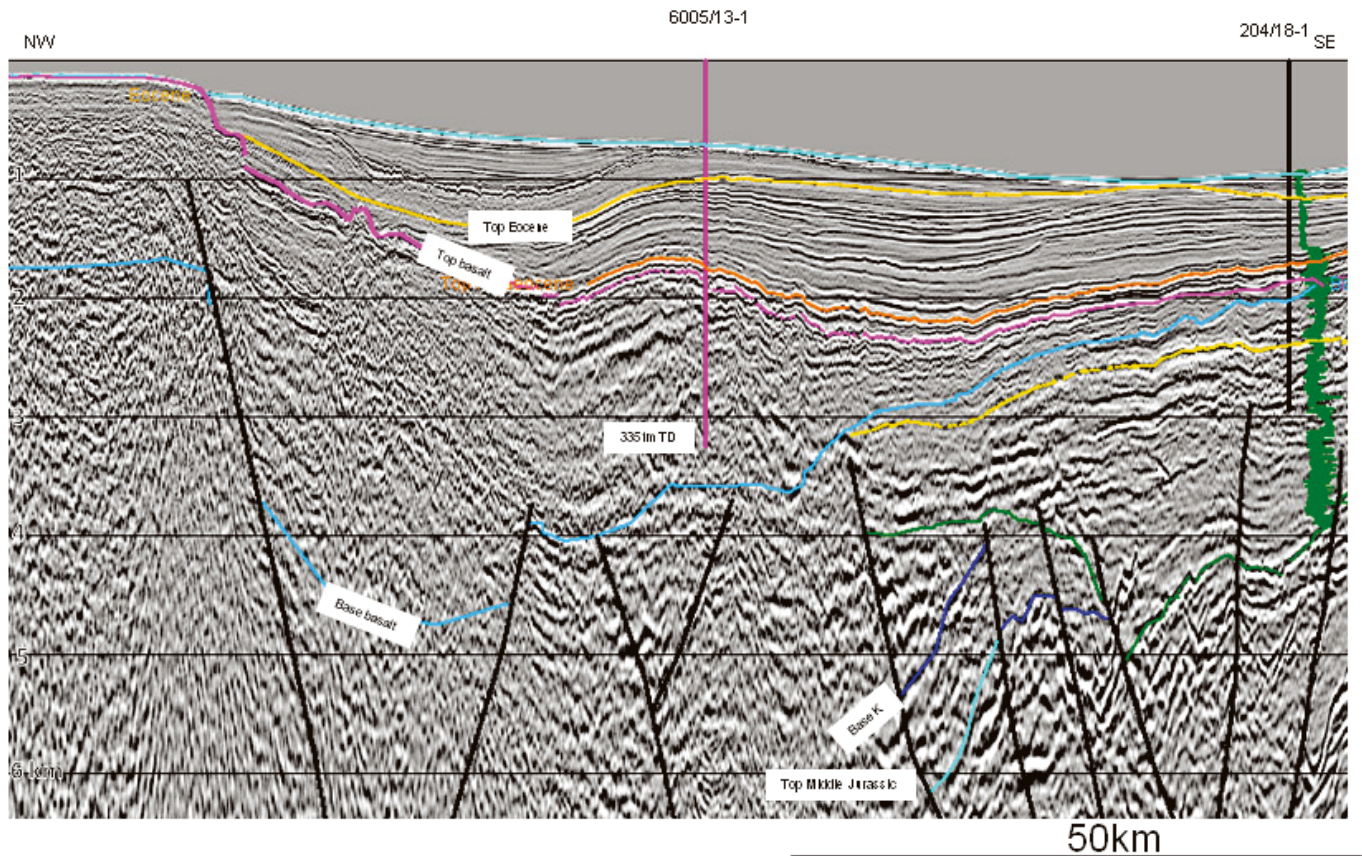
- a) Seismic line across the Faroes Shetlands basin showing the complete crustal structure. A large anticline is imaged below the basalts which may constitute a large potential trap in Mesozoic strata. This line is shot along the same location as the iSIMM line (White et al. 2005).
 b) Velocity model used to produce the PSDM image in a)

Faroes-Shetlands and Møre Basins

The Faroes-Shetlands and Møre Basins were proved to be important new hydrocarbon provinces when the Foinaven, Scheihallion and Ormen Lange Fields were discovered in the 1990s. However, subsequent discoveries have been smaller and more difficult to find. Large areas of the UK and Faroes sector are covered by thick Palaeogene volcanics from the Iceland plume eruptions. The Lopra-1 well on the Faroe Islands drilled 3565m of the lower basalt series (Ziska and Andersen 2005). Their total thickness is not known but seismic data suggest perhaps > 6 km of basalt may be present below the Faroe Islands (White et al. 2005). Only two wells (6104/21-1, 6005/13-1) have been drilled in the Faroes continental shelf into thick basalt and the results have not yet been released into the public domain. The volcanics have a high seismic absorption (Q factor, Maresh et al. 2006; Shaw et al. 2008) caused by interbedded lower velocity sediments, coals, and paleo-sol horizons which

produce a large number of internal reflections. Several seismic lines have been shot with wide aperture arrays or sea-bottom seismometers that have produced encouraging results with sub-basalt imaging (Richardson et al. 1999, White et al. 2003 and 2005). However, until now there have been no basin wide seismic data sets available which image the whole crustal structure.

The base basalt is interpreted to be a transitional zone of thinly bedded tuffs lavas and sediments as there is usually no clear seismic event associated with the base (Fig. 3). However, deeper stratal reflections are clearly imaged in many areas below a weakly reflective zone. Large-scale Cenozoic age folds (Munkagrundur Ridge and Fugloy Ridge; Ritchie et al. 2008) and Mesozoic rotated fault blocks constitute major potential traps with Mesozoic reservoirs and Jurassic source rocks present (Lamers & Carmichael 1999; Figs. 2,3 & 4). The Cenozoic folds are cored by Late Cretaceous strata (well 6004/16-1; Smallwood, 2009) and are laterally

**Fig. 3.**

Seismic section through the 6005/13-1 illustrating the problem with definition of the base basalt. The underlying rotated fault blocks are well imaged and may be viable exploration targets situated at >4 km depth. The well log for 204/18-1 is the sonic curve.

extensive. More closely-spaced seismic lines will be required to properly define structural closures on these long ridges. To our knowledge many of these deeper sub-basalt structures were not visible on previous seismic data. They are located in water depths up to 2 km, at drillable depths, and in the oil window (4-6 km); making them viable exploration plays, which have yet to be tested.

The onset of volcanism started at 61-62 Ma (Saunders *et al.* 1997), with the main period of extrusion following at 56-53 Ma which accompanied seafloor spreading of the North East Atlantic (Chron 24). Magmatism along the North East Atlantic produced a thick Seaward Dipping Reflector (SDR) zone which is oriented parallel to the ocean-continental crust boundary (Spitzer *et al.* 2008; Fig.1). The individual SDRs extend for some 20 km horizontally in a NW-SE direction orthogonal to the spreading centre. This length of lava run-out is thought to be indicative of sub-aerial volcanism (Fig. 2). This suggests that

the mid-ocean ridge was sub-aerial at this time due to the anomalously high mantle potential temperature created by the Iceland plume. The ocean-continent crust boundary is here defined as the seaward end of the last visible rotated fault block below the thick basalts (>2 km), which lies close to the landward edge of the main SDR sequence (Fig. 2).

The SDRs reach up to approximately 6-7 km maximum thickness. We interpret this to be new oceanic crust erupted above sea-level when they reach this thickness. The basalts flowed landward from the spreading centre to produce lava deltas, with eastward prograding foresets. These reach up to 1 km in vertical relief indicating the probable palaeo-water depths (Fig. 4). Below the landward (eastern) edge of the basalts, rifted sedimentary strata are clearly imaged which reach up to 5-7 km in thickness these can be traced for a short distance below the basalts (Figs. 2 and 4). Stratal reflections are imaged, even under the

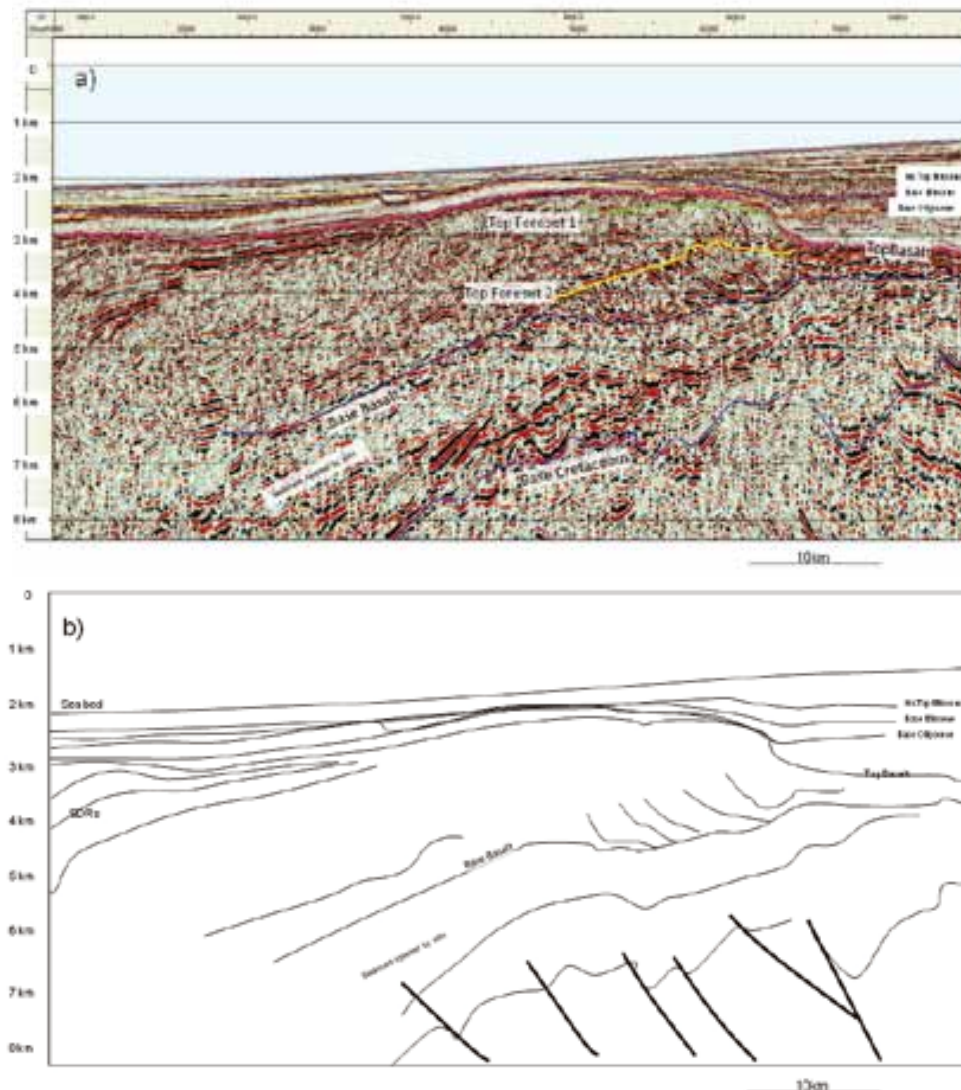


Fig. 4.

Seismic section through the Møre Basin showing a lava pile with a landward-dipping scarp edge. Two foreset sequences of lavas are piled one top of the other. The lavas overlie rifted fault blocks which have been injected by sills (bright reflections). These rifted basins are buried to depths of 5-6 km and reservoir quality is predicted to be poor at these depths.

thickest sequence of SDRs, however these are apparently injected by a large number of magmatic sills (Fig. 4). Strong irregular reflections occur at depths of 12-16 km below the SDRs, and are interpreted to be basic intrusions injected into Lewisian basement (Fig. 2). Prominent intrusions are occasionally imaged where there is a strong landward (eastward) dipping reflection imaged at depths of 10-15 km (Fig. 2).

Many igneous sills were injected into the crust during the Paleogene (Fig. 4) and 70 such sills have been intersected by the well 164/7-1; these sills range in thickness from 2 to 150m in thick-

ness (Linnard and Nelson, 2005). The sills are mainly injected around 1-3 km below the level of the Palaeocene basalts, where the depth of sill injection was determined by the depth at which the magmatic pressure exceeded the prevailing lithostatic pressure. Below this depth, the magmatic pressure was not great enough to overcome the lithostatic pressure and only dyke injection would have occurred. Hence, the deeper Jurassic source rocks may have escaped pervasive heating from sill injection as they are situated at greater depths than 3 km below base basalt.

Structural inversion and uplift occurred in

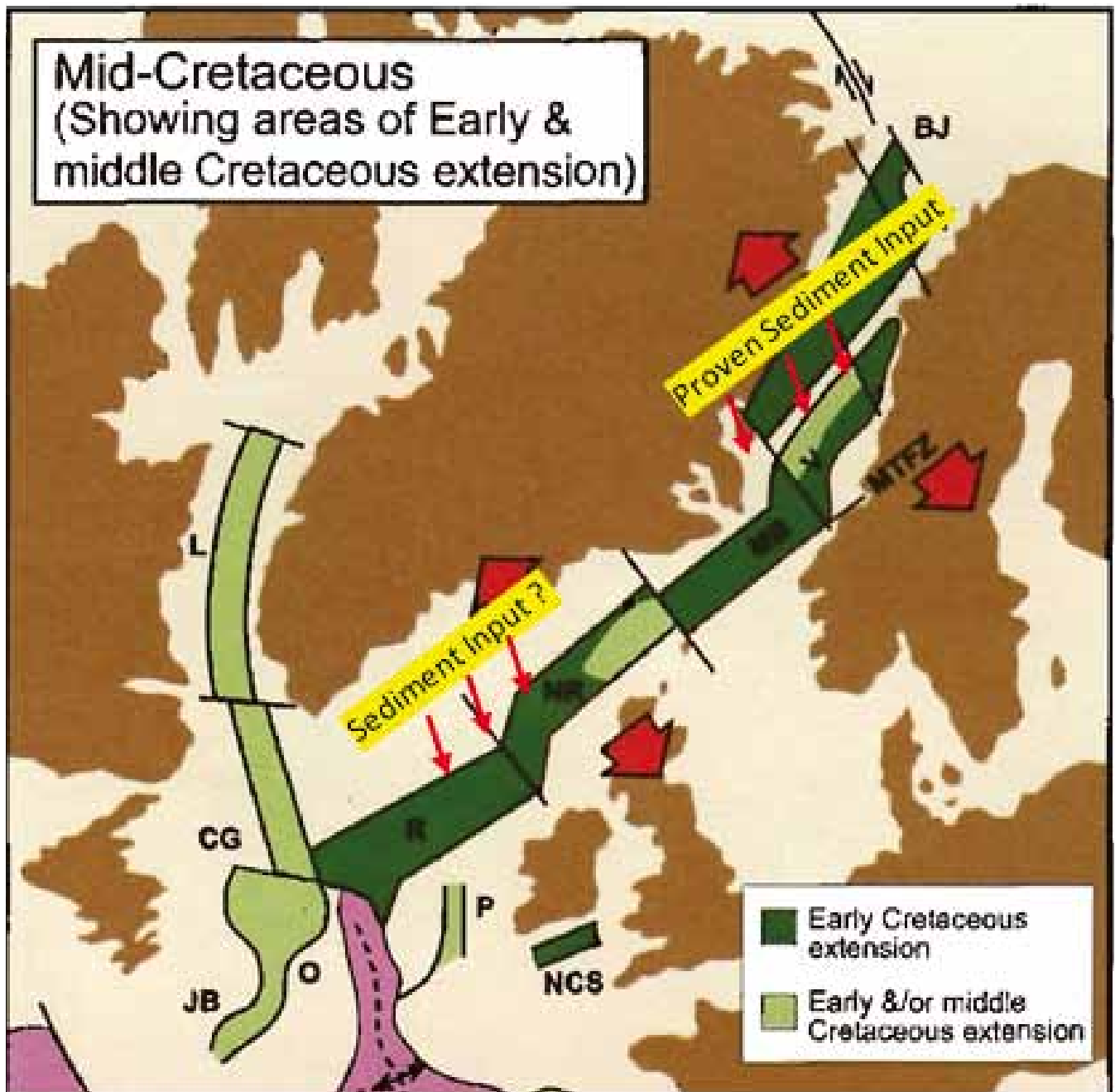


Fig. 5.

Map of the NE Atlantic showing a pre-drift reconstruction (Doré *et al.* 1999). Cretaceous strata are derived from Greenland in the Vøring Basin and Palaeocene strata were derived from Greenland in the Shetlands area. We suggest here that Late Cretaceous sandstones in the Rockall could also be sourced from west of the basin.

the Middle to Upper Eocene when the Wyville – Thompson and Ymir Ridges were formed. These structures were also reactivated in the Miocene (Ritchie *et al.* 2003). These ridges are capped by basalts and it has never been possible to image the sedimentary packages below the basalt. The new data presented here indicate that underlying layered Mesozoic strata are present and that

rotated fault block geometries are present in the anticlines. Consequently, large structural closures may be expected below the basalts.

Conclusions

New PSDM seismic data along the NE Atlantic margin have successfully imaged the sub-basalt

strata and indicate that there are many potentially large structural closures located below the Paleogene basalts and sills in the Faroes-Shetlands and Møre basins. The base of the Paleogene volcanics is not clearly imaged on the seismic data, probably because this is a transitional contact with little acoustic impedance contrast; the thickness of basalts is therefore very difficult to predict. This increases the (drilling) risk of exploration targets but this does not preclude exploration as the deeper Mesozoic half grabens are well-imaged 1-2 km below the presumed base of the volcanics.

The new seismic data presented here have helped to identify many potential structures below the basalts, which are some of the largest undrilled structures in NW Europe. Deep – towed long streamer seismic data have successfully imaged below the basalt province.

Acknowledgments

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The Wyville Thomson Ridge Complex located in the NE Atlantic – Aspects of the Tertiary development

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Abstract

The Wyville Thomson Ridge Complex (WTRC) located in the NE Atlantic is a major anticline compound that has been investigated by interpreting modern commercial 2D digital reflection seismic data. The complex constitutes of Wyville Thomson Ridge (WTR) and the Ymir Ridge (YR) and is so far undrilled. It is proposed that the orientation of the apex alongside the WTR and the YR changes direction; the WTRC tilted towards the southeast; the WTRC experienced a clockwise rotation from Late Paleocene until Middle Miocene; the WTRC is segmented by two ENE/WSW trending fissures and adjacent NE/SW trending transfer faults; beneath the south eastern part of the Ymir Ridge a transcurrent fault ends as a listric fault in the Rockall Basin. Four compressional phases affected the WTRC 1) Late Paleocene - Early Eocene 2) Early Eocene 3) Early Oligocene and 4) Middle Miocene. Based on the seismic interpretations a structural model is presented.

Introduction

The objective of this paper is to present a simple structural model accounting for all the diverse stress regimes and resulting strain that have been involved in the post volcanic evolution of the Wyville Thomson Ridge Complex (WTRC) (Fig. 1). This was obtained by resolving the Tertiary movement of WTRC, by interpretation of the post basalt sediments from 2D modern reflection seismic profiles (Fig. 1). The seismic interpretation was correlated to the nearby areas. In order to construct the model the stratigraphic development of the basins next to the Wyville Thomson Ridge (WTR) and Ymir Ridge (YR) was analysed.

Geological Setting

Since the collapse of the Caledonides during the late Silurian – Early Devonian (Archer *et al.*, 2005) several episodic, non-magmatic extension phases occurred in the area with the NE Atlantic margin resulting in a wide rifted region (Lundin and Doré, 2005) bearing the NE-SW Caledonian trend (e.g. Archer *et al.*, 2005). Within the Faroe-Shetland, Rockall and Porcupine areas at least six main phases of rifting events have been described (Dean *et al.*, 1999): (1) Devonian – Carboniferous (2) Permian – Triassic (3) Middle Jurassic (4) Late Jurassic (5) Cretaceous and (6) Paleocene.

During Hauterivian (Early Cretaceous c. 135-

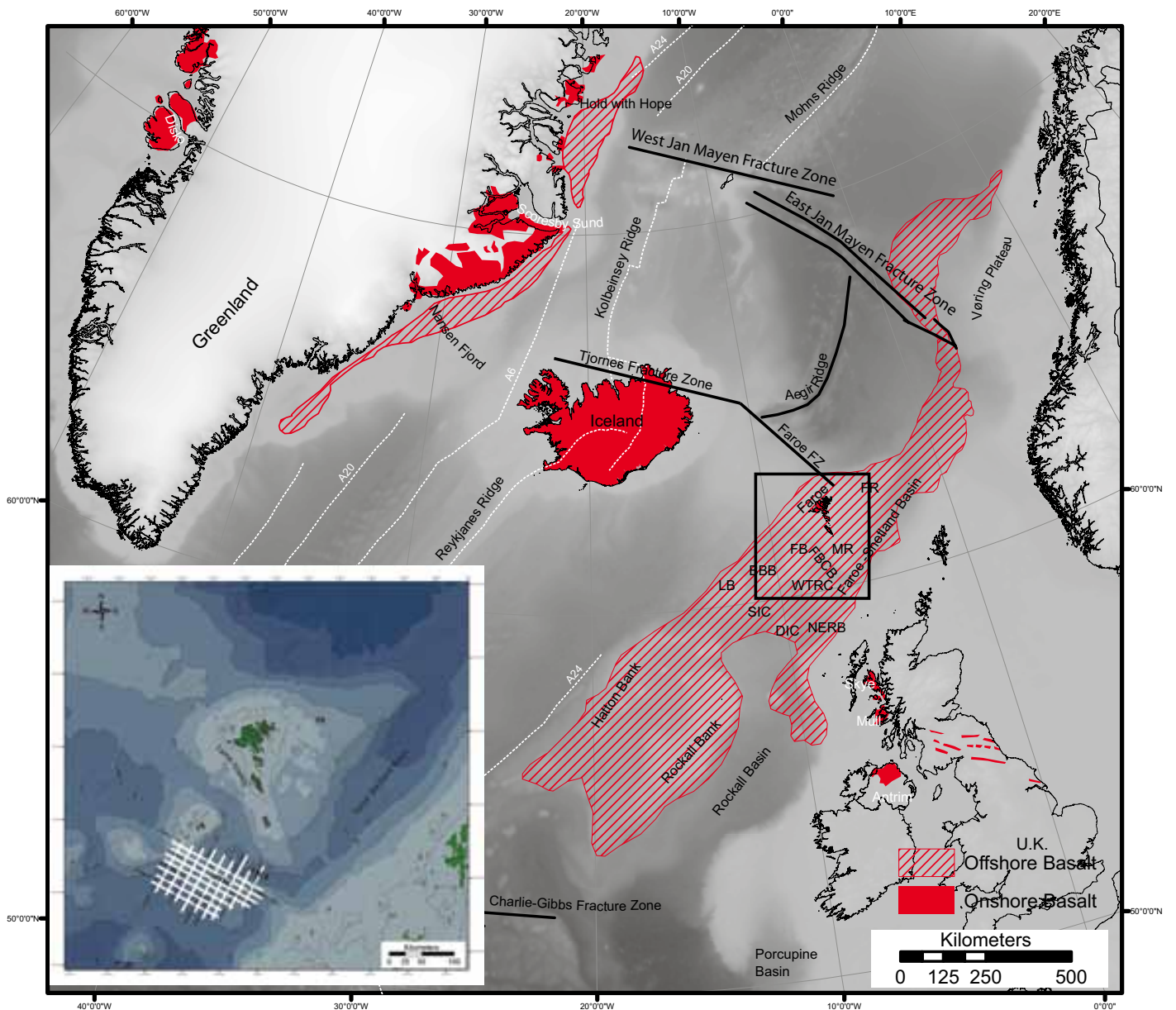


Fig. 1.

Geographical map of the North Atlantic including structural features in the region: the North Atlantic Igneous Province (NAIP) extends from Baffin Islands (not shown) and to the northern part of the British Isles (Adapted from Ritchie *et al.*, 1999, Frei, 2005 and Ganerød *et al.* 2008). The bathymetry data is the ETOP01 data set (Amante and Easkins, 2008). Faroe Fracture Zone and Tjörnes Fracture Zone (redrawn from Kimbell *et al.*, 2005). BBB, Bill Bailey Bank; DIC, Darwin Igneous Centre; FB, Faroe Bank; FBCB, Faroe Bank Channel Basin; FR, Faroe Ridge; LB, Lousy Bank; MR, Munkagrannur Ridge; NERB, North East Rockall Basin; SIC, Sigmundur Igneous Centre; WTRC, Wyville Thomson Ridge Complex. The seismic survey YMR97 investigated in this study is visualized in the insert figure.

132 Ma) the extensional stress direction shifted from E-W to NW-SE (Doré *et al.*, 1999) when the southerly propagating Arctic rift (consisting of the Faroe-Shetland-Møre basins) united with the northerly propagating Atlantic rift (consisting of the Rockall Basin) forming a single, linked

rift system (Roberts *et al.*, 1999) (Fig. 1). Oceanic spreading and voluminous and widespread volcanism accompanied the rift phases culminating with the separation of Greenland and Eurasia (e.g. Saunders *et al.*, 1997). The North Atlantic seafloor spreading started in Aptian time (middle

Early Cretaceous c. 118 Ma) between Newfoundland and Bay of Biscay/Iberia representing a weak NE-SW extensional stress direction (Lundin and Doré, 2005).

Rifting between Greenland and Eurasia, including the Faroe-Rockall Plateau (Boldreel and Andersen, 1993) plus the Jan Mayen microcontinent (Hinz *et al.*, 1993; Eldholm and Thiede, 1980) initiated in Paleocene (57–56 Ma Chron 24/25) (Larsen, 1988) or (c. 56–53 Ma) (Saunders *et al.*, 1997). The simultaneous rifting on each side of Greenland, in the Labrador Sea/Baffin Bay and the NE Atlantic, was linked at a triple junction south of Greenland (Lundin and Doré, 2005). The ocean spreading in the Labrador Sea slowed and stopped in Middle – Late Eocene between Chrons 21 and 13 (c. 50 Ma until 36 Ma) (Saunders *et al.*, 1997) but Chron 13 also marks the initiation of a continuous spreading axis between the Arctic and the NE Atlantic (Lundin and Doré, 2005). Oceanic spreading between the Faroe Fracture Zone (Bott, 1985) and the Jan Mayen Fracture Zone transferred north-westwards from the Aegir Ridge to the Kolbeinsey Ridge crossing the Jan Mayen Ridge (Fig. 1) isolating the Jan Mayen microcontinent (e.g. Nunns, 1983). Some disagreement exist on the timing of this event: from the initiation of break-up in Paleocene – Early Oligocene (Chron 24 c.54 Ma until Chron 12 c.32 Ma) when the Aegir Ridge was abandoned (Lundin and Doré, 2005), Middle Eocene – Early Miocene (Chron 22–20 until Chron 6) (Larsen, 1988), Late Eocene – Oligocene until Early Miocene (Chron 20 until Chron 7) (e.g. Nunns, 1983) or Late Oligocene (Chron 7 c.25 Ma until Chron 6) at the culmination of the Jan Mayen microcontinent separating from Greenland (Hinz *et al.*, 1993; Eldholm and Thiede, 1980).

This ridge transition caused the Jan Mayen microcontinent to rotate approximately 8° anti-clockwise relative to East Greenland and NW Europe (Boldreel and Andersen, 1993) and is well recorded by the difference in the orientation between the Tjörnes and Faroe Fracture Zones (Kimbell *et al.*, 2005) and between the West- and East Jan Mayen Fracture Zone making up the Jan

Mayen Fracture Zone (Fig. 1) (Talwani and Eldholm, 1977). This ridge transition resulted in a change in the relative motion of Greenland and Eurasia from approximately normal to the mid-ocean ridge, constituting of the Reykjanes Ridge south of Iceland and the Kolbeinsey Ridge to the north, to the spreading that persists to the present day. This spreading ridge was accomplished by utilization of the Caledonian suture zone and can be viewed as the natural consequence of the Pangaeon break-up (Lundin and Doré, 2005). The rotation of the Jan Mayen microcontinent was manifested on the Faroe-Rockall Plateau (Boldreel and Andersen, 1993) where the Late Eocene unconformity developed within the Hatton Bank area (Johnson *et al.*, 2005), the Rockall Basin, the north Faroe-Shetland and North Sea Fan-Vøring areas as the intra-Miocene unconformity (constituting of both Late Eocene and Base Neogene due to stratigraphic inter-correlation uncertainty over the WTRC) (Stoker *et al.*, 2005).

The rifted continental margin between Greenland and Eurasia experienced the influence of the Icelandic hotspot around 64 Ma (Archer *et al.*, 2005) that produced abundant igneous activity (e.g. Smith *et al.*, 2005) with an overall melt estimation of $5\text{--}10 \times 10^6 \text{ km}^3$ that was generated in only 2 Ma (White *et al.*, 1987) and is referred to as the North Atlantic Igneous Province (NAIP) (e.g. Saunders *et al.*, 1997; Keser Neish and Ziska, 2005). The volcanic activity on Faroe Islands lasted 4 to 7 million years mainly in the Thanetian and ceased before the opening of the NE Atlantic during Chron 24R close to the Paleocene/Eocene boundary (Waagstein, 1988). The Faroese onshore basalts are subdivided into three separate formations: The Beinisvørð Fm (Passey and Bell, 2007) (>3000 m) that extruded during Chron 26R to Chron 25N and is present over most of the Faroe-Rockall Plateau (Waagstein, 1988). The Beinisvørð Fm is overlain by the Malinstindur Fm (Passey and Bell, 2007) (~1400 m) (Waagstein, 1988) that is overlain by the Enni Fm (Passey and Bell, 2007) (~900 m) (Waagstein, 1988) and both extruded during Chron 24R and are confined to a relative narrow area along the

Stratigraphic Correlation Diagram around the WTRC							
Author/Area	Present Article	STRATAGEM Partners, 2002	Sørensen, A. B., 2003	Andersen et al, 2000	Tate et al., 1999	Boldreel & Andersen, 1993	
Epoch	WTRC	South Western Faroe-Shetland Channel	Faroe area	Eastern Faroe region	NE Rockall Basin	Faroe-Rockall area	
Pleistocene	16	FSN-1	D3	FPC-D.3	9	Pliocene – Pleistocene	
	12		Glacial	CN-050			
Pleistocene – Middle Pliocene	15 14 *		D2	FPC-D.2			
	11	GU/INU	Pliocene	CN-040		Late Miocene	
Early Pliocene – Late Miocene	12 10 11	FSN-2	D1a + D1b	FPC-D.1		Late Mid Miocene	
	9		Mid Miocene	CN-030		Mid Miocene	
Early Miocene	10		C	FPC-C		Early Miocene	
	8	TPU	Top Oligocene	CN-010		Mid Oligocene	Late Oligocene
Early Oligocene	8 7 7	FSP	Oligocene	FPC-B.4		8	Late Eocene – Oligocene
	6		Top Eocene	CP-100		Top Eocene	Eocene – Oligocene
Middle Eocene	6		Mid Eocene	FPC-B.3	7	Mid – Late Eocene	
	5		Mid Eocene	CP-060	Mid Eocene	Intra Eocene	
Early Eocene	5		Early Eocene	Early Eocene	FPC-B.2	6	Early Eocene
	4						
	4						
	3						
	3						
	2						
	2						
	1						
	1						
	Top Basalt						
	Late Paleocene, SS, DS, FBCK and Drekaeyga Intrusion		Top Paleocene, T40 – T45		Brown	Paleocene – Eocene	
Late Paleocene			Middle & Upper Basalt Series, T38 – T40		4b basalts	Middle & Upper Basalt Series	
					Base Basalt		
					3		

Fig. 2.

A stratigraphic correlation diagram visualizing geological time versus geological locations adjacent the investigated area interpreted by different authors, showing successions (white) and their bounding unconformities or conformable reflectors/seismic key markers (pale green). T numbers are BP sequences. Numbers in the column “Present Article” refer to the interpreted seismic stratigraphic units in the present study. *Unit 14 is deposited in the AB and due to erosion the lower boundary is reflector 9. See section „Summarizing the three Sub Areas“ for further details.

continental margin. The Prestfjall Fm (Passey and Bell, 2007) previously called the coal-bearing sequence (Waagstein, 1988) and the Hvannhagi Fm (Passey and Bell, 2007) also belongs to the Chron 24R and separates the Beinisdvörð Fm and the Malinstindur Fm (Waagstein, 1988).

It is generally accepted that the Icelandic hot-

spot was the location for the break-up in the NE Atlantic, but Lundin and Doré (2005) investigated alternative origins as a consequence of plate tectonics. The NAIP was divided into two pulses (e.g. Saunders *et al.*, 1997) due to the timing and amount of the magmatic eruption, plus that the location of these magmatic areas for a specific

time interval trended almost perpendicular to one another.

The first pulse was in Middle Paleocene (c. 62–58 Ma) and mainly confined to continent-based magmatism in the form of intense dyke swarms in the British Tertiary Igneous Province (Archer *et al.*, 2005) also called the British Volcanic Province (BVP), West Greenland and eastern Baffin Bay governed by a short-lived attempt at seeking a new rift path (Fig. 1). This NW–SE belt called the Thulean Volcanic Line (Hall, 1981) indicate a NE–SW extensional stress field during part of the Paleocene (England, 1988) probably enhanced by the Pyrenean phase of the Alpine collision that was replaced as stretching and subsequent separation refocused on the NE Atlantic margin in the later Paleocene – Early Eocene (Archer *et al.*, 2005).

The second pulse was in latest Paleocene – earliest Eocene (c. 56–53 Ma) with voluminous magmatism along the NE Atlantic margins related to the final break-up of Pangaea, exploiting the collapsed Caledonian fold belt (Archer *et al.*, 2005).

Several post break-up compressional phases took place in the NE Atlantic margin and generally the Eocene – Miocene compressional phases, in form of basin inversion observed in major basins in NW Europe, are assumed to be related to the Alpine deformation (e.g. Ziegler 1987). Within the Faroe-Shetland and Hatton-Rockall areas three compressional/transpressional phases, developing major anticlinal folds, are identified: (1) Late Paleocene (Thanetian) – Early Eocene, gravitational ridge push from the N (Boldreel and Andersen, 1993) forming the WNW-trending WTRC and the NNW-trending Munkagrunnur Ridge, and the ENE-trending Fugloy Ridge might also have been active (Boldreel and Andersen, 1994) (2) Oligocene, gravitational ridge push from the NNW (Boldreel and Andersen, 1993) developing NE- to ENE-trending fold axis to the east of the Faroe Islands and in the area between Faroe Islands and Hatton Bank (Boldreel and Andersen, 1994), and at the WTRC erosional truncations are seen on the Late Oligocene unconformity

(Boldreel and Andersen, 1993) and (3) Middle – Late Miocene, gravitational ridge push from the NW developing WSW–ENE to SW–NE-trending anticlines in the FSB plus almost E–W-trending structures in the Faroe Rockall Plateau (Boldreel and Andersen, 1993). However, Andersen *et al.* (2002) only recognised two major compressional phases in the Faroe Platform area (1) Middle Eocene and (2) Middle – Late Miocene times. Many scenarios have been described in the attempt to explain the genesis of the WTR and the YR and therefore also of the WTRC e.g. Boldreel and Andersen (1993), Johnson *et al.* (2005), Keser Neish and Ziska (2005), Kimbell *et al.* (2005), Smith *et al.*, (2009), Stoker *et al.* (1993) and Tate *et al.* (1999).

Materials and Methods

In 1997 Fugro-Geoteam AS recorded a 2D seismic survey YMR97 (1782 line km) SW of the Faroe Islands. The source type was a G-Gun 3040 Cu. inch array fired at a depth of 8 m below the sea surface and at an interval of 25 m. The 6 km long cable, placed in 10 m depth in the water with an offset of 140 m, contained 480 channel groups with a group length of 12.5 m and the CDP interval was 6.25 m. The recorded length was 8 sec. with a 2 msec. sample rate, filters of a Low-cut 4 Hz and a High-cut 204 Hz, and the data were stored as normal polarity in a SEG-D 8015 format. Afterwards Robertson Research International LTD. processed the data to zero phase signal with a CDP gather of 120 fold and used the Kirchhoff summation migration method.

The criteria applied for reflector interpretation has been to define the base of each seismic unit on the 16 profiles used. The seismic profiles were thus thoroughly interpreted as a large numbers of reflectors and units have been analysed.

The investigated area was divided into three sub areas named the Rockall Basin (RB), Auðhumla Basin (AB) and Faroe Bank Channel Basin (FBCB) located from the south towards the north (Fig. 1). The three sub areas were interpreted independently and afterwards the

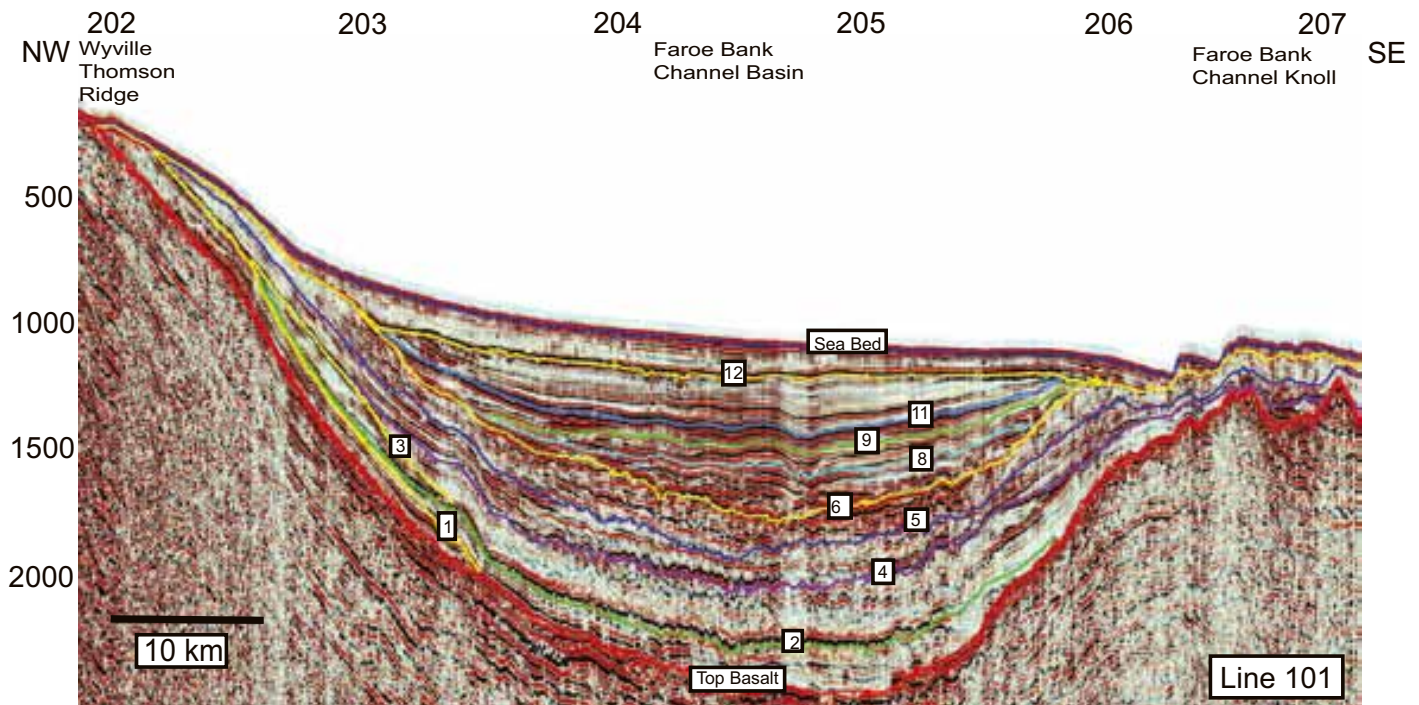


Fig. 3.

The key profile chosen in the FBCB is the WNW-ESE seismic line 101 approximately 161 km long. 11 reflectors were interpreted and 11 units analysed and are referred to in the text. The depth in ms TWT and the crossing seismic lines are shown. To the NW the upper part of the WTR is eroded. On the WTR north western flank adjacent to line 203 two small bulges are expressed at the Top Basalt that continues until the reflector 6. The sedimentary units below reflector 6 display a chaotic seismic pattern whereas the younger units show a non chaotic seismic character with a similar geographical extent of the depositional area.

interpretation was merged from south to north. A direct correlation can be established between the interpreted reflectors in the RB and the AB, but a similar connection is not possible between the AB and the FBCB. To assign age relation to the interpreted reflectors STRATAGEM Partners (2002), Sørensen (2003), Andersen *et al.* (2000), Tate *et al.* (1999) and Boldreel and Andersen (1993) were used for correlation (Fig. 2).

Three interpreted key seismic profiles (Figures 3, 4 and 5) representing each sub area and one profile (Fig. 6) that intersects the sub areas are shown in order to illustrate the connection and correlation of the development in the sub areas. The profiles that represent the RB and the FBCB are located high on the flanks of the ridges, and therefore reflect the deposition of the sediment packages located on the flanks rather than in the basins. The profile in the AB is located in the centre of the basin. The seismic units are named after

the upper reflector demarcating the units.

Results

Top Basalt

The top of the basalt is a high amplitude peak reflector in the entire survey and a depth map to the surface of the basalt is produced (Fig. 7). The interpretation shows that the basalt in places is eroded at the apex of the WTR and the YR, and that the orientation of the apex alongside both ridges changes.

Faroe Bank Channel Basin - Observations of the Seismic Units

In the Faroe Bank Channel Basin the seismic line 101 (Fig. 3) was chosen as the key profile.

FBCB Unit 1– FBCB Unit 5

FBCB Unit 1 is at the lower boundary limited by reflector Top Basalt and at the upper boundary by reflector 1. Reflector 1 is characterized as a high amplitude peak that downlaps the Top Basalt reflector in the basinwards direction. The external form of FBCB Unit 1 is a prograding wedge. The internal reflection pattern consists of subparallel downlapping reflectors of high amplitude and frequency and truncations occur at the top of FBCB Unit 1 towards the WTR. The unit has a limited extent as it is present only at the intersecting of line 202, 203 and 101.

FBCB Unit 2 is at the lower boundary limited by reflector Top Basalt and at the upper boundary by reflector 2. Reflector 2 is a high amplitude trough that onlaps the Top Basalt reflector and downlaps the Top Basalt reflector basinwards. The external form of FBCB Unit 2 is a basin fill. In the NW the internal pattern is at the lower part represented by onlapping high amplitude and frequency subparallel continuous reflectors and truncation occurs in the upper part. Basinwards the pattern becomes chaotic to contorted and lower amplitude and frequency is observed as compared to the NW. The unit is found in most of the FBCB.

FBCB Unit 3 is at the lower boundary limited by reflector 2 and at the upper boundary by reflector 3. Reflector 3 is a high amplitude trough and onlaps reflector 1 towards the WTR and downlaps reflector 2. The external form of FBCB Unit 3 is a prograding wedge. The internal pattern is continuous subparallel reflectors of medium to high amplitude and low to high frequency. The unit is limited to the intersection of line 203 and 101.

FBCB Unit 4 is at the lower boundary limited by reflector 2 and at the upper boundary by reflector 4. Reflector 4 is a high amplitude undulating trough that onlaps the older boundaries. The external form of FBCB Unit 4 is a basin fill. The internal pattern is onlapping subparallel continuous reflectors of medium to high amplitude and low to medium frequency that become disrupted in the central part of the FBCB. The unit is found in FBCB apart from FBCK.

FBCB Unit 5 is at the lower boundary limited by reflector 4 and at the upper boundary by reflector 5. Reflector 5 is a low amplitude peak that onlaps reflector 4 and in the north western area reflector 3. The external form of FBCB Unit 5 is a basin fill. The internal pattern consists of onlapping subparallel continuous reflectors of medium to high amplitude and low to medium frequency. The unit is found in FBCB apart from the FBCK.

FBCB Unit 6

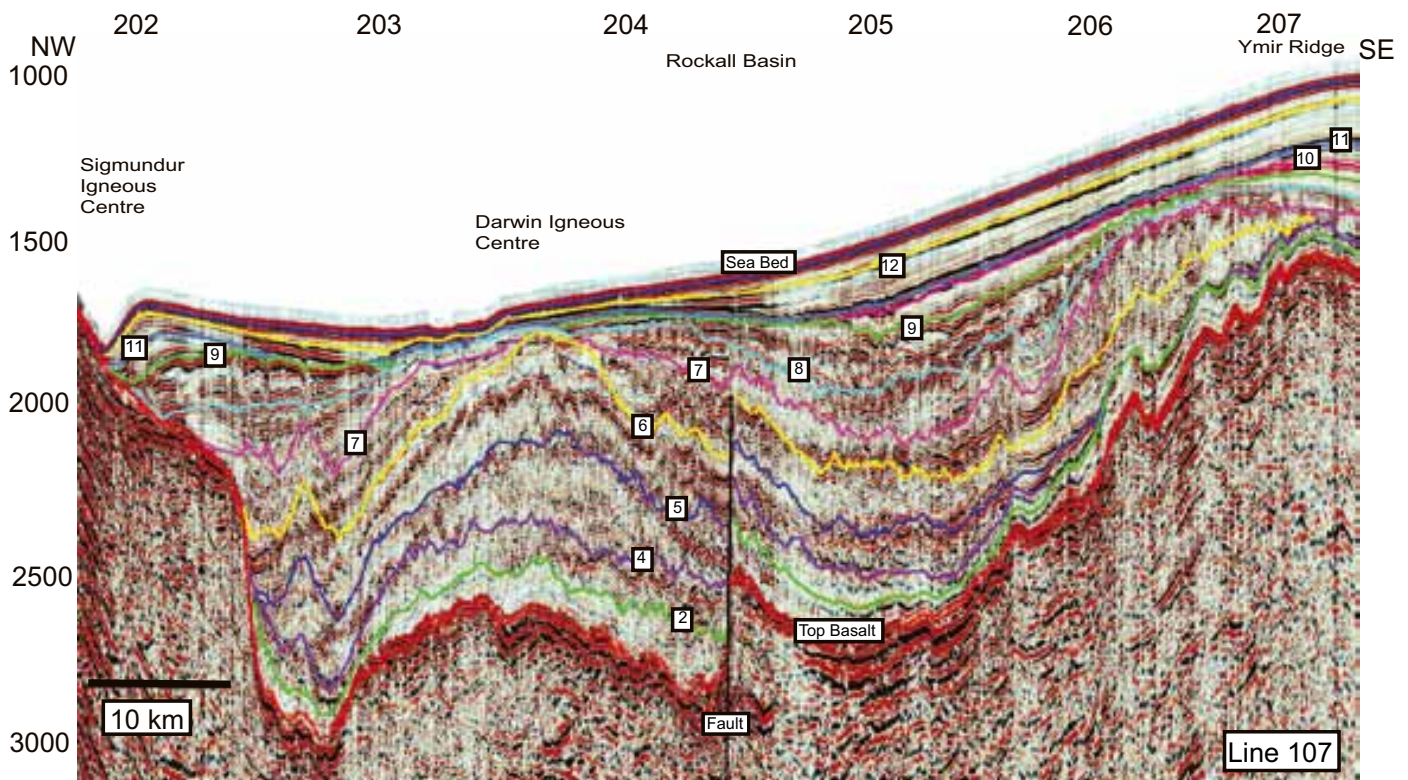
FBCB Unit 6 is at the lower boundary limited by reflector 5 and at the upper boundary by reflector 6. Reflector 6 is a high amplitude peak that is truncated at the top of FBCB Unit 6 in the NW. The external form of FBCB Unit 6 is a basin fill. The internal pattern is onlapping subparallel continuous reflectors of medium to high amplitude and low to medium frequency. The unit is found in FBCB apart from the FBCK.

FBCB Unit 7

FBCB Unit 7 is at the lower boundary limited by reflector 6 and at the upper boundary by reflector 8. Reflector 8 is a high amplitude peak that onlaps reflector 6. The external form of FBCB Unit 7 is a basin fill changing into a prograding wedge on line 206. The internal pattern shows high amplitude and medium to high frequency continuous subparallel reflectors. The unit is present in the area where lines 202-206 intersect 101.

FBCB Unit 8

FBCB Unit 8 is at the lower boundary limited by reflector 8 and at the upper boundary by reflector 9. Reflector 9 is a high amplitude peak that onlaps reflector 8 towards the NW and seems to onlap reflector 6 close to the FBCK. The external form of FBCB Unit 8 is a prograding basin fill changing into a prograding wedge close to the FBCK. The internal pattern is characterized of high amplitude and medium to high frequency by onlapping continuous subparallel reflectors, although downlap is observed adjacent the FBCK. The unit is found in the area where lines 203-206 intersect 101.

**Fig.4.**

The key profile chosen in the RB is the approximately 120 km long WNW-ESE seismic line 107 along with 11 reflectors interpreted and 11 units analysed. The depth in ms TWT and the crossing seismic lines are shown. The Sigmundur Igneous Centre (SIC) is located in the NW with the adjacent Darwin Igneous Centre (DIC) and ending with the YR in the SE. The SIC is demarcated by an escarpment and the flank of DIC towards the north is demarcated by a fault being active until Early Oligocene. The lava flows that originate from the SIC show onlap to the lava flows from the DIC in a SE direction. Above the DIC, between the SIC and the fault, the sediments up to Top Oligocene are squeezed and bulging upwards as a dome indicating compression. SE of the fault the only indication of compression is a few bulges adjacent to line 206 at the Top Basalt that continues in the sedimentary units until the Top Oligocene.

FBCB Unit 9

FBCB Unit 9 is at the lower boundary limited by reflector 9 apart from the NW where the lower boundary is reflector 6, and is at the upper boundary limited by the reflector 11. Reflector 11 is a high amplitude trough that onlaps the lower boundary. The external form of FBCB Unit 9 is an onlapping basin fill changing into a prograding wedge close to the FBCK. The internal pattern constitutes of onlapping continuous subparallel reflectors of high amplitude and medium to high frequency, however downlap is observed adjacent the FBCK. The unit is found in the area where the lines 203-206 intersect 101.

FBCB Unit 10

FBCB Unit 10 is at the lower boundary limited by reflector 11 and at the upper boundary by reflector 12. Reflector 12 is a high amplitude trough that onlaps the lower boundary. The external form of FBCB Unit 10 is a prograding basin fill changing into a prograding wedge close to the FBCK. In the NW the internal pattern is represented by onlapping continuous subparallel reflectors of medium to high amplitude and low to high frequency and truncations are found at the top of FBCB Unit 10 in the NW. Adjacent the FBCK the internal pattern shows onlapping reflectors that basinwards downlap the lower boundary of medium ampli-

tude and low to medium frequency. The unit is present in the area of intersecting lines 203-209 with the 101 apart from the FBCK.

FBCB Unit 11

FBCB Unit 11 is at the lower boundary limited by reflector 12 and at the upper boundary by reflector Sea Bed. Reflector Sea Bed is a high amplitude peak in the entire survey. The external form of FBCB Unit 11 is a basin fill. The internal pattern consists of onlapping continuous subparallel reflectors of medium to high amplitude and low to high frequency. The unit is present in the whole FBCB.

Interpretations of the Seismic Units

Early Eocene

FBCB Unit 1 has a limited areal distribution and is reasonable thin (Fig. 3). This indicates that a small amount of clastic material, as judged from the internal reflector pattern, may originate from the north western part of the WTR. It is interpreted that the relative sea level fell to allow for erosion of the flood basalt in the NW of the WTR. This relative sea level fall indicates the first uplift of the WTR as the unit is found only in the NW. This shows that the WTR did not act as one continuous ridge and thus it seems to be segmented.

FBCB Unit 2 consists of material that seems to originate from the WTR. The unit is extensive distributed in the basin and indicates that abundant depositional material was available (Fig. 3). To the NW the unit is deposited higher up the flank of the WTR than in the SE indicating that the WTR is segmented. The distribution of the unit indicates a relative sea level rise due to basin subsidence before and during the deposition.

FBCB Unit 3 has a limited areal distribution (Fig. 3) and indicates that a small amount of material which may well originate from the NW of the WTR. Before and during deposition of the unit the relative sea level fell to allow for erosion of the flood basalt in the NW of the WTR. The relative sea level fall indicates the second uplift of

the WTR. The extent of the unit shows that the WTR is segmented, and that the basin subsided towards the SE.

FBCB Unit 4 is found in the whole FBCB apart from the FBCK and indicates that abundant material, as indicated by the internal reflector pattern, may originate from the WTR (Fig. 3). To the SE the unit is deposited higher up the flank of the WTR than to the NW indicating that the WTR is segmented. The distribution of the unit indicates a relative sea level rise caused by basin subsidence before and during the deposition.

FBCB Unit 5 consists of material that seems to originate from the WTR and onlaps reflector 1 near the sea bed to the NW as judged from the internal reflector pattern (Fig. 3). The Early Eocene reflector represent the first compressional phase, as all the previous units can be seen as an continuation of the same development resulting in a compressional phase.

Middle Eocene

FBCB Unit 6 has the widest distribution among all the units. Approximately midway on the inclination of the WTR the reflectors show progradation and aggradation from the WTR (Fig. 3) signifying a fast depositional rate. Before and during the deposition of the unit the relative sea level rose due to basin subsidence.

After deposition of the reflector 6 the relative sea level fell and caused a break in sedimentation indicated by the erosion of the entire upper boundary of FBCB Unit 6. This relative sea level fall marks the Top Eocene unconformity and indicates the third uplift of the WTR.

Early Oligocene

FBCB Unit 7 is recognized in the deepest part of the basin (Fig. 3) and may originate from the WTR. The internal reflectors onlap reflector 6 and indicate an enhanced accommodation space during deposition of the unit caused by basin subsidence and a relative sea level rise.

Early Miocene

FBCB Unit 8 is found in the deepest part of the

basin and may originate from the WTR as judged from the internal reflector pattern (Fig. 3). The unit thickens basinwards and indicates that the accommodation space and the relative sea level were about the same or slightly enlarged as in the deposition of FBCB Unit 7.

Late Miocene – Early Pliocene

FBCB Unit 9 appears in the deepest part of the basin adjacent the FBCK and the internal reflector configuration is downlap (Fig. 6). The relative sea level was about the same or slightly rising as the basin continued subsiding leading to a slightly enhanced accommodation space.

Middle Pliocene – Pleistocene

FBCB Unit 10 is recognized in the deepest part of the basin and as based on the seismic interpretations originated from the WTR. The internal reflectors onlap reflector 11 in the NW (Fig. 3) but downlap on reflector 11 in the SE (Fig. 6). The accommodation space was slightly enhanced as the unit reaches further up the WTR than the previous unit and indicates a relative sea level rise caused by basin subsidence.

After deposition of FBCB Unit 10 a relative sea level fall occurred causing a break in sedimentation indicated by the erosion of the entire upper boundary of the unit (Fig. 3). The erosion was more severe in the NW where all reflectors show truncation and may be caused by one or more of the following 1) the accommodation space was filled up 2) strong sea currents initiated 3) glaciation began 4) only the NW of the WTR was uplifted indicating that the WTR is separated by faults or transfer zones. The relative sea level fall marks the Glacial Unconformity and indicates the fourth uplift of the WTR.

Pleistocene

During deposition of FBCB Unit 11 the basin subsided and this indicates a relative sea level rise.

Rockall Basin - Observations of the Seismic Units

In the Rockall Basin the seismic line 107 (Fig. 4) was chosen as the key profile.

RB Unit 1 – RB Unit 3

RB Unit 1 is at the lower boundary limited by reflector Top Basalt and at the upper boundary by reflector 2. Reflector 2 is an onlapping undulating trough. The external form of RB Unit 1 is a basin fill. The internal pattern is chaotic to contorted onlapping reflectors of medium amplitude and low to medium frequency. The unit is present in the RB apart from a fault in an area where the line 106 intersects line 204.

RB Unit 2 is at the lower boundary limited by reflector 2 and at the upper boundary by reflector 4. Reflector 4 is a high amplitude undulating trough that onlaps the top basalt. The external form of RB Unit 2 is a basin fill. The internal pattern is represented by onlapping chaotic to contorted reflectors of medium amplitude and low to medium frequency. The unit is present in the RB apart from the area where the lines 204 and 106 intersect due to a fault and in the area with intersecting lines 206-207 with 106-107 as a compression bulge is found here (See Fig.1 for seismic line location).

RB Unit 3 is at the lower boundary limited by reflector 4 and at the upper boundary by reflector 5 (Fig. 4). Reflector 5 is a low to medium amplitude undulating peak onlapping the lower boundary. The external form of RB Unit 3 is a basin fill. The internal pattern is chaotic to contorted onlapping reflectors of low to high amplitude and frequency. The unit is found in the whole RB apart from the SE where line 207 intersects line 107.

RB Unit 4

RB Unit 4 is at the lower boundary limited by reflector 5 and at the upper boundary by reflector 6. Reflector 6 is a low to high amplitude undulating peak that onlaps the top basalt at the YR and reflector 5 in the SE. The external form of RB Unit 4 is a basin fill. The internal pattern is chaotic to contorted onlapping reflectors of low to

high amplitude and frequency. The unit is present in the whole RB.

RB Unit 5 – RB Unit 6

RB Unit 5 is at the lower boundary limited by reflector 6 and at the upper boundary by reflector 7. Reflector 7 is a low to medium amplitude undulating peak that onlaps the top basalt towards the YR and reflector 6 to the SE. The external form of RB Unit 5 is a basin fill. The internal pattern shows onlapping chaotic to contorted reflectors of low to high amplitude and frequency. The unit is present in the RB apart from minor patches.

RB Unit 6 is at the lower boundary limited by reflector 7 and at the upper boundary by reflector 8. Reflector 8 is a low to medium amplitude undulating peak that onlaps the top basalt towards the YR and reflector 7 to the SE (Fig. 4). The external form of RB Unit 6 is a basin fill. The internal pattern shows onlapping reflectors that vary from subparallel and chaotic to contorted, all of low to high amplitude and frequency. The unit is present in the RB apart from an area where lines 205-207 intersect line 106 (Fig. 1).

RB Unit 7

RB Unit 7 is at the lower boundary limited by reflector 8 and at the upper boundary by reflector 9. Reflector 9 is a high amplitude peak (low to high in the NW) that onlaps the top basalt towards the YR (Fig. 4), and reflector 8 in the central and SE part of the RB. The external form of RB Unit 7 is a basin fill. The internal pattern shows onlapping subparallel reflectors that are chaotic to contorted in the NW all of medium to high amplitude and low to high frequency. Truncations are found in the central and north western part except adjacent the Sigmundur Igneous Centre (SIC) and the YR. The unit is present in the RB apart from small patches.

RB Unit 8 – RB Unit 9

RB Unit 8 is at the lower boundary limited by reflector 9 and at the upper boundary by reflector 10. Reflector 10 is a medium to strong amplitude peak. The external form of RB Unit 8 is an onlap-

ping and migrating wave basin fill. The internal pattern shows one reflector changing to three subparallel reflectors in the SE but changes to continuous subparallel reflectors in the NW all with low to medium amplitude and frequency. The unit is only found to the SE where lines 205-207 intersect line 107 (Fig. 4).

RB Unit 9 is at the lower boundary limited by reflector 10 and at the upper boundary by reflector 11. Reflector 11 is a high amplitude trough that in the NW on line 107 (Fig. 4) onlaps reflector 9. The external form of RB Unit 9 is an onlapping and migrating wave basin fill. The internal pattern shows one to two continuous subparallel reflectors in the SE changing to abundant reflectors in the NW of medium to high amplitude and low to high frequency. The unit is present in the RB apart from some minor areas.

RB Unit 10

RB Unit 10 is at the lower boundary limited by reflector 11 and at the upper boundary by reflector 12. Reflector 12 is a high amplitude trough. The external form of RB Unit 10 is a downlapping and migrating wave basin fill. The internal pattern shows continuing subparallel downlapping reflectors of medium to high amplitude and low to high frequency. The unit is present in the RB apart from an area with intersecting lines 205-207 with line 106 (Fig. 1).

RB Unit 11

RB Unit 11 is at the lower boundary limited by reflector 12 and at the upper boundary by reflector Sea Bed (Fig. 4). The Sea Bed reflector is a high amplitude peak. The external form of RB Unit 11 is a basin fill. The internal pattern shows one or two continuous subparallel downlapping reflectors of medium to high amplitude and low to high frequency. The unit is present all over the RB.

Interpretations of the Seismic Units

Early Eocene

RB Unit 1 may originate from the WTR and other surrounded elevated basaltic areas and has been exposed to compression (Fig. 4). The relative sea level fell after the volcanic activity ceased to allow for erosion of the flood basalt of the WTR and other surrounded elevated areas indicating the first uplift of the YR.

A NW/SE trending normal fault is located along the northern flank of the Darwin Igneous Centre (DIC) (Fig. 8) and RB Unit 1 is thicker to the SW than to the NE of the fault. The unit is missing in the area between the fault and YR on line 204 and on line 106 in the area from the fault to midway between lines 204 and 205. SE of line 204 the normal fault is outside the seismic survey. Located between the normal fault and the YR, a small scale foreland compressional belt is present (Fig. 8). The belt dies out or is located outside the survey on line 207 where the YR is widest. This compressional foreland belt is much smaller in size to the NW than to the SE and indicates that the two volcanic centres may have absorbed much of the compression.

RB Unit 2 may originate from the WTR and other surrounded elevated basaltic areas (Fig. 4). The unit has been subjected to compression. The relative sea level rose, but in the SE the relative sea level fell likely due to the compression that elevated the southern part of the YR thus indicating that the YR is a segmented ridge.

During deposition of RB Unit 2, deposition did not occur in two areas. The first area is described above in RB Unit 1. The second area is to the SE and is located on line 106 and 107 at the apex of YR approximately between line 206 and 207 (Fig. 8). On line 107 (Fig. 4) the unit is missing just NW of the apex indicating that the YR was exposed to a compression from the north, northeast, east or southeast resulting in a rotation of the apex towards the south, southwest, west or northwest. On line 106 the unit is missing on top of the YR and on top of the compressional foreland belt located alongside the YR further to the

NW. On the intersecting line 207 the compressional foreland belt is located outside the survey but the RB Unit 2 is missing at the apex of the YR. Rotation seems not to have taken place at the location of line 207 (Fig. 4) as reflector 4 is onlapping the reflector 2 approaching the area from the RB in the SW but downlapping the reflector 2 coming from the AB in the NE. However, this area without deposition, was the apex of the YR during deposition of the RB Unit 2 and therefore, the reflector had to be deposited as an onlapping reflector. By a rotation an onlapping reflector changes into a downlapping reflector in the direction of the compression being the north, northeast or east. Where line 206 (Fig. 6) intersects the YR perpendicularly, the apex has turned towards the SW (Fig. 7) indicating that the compression originated from the NE. This NE compression is also confirmed by the flanks of the compressional belt indicating that the NE-SW direction is parallel to the compression direction. Therefore, this could correspond to the first compressional phase described as a ridge push from the north in Late Paleocene – Early Eocene (Boldreel and Andersen, 1994) although the compression direction differs slightly.

It is suggested that a transcurrent fault being an extension of the Ymir/Ness Lineament is located underneath and alongside the YR (Fig. 8) causing the ridge to wiggle about the transcurrent fault as if it was a hinge. Therefore, the upper boundary of RB Unit 1 downlaps the Top Basalt reflector coming from the RB in the SW, and reflector 4 downlaps reflector 2 coming from the AB in the NE. This indicates that after deposition of RB Unit 1 the apex of the YR was pushed towards the NE and after deposition of RB Unit 2 the apex of the YR was pushed towards the SW. Also the compression that originated from the SW had to be stronger than the compression originated from the NE as reflector 2 retained being a downlapping reflector.

The area SW of the normal fault and between the two volcanic centres (Fig. 8) subsided more than the area to the NW of the normal fault. This might be due to the location of one or several fis-

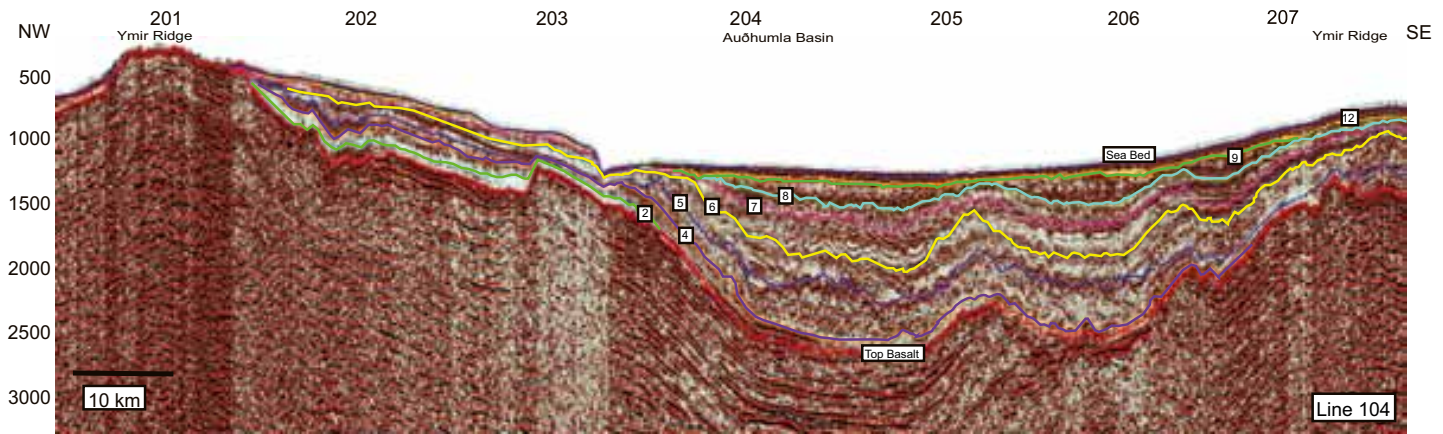


Fig.5.

The key profile chosen in the AB is the WNW-ESE seismic line 104 approximately 152 km long with 9 reflectors interpreted and 9 units analysed. The depth is ms TWT and the crossing seismic lines are shown. The bending YR is situated to the NW and the SE with the AB placed in between. In the NW an escarpment is located at the Top Basalt and between lines 205-207 two larger bulges are seen interpreted as the Drekaeyga Intrusion (DI). The sedimentary units up until reflector 9 follow the contours of the Top Basalt, and overall the sedimentary units onlap in the NW but overstep the YR in the SE. Units deposited after reflector 9 comprises sinusoidal sediment waves and contourite deposits. To the NW an erosional channel is present where sediments were eroded down into the Top Eocene unit by the Glacial Unconformity.

tures as well as lava flows originating from the DIC spreading out and reaching the YR. Indications that the normal fault released much of the compressional tensions are seen.

RB Unit 3 may originate from the WTR and surrounded elevated areas. The normal fault mentioned previously in RB Unit 2 was active during the deposition, as RB Unit 3 is slightly thicker SW of the normal fault than to the NE (Fig. 4). From line 205 towards the NW the RB Unit 3 is thicker and reaches slightly further up the YR than in the RB Unit 1 and RB Unit 2 (Fig. 4), indicating an area of subsidence that caused a relative sea level rise.

Middle Eocene

RB Unit 4 has the widest distribution among the units interpreted. The depositional material may originate from the WTR and surrounded elevated areas. The normal fault described in RB Unit 2 seems to be a strike-slip fault that ends as a listric fault in the NW (Fig. 8). During deposition of RB Unit 4 the strike-slip fault was active and the area was subjected to subsidence but not compression that caused a relative sea level rise.

After deposition of RB Unit 4 the Top Eocene unconformity occurred and eroded the entire apex of the YR due to a pronounced fall in the relative sea level and indicates the second uplift of the YR. A moat present alongside the YR is caused by erosive sea currents which may have been active at a later stage.

Early Oligocene

RB Unit 5 may originate from the WTR and surrounded elevated areas. During deposition of RB Unit 5, lack of deposition occur in the same area as described in RB Unit 1 and at the top of the dome on line 107 (Fig. 4) where the RB Unit 5 onlaps reflector 6. Thickness analyses show that the strike-slip fault was active at a time period that corresponds to about three quarters of the thickness of RB Unit 5. The lows created by the compression and erosion at the Top Eocene unconformity were infilled as the accommodation space was enhanced compared to RB Unit 4 indicating a relative sea level rise due to basin subsidence.

RB Unit 6 seems to originate from the WTR and surrounded elevated areas. Only at the top of

the YR in the SE deposits lack (Fig. 4). The RB Unit 6 oversteps the previously mentioned dome and reaches higher up the SIC and YR indicating a relative sea level rise due to basin subsidence during the deposition.

After deposition of RB Unit 6 the upper boundary was eroded due to compression, indicated by the Top Oligocene unconformity and the third uplift of the YR. In the SE part of line 107 (Fig. 4) between line 206 and 207 the apex of the YR rotated once again towards the SW and the same processes may be active as described previously in RB Unit 2. This may correspond to the second compressional phase described as a ridge push from the north-north-west in Oligocene (Boldreel and Andersen, 1994) although the compressional direction is not the same.

Early Miocene

RB Unit 7 may originate from the WTR and surrounded elevated areas. The RB Unit 7 onlaps the dome previously mentioned and the SIC western flank higher up than in RB Unit 6 (Fig. 4) and thus indicate a relative sea level rise due to basin subsidence during the deposition.

After deposition of RB Unit 7 erosion occurred and truncated the upper boundary NW of the dome due to compression and indicates the fourth uplift of the YR and a relative sea level fall. The sediments are eroded by sea bottom currents that transported and deposited the sediments as contourites and sediment waves towards the YR and the moat mentioned previously in RB Unit 4. This is the Middle Miocene unconformity and could correspond to the third compressional phase described as a ridge push from the NW in Middle – Late Miocene (Boldreel and Andersen, 1994 and Andersen *et al.*, 2002).

Late Miocene – Early Pliocene

RB Unit 8 is deposited by sea bottom currents that eroded the RB Unit 7 and transported the sediments adjacent the SIC and the YR (Fig. 4) as sinusoidal sediment waves or contourite deposits filling up the moat. During deposition of RB Unit 8 the relative sea level rose due to basin subsidence but the same area as described in RB Unit 1

is left without deposition.

RB Unit 9 is deposited by sea bottom currents and transported the sediments adjacent the SIC and the YR as sinusoidal sediment waves or contourite deposits filling up the moat (Fig. 4). During deposition of the RB Unit 9 the relative sea level rose due to basin subsidence.

After the contourite deposition the sea bottom currents changed their direction or strength and a new moat developed at the flank of the YR in the NW on line 201 and 202.

Middle Pliocene – Pleistocene

RB Unit 10 seems to originate from the previously deposited units, YR, WTR and other surrounded elevated areas. The RB Unit 10 is deposited by sea bottom currents that transported the sediments adjacent the SIC and the moat at the flank of the YR as sinusoidal sediment waves or contourite deposits filling up the moats (Fig. 4). During deposition of RB Unit 10 the relative sea level rose due to basin subsidence.

Pleistocene

RB Unit 11 may originate from the WTR and other surrounded elevated areas. During deposition of RB Unit 11 the relative sea level rose due to basin subsidence.

Auðhumla Basin - Observations of the Seismic Units

In the Auðhumla Basin the seismic line 104 (Fig. 5) was chosen as the key profile.

AB Unit 1 – AB Unit 3

AB Unit 1 is at the lower boundary limited by reflector Top Basalt and at the upper boundary by reflector 2. Reflector 2 is a low to high amplitude trough and oversteps the YR in the NW on line 202 and on line 206 (Fig. 6) and line 207 into the RB. Reflector 2 onlaps reflector Top Basalt at the YR in the NW, but downlaps reflector Top Basalt between line 203 and 204 (Fig. 5). The external

form of AB Unit 1 is a basin fill. The internal pattern shows onlapping subparallel continuous reflectors in the NW with high amplitude and frequency that changes into chaotic to contorted reflectors with lower amplitude and frequency towards the SE. The unit is present in the entire AB.

AB Unit 2 is at the lower boundary limited by reflector 2 and at the upper boundary by reflector 4. Reflector 4 is a high amplitude trough, that onlaps reflector 2 in the NW and reflector Top Basalt in the SE. The external form of AB Unit 2 is a basin fill. The internal pattern shows onlapping subparallel continuous reflectors that changes into chaotic to contorted reflectors towards the SE of low to high amplitude and frequency. The unit is present in the entire AB.

AB Unit 3 is at the lower boundary limited by reflector 4 and at the upper boundary by reflector 5. Reflector 5 is a low to medium amplitude undulating peak. The external form of AB Unit 3 is a basin fill. The internal pattern consists of onlapping subparallel reflectors that changes into chaotic to contorted reflectors towards the SE of low to high amplitude and frequency. In the NW on line 203 a moat is present. The unit is present all over the AB.

AB Unit 4

AB Unit 4 is at the lower boundary bounded by reflector 5 and at the upper boundary by reflector 6. Reflector 6 is a low to medium amplitude undulating peak. The external form of AB Unit 4 is a basin fill. The internal pattern shows onlapping subparallel reflectors of low to high amplitude and frequency that changes into chaotic to contorted reflectors towards the SE. Truncations are found at the upper boundary and an erosional channel is observed on line 104 (Fig. 5) between line 203 and 204. The unit is present in the AB except in one minor area.

AB Unit 5 – AB Unit 6

AB Unit 5 is at the lower boundary limited by reflector 6 and at the upper boundary by reflector 7. Reflector 7 is a low to medium amplitude undulating peak.

The external form of AB Unit 5 is a basin fill. The internal pattern shows onlapping subparallel and chaotic to contorted reflectors of medium to high amplitude and frequency. Truncations are found and the erosional channel is observed on line 104 (Fig. 5). The unit is present in the AB apart from two minor patches.

AB Unit 6 is at the lower boundary limited by reflector 7 and at the upper boundary by reflector 8. Reflector 8 is a medium amplitude undulating peak. The external form of AB Unit 6 is a basin fill. The internal pattern shows onlapping subparallel and chaotic to contorted reflectors of medium to high amplitude and frequency. Truncations are found. The unit is found in the central and SE part of the AB.

AB Unit 7

AB Unit 7 is at the lower boundary limited by reflector 8 and at the upper boundary by reflector 9. Reflector 9 is a medium to high amplitude peak. The external form of AB Unit 7 is a basin fill. The internal onlapping subparallel and chaotic to contorted reflectors of pattern shows medium to high amplitude and frequency. Truncations are found. The unit is found in the central and SE part of the AB.

AB Unit 8

AB Unit 8 is at the lower boundary limited by reflector 9 and at the upper boundary by reflector 12. Reflector 12 is a high amplitude trough that downlaps the erosional channel and the moat in the NW (Fig. 4). The external form of AB Unit 8 is a basin fill. The internal pattern shows one to three reflectors of high amplitude and frequency. Truncations are found. The unit is found in the entire AB.

AB Unit 9

AB Unit 9 is at the lower boundary limited by reflector 12 and at the upper boundary by reflector Sea Bed. Reflector Sea Bed is a high amplitude peak. The external form of AB Unit 9 is a basin fill. The internal pattern shows infilling and onlapping subparallel reflectors of high

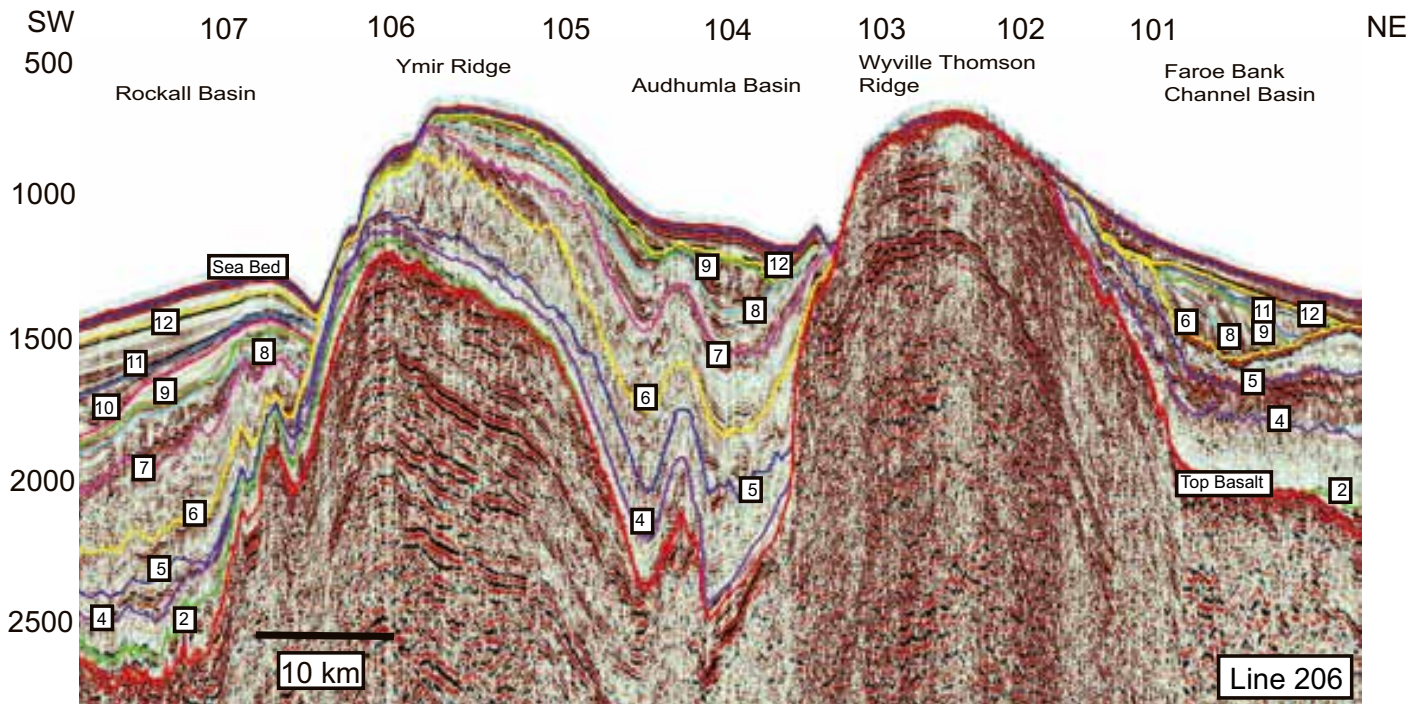


Fig. 6.

The key profile chosen for connecting the three sub areas is the NNE-SSW seismic line 206 that is approximately 103 km long where 14 reflectors were interpreted and 16 units analysed. The Depth in ms TWT and the crossing seismic lines are shown. The FBCB is different from the RB and AB and characterized by rather calm depositional conditions and was hardly affected by compressional forces. Approaching the study area from the south (RB) the four lower most units overstep the YR and continues into the AB whereas the WTR acts as a barrier and no units are directly connected to the FBCB. In the FBCB the seismic picture of the four lower most and four upper most units look similar to the four lowermost and four uppermost units in the RB. The top of the WTR show truncations and compressional structures are found in the RB and AB being active until Middle Miocene. The following pelagic units were deposited by sea bottom currents that show onlapping and migrating or sinusoidal sediment waves and contourite deposits.

amplitude and frequency. The unit is found in the entire AB.

Interpretations of the Seismic Units

Early Eocene

AB Unit 1 onlaps the Top Basalt reflector and has a limited extension in the NW and continues like a tongue along the south western side of the basin. This indicates that a small amount of likely clastic depositional material that may originate from the WTR and surrounded elevated areas was available. The relative sea level fell to allow for erosion of the flood basalt and indicates the first ridge uplift.

The deposition of AB Unit 1 indicates that the

compression came from the NE, the basin was at a higher level in the NE and the compressional belt began to develop before or during the deposition of AB Unit 1. The area to the NW of the Drekaeyga Intrusion (DI) was more exposed and responsive to compression than the area SE of the DI. The deposition area of AB Unit 1 included the YR and continued into the RB. However the unit was removed by later erosion exposing the Top Basalt. The WTR was more elevated than the YR.

AB Unit 2 seems to originate from the WTR and surrounded elevated areas. It is interpreted that the AB Unit 2 constitutes of clastic materials. During deposition of AB Unit 2 the relative sea level rose causing an enhanced accommodation space due to basin subsidence.

The deposition of AB Unit 2 indicates that both

sides of the basin were at the same elevation, the basin subsided towards the SE or the area tilted towards the SE. The downlap of AB Unit 2 on reflector 2 on line 207 adjacent to line 106 (See trend on Fig. 6 and Fig. 8) indicates a compression from the NE or that the YR is divided into sections by faults or fissures.

AB Unit 3 could originate from the WTR and surrounded elevated areas. The depositional distribution is similar to AB Unit 1 and signifies that YR was not highly elevated. Compared to AB Unit 2 (Fig. 5) the accommodation space was slightly enhanced in the NE and enlarging towards the SW due to basin subsidence which indicates a relative sea level rise.

After deposition of AB Unit 3 compression commenced causing the YR and all the above deposited sedimentary units to bend upwards indicating the second ridge uplift (Fig. 6). In the NW towards the SE until line 205, erosion exposed the Top Basalt reflector indicating that this part of YR is more responsive to compression than in the SE (Fig. 8). On line 207 adjacent to the intersecting of line 106 the AB Unit 3 downlaps reflector 2 towards the SW, indicating a compression from the NE and causing a SW rotation of the YR apex. This compression may correspond to the first compressional phase described as a ridge push from the north in Late Paleocene – Early Eocene (Boldreel and Andersen, 1994). The relative sea level fell during the compression and allowed erosion of the newly uplifted areas, and thus AB Unit 3 was removed above YR.

Middle Eocene

AB Unit 4 seems to originate from the WTR and surrounded elevated areas. The AB Unit 4 has the widest distribution (Fig. 5) and during the deposition the relative sea level rose enhancing the accommodation space due to basin subsidence.

After deposition of the AB Unit 4 the Top Eocene unconformity commenced causing a pronounced relative sea level fall and indicates the third uplift of the ridges. In shallow areas truncations are found and an erosional channel is observed (Fig. 5). On lines 205-207 (Fig. 8) a

precursor to the moat located adjacent to line 106 developed due to erosive sea currents. Originally AB Unit 4 was deposited above the YR and continued into the RB but later erosion removed the AB Unit 4 from line 205 towards the NW. This indicates that the first compressional phase continued until the Top Eocene or that the second compressional phase commenced from the NE.

Early Oligocene

AB Unit 5 may originate from the WTR and surrounded elevated areas. At the WTR southern flank AB Unit 5 onlaps the lower reflector except in the interval from line 206 and line 207 (Fig. 8) where AB Unit 5 onlaps reflector Top Basalt. This indicates a relative sea level rise during the deposition due to basin subsidence or sediment compaction enhancing the accommodation space.

After deposition of AB Unit 5 erosion occurred and truncated the upper boundary (Fig. 5). There are no signs of compression and thus the erosion is caused by the accommodation space being filled up.

AB Unit 6 could originate from the WTR and surrounded elevated areas. The AB Unit 6 is present to the SE of line 203 and onlaps reflector 7 before the erosional channel and the moat (Fig. 5). There are indications of a relative sea level rise towards the SE due to basin subsidence or sediment compaction during deposition of AB Unit 6.

The lack of deposition of AB Unit 6 in the NW is due to erosion caused by the Top Oligocene unconformity caused by compression that elevated the area in the NW signifying the fourth ridge uplift and a relative sea level fall. This compression indicates the third compressional phase which may correspond to the second compressional phase described as a ridge push from the north-north-west in Oligocene (Boldreel and Andersen, 1994). This implies that the ridge is divided into segments where the segment SE of line 203 experienced basin subsidence or sediment compaction.

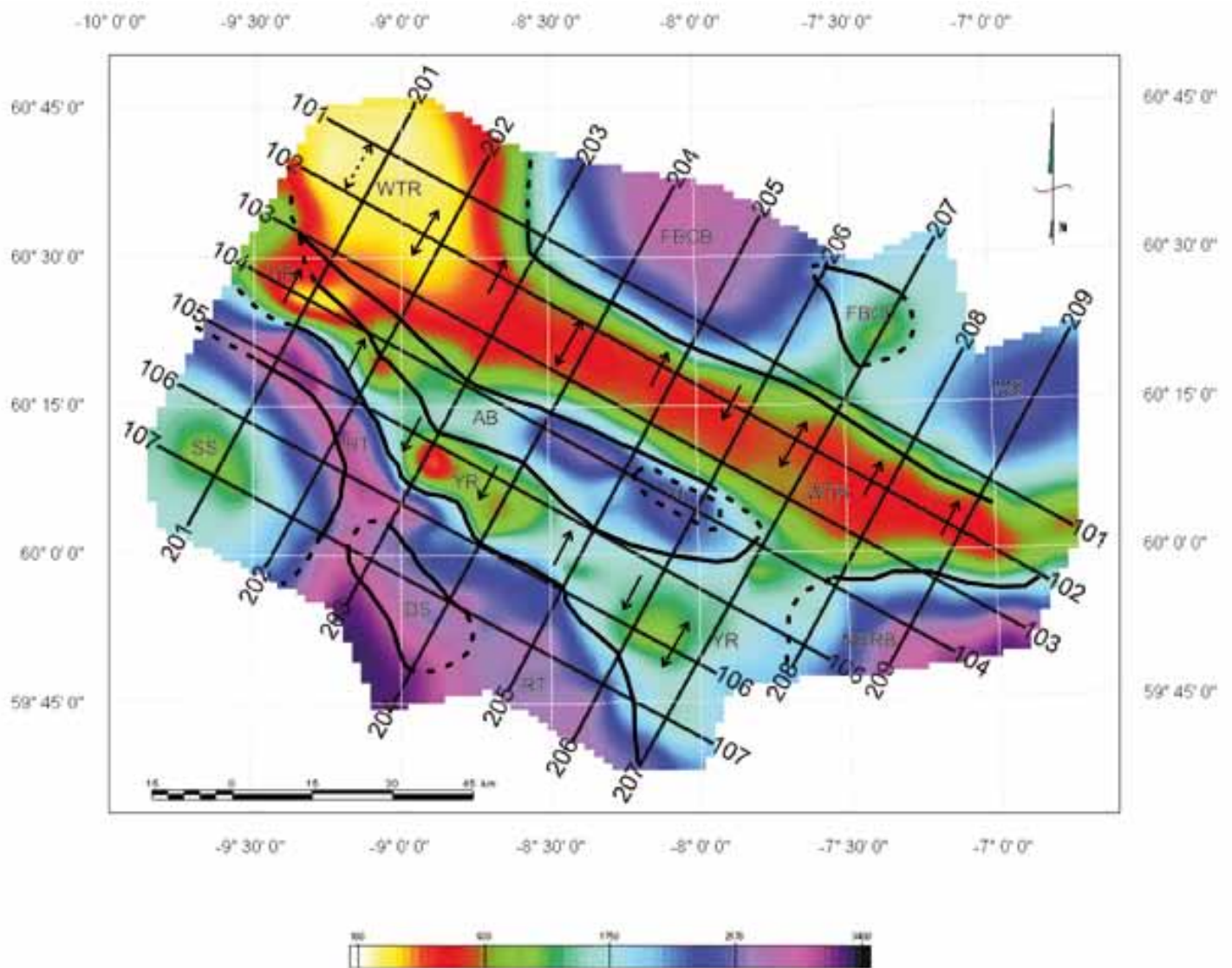


Fig. 7.

An isopach map of the interpreted Top Basalt surface to the seafloor that outlines the interpreted geological structures. Solid lines are interpreted from the YMR97 survey and dotted lines are tentative interpretations. Arrows indicate the present apex facing directions of the WTR and YR being the resultant compressional force affecting the WTRC. Abbreviations: AB, Auðhumla Basin; DIC/DS Darwin Igneous Center; DI, Drekaeyga Intrusion; FBCB, Faroe Bank Channel Basin; FBCN, Faroe Bank Channel Knoll; MB, Munkur Basin; NERB, Northeast Rockall Basin; RB, Rockall Basin; SIC/SS, Sigmundur Igneous Centre; WTR, Wyville Thomson Ridge; YR, Ymir Ridge.

Early Miocene

AB Unit 7 seems to originate from the WTR and surrounded elevated areas. The distribution of AB Unit 7 indicates that during the deposition the accommodation space enhanced more in the SE than in the NW (Fig. 5) due to basin subsidence or sediment compaction resulting in a relative sea level rise.

After deposition of AB Unit 7 the Middle

Miocene unconformity commenced and the sea currents eroded the upper part of the AB Unit 7 (Fig. 5). A massive compression from the NE rotated the apex of the YR and all the above deposited units towards the SW (Fig. 6) and indicates the fifth uplift of the YR resulting in a relative sea level fall. This compression signifies the fourth compressional phase which may correspond to the third compressional phase (Boldreel

and Andersen, 1994) and the second compressional phase by Andersen *et al.* (2002).

Late Miocene – Early Pliocene or Middle Pliocene – Pleistocene

AB Unit 8 may originate from the WTR and surrounded elevated areas and is deposited during a relative sea level rise due to basin subsidence.

After deposition of AB Unit 8 the Glacial Unconformity commenced and the relative sea level fell. The previously mentioned erosional channel on line 104 (Fig. 5) began to develop at the Glacial Unconformity and eroded down into reflector 6 (AB Unit 4) being the NE part of the moat located on line 203.

Pleistocene

AB Unit 9 may originate from the WTR and surrounded elevated areas and is deposited during a relative sea level rise due to basin subsidence.

Summarizing the three Sub Areas

Discussion/Conclusion

The three sub areas are connected using seismic line 206 (Fig. 6). In the text the units are to be found in Figures 3, 4 and 5.

Early Eocene

The Early Eocene comprises five units (Fig. 2): unit 1 (Top Basalt – reflector 1) located in the FBCB, unit 2 (Top Basalt – reflector 2) found in all three basins, unit 3 (between reflectors 2 - 3) present in the FBCB, unit 4 (between reflectors 2 - 4) and unit 5 (between reflectors 4 - 5) which both are found in all three basins.

The WTR started to elevate during the Early Eocene and the units deposited in the FBCB did not continue into the AB. The YR was not highly elevated and therefore the units are found in both the AB and the RB. Overall the WTRC tilted towards the SE and the ridges are segmented by faults having a NE-SW direction bearing the Caledonian trend or by fissures that connect the volcanic centres (Fig. 8). The volcanic centres, the

extension of the lava flows and the transcurrent fault, which is the north western part of the Ymir/Ness Lineament, located underneath and alongside the YR controlled much of the depositional distribution and the localization of the compression. Therefore the effect of the compression is more intense in the NW of the AB and in the SE in the RB. During Early Eocene the YR started to elevate and the WTR continued to elevate due to compressional forces having a NE-SW direction. The first compression originated from the NE (during deposition of unit 1), the second from the SW (taken place during deposition of unit 2) and the third also from the NE (during deposition of unit 3, 4 and 5).

The compression from the SW may be ascribed to the seafloor spreading between Greenland and Eurasia linked at a triple junction south of Greenland (Lundin and Doré, 2005) (Fig. 1) that initiated in Paleocene (57-56 Ma Chron 24/25) (Larsen, 1988) or (c.56-53 Ma) (Saunders *et al.*, 1997). However, the oceanic spreading between the Faroe Fracture Zone (Bott, 1985) and the Jan Mayen Fracture Zone transferring to the NW (Fig. 1) isolating the Jan Mayen microcontinent (e.g. Nunns, 1983) also had an affect on the WTRC. Many suggestions exist on the timing of the Aegir Ridge transition and judged from the internal reflector pattern outlined in this study the period from the initiation of break-up in Paleocene – Early Oligocene (Chron 24 c.54 Ma until Chron 12 c.32 Ma) when Aegir Ridge was abandoned (Lundin and Doré, 2005) is suggested. This ridge transition is manifested on the Faroe-Rockall Plateau (Boldreel and Andersen, 1993) and is the compressional force from the NE observed in the WTRC. A clockwise rotation of the WTRC is suggested by the opposed compressional directions and the different facing directions of the apex of the ridges (Fig. 7), and the strength of the affect depends on the width of the ridges. This could correspond to the first compressional phase described as a ridge push from the north in Late Paleocene – Early Eocene by Boldreel and Andersen (1994) and as the first compressional phase in the Middle Eocene by Andersen *et al.*

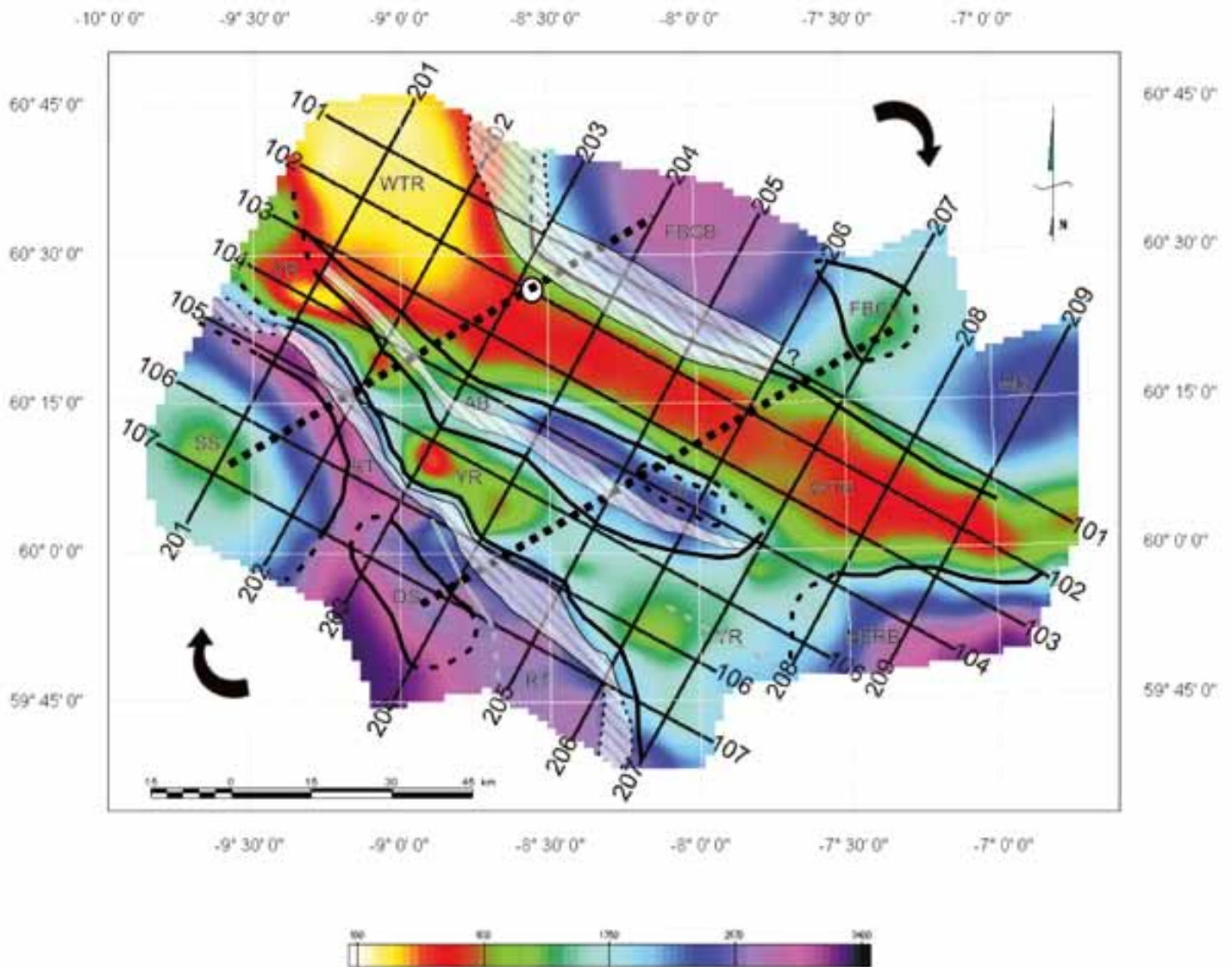


Fig. 8.

A suggested structural model of the WTRC as judged from the seismic interpretation of the seismic survey YMR97. Solid and dotted grey lines are parts of the transcurrent faults Ymir/Ness Lineament. Shaded areas show the compressional belts. Question mark relates to the shaded area in the FBCB. White dot indicates a small basin located at the Top Basalt on the WTR. Heavy dotted black lines show fissures and adjacent faults. Solid black arrows indicate clockwise rotation of the WTRC.

(2002). The Thulean Volcanic Line (Hall, 1981) (Fig. 1) located south of the WTRC, indicates a NE-SW extensional stress field (England, 1988) in the Middle Paleocene (c. 62-58 Ma) (Archer et al., 2005) that together with the mafic laccolithic located in the NE Rockall Basin could have affected the subsidence in the SE due to thermal cooling.

Overall the relative sea level rose towards the SE and enhanced the accommodation space but fell in the NW allowing erosion of the uplifted

areas that removed all or parts of the units at the YR southern flank.

Middle Eocene

The Middle Eocene is comprised of one unit (Fig. 2): unit 6 (between reflectors 5 - 6) deposited in all three basins and has the widest geographical distribution of all the mapped units.

The relative sea level rose during the deposition of the unit caused by basin subsidence that enhanced the accommodation space. The

transcurrent fault in the RB that branches out as a listric fault in the NW direction was active.

After deposition of the unit the relative sea level fell and the Top Eocene unconformity evolved which caused a sedimentation stop and erosion of the upper boundary. This indicates that the first compressional phase continued or that the second compressional phase started from the NW. This compression may be due to the continuous spreading axis between the Arctic and the NE Atlantic in the Late Eocene (Chron 13) (Lundin and Doré, 2005) that initiated after the ocean spreading in the Labrador Sea slowed and stopped in Middle – Late Eocene (c. 50 Ma until 36 Ma Chrons 21-13) (Saunders *et al.*, 1997). Also the rotation of the Jan Mayen microcontinent manifested on the Faroe-Rockall Plateau and the WTRC contributed to the development of the Top Eocene unconformity (e.g. Johnson *et al.*, 2005 and Stoker *et al.*, 2005). These tectonic movements caused a further clockwise rotation of the WTRC.

Early Oligocene

The Early Oligocene is comprised of three units (Fig. 2): unit 7 (between reflectors 6 – 7) and unit 8 (between reflectors 7 – 8) deposited in the RB and the AB, and unit 9 (between reflectors 6 – 8) deposited in the FBCB.

The lows created at the Top Eocene unconformity became filled as the accommodation space enhanced due to basin subsidence or sediment compaction indicating a relative sea level rise. The transcurrent fault in the RB (Fig. 8) was active at the beginning of the period.

After deposition of the units the Top Oligocene unconformity commenced and the upper boundary was eroded which indicates a relative sea level fall caused by compression. The north western area of the ridges was elevated but the south eastern area subsided signifying that the ridges are segmented or that the clockwise rotation of the WTRC was reactivated (Fig. 8). In the NW of the AB and at the YR southern flank the entire unit was removed. This is caused by the third compressional phase which could correspond to the

second compressional phase described as a ridge push from the north-north-west in Oligocene by Boldreel and Andersen (1994).

Early Miocene

The Early Miocene is comprised of one unit (Fig. 2): unit 10 (between reflectors 8 – 9) deposited in all three basins.

During deposition of the unit the accommodation space enhanced due to basin subsidence or sediment compaction that indicates a relative sea level rise, but in the AB the unit only exists SE of the erosional channel and the moat (Fig. 5).

After deposition of the unit the Middle Miocene unconformity commenced eroding and truncating most of the upper boundary by sea bottom currents in the AB and RB and transported the sediments as contourites and sinusoidal sediment waves (Fig. 7 and 4). This signifies the fourth compressional phase which could correspond to the third compressional phase described as a ridge push from the north-west in Middle – Late Miocene (Boldreel and Andersen, 1994) and as the second compressional phase by Andersen *et al.* (2002). This last compressional phase originating from the NW may be caused by the remnants of the structural adjustment of the rotation of the Jan Mayen microcontinent. The effect of the compressional force is that the apex of the YR and all the above deposited units turned towards the SW (Fig. 7) and indicates that the ridges are segmented and the relative sea level fell due to uplift of the ridges.

Late Miocene – Early Pliocene

The Late Miocene – Early Pliocene comprises three units (Fig. 2): unit 11 (between reflectors 9 - 10) and unit 12 (between reflectors 10 - 11) deposited in the RB, and unit 13 (between reflectors 9 - 11) deposited in the FBCB.

In the FBCB deposition took place during calm conditions, as compared to the RB, and the relative sea level was about the same as in the previous period or rose slightly enhancing the accommodation space due to basin subsidence (Fig.3). In the RB (Fig. 4) the relative sea level

rose due to basin subsidence and sea bottom currents that eroded the lower units and transported the sediments adjacent to the escarpment and flank of the YR as sinusoidal sediment waves or contourite deposits filling up the moat and lows.

After the contourite deposition the Pliocene unconformity commenced where the sea bottom currents changed their direction or strength and developed a new moat at the YR southern flank in the NW.

Middle Pliocene - Pleistocene

The Middle Pliocene – Pleistocene comprises two units (Fig. 2): unit 14 (between reflectors 9 - 12) deposited in the AB, and unit 15 (between reflectors 11 - 12) deposited in the RB and FBCB.

During deposition of the unit the relative sea level rose due to basin subsidence. In the FBCB the unit was deposited under more calm conditions than in the AB and the RB (Fig. 6).

After deposition of the unit erosion took place indicated by the Glacial Unconformity and a relative sea level fall. This signifies that the accommodation space was filled up and the onset of the glaciation. In the RB there are no indications of the Glacial Unconformity (Fig. 4), but in the AB an erosional channel developed (Fig. 5) that eroded down into the Top Eocene period being the NE part of the moat. Therefore, unit 14 (between reflectors 9 - 12) might be part of the Late Miocene – Early Pliocene period as the top of the unit has been severely eroded. Hence the Middle Pliocene – Pleistocene is not present in the AB. In the FBCB the entire upper boundary is eroded and truncated in the NW (Fig. 3).

Pleistocene

The Pleistocene is comprised of one unit (Fig. 2): unit 16 (reflector 12 – Sea Bed) deposited in all three basins. During deposition of the unit the relative sea level rose due to basin subsidence where the depositional conditions overall were more calm than in the two previously periods.

Structural Model of the WTRC

The WTRC constitute of the two NW-SE striking compressional ridges, the WTR and the YR with the intervening AB. To the NW and SE the ridges merge and at these locations the ridges have their largest width. The width of the WTR is rather consistent whereas alongside the YR the width changes (Fig. 7).

The apex of the ridges faces in different directions (Fig. 7) and this is caused by the combined effect of approximately perpendicular directed stress towards the ridges and the location of the volcanic centres thought of as being connected by two ENE/WSW trending fissures (Fig. 8). The western fissure connects the SIC and a small sediment basin on top of the WTR apex whereas the eastern fissure connects the DIC, DI and the FBCK. The fissures divide the investigated area into three parts where the NW segment had a resultant compressional direction towards the NE, the central segment towards the SW and the SE segment towards the NE (Fig. 7). Two facing directions deviate from this: in the central part of the WTR the apex faces towards NE (line 205) and in the SE the apex of the YR faces towards the SW (line 206). This local effect is caused by the width of the ridges and the location of the DI, as it is speculated that the volcanic centres absorbed and stopped much of the compressional forces and thereby locally controlled the development of the ridges. Adjacent to the fissures two NE/SW trending transfer faults that bear the Caledonian trend are interpreted (Fig. 8). In addition a transcurrent fault, which extends from the Ymir/Ness Lineament, is located underneath the south eastern part of the YR that was active until late Early Eocene. In the RB the transcurrent fault is located alongside the southern flank of YR and ends as a listric fault between the DIC and the central part of the YR that was active until Early Oligocene. The WTRC tilted towards the SE due to thermal cooling of the Thulean Volcanic Line located south of the WTRC and the mafic laccolithic in the NE Rockall Basin. Four compressional phases, listed below, affected the complex and caused compressional belts along-

side the flanks of the ridges and a clockwise rotation.

1. Late Paleocene – Early Eocene: the compressional force has a NE-SW direction and constitutes of three smaller compressional forces that overall resulted in a clockwise rotation of the WTRC. The first compressional force originated from the NE may be due to oceanic spreading between the Faroe Fracture Zone and the Jan Mayen Fracture Zone transferring north westwards and isolated the Jan Mayen microcontinent. This ridge transition caused the Jan Mayen microcontinent to rotate anticlockwise. The second compressional force originated from the SW may be due to the seafloor spreading between Greenland and Eurasia. The third compressional force originated from the NE is a continuation of the first compressional force due to the ridge transition that caused the anticlockwise rotation of the Jan Mayen microcontinent.
2. Late Eocene: the compressional force that originated from the NW may be due to the new continuous spreading axis between the Arctic and NE Atlantic and the rotation of the Jan Mayen microcontinent.
3. Late Oligocene: the compressional force that originated from the NW caused a reactivation of the clockwise rotation of the WTRC.
4. Middle Miocene: the compressional force originated from the NW is due to the remnants of the structural adjustment of the anticlockwise rotation of the Jan Mayen microcontinent.

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- Figure 3: WNW-ESE seismic line 101 located in the FBCB.
- Figure 4: WNW-ESE seismic line 107 located in the RB.
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List of Abbreviations

- Auðhumla Basin = AB
 Bill Bailey Bank = BBB
 British Volcanic Province = BVP *Continent-Ocean Boundary* = COB
 Darwin Igneous Centre = DIC Drekaeyga Intrusion = DI
 Faroe Bank Channel Basin = FBCB Faroe Bank Channel Knoll = FBCK
 Faroe Shetland Basin = FSB *Greenland – Faroës Ridge* = GFR
 Large Igneous Province = LIP Munkur Basin = MB
 Munkagrannur Ridge = MR North Atlantic Igneous Province = NAIP
 Northeast Rockall Basin = NERB *Norwegian Sea Deep Water* = NSDW
 Rockall Basin = RB Seaward Dipping Reflectors = SDR
 Sigmundur Igneous Centre = SIC Wyville Thomson Ridge = WTR
 Wyville Thomson Ridge Complex =WTRC Ymir Ridge = YR

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Fracture orientations onshore Faroe Islands (North Atlantic); evidence for dual rifting episodes in the Palaeogene?

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Abstract

Analysis of aerial and satellite photographs with respect to faults, fractures and dykes onshore Faroe Islands has revealed some preferred orientations. Investigations of these have been made in order to gain a better understanding of the tectonic evolution of the Faroese Platform during the Palaeogene.

The project had access to aerial photographs over roughly 70% of the country, and good quality satellite images covered most of the remainder. This enabled a thorough analysis of lineaments that are interpreted to represent faults, fractures and/or dykes. The analysis is performed in a consistent manner in order to avoid bias towards areas covered by the higher quality aerial photographs compared to the Satellite photograph covered areas.

The results show a distinct difference in the dominant trend of the interpreted features (faults, fractures and dykes) in the oldest lava formation (NW/SE-NNW/SSE) compared to the younger formations (E/W). The older formation is exposed on the west side of the Faroe Islands, which is adjacent to a proposed parallel early Palaeocene transient rift. The E/W-ENE/WSW orientation is parallel to minimum stress direction associated with the continental breakup between Greenland and Eurasia.

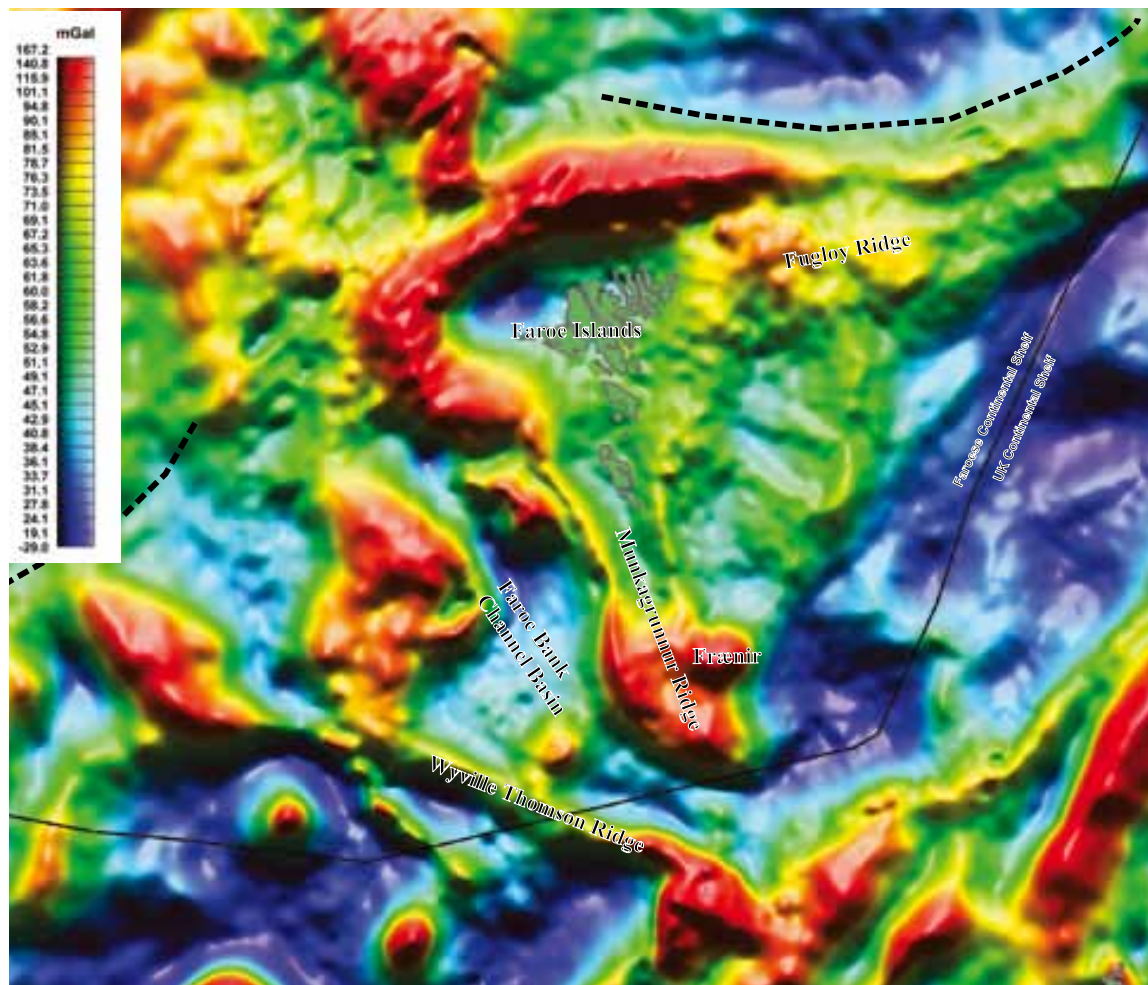
This work does thus show that there were two extensional events in the Palaeogene, and that these were offset both in time and space. The former was early in the Paleocene and located west of the Faroese Platform and had a preferred NW/SE-NNW/SSE orientation. The latter was in the latest Palaeocene to early Eocene, was located north of the Faroese Platform and had a preferred E/W-ENE/WSW orientation.

Introduction

The Faroe Islands (Fig. 1) are located on the European Atlantic Margin between the British Isles and Iceland (Fig. 2). They are just landwards of the breakup line between Greenland and Eurasia, and are wholly covered by flood basalts of the North Atlantic Igneous Province e.g. (Larsen *et al.*, 1999a). The location of the Islands just south

of the final breakup line, and between areas with dominant NW/SE and NE/SW trends respectively (Fig. 1), indicates that the Faroe Islands is a logical place to search for clues regarding the tectonic history of the region during the Palaeogene and Neogene.

The basalts on the Faroe Islands are assigned to the Faroe Islands Basalt Group (Fig. 6), which are

**Fig. 1.**

Gravity data showing dominant trends on the Faroese Continental Shelf.

Dashed line: approximate location of Continent Ocean Transition.

part of the greater North Atlantic Igneous Province (red area on Fig. 2). The Faroe Islands Basalt Group has been subdivided into seven formations (Passey and Jolley, 2009), with four of these being sedimentary in nature (Lopra, Prestfjall, Hvannhagi and Sneis Formations), while the other three formations are basaltic (Beinissvørð, Malinstindur and Enni Formations). The thickness of the sedimentary sections ranges from a few meters to less than a meter, while the lava formations are up to more than a kilometre thick. The oldest formation is the Lopra Formation which is only found in the Lopra-1 well (Fig. 6). It is a hyaloclastic sequence which is most likely a result of lavas flowing into depressions in the pre-volcanic landscape (Ellis *et al.*, 2002). This is overlain by the on average 20-50 m thick simple flows of the

Beinissvørð Formation with a total thickness of 3450 m of which 900 m is exposed and another 2550 m were drilled in the Lopra-1 well (Christie *et al.*, 2006). There was volcanic quiescence after the emplacement of the Beinissvørð Formation. The Prestfjall Formation, which contains coals within a clay-matrix, total thickness being only a few meters, was deposited during a period with no local volcanism. The formation represents, based on Milankovitch cycles, 1.42 Ma of deposition (Mørk, 2007). Renewed volcanism started with the deposition of the Hvannhagi Formation, a volcanic conglomeratic sequence thought to have originated from the explosive re-initiation of volcanism (Rasmussen and Noe-Nygaard, 1969). This later volcanism resulted in the emplacement of the Malinstindur and Enni

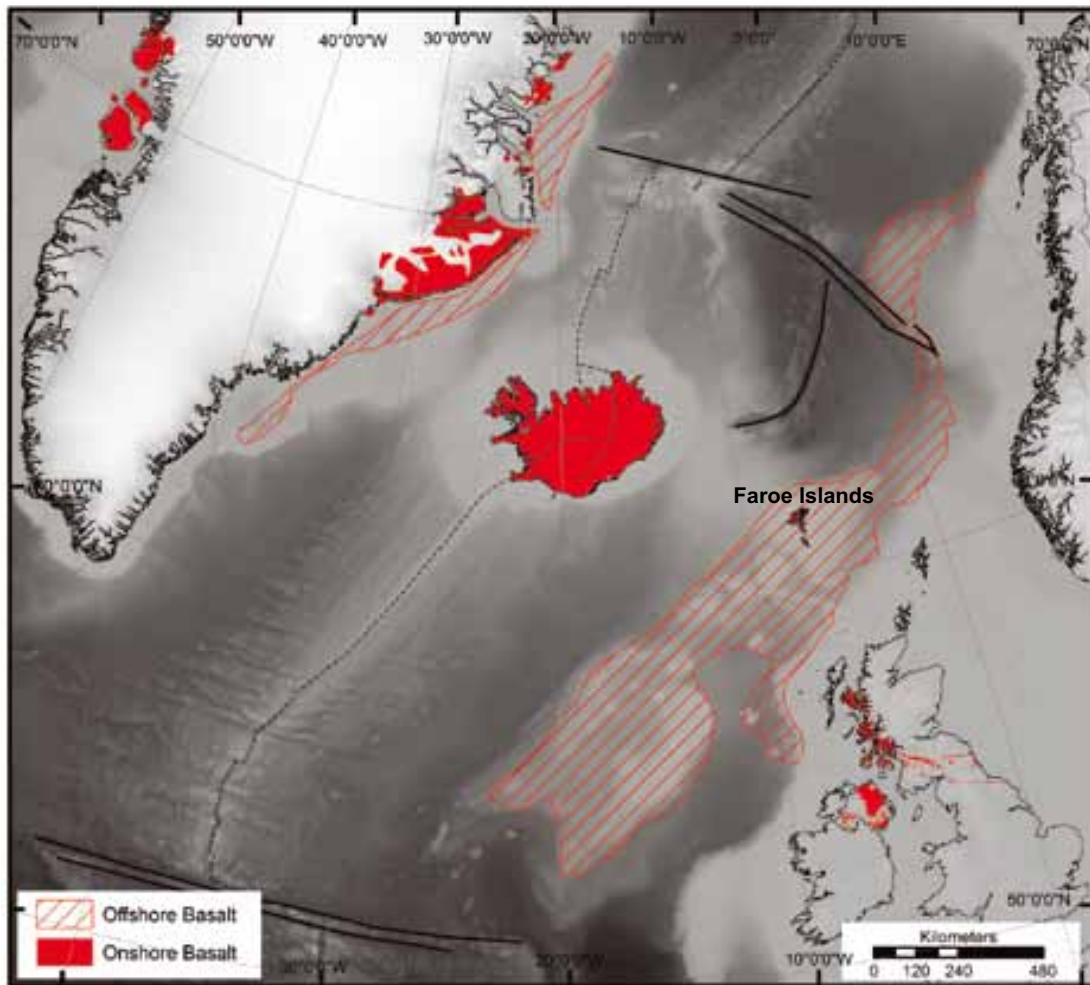


Fig. 2.

North Atlantic igneous Province (red) with bathymetry and topography, adapted from (Ritchie et al., 1999), (Frei et al., 2005) and (Saunders et al., 1997)

Formations, separated by the sedimentary Sneis Formation (Passey and Jolley, 2009).

England (1988) found variations in the stress regime in NW Scotland, that showed changes in the minimal stress axis going from NE/SW in the early Palaeocene to a NW/SE minimal stress direction during the Palaeogene. Dewey (2002) proposed rapid NE/SW stretching in a corridor from SE Greenland, through Faroes to NW Scotland, which was followed by extension and subsequent continental breakup along an E/W to NE/SW line north and west of the Faroese Platform.

I would like to test the hypothesis that this change in stress regime is also represented on the Faroe Islands. This will be done through the analysis of preferred fracture orientations as

seen on aerial and satellite images of the islands. The assumption is that fractures and dykes will align themselves at right angles to the maximal tensional stress direction as documented in the literature e.g. (England, 1988; Gudmundsson, 2002; Nakamura, 1977). The ages and hence period of the emplacement of the Beinissvørð Formation is disputed, with palynological ages suggesting about one million years, while radiometric dating suggest a longer period of volcanism (Jolley, 2009). The Malinstindur and Enni Formation lavas were emplaced possibly in less than one million years (Jolley, 2009).

Mapping on the Faroe Islands e.g. (Rasmussen and Noe-Nygaard, 1969) indicates a difference in orientation of dykes on Suðuroy compared to the islands further north. This coincides with

the exposure of the oldest lava formation vs. the younger lava formations. A likely change in stress regime is thus most likely found in differences in fracture and dyke orientations between the older lava formations (Beinisvørð Formation) and the younger lava formations (Malinstindur and Enni Formations), where the only significant volcanic hiatus, represented by the Prestfjall and Hvannhagi Formations, is located.

The results of a detailed analysis of dyke and fracture orientations on aerial and satellite photographs are discussed in relation to the tectonic and volcanic evolution of the Faroese Continental Shelf during the Palaeogene. A brief discussion of the implications for hydrocarbon exploration will also be included.

Analysis

The work in this study is primarily based on analysis of aerial photographs covering most of the islands towards the north, and satellite photographs covering the southernmost island and the two islands furthest to the northeast (Fig. 3). The quality of the photos enables detailed analysis of features in the landscape down to a few meters (Fig. 4). Difference in colour of the aerial and satellite photograph is mostly due to the images being taken at different time of year. The resolution of the satellite images and aerial photographs is comparable, and both have been included with the same weight in this project.

The aerial and satellite photographs were loaded into a GIS environment, where they were geo-referenced. The overall coverage can be seen in Figure 3.

There are many linear features in the landscape, some of which do not relate to the primary target for this work, which are fractures zones, faults and dykes. These features are thus potential pitfalls when tracing tectonic lineaments on the aerial photographs. These include: drainage patterns river beds, lava benches forming „stairs“ in the landscape and man made features like footpaths, fences etc. (Fig. 4b). The drainage patterns are the biggest challenge as these do exploit frac-

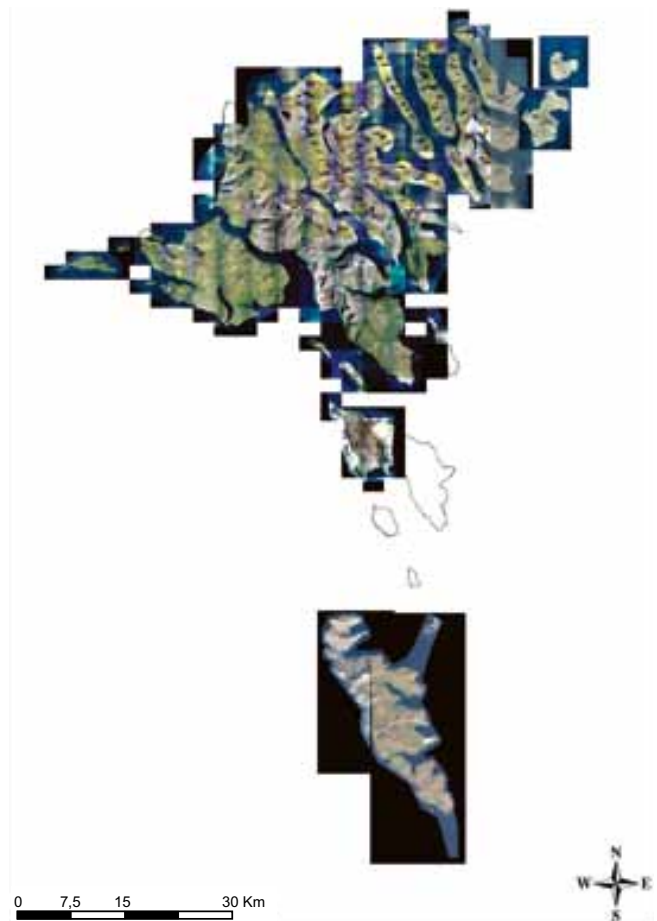


Fig. 3.
Data coverage.

tures in some locations, while they cut across features in other places. The drainage patterns / fractures challenge was dealt with on a case by case basis, where issues such as angle relative to maximum dip, straight sections etc were used to discriminate between the recent erosive feature and the possibly underlying fractures.

Linear features that are interpreted to represent faults, dykes, fractures or fracture zones were then traced on the images. Examples of the visual appearance of these in the field are shown in Figure 4.

Aerial and satellite photographs do not permit distinction between faults, fractures, dykes or dyke intruded fracture zones, and a consistent ground-truthing of all the mapped features is beyond the scope of this paper. All the mapped features will therefore be referred to as fractures throughout the remainder of this paper.

Some of the fractures do coincide with large

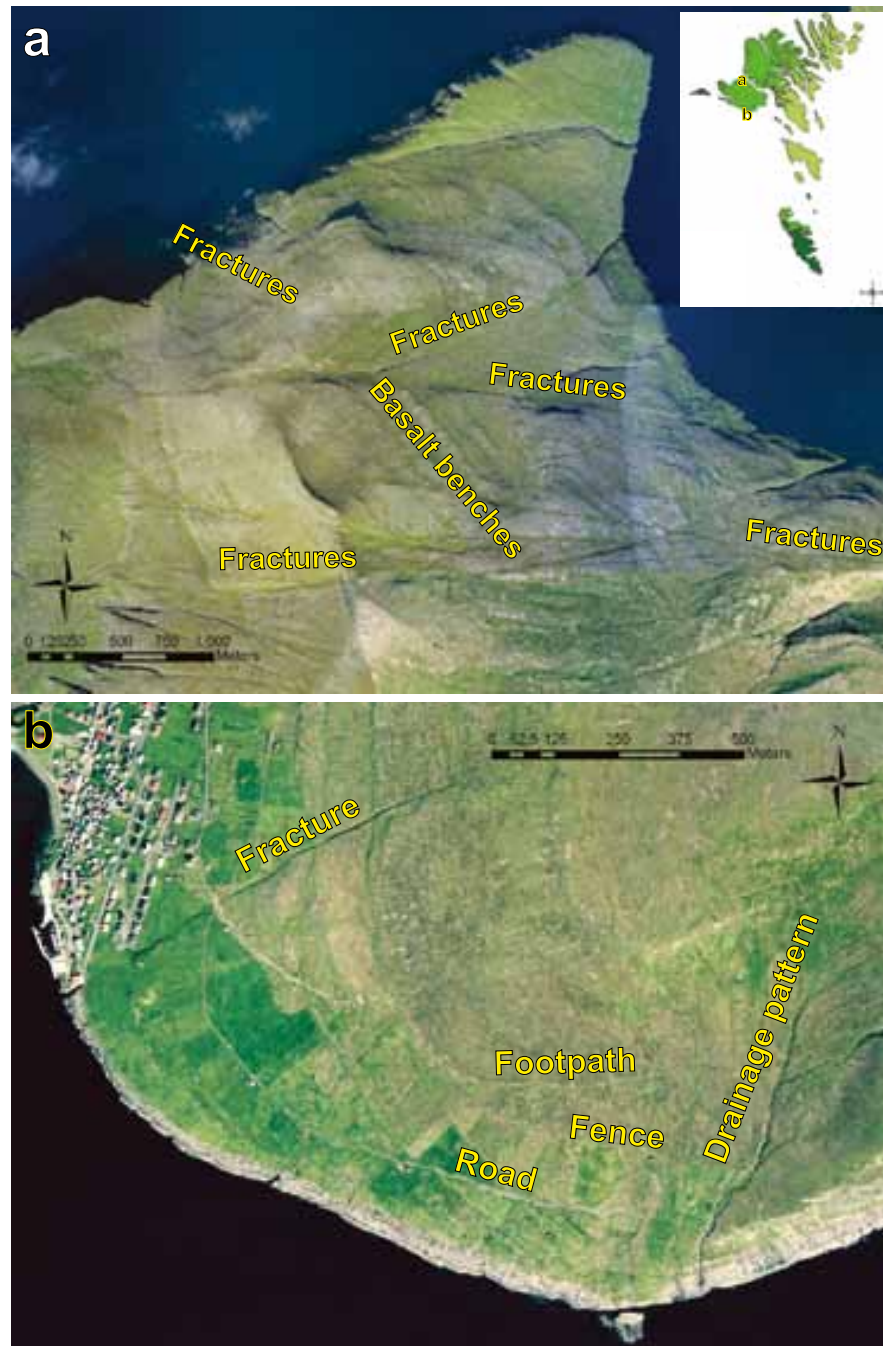


Fig. 4.

Aerial Photographs. A: Fractures and basalt benches B: Fracture, pitfalls: footpaths, drainage patterns, fence and road.

crevasses (Fig. 5d) that have been shown in the field to be caused by the erosion of crushed zones associated with pervasive fracturing. Others are associated with dykes (Fig. 5b), and in some instances dykes which have invaded fractured zones. Examples of faults can be seen in Figures 5a,c. Faults have also been documented in the literature e.g. (Rasmussen and Noe-Nygaard, 1969).

There seem to be obvious differences in the

general trends in the northern part, where most faults have a preferred East/West orientation, while the dominant trend on the southernmost island (Suðuroy) is more NW/SE (Fig. 6). The fractures (purple lines) on Figure 6 are the interpreted features. Features at right angles to the dominant morphological trends are easiest to pick up, while features that have the same overall trend as the dominant morphological features

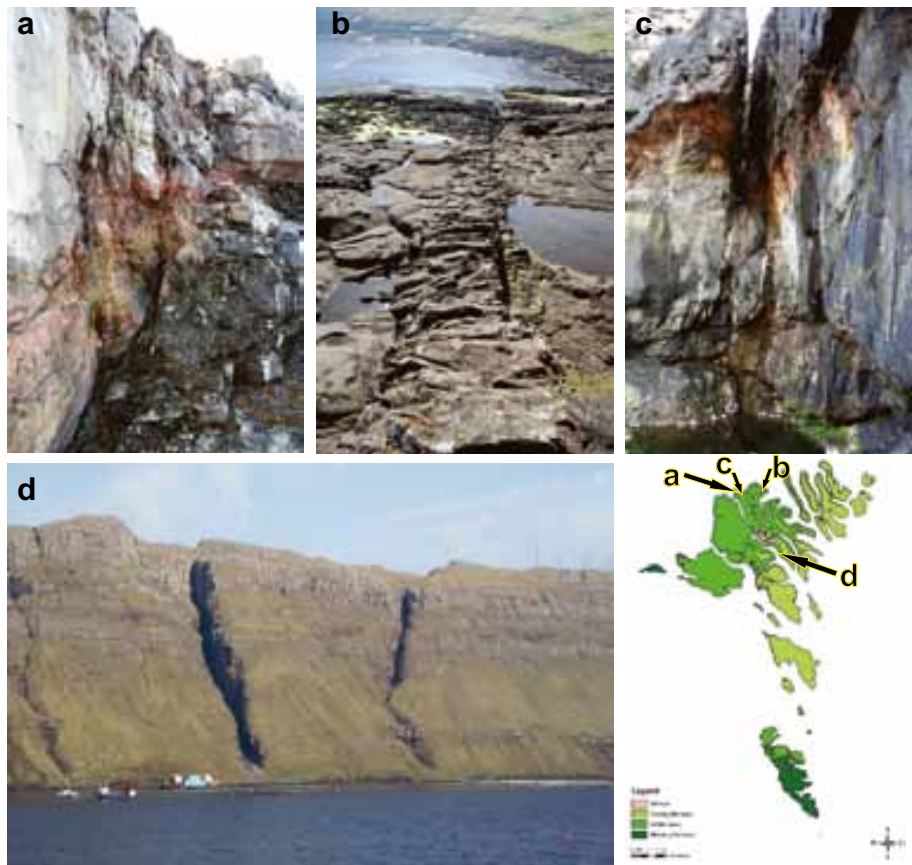
**Fig. 5.**

Image examples of lineaments. A: Fault with crushed zone, B: Dyke, C: Fault with small offset, D: Crevasses in the landscape.

are harder to pick up. The primary morphological features in question are the fjords and sounds between the islands, most of which are oriented NW/SE. These could be speculated to be related to fractures or faults, but sub-sea tunnels have shown that little fracturing along the fjords is observed in the Malinstindur Formation (Madsen, 2006). The overall assumption is thus that the interpreted fractures are representative for the Faroe Islands as a whole.

The fractures are plotted into rose diagrams (Fig. 7). The aim of this work is to investigate variations in the fracture-orientations temporally and aurally onshore Faroe Islands, which means that there was no reason to plot all in the same rose-diagram. This would also, considering the obvious differences in orientations on Suðuroy compared to the islands to the North, be of limited value. The islands are therefore grouped. The presented groupings are based on

preliminary plots and a visual inspection of the interpreted fracture orientations, where the aim was to present groupings that demonstrates variations in dominant fracture orientations.

The fractures for the different groups/islands are then plotted in Rose diagrams. The chosen groupings represented in Figure 7 are:

- The six islands towards the northeast (Fugloy, Svínø, Viðø, Borðø, Kunoy and Kalsoy)
- The two main islands with three small islands in the middle (Eysturoy, Streymoy, Koltur, Hestur and Nólsoy)
- The islands towards the west (Vágø and Mykines)
- Suðuroy.

Sandoy is not included as it is only partly covered by aerial photographs, and the total number of fractures mapped on the area covered was not

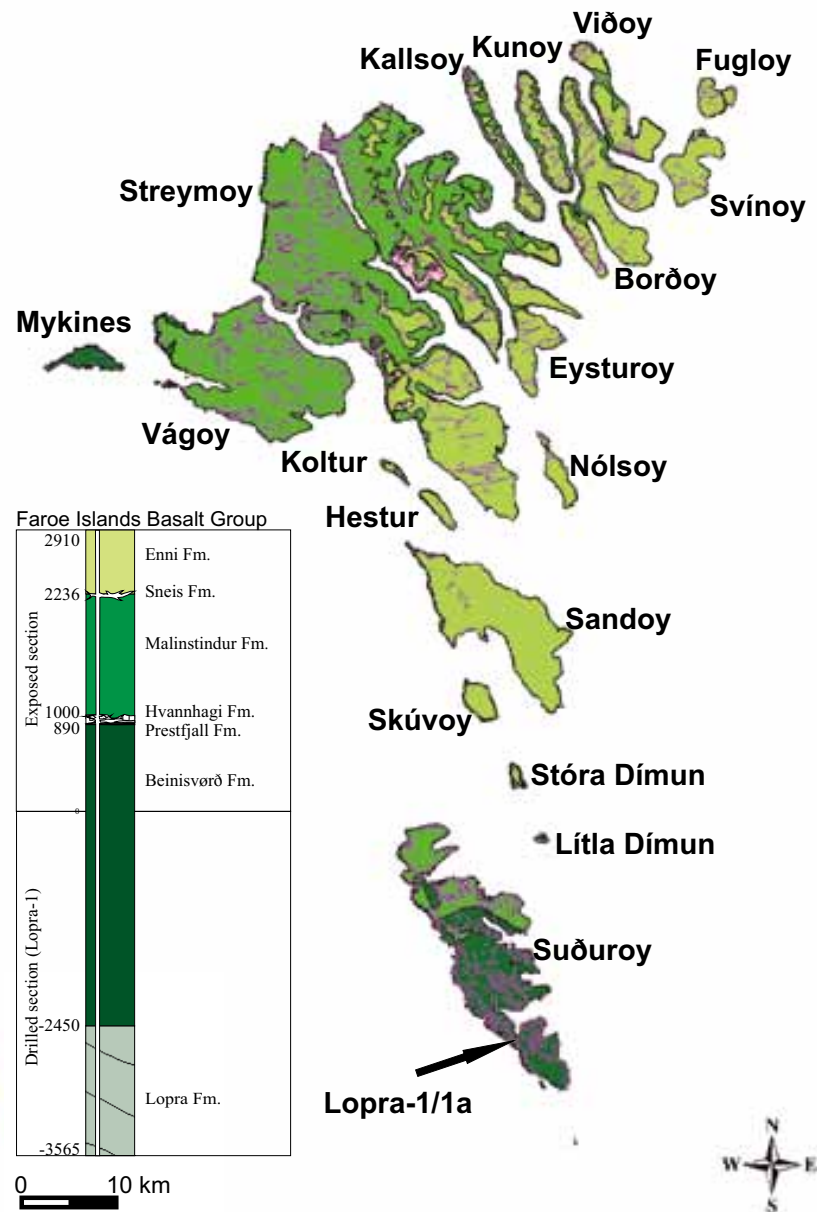


Fig. 6.

Geological Map of the Faroe Islands with interpreted fractures. Stratigraphic column of the Faroe Islands Basalt Group as inset.

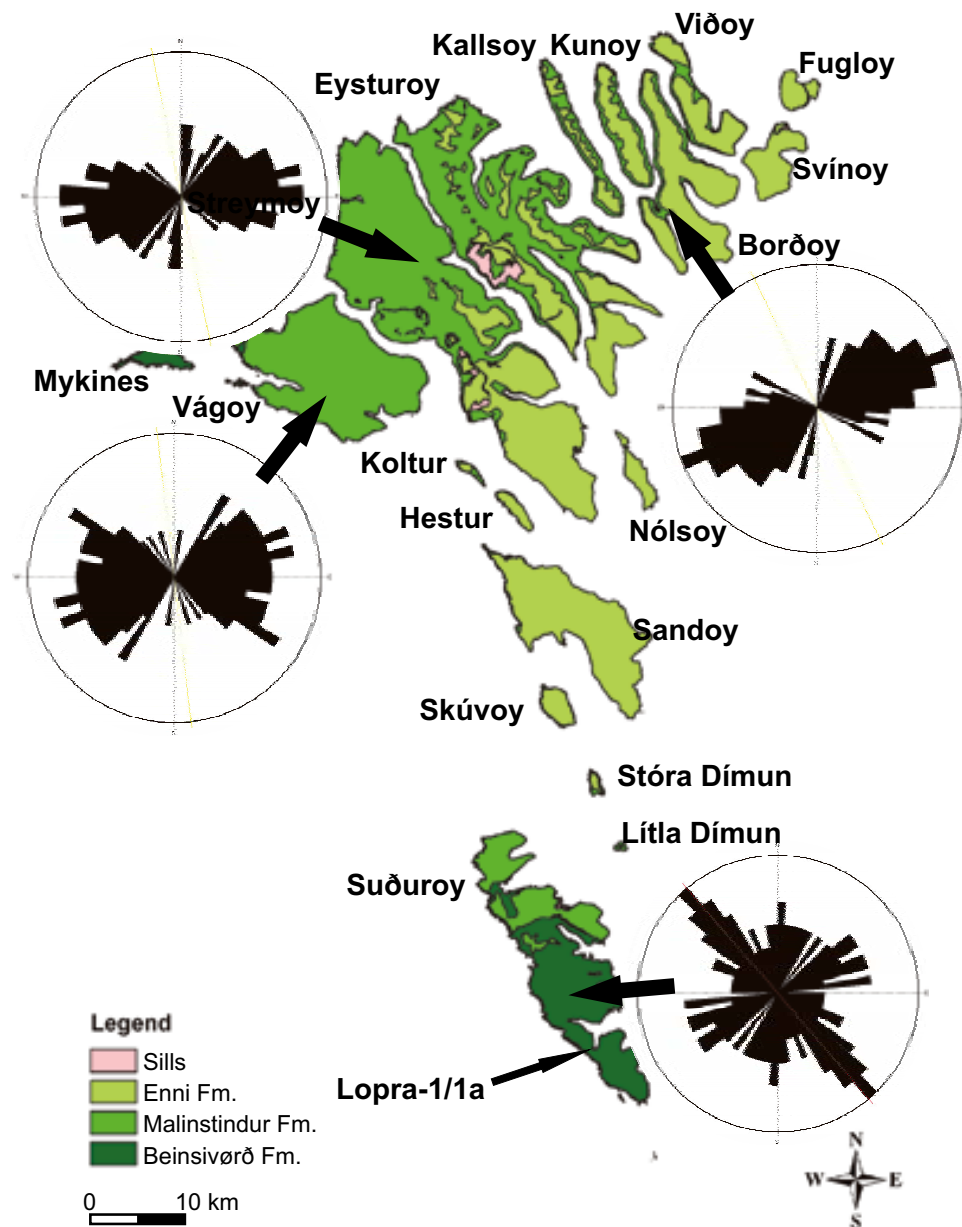
sufficient to give a valid distribution. Figure 7 shows the dominating fracture orientations on the islands as they have been grouped here.

Aerial variations in fracture trends

The observed fracture orientations on the northern islands are in a broad sense similar with a dominant E/W trend, as opposed to the observed orientation on Suðuroy where the dominating trend is NW/SE (Fig. 7).

The dominant fracture orientation on the six

small islands to the northeast (Fugloy, Svínø, Viðoy, Borðoy, Kunoy and Kalsoy) is ENE/WSW, while the dominant orientation on the two main islands (Streymoy and Eysturoy) is E/W to ENE/WSW. The dominant trend on the islands to the west (Mykines and Vágoy) is also E/W, but the distribution is broader, and lacks the increase in frequency towards the centre of the distribution (Fig. 7). There are very few fractures on any of the islands to the north which are oriented in directions that are significantly different from the dominant trends. There are, however a few

**Fig. 7.**

Geological map with rose diagrams for groups of islands. Note that no Rose diagram is shown for the islands in the centre.

dykes, and two of these are cut by younger E/W oriented dyke filled fracture zones (Rasmussen and Noe-Nygaard, 1969), which offset the older dykes. This demonstrates a temporal relationship between these two primary orientations.

Suðuroy does show a very different dominant fracture orientation compared to the other islands, with a dominant NW/SE trend and a broad distribution of other orientations at a lower frequency.

It is unfortunate that no data were available for the islands Sandoy and Skúvoy (Fig. 3 and 6) north of Suðuroy, as a full analysis of these

would give a better understanding of the transition between the dominant trends observed in the north and south of Faroe Islands respectively. This should be addressed in future work.

A more detailed analysis of Suðuroy and Vágoy/Mykines has been done in order to investigate whether the observed broader distribution of fracture orientations (Fig. 7) can be broken into different events. These analyses are discussed below.

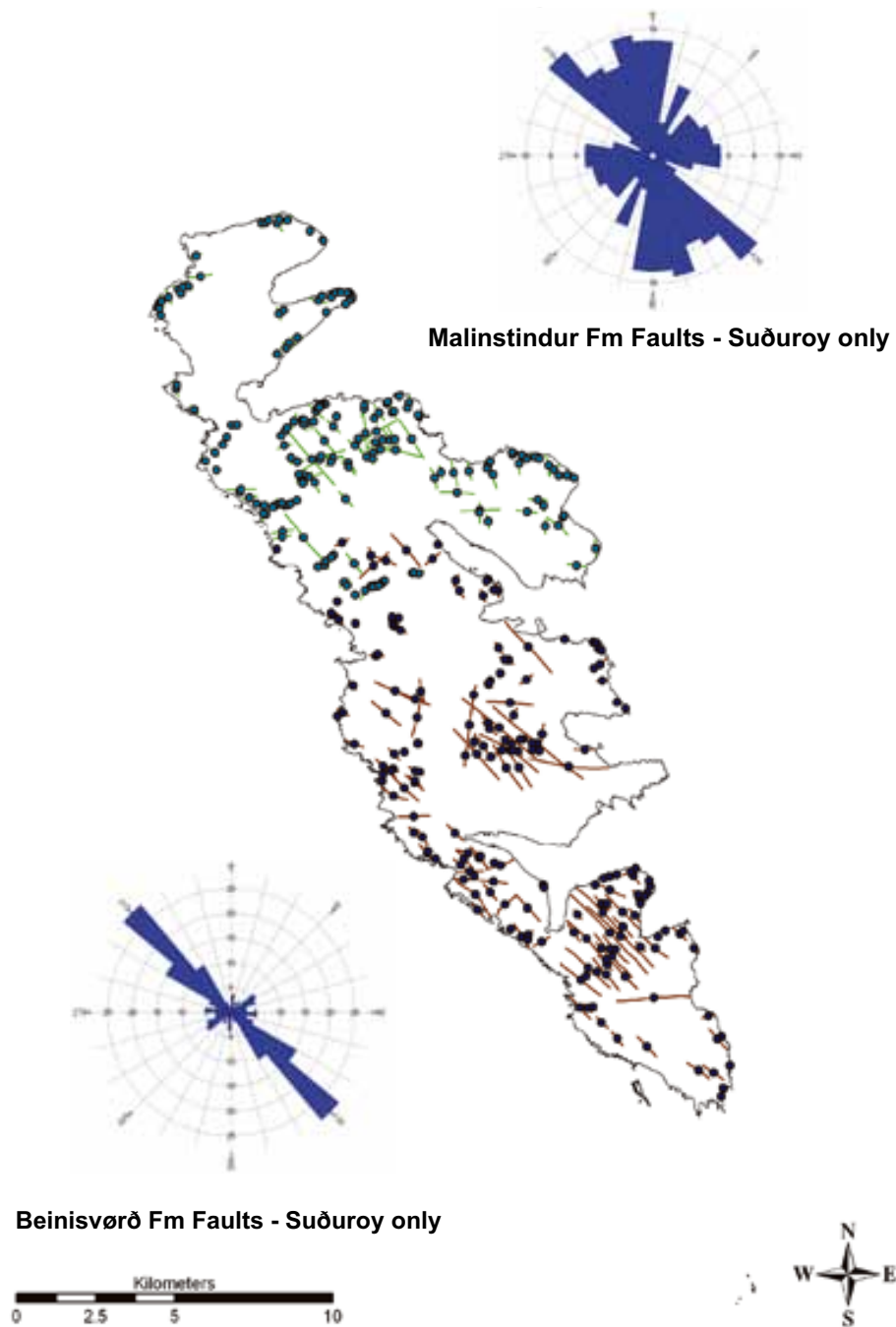


Fig. 8.

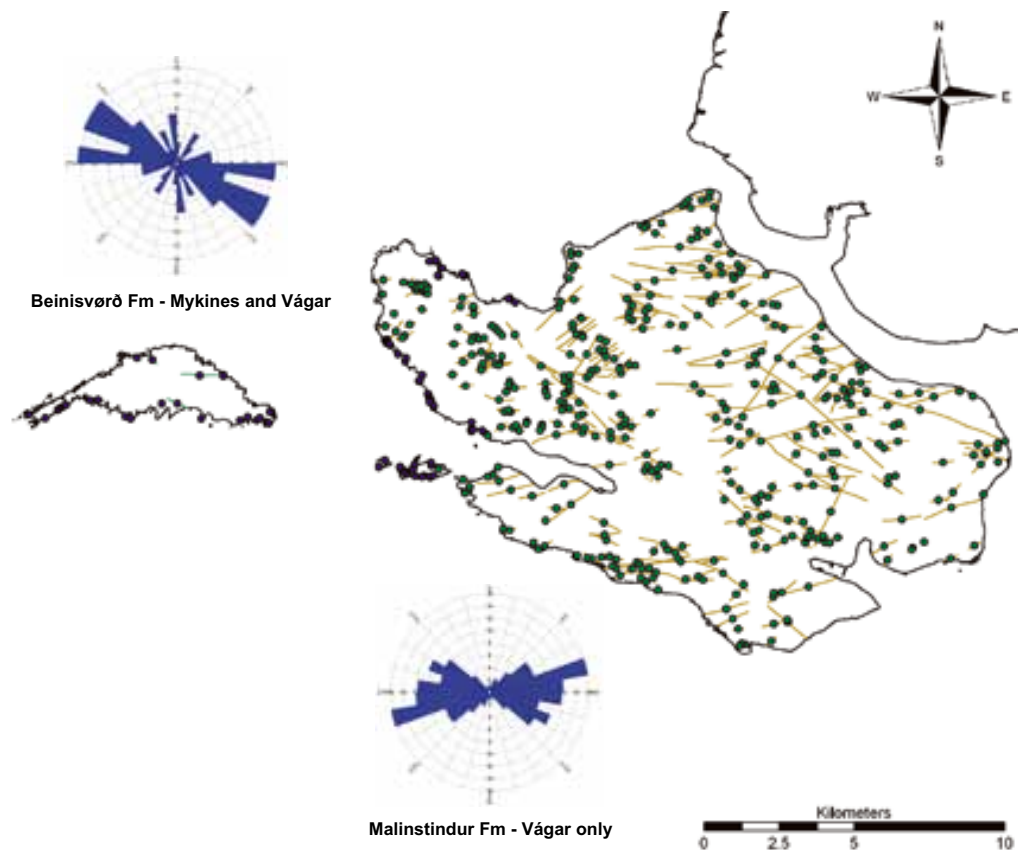
Interpreted fractures in the Beinisvørð (purple) and Malinstindur (green) Formations, and rose diagrams for both groups on Suðuroy. Small circles show centre of each individual mapped fracture.

Temporal variations in fracture trends

The southern part of Suðuroy (Fig. 6 and 7), i.e. where the oldest basalt formation is exposed, shows a more uniform NW/SE trend, with little variations, compared to a more varied directional preference in the northern part of the island, where the younger volcanic episode is represented by the Malinstindur Formation. The fractures on

Suðuroy are therefore plotted in two rose-diagrams (Fig. 8), one where all the fractures that are observed only in the older basalts (Beinisvørð Formation) are plotted, and one where fractures that can be traced into the younger basalts (Malinstindur Formation) are plotted.

The figure shows that all fractures which are only observed in the Beinisvørð Formation are

**Fig. 9.**

Interpreted fractures in the Beinisvørð (purple) and Malinstindur (green) Formations, and rose diagrams for both groups on Vágoy and Mykines. Small circles show centre of each individual mapped fracture.

oriented in a NW/SE direction. Fractures in the Malinstindur Formation have a similar dominating fracture orientation, but with a much broader distribution, and with an added E/W to SSW/NNE component at a lower frequency.

A similar analysis was performed on Mykines and Vágoy where both the Beinisvørð and Malinstindur formations are exposed (Fig. 6). The result can be seen on Figure 9. The broad distribution when looking at all fractures (Fig. 7) is dissolved into two more narrow distributions. The Beinisvørð Formation is dominated by WNW/ESE orientations, while the Malinstindur Formation is dominated by a WSW/ESE orientation.

There is thus a consistent difference in dominant fracture orientation in the Beinisvørð Formation compared to the Malinstindur Formation, which is significant as the only documented volcanic quiescence in the Faroe Islands Basalt Group is represented by the sedimentary units separating the Beinisvørð Formation from the

younger basalt formations. This work shows that the two primary fracture orientations onshore Faroe Islands are offset in time, and are thus unlikely to be caused by the same event.

The slight rotation in the dominant fracture orientation on Vágoy/Mykines compared to Suðuroy (WNW/ESE vs. NW/SE) is similar to variations observed on the Isle of Sky over comparable distances (England, 1988). England concluded that the variations are primarily due to local differences in the regional stress field.

There are no observations of fractures being confined to either the Malinstindur or Enni Formations, which prevents discrimination between these formations in subsequent analyses. This observation is also supported by the knowledge that these formations were emplaced in less than one million years in total (Jolley, 2009; Waagstein *et al.*, 2002). They do temporally seem to overlap, and it is therefore less likely that there are significant tectonic changes during this period.

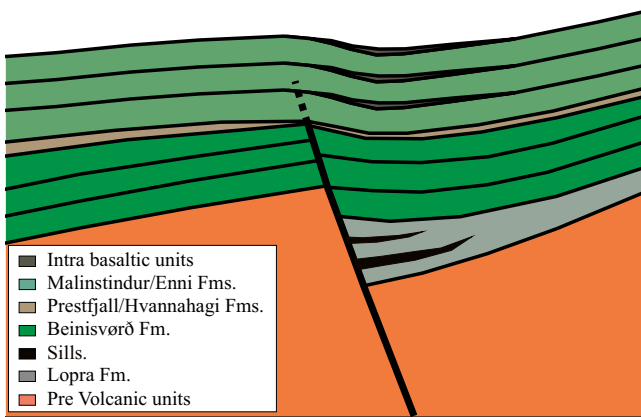


Fig. 10.

Conceptual cross-section showing the proposed link between faulting in the Beinissvørð Formation and geological development of the Malinstindur and Enni Formations.

Discussion

The observed fracture orientations onshore Faroe Islands do give some indications regarding the tectonic evolution of the Faroese Platform, which in turn reflects the tectonic evolution on a more regional scale.

It has been suggested by a number of authors i.e. (England, 1988; Gudmundsson, 1995, 2002; Nakamura, 1977) that regional stress regimes in volcanic areas are the primary controlling factor regarding orientations of fractures, faults and regional dykes. The distinctly parallel NW/SE fracture orientation of the observed fractures on Suðuroy could therefore be taken as a clear indication that the minimum regional stress orientation was SW/NE during the emplacement of the Beinissvørð Formation. This is previously documented by Rasmussen and Noe-Nygaard (1969) who observed a NW/SE oriented fault that clearly predates the emplacement of the Malinstindur Formation. Where several faults are found, there is a tendency for the faults to be downthrown in the same direction (Rasmussen and Noe-Nygaard, 1970). NW oriented dykes are observed towards the west on the British Isles and extending north- and westwards (Fig. 2). Dewey and Windley (1988) refer to the zone where these are observed as an „Extensional magmatic zone,“

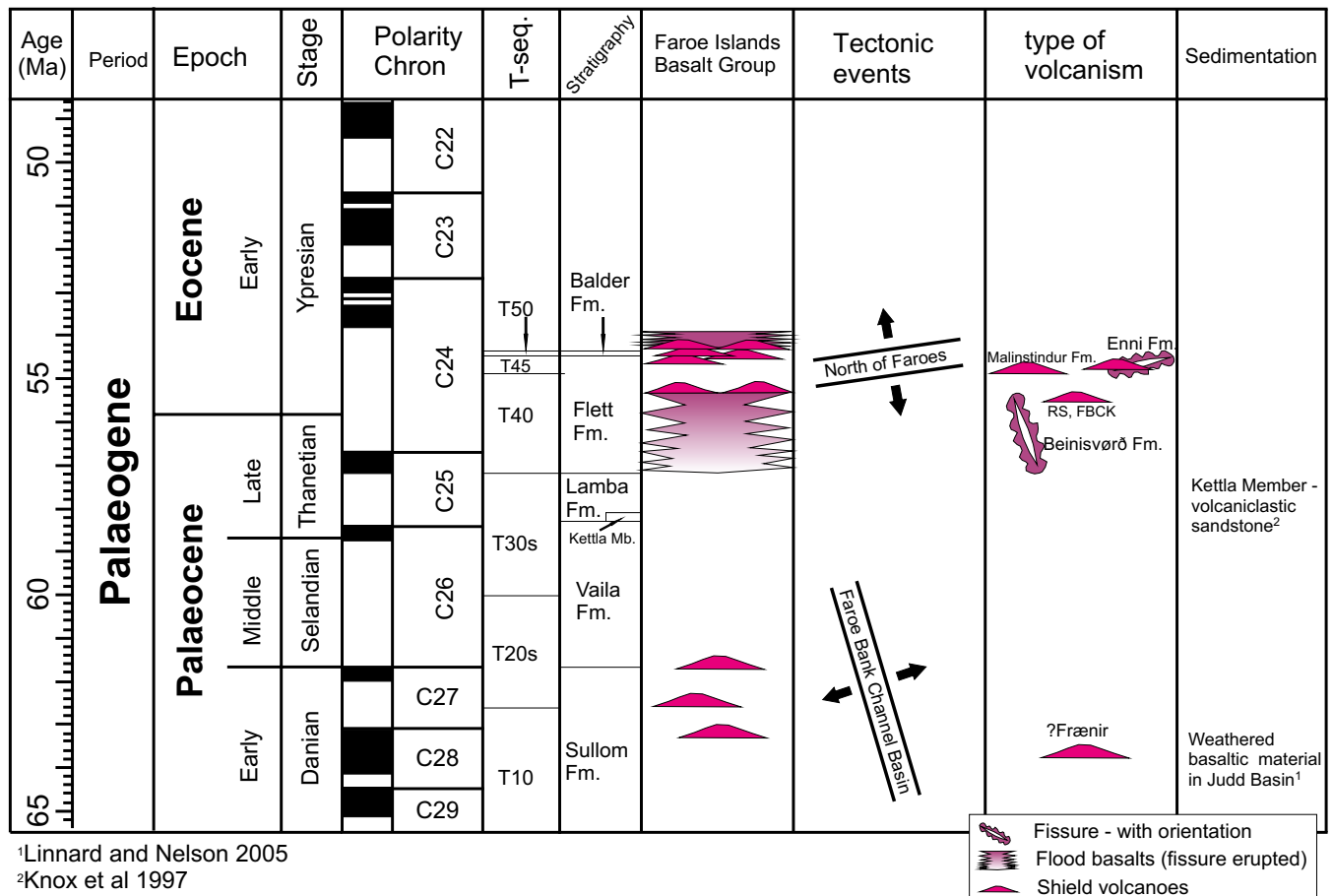
which are caused by rapid stretching (Dewey, 2002).

The younger volcanics of the Malinstindur and Enni Formations are on the northern islands cut by fractures that have dominant E/W orientation. This can again be taken as evidence for a regional stress regime with the minimal stress direction being N/S during the emplacement of these younger basaltic units.

Work by Geoffroy (1994) on the Faroe Islands did show a similar change in regional stress directions. The timing suggested by Geoffroy is however different. He suggests that the SW/NE rift event is concurrent with the emplacement of the Malinstindur Formation, while a N/S extensional regime was present during the emplacement of the Enni Formation. There is thus a good agreement in the sequence of extensional events in the work done by Geoffroy (1994) and this work.

The Malinstindur and Enni Formations were emplaced in less than one million years (Waa-gstein *et al.*, 2002). No change in fracture orientation between the Malinstindur and Enni Formations is observed during the work presented in this paper. This work has demonstrated that the fracture orientation in the Beinissvørð Formation is distinctly different from the Malinstindur and Enni Formations (Fig. 8 and 9). There are, however a few NW/SE oriented fractures present in the Malinstindur Formation on Vágoy and Suðuroy. These can most likely be attributed to re-activation of the older rift related faults during later tectonic events (Fig. 10). The decreasing number of such fractures further northeast is likely linked to the distance from the early Palaeocene transient rift. Dip of the basaltic units and an increasing thickness of younger volcanics towards the west (Rasmussen and Noe-Nygaard, 1969) can also influence the frequency of reactivated older features.

The older faults were most likely controlled by the regional stress regime during the extensional phase that resulted in the emplacement of the Beinissvørð Formation. This work has shown that the NW dominant trend is mostly confined

**Fig.11.**

Palaeogene evolution: Timing and style of magmatism, discussed rift events and volcanically derived sedimentation.

to the older lava formations, which is supported by Rasmussen and Noe-Nygaard (1969) who observed a fault in the Beinisdvørð and Prestfjall Formations, which did not reach into the overlying Malinstindur Formation. Some of these faults can however have been reactivated during later events and thus explain the small number of NW/SE oriented faults in the younger basalt formation. In addition flexures resulting from the compaction of syn-rift (pre-and syn-volcanic) sedimentary deposits (Fig. 10) can have resulted in small synclines that can have acted as sediment traps and/or been the focus point of erosion, and thus could be speculated to be the underlying reason for the dominating topographic NNW/SSE trend observed onshore Faroe Islands (Fig. 7).

The younger formations are consistently fractured by predominant east/west oriented fractures, and thus parallel to the continental breakup

line north of the Faroese Platform. The extensional regime associated with these is therefore correlateable to the event which led to continental breakup between Eurasia and Greenland in the Eocene.

The fractures observed onshore Faroe Islands do thus indicate that there was a distinct shift in the dominating minimum stress direction (Fig. 11) from the emplacement of the Beinisdvørð Formation (minimum stress: SE/NW) to the emplacement of the Malinstindur/Enni Formation lavas (minimum stress: N/S).

Regional tectonic development

Pre-flood basalt volcanism

The two extensional events exemplified by the analysis of onshore fracture trends need to be dis-

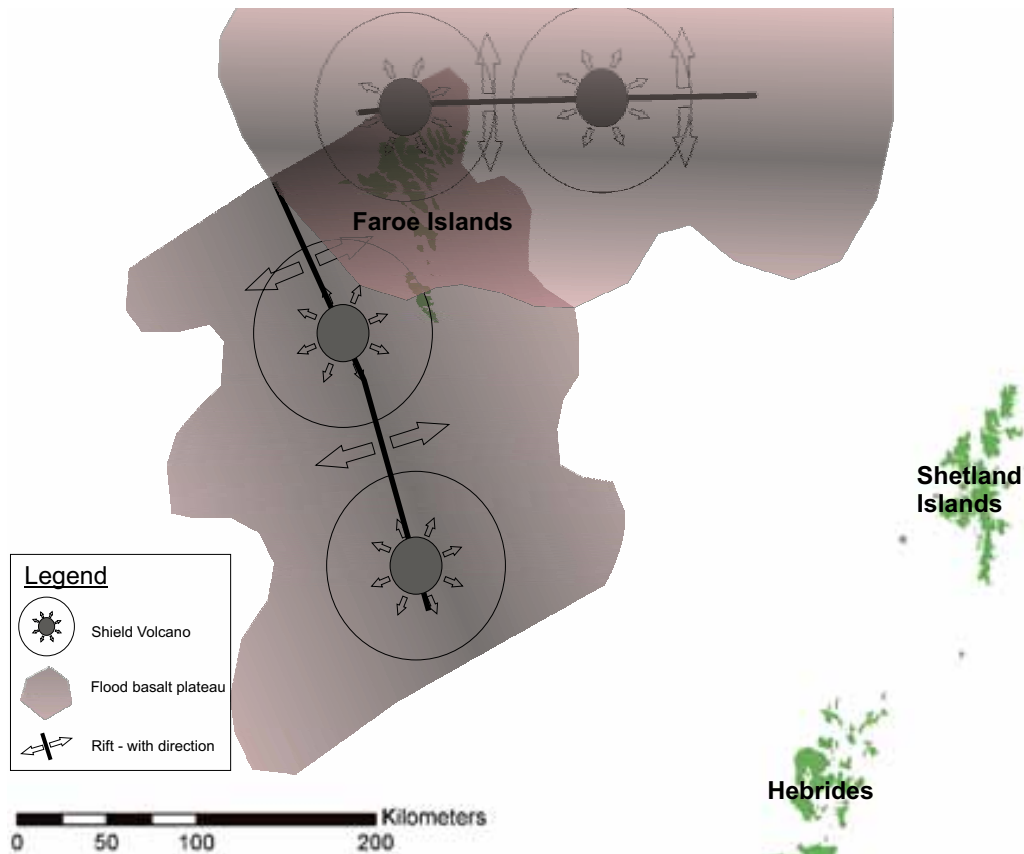


Fig. 12.

Rift events, with suggested thinning of volcanics being shown as lighter coloured areas. Circles show suggested or known locations of igneous centres.

cussed in a more regional context, where there unfortunately is a shortage of literature regarding the early Palaeocene transient rift event. Some authors have, however, discussed this option.

Dewey (2002) suggested „rapid stretching across a NW-trending line“ running from SE Greenland, Rockall, Faroe Islands, northern part of Ireland and NW Scotland. Waagstein (1988) proposed, based on geochemical data, a transient rift as the primary cause of the Beinissvørð Formation lavas. This rift had a NE/SW orientation, and was followed by a parallel rift in east Greenland. Lundin and Doré (2005) also proposed, based on regional occurrences of volcanism and tectonic trends, that there was a transient rift in the Early Palaeocene. This rift had a NNW/SSE preferred orientation across the Faroese area and was located just west of the island, and coincides with the area Dewey (2002) previously had suggested was subjected to rapid stretching during

the Palaeocene. Evidence for the stresses associated with this event are observed onshore northwest Britain (England, 1988). Ziska and Varming (2008) presented evidence for such an event based on reflection seismic data. Funck *et al* (2008) have performed wide aperture modelling across the North Faroe Bank Channel Basin. Their results show crustal thinning across the North Faroe Bank Channel Basin, which they, due to the lack of evidence for a major extension in SW-NE direction within the literature, proposed to be a result of transform fault movements.

Massive input of sediments from the west into the Judd Basin (Varming, 2009), some of which contain eroded basaltic material (Linard and Nelson, 2005) suggests that the rifting started in the Early Palaeocene (Fig. 11). This is supported by the presence of early Palaeocene igneous activity in Scotland, e.g. Mull, Skye, Rhum-Muck, Jura-Islay and Arran (Bell and

Williamson, 2002) where, in addition to central igneous complexes there are a number of NW/SE to NNW/SSE oriented regional dykes (Fig. 2 and (Dewey, 2002)). The igneous activity has in some areas resulted in a 35% extension of the crust (Bell and Williamson, 2002) indicating a minimum stress direction that is SW/NE, i.e. parallel to the proposed transient rift in the Faroe Bank Channel Basin. Parallel dykes can be traced on magnetic data from Scotland and Northern Ireland towards the NW, where some of them line up with the Faroe Bank Channel Basin (Linnard and Nelson, 2005). This is on trend with Nagsugtoqidian/Ammasalik to Lewisian primary trends (Buchan *et al.*, 2000; Whitehouse and Bridgewater, 2001) and indicates that the rift has exploited pre-existing weaknesses caused by the mentioned orogenic phases.

The presence of pre-flood basalt volcanics on the Faroese Continental Shelf has not been proven through sampling, but the presence of volcanoclastic units such as the Kettla Member (Fig. 11) that increases in thickness towards the Munkagrinnur Ridge (Knox *et al.*, 1997) suggests that there are pre-flood basalt volcanic units present under the Beinisvørð Formation. One such candidate is the Frænir Igneous Centre (Fig. 11) at the south western edge of Munkagrinnur Ridge (Fig. 1). This feature has a distinct potential field anomaly (Fig. 1), but is not visible on seismic data within or above the basalt. The deeper section can not be mapped on available seismic data.

Seafloor spreading into the Labrador Sea, north of the Charlie Gibbs Fractures, zone was initiated in the Cretaceous and continued into the Palaeocene e.g. (Tate *et al.*, 1999), and the latter stages are thus concurrent with the proposed transient rift transecting the Faroese Continental Shelf. It is thus likely that the transient rift is linked to the seafloor spreading in the Labrador Sea – Baffin Bay, potentially as an attempt to open a new rift path, after further rifting in the Baffin Bay stopped, as proposed by Lundin and Doré (2005). This ultimately failed attempt to open a new NW/SE oriented rift from Baffin Bay across Greenland, Faroes and northern Scotland seized, when

a new rift was opened at the Charlie Gibbs Fracture Zone, reaching north-eastwards in the latest Palaeocene and into the Eocene. This resulted in continental breakup between Greenland and Eurasia in the Eocene e.g. (Larsen *et al.*, 1999b), which is contemporaneous with the extrusion of the Malinstindur and Enni Formation lavas.

Lopra and Beinisvørð Formations

The initial rift event, that resulted in the extrusion/intrusion of the early Palaeocene igneous units was followed by fissure erupted flood basalt volcanism. Rasmussen and Noe-Nygaard (1969) proposing that the fissures were oriented NW/SE, which is parallel to the observed dominant fracture trend in the Beinisvørð Formation (above).

The absolute timing of the initiation of the flood basalt volcanism is disputed with dates based on radiometric dating being 58,8 Ma (Waagstein *et al.*, 2002) and on palynological correlation being 54,9-57 Ma (Jolley, 2009). This is compared to the 61-58 Ma years assigned for most of the Palaeogene Igneous rocks in Scotland (Ganerød *et al.*, 2008). The flood basalts do thus postdate the earliest volcanics, including the observed eroded material found in the Judd Basin (Linnard and Nelson, 2005) by up to several million years. Such a delay has been observed in other volcanic provinces (Jerram and Widdowson, 2004).

What is more striking is the area covered by these flood basalts i.e. Beinisvørð Formation on the Faroese Continental Shelf and the Nansen Fjord Formation in East Greenland (Larsen *et al.*, 1999a). Larsen *et al.* propose that the Beinisvørð Formation is primarily constrained to an elongated NW/SE transect from East Greenland to the Faroe Islands. This would be along the transient rift zone discussed in this paper, but the lack of firm data points, which is acknowledge by Larsen *et al.* (1999a) highlights the need for more research into the issue.

The mapped fractures within the Beinisvørð Formation are distinctly parallel to the proposed main rift axis in the Faroe Bank Channel Basin. The regional stress causing faulting of the Beini-

svørð Formation seems to be terminated prior to the initiation of renewed volcanism as demonstrated by Rasmussen and Noe-Nygaard (1969) who observed fault movement on NW/SE oriented faults terminating prior to emplacement of the younger volcanics.

Malinstindur and Enni Formations.

The volcanically quiet period was followed by renewed volcanism. The first expression of this phase is the Malinstindur Formation, which was most likely emplaced as shield volcanoes (Passey and Jolley, 2009). Such volcanoes are normally located near the main rift axis, which at this time was the north Atlantic rift that had propagated to the area north of the Faroese Platform at this time. It is thus to be expected that the thickness of the Malinstindur Formation lavas is more localised and thins away from the rift, as also speculated by Kjørboe (1999). This is in line with observations that show 1400 m present on central Streymoy (Rasmussen and Noe-Nygaard, 1969), while little to no lavas are expected to have been present on southern Suðuroy (Jørgensen, 2006). The second part of this volcanic phase is represented by the Enni Formation, which is partly point source and partly fissure erupted (Passey and Jolley, 2009). The fractures found in this section do support a regional minimum stress direction that is N/S, and thus creating an E/W oriented rift axis. This rift evolved into continental breakup between Eurasia and Greenland in the Eocene. Fault movements do show some degree of strike slip movement of up to 100 m (Rasmussen and Noe-Nygaard, 1969), which can be seen as a consequence of the initial E/W movement of Greenland relative to Eurasia in the Eocene (Gaina *et al.*, 2009).

Temporal evolution and its implications for the thickness of lava pile

The overall evolution of the area is summarised in Figure 12, which shows the initial rift west of the Faroese Platform, which was most likely

associated with erosion of the rift-shoulders. Indications of this is seen in the massive input of sediments in the Judd basin (Varming, 2009). The flood basalt volcanism associated with this rift is extruded through rift-parallel dykes (Rasmussen and Noe-Nygaard, 1969), as is seen in other volcanic rifts (Gudmundsson, 2002). The volcanism is expected to be most pronounced near the rift, and thus thinning in either direction away from the rift (Fig. 12), as indicated by the distribution of the older flood basalt formation (Beinissvørð Formation) in Larsen *et al* (1999a). A full evaluation of this is, however beyond the scope in this paper.

The early volcanism is followed by a volcanically quiet period, in which onshore Faroe Islands is subjected to erosion (Rasmussen and Noe-Nygaard, 1969), where the erosive products are re-deposited down dip. There is thus a likelihood that the aerially constrained location of the early volcanism (i.e. along the rift axis) means that sediment transport from Greenland could have continued through this period in the areas further away from the rift, i.e. eastern part of the Fugloy Ridge (Fig. 1).

The different areas of the Faroese Continental Shelf were thus covered by volcanic rocks at different times and from different sources. This means that we are looking at a more complex evolution of the North Atlantic Igneous Province on the Faroese Continental Shelf than previously expected.

Implications for prospectivity

The most prolific reservoir rocks in the Faroe Shetland Basin have to date been Palaeocene sands, mainly in the Vaila Formation e.g. (Lamers and Carmichael, 1999). These are deposited after the transient rift was initiated and before the final continental breakup and final cessation of volcanism. This shows the importance of a full understanding of the areal and temporal evolution of the two rifts and volcanism during the Palaeocene. The potential effects that are directly relevant are increased heat flow, rapid burial due

to loading of basalt and potential contamination of siliciclastic sediments by eroded basaltic material.

The first phase of volcanism was most likely constrained to discrete shield volcano type eruptions in the vicinity of the rift. These are not expected to have a significant effect on large areas, but erosive products are likely to contaminate potential reservoir units, and in some cases become reservoirs (Ólavsdóttir and Ziska, 2009). The influence of these erosive products is expected to be most pronounced near the rift, as indicated by the thickening of the debris flow deposited Kettla Member (Fig. 11) towards the Munkagrunnur Ridge (Knox *et al.*, 1997).

The offset in time between the two rifts means that areas were affected differently by the two volcanic phases. This is quite different from previous assumptions where the discussion has been limited to which volcanic phase reached furthest out into the Faroe Shetland Basin e.g. (Ritchie *et al.*, 1999; Smythe, 1983). The direct implication for hydrocarbon exploration is that the eastern part of the Faroe Shetland Basin is likely less affected by the first volcanic phase due to the focus for volcanism being in the Faroe Bank Channel Basin. The probably late cut-off of sediment input from Greenland is likely to have had a positive effect on the overall prospectivity in this area. This does however, not take into consideration the volcanism associated with the Brendan and Erlend igneous centres, which are expected to be contemporaneous with the Beinisvørð Formation lavas (Jolley and Bell, 2002).

The Faroe Bank Channel Basin area on the other hand is located on the early Palaeocene rift, and is thus likely to have experienced an increased heat flow during the extensional phase. How this has affected the hydrocarbon system is hard to evaluate due to lack of data regarding presence and paleo- and current depth of the source rock. The close proximity of the area to the second rift, i.e. breakupline, means that the second rift phase has also affected this area, and there is consequently a risk that both volcanic phase associated with the transient rift and the one associated with continental breakup are present quite far south

and the area thus is affected by both phases to some degree. Whether the evidence presented by Jørgensen (2006) regarding the non-emplacement of lavas on the southern part of Suðuroy indicates a maximum southward extent or is a result of the Munkagrunnur Ridge being a positive feature at the time (Ziska and Varming, 2008) is not something present data give the basis to conclude on.

Conclusions

Distinctly observed fracture orientations in the older basalt formations compared to the younger formations shows that the area was subjected to different stress regimes during the two volcanic episodes.

The rifts that resulted in the emplacement of the Beinisvørð and Malinstindur/Enni Formations were offset in both space and time.

The north-eastern part of the Faroese Continental Shelf area seems to be less affected by the early volcanism due to its distance from the earlier rift event, allowing for continued sediment input from a potential Greenland source.

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Heavy mineral constraints on Paleocene sand transport routes in the Faroe-Shetland Basin

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Abstract

An integrated study of Paleocene sand provenance in the Faroe-Shetland Basin, involving heavy mineral analysis, garnet geochemistry, rutile geochemistry, zircon age dating and palynology, has revealed the presence of four distinct sand types with different provenances. Three of these sand types (FSP1, FSP2 and FSP4) can be tied back to sources on the Orkney-Shetland Platform to the east, but sand type FSP3 is considered to have a western origin. Although FSP3 has mineralogical similarities with coeval sandstones in the Kangerlussuaq area of East Greenland, differences in zircon age spectra rule out this region as an entry point for FSP3 sands. The sub-basaltic basement of the Blosseville Kyst and the Faroe Islands, together with intrabasinal highs, are therefore considered more likely sources. Only 10% of samples in the data set have FSP3 characteristics, which contrasts with previous sediment volume balancing calculations that suggested that 30% of sediment was supplied from the west. Sampling bias (controlled by well location) is the main reason for this discrepancy, since wells in Faroes sector contain a much higher proportion of FSP3 sediment (> 35%).

The easterly-derived sandstones show marked evolution in provenance characteristics with time. Most of the sandstones in the T10-T34 succession were recycled from pre-existing sandstones on the Orkney-Shetland Platform, whereas basement sources were prevalent in the T35-36 and T38 successions. The change in provenance coincides with a change in depositional style, which has been linked to thermal doming, rifting and volcanism in the NE Atlantic. The shift from recycled to first-cycle provenance is marked by a short-lived influx of sand type FSP3 in conjunction with volcanoclastic material, indicating a link between changes in provenance, magmatism and depositional style in the region.

Introduction

An understanding of the distribution of Paleocene sandstones in basins west of the Shetland Isles is of crucial importance for hydrocarbon exploration in both the Faroes and UK sectors of the NE Atlantic. With the exception of the Clair Field, hosted by Devonian-Carboniferous clastics (Allen and Mange-Rajetzky, 1992), virtually all the significant hydrocarbon discoveries in the area to date are hosted by Early Tertiary sandstones, such as the Foinaven and Schiehallion fields (Cooper *et al.*, 1999; Leach *et al.*, 1999) and the Marjun and Rosebank discoveries (Smallwood *et al.*, 2004; Helland-Hansen, 2009). Sandstones in this stratigraphic interval remain the principal target for exploration in the area.

During the Paleocene, prior to the Early Eocene separation of Greenland from NW Europe, a relatively narrow restricted marine basin occupied the Faroe-Shetland area (Smallwood and White, 2002). Consequently, during this time there was potential for sediment derived from landmasses to the west (Greenland) and the east (Orkney-Shetland Platform) to reach the basin centre (Smallwood, 2005). Lamers and Carmichael (1999) suggested that most of the Paleocene clastic sediment in the UK sector was sourced from the east of the basin, the British Isles having been a long-lasting emergent provenance area. This view is supported by previous provenance studies (Morton *et al.*, 2002; Jolley *et al.*, 2005; Jolley and Morton, 2007). However, Jones *et al.* (2002) considered that the volume of material denuded from the eastern margin is insufficient to account for the sediment in the Faroe-Shetland Basin. According to Smallwood (2008), sediment volume balancing suggests that around 30% of the Paleocene sediments currently in the basin were sourced from a westerly provenance area, the most likely candidates being East Greenland and the pre-basaltic succession of the Faroes Platform (Fig. 1). Successful exploration of Paleocene targets in the western part of the Faroe-Shetland Basin is dependent on the presence of arenaceous clastic sediment shed from sources other than the Orkney-Shetland Platform to the east.

In this paper, we present the results of an integrated provenance study of Paleocene sandstones in the Faroe-Shetland Basin, and consider the implications regarding sediment source locations and sand transport routes. The provenance framework is provided by heavy mineral data, major element garnet geochemistry, trace element rutile geochemistry, zircon age dating and palynology. A similar approach proved successful in identifying sand sourcing and constraining transport routes in the Cretaceous-Tertiary of the Norwegian Sea (Morton *et al.*, 2005) and the Kangerlussuaq area of East Greenland (Whitham *et al.*, 2004).

Geographical and Stratigraphic Scope

The data set comprises results from over 300 samples from seventeen hydrocarbon exploration wells, fourteen of which are located in the UK sector and the remaining three in the Faroes sector (Fig. 2). The wells were chosen in order to maximise geographical coverage across the region. Extension of the study into UK Quadrant 209 and the northern part of Quadrant 208 was precluded by the scarcity of sand in the Paleocene of this region. The majority of data were collected from ditch cuttings, but core and sidewall core samples were analysed wherever possible.

The stratigraphic interval discussed herein comprises the Sullom, Vaila and Lamba formations (T10-T38 as defined by Ebdon *et al.*, 1995), as shown in Fig. 3. The sandstones in these formations are predominantly of mass-flow submarine fan origin (Lamers and Carmichael, 1999), deposited in relatively deep water (Mudge and Bujak, 2001). The overlying Flett Formation (T40-T45) is not discussed here, although preliminary heavy mineral studies suggest there is a distinct shift in provenance patterns at this level. This change is attributed to a marked shallowing of water depth, with deposition taking place in shallow marine and paralic environments (Mudge and Bujak, 2001). The change in water depth is attributable to regional plate uplift associated with the Icelandic plume (Mudge and Jones, 2004; Rudge *et al.*,

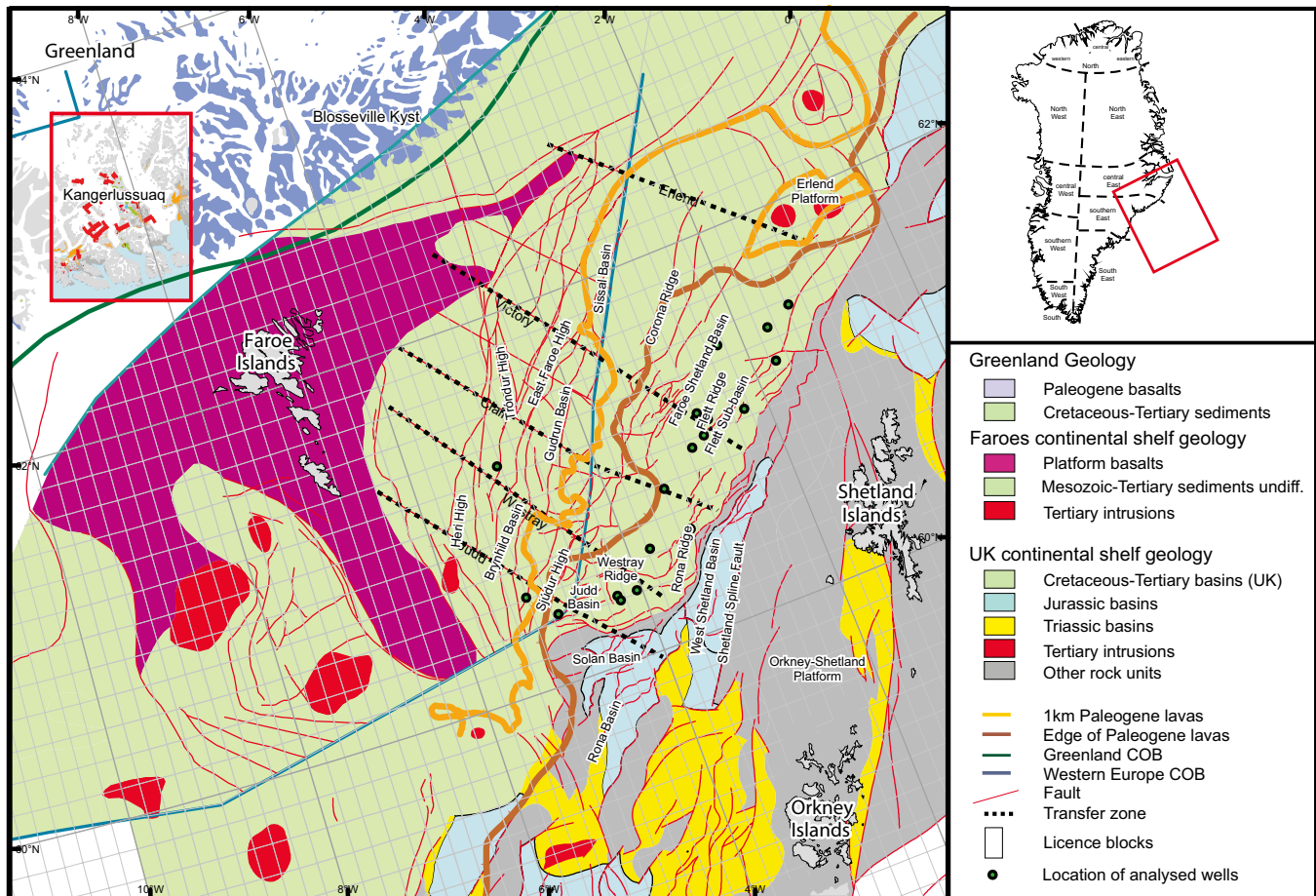


Fig. 1.

Best-fit pre-rift reconstruction of the Faroe-Shetland region immediately prior to opening of the NE Atlantic in the Early Eocene (magnetic anomaly 24), adapted from Larsen and Whitham (2005), showing main structural features. Boxed area is the Kangerlussuaq region of southern East Greenland (Whitham *et al.*, 2004). Green circles denote wells discussed in the paper (see Fig. 2 for detail).

2008). The change from localised lava eruption to more widespread basaltic volcanism and the establishment of extensive subaerial basalt fields in the western part of the basin, which took place in T40 (Jolley and Bell, 2002) is another factor that had the potential to affect provenance patterns at this time. The onset of major basaltic volcanism in East Greenland also took place in T40 (Jolley and Whitham, 2004).

Analytical Scope and Methods

The basic framework for discrimination of sand provenance in the Paleocene of the Faroe-Shetland Basin is provided by conventional (petrographic) heavy mineral data, in particular provenance-sensitive heavy mineral ratios,

together with garnet and rutile geochemical data. Sandstone provenance has been further characterised by acquisition of zircon age data. In addition, sediment input from both the eastern (Orkney-Shetland) and western (Greenland) margins of the basin has been identified on the basis of studies of terrestrially-derived palynoflora (Jolley *et al.*, 2005; Jolley and Morton, 2007).

Heavy mineral analysis

Core samples were gently disaggregated by use of a pestle and mortar, avoiding grinding action. This was not necessary for cuttings samples because they were already disaggregated through the action of the drill bit. Chemical pre-treatment was not used, thereby avoiding the

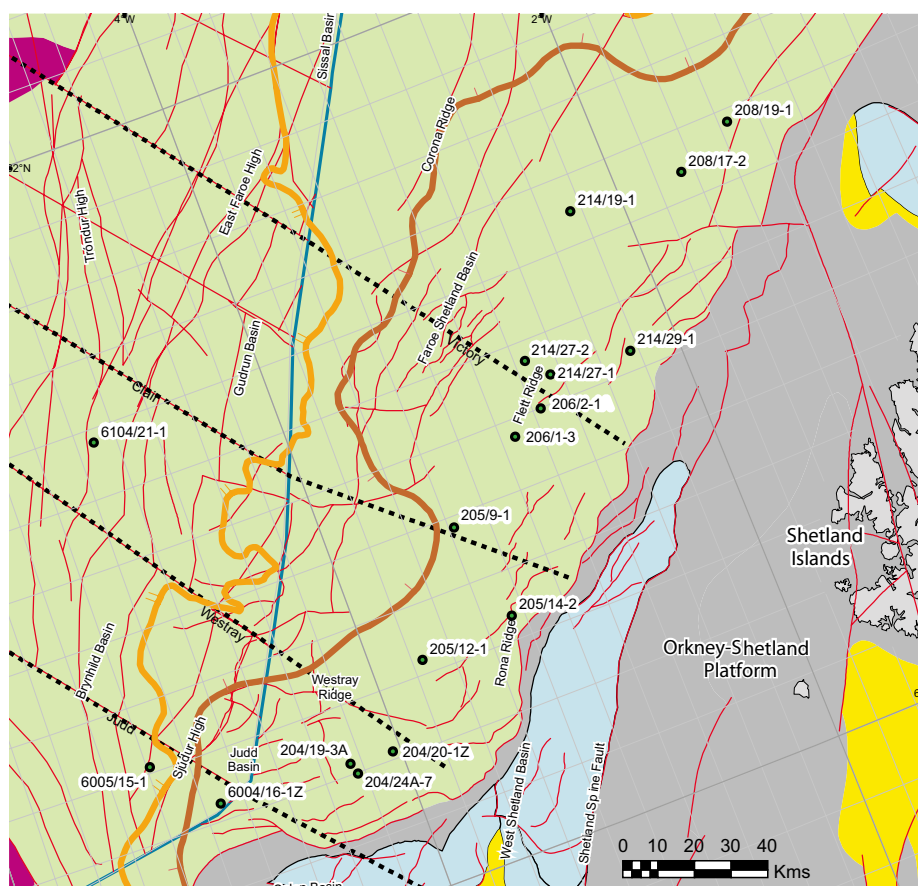


Fig. 2.
Location of wells discussed in the paper.

possibility of modifying assemblages in the laboratory. Following disaggregation, the samples were immersed in water and cleaned by ultrasonic probe to remove and disperse any clay that was adhering to grain surfaces. The samples were then washed through a 63 μm sieve and resubjected to ultrasonic treatment until no more clay passed into suspension. Following this, the samples were wet sieved through the 125 and 63 μm sieves, and the resulting $>125 \mu\text{m}$ and 63–125 μm fractions were dried in an oven at 80°C. The 63–125 μm fraction was placed in bromoform with a measured specific gravity of 2.8. Heavy minerals were allowed to separate under gravity, with frequent stirring to ensure complete separation. The heavy mineral residues were mounted under Canada Balsam for optical study using a polarising microscope. Where possible, a split was retained for mineral chemical analysis.

Heavy mineral proportions were estimated

by counting 200 non-opaque detrital grains using the ribbon method described by Galehouse (1971). Identification was made on the basis of optical properties, as described for grain mounts by Mange and Maurer (1992). A qualitative assessment was also made of other components, such as diagenetic minerals, opaques and mica. Provenance-sensitive mineral ratios (Morton and Hallsworth, 1994) were determined on the basis of a 200 grain count per mineral pair (where heavy mineral recovery allowed).

Garnet geochemical analysis by electron microprobe

Garnet grains were picked with a needle from dry residues during optical examination under the polarising microscope, placed on double-sided adhesive tape, coated with carbon, and analysed using a Link Systems AN 10/55S energy-dis-

Eocene	Ypresian	T60	
		T50	Balder Fm
		T45	Flett Fm
		T40	
	Thanetian	T38	Lamba Fm
		T36	
	Selandian	T35	Vaila Fm
		T34	
		T32	
		T31	
		T25-28	
		T22	
Paleocene	Danian	T10	Sullom Fm

Fig. 3.

Paleocene-Early Eocene stratigraphic framework of the Faroe-Shetland Basin, adapted from Ebdon *et al.* (1995), Knox *et al.* (1997), Mudge and Bujak (2001), Sørensen (2003) and Jolley *et al.* (2005).

persive x-ray analyser attached to a Cambridge Instruments Microscan V electron microprobe at the University of Aberdeen. The count time was 30 seconds for each grain. Data reduction used the ZAF4 programme. Results from poorly-oriented or rough grain surfaces were identified by low analytical totals and/or deviations from ideal

stoichiometry, and were discarded. Garnet compositions are expressed in terms of the molecular proportions of Mg, Fe²⁺, Ca and Mn, calculated on the basis of 24 oxygens, and normalised to total Fe+Mg+Ca+Mn, with all Fe calculated as Fe²⁺, as recommended by Droop and Harte (1995).

Rutile geochemical analysis by laser ablation inductively coupled plasma-mass spectrometry (LA-ICP-MS)

Rutile grains were picked with a needle from dry residues during optical examination under the polarising microscope, placed on double-sided adhesive tape, and analysed using a VG (Thermo) Elemental PlasmaQuad 3 and a VG (New Wave Research) MicroProbe II 266nm Nd:YAG laser at the University of Bristol. Laser beam diameter was 30µm and laser repetition rate 5Hz. Helium gas was the ablation medium. A carrier gas of helium, then helium plus argon, transported the sample to the ICP-MS. Time-resolved analysis (TRA) data acquisition software was used with a total acquisition time of 120s per analysis, allowing about 60s for background followed by 60s for laser ablation. Data reduction used GLITTER (LA-ICP-MS software package) and internal standards of Si for the glass standards and Ti for rutile (98%). Signals were checked for contamination from other phases or the mounting medium and the integration intervals adjusted accordingly. NIST 610 glass was used for instrument calibration, with NIST 612 used as a secondary check standard.

The great majority of rutiles are derived from either metamafic or metapelitic sources (Force, 1980; Zack *et al.*, 2004). Discrimination of rutiles derived from these two sources is achieved using their Cr and Nb contents (Zack *et al.*, 2004; Triebold *et al.*, 2007; Meinhold *et al.*, 2008). In this paper, we have classified rutiles with > 800 ppm Nb and Cr<Nb as metapelitic, and all others as metamafic, following the criteria proposed by Meinhold *et al.* (2008). Rutile formation temperatures have been estimated using the Zr content following the formula proposed by Watson *et al.* (2006).

Zircon age dating

Zircon grains were separated from heavy residues using heavy liquid (density 3.3 gm/cm³) and paramagnetic procedures. The zircon-rich heavy mineral concentrates were poured onto double-sided adhesive tape, mounted in epoxy together with chips of the reference zircons (Temora and SL13), sectioned approximately in half, and polished. Reflected light photomicrographs, transmitted light photomicrographs and cathodoluminescence (CL) scanning electron microscope (SEM) images were prepared for all zircon grains. The CL images were used to decipher the internal structures of the sectioned grains and to ensure that the ~20µm SHRIMP spot was wholly within a single age component (usually the youngest) within the sectioned grains.

The U-Th-Pb analyses were made using SHRIMP II and SHRIMP RG at the Research School of Earth Sciences, The Australian National University, Canberra, Australia, following the method described by Williams (1998, and references therein). For each sample, zircons on the mount were analysed sequentially and randomly until a total of at least 60 grains for each sample was reached. Each analysis consisted of 4 or 5 scans through the mass range, with a reference zircon analysed for every five unknown zircon analyses. The data have been reduced using the SQUID Excel Macro of Ludwig (2001). The U/Pb ratios have been normalized relative to a value of 0.0668 for the Temora reference zircon, equivalent to an age of 417 Ma (see Black *et al.*, 2003). Uncertainties given for individual analyses (ratios and ages) are at the one sigma level. Correction for common Pb was made using the measured ²⁰⁴Pb/²⁰⁶Pb ratio, except for grains younger than ~800 Ma (or those low in U and therefore low in radiogenic Pb), where the ²⁰⁷Pb correction method has been used (see Williams, 1998). When the ²⁰⁷Pb correction is applied, it is not possible to determine radiogenic ²⁰⁷Pb/²⁰⁶Pb ratios or ages. In general, for areas younger than ~800 Ma (and for areas that are low in U and therefore low in radiogenic Pb), the radiogenic ²⁰⁶Pb/²³⁸U age has

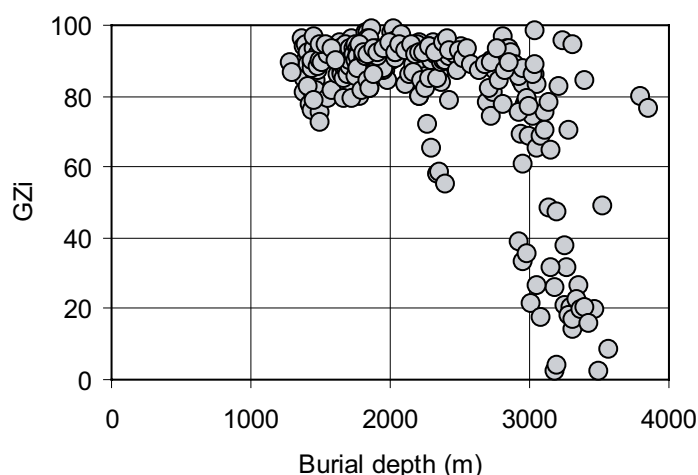


Fig. 4.

Relationship between garnet:zircon index (GZi) and present-day burial depth in Paleocene samples from the Faroe-Shetland Basin.

been used for the probability density plots. The ²⁰⁷Pb/²⁰⁶Pb age is used for grains older than 800 Ma, or for younger grains enriched in U. Tera and Wasserburg (1972) concordia plots and probability density plots with stacked histograms were generated using ISOPLOT/EX (Ludwig, 2003).

Palynology

Although the principal aim of palynological analysis of the Faroe-Shetland Basin fill has been to provide constraints on stratigraphy, the technique has also provided important information on sediment provenance. This is because the terrestrial palynomorphs (pollen and spores) that are preserved in marine basins are derived from vegetation growing within catchment areas at the basin margins. In marine basins with salinity stratification, or where there is major input from large fluvial systems, the pollen and spore load may be transported for considerable distances offshore. Such conditions occurred in the NE Atlantic during the Paleocene and Early Eocene, resulting in the offshore transfer of large volumes of terrigenous plant debris (Naylor *et al.*, 1999).

Regional differences in ecological conditions around the basin margins enabled Jolley *et al.* (2005) and Jolley and Morton (2007) to differentiate a westerly-derived palynoflora from

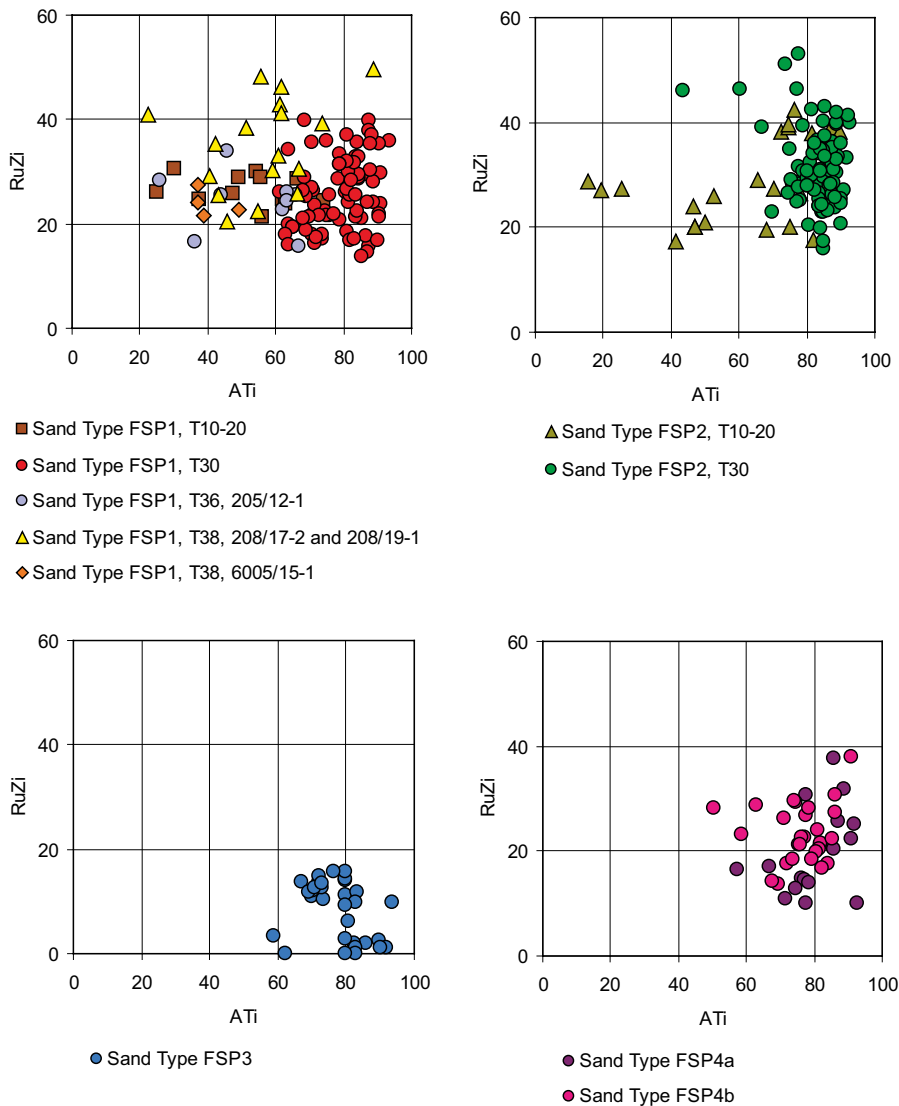


Fig. 5. Plots of RuZi and ATi for sand types FSP1, FSP2, FSP3 and FSP4.

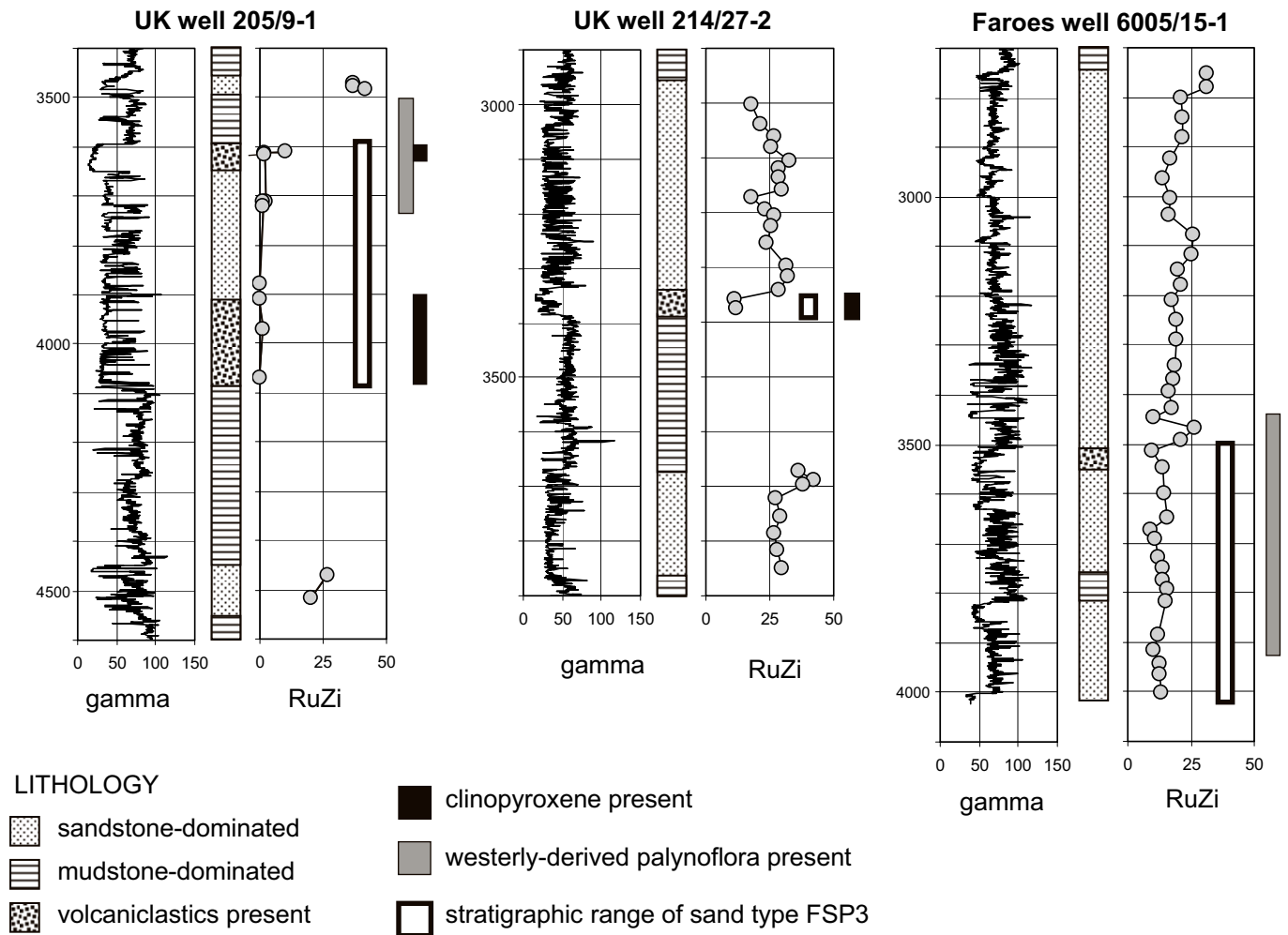
palynofloras sourced from the east. Palynofloras derived from the westerly source are characterised by the common occurrence of *Cupuliferoipollenites* and *Cupuliferoideaepollenites* (Fagaceae) in association with *Momipites* species (Juglandaceae) (Jolley *et al.*, 2005; Jolley and Morton, 2007). However, since terrestrially-derived palynomorphs fall into the size range of 5-60 μm , they have hydrodynamic properties comparable with silt size sedimentary particles. In consequence, it is likely that westerly-derived palynoflora will be more widely distributed within the Paleocene succession of the Faroe-Shetland Basin compared with westerly-derived arenaceous material.

Discrimination of sand types

Four distinct sand types have been identified in the Paleocene succession of the Faroe-Shetland Basin, on the basis of variations in heavy mineral assemblages, mineral chemistry, and terrestrially-derived palynoflora. These were termed sand types I-IV by Jolley and Morton (2007), but in this paper they have been renamed as FSP1-4 in order to associate them with their specific location (Faroe-Shetland Basin) and age (Paleocene).

Conventional heavy mineral data

Heavy mineral assemblages are controlled not only by provenance, but also by processes that

**Fig. 5.**

RuZi profiles for Faroes well 6005/15-1 and UK wells 205/9-1 and 214/27-2, and the relationship with the presence of clinopyroxene and westerly-derived palynoflora. Note that palynological data are not available for UK well 214/27-2. Also note that although clinopyroxene is present in Faroes well 6005/15-1, it is uncertain whether this mineral is present as a detrital phase or whether it was derived from interbedded basalt flows, volcaniclastics or intrusions. Vertical axis is measured depth in metres.

operate during the sedimentary cycle, the most important being diagenesis during deep burial, hydrodynamic processes during transport and deposition, and weathering (Morton, 1985; Mange and Maurer, 1992; Morton and Hallsworth, 1999; Mange and Wright, 2007; Garzanti *et al.*, 2008). The effects of the processes operative during the sedimentary cycle can be minimised by determining the relative abundance of minerals that are hydrodynamically-equivalent and are stable within the context of the study (Morton and Hallsworth, 1994).

In the Paleocene of the Faroe-Shetland Basin, major variations are seen in three ratios, garnet:zircon (GZi), rutile:zircon (RuZi), and apatite:tourmaline (ATi), as defined by Morton and Hallsworth (1994). Variations in GZi, however, appear to be related primarily to burial depth (Fig. 4), implying that garnet has undergone partial or complete dissolution through the action of high-temperature pore waters. Evidence for garnet dissolution is also provided by the presence of surface corrosion textures on garnet similar to those illustrated by Turner and Morton (2007). Hence, GZi values cannot be used to reli-



Plots of RuZi vs. ATi values (Fig. 5) show that sand types FSP1, FSP2 and FSP4 are characterised by relatively high RuZi values, whereas sand type FSP3 has low values. FSP1, FSP2 and to a lesser extent FSP4 have wide ranges in ATi, whereas FSP3 has uniformly high ATi. Hence, FSP1, FSP2 and FSP4 are indistinguishable in terms of provenance-sensitive ratio characteristics, but can be distinguished from FSP3 on the basis of their RuZi values.

is interpreted as having a western provenance. However, since clinopyroxene is one of the most unstable heavy minerals during burial diagenesis (Morton and Hallsworth, 2007), the absence of clinopyroxene does not preclude derivation from the west.

Differences in garnet assemblages in Paleocene-Eocene sandstones in the Faroe-Shetland Basin are manifested by changes in the relative proportions of low-Ca, high-Mg garnets compared with low-Mg garnets containing variable amounts of Ca and Mn (respectively, types A and B in the terminology of Morton *et al.*, 2004). High-Ca, high-Mg garnets (Type C) form a minor component of the garnet populations and show no discernible regional or stratigraphic pattern. As noted by Jolley and Morton (2007), there are distinct variations in sandstones rich in Type A garnet, some of which are dominated by Type Ai garnets (those with molecular proportions of Mg

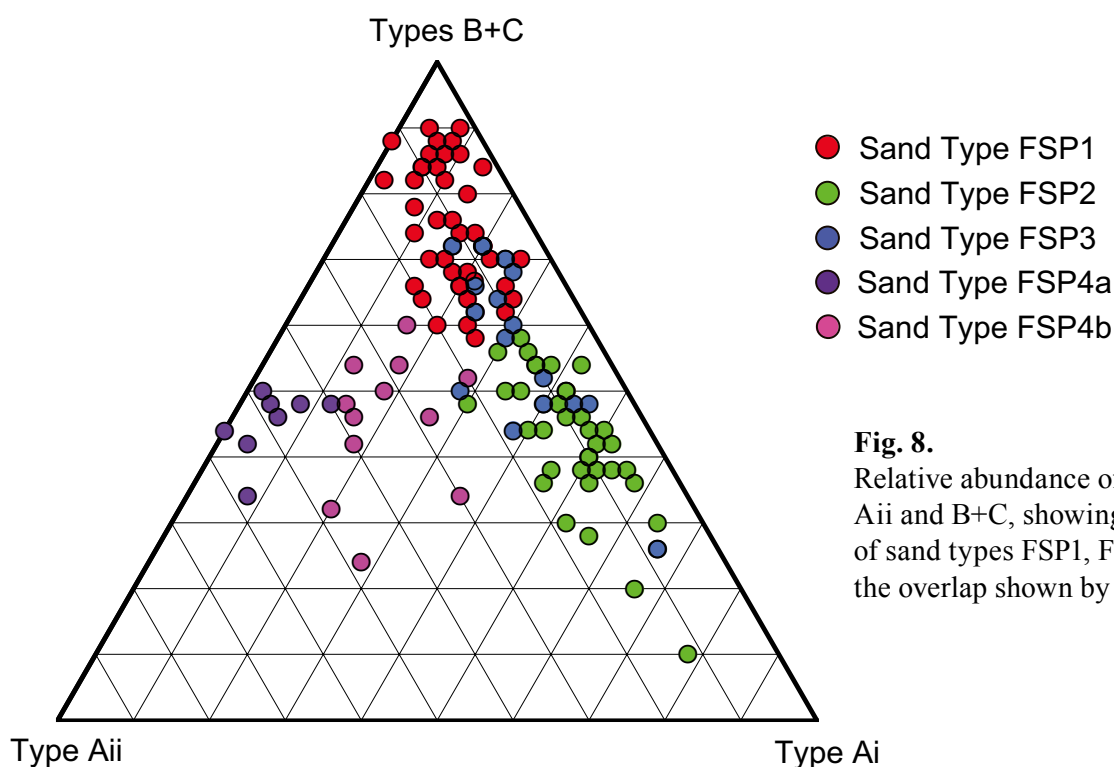


Fig. 8.
Relative abundance of garnet types Ai, Aii and B+C, showing the differentiation of sand types FSP1, FSP2 and FSP4, and the overlap shown by FSP3.

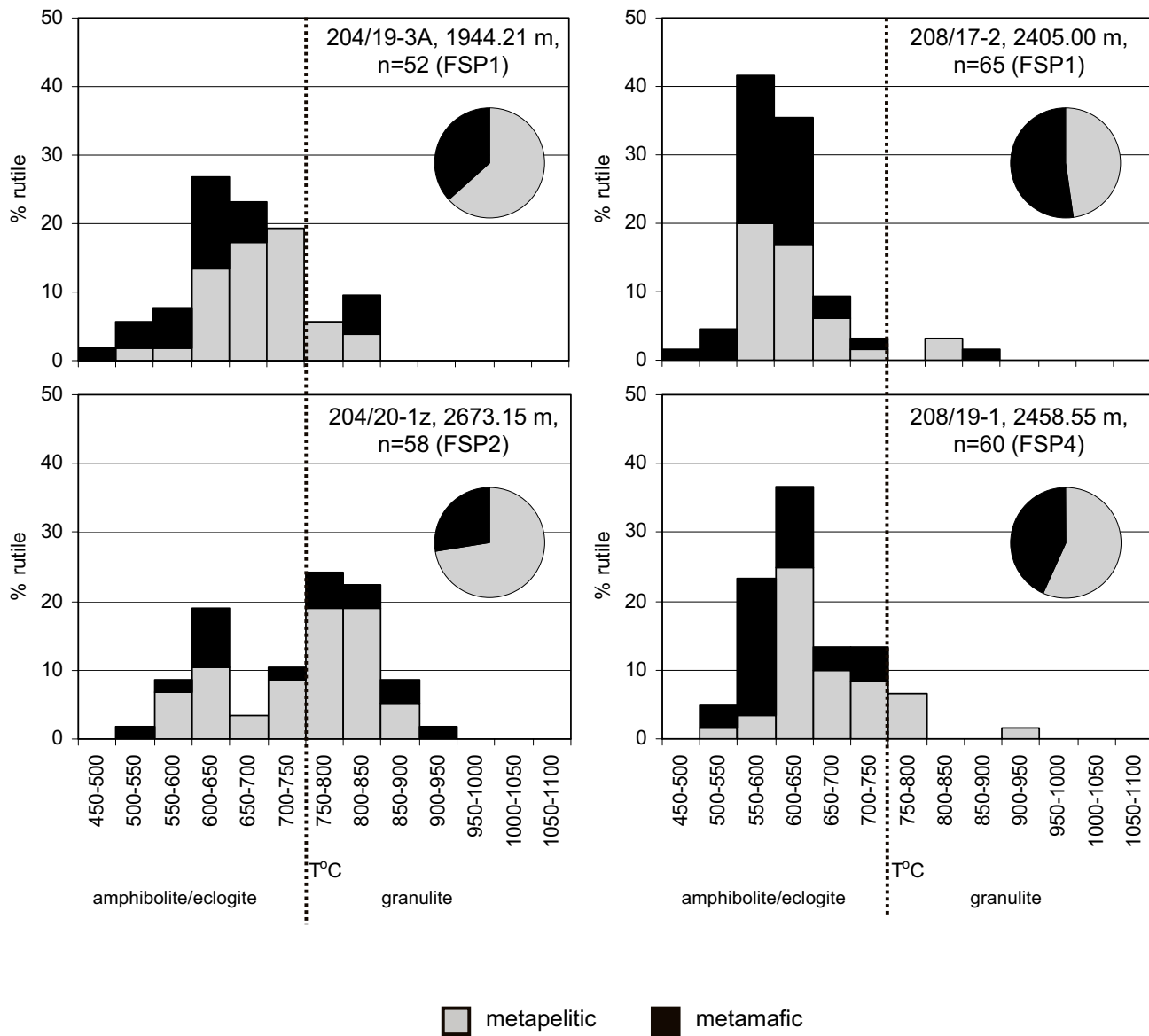
> 30%), whereas others are dominated by Type Aii (garnets with molecular proportions of Mg between 20% and 30%). Representative ternary diagrams illustrating variations in abundance of Type Ai, Type Aii and Type B garnets within the data set are shown in Fig. 7.

Garnet geochemistry is the key parameter for distinguishing sand types FSP1, FSP2 and FSP4. These sand types cannot be differentiated on the basis of their RuZi or ATi values (Fig. 5), but have distinctly different garnet assemblages defined by variations in the relative abundance of types Ai, Aii and B+C garnets (Fig. 8). Sand type FSP1 is dominated by Type B garnet, with B+C > ~70%. Sand type FSP2 is characterised by common Type Ai garnet, which forms between 30-60% of the populations in most cases. A small number of samples have higher Type Ai contents, but these are all from deeply-buried sandstones that have undergone partial garnet dissolution. Because Ca-rich garnets are less stable than Ca-poor varieties during burial diagenesis (Morton, 1987; Morton and Hallsworth, 2007), the relative abundance of stable garnets (such as Type A) tends to increase as burial diagenesis proceeds. Sand type FSP4 also has common Type A garnet

(> 30%), but in this case they are predominantly Type Aii. Sand type FSP4 has been subdivided into two subtypes, FSP4a and FSP4b, on the basis of their Ai garnet content. FSP4a has low proportions of Type Ai garnet, whereas FSP4b has higher Type Ai contents. Finally, sand type FSP3 cannot be distinguished from FSP1 and FSP2 on the basis of garnet geochemistry, since FSP3 sandstones have garnet assemblages that overlap the FSP1 and FSP2 fields.

Rutile geochemistry

There are variations in rutile provenance in the Paleocene of the Faroe-Shetland Basin. These differences are expressed in terms of both source lithotype (metamafic or metapelitic) and metamorphic grade (Fig. 9). Sand type FSP2 is distinguished from FSP1 and FSP4 by high abundances of granulite-facies metapelitic rutile (Fig. 10). FSP1 and FSP4 do not have markedly different characteristics, although the available data suggest FSP1 has somewhat higher abundances of granulite-facies rutile compared with FSP4. However, the FSP1 sandstones in the T38 interval from wells 208/17-2 and 208/19-1, located in

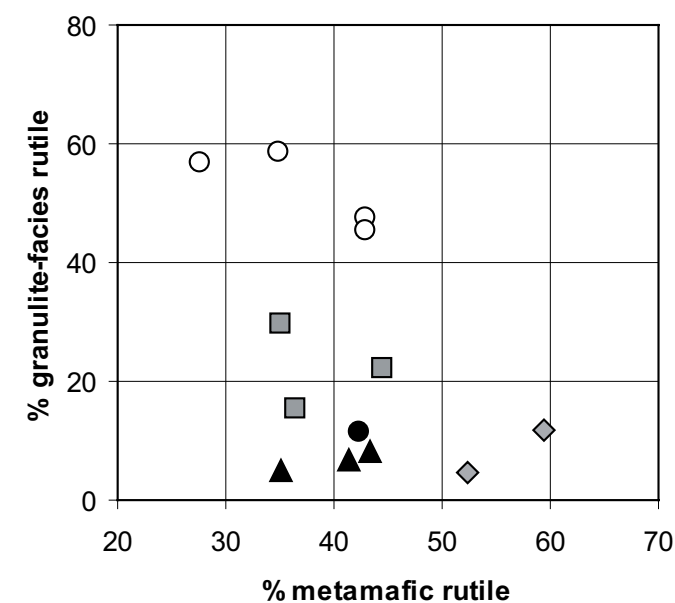
**Fig. 9.**

Rutile geochemistry for sand types FSP1, FSP2 and FSP4. Note the contrast in rutile provenance between the two FSP1 samples, one from the southern part of the area (UK well 204/19-3A) and one from the north (UK well 208/17-2).

the northeastern part of the study area (Fig. 2), are different in having much higher abundances of metamafic rutile (Fig. 10). These sandstones were clearly derived from a different source compared with the FSP1 sandstones further south, and this may also account for their lower ATi values (Fig. 5). Rutile geochemical data cannot be used to discriminate sand type FSP3 owing to the relative scarcity of rutile.

The abundance of granulite-facies metapelite rutile in sand type FSP2 correlates with the abundance of Type Ai garnet (Fig. 11). The main

source of Type A garnet is believed to be high-grade metasediments and charnockites (Sabeen *et al.*, 2003; Mange and Morton, 2007), and a link between abundance of Type A garnet and granulite-facies metapelite rutile was recently established in a study of Jurassic-Paleocene sandstones in the Norwegian Sea (Morton and Chenery, 2009). However, the rutiles in sand type FSP4, which are characterised by Type Aii garnet, are predominantly metamafic and formed under amphibolite-facies (principally lower amphibolite-facies) conditions. The results of



■ FSP1 ◆ FSP1 (north) ○ FSP2 ▲ FSP4a ● FSP4b

Fig. 10.

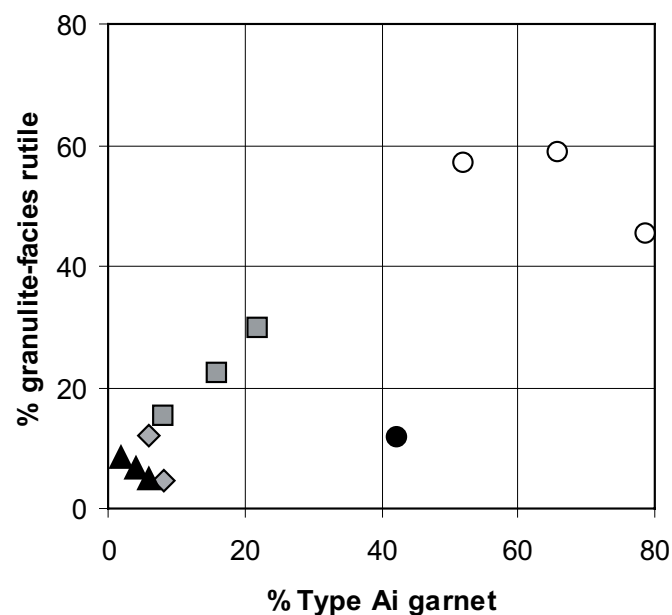
Variations in abundance of metamafic rutile and granulite-facies rutile in sand types FSP1, FSP2 and FSP4.

the current study therefore show that Type Aii garnets do not necessarily form under high-grade (granulite-facies) conditions, and may on occasions have a metamafic provenance.

Zircon age data

Differences in Paleocene sand provenance in the Faroe-Shetland Basin have also been explored using detrital zircon age data. Seven samples covering the four sand types were included in the zircon analytical programme (Fig. 12). In broad terms, the seven samples have similar age spectra, in that all contain a well-defined Archaean group peaking in the 2700-2800 Ma range, together with a range of Paleoproterozoic-Mesoproterozoic zircons (~1000-1900 Ma) that generally lack a well-defined age structure. In addition, there is a small number of Early Paleozoic zircons corresponding to the Caledonian Orogeny, and in some cases a small number of grains with Permo-Triassic ages are present.

The main variation within the zircon data set



■ FSP1 ◆ FSP1 (north) ○ FSP2 ▲ FSP4a ● FSP4b

Fig. 10.

Relationship between abundance of granulite-facies rutile and Type Ai garnet in the Paleocene of the Faroe-Shetland Basin.

relates to the abundance of Archaean grains. In most cases, Archaean zircons form ~30-40% of the populations, but in two samples, they form a significantly higher proportion (63-65%). These two samples (UK well 205/9-1, 3712.40 m; Faroes well 6104/21-1, 3777 m) have FSP3 mineralogy, being characterised by low RuZi values. The zircon age data therefore confirm the heavy mineral evidence for a difference in provenance between FSP3 and FSP1, 2 and 4.

Palynology

Palynological data are available for two of the wells that contain sand type FSP3, UK well 205/9-1 and Faroes well 6005/15-1 (Jolley *et al.*, 2005; Jolley and Morton, 2007). In UK well 205/9-1 (Fig. 6), the westerly-derived palynoflora, recognised by the common occurrence of *Cupuliferoipollenites* and *Cupuliferoidaepollenites* (Fagaceae) in association with *Momipites* species (Juglandaceae), characterises the upper part of the thick low-RuZi sand of T35-T36 age

(Jolley and Morton, 2007), whereas in Faroes well 6005/15-1 (Jolley *et al.*, 2005), it occurs in the T31-T34 section. Other records of the west-erly-derived palynoflora, notably those found in the Vaila and Lamba Formations of the Foinaven area (Jolley and Morton, 2007), are not associated with the FSP3 sand type. These occurrences indicate a decoupling of sediment transport, with coarse clastic sediment being supplied from the east and at least some of the fine-grained sediment being transported from the west (Jolley and Morton, 2007). Clearly, this requires two different depositional mechanisms. The Paleocene sandstones in the Foinaven area are known to represent the products of submarine fan systems derived from the Orkney-Shetland margin (Lamers and Carmichael, 1999), but the pollen and spores, being hydrodynamically-equivalent to silt-size material, are believed to represent the distal products of fluvial systems that entered the basin from both margins.

The palynoflora have also proved useful for established lateral differences in provenance characteristics of sediments derived from the Orkney-Shetland margin, substantiating evidence for differences in heavy mineral assemblages. The palynoflora evidence is especially important in revealing sediment recycling and constraining the nature of the sediment source. For example, sand type FSP4 is associated with reworked Namurian-Westphalian palynofloras (Jolley and Morton, 2007), suggesting that at least some of the FSP4 sands were recycled from the Carboniferous of the Orkney-Shetland Platform.

Provenance and distribution of Paleocene sandstones

Sand type FSP1

Sand type FSP1 is characterised by high abundances of Type B garnet, moderate RuZi and high GZi (except for the most deeply-buried sandstones, where garnet dissolution has been pervasive). ATi is normally high, although it is lower in most of the T10-T20 sandstones (Fig.

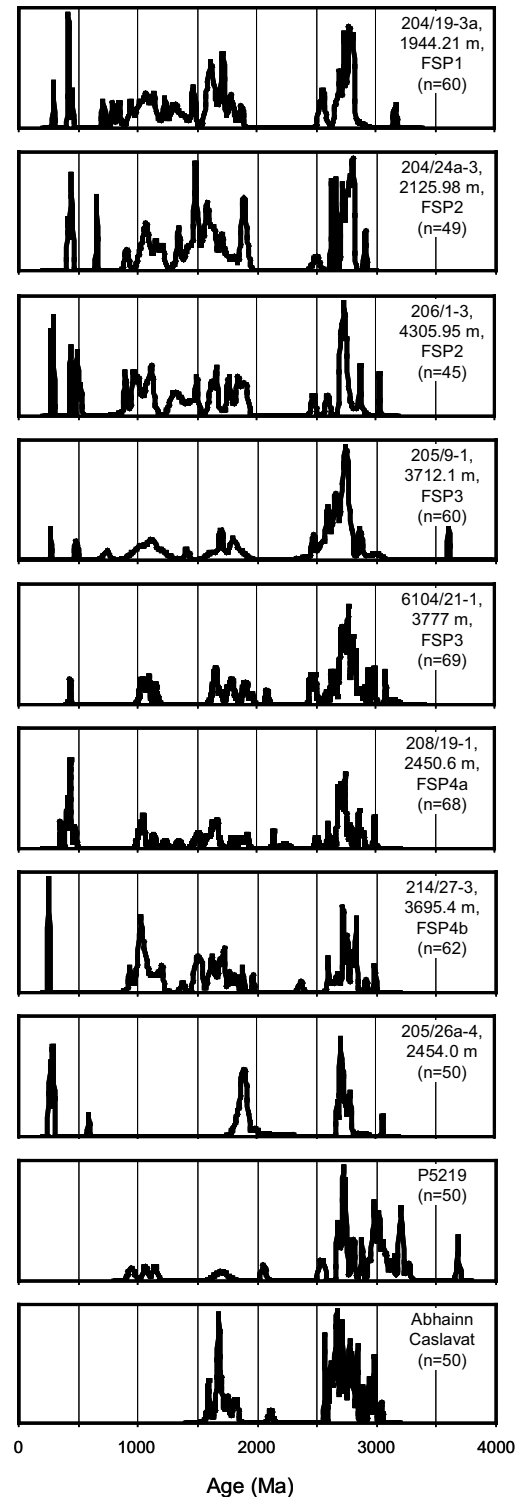


Fig.12.

Zircon age spectra in FSP1, FSP2, FSP3 and FSP4 sandstones displayed as relative probability plots. Also shown is the zircon age spectrum for the Foula Formation (Triassic) in UK well 205/26a-4 (Morton *et al.*, 2007), CASP sample P5219 from the Eocene succession in Kangerlussuaq, East Greenland (Whitham *et al.*, 2004) and a river sand (Abhainn Caslavat) sourced from the Tarbert Terrane (Lewisian) on the Isle of Lewis (Outer Hebrides).

5), and low values also occur locally in the T36-T38 interval. Rutile compositions indicate supply from both metapelitic and metamafic rocks, with metapelitic rutiles more abundant than metamafic types. Most of the rutiles were derived from amphibolite-facies rocks, but a small proportion (15-30%) originated in granulite-facies rocks. FSP1 sandstones that have not undergone deep burial have high proportions of unstable heavy minerals, including amphibole, epidote and titanite. Abundant unstable minerals in FSP1 sandstones have been observed both in the Foinaven area (Morton *et al.*, 2002) and further north, for example in the T36 interval of 205/9-1 and 214/27-2. The presence of abundant unstable minerals is a strong argument for predominantly first-cycle derivation from basement rocks, and although the presence of some recycled material also seems likely.

The high abundance of Type B garnet is compatible with derivation from the metamorphic basement that forms the platform areas along the Orkney-Shetland margin (Fig. 1). The basement in this area is comparatively poorly known, but comprises equivalents of the Lewisian Gneiss and Moine/Dalradian metasediments of onshore Scotland and Shetland (Stoker *et al.*, 1993). Garnets in modern river sediments derived from these basement lithologies belong almost exclusively to Type B (Morton *et al.*, 2004). Rutile compositions suggest that both Lewisian and Moine/Dalradian sources were involved, since the Moine and Dalradian consist mainly of metasediments, and are therefore unlikely to supply a substantial number of metamafic types.

The majority of zircons are Paleoproterozoic-Mesoproterozoic in age (Fig. 12), and these are interpreted as indicating derivation from Moine and/or Dalradian metasediments, since most zircons in Moine and Dalradian metasediments from onshore Scotland fall in this range (Cawood *et al.*, 2003; Cawood *et al.*, 2004). However, a significant number of zircons (33%) were derived from the Archaean, either directly or from younger metasedimentary basement with a substantial Archaean detrital component. The

Moine of northern Scotland lacks Archaean zircons (Cawood *et al.*, 2004), but Archaean grains are common in parts of the Dalradian Supergroup of Scotland (Cawood *et al.*, 2003) and Shetland (Cutts *et al.*, 2009).

The provenance of FSP1 sandstones in the northern part of the study area (sampled in UK wells 208/17-2 and 208/19-1) appears to be significantly different to those in the south, since they have higher proportions of metamafic rutile. In the absence of zircon age data, it is impossible to speculate whether this is related to changes in the relative supply from Lewisian and Moine/Dalradian sources. However, there is some evidence that the basement in the northern part of the Shetland Platform contains metamafic rocks, such as those drilled in UK well 220/26-1 and BGS borehole 81/17 (Stoker *et al.*, 1993).

The low ATi values seen in many FSP1 sandstones from the T10-20 interval are interpreted as reflecting supply of sediment eroded from the weathered Late Cretaceous land surface (Morton *et al.*, 2002). Lowering of ATi takes place during surficial weathering because apatite is highly unstable under such conditions, whereas tourmaline is stable (Morton and Hallsworth, 1999). The change to consistently high ATi values in the T31-T34 interval is interpreted as indicating removal of the weathered material and erosion of fresh bedrock. The trend from erosion of weathered to unweathered source lithologies is best illustrated with UK well 204/24a-7, which shows a clearly-defined upward increase in ATi from T10 into T30 (Fig. 13). Deep kaolinitic regoliths on Archaean basement have been found in BGS boreholes along the Hebridean margin, the thickest (> 18 m) being in borehole 77/7 (Evans *et al.*, 1997).

Some FSP1 sandstones show a return to low ATi characteristics in the T36-T38 interval (Fig. 5). Low ATi values occur in T36 sandstones in UK well 205/12-1, T38 sandstones in UK wells 208/17-2 and 208/19-1, and T38 sandstones in Faroes well 6005/15-1. It is difficult to attribute these low ATi values to supply from the weathered Late Cretaceous land surface, since the

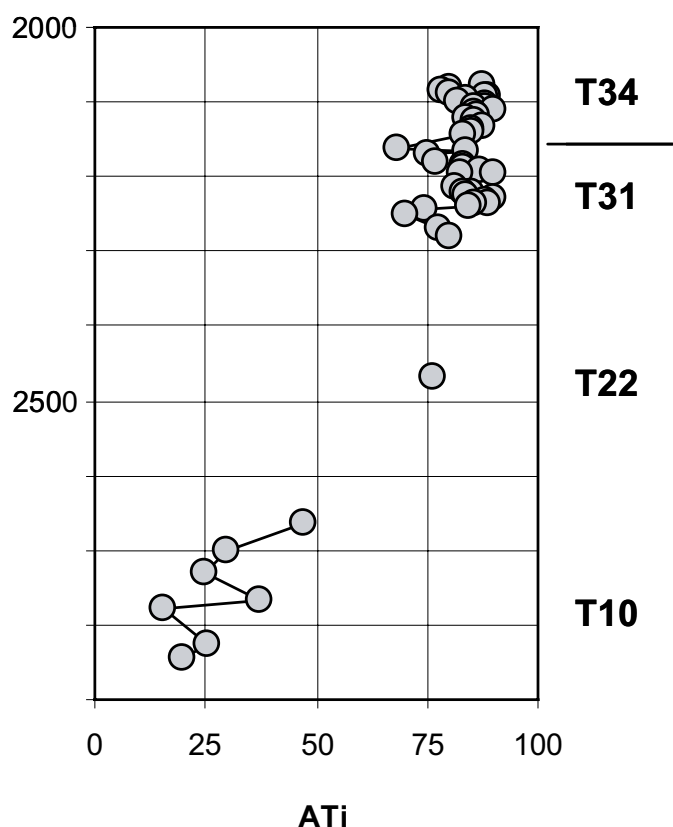


Fig.13.

Apatite:tourmaline index (ATi) profile through the T10-T34 succession in UK well 204/24a-7, showing an upward increase in ATi indicative of the gradual replacement of weathered detritus by fresh detritus. Vertical axis is measured depth in metres.

absence of low ATi values in the T31-T34 interval suggests this had already been stripped off. In UK wells 208/17-2 and 208/19-1, the difference in rutile geochemistry described above indicates the T38 sandstones had a different provenance to the FSP1 sandstones further south, and this could also account for their lower ATi. However, there is no evidence that the T36 sandstones in UK well 205/12-1 or the T38 sandstones in Faroes well 6005/15-1 had a different source, and the most likely explanation for the low ATi in these wells is that the sediment underwent weathering during periods of alluvial storage prior to entering the marine environment.

FSP1 sandstones are of relatively minor occurrence in the T10-T20, T31-32 and T34 intervals being found only in a small number of wells from the southern part of the area (Fig. 14). The sand type becomes more widespread in T35-

T36, being recorded in virtually every well that encountered sand in this stratigraphic interval. This pattern is maintained in the T38 interval, during which time the sand type shows its maximum regional distribution, extending north into UK wells 208/17-2 and 208/19-1. However, as discussed above, the FSP1 sandstones in these two wells have a different basement provenance to those found further south, manifested by a difference in rutile composition.

Sand type FSP2

Sand type FSP2 is characterised by abundant Type Ai garnet, moderate RuZi and high GZi (except for the most deeply-buried sandstones, where garnet dissolution has been pervasive). ATi is generally high, but many of the T10-T20 sandstones have lower values (Fig. 5). Rutile assemblages are dominated by granulite-facies metapelitic types, in contrast to the rutiles associated with sand types FSP1 and FSP4. Unstable minerals such as amphibole, epidote and titanite are relatively uncommon in FSP2 sandstones, although this may be partly because FSP2 sandstones are most widespread in the earlier part of the Paleocene, and hence tend to be more deeply buried than FSP1.

The abundance of Type Ai garnet and of granulite-facies metapelitic rutile indicates that a high proportion of the detritus forming the FSP2 sand type was derived from high-grade (granulite facies) metasedimentary rocks. Garnet geochemical studies of modern river sediments indicate that Type Ai garnets are scarce in the Lewisian, Moine and Dalradian of northern Scotland and Shetland (Morton *et al.*, 2004), precluding derivation from such lithologies. Although it is possible that granulite facies metasedimentary rocks are present in submerged areas of basement of the NW European margin, it is more likely that the Type Ai garnets and granulite-facies rutiles were recycled from the Triassic, based on data from the Foula Formation in the Strathmore Field close to the margin of the Faroe-Shetland Basin. Sandstones of the Foula Formation, which is

Ladinian-Carnian in age (Swiecicki *et al.*, 1995), have garnet populations rich in the Type Ai component (Morton *et al.*, 2007). Furthermore, they have rutile populations with abundant granulite-facies metapelitic types (Fig. 14). The Foula Formation is believed to have a source to the west, the most likely area being in East Greenland, although submerged highs in the Faroe-Shetland Basin cannot be ruled out as potential sources (Morton *et al.*, 2007).

Having identified the Foula Formation as a significant contributor to sand type FSP2, it is necessary to invoke additional supply from other sources. Input from a zircon-rich source is required since GZi values in the Foula Formation are extremely high, with a mean value of 98. This is significantly higher than in FSP2, which has a mean value of 89 (excluding sandstones that have undergone significant garnet dissolution). Likewise, RuZi values are higher in the Foula Formation than in FSP2 (54, compared with 31). The most likely source of zircon-rich detritus is the Devonian Old Red Sandstone (ORS), which is exposed widely over the Orkney-Shetland Platform (Stoker *et al.*, 1993). The ORS of this area has diagenetically-stable heavy mineral assemblages that are generally devoid of garnet, and also has low RuZi values (unpublished HM Research data). It is also necessary to invoke a source with Type B garnets, since the Foula Fm has higher proportions of Type Ai garnet (mean of 65%) than FSP2 (mean of 47%). The basement lithologies that are the dominant source for FSP1 are the most likely sources of such material.

The zircon age spectra associated with FSP2 (Fig. 12) are similar to that for FSP1, having a dominant wide-ranging Paleoproterozoic-Mesoproterozoic group and a subordinate Archaean group. This reflects a broadly similar ultimate provenance, although in this case much of the sediment is believed to be recycled from the ORS and the Foula Fm. The Foula Fm has a distinctive zircon age spectrum (Fig. 12), characterised by an Archaean group, a well-defined Paleoproterozoic peak at ~1750-1900 Ma, and a Permo-Triassic group. These three peaks are pre-

sent in the FSP2 spectra, although the 1750-1900 Ma and Permo-Triassic peaks are more muted owing to the abundance of detritus from other sources. The ORS recycling hypothesis cannot be tested by zircon age data at present, owing to the lack of published information on the succession in this area.

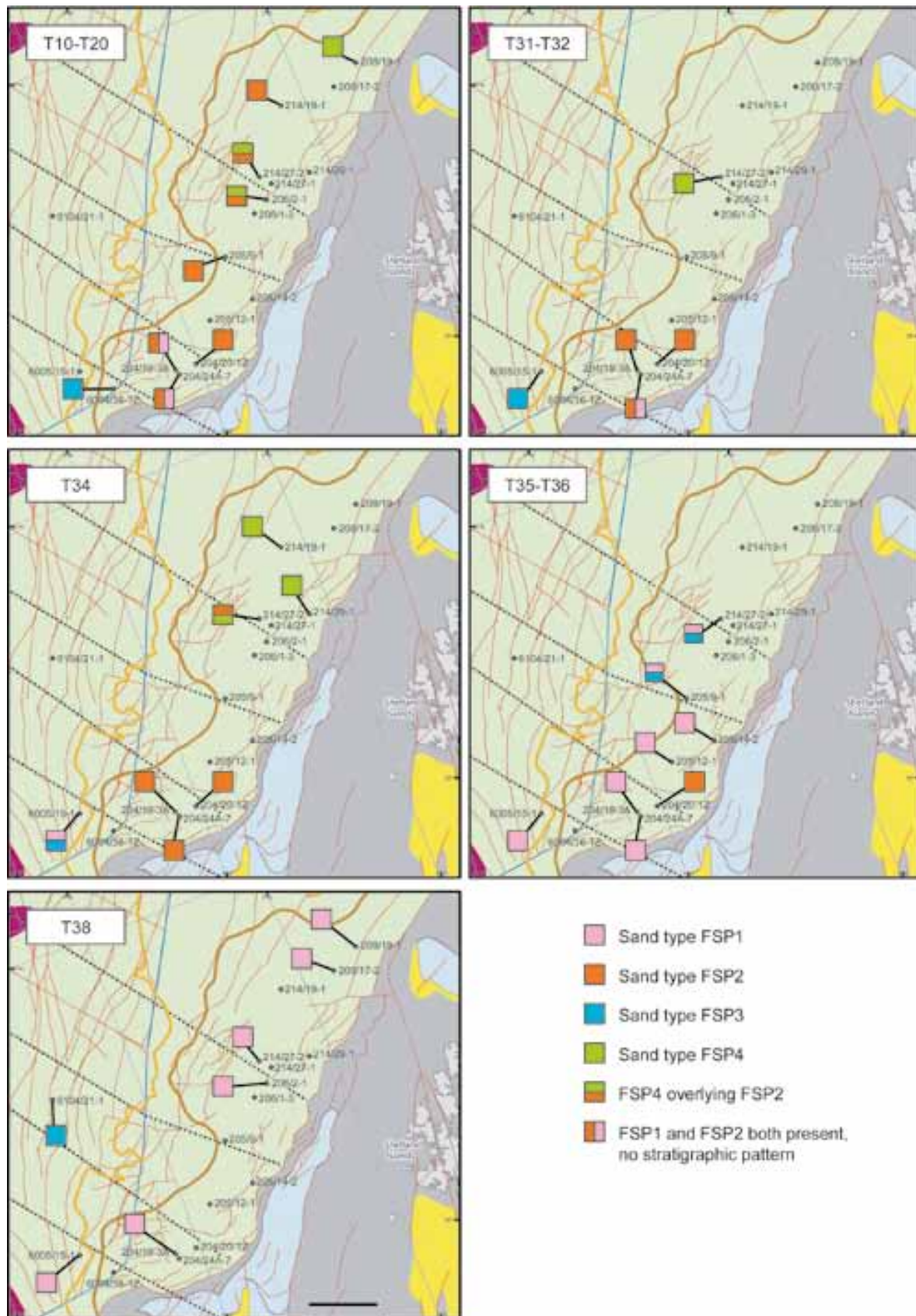
The low ATi values seen in some T10-T20 FSP2 sandstones are interpreted as reflecting supply of sediment eroded from the weathered Late Cretaceous land surface (Morton *et al.*, 2002), as described above for FSP1 sandstones in the same stratigraphic interval.

Sand type FSP2 is widely distributed across the study area in the T10-20, T31-32 and T34 intervals (Fig. 14), but becomes much less common in T36 (present only in UK well 204/20-1z), and has not been identified in any T38 sandstones.

Sand type FSP3

Sand type FSP3 is distinguished from FSP1, 2 and 4 on the basis of low RuZi values, the association with clinopyroxene, the presence of the westerly-derived palynoflora, and relatively high proportions of Archaean zircons. The abundance of Type B garnet is relatively high, but garnet populations overlap with those associated with sandstone types FSP1 and FSP2, and are therefore not diagnostic. In addition to clinopyroxene, other unstable minerals such as epidote and calcic amphibole are common in sandstones that have not suffered severe diagenetic modification.

The combination of low RuZi and high Type B garnet content is consistent with derivation from predominantly gneissic basement, and the zircon age data indicate that a large proportion of the detritus is Archaean. The presence of epidote and amphibole also suggests input from basement rocks. This combination of lithology and age is superficially consistent with derivation from Lewisian rocks of NW Scotland and the Outer Hebrides. However, it is now known that the onshore Lewisian comprises a number of terranes, each with a different geological history (Kinny *et al.*, 2005). Many of these terranes have

**Fig. 14.**

Distribution of sand types FSP1, FSP2, FSP3 and FSP4 in the T10-T20, T31-T32, T34, T35-T36 and T38 sequences of the Faroe-Shetland Basin. See Figs 1 and 2 for detail of structural features. Scale bar (T38) is 40 km.

age ranges that do not correspond with the main peak (~2700–2800 Ma) seen in the FSP3 sandstones. For example, one of the largest terranes in the Lewisian Complex, the Tarbert Terrane (Outer Hebrides) contains plutonic rocks of tonalite-trondhjemite-granodiorite (TTG) affinity dated as ~2850–3100 Ma and granites dated as ~1675 Ma (Kinny *et al.*, 2005). The zircon age spectrum in river sand derived from the Tarbert Terrane (Fig. 12) displays evidence for both the ~2850–3100 Ma and ~1675 Ma events identified by Kinny *et al.* (2005), but also has a large group between ~2550 Ma and ~2850 Ma, and bears little direct resemblance to the zircon age spectrum in FSP3 sandstones.

The lack of detailed information on the geochronology of the offshore equivalents to the Lewisian is a major problem in assessing if FSP3 sandstones could have been derived from the Archaean rocks of the Orkney-Shetland margin. Rb-Sr whole-rock analyses of basement rocks around the Rona Ridge yielded a latest Archaean age (2527 ± 73 Ma) and unspecified 'Lewisian' ages have been reported from other wells in the same vicinity (Ritchie and Darbyshire, 1984; Ritchie *et al.*, 1987), but this information is inadequate to enable comparisons with the detrital zircon age spectra. A reassessment of the geochronology of basement rocks in the Faroe-Shetland area, concentrating on zircon age data, is urgently needed to address this knowledge gap.

Although the available data do not rule out derivation from the eastern margin of the basin, the association of sand type FSP3 with clinopyroxene of basic igneous origin and with a westerly-derived palynoflora indicates that the source is much more likely to have lain to the west. Furthermore, there is no obvious sand entry point on the Orkney-Shetland Platform margin for the T35–T36 sandstones in UK well 205/9-1 (Smallwood *et al.*, 2004), which represents the thickest development of FSP3 sandstones in the wells analysed to date. Magnetic fabric measurements on the FSP3 sandstones in UK well 205/9-1 suggested transport from the NE (Smallwood *et al.*,

2004). However, since there is no obvious detrital source to the NE, Jolley and Morton (2007) considered that the magnetic fabrics in UK well 205/9-1 relate to local basin-floor topography and do not give any useful guide to source location.

Sandstones with low-RuZi mineralogy, similar to FSP3, have been found in the Paleogene of the Kangerlussuaq area of East Greenland (Whitham *et al.*, 2004). However, although the low-RuZi sandstones in the Kangerlussuaq area have predominantly Archaean zircons, there are significant differences in their age distribution compared with that in FSP3 (Fig. 12). The Kangerlussuaq low-RuZi sandstones have two main groups at ~2700–2750 Ma and ~2950–3100 Ma, with a subsidiary peak at ~3200 Ma, and over 50% of the Archaean grains are older than 2950 Ma (Whitham *et al.*, 2004). The ~2950–3100 Ma and ~3200 Ma peaks are either absent or very poorly developed in the FSP3 sandstones, and zircons older than 2950 Ma form only 5–11% of the Archaean population. Derivation of FSP3 sandstones from the Kangerlussuaq area of East Greenland can therefore be ruled out. Alternative locations towards the west must therefore be considered. Potential source areas include the sub-basaltic basement of the Blosseville Kyst of East Greenland and the Faroe Islands, together with intrabasinal highs such as the Corona Ridge, the Judd High, the Munkagrunnur Ridge, the Wyville-Thomson Ridge and the Rockall Plateau. Basement gneisses to the north of the Blosseville Kyst (between ~70°N and 72.5°N) appear to be a suitable source for the Archaean zircons in FSP3, since these are believed to have formed between ~2700–2800 Ma (Thrane, 2002).

Sandstone samples with FSP3 characteristics form a relatively small proportion (~10%) of the total number of samples in the data set. They have been found in the T10 succession in Faroes well 6004/16-1z, the T31–T34 succession in Faroes well 6005/15-1, the T35–T36 in UK well 205/9-1, the T36 succession in 214/27-2, and the T38 succession in 6104/21-1 (Fig. 14). It is noteworthy that FSP3 sandstones have not been encountered

in wells located close to the eastern margin of the basin, providing further evidence that the source area/areas lay towards the west.

Sand type FSP4

Sand type FSP4 is characterised by abundant Type Aii garnet, relatively low RuZi, high ATi and high GZi (Fig. 5). Two subtypes have been recognised on the basis of the abundance of Type Ai garnet, which is higher in subtype FSP4b than in FSP4a (Fig. 8). Rutile assemblages are dominated by amphibolite-facies types with metapelitic rutiles more abundant than metamafic varieties (Fig. 10). Unstable minerals are very scarce, even in the least deeply buried sandstones. Zircon age spectra contain fewer Archaean grains (31–41%) than FSP3 sandstones, but are virtually indistinguishable from those of FSP1 and 2 (Fig. 12). Recycled Namurian-Westphalian palynomorphs are found in FSP4 sandstones (Jolley and Morton, 2007).

FSP4 sandstones occur only in the northern part of the study area, in UK Quadrants 208 and 214 (Fig. 14). FSP4a and FSP4b sandstones have different geographical distributions, with FSP4b being found further south compared with FSP4a (Fig. 14). The distribution of FSP4 sandstones constrains their source location to the northern part of the Orkney-Shetland Platform, and the presence of recycled Namurian-Westphalian palynomorphs suggests recycling of mid to late Carboniferous sandstones. There is presently no record of mid to late Carboniferous sandstones on the Orkney-Shetland Platform, and it is therefore considered likely that these were removed during the Paleocene depositional phase. Another possibility is that FSP4 sandstones were recycled from Mesozoic sandstones, which were themselves derived by recycling of Carboniferous sandstones. Type Aii garnets are common at certain levels in the Clair Group (Devonian–Early Carboniferous) succession of the Clair Field (Morton *et al.*, 2010), and garnet assemblages rich in the Type Aii component typify north-

erly-derived sandstones of the Pennine Basin of England (Hallsworth *et al.*, 2000; Hallsworth and Chisholm, 2008). The increased abundance of Type Ai garnets in FSP4b suggests that these sandstones contain some sediment derived from the same source as FSP2. This suggestion is consistent with their geographical distribution, since they lie in an intermediate position between FSP4 sandstones in the north and FSP2 sandstones in the south (Fig. 13). FSP4 sandstones are present in T10–T34, but have not been found in the subsequent T35–T38 interval (Fig. 14).

Discussion

Four distinct sand types have been identified in the Paleocene succession of the Faroe-Shetland Basin on the basis of a number of criteria. The key heavy mineral parameters are the rutile:zircon index (RuZi), distribution of clinopyroxene, garnet geochemistry, rutile geochemistry and detrital zircon age patterns. Terrestrially-derived and recycled palynoflora place important additional constraints of provenance, although there is evidence for decoupling of pollen and sand provenance, related to their difference hydrodynamic behaviour.

The integrated approach followed in this study has been crucial in the recognition of these four sand types and to determine their provenance, because they cannot be distinguished on the basis of one single parameter. RuZi values distinguish FSP3 from FSP1, FSP2 and FSP4. Garnet geochemistry effectively distinguishes FSP1, FSP2 and FSP4, but FSP3 garnet populations overlap those of FSP1 and FSP2. Rutile geochemistry clearly differentiates FSP2 from FSP1 and FSP4, and may help distinguish FSP1 from FSP2, although insufficient data are presently available to test this possibility. Rutile geochemistry is difficult to apply to FSP3 owing to its low rutile content. Zircon geochronology distinguishes FSP3 from FSP1, 2 and 4, but do not readily enable differentiation of the other three sand types from one another.

Three of the four sand types (FSP1, FSP2 and FSP4) were derived from the eastern margin of the Faroe-Shetland Basin. By contrast, FSP3 has a westerly source, but despite some similar heavy mineral characteristics cannot be matched to a source in the Kangerlussuaq area of East Greenland, because of differences in detrital zircon ages. Possible sources include the sub-basaltic basement of the Blosseville Kyst and the Faroe Islands, and intrabasinal highs such as the Wyville-Thomson Ridge, the Judd High, the Corona Ridge, the Munkagrannur Ridge and the Rockall Plateau. Given the variable stratigraphic and geographical distribution of FSP3 sandstones, it seems likely that more than one source area was involved.

Smallwood (2008) estimated that 30% of the Paleocene sediments in the basin were sourced from the west on the basis of sediment volume balancing calculations. However, only 10% of the samples in the data set have FSP3 characteristics. This apparent contradiction can be mostly explained by the strong sampling bias due to the distribution of wells, which are mostly located close to the eastern margin of the basin. In the Faroes sector, the proportion is considerably higher: in Faroes well 6005/15-1, approximately 35% of the samples have FSP3 characteristics, and the T10 sandstones in Faroes well 6004/16-1z are also classified as FSP3. The scarcity of FSP3 sandstones in the analysed parts of the basin can be attributed to structural controls, with the axial Corona Ridge preventing westerly-derived sand from reaching the eastern parts of the basin, especially during the T10-T35 (Sullom and Vaila) interval. The only bypass potential in this interval is at the SW end of the Corona Ridge, where many of the wells with FSP3 sandstones are located. There is also a lineament adjacent to UK well 205/9-1 that could have offered a bypass route through the Corona Ridge allowing FSP3 sandstones to be deposited during T35-T36.

Another contributing factor to the observed bias is that heavy mineral data concentrate only on the coarse clastic component of the sediment

fill, and it is evident from studies of terrestrially-derived palynoflora that westerly-derived material is more widely distributed in the finer-grained components (siltstones and claystones) of the succession (Jolley and Morton, 2007). For example, the 'Greenland' flora is found in UK well 204/19-3A, but the sandstones in this well are classed either as FSP1 or FSP2.

There is a distinct difference between sand sourcing in the T10-T34 interval compared with T35-T38. During the T10-T34 interval, most of the sand shed from the Orkney-Shetland margin has FSP2 characteristics in the south and FSP4 characteristics in the north, both sand types being primarily recycled from Mesozoic and Paleozoic sandstones. By contrast, most of the T36-T38 sandstones have FSP1 mineralogy, principally sourced from basement lithologies (Moine/Dalradian and Lewisian and their offshore equivalents). The changeover from mostly recycled to mostly first-cycle detritus is associated with an influx of FSP3 sandstones and associated volcanoclastic detritus in the northern part of the area (UK wells 205/9-1 and 214/27-2). This change coincides with a change in depositional style, the T36-T40 sequences having a progradational character with the position of the shelf/slope break advancing significantly beyond the positions of underlying Cretaceous fault scarps (Ebdon *et al.*, 1995). Ebdon *et al.* (1995) related this change in depositional character to thermal doming and rifting in the North Atlantic area, leading to volcanism, changes in provenance, and the deposition of extensive base-of-slope fan systems.

Conclusions

The study of Paleocene sandstone provenance in the Faroe-Shetland Basin has had two important outcomes, one concerning transport directions and the other concerning stratigraphic evolution.

Most of the sandstones analysed in the course of the study can be tied back to sources on the Orkney-Shetland Platform to the east (sand types FSP1, FSP2 and FSP4), with only 10% of samples

having characteristics that suggest input from the west (sand type FSP3). This contrasts with the outcome of sediment volume balancing calculations, which suggest that 30% of sediment was supplied from the west (Smallwood, 2008). This is partly explained by sampling bias, since the data set is controlled by the distribution of wells, most of which are closer to the Orkney-Shetland margin than the Faroe-Greenland margin. In the Faroes sector, over 35% of the samples have FSP3 characteristics. The scarcity of FSP3 sandstones in the analysed parts of the basin is due to the presence of the axial Corona Ridge, which prevented westerly-derived sand from reaching the eastern parts of the basin during the T10-T35 (Sullom and Vaila) interval, except at the SW end of the ridge, where there was significant bypass potential. It is also possible that the imbalance may be due in part to differences in the nature of westerly- and easterly-derived sediment. Heavy mineral data, which provide information only on the coarse clastic components, indicate that most of the sand was derived from the east. However, palynological studies indicate that westerly-derived palynoflora are more widely distributed in the finer-grained components of the succession compared with the sand-rich intervals, implying that a larger proportion of the mud in the basin fill is of western origin. Although sand type FSP3 has a western source, it was not related to sandstones in the Kangerlussuaq area on account of differences in the zircon age distribution. The sub-basaltic basement of the Blossøeville Kyst and the Faroe Islands and intrabasinal highs such as the Corona Ridge, Judd High, the Wyville-Thomson Ridge, the Munkagrannur Ridge and the Rockall Plateau, must be considered as possible sources of FSP3 sediment.

Exploration for sub-basaltic Paleocene plays in the Faroe-Shetland Basin depends on the existence of sand fed from the west, since easterly-derived systems will tend to become less prospective towards the west. The available data suggest that westerly-derived sands are most common in the SW part of the basin (Faroes wells

6004/16-1z and 6005/15-1), although they have been found locally elsewhere (notably UK well 205/9-1). However, the data set is geographically biased because of the distribution of exploration wells, and it is therefore possible that the picture will change as more wells are drilled in the western part of the basin.

The easterly-derived sandstones show marked evolution in provenance characteristics with time. Most of the sandstones in the T10-T34 succession were recycled from pre-existing sandstones on the Orkney-Shetland Platform, with the Triassic and Old Red Sandstone supplying sand type FSP2 in the south and the Carboniferous (possibly recycled through the Mesozoic) supply sand type FSP4 in the north. Basement sources, which are poorly-represented in the T10-T34 succession, become prevalent in the T35-36 and T38 successions. The change in provenance coincides with a change in depositional style, the T36-T40 sequences being more progradational than the T10-T34 sequences, believed to be related to thermal doming, rifting and volcanism in the NE Atlantic (Ebdon *et al.*, 1995). The change in provenance coincides with a short-lived influx of sediment with FSP3 character, associated with volcanoclastic material, seen in the T35-T36 interval in 205/9-1 and 214/27-2, providing credence for a link between changes in provenance and magmatism in the region.

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Exploring for hydrocarbons in a volcanic province – A review of exploration on the Faroese Continental Shelf

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Abstract

From the initial focus on deep water Paleocene stratigraphic/combined play types chased in the Judd sub-basin, exploration effort in the Faroese area has changed its attention to the sub-volcanic structural play types. Post-well analyses of the first Judd sub-basin wells concluded that some of the amplitude driven plays drilled were caused by poor lithology calibration, but an example of an amplitude anomaly in the Vaila Fm, with well control, from the Judd sub-basin caused by gas, is presented.

This review will give a short summary of the technical advancements in seismic imaging and drilling performance achieved over the last 5-10 year period and though only a few wells has targeted sub-volcanic plays and their results largely remains confidential, some outcomes have been released and some geological issues, which these released results have raised, are qualitatively discussed as well as some speculations on their impact on prospectivity seen in the light of both volcanic and non-volcanic analogues, from other parts of the world.

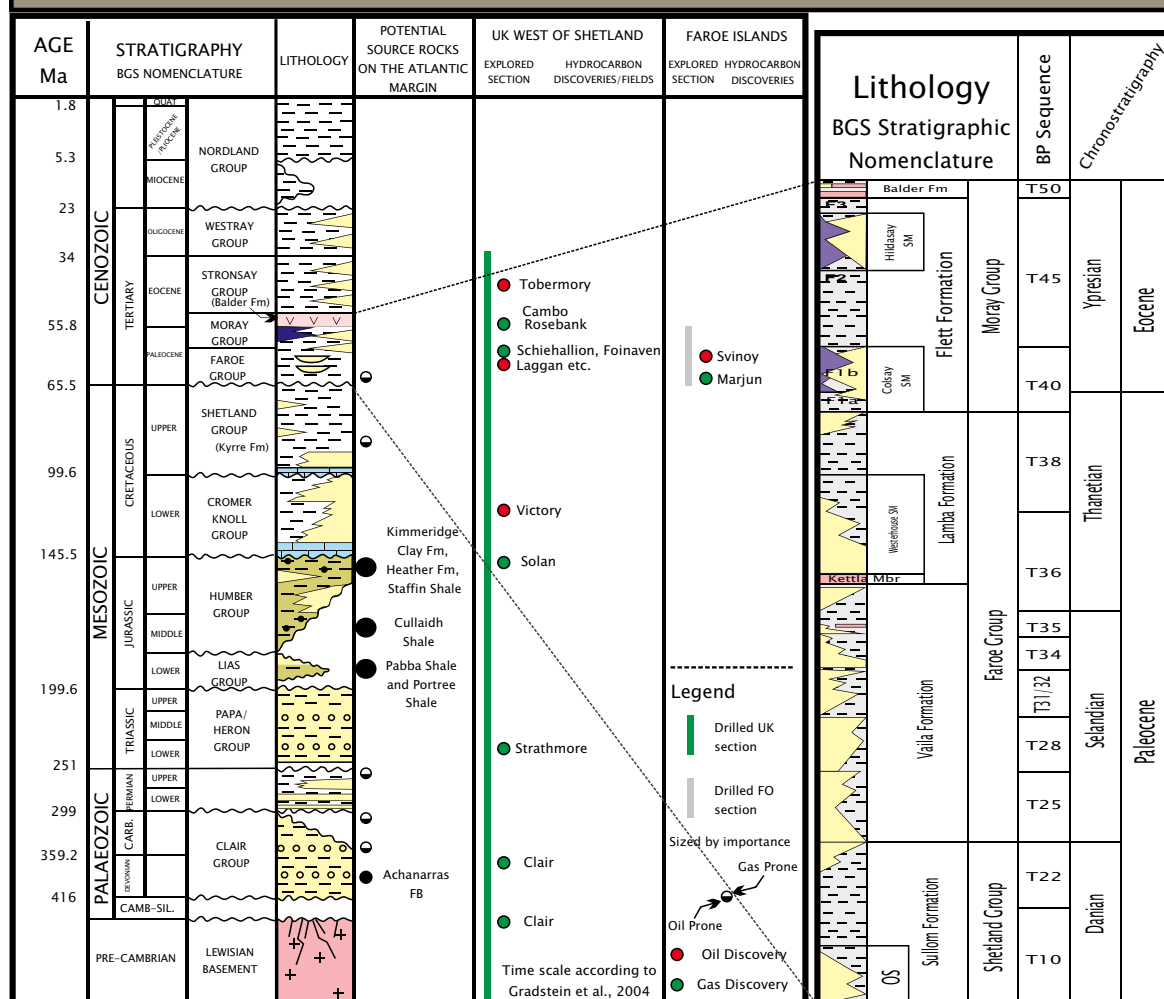
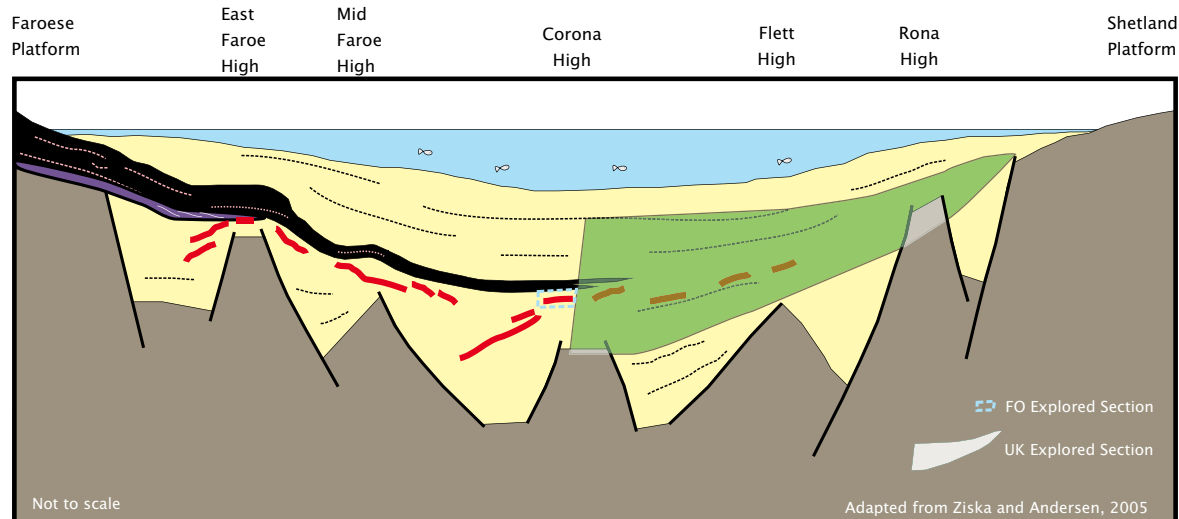
Introduction

Exploration in the UK Faroe-Shetland Basin commenced in the mid-70's with the first discovery of the Clair field on the Rona Ridge in 1977. Since then more than 400 wells (includes exploration, appraisal and development wells, 2010 numbers from DECC, the number also includes the North Rockall area) have been drilled with discoveries encountered in most stratigraphic levels from Pre-Cambrian Lewisian basement in the Clair Field to the Eocene reservoirs of Tobermory (Fig. 1) and with all major intrabasinal highs targeted. This contrasts the Faroese side where only the Paleocene section has been penetrated and with only two of the major intrabasinal highs targeted (Fig. 1), though the results of the last well 6004/8a-1a remains confidential.

Since the discovery of Foinaven in the early 1990's exploration wells have been positioned on Paleocene prospects with traps formed from a combination of structural and stratigraphic components where the Paleocene sandstone reservoir either pinches out or shales out updip or the juxtaposition of oil-bearing T30 sandstones and T20 shales (see Fig. for the stratigraphic scheme used in this paper) but in the last decade a shift to a strategy of targeting structural plays further out in the basin on and around the Corona High has been successful with the discoveries of Rosebank/Lochnagar and Cambo.

In Faroese waters, exploration has focused on the Faroe-Shetland Basin, of which more than 90% percent of the area within Faroese waters is covered by Early to Mid Palaeogene volcanics

Conceptual Cross Section – Faroe-Shetland Basin



Adapted from Scotchman and Doré, 1995; Scotchman et al., 1998; Scotchman and Carr, 2005 and Ziska and Andersen, 2005

Fig.1. Conceptual cross section across the Faroe Shetland Channel illustrating the difference in drilled sections on the UK side and the Faroese side. Full stratigraphic column for the explored section on the UK side, including potential source rocks on the North Atlantic Margin (Scotchman and Doré, 1995; Scotchman et al., 1999 and Scotchman and Carr, 2005) and discoveries. Stratigraphic column for the Paleocene-Earliest Eocene in the West Shetland area. The BGS nomenclature is from Knox et al. (1995) and Lamers and Carmichael (1999). The T-sequence is based on Ebdon et al. (1995).

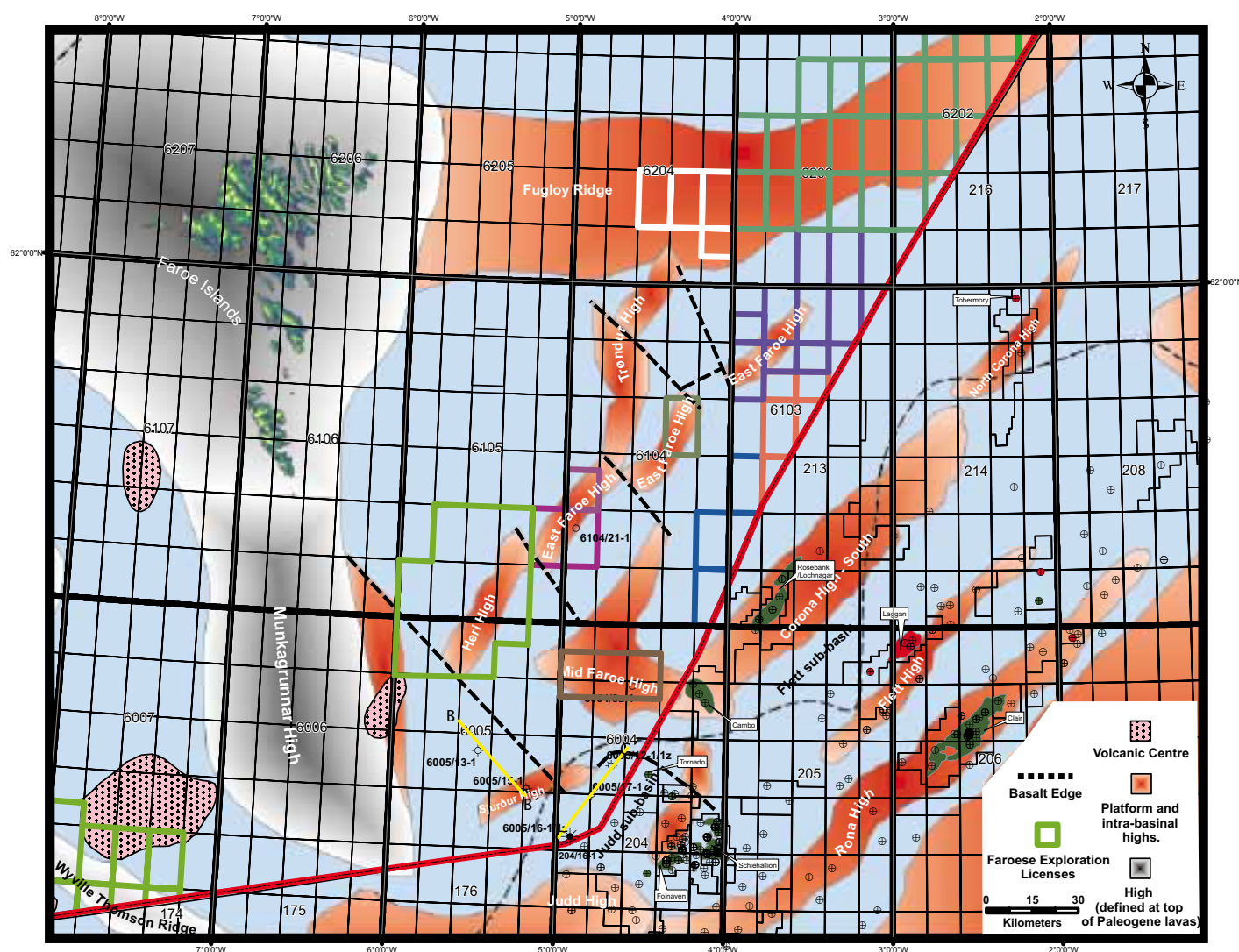


Fig. 2. Location map with structural elements, exploration licenses, oil fields and quadrants (numbered). The yellow lines indicate location of seismic profiles in Figs 3, 4 and 8.

and exploration is therefore likely to be influenced by the volcanic section (Fig. 2). Three of the seven wells drilled in the Faroese area have been targeting Paleocene prospects with a strong stratigraphic component and Fig. 3 shows a semi regional seismic line running close to three wells, 6004/16-1z, 6004/17-1 and 6004/12-1z exemplifying the stratigraphic nature of the drilled prospects.

Although the drillings in the Judd sub-Basin resulted in two discoveries and oil shows in a third well (Varming, 2009), the failure to encounter hydrocarbon accumulations which at present seem economically feasible was mainly related to the lack of sealing lithologies, with high net sand to gross interval thickness ratios

encountered in the Vaila Formation in the Faroese wells (Smallwood, 2005; Woodfin *et al.*, 2005; Varming, 2009). This contrasts the results from the UK side, where post-drilling analyses has shown that poor trap definition, reservoir presence and quality is a common failure component of the stratigraphic Paleocene plays (Loizou *et al.*, 2006; Lamers and Carmichael, 1999; Smallwood, 2005). Only in one of the Faroese wells were stratigraphic levels older than the Vaila Formation penetrated, the Marjun discovery was encountered in the Early Paleocene Sullom Formation, which in this location shows low permeabilities.

Given that true stratigraphic traps have little or no structural control, location of the drilled Paleo-

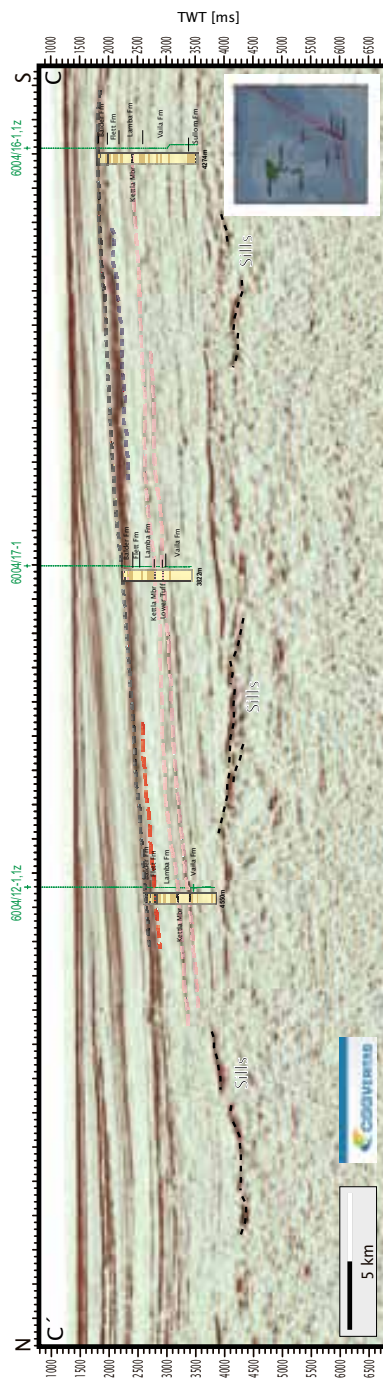


Fig. 3. Seismic section striking along the Faroe-Shetland Basin length axis running close to the three wells 6004/16-1z, 6004/17-1 and 6004/12-1z (from south to north). The furthest away from the seismic profile is located ~2.5 km away (6004/16-1z). The simplified lithological columns for the well have been constructed from a combination of well site lithological descriptions and wireline logs. Yellow indicates sandstone dominated lithologies, brown indicates mudstone dominated lithologies, red indicates igneous material and pink indicates tuffaceous material. Aero-offset check shot data has been used convert well depth to Two-Way-Time. Seismic data reproduced with kind permission from CGGVeritas.

cene UK targets has relied extensively on additional geophysical techniques, such as direct hydrocarbon indicators (DHI's), i.e. flat spots, amplitude anomalies and amplitude variations with offset (AVO) (Loizou *et al.*, 2006; Cooper *et al.*, 1999; Lamers and Carmichael, 1999; Leach *et al.*, 1999).

Post-drilling analyses of UK and Faroese wells (Loizou *et al.*, 2006; Varming, 2009) targeting Paleocene prospects with a strong stratigraphic component show that lithology effects induced by the volcanic activity (e.g. igneous intrusions and tuffaceous deposits) led to erroneous interpretations of the amplitude and AVO anomalies supporting some of the chased plays. However, an example of an amplitude anomaly which seems to relate to changes in fluid composition (gas) is seen in the deepening of the well 6004/12-1/z, where the Paleocene (T31) Fugloy prospect was targeted. The interval contained a thin (~5 m) gas accumulation (Fig. 4c) and petrographic analysis (from ditch cuttings and SWC) of the section confirms that the interval is devoid of volcanic detritus, while the upper (T35.3 and younger) section in the well displays an influx of volcanic material (e.g. shards of volcanic glass and basaltic fragments). The discovery was mapped on 3D seismic (an in-line example is shown in Fig. 4a) and amplitude extract of the interval T34 – T28 was subsequently produced on the far-stack data. The resulting amplitude map is shown in Fig. 4b.

After the first, less successful, exploration efforts on the Faroese Continental Shelf (FoCS), focus has primarily changed from the solely stratigraphic or combined stratigraphic/structural trapped deep water Paleocene play types in the non-basalt covered Judd sub-Basin to structural play types along the East Faroe High and over the Mid Faroe High of both post-, intra- and sub-volcanic plays (Fig. 1 & 2).

Moving the exploration effort into the volcanic covered areas, another set of exploration challenges emerges, and some of these will be discussed in this paper. Some examples of exploration challenges in a volcanic province are mapping and distribution of volcanoclastic intra-basalt sedimentation, sill intrusions and their

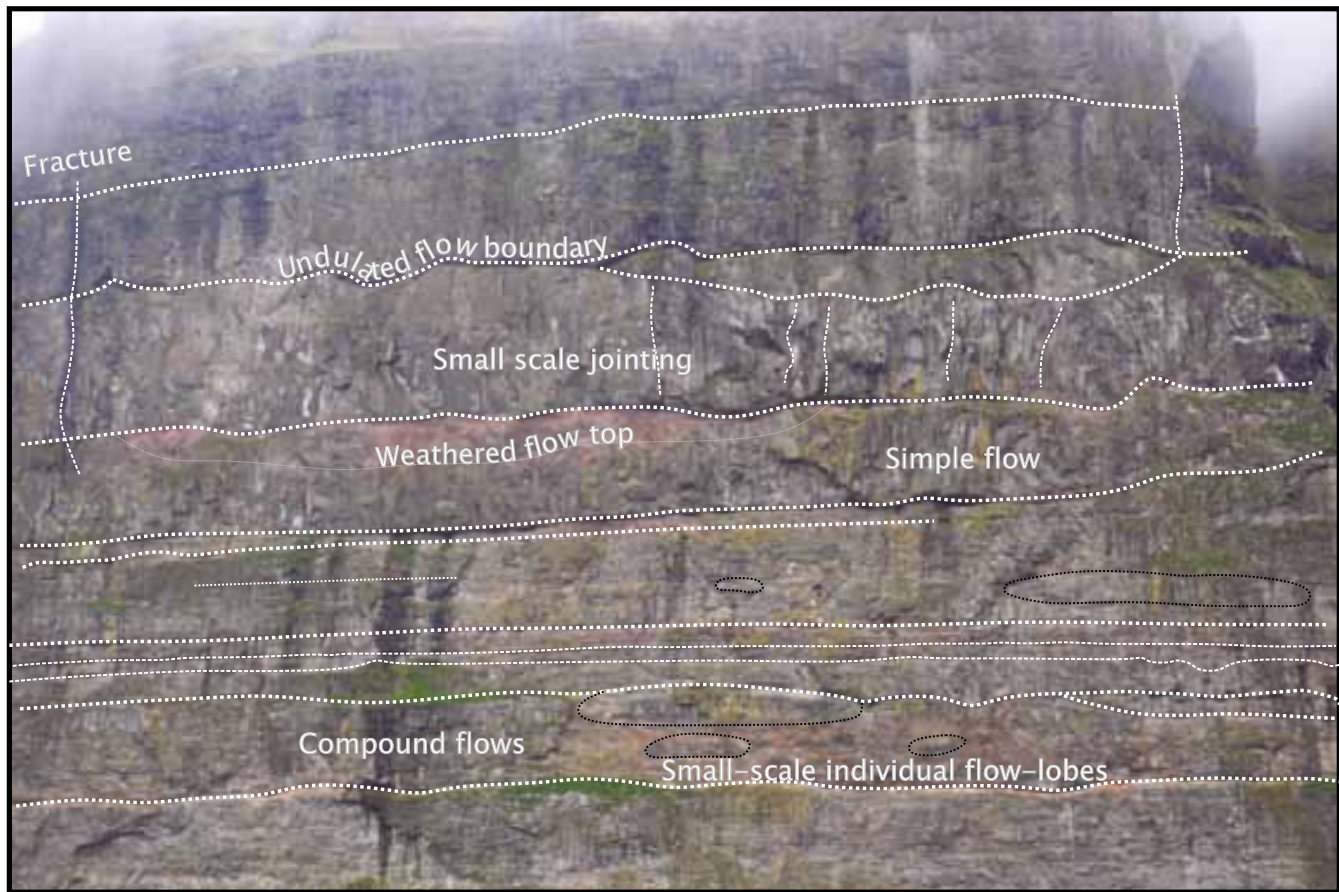


Fig. 5. Photo of cliff section displaying some of the features likely to contribute to the poor seismic data quality. Note the variability in flow thickness, lateral changes in boundary characteristics and the undulating geometry of some of the boundaries as well as the geometry of the individual flow lobes. All these heterogeneities can cause scattering and distortion of the seismic wavefield and induce random noise. The height of the visible section is not known.

the acquisition and processing phase) has helped de-risking the sub-volcanic plays by better definition of intra and sub-volcanic features and enhancing coherent reflectors in the sub-volcanic section. Improvements in drilling performance has lowered the risk of technical failure and drilling costs. The primary challenges remaining is understanding which stratigraphy these sub-volcanic reflectors represent, the age of sub-volcanic structures and what influence the volcanic section has on source rock maturity, reservoir and migration timing.

The volcanic sequence offshore has traditionally been referred to as a basalt sequence, but in this publication such a specific term will not be applied. This is because the sequence is known from onshore exposures, onshore drill cores and drilling results to include a variety of volcanic

facies, including sub-aerially emplaced basalt lava flows, hyaloclastite deltas, volcaniclastic sandstones and inter-lava units such as coals, paleosols and other sedimentary units (e.g. Rasmussen and Noe-Nygaard, 1969; Passey and Jolley, 2009). This does not include the numerous sills typically imaged below the volcanic sequence, and found onshore Faroe Islands within the volcanic sequence.

Seismic Imaging

With the change in exploration focus towards the volcanic covered structural highs of the FoCS, the seismic imaging problem and methods to enhance geophysical data quality to de-risk sub-volcanic plays has been a major concern and though the problem still persists, there has been some major

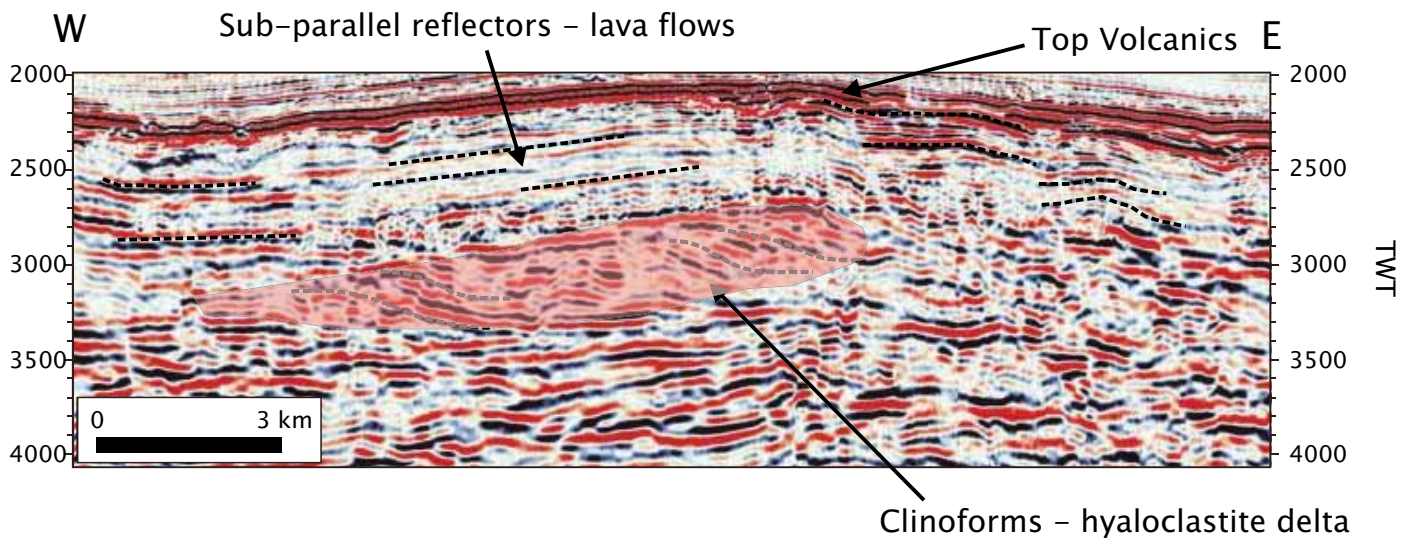


Fig. 6. The figure shows an example of a seismic line with what is interpreted as a hyaloclastic delta with clinoform structures (pink overlay), overlain by parallel reflectors interpreted as parallel bedded lava flows.

advancements in both seismic data acquisition (e.g. over/under acquisition; Leathard *et al.*, 2009, deep towing, bubble-tuned, long offsets; White *et al.*, 2005) and processing techniques (e.g. Gallagher and Dromgoole, 2008; Hobbs *et al.* 2009; Bean and Martini, 2010) over the last 5-10 years. Today confident mapping of large structures below the volcanic section is possible due to a marked improvement in the imaging of coherent reflections below the volcanic section, something that the first sub-volcanic well verified, by penetrating the shaly Lamba Fm (Ellis, 2009; Allinson, 2009).

Previously it has been the general idea that the main mechanism for poor sub-volcanic imaging was related to a high impedance contrast between basaltic lavas juxtaposing lower density and/or lower velocity rocks (e.g. sandstone or shales), but recent investigations in seismic wave field behaviour have shown a more complex set of mechanisms, emphasizing the influence of the internal heterogenic nature of a volcanic pile e.g. inter-layered volcanoclastic sediments, eroded intra-layered surfaces, vesicular flow margins, jointing, fracturing and irregular faces; Fig. 5 shows an example of how some of these features looks like in a cliff section on Svínøy, which notably adds to the picture of distorting and scattering the seismic wave field (Pujol and Smithson, 1991;

Gibson and Levander, 1998; Hestholm and Ruud, 2000; Martini *et al.*, 2001; Boldreel, 2003; Bean and Martini, 2010). This scattering can be both coherent, where wave field reverberate within individual layers, and incoherent, producing random noise through which deeper sub-volcanic reflections are masked (Martini and Bean, 2002).

From the many published works on seismic imaging within the Faroese area, a few conclusions can be drawn with respect to seismic data acquisition and processing (White *et al.*, 2005; Gallagher and Dromgoole, 2008; Hobbs *et al.*, 2009; Leathard *et al.*, 2009). One of the main conclusions is the significance of deep towing (≥ 15 m) source and streamer in the acquisition phase, to band limit the data, as well as focusing on putting more of the available energy into the low end of the frequency spectrum (Hobbs, 2002; Martini and Bean, 2002; Ziolkowski *et al.*, 2003; White *et al.*, 2005). The very long offsets of 12 km and above tested in the early 90's have not necessarily been beneficial and apertures of 6-8 km are now regarded as being appropriate (e.g. Gallagher and Dromgoole, 2008; Hobbs *et al.*, 2009). The streamer length shall be long enough to obtain information to allow for confident demultiple and accurate velocity analysis, which will depend on the depth to the top of the volcanics, their thickness and the depth to target.

In the seismic acquisition phase, the industry has come up with a fairly robust setup while the biggest uplift has been in the seismic processing stage, where particular emphasis on removing the high frequencies early in the processing stage has proven beneficial, careful noise attenuation, demultiple (e.g. SRME and Radon Transform demultiple), iterative geological interpretation derived velocity analysis of the volcanic section and pre-stack beam steering migration has given a remarkable improvement in the seismic imaging of the deeper sub-volcanic levels (e.g. Gallagher and Dromgoole, 2008; Hobbs *et al.*, 2009). Inspection of older speculative seismic surveys has demonstrated that some of these data sets also contain the necessary information in the lower frequencies and reprocessing of these surveys using the above processing work-flow has uplifted these data set's ability to image the sub-volcanic section (e.g. Gallagher and Dromgoole, 2008; Hobbs *et al.*, 2009).

Drilling of volcanics

Another issue that has been conceived a hindrance for exploration within the volcanic covered areas of the FoCS is the hard rock drillings, which previously has been perceived too difficult, too time consuming and ultimately too expensive. An example is the well 164/7-1 in the UK sector of the Rockall Trough, which drilled 1.2 km of Paleocene basaltic lavas (Archer *et al.*, 2005). From the End of Well report, penetration rates of around 3-4 m/hr in the upper weathered basaltic section is documented, while the deeper Late Cretaceous dolerite sill intruded mudstone section, displays even lower penetration rates (~1.2 m/hr) and average drill bit runs of ~125 m through the Paleocene basaltic section. An earlier example of basalt drilling is the deepening of the Lopra-1/1a onshore borehole in 1996. In this borehole; 1400 m of volcanic strata was drilled using a conventional tri-cone bit. In all 24 6½ inch bits were used, each drill bit drilling 58 m in average (range 4 - 170 m) and with an average rate of penetration of 2.18 m/hr (range 0.74 – 3,5 m/hr) per bit.

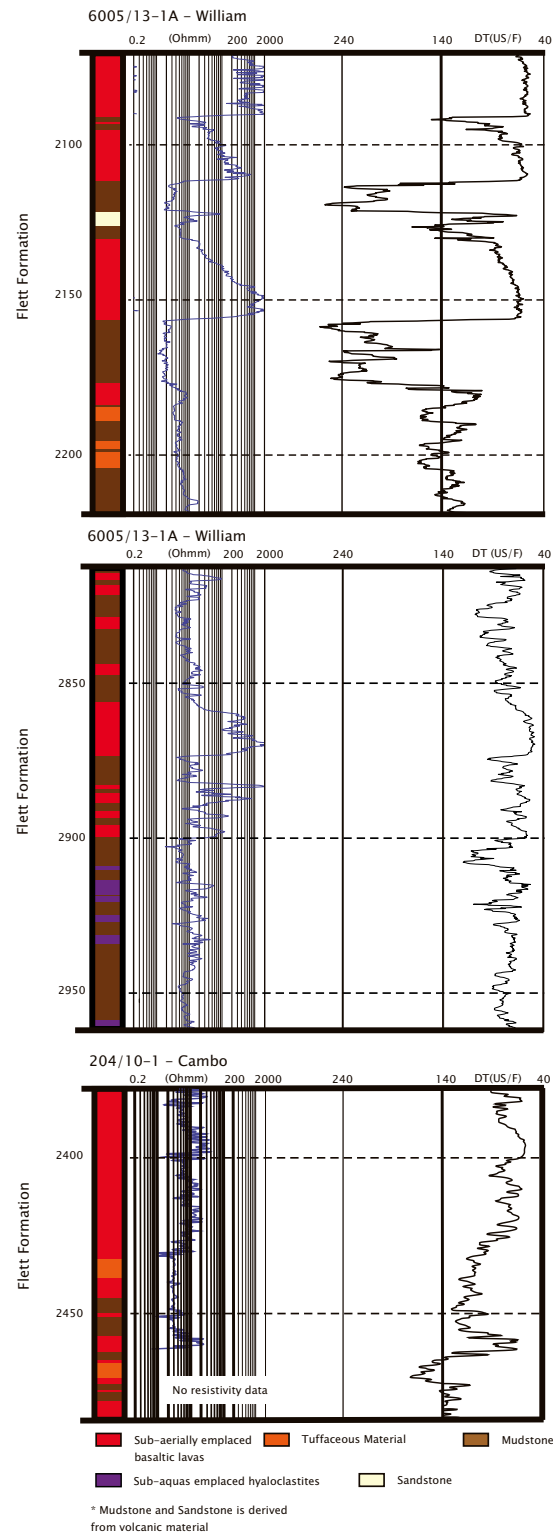


Fig. 7. Log traces from 204/10-1 and 6005/13-1A for the volcanic section in the Flett Formation (only upper part of 6005/13-1A shown) to show differences in the resistivity and sonic log trace appearances. Note the difference in the log appearance of the upper part of well 6005/13-1A, with very sharp boundaries between the lavas and the volcanoclastic mudstone, giving the log trace a box-like shape, while the lower section of 6005/13-1A and 204/10-1 displays more serrated log traces.

It was therefore, from these rather discouraging results, necessary to increase drilling performance prior to drilling the first well targeting a sub-volcanic play on the FoCS. Fortunately testing and development of new drill techniques and bits on the Rosebank discovery (drilled in 2004) led to substantial advancements in drilling performance (Close *et al.*, 2005).

The results from the drilling showed a direct proportionality between rotational speeds and rates of penetration. To achieve high rotations per minute, a combination of diamond impregnated bits, a turbine system and powerful rig pumps were utilised. Comparison between offset well costs and a post-well cost analysis revealed that these improvements led to a lowering of the cost-per-drilled-foot by a factor 2.5 (Close *et al.*, 2005).

The technology was adopted by Statoil when drilling well 6104/21-1. Penetration rates around 10 m/hr have been reported while one single diamond impregnated bit drilled over 750 m of volcanics in the well (Allinson, 2009). The same technique was also successfully used when well 6005/13-1 was drilled, here a single drill bit also drilled more than 700 m of volcanics, though the penetration rates were lower (~ 4 m/hr; ROP values obtained from the End of Well Report). The conclusions from post-well analyses suggests that drilling of hard volcanics doesn't pose a problem, while a section of mixed thinly bedded hard and soft sections e.g. a sill intruded shaly section or a volcanic section with a thick, deeply weathered soft upper crust and a hard brittle flow core often leads to slower rates of penetration and vibrations in the drill stem.

But all together, the recent advancements in drilling performance and lower cost per drilled meters of volcanics; should increase the impetus for sub-volcanic exploration on the Faroese Continental Shelf.

Reservoir

Pre-volcanism reservoirs

The Vaila interval encountered in the Faroese part of the Judd sub-basin proved to be sand prone (Smallwood, 2005; Woodfin *et al.*, 2005; Varming, 2009), while the necessary sealing lithologies, were for the most part deficient, with the only encountered competent seal (Pre-Vaila) being the mudstones of the Early Paleocene Sullom Formation encountered in the well 6004/16-1z and the shaly Lamba Formation (post well analyses, Woodfin *et al.*, 2005; Smallwood, 2005b). The Sullom Formation reservoir section (average porosity of 11 % and permeability < 1 mD) consisted of two reservoir sections separated by a 10 m thick shale section and a 8 m thick dolerite section just above related to an apophysis from a sill complex south west of the well. Petrographic studies of the reservoir section have shown an absence of garnet grains and a presence of high temperature authigenic phases including rhodochrosite, lamellar twinned calcite and sericite (a fine grained form of mica), all indicative of elevated temperatures. Though the reservoir section displayed low bulk permeability values (higher horizontal permeability values), its porosities appear to be largely unaffected (Smallwood and Harding, 2009; see later section in this paper), thus suggesting that the Sullom Fm could be a viable reservoir in other non-intrusive affected areas.

The Paleocene interval in the UK part of the Faroe-Shetland Basin contains excellent quality sandstone reservoirs within basin-floor, slope fan and shelfal facies (e.g. Lamers and Carmichael, 1999; Naylor *et al.*, 1999). Porosities of more than 25 % and permeabilities greater than 100 mD is often observed where the sandstones are buried less than 2500 m below seabed in the Judd sub-basin (e.g. Ebdon *et al.* 1995; Lamers and Carmichael, 1999; Cawley *et al.*, 2005) while in the Flett sub-basin the porosities and permeabilities are generally lower due to their deeper burial (Lamers and Carmichael, 1999), though e.g. the wells 205/9-1, 206/1-2 and 214/27-2 close to the Flett High display porosities of 25 % and

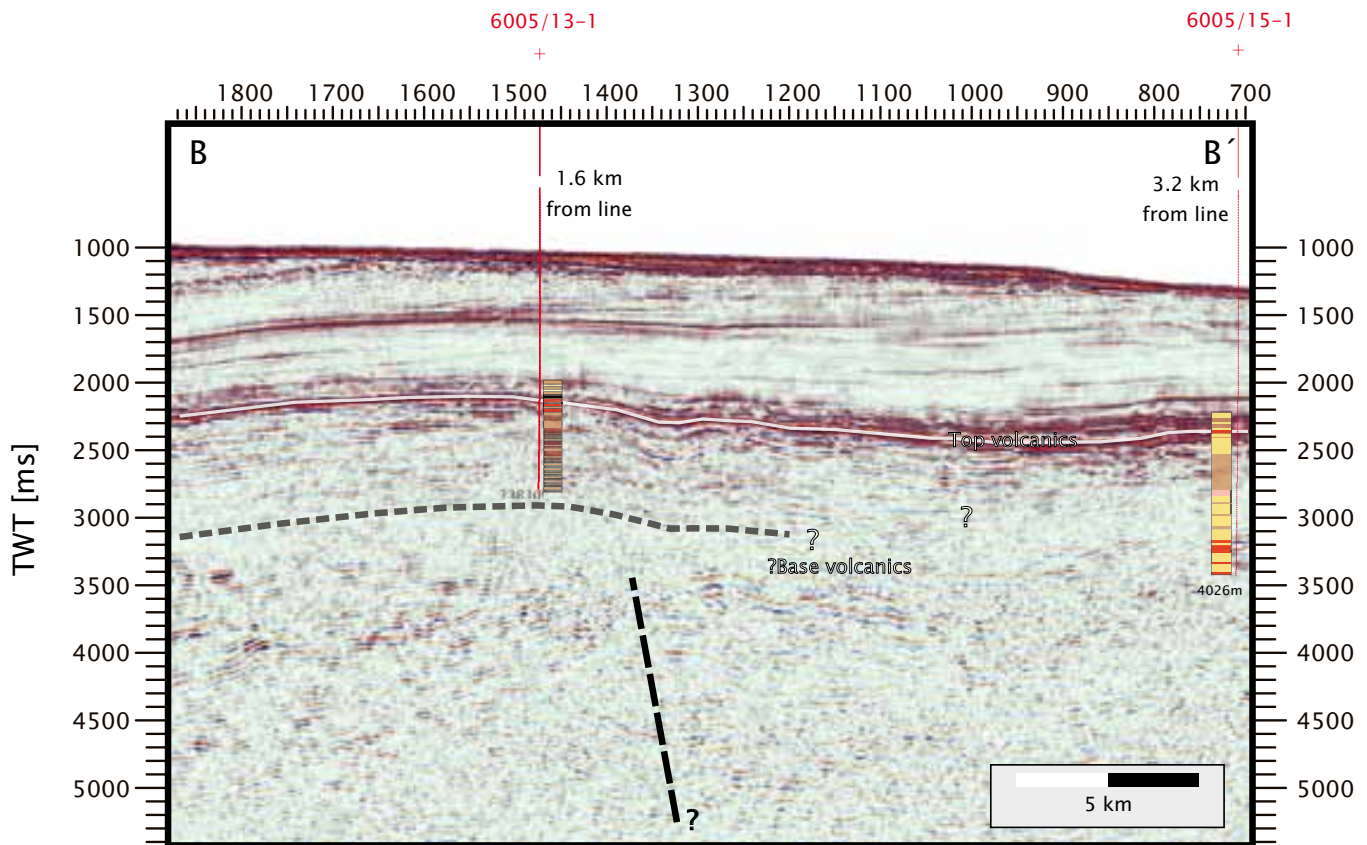


Fig. 8. 2D seismic line connecting the two wells 6005/15-1 and 6005/13-1. The well 6005/15-1 encountered 30 m of basaltic lava flows and 55 m of volcanoclastic sandstone/siltstone, while the well 6005/13-1 TD'd within the volcanic section encountering a thickness of 1475 m of basaltic lavas and hyaloclastites. Even with well ties, it is difficult on the seismic section, to reconcile how the large differences in thickness of the encountered volcanic section can arise, but also how these thickness differences manifest itself so poorly in the seismic data.

permeabilities of several hundreds of mD in T31-T34 reservoirs at depths greater than 3000 m below seabed. The precipitation of early chlorite cementation on sandstone grains is considered the porosity preserving mechanism in these wells (Sullivan *et al.*, 1999) though the presence of increased aquifer pressure (350-600 psi overpressure) in the T31-T35 sequence over most of the Flett sub-basin (Loizou *et al.*, 2006) could also have an influence on the favourable reservoir characteristics.

The high net to gross interval thickness ratio (60-70% for the T36-T28 sequences in wells 6004/12-1z and 6004/16-1z) of the Vaila Formation encountered in the Faroese wells (Smallwood, 2005b; Woodfin *et al.*, 2005), contrasts some of the findings on the UK side (Smallwood, 2005b), e.g. is the average net to gross for the Upper Paleocene (T31-T34, Vaila Fm) reservoir section

for the Foinaven field as a whole 55 % (Carruth, 2003) was not anticipated.

Heavy mineral and palynological data suggest that a sediment supply sourced from the west has influenced the Judd sub-basin through the Paleocene. A calculation and comparison of the volume of Paleocene sediments in the Judd sub-basin (Smallwood, 2005a; Smallwood, 2005b; Smallwood, 2008) and the volume denudated from the British Isles during the entire Cenozoic (Jones *et al.*, 2002) display a discrepancy, with the Paleocene sediments in the basin being 30-40% larger than the volume denudated from the British Isles (Smallwood, 2008). Thus it suggests that an alternative source, other than the British Isles, is needed to account for the Paleocene sedimentary volume in the basin. Possible contributions to explain the volume discrepancy could be temporal variations in drainage divides onshore

UK (meaning that the sediment catchment area changes/expands over time) or sediments from a westerly source.

Temporal differences in the onshore drainage catchment area have been demonstrated and Jones and White (2003) infer a much larger catchment area at the end of the Paleocene than at present day. Several lines of evidence from heavy mineral analysis and palynology has indicated (e.g. Jolley *et al.*, 2005; Jolley and Morton, 2007; Morton *et al.*, this issue) an input of westerly (non-British Isles) sourced material is likely to have taken place prior to the onset of volcanism, but the provenance area from which it has been derived from remains largely uncertain, with proposed source areas being the Kangerlussuaq area, East Greenland (Larsen *et al.*, 1999; Larsen and Whitham, 2005), sub-volcanic basement of Blosseville Kyst or the Faroe Islands and intrabasinal highs within the Faroe-Shetland Basin e.g. Corona High, Judd High and Wyville-Thomson Ridge (e.g. Ziska and Andersen, 2005; Smallwood, 2008; Ziska and Varming, 2008; Morton *et al.*, this issue). Heavy mineral observations from the Faroese wells suggest the influx of non-British Isles material is most prominent in the early Paleocene and wanes up through the Paleocene (Varming, 2009). Whether this is due to an overall increase of easterly sourced material during the Paleocene, a shut-off of material from the west or regional changes in bottom currents and distribution patterns remain speculative.

Reservoir sections older than Paleocene have not been penetrated within the Faroese area; but the Lochnagar discovery has shown that sub-volcanic prospective reservoirs of Jurassic age exist close to the Faroes-UK border. In areas of a thick volcanic cover sections older than the Paleocene is expected to be challenged due to their likely deep burial within the basinal parts of the Faroese area, but in areas of a thinner volcanic cover and on the intra-basinal highs pre-Paleocene sections might still be viable targets, though it is unknown if older sections than Paleocene exists in these settings.

Syn- and post-volcanism reservoirs

The intra-basalt discovery of Rosebank (213/27-1) and the post-basalt discovery of Cambo (204/10-1) close to the Faroese border, has led to an emerging focus on the reservoir properties of the younger part of the Paleocene section especially the influence of volcanic derived sediments and their impact on reservoir properties. The intra-basalt Rosebank discovery is found in the Early Eocene Colsay sandstones within the Flett Formation and contains at least three reservoir sections with autonomous oil and gas accumulations (Helland-Hansen, 2009). The deepest reservoir in the volcanic section is siliciclastic, indicating that sedimentation of siliciclastic material co-existed during the early stages of volcanism. In the higher reservoir section, the reservoir consists of an interbedded combination of volcanoclastic and siliciclastic sediments, with little or no interaction between the two lithologies, indicating two interfingering sedimentary systems competing for the same accommodation space (Helland-Hansen, 2009).

The End of Well Report descriptions of the ditch cuttings and side wall cores for the reservoir section of the post-basalt Cambo discovery in the Hildasay Sandstone Member of the Flett Formation, seems to indicate a relative clean siliciclastic sandstone in the reservoir zone, while the main influence of volcanic derived material (tuffaceous material and glass shards) seems to be concentrated above and below the main reservoir section.

With the findings of the Rosebank and Cambo discoveries and the occurrence of laterally extensive mafic volcanoclastic fluvial sandstones onshore the Faroe Islands (Ólavsdóttir and Ziska, 2009; Passey, 2009; Passey and Jolley, 2009), a focus on the reservoir properties of these volcanoclastic sandstones has emerged (e.g. Ólavsdóttir and Stoker, 2006; Ellingsgaard *et al.*, 2009; Ólavsdóttir, 2009; Ólavsdóttir and Ziska, 2009; Passey, 2009 and Passey and Jolley, 2009). Preliminary results from cored samples for some of the volcanoclastic sandstones indicate average porosities of 18-20 % and generally

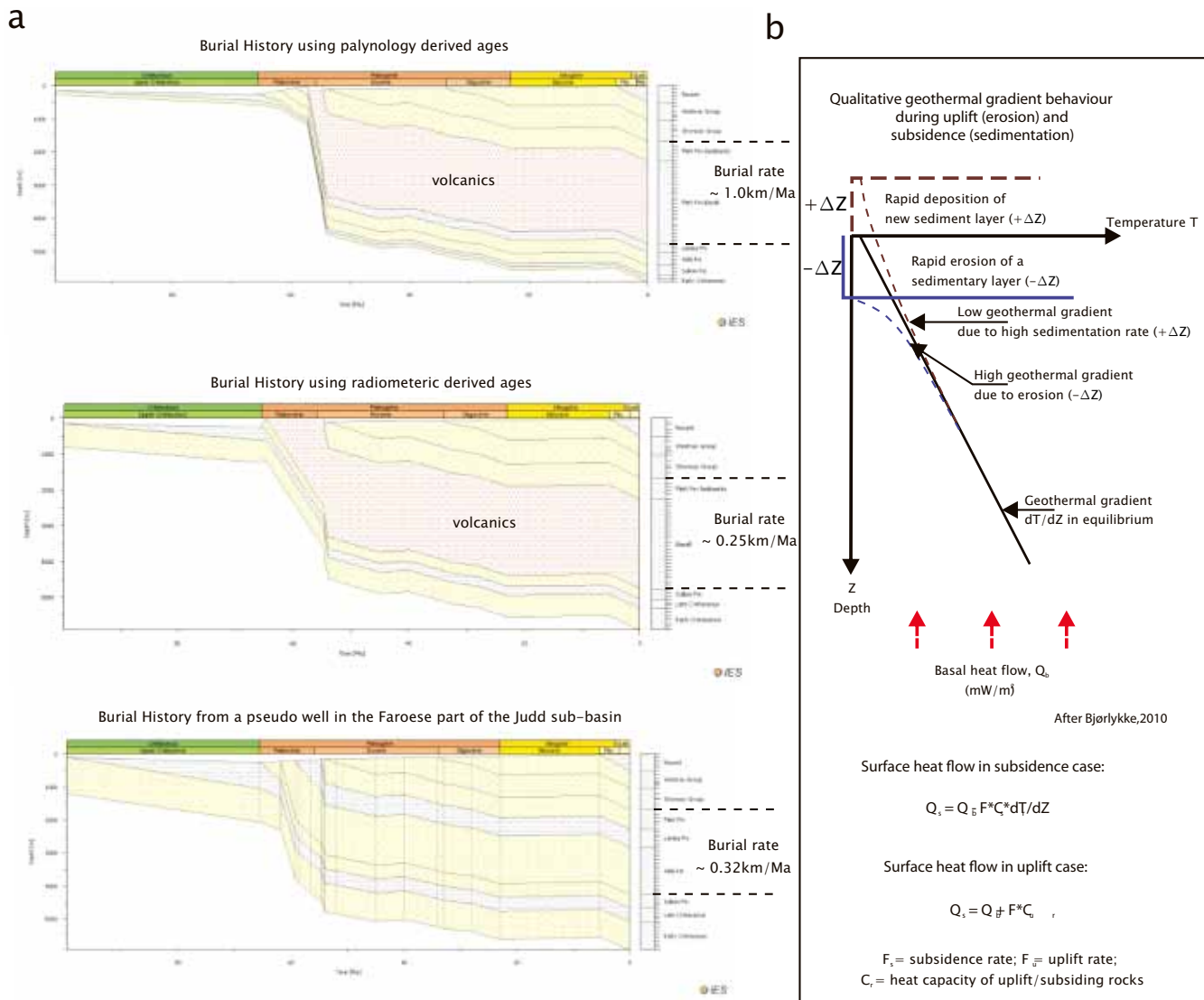


Fig 9. a) Three burial history scenarios derived from different age dating methods for the volcanic interval using a 1D basin model. In the first two examples, the volcanic section is 2.5 km thick, while the third example is from a well drilled in the Faroese part of the Judd sub-basin. The burial rate is calculated from the stratigraphic interval - Top Flett Fm to Base Vaila Fm.

b) Geothermal gradients as a function of rapid uplift (erosion) and subsidence (sedimentation). During subsidence some of the heat flow is used to heat the subsiding sediments and underlying basement, reducing the geothermal gradient, forming a cold basin. During uplift, the heat from cooling rocks will add to the heat flux, producing a steeper geothermal gradient Adapted from Bjørlykke, 2010.

low, but very variable permeabilities of 0.12-13 mD (Ellingsgaard *et al.*, 2009 and Ólavsdóttir, 2009). The variability in the permeability seems to be related to lithology (Ellingsgaard *et al.*, 2009; Ólavsdóttir, 2009). Onshore mapping of some of these volcanoclastic lithologies suggest a change of facies, from conglomerate to sandstone in a North-South direction (Ellis *et al.*, 2009; Passey, 2009), opening up for the possibil-

ity for sorting away unstable minerals and lithic components (e.g. volcanic glass) and mixing or interfingering with siliciclastic components at the distal edge of the lava field (Ólavsdóttir and Ziska, 2009). This mixing or interfingering of different sediment types are also observed in the post-volcanic section in the Faroe-Shetland Basin deposited during Eocene time (Ólavsdóttir and Ziska, 2009; Ólavsdóttir *et al.*, 2010), where map-

ping of sedimentary dispersal systems indicate a sedimentary system originating from the west going eastwards and thought to have a volcanic hinterland while a sedimentary system prograding in a north-westward direction from the NW British shelfal area thought to be of siliciclastic in composition (Ólavsdóttir and Ziska, 2009).

Indications for a regionally working sedimentary distributary system, during the time interval encompassing the period of volcanic activity, are exemplified in the onshore volcanoclastic section of the Sund Beds, where a diverse assemblage of Late Cretaceous reworked dinoflagellate cysts has been observed (Passey and Jolley, 2009), indicating that an influx of older material into the sedimentary system took place during the eruptive hiatuses.

Source-rock and maturity

From the first wells in the non-volcanic covered Judd sub-Basin it has been established that a working hydrocarbon system exists within the Faroese area. Organic geochemistry and biomarker analysis indicates that the hydrocarbon system and charge is similar to what is seen in Foinaven and Schiehallion fields (Cawley *et al.*, 2005). This means a dual source-rock system of Mid-to-Late Jurassic age (Heather Formation and Kimmeridge Clay Formation) and at least two episodes of charging (Scotchman *et al.*, 1998). Moving into the basalt-covered areas, only a few non-conclusive indications from seep-surveys have suggested the presence of an efficient hydrocarbon system, so when drilling the first well targeting a sub-volcanic prospect, the existence of an efficiently working hydrocarbon system was one of the main risks. Furthermore were the maturity and hydrocarbon phase little known and the effect of volcanic overprinting on the hydrocarbon system was uncertain. Previous understanding suggests the most likely phase would be dry gas, but thermal modelling incorporating pressure retardation (Carr and Scotchman, 2003; Scotchman and Carr, 2005) has indicated that the Kimmeridge Clay Formation is still

mature for oil generation at 8 km depth in overpressured basins and that onset of expulsion from the source rock is delayed in a overpressured system compared to a normal pressured system.

Post-well analysis of hydrocarbon indicators (from FIT analysis, Fluid Inclusions Technique) within well 6104/21-1 demonstrated a working hydrocarbon system (Ellis, 2009) within the basalt covered area, reducing one of the risks existing prior to drilling.

Aspects of thermal effects from volcanic activity

The thermal effects of volcanic activity can be seen on several scales, from a rise in regional basal heat flow during break-up, thermal effects on reservoir sections from igneous intrusions, local heating of source rocks, and modification of previously generated hydrocarbons.

Rifting and break-up is commonly associated with a regional rise in the basal heat flow from the rising of the mantle due to thinning of the lithosphere (Kusznir and Park, 1987), the heat flow rise is generally dependant on stretching factor, strain rate, time interval of the rifting episode, initial mantle temperature and thermal properties of basin fill.

The relationship between volcanic activity, rifting and ultimately break-up has been much debated, but it is generally accepted that the North Atlantic Igneous Province (NAIP) is a consequence of two major magmatic phases: a pre-break-up phase (~62-58 Ma) and a syn-break-up phase (~56-54 Ma) contemporaneous with the onset of the North Atlantic sea floor spreading (Saunders *et al.*, 1997; Larsen *et al.*, 1999; Torsvik *et al.*, 2001; Jolley and Bell, 2002; Waagstein *et al.*, 2002; Meyer *et al.*, 2007; Storey *et al.*, 2007). The first pre-break-up phase of volcanism was constrained to a NW-SE band extending from the UK, across Faroe Islands and Greenland (East and West) and on to Baffin Island in East Canada (e.g. Saunders *et al.*, 1997; Doré *et al.*, 1999; Lundin and Doré, 2005; Ziska and Varming, 2008; Ziska, this volume), while the second phase of

volcanism is directly associated with the breakup of the North Atlantic in the Early Eocene (Saunders *et al.*, 1997; Doré *et al.*, 1999; Jolley and Bell, 2002).

The relationships between igneous rocks and hydrocarbons have been well documented from several places around the world, e.g. in the Liaohe Basin in China (Zhenyan *et al.*, 1999), the intra-cratonic Paraná Basin of Brazil (Araújo *et al.*, 2000), the Gunnedah Basin onshore East Australia (Othman *et al.*, 2001) and the Neuquén Basin in Argentina, where a close relation between source rock maturity and distance to igneous intrusions (sills/laccoliths) is observed (e.g. Delpino and Bermúdez, 2010; Rodríguez Monreal *et al.*, 2009; Rodríguez *et al.*, 2007), though the effects of intrusive rocks on source rock maturity can be difficult to distinguish from those related to burial generation in areas where both processes coexist (e.g. Bishop and Abbott, 1995).

Examples from the North Atlantic region has shown that igneous intrusion can act as a heat source, which can induce local thermal maturity for hydrocarbon generation, is e.g. seen in the stratigraphic borehole Umiivik-1 on Svartenhuk Halvø, West Greenland, where vitrinite reflectance values (R_o) and T_{max} values display a clear bell shape around the thickest intrusions (Dam *et al.*, 1998), in the Midland Valley of Scotland (Raymond and Murchison, 1992) where Carboniferous coals demonstrate extensive modifications to molecular distributions caused by rapid heating from igneous activity, and on the Isle of Skye (Hudson and Andrews, 1987; Bishop and Abbott, 1995; Rohrman, 2007) where a comparison between different maturity proxy data sets (e.g. vitrinite reflectance and molecular geochemical ratios) displays different responses to the heating effect from igneous dykes.

In addition to the effects igneous intrusions can have on source rock maturity and hydrocarbons generated, they have also been observed to severely alter reservoir properties, e.g. in 6004/16-1z were the elevated temperatures and convection of hot fluids driven by the heat input from a sill

complex into the reservoir to the south-west has degraded the reservoir permeability, while mainly maintaining its porosity. The porosities plot on the same normal compaction trend curve for other, non-intrusive affected, sands in the well (e.g. Smallwood and Harding, 2009). Studies in the Huab Basin, NW Namibia have also shown how sills and dykes can compartmentalize the reservoir section by well-cemented induration zones of 1-2 m's width away from the intrusion (Jerram, 2006).

Production of hydrocarbons from reservoirs deposited in volcanic settings or affected by volcanism is seen several places around the world e.g. offshore Namibia in the Orange sub-basin, where the Kudu gas field is producing from a deeply buried (4400 m bsl) fluvio-aeolian sandstone syn-rift reservoir of Barremian age. The reservoir section comprises sedimentary layers, varying from 12 m to 38 m in thickness, interleaved between basalt flows. The reservoir interval consists of quartzitic and volcanoclastic sandstones with a mean porosity of 12 % (Stanistreet and Stollhofen, 1999)

In the onshore Barmer Basin, North East India, the Mangala field is producing oil from the Late Cretaceous to Early Paleocene Fatehgarh Fm, a clean quartzose sandstone in the north of the Barmer Basin with high porosity (20-34 %) and multi-Darcy permeability, but further to the south it incorporates more volcanoclastic material derived from the Deccan Traps causing reservoir quality and thickness of net sand to deteriorate (Compton, 2009).

Finally; is concentrically fractured sills seen to act as reservoirs in the Neuquén Basin in Argentina, producing hydrocarbons for up to 20 years (Bermúdez and Delpino, 2008).

Some issues in estimating the thickness of the volcanic section

Though a clear uplift in the seismic imaging quality has emerged during the last decade, tying down which seismic reflector represents the base of the volcanic section is still a difficult task

and hence the thickness of the volcanic section can still be difficult to estimate. Reasons for this are often multiple and besides the obvious lack of well control and hence the long distance well correlations have to be carried, one aspect is that the base of the volcanics are often not a sharp boundary between lithologies, but instead a gradational transition (Fig. 6), often initiating (from the bottom of the volcanic section) with a prograding hyaloclastite section grading up into parallel bedded lava flows possibly with intra-lava volcanoclastic sedimentary sections (sandstones and mudstones) and paleosols.

The two wells on the Faroese Continental Shelf, which to date has drilled substantial thicknesses (>1000m) of volcanics, have shown similar volcanic sections. Within a volcanic system, the variability in environments can be substantial and depending on location in the volcanic system (which might change through time), weathering of the lava flows, and the influence of fluvial sedimentary systems, marine incursions and lacustrine settings all influence the drilled volcanic section. It is generally accepted that the morphology of lava flows is dependant on the emplacement mechanism (Walker, 1972; Walker, 1973; Self *et al.*, 1997; Jerram, 2002; Passey and Bell, 2007), e.g. fissure eruptions versus shield volcanoes and the shift in distribution through time of these two end-member eruptive mechanisms, but also in other ways do the proximity to the eruptive centres influence the volcanic section, for instance may a distal placement within the lava field result in large periods of non-eruption, leading to more profound weathering and/or erosion of the lava flows with the development of significant paleosols.

This leads to a very heterogeneous section without a sharp lithological basal boundary between the volcanic section and the underlying non-volcanic section; complicating pinpointing the first reflector which would represent a horizon being devoid of a volcanic influence and hence complicate estimating the thickness (in the first instance as two-way-time) of the volcanic section.

Therefore mapping is normally done on packages/sections on the seismic lines and changes in the seismic character (e.g. coherent parallel bedded reflectors versus chaotic incoherent reflectors) of the individual packages rather than mapping of a specific reflector (Fig. 6).

With the possible scenario of a heterogeneous volcanic section, a large variability of the velocities might also be expected, with deeply weathered lava flows with thick associated paleosols being at the low end of the velocity spectrum, while e.g. intrusions or thick tabular non-weathered lava flows would have higher velocities. Examples of this are for instance seen in the zero-offset VSP derived interval velocities from 204/10-1, 6005/15-1 and 6005/13-1/1a, where velocities for the whole of the volcanic section in 204/10-1 is 3365 m/s while it is significantly higher in the two released Faroese wells 6005/15-1 and 6005/13-1/1a (~ 4000 - 4200 m/s). By comparing the well site lithology descriptions and the wireline log signatures for the volcanic section (gamma ray, resistivity, neutron density and sonic wireline log data) in the three wells, the results from 204/10-1 suggests a volcanic interval that has experienced more pronounced weathering than in the other two wells. In well 204/10-1 the log signatures in the volcanic section appear more erratic, lacking the typical box-like log signature of basaltic flows with a solid core and a weathered top or vesicular crust. Wireline log traces of resistivity and sonic data from the well 204/10-1 and 6005/13-1 are shown in Fig.7, supporting the concept that the more distal parts of the volcanic influenced area have had larger eruptive breaks allowing for more time to weather down the lavas.

An example of the difficulties in confidently mapping the base of the volcanics, even with well calibrations, is shown in Fig.8 where a 2D seismic line running close (perpendicular distance of 1.6-3.2 km) to the two wells 6005/13-1 and 6005/15-1, in which the encountered thickness of the volcanic section is 1475 m and 95 m (30 m of sub-aerial emplaced lavas and 65 m of volcanoclastic sandstones) respectively. Even with two

well calibration points, it is difficult to reconcile how the thickness of the volcanic section can vary so considerably over a distance of 20 km and yet manifest itself so poorly in the seismic data and though it is possible to map coherent reflectors between the two wells, the transition between the extrusive volcanics in 6005/13-1 and the siliciclastic sedimentary system in 6005/15-1 is not possible to observe in the seismic data.

Dating of the volcanic section and some implications for exploration

Traditionally lavas of basaltic composition has been dated using Ar/Ar or K/Ar radiometric dating techniques and the same methods have also been utilised on the exposed and drilled sections on the Faroes (e.g. Tarling and Gale, 1968; Fitch *et al.*, 1978; Waagstein *et al.*, 2002; Storey *et al.*, 2007). The different Ar/Ar and K/Ar radiometric datings have resulted in quite diverse results leading to a maximum end-member eruption time interval from 53.2 Ma to 64.8 Ma (an interval of 11.6 Ma from Ar/Ar radiometric data; Waagstein *et al.*, 2002) while others report of a time interval only spanning half of that e.g. 6.4 Ma (54.2–60.6 Ma) from the Ar/Ar data results of Storey *et al.* (2007) and Waagstein *et al.* (2002) reports from their K/Ar results leading to a interval of 5.8 Ma (54.5–60.3 Ma). Comparing these results with results derived from pre-syn and post-volcanic age-specific palynological data of Jolley *et al.* (2002) which estimates the total time interval for the exposed and drilled volcanic section on the Faroe Islands to be 2.6 Ma (54.6–57.2 Ma); a factor of almost 4.5 less than the time interval from the Ar/Ar data of Waagstein *et al.* (2002). The discrepancy between the two results lies in the onset of volcanism and not in the eruptive hiatuses, onshore represented by the coal-bearing Prestfjall Fm, or the start of syn-rift volcanism. The authors of this paper are in no position to dispute any of the results, or any objections to the results, but only want to point to possible consequences of this.

Besides the obvious uncertainty in the onset

of volcanism; since the vast majority of the volcanic section has been emplaced sub-aerially, it induces a significant uncertainty in the burial rate. Depending on the available time interval, the resulting burial rate could have been more than twice as high compared to other locations outside the volcanic covered areas. Figure 9a shows examples of three different burial history scenarios. In the two first examples, pseudo wells have been constructed with a thickness of a sub-aerially emplaced volcanic section of 2.5 km being used (equivalent to the thickness of volcanics encountered in 6104/21-1), in the first pseudo well, the minimum time interval (2.6 Ma; ~1 km of burial per 1 Ma) for the emplacement of the volcanic section is utilised (using the age-diagnostic palynological data of Jolley *et al.*, 2002), while the second pseudo well is using the maximum time interval (11.6 Ma; Ar/Ar radiometric data; burial rate ~0.25 km/Ma). The third example displays the burial history for well 6004/16-1/1z drilled in the non-volcanic covered part of the Judd sub-basin (burial rate ~0.32 km/Ma for the stratigraphic interval Top Flett Fm – Base Vaila Fm).

Since sub-aerially emplaced lavas rapidly cool and solidify on a geological time scale (from days, weeks and up to a few years; Hon *et al.*, 1994; Self *et al.* 1996 and estimates of individual flow cooling rates from onshore Faroes in Passey and Bell, 2007), rapid burial of a thick volcanic pile, which has near-surface temperature, will act as a perturbation on the thermal system, since the volcanic pile would not initially be in thermal equilibrium with its surroundings at depth. The volcanic section would then act as a block of initially colder material; drawing heat from its surroundings until thermal equilibrium is reached, Fig. 9b (adapted from Bjørlykke, 2010) depicts, simplified and qualitatively only, how the effect of rapid burial/uplift will affect the geothermal gradient and heat flow.

This effect, commonly termed „thermal blanketing“ (lowering of the surface heat flow), has been modelled on the Vøring Margin, where high deposition rates of several hundred meters

per million years (the highest deposition rates occurring in Late Cretaceous of 1.2 km/Ma) of cold sediments with low heat conductivity during the mid-Cretaceous to Paleocene (Wangen *et al.*, 2008) causes a temporal reduction in the surface heat flow. Also at basement level is the thermal blanketing effect apparent in the heat flow, but with more subdued and long waved temporal fluctuations than in the surface heat flow (Wangen *et al.*, 2008). Considering that the Faroese lavas have low heat conductivity (1.5-2.0 Wm/K; Balling *et al.*, 1984; Balling *et al.*, 2006) the same thermal blanketing effect might also happen if rapid emplacement of the volcanic section takes place as evoked by the age-diagnostic palynological data. It should also be noted that the overall strain rate for the basin is influencing the effect of thermal blanketing, a slow strain rate would allow for more time to cool the raised mantle compared to a higher strain rate (if everything else is kept equal).

The pre-volcanic structuration, topography and the contrast between low thermal conductivity sediments and higher conductive basement also play an important role in the thermal structure, with the heat flow from depth preferentially being conducted into flanking basement highs (Jaeger, 1965; Giles *et al.*, 1999; Ritter *et al.*, 2004). In some instances does this create heat refraction (e.g. Pismo Basin, California, Heasler and Surdam, 1985; Vøring Basin, Ritter *et al.*, 2004), reducing the heat flow in the centre of the basin. The consequence of this on the temperature distribution within the basins is not trivial and will depend (but not exclusively) on basin geometry, subsidence history, thermal properties and the thermal evolution at basement level.

Another subject, which could arise from rapid burial, is the development of overpressured sections below the volcanic sequence, if a competent seal is present (e.g. the shales of the Lamba Formation or Upper Cretaceous shales) and adequate lateral compressive stresses (e.g. by vertical barriers like sealing faults) is hindering the pressure build-up to dissipate out laterally (Osborne and Swarbrick, 1997). This mechanical compaction

might result in the development of an undercompacted section. Two effects commonly observed from undercompaction are higher porosities and lower velocities, but also a lowering of thermal conductivities (e.g. Gulf of Mexico; Mello and Karner, 1996).

If transient overpressure builds up in the deeper sections e.g. at source rock level, the mechanism of pressure retardation of the source rock might take place (Carr and Scotchman, 2003; Scotchman and Carr, 2005). Using vitrinite reflectance as a proxy for source rock maturation and hydrocarbon generation, Scotchman and Carr (2005) used a pressure-dependent vitrinite reflectance basin model with a multiple heat flow history based on both the tectonic and volcanic history for the UK part of the Faroe-Shetland Basin to predict hydrocarbon generation in well 204/19-1. The effect of including a temperature and pressure dependant vitrinite reflectance model suggests that maturation and expulsion of hydrocarbons is stalled as overpressure builds-up and resumes when overpressure is reduced e.g. at the Late Oligocene-Mid Miocene compressive event. Furthermore, overpressure influences the hydrocarbon phase generated and their timing; pressure retardation extends the temperature regime at which hydrocarbons in the oil phase is generated compared to a normal pressured situation (Scotchman and Carr, 2005). As a consequence, besides the added complications of drilling at augmented pressure regimes, overpressure could potentially enhance prospectivity by the extended temperature regime for oil generation, but to date only the T10 reservoir sands of the Sullom Formation (only penetrated by well 6004/16-1z) has shown evidence of overpressure, where MDT data indicates an overpressured section of about 1500 psi compared to a normally pressured gradient (from the End of Well Report).

Conclusions

With only seven offshore exploration wells drilled on the Faroese Continental Shelf, the region is still in its infancy and still a frontier

area. What the well results have demonstrated is the presence of a functioning hydrocarbon system, existence of clean reservoir sections within the Vaila Fm but with a lack of competent sealing lithologies in the Vaila Fm, a vital constituent in the stratigraphic/combined play types targeted by the first four wells drilled in the non-volcanic covered Judd sub-basin. Since then the exploration effort has changed to the structural sub-volcanic plays along the intrabasinal highs. The primary risk lies in what effect the volcanics of the North Atlantic Igneous Province has on the prospectivity of these sub-volcanic plays. The confirmation of the presence of the Lamba Formation below the volcanic section in the first sub-volcanic well and the thick volcanic section encountered in the well suggests a rapid emplacement of the volcanic section and hence large burial rates (~ 1 km/Ma). Such large burial rates have in other parts of the world led to overpressured sections and may also lead to overpressured sub-volcanic sections with the possibility of improved preservation of sub-volcanic reservoir properties and delay in source rock maturity through pressure-retardation.

Even with well control is seismic mapping of the volcanic section and especially the base of the volcanic section very difficult, commonly does the base of the volcanic section not appear as a sharp boundary/strong reflector on the seismic data, making mapping of the base of the volcanics and hence estimating the thickness of the volcanic section, over even short distances between well control (15-20 km) difficult. This is exemplified by the ambiguities in tying the volcanic thickness in the two wells 6005/13-1 and 6005/15-1 where the encountered volcanic thicknesses in the wells are 1475 m and 95 m, respectively. A second implication is that mapping of the interfingering volcanic and siliciclastic system is difficult if the mapping is only based on tracing reflectors on seismic sections.

Studies of reservoir quality of the volcaniclastic sediments exposed onshore Faroe Islands have shown that though the porosity in these sandstones /mudstones can be up to 18-20 %, the permeability is quite variable and generally

being on the low side, 0.1-1 mD making these intra-basalt sediments likely to be only secondary targets.

Therefore a firm establishment of the presence of reservoir sections, their quality and age, pressure regime and fluid composition and in a discovery case, hydrocarbon phases, underneath the volcanics is needed. With the improvements in the technical aspects i.e. sub-volcanic seismic imaging and drilling performance, the opportunity for a success should be closer to materialise itself.

At present time, two wells have been committed to be drilled over the next couple of years, one on license L008 and one well on license L006.

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