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Magmatic and tectonic evolution of the North Atlantic

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Abstract: The primary aim of the present paper is (1) to review the tectonomagmatic evolution of the North Atlantic, and (2) constrain evolutionary models with new lithosphere strength estimates and interpretation of potential field data north of Iceland. Our interpretations suggest that the breakup along the entire eastern Jan Mayen Ridge occurred at *c*. 55 Ma. Calculations of lithospheric yield strength indicate that the continental rifting in East Greenland, which led to oceanic crustal formation west of the Jan Mayen Ridge at *c*. 25 Ma, could have started at *c*. 42.5 Ma. Symmetrical V-shaped gravimetric ridges, which can be traced back to *c*. 48 Ma, document large-scale asthenospheric flow both north and south of Iceland. Such flow is predicted by geodynamic models of mantle plumes, but has yet to be predicted by other mechanisms. The results from the compartments north of Iceland, viewed in a regional context, strengthen the hypothesis attributing the anomalous magmatism in the North Atlantic area from *c*. 70 Ma to the present to the Icelandic plume.

Following the breakup of Pangaea at *c*. 55 Ma the North Atlantic has experienced extensive volcanism. As one of the large igneous provinces of the world, this region is influenced by the Iceland hotspot. This hotspot is often considered to be the surface expression of a convection plume of anomalously hot material rising from the deep mantle (e.g. Morgan 1971). However, the Icelandic plume hypothesis has recently been questioned and other mechanisms such as small-scale convection (e.g. Mutter *et al*. 1988) or fertile mantle melting (e.g. Foulger *et al*. 2005) have been proposed. Continental rifting and sea-floor spreading have also been integral to the evolution of this basin and therefore the temporal and spatial linkages between tectonic and magmatic events provide important clues to the origin of the Icelandic hotspot.

The sea-floor compartment NE of Iceland, bounded by the Iceland–Faeroes and the Jan Mayen fracture zones, is complex (Fig. 1; Talwani & Eldholm 1977). Here, oceanic crust accreted along the now extinct Aegir Ridge while continental stretching was occurring in East Greenland. The process rifted off the continental Jan Mayen Ridge, and oceanic spreading commenced along the Kolbeinsey Ridge at *c*. 25 Ma, when the Aegir Ridge became extinct (e.g. Vogt *et al*. 1980). Understanding of this compartment has improved significantly recently (e.g. Kodaira *et al*. 1997, 1998; Breivik *et al*. 2006; Mjelde *et al*. 2007*b*), and we will argue that the Iceland–Jan Mayen area holds many of the key observations needed to develop a coherent tectonomagmatic model for the Tertiary evolution of the North Atlantic. We will review recent work in the area, primarily based on wide-angle, ocean bottom seismic (OBS) data, and constrain the results by new interpretations of potential field data. Furthermore, we will use information from the OBS data to constrain the composition of the lithosphere and determine its mechanical strength. Finally, we will calculate the regional variations in oceanic magmatic production since breakup.

Regional background

The area's plate tectonic evolution has been summarized by Torsvik *et al*. (2001). All the continents moved towards the NNE during the Late Cretaceous, until Early Campanian time (*c*. 80 Ma), when there was an abrupt change to NW movement corresponding to the onset of the final rifting of the NE Atlantic (e.g. Skogseid *et al*. 2000). The breakup at *c*. 55 Ma is marked by a clear NE-directed kink in the motion of the Eurasian plate. From anomaly 13 (*c*. 35 Ma), the Greenland plate became attached and began moving with the North American plate, and its more NW-directed movement allowing the Atlantic to continue its opening further northwards towards the polar basin. The fusing of the Greenland plate to the North American plate appears to have started earlier at its southernmost tip (anomaly 17), than further north (anomaly 13, Jan Mayen Fracture Zone).

The Iceland hotspot (Fig. 2) is inferred to have been located at the Greenland shelf edge at *c*. 40 Ma, when focused magmatism occurred there (Torsvik & Cocks 2005). The northern Mid-Atlantic Ridge has gradually moved westward relative to the hotspot since continental breakup. The current westward motion of both the North American plate (*c*. 27 mm a^{-1}) and the European plate (*c*. 14 mm a^{-1}) indicates that the ridge axis is still moving west relative to the Iceland hotspot at the present day (Gripp & Gordon 2002).

Several features define the northern boundary of the Greenland–Faeroe Ridge north and east of Iceland. The Iceland–Faeroe Fracture Zone marks an abrupt northeastward decrease in crustal thickness (Kimbell *et al*. 2004). The Clair Lineament is one feature that probably formed the precursor to the Iceland–Faeroe Fracture Zone. Modelling of OBS data has shown that the area to the north of the Clair Lineament represents the continuation of the Palaeozoic–Early Mesozoic North Sea rift, whereas the region to the south of the lineament is dominantly related to the NE-propagating Late Mesozoic rift (Raum *et al*. 1995). North of Iceland, the Tjørnes Fracture Zone represents a 4 Ma old and 150 km lateral shift linking the spreading along the Northern Volcanic Zone in Iceland to the Kolbeinsey Ridge (Sigmundsson 2006).

The Jan Mayen Fracture Zone is the most significant NW-trending oceanic fracture zone segmenting the NE Atlantic

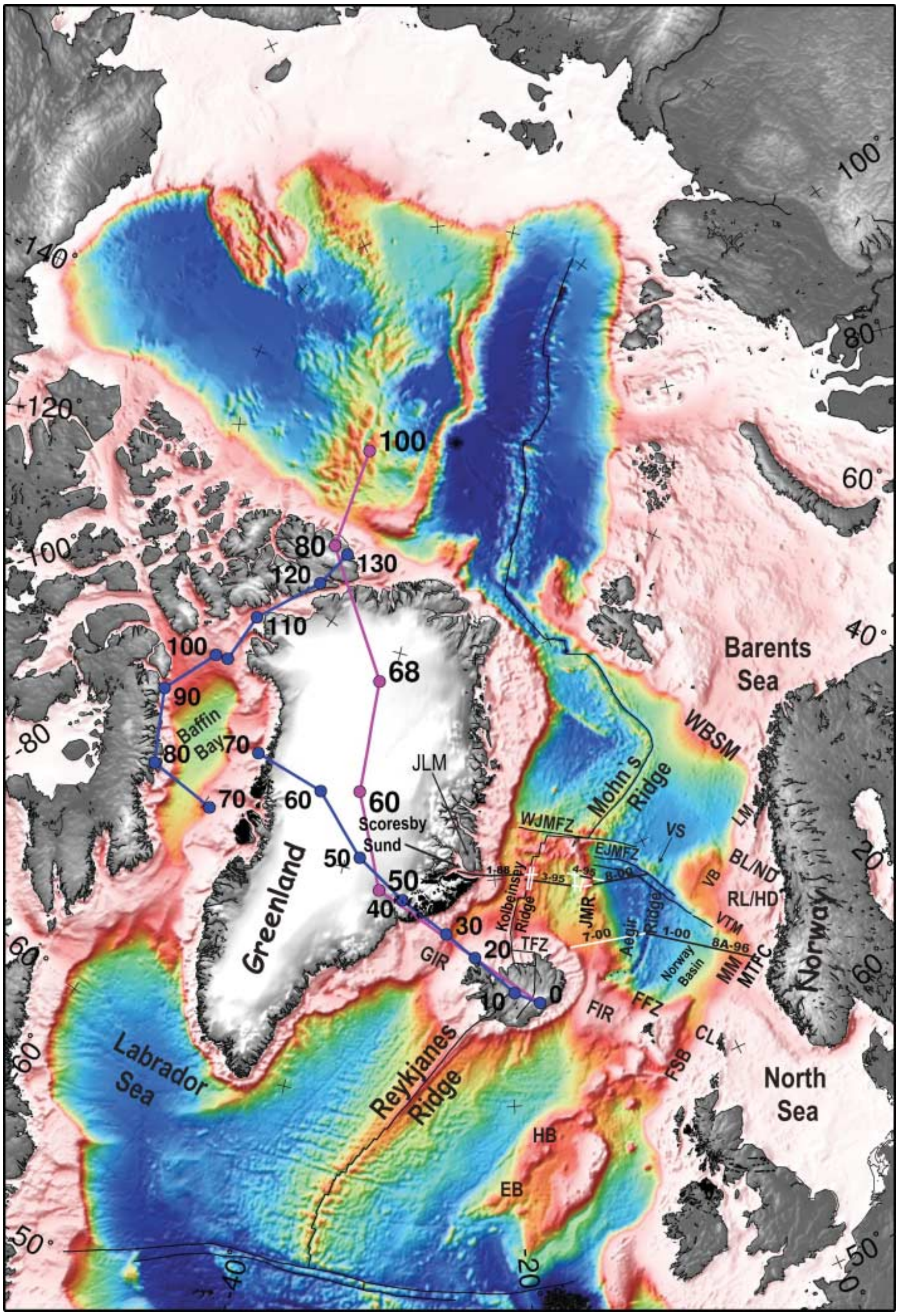


Fig. 1. Location map of the North Atlantic and Arctic. ETOPO-2 shaded relief bathymetry and topography are based on data from Sandwell & Smith (1997). Hotspot tracks proposed by Lawver & Müller (1994; blue) and Forsyth *et al*. (1986; purple) are included. Black areas onshore Greenland and the British Isles indicate early Tertiary basalt flows or dykes. These belong to the North Atlantic Volcanic Province, which also includes breakup magmatism from Edoras Bank (EB) to the Western Barents Sea Margin (WBSM). GIR, Greenland–Iceland Ridge; FIR, Faeroe–Iceland Ridge; FSB, Faeroe–Shetland Basin; CL, Clair Lineament; FFZ, Iceland–Faeroe Fracture Zone; HB, Hatton Bank; TFZ, Tjørnes Fracture Zone; JMR, Jan Mayen Ridge; WJMFZ, West Jan Mayen Fracture Zone; EJMFZ, East Jan Mayen Fracture Zone; VTM, Vøring Transform Margin; VS, Vøring Spur; VB, Vøring Basin; BL/ND, Bivrost Lineament–Nesna Detachment; MM, Møre Margin; MTFZ, Møre–Trøndelag Fault Complex; RL/HD, Rån Lineament–Høybakken Detachment; LM, Lofoten Margin. The locations of OBS profiles are shown as black and white profile lines. Black profile lines are included in the transect discussed by Mjelde *et al*. (2007*b*). Profiles 1-00 and 8A-96 are from Breivik *et al*. (2006), 7-00 and 8-00 from Breivik *et al*. (unpublished), 4-95 from Kodaira *et al*. (1998), 3-95 from Kodaira *et al*. (1997), and 1-88 from Weigel *et al*. (1995).

Margin (Doré *et al*. 1999). Its easternmost part defines the Vøring Transform Margin (e.g. Raum *et al*. 2007). The East Jan Mayen Fracture Zone marks the trend of the initial opening of the North Atlantic between the Eurasian and Jan Mayen blocks up to the Early Oligocene jump in spreading axis, whereas the West Jan Mayen Fracture Zone records the movement between Eurasia and Greenland. There are no major oceanic fracture zones between the Jan Mayen Fracture Zone and the western Barents Sea Transform Margin, which marks the northern border of the Early Tertiary Atlantic Ocean (Faleide *et al*. 1993; Olesen *et al*. 2007).

Results

Interpretation of potential field data for the Kolbeinsey/Aegir Ridge compartment

Figure 3a–d showsthe bathymetry, free-air gravity map, Bouguer gravity map and magnetic anomaly map of the area around the Jan Mayen Ridge. These data and available OBS profiles will be used to interpret the continent–ocean boundary.

The continent–ocean boundary is relatively well constrained on both sides of the ridge along OBS profile 7-00 and on the eastern side of the ridge along profile 8-00 (Breivik *et al*.

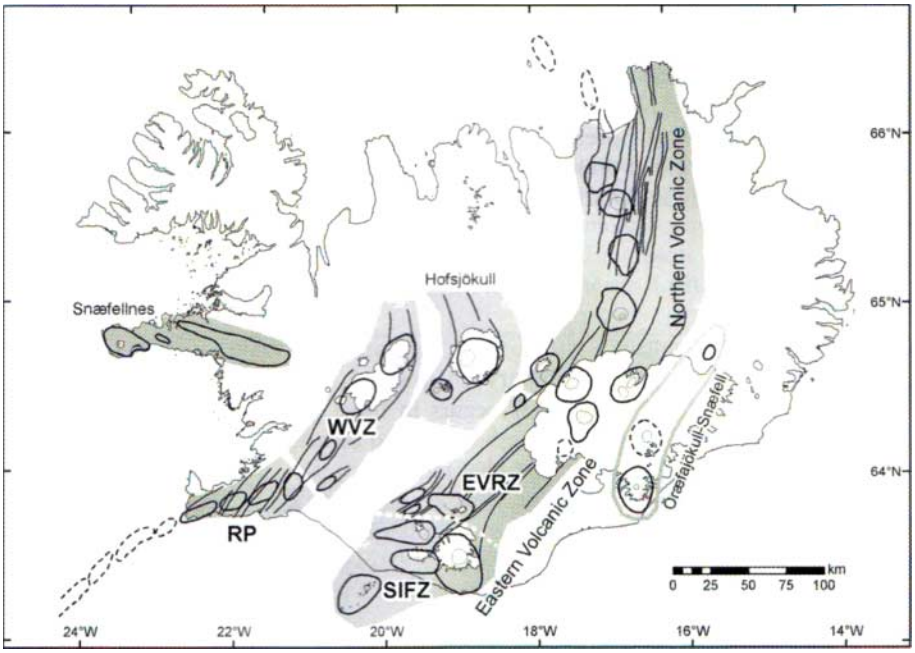


Fig. 2. Volcanic zones of Iceland (from Sigmundsson 2006). RP, Reykjanes Peninsula oblique rift; WVZ, Western Volcanic Zone; SIFZ, South Iceland Flank Zone; EVRZ, Eastern Volcanic Rift Zone.

unpublished). Between the two OBS profiles along the eastern side of the Jan Mayen Ridge the continent–ocean boundary is also marked by a steep gradient in the bathymetry, Bouguer map and free-air gravity map. Its imprint in the magnetic data is not clear because of poor data coverage, but the interpreted continent–ocean boundary roughly marks the boundary between clear magnetic anomalies formed along the Aegir Ridge and more irregular pattern of anomalies further westwards.

The continental part of the Jan Mayen Ridge terminates northwards at the southern trace of the East Jan Mayen Fracture Zone, which we extend from the reinterpreted location north of the Aegir Ridge (Breivik *et al*. 2006) to the offset in the Kolbeinsey Ridge. Interpretation of the East Jan Mayen Fracture Zone implies that the area between it and the West Jan Mayen Fracture Zone, characterized by shallow bathymetry, low Bouguer anomaly and positive free-air anomalies, represents an oceanic plateau related to the Jan Mayen magmatism. That magmatism has been attributed to a hotspot (e.g. Neumann & Schilling 1984), or to melting of metasomatized mantle (e.g. Trønnes *et al*. 1999). We infer that the magmatism is linked to the Icelandic hotspot through shallow transport of hot asthenosphere along the Kolbeinsey Ridge.

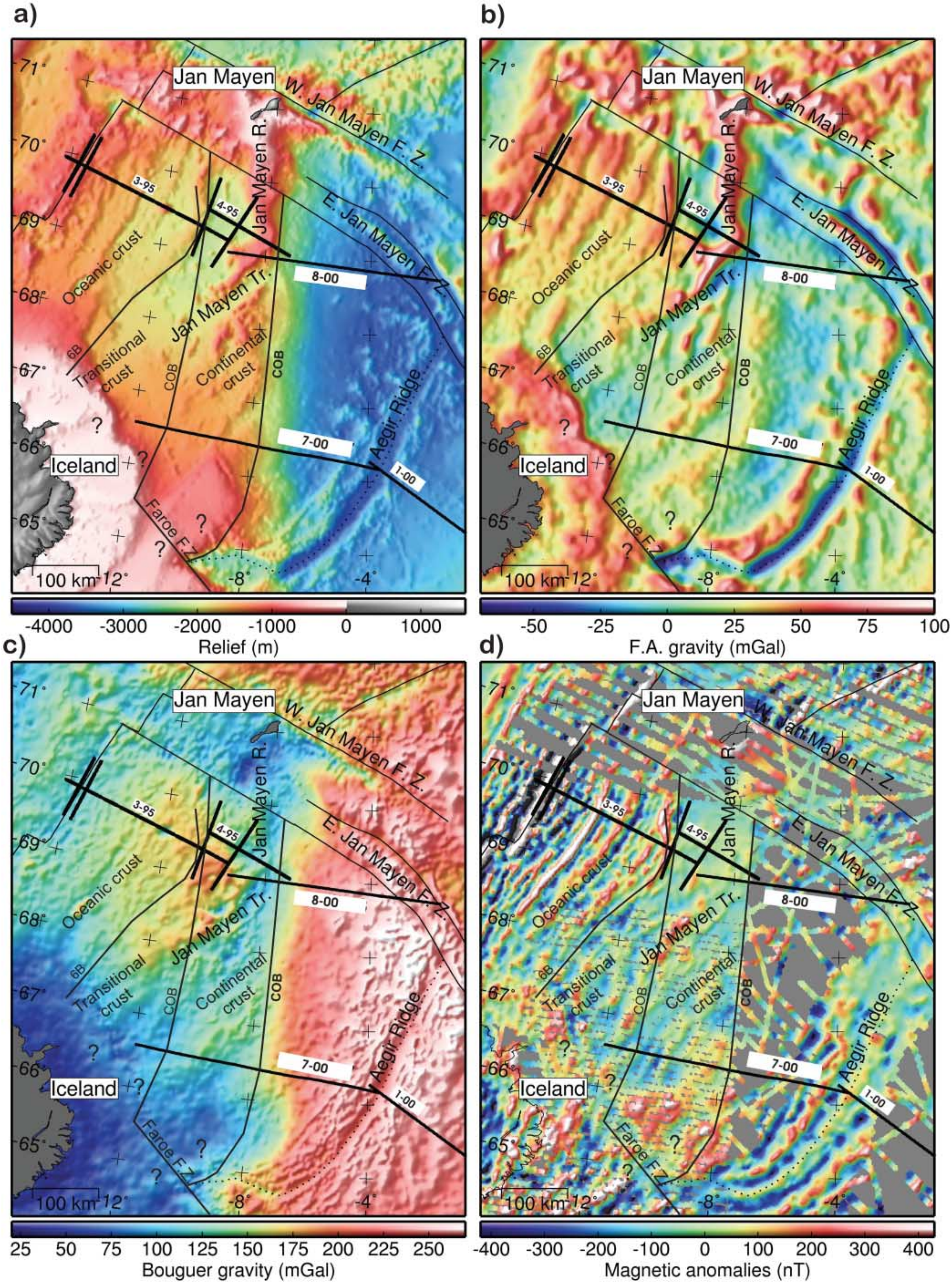
The southernmost part of the Jan Mayen Ridge is characterized by a bathymetric high (Fig. 3a), but its southward extent is poorly resolved. Potential field data show that the continent–ocean boundary on the eastern side of the ridge turns more southwesterly towards Iceland south of profile 7-00. We infer that the Jan Mayen Ridge terminates southwards at the inferred Iceland–Faeroe Fracture Zone (Kimbell *et al*. 2005), but the continental ridge may continue further southwestwards beneath the Icelandic shelf (e.g. Fedorova *et al*. 2005).

The continent–ocean boundary on the western side of the Jan Mayen Ridge is also uncertain. The boundary is well defined along profile 3-95, but the strong magmatic influence from Iceland along profile 7-00 makes interpretation difficult. Between the two OBS profiles, the continent–ocean boundary cannot be identified with certainty in the potential field data. However, the spreading along the Kolbeinsey Ridge is clearly expressed as magnetic anomalies from anomaly 6B (*c*. 23 Ma; Fig. 3d). The area between anomaly 6B and western continent–ocean boundary (Fig. 3), and NE of the Iceland shelf, is characterized by small-amplitude bathymetric ridges and a diffuse pattern of short gravimetric and magnetic anomalies. We interpret this area as a mixture of oceanic and continental crust, where the proto-Kolbeinsey Ridge propagated from Iceland and northwards into the strongly thinned western part of the Jan Mayen Ridge, possibly from Late Eocene time. This interpretation implying southward widening of the Jan Mayen Ridge as a result of progressively increased continental stretching in that direction agrees with the interpretation of multichannel seismic data (e.g. Kuvaas & Kodaira 1997). The enhanced stretching could indicate a broad zone of thin lithosphere associated with the Icelandic hotspot (Müller *et al*. 2001).

A similar model involving a propagating ridge axis as invoked here has been well documented in the Woodlark Basin in the western Pacific. In that non-volcanic area, the dominant form of rifting-to-spreading transition is not stress concentration at the tip of a propagating ridge axis, but stepwise spreading nucleation related to rheologically weak zones (Taylor *et al*. 1999). The rheological weak zones north of Iceland can partly be related to lateral heat transfer from the hotspot. Increased Icelandic hotspot activity at *c*. 23 Ma seems to have changed the system from diffuse oceanic accretion to concentrated spreading at the Kolbeinsey Ridge within the entire compartment.

Mechanical arguments for rift initiation west of the Jan Mayen Ridge

The timing of initial continental rifting and its cause were constrained by a series of calculations of lithospheric strength. Rifting occurs when the tectonic tensile stress is sufficient to permanently stretch the lithosphere. The yield stress at a given depth z is the minimum horizontal tensile stress in excess of lithostatic stress required to cause permanent (non-elastic) extension. We assume that this extension occurs by a combination of both brittle failure, following Byerlee's law, and ductile deforma-


a)
b)
c)
d)
Jan Mayen
W. Jan Mayen F. Z.
E. Jan Mayen F. Z.
Jan Mayen R.
Jan Mayen Tr.
Oceanic crust
Transitional crust
Continental crust
COB
6B
3-95
4-95
8-00
7-00
1-00
Aegir Ridge
Iceland
Faroe F.Z.
100 km
Relief (m)
-4000 -3000 -2000 -1000 0 1000
F.A. gravity (mGal)
-50 -25 0 25 50 75 100
Bouguer gravity (mGal)
25 50 75 100 125 150 175 200 225 250
Magnetic anomalies (nT)
-400 -300 -200 -100 0 100 200 300 400

tion, using a temperature-dependent, non-Newtonian rheology (e.g. Burov & Diament 1995). The yield stress is the weaker of the two deformation mechanisms and varies with depth, lithology, and temperature (Fig. 4d and e).

Temperature, T, beneath the continental lithosphere west of Jan Mayen is assumed to follow a conductive geotherm; T increases linearly from $T_o = 0°C$ at the surface to $T_{mantle} = 01300°C$ at the lithosphere–asthenosphere boundary and remains uniform below the that boundary. Beneath the Aegir Ridge, the geotherm is computed by solving the 1D, steady-state, advection–diffusion equation, assuming the spreading plates passively draw hot asthenosphere upward beneath the ridge at a rate v equal to the half-spreading rate. The solution as derived in the Appendix is

$$T = T_{mantle}\left[1 - \exp\left(-\frac{v}{\kappa}\right)z\right] + T_o\exp\left(-\frac{v}{\kappa}\right)z$$

where κ is the thermal diffusivity.

Lithology as a function of depth at the two localities is constrained from the seismic profiles (Mjelde *et al*. 2007*a*). Table 1 lists the rheological parameters used. Crustal thicknesses as well as spreading rate v at the Aegir Ridge are those estimated by Breivik *et al*. (2006) (see Fig. 4a and b). Continental crustal thickness is *c*. 17.5 km prior to rifting at 50 Ma and *c*. 5 km after the onset of rifting at *c*. 25 Ma (Mjelde *et al*. 2007*a*). From the computed geotherms and lithological parameters, we calculate yield stress and then integrate over depth to compute the total strength of the lithosphere at the two locations.

The least constrained parameter in these calculations is the depth of the lithosphere–asthenosphere boundary beneath the continent. Geophysical evidence shows that coeval rifting between the two locations was already established by 25 Ma, and we use this point to define the lithosphere–asthenosphere boundary at this time. Assuming coeval rifting can only occur if the strength of the continental lithosphere is approximately equal to that beneath Aegir Ridge, the predicted continental lithosphere–asthenosphere boundary at 25 Ma was at a depth of 18.5 km (Fig. 4d and e). Earlier strengths of the rifting continent assume that the crust thinned at a constant rate prior to 25 Ma and that the lithospheric mantle thinned either at the same fractional rate as the crust or more slowly.

The first important finding is that if the vertical strain rate in the crust and mantle portions of the continental lithosphere were equal, at no time prior to 25 Ma could coeval rifting have occurred: the continental lithosphere would always have been too strong for rifting to have initiated prior to 25 Ma (Fig. 4c). We

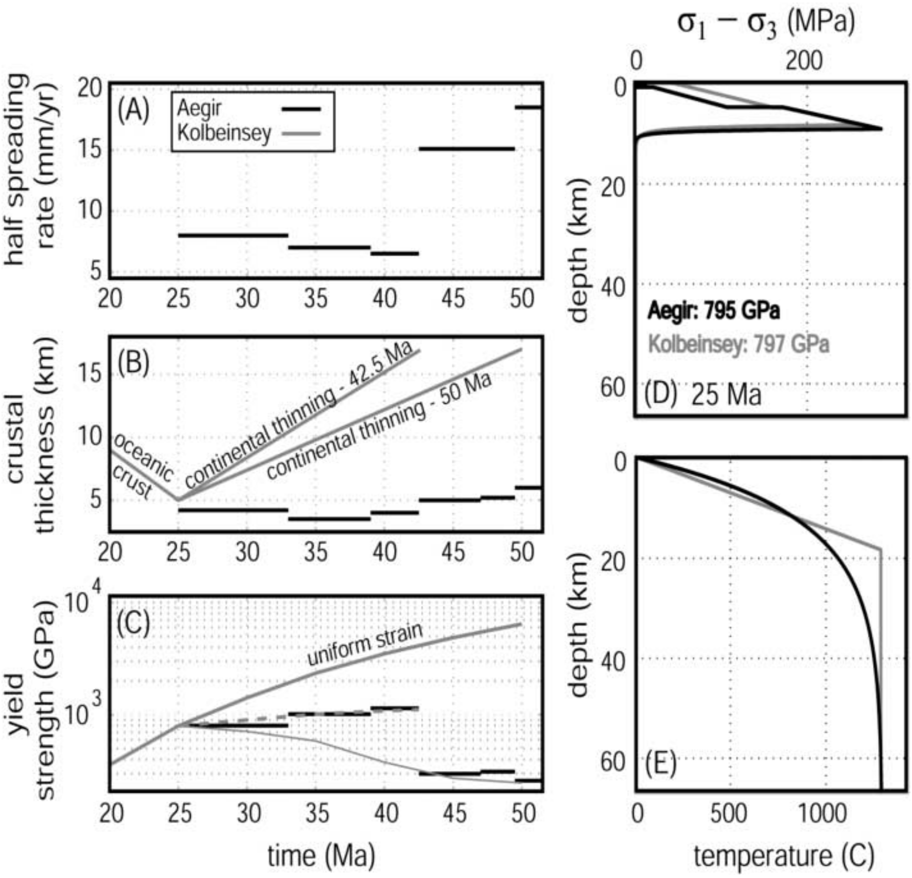


Fig. 4. (**d**) Yield stress envelopes for both the Aegir Ridge (black line) and the area west of the Jan Mayen Ridge (grey line) at 25 Ma. Horizontal axis is tensile stress in excess of lithostatic stress. Similar calculations for earlier than 25 Ma are based on estimates of (**a**) spreading rate, (**b**) crustal thickness, (**e**) computed temperature profiles and seismically interpreted lithologies at both locations (Table 1). Integrating the stress envelopes over depth, estimates of (**c**) lithospheric strength are calculated for the Aegir Ridge and for three models of continental thinning west of the Jan Mayen Ridge. Thinning of the continental crust is assumed to be constant after rifting began, and simultaneous thinning of the mantle portion of the lithosphere is assumed to occur at an equal rate beginning at 50 Ma (thick grey line), or at a slower rate with rifting beginning at either 50 Ma (thin grey line) or 42.5 Ma (dashed grey line).

Fig. 3. Our interpretation of the continent–ocean boundary and oceanic fracture zones are shown on different datasets. Shaded relief is illuminated from the west. (**a**) Bathymetry in the Jan Mayen Ridge area, with location of OBS profiles indicated. (**b**) ERS-1 (KMS version) based free-air gravity map of the Jan Mayen Ridge area (Andersen & Knudsen 1998). (**c**) Bouguer gravity map made from the ERS-1 satellite map and the bathymetry. (**d**) Magnetic map of the Jan Mayen Ridge area (Verhoef *et al*. 1996).

Table 1. *Lithological and rheological parameters used in the yield strength calculations shown in Figure 4*

Lithology	Δz (km) [25/45/50 Ma]	ρ (kg m^{-3})	A (Pa^{-n} s^{-1})	n	Q (J mol^{-1})	C_0 (Pa)	Reference
Oceanic							
Diabase	5.0/5.0/5.0	3000	6.31e−20	3.05	2.76e5	20e6	1
Dry peridotite		3300	2.5e−2	3.5	5.32e5	20e6	2
Continental							
Sandstone	0.2/1.4/1.4	2000	4e3	2.1	0.4e5	0	3
Shale	0.8/5.6/5.6	2000	4e3	2.1	0.4e5	0	3
Granite	4.0/10/10	2700	8.0e−9	3.1	2.43e5	0	4
Dry peridotite		3300	2.5e−2	3.5	5.32e5	20e6	2

References: 1, Carter & Tsenn (1987); 2, Chopra & Paterson (1984); 3, Hsu & Nelson (2002); 4, Wilks & Carter (1990). The crustal thickness is given for 25, 45 and 50 Ma (Kodaira *et al*. 1998; Breivik *et al*. 2006). The continental sand/shale ratio (20% sandstone) and crystalline composition (felsic) have been derived from V_p/V_s measurements (Mjelde *et al*. 2007*a*). Parameters are (empirical values): A, pre-exponential constant; n, power-law rheology component (controls the non-linear relationship between stress and strain rate); Q, activation enthalpy; C_0, cohesion of the material.

therefore require the mantle portion of the continental lithosphere to have thinned at a slower fractional rate than the crust.

If continental rifting initiated at 50 Ma, a lithosphere–asthenosphere boundary depth of 35 km is required to produce an equally weak continental lithosphere (Fig. 4b). However, sea-floor spreading along the Aegir Ridge is predicted to stop at *c*. 42.5 Ma owing to the modelled large increase in strength associated with the reduction in both oceanic crustal thickness and spreading rate (from *c*. 15 to *c*. 8 mm a^{-1}). Alternatively, if we assume that continental thinning initiated at 42.5 Ma (31 km depth to lithosphere–asthenosphere boundary), the yield strengths of both localities remain nearly identical until 25 Ma. Shortly after 25 Ma, all extension is predicted to shift to the Kolbeinsey Ridge, as a result of it forming thick oceanic crust and corresponding weak lithosphere. Thus the initiation of rifting at 42.5 Ma is most likely. A lithosphere–asthenosphere boundary depth of 31 km just prior to continental rifting is much shallower than the present-day 70 km depth beneath east Greenland (Kumar *et al*. 2005), and its exposure to a high mantle temperature anomaly associated with the Iceland hotspot is the likely cause.

Discussion

Crustal thickness

The Edoras Bank–mid-Norwegian margin and its conjugate East Greenland margin are characterized by ‘breakup-related’, sea-ward-dipping wedges of basalt and thick igneous crust (Morgan *et al*. 1989; Mjelde *et al*. 2005). South of Iceland, the maximum igneous crustal thickness accreted to each margin at distances greater than 500 km from the hotspot track is *c*. 18 km (Table 2), which is significantly thicker than the world average of 7.1 km for crust formed at mid-oceanic ridges (White *et al*. 1992). The thickness of the oceanic crust decreases to 8–10 km around magnetic anomaly 21 (Hopper *et al*. 2003). Closer to the Greenland–Iceland–Faeroe Ridge, the maximum oceanic thickness at breakup exceeds 30 km. Along the Reykjanes Ridge, the crustal thickness is estimated to decrease gradually southwards from *c*. 20 km close to Iceland, to *c*. 7 km at 700 km distance from Iceland (Sigmundsson 2006).

Beneath central Iceland, the maximum crustal thickness is *c*. 40 km, and beneath Iceland's active rift zones the crustal thickness is generally of the order of 20 km (e.g. Brandsdóttir *et al*. 1997). A maximum igneous crustal thickness of *c*. 40 km near the East Greenland margin decreases to an average of *c*. 30 km along the Greenland–Iceland–Faeroe Ridge (Holbrook *et al*. 2001). On the conjugate margin, the maximum igneous thickness is estimated to be *c*. 35 km, decreasing to *c*. 25 km near the Icelandic shelf edge (Smallwood *et al*. 2002). The increased crustal thickness along the entire Greenland–Iceland–Faeroe Ridge is generally attributed to increased magmatism caused by the Icelandic hotspot.

North of Iceland, the oceanic crustal thickness adjacent to the Møre marginal high is modelled to be 11 km near the location of breakup, 6 km at 51 Ma and *c*. 5 km from anomaly 22 (Breivik *et al*. 2006). At the Kolbeinsey Ridge just north of the Iceland shelf, crustal thicknesses are *c*. 9.5 km (Hooft *et al*. 2006). On the Iceland shelf, the crustal thickness increase uniformly south-wards to *c*. 12 km. Further northwards along the Kolbeinsey Ridge, Kodaira *et al*. (1997) found that the crustal thickness has remained relatively stable at 8–9.5 km since *c*. 23 Ma.

On the Vøring Margin, the oceanic crustal thickness immediately after breakup at *c*. 55 Ma is *c*. 23.5 km, decreasing to *c*. 8 km at anomaly 22 (Mjelde *et al*. 2005). At the present day, crustal production along the Mohns Ridge is much reduced, with an estimated crustal thickness of only 4 km (Klingelhöfer *et al*. 2000).

Full spreading rates

South of Iceland, Smallwood & White (2002) estimated the full spreading rates to be 25–30 mm a^{-1} immediately after breakup, decreasing to an average of *c*. 20 mm a^{-1} after anomaly 21 (Table 2). For the first few million years of sea-floor spreading, spreading rates as high as 88 mm a^{-1} have been estimated off East Greenland (Larsen & Saunders 1998). The present spreading rate in Iceland is 19 mm a^{-1} (Sigmundsson 2006).

The spreading velocities along the Aegir Ridge are estimated to be 60 mm a^{-1} at 53–51.5 Ma, 35 mm a^{-1} at 51.5–49 Ma and *c*. 30 mm a^{-1} from anomaly 22 (Breivik *et al*. 2006). The spreading rates along the Kobeinsey Ridge are calculated to be 15–20 mm a^{-1} (Appelgate 1997). North of the Jan Mayen Fracture Zone, estimates of the spreading rates between anomaly 24B and 23 vary from *c*. 40 to 50 mm a^{-1}, decreasing to *c*. 30 mm a^{-1} at anomaly 22 (Eldholm *et al*. 1984; Tsikalas *et al*. 2002; Torsvik & Cocks 2005). The average spreading rate since anomaly 21 is estimated to be *c*. 17 mm a^{-1} (Torsvik & Cocks 2005).

Table 2. *Full (average) spreading rate, total thickness of the (oceanic) magmatic crust and total magmatic production for the North Atlantic segments as a function of time*

Variable	Mohns (Lofoten)	Mohns (Vøring)	Aegir	Greenland–Faeroe	Reykjanes
Full spreading rate at breakup (mm a^{-1}, 55 Ma)	45	45	60	45	60
Full spreading rate at anomaly 23 (51 Ma)	35	35	35	30	35
Full spreading rate at anomaly 22 (49 Ma)	30	30	32	25	32
Full spreading rate at anomaly 20 (43 Ma)	17	17	30	20	30
Magmatic thickness at breakup (km)	9	21	11	38	17
Magmatic thickness at anomaly 23 (km)	6.5	12	8	33	11
Magmatic thickness at anomaly 22 (km)	5.5	7	5.5	31	9
Magmatic thickness at anomaly 20 (km)	5.5	5.5	5.5	27.5	5.5
Total magmatic production at breakup (km^3 km^{-1} Ma^{-1})	400	950	660	1710	1020
Total magmatic production at anomaly 23 (km^3 km^{-1} Ma^{-1})	230	420	280	990	390
Total magmatic production at anomaly 22 (km^3 km^{-1} Ma^{-1})	170	210	180	780	290
Total magmatic production at anomaly 20 (km^3 km^{-1} Ma^{-1})	90	90	160	550	160

Magmatic productivity

Close to the Greenland–Iceland–Faeroe Ridge south of Iceland, the estimated crustal thicknesses and average spreading rates imply a total magmatic productivity for the two conjugate margins of *c*. 1700 km^3 km^{-1} Ma^{-1} (Table 2). The corresponding value at distances >500 km from the hotspot track is 700 km^3 km^{-1} Ma^{-1}. The generation of the thick crust in Iceland requires magma production averaging about 570 km^3 km^{-1} Ma^{-1} (Sigmundsson 2006).

For the Aegir Ridge compartment, the conjugate margin productivity is estimated to be *c*. 660 km^3 km^{-1} Ma^{-1} near breakup, decreasing to *c*. 280 km^3 km^{-1} Ma^{-1} at 51.5 Ma and 180 km^3 km^{-1} Ma^{-1} from anomaly 22. At the Kolbeinsey Ridge just north of the Iceland shelf, a magma production rate of *c*. 190 km^3 km^{-1} Ma^{-1} is calculated from a full spreading rate of *c*. 20 mm a^{-1} (Hooft *et al*. 2006). On the Iceland shelf, the melt flux of the Kolbeinsey spreading centre increases uniformly southwards to values of 240 km^3 km^{-1} Ma^{-1}. At radial distances of 200–500 km from the Iceland hotspot centre, the crust is 2.0–2.5 km thinner along the Kolbeinsey Ridge than to the south along the Reykjanes Ridge. This difference corresponds to an increase in melt flux of about 20% at the Reykjanes Ridge relative to the Kolbeinsey Ridge, which Hooft *et al*. (2006) interpreted as indicative of a tilted plume. Further northwards along the Kolbeinsey Ridge, the melt flux is estimated to be 120–190 km^3 km^{-1} Ma^{-1}.

North of the Jan Mayen Fracture Zone, the conjugate margin production is calculated to be *c*. 950 km^3 km^{-1} Ma^{-1} during anomaly 24B-23, decreasing to *c*. 210 km^3 km^{-1} Ma^{-1} around anomaly 22. The corresponding value for the present-day spreading along the Mohns Ridge is *c*. 7 km^3 km^{-1} Ma^{-1}.

Potential field V-shaped ridges and outward flow from Iceland

The most prominent features south of Iceland are the V-shaped ridges clearly observed in the gravity field, and generally related to a pulsating plume (e.g. Vogt 1971). The angle of the V-shaped ridges with respect to the Reykjanes Ridge can be explained by a localized zone of excess crustal production propagating southward along the ridge at a rate of 100–250 mm a^{-1}. The V-shaped ridges can be traced back to earliest Oligocene time (*c*. 37 Ma).

The V-shaped ridges in the gravity field have also been observed near the Kolbeinsey Ridge. Jones *et al*. (2002) indicated that the asthenospheric transport rate here is similar to that south of Iceland. Recently, Breivik *et al*. (2006) have described similar V-shaped ridges also along the Aegir Ridge, which record outward flow from the Icelandic hotspot from anomaly 21 (*c*. 48 Ma), about 2–3 Ma after the strong pulse of breakup magmatism ceased here. The asthenospheric flow velocity along the Aegir Ridge was as low as 3–6 mm a^{-1}.

Breivik *et al*. (2008) recorded the presence of a significant oceanic plateau, the Vøring Spur, located north of the East Jan Mayen Fracture Zone along a northward projection of the Aegir Ridge. The maximum crustal thickness beneath the plateau is *c*. 16 km. The plateau is located on Middle–Late Eocene oceanic crust, but an uplift-related unconformity suggests Late Miocene formation (Breivik *et al*. 2008). The formation of the plateau is attributed to northeastward flow of asthenosphere away from the Icelandic hotspot, in some kind of interaction with the Aegir Ridge

Ridge jumps

Further evidence for variation in hotspot influence south of Iceland is found in the magnetic data. Excepting a ridge jump in the northern part of the area between anomaly 22 and 21, the oldest anomalies are smoothly aligned along the trend of continental breakup (e.g. Smallwood & White 2002). A change in plate motion led to oblique spreading and formation of a segmented ridge from south to north with 30–80 km long segments south of Iceland, at anomalies 17 to 13. Increased activity from the Iceland hotspot, commencing between anomaly 9 and 6, produced a southward progressing shift from segmented to continuous ridge axis without any transform offsets (White 1997).

Mapping in Iceland has shown repeated eastward jumps of the rift zones. The rift zone at Snæfellsnes was the main locus of spreading from *c*. 15 Ma to *c*. 7 Ma (Fig. 2; e.g. Sigmundsson 2006). During that time, the Snæfellnes–Skagi Rift Zone linked directly to the Kolbeinsey Ridge north of Iceland. The Western Volcanic Zone and the Northern Volcanic Zone have been active for 6–7 Ma, whereas the activity in the Eastern Volcanic Zone began 2–3 Ma ago. Lithospheric heating by the penetrating magma is sufficient to cause the observed ridge jumps (Mittelstaedt *et al*. 2007). The Reykjanes Ridge is linked to the Western Volcanic Zone through the Reykjanes Peninsula oblique rift, whereas the 4 Ma old Tjörnes Fracture Zone represents a 150 km lateral shift in the spreading axis from the Northern Volcanic Zone to the Kolbeinsey Ridge north of Iceland. Jones *et al*.

(2002) suggested that time-dependent flow in the Iceland plume, generating the V-shaped ridges along the Reykjanes Ridge, triggered the rift jumps in Iceland.

Numerous rift jumps have been inferred along the Greenland–Iceland–Faeroe Ridge, throughout its existence. The many rift jumps produce a pattern of magnetic anomalies along the Greenland–Iceland–Faeroe Ridge that defies reliable interpretations (Smallwood & White 2002). From *c*. 30 Ma, the local axis of spreading has remained centred over the hotspot while North America and the rest of the mid-Atlantic Ridge have steadily migrated WNW.

Symmetry

Two of the three existing conjugate margin pairs in the compartment south of Iceland indicate symmetrical margin segments, whereas one model suggests that nearly double the volume of oceanic crust was emplaced near the Greenland margin (Hopper *et al*. 2003). The latter observation may possibly be related to movements of the plates: the Eurasian plate has had small relative movement to the plume since breakup compared with the North American plate (Gripp & Gordon 2002). The faster movement of the North American plate might have induced larger amounts of decompression melting near the East Greenland margin.

Sea-floor spreading appears to be asymmetric about the Aegir Ridge throughout most of its development, with the western plate moving more slowly away from the ridge than the eastern plate (Breivik *et al*. 2006). The condensed *c*. 50 km of oceanic crust along the western side of the Aegir Ridge represents the amount of contemporaneous continental stretching west of the Jan Mayen Ridge. North of the Jan Mayen Fracture Zone the spreading has been uniform and symmetrical along the Mohns Ridge since the opening (Olesen *et al*. 2007).

Mantle plume v. non-plume magmatic hypotheses

Supporting arguments for the Iceland hotspot representing a mantle plume (e.g. Saunders *et al*. 1997) are as follows.

(1) The North Atlantic region is characterized by a strong geoid anomaly, with an equipotential surface *c*. 40 m above the Earth's reference ellipsoid. Geodynamical modelling indicates the presence of a low-density (high-temperature) anomaly in the lower mantle in this region (e.g. Marquart & Schmeling 2004).

(2) Dynamic modelling has shown that a plume with 100 km radius and excess temperature of *c*. 180 °C is capable of reproducing observed variations in along-axis crustal thickness, bathymetry and gravity field around Iceland (Ito *et al*. 2003).

(3) Evidence for Palaeogene transient uplift around Britain and Ireland can be readily explained by a sub-lithospheric thermal anomaly (Jones & White 2003).

(4) Tomographic experiments designed to resolve the upper mantle velocities reveal that these are lower than normal, generally attributed to elevated temperatures and the presence of melt (Tryggvason 1964; Wolfe *et al*. 1997; Foulger *et al*. 2001). The results may be interpreted as a vertical plume conduit between *c*. 400 and 200 km depth, and a horizontal plume head above *c*. 200 km (Allen *et al*. 2002).

(5) Studies of P-to-S converted waves from primary discontinuities at 410 and 660 km depth beneath Iceland have shown that this transition zone is anomalously thin, which is taken as an indication of mantle upwelling (e.g. Shen *et al*. 2002). The centre of the thin zone is located at least 100 km south of the upper mantle anomaly, indicating that the plume may be tilted *c*. 10° from vertical.

(6) Whole-mantle tomography suggests that the Iceland plume extends down to the core–mantle boundary (Bijwaard & Spakman 1999; Zhao 2004), and Burke & Torsvik (2004) inferred that the Iceland hotspot overlies a V_s anomaly of *c*. −0.5% in the D″ zone (the lowest 300–400 km of the mantle).

(7) Geochemical studies on the concentration of rare earth elements have revealed gradients in He, Hf, Nd, Pb and Sr ratios along the Mid-Atlantic Ridge for several hundred kilometres away from the centre of the Icelandic hotspot (e.g. Schilling 1973, 1999; Ito *et al*. 2003; Blichert-Toft *et al*. 2005). These observations suggest mixing of melts from a mantle plume source and an upper asthenospheric source depleted in light rare earth elements (e.g. Schilling 1973). Geochemical enrichment along the Mid-Atlantic Ridge is strongly asymmetric about Iceland.

However, the existence of volcanic margins at long distances from known hotspots has led many researchers to suggest that the production of excess magmatism is intrinsic to the dynamics of the rifting process itself and does not require a mantle plume. Buck (1986) and Mutter *et al*. (1988) proposed that strong focusing of rifting above a region of normal mantle temperatures produces small-scale convection. This would allow larger amounts of mantle to be fluxed through the melting region, generating significantly more melt than passive upwelling alone.

The mantle may be hotter than average under large, thick continents, as a result of insulation (e.g. Gurnis 1988). Elder (1976) demonstrated how a pulsating upflow of hot fluid could exist near the edge of a continent, and King & Anderson (1998) discussed the complex interaction between the thermal anomalies in the mantle beneath a continent and at the craton boundary.

Foulger *et al*. (2005) and Lundin & Doré (2005) attributed the observed enhanced magmatism in the Iceland region to high local mantle fertility leading to the formation of anomalously large volumes of melt. The source of the fertile region according to this model is eclogitized oceanic crust subducted during the Caledonian collision. The geochemical part of the model relates the observed compositional variations to various degrees of melting of eclogite and fractional remelting of abyssal gabbro.

Magmatism and extension prior to breakup

The earliest magmatism related to the North Atlantic Volcanic Province is found in seamounts in the Rockall Trough (70 Ma; O'Connor *et al*. 2000). Subsequently, extensive magmatism occurred in the NW British Isles, with a peak in magmatism from *c*. 62 to *c*. 52 Ma (e.g. Saunders *et al*. 1997). Contemporaneous magmatism is documented from eastern Baffin Island–West Greenland (peak: 62–60 Ma; Storey *et al*. 1998) and from East Greenland (peak: 58–48 Ma; e.g. Saunders *et al*. 1997).

Pre-breakup magmatism, NE of the Labrador Sea rift, is generally considered to be related to the Icelandic plume (e.g. White & McKenzie 1989; Saunders *et al*. 1997), but Lundin & Doré (2005) viewed it as a by-product of plate breakup. They interpreted the NW–SE alignment of magmatism as complementary to the contemporaneous NE–SW Palaeocene motion in the southern North Atlantic, Labrador Sea and Baffin Bay. However, rift-related magmatism would be concentrated directly along the zone of Labrador Sea rift and spreading, not within the thicker lithosphere to the NE. We thus attribute the extensive Early Tertiary magmatism extending from the British Isles to West Greenland to the Icelandic plume. The magmatism inferred landward of the Early Eocene line of breakup on the Vøring

Margin is well documented as intrusions in the sedimentary rocks and lower crust extending several hundred kilometres landward of the continent–ocean boundary.

Deep seismic crustal-scale transects and OBS profiles along the Møre and Vøring margins may indicate that the mid-Norwegian margin switches from simple-shear, lithospheric lower plate configuration on the Møre Margin to simple-shear upper plate on the Vøring Margin (Osmundsen *et al*. 2002; Mjelde *et al*. 2007*d*). In spite of these changes in structural style along strike, the arrival of the Icelandic plume apparently caused the system to break up at the western edge of the extensional basin, from the Edoras Bank to the Bivrost Lineament (see Fig. 1). The limited influence from the Iceland plume north of the Bivrost Lineament allowed the breakup to be dominantly focused by the simple-shear, tectonic style here.

Magmatism from breakup to present

Breakup magmatism follows a symmetrical pattern about Iceland, with thickest magmatic crust in the proximity of the plume. This is best documented along the East Greenland margin, where crustal thickness and active mantle upwelling along the margin is found to decrease with distance from the plume (Holbrook *et al*. 2001). Along the Greenland–Iceland Ridge, Holbrook *et al*. (2001) inferred an upwelling ratio about four times passive, which is in agreement with upwelling rates in fluid dynamical models of a ridge-centred hot plume (Ito *et al*. 1999). South of the Greenland–Iceland Ridge the thermal anomaly was exhausted within 6–12 Ma, and active upwelling was restricted to margin segments not exceeding *c*. 500 km distance from the Greenland–Iceland Ridge. Closer to the plume the breakup magmatism continued to anomaly 20 (43 Ma); that is, all the way to the present-day Iceland shelf edge along the Faeroe–Iceland Ridge. The Faeroe–Iceland Ridge thus consists of rocks emplaced during the breakup anomaly, whereas the Greenland–Iceland Ridge represents a mixture of breakup-related magmatism and later emplacement related to the hotspot track. The asymmetry of the Faeroe–Iceland Ridge, with the crustal thickening extending further to the SW than to the NE (Kimbell *et al*. 2005), may indicate that the plume has been tilted southwards (Shen *et al*. 2002) since its origin. Magmatism along the Møre and Vøring margins can be modelled by pure passive upwelling, or modestly active upwelling (Breivik *et al*. 2006; Mjelde *et al*. 2007*c*).

If the intrepretations are correct about a localized zone of active upwelling, then a likely source of buoyancy is excess temperature, which is confirmed by dynamic modelling (Nielsen & Hopper 2004). Enhanced upwelling associated with continental breakup, with normal mantle temperatures (Van Wijk *et al*. 2001), can explain crustal thicknesses in some locations along this margin (e.g. the Vøring plateau), but these models predict almost all of the melt to have erupted during *c*. 4 Ma prior to breakup, which is not confirmed by observations. Furthermore, the large basin width of at least 300 km near breakup (Mjelde *et al*. 2007*b*) indicates that significant small-scale convection cells probably did not exist at that time.

Weaknesses of the fertile mantle (Foulger *et al*. 2005; Meyer *et al*. 2007) being the main cause of excess magmatism are as follows.

(1) If subducted eclogites, generally assumed to be formed during the Caledonian orogeny (e.g. Christiansson *et al*. 2000), were the main cause of the Icelandic hotspot, we would have expected increased magmatism along the entire track of the Caledonian subduction zone, which is not observed.

(2) Bodies of lower crustal eclogitized rocks (V_p *c*. 8.4 km s^{-1}) have been inferred close to the mainland in the North Sea, in the Møre Basin, and on the Vøring Margin (e.g. Olafsson *et al*. 1992). The eclogites are not associated with excess magmatism in any of these areas. Lower crustal eclogites have been inferred at one location close to the line of breakup; in the southern Vøring Basin (Raum *et al*. 2007), and the eclogites in that area are associated with less magmatism compared with the surrounding areas. This observation can be explained by increased lithospheric thickness beneath the eclogites, leading to diminished decompression melting.

(3) Modelling of small-scale convection has shown that the period of increased magmatism is limited (e.g. Nielsen & Hopper 2004), and fertile upper mantle is unlikely to maintain the long period of excessive magmatism related to the Icelandic hotspot.

(4) A compositional anomaly cannot explain the well-documented asthenospheric flow both north and south away from the Iceland hotspot.

(5) Pyroxenite or eclogite is likely to form significant amounts of garnet in the mantle and thus be denser than lherzolite (e.g. Irifune *et al*. 1986), thus it is difficult for these rocks to rise and to melt. The presence of a hotspot centre is even more problematic in that a localized zone of mantle enriched in garnet would be denser, and thus it would be even harder to make it rise and melt.

Observations of mantle depletion and volatile enrichment in the Lau Basin, western Pacific, can help us understand how a system can change quickly from generating anomalously thick oceanic crust to very thin crust as observed on the Møre Margin. Spreading centres in the Lau back-arc basin exhibit enhanced and lower than normal magma supply, which correlates with distance from the volcanic front but not with spreading rate, as noted by Martinez & Taylor (2002). Those workers proposed that depleted upper mantle material, generated by melt extraction in the mantle wedge, is overturned and reintroduced beneath the back-arc basin. The spreading centres experience enhanced melt production near the volcanic front, and diminished melting within the overturned depleted mantle farther away. The formation of anomalously thin crust at the Aegir Ridge starting at about anomaly 21, could thus be explained by the mantle available for melt generation being partly depleted by melt extraction during the vigorous breakup magmatism, and therefore less fertile for the ensuing oceanic crustal accretion. Furthermore, indications that the magma budget available for generating sea floor was lower in the southern Norway Basin near the Faeroe–Iceland Ridge suggest the presence of enhanced mantle depletion beneath the southern part of the Aegir Ridge related to the construction of the Faeroe–Iceland Ridge (Breivik *et al*. 2006).

Summary and conclusions

Results from recent OBS surveys between Iceland and the Jan Mayen fracture zones are reviewed, and the geological evolution of this part of the North Atlantic is constrained by new interpretations of potential field data and calculation of yield strength. Continental breakup occurred at *c*. 55 Ma throughout the area. The last period of continental rifting and the first phase of oceanic spreading was associated with excess magmatism, caused by the Icelandic plume. From *c*. 42.5 Ma, thin oceanic crustal formation along the Aegir Ridge occurred simultaneously with continental stretching in East Greenland. This stretching was most rapid in the south and caused continental breakup to propagate northwards along the western side of the Jan Mayen Ridge, which in the northern part of the area was split off East Greenland at *c*. 25 Ma. From 23 Ma to the present day, the Kolbeinsey Ridge has generated anomalously thick oceanic crust.

Our preferred model for explaining the Early Tertiary North Atlantic volcanism is as follows. A plume head of radius *c*. 300 km (Holbrook *et al*. 2001) and ΔT of 100–200 °C affected western Greenland around 70 Ma. NE–SW extension facilitated intrusions of dykes from eastern Baffin Bay to the British Isles. The magmatism was at its peak from *c*. 57 to 52 Ma when the plume was located beneath central Greenland, and a *c*. 50 km thick layer of hot material (Nielsen & Hopper 2004) flowed beneath the thinnest portions of the lithosphere, generating extensive margin magmatism coinciding with breakup at *c*. 55 Ma. By *c*. 48 Ma most of the plume head, and consequently the distal warm layer, were exhausted, and excess magmatism was thereafter confined to a narrow, *c*. 100 km radius, zone immediately above the Iceland plume stem. The increased magmatism observed along the Kolbeinsey Ridge at *c*. 23 Ma may indicate direct plume–ridge interaction from that time (Torsvik & Cocks 2005), and the magmatic episode that formed the Vøring Spur north of the Aegir Ridge may be interpreted as evidence for increased plume activity at *c*. 10 Ma.

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Derivation of thermal profile beneath a spreading ridge assuming a uniform vertical velocity

A well-established spreading centre will be in steady state and a simple advection–diffusion equation will describe the geotherm directly beneath the ridge axis. If the vertical velocity, v, is assumed to be constant with depth, z, and equal to the half-spreading rate then the temperature, T, will be controlled by (1D)

$$\frac{\mathrm{d}^2 T}{\mathrm{d}z^2} + \frac{v}{\kappa}\frac{\mathrm{d}T}{\mathrm{d}z} = 0 \qquad (1)$$

where κ is the thermal diffusivity, z is positive downward and v is in the negative z direction. Using reduction of order, equation (1) is reduced to

$$\frac{\mathrm{d}u}{\mathrm{d}z} + \frac{v}{\kappa}u = 0 \qquad (2)$$

where $u = \mathrm{d}T/\mathrm{d}z$. The solution of equation (2) is simply

$$u = C_1 \exp\left(-\frac{v}{\kappa}\right)z. \qquad (3)$$

Replacing u with the vertical derivative of T and integrating we can obtain the solution of equation (3) with constants C_1 and C_2:

$$T = -\frac{\kappa}{v}C_1 \exp\left(-\frac{v}{\kappa}\right)z + C_2. \qquad (4)$$

Appropriate boundary conditions can now be applied to achieve the final solution. The temperature is assumed to go to T_{mantle} as z approaches infinity and the temperature is set to T_{o} at $z = 0$. This gives the final solution for the thermal profile beneath the ridge axis as

$$T = T_{\mathrm{mantle}}\left[1 - \exp\left(-\frac{v}{\kappa}\right)z\right] + T_{\mathrm{o}}\exp\left(-\frac{v}{\kappa}\right)z. \qquad (5)$$

The assumed uniform upwelling rate is a first-order approximation and tends to produce higher temperatures than the more realistic case in which v decreases toward the surface.

References

Allen, R.M. Nolet, G., Margan, W.J., *et al.* 2002. Imaging the mantle beneath Iceland using integrated seismological techniques. *Journal of Geophysical Research*, **107**, doi:10.1029/2001JB000595.

Andersen, O.B. & Knudsen, P. 1998. Global marine gravity field from the ERS-1 and Geosat geodetic mission altimetry. *Journal of Geophysical Research*, **103**, 8129.

Appelgate, B. 1997. Modes of axial reorganization on a slow-spreading ridge: The structural evolution of Kolbeinsey ridge since 10 Ma. *Geology*, **25**, 431–434.

Bijwaard, H. & Spakman, W. 1999. Tomographic evidence for a narrow whole mantle plume below Iceland. *Earth and Planetary Science Letters*, **166**, 121–126.

Blichert-Toft, J., Agranier, A., Andres, M., Kingsley, R., Schilling, J.-G. & Albarède, F. 2005. Geochemical segmentation of the Mid-Atlantic Ridge north of Iceland and ridge–hotspot interaction in the North Atlantic. *Geochemistry, Geophysics, Geosystems*, **6**, doi:10.1029/2004GC000788.

Brandsdóttir, B., Menke, W., Einarsson, P., White, R.S. & Staples, R.K. 1997. Faroe–Iceland Ridge experiment. 2: Crustal structure of the Krafla central volcano. *Journal of Geophysical Research*, **102**, 7867–7886.

Breivik, A.J., Mjelde, R., Faleide, J.I. & Murai, Y. 2006. Rates of continental breakup magmatism and seafloor spreading in the Norway Basin—Iceland Plume interaction. *Journal of Geophysical Research*, **111**, B07102, doi:10.1029/2005JB004004.

Breivik, A.J., Faleide, J.I. & Mjelde, R. 2008. MORB hijacking by the Iceland Plume. *Geochemistry, Geophysics, Geosystems*, in press.

Buck, R.W. 1986. Small-scale convection induced by passive rifting; the cause for uplift of rift shoulders. *Earth and Planetary Science Letters*, **77**, 362–372.

Burke, K. & Torsvik, T.H. 2004. Derivation of large igneous provinces of the past 200 million years from long-term heterogeneities in the deep mantle. *Earth and Planetary Science Letters*, **227**, 531–538.

Burov, E.B. & Diament, M. 1995. The effective elastic thickness (*T*e) of continental lithosphere: What does it really mean? *Journal of Geophysical Research*, **100**, 3905–3927.

Carter, N.L. & Tsenn, M.C. 1987. Flow properties of continental lithosphere. *Tectonophysics*, **136**, 27–63.

Chopra, P.N. & Paterson, M.S. 1984. The role of water in the deformation of dunite. *Journal of Geophysical Research*, **89**, 7861–7876.

Christiansson, P., Faleide, J.I. & Berge, A.M. 2000. Crustal structure in the northern North Sea: an integrated geophysical study. *In:* Nøttvedt, A. (ed.) *Dynamics of the Norwegian Margin*. Geological Society, London, Special Publications, **167**, 15–40.

Doré, A.G., Lundin, E.R., Jensen, L.N., Birkeland, Ø., Eliassen, P.E. & Fichler, C. 1999. Principal tectonic events in the evolution of the northwest European Atlantic margin. *In:* Fleet, A.J. & Boldy, S.A. (eds) *Petroleum Geology of Northwest Europe: Proceedings of the Fifth Conference*. Geological Society, London, 41–62.

Elder, J. 1976. *The Bowels of the Earth*. Oxford University Press, Oxford.

Eldholm, O., Sundvor, E., Myhre, A.M. & Faleide, J.I. 1984. Cenozoic evolution of the continental margin off Norway and western Svalbard. *In:* Spencer, A.M. (ed.) *Petroleum Geology of the North European Margin*. Norwegian Petroleum Society, Special Publication, **5**, 3–18.

Faleide, J.I., Va°gnes, E. & Gudlaugsson, S.T. 1993. Late Mesozoic–Cenozoic evolution of the south-western Barents Sea in a regional rift-shear tectonic setting. *Marine Petroleum Geology*, **10**, 186–214.

Fedorova, T., Jacoby, W.R. & Wallner, H. 2005. Crust–mantle transition and Moho model for Iceland and surroundings from seismic, topograpy and gravity data. *Tectonophysics*, **396**, 119–140.

Forsyth, D.A., Morel, A.L., Huissier, P., Asudeh, I. & Green, A.G. 1986. Alpha Ridge and Iceland; products of the same plume? *Journal of Geodynamics*, **6**, 197–214.

Foulger, G.R., Pritchard, M.J. & Julian, B.R. *et al.* 2001. Seismic tomography shows that upwelling beneath Iceland is confined to the upper mantle. *Geophysical Journal International*, **146**, 504–530.

Foulger, G.R., Natland, J.H. & Anderson, D.L. 2005. A source for Icelandic magmas in remelted Iapetus crust. *Journal of Volcanology and Geothermal Research*, **141**, 23–44.

Gripp, A.E. & Gordon, R.G. 2002. Young tracks of hotspots and current plate velocities. *Geophysical Journal International*, **150**, 321–361.

Gurnis, M. 1988. Large-scale mantle convection and the aggregation and dispersal of supercontinents. *Nature*, **332**, 695–699.

Holbrook, W.S., Larsen, H.C. & Korenaga, J. *et al.* 2001. Mantle thermal structure and active upwelling during continental breakup in the North Atlantic. *Earth and Planetary Science Letters*, **190**, 251–266.

Hooft, E.E.E., Brandsdottir, B., Mjelde, R., Shimamura, H. & Murai, Y., 2006. Asymmetric plume–ridge interaction around Iceland: the Kolbeinsey Ridge Iceland Seismic Experiment. *Geochemistry, Geophysics, Geosystems*, **7**, doi:10.1029/2005GC001123.

Hopper, J.R., Dahl-Jensen, T., Holbrook, W.S., *et al.* 2003. Structure of the SE Greenland margin from seismic reflection and refraction data: Implications for nascent spreading centre subsidence and asymmetric crustal accretion during North Atlantic opening. *Journal of Geophysical Research*, **108**, doi:10.1029/2002JB001996.

Hsu, S.C. & Nelson, P.P. 2002. Characterization of the Eagle Ford Shale. *Engineering Geology*, **67**, 169–183.

Irifune, T., Sekine, T., Ringwood, A.E. & Hibberson, W.O. 1986. The eclogite–garnetite transformations at high pressure and some geophysical implications. *Earth and Planetary Science Letters*, **77**, 245–256.

Ito, G., Shen, Y., Hirth, G. & Wolfe, C. 1999. Mantle flow, melting and dehydration of the Iceland mantle plume. *Earth and Planetary Science Letters*, **165**, 81–96.

Ito, G., Lin, J. & Graham, D. 2003. Observational and theoretical studies of the dynamics of mantle plume–mid-ocean ridge interaction. *Reviews of Geophysics*, **41**, doi:10.1029/2002RG000117.

Jones, S.M. & White, N. 2003. Shape and size of the starting Iceland Plume swell. *Earth and Planetary Science Letters*, **216**, 271–282.

Jones, S.M., White, N. & MacLennan, J. 2002. V-shaped ridges around Iceland: Implications for spatial and temporal patterns of mantle convection. *Geochemistry, Geophysics, Geosystems*, **3**, 1059, doi:10.129/2002GC000361.

Kimbell, G.S., Gatliff, R.W., Ritchie, J.D., Walker, A.S.D. & Williamson, J.P. 2004. Regional three-dimensional modeling of the NE Atlantic margin. *Basin Research*, **16**, 259–278.

Kimbell, G.S., Ritchie, J.D., Johnson, H. & Gatliff, R.W. 2005. Controls on the structure and evolution of the NE Atlantic margin revealed by regional 3D gravity modelling. *In:* Doré, A.G. & Vining, B.A. (eds) *Petroleum Geology: North-West Europe and Global Perspectives—Proceedings of the 6th Petroleum Geology Conference*. Geological Society, London, 933–947.

King, S.C. & Anderson, D.L. 1998. Edge-driven convection. *Earth and Planetary Science Letters*, **160**, 289–296.

Klingelhöfer, F., Geli, L. & White, R.S. 2000. Geophysical and geochemical constraints on crustal accretion at the very-slow spreading Mohns Ridge. *Geophysical Research Letters*, **27**, 1547–1550.

Kodaira, S., Mjelde, R., Shimamura, H., Gunnarsson, K. & Shiobara, H. 1997. Crustal structure of the Kolbeinsey Ridge, N. Atlantic, obtained by use of Ocean Bottom Seismographs. *Journal of Geophysical Research*, **102**, 3131–3151.

Kodaira, S., Mjelde, R., Gunnarsson, K., Shiobara, H. & Shimamura, H. 1998. Structure of the Jan Mayen micro-continent and implications for its evolution. *Geophysical Journal International*, **132**, 383–400.

Kumar, P., Kind, R. & Hanka, W. *et al.* 2005. The lithosphere–asthenosphere boundary in the North-West Atlantic region. *Earth and Planetary Science Letters*, **236**, 249–257.

Kuvaas, B. & Kodaira, S. 1997. The formation of the Jan Mayen microcontinent: the missing piece in the continental puzzle between the Møre–Vøring Basins and East Greenland. *First Break*, **15**, 239–247.

Larsen, H.C. & Saunders, A.D. 1998. Tectonism and volcanism at the southeast Greenland rifted margin: A record of plume impact and later continental rupture. *In:* Saunders, A.D., Larsen, H.C. & Wise, S.W. Jr. (eds) *Proceedings of the Ocean Drilling Program, Scientific Results, 152*. Ocean Drilling Program, College Station, TX, 503–533.

Lawver, L.A. & Müller, R.D. 1994. Iceland Hotspot track. *Geology*, **22**, 311–314.

Lundin, E. & Doré, A.G. 2005. NE Atlantic breakup: a re-examination of the Iceland mantle plume model and the Atlantic–Arctic linkage. *In:* Doré, A.G. & Vining, B.A. (eds) *Petroleum Geology: North-West Europe and Global Perspectives—Proceedings of the 6th Petroleum Geology Conference*. Geological Society, London, 739–754.

Marquart, G. & Schmeling, H. 2004. A dynamic model for the Iceland Plume and the North Atlantic based on tomography and gravity data. *Geophysical Journal International*, **159**, 40–52.

Martinez, F. & Taylor, B. 2002. Mantle wedge control on back-arc crustal accretion. *Nature*, **416**, 417–420.

Meyer, R., Van Wijk, J. & Gernigon, L. 2007. North Atlantic Igneous Province: A review of models for its formation. *In*: *The Origins of Melting Anomalies: Plates, Plumes and Planetary Processes*. Geological Society of America, Special Papers, in press.

Mittelstaedt, E., Ito, G. & Behn, M. 2007. Mid-ocean ridge jumps associated with hotspot magmatism. *Earth and Planetary Science Letters* (in press).

Mjelde, R., Raum, T., Breivik, A.J., Shimamura, H., Murai, Y., Takanami, T. & Faleide, J.I. 2005. Crustal structure of the Vøring Margin, NE Atlantic: a review of geological implications based on recent OBS data. *In:* Doré, A.G. & Vining, B.A. (eds) *Petroleum Geology: North-West Europe and Global Perspectives—Proceedings of the 6th Petroleum Geology Conference*. Geological Society, London, 803–814.

Mjelde, R., Eckhoff, I., Solbakken, S., *et al.* 2007a. Gravity and S-wave modeling across the Jan Mayen Ridge, North Atlantic; implications for crustal lithology and continental break-up processes. *Marine Geophysical Researches*, doi:10.1007/s11001-006-9012-3.

Mjelde, R., Raum, T., Breivik, A.J. & Faleide, J.I. 2007b. Crustal transect across the North Atlantic. *Earth and Planetary Science Letters*, in press.

Mjelde, R., Raum, T., Murai, Y. & Takanami, T. 2007c. A tectono-magmatic model of the Vøring Plateau, NE Atlantic, Continent–Ocean-Transition. *Journal of Geodynamics*, **43**, 374–392.

Morgan, W.J. 1971. Convection plumes in the lower mantle. *Nature*, **230**, 42–43.

Morgan, J.V., Barton, P.J. & White, R.S. 1989. The Hatton Bank continental margin—III: Structure from wide-angle OBS and multichannel seismic refraction profiles. *Geophysical Journal International*, **98**, 367–384.

Müller, R.D., Gaina, C., Roest, W.R. & Lundbek-Hansen, D. 2001. A recipe for microcontinent formation. *Geology*, **29**, 203–206.

Mutter, J.C., Buck, W.R. & Zehnder, C.M. 1988. Convective partial melting, a model for the formation of thick basaltic sequences during the initiation of spreading. *Journal of Geophysical Research*, **93**, 1031–1048.

Neumann, E.-R. & Schilling, J.-G. 1984. Petrology of basalts from the Mohns–Knipovich Ridge; the Norwegian–Greenland Sea. *Contributions to Mineralogy and Petrology*, **85**, 209–223.

Nielsen, T.K. & Hopper, J.R. 2004. From rift to drift: Mantle melting during continental breakup. *Geochemistry, Geophysics, Geosystems*, **5**, doi:10.1029/2003GC000662.

O'Connor, J.M., Stoffers, P., Wijbrans, J.R., Shannon, P.M. & Morrissey, T. 2000. Evidence from episodic seamount volcanism from pulsing of the Iceland Plume in the past 70 Myr. *Nature*, **408**, 954–958.

Olafsson, I., Sundvor, E., Eldholm, O. & Grue, K. 1992. Møre Margin: crustal structures from analysis of expanded spread profiles. *Marine Geophysical Researches*, **14**, 137–162.

Olesen, O., Ebbing, J., Lundin, E., *et al.* 2007. An improved tectonic model for the Eocene opening of the Norwegian–Greenland Sea: use of modern magnetic data. *Marine and Petroleum Geology* **24**, 53–66.

Osmundsen, P.T., Sommeruga, A., Skilbrei, J.R. & Olesen, O. 2002. Deep structure of the Mid Norway rifted margin. *Norwegian Journal of Geology*, **82**, 205–224.

Raum, T., Mjelde, R. & Berge, A.M. *et al.* 1995. Sub-basalt structures east of the Faroe Islands revealed from wide-angle seismic and gravity data. *Petroleum Geoscience*, **11**, 291–308.

Raum, T., Mjelde, R., Shimamura, H., *et al.* 2007. Tectono-magmatic evolution of the southern Vøring Basin and Vøring Transform Margin, NE Atlantic. *Tectonophysics* **415**, 167–202.

Sandwell, D.T. & Smith, W.H.F. 1997. Marine gravity anomaly from Geosat and ERS 1 satellite altimetry. *Journal of Geophysical Research*, **102**, 10039–10054.

Saunders, A.D., Fitton, J.G., Kerr, A.C., Norry, M.J. & Kent, R.W. 1997. The North Atlantic Igneous Province. *In:* Mahoney, J.J. & Coffin, M.F. (eds) *Large Igneous Provinces*. Geophysical Monographs, American Geophysical Union, **100**, 45–93.

Schilling, J.-G. 1973. Iceland mantle plume. *Nature*, **246**, 141–143.

Schilling, J.-G. 1999. Dispersion of the Jan Mayen and Iceland mantle plumes in the Arctic: A He–Pb–Nd–Sr isotope tracer study of basalts from the Kolbeinsey, Mohns and Knipovich Ridges. *Journal of Geophysical Research*, **104**, 10543–10569.

Shen, Y., Salomon, S.C. & Bjarnason, I.T. *et al.* 2002. Seismic evidence for a tilted mantle plume and north–south mantle flow beneath Iceland. *Earth and Planetary Science Letters*, **197**, 261–272.

Sigmundsson, F. 2006. *Iceland Geodynamics: Crustal Deformation and Divergent Plate Tectonics*. Springer, Berlin.

Skogseid, J., Planke, S., Faleide, J.I., Pedersen, T., Eldholm, O. & Neverdal,

F. 2000. NE Atlantic continental rifting and volcanic margin formation. *In:* Nøttvedt, A. (ed.) *Dynamics of the Norwegian Margin*. Geological Society, London, Special Publications, **167**, 295–326.

Smallwood, J.R. & White, R.S. 2002. Ridge–plume interaction in the North Atlantic and its influence on continental breakup and seafloor spreading. *In:* Jolley, D.W. & Bell, B.R. (eds) *The North Atlantic Igneous Province: Stratigraphy, Tectonic, Volcanic and Magmatic Processes*. Geological Society, London, Special Publications, **197**, 15–37.

Storey, M., Duncan, R.A., Pedersen, A.K., Larsen, L.M. & Larsen, H.C. 1998. $^{40}Ar/^{39}Ar$ geochronology of the West Greenland Tertiary volcanic province. *Earth and Planetary Science Letters*, **160**, 569–586.

Talwani, M. & Eldholm, O. 1977. Evolution of Norwegian–Greenland Sea: Recent results and outstanding problems. *Geological Society of America Bulletin*, **88**, 969–999.

Taylor, B., Goodliffe, A.M. & Martinez, F. 1999. How continents break up: Insights from Papua New Guinea. *Journal of Geophysical Research*, **104**, 7497–7512.

Torsvik, T.H. & Cocks, L.R.M. 2005. Norway in space and time: A Centennial cavalcade. *Norwegian Journal of Geology*, **85**, 73–86.

Torsvik, T.H., Van Der Voo, R., Meert, J.G., Mosar, J. & Walderhaug, H.J. 2001. Reconstructions of the continents around the North Atlantic at about the 60th parallel. *Earth and Planetary Science Letters*, **187**, 55–69.

Trønnes, R.G., Planke, S., Sundvoll, B. & Imsland, P. 1999. Recent volcanic rocks from Jan Mayen: Low-degree melt fractions of enriched northeast Atlantic mantle. *Journal of Geophysical Research*, **104**, 7153–7168.

Tryggvason, E. 1964. Arrival times of P waves and upper mantle structure. *Bulletin of the Seismological Society of America*, **54**, 727–736.

Tsikalas, F., Eldholm, O. & Faleide, J.I. 2002. Early Eocene sea floor spreading and continent–ocean boundary between Jan Mayen and Senja fracture zones in the Norwegian–Greenland Sea. *Marine Geophysical Researches*, **23**, 247–270.

Van Wijk, J.W., Huismans, R.S., Voorde, M.T. & Cloetingh, S.A.P.L. 2001. Melt generation at volcanic continental margins: no need for a mantle plume? *Geophysical Research Letters*, **28**, 3995–3998.

Verhoef, J., Roest, W.R., Macnab, R. & Arkani-Hamed, J. 1996. *Magnetic anomalies of the Arctic and North Atlantic oceans and adjacent areas*. Geological Survey of Canada, Technical Report.

Vogt, P.R. 1971. Asthenospheric motion recorded by the ocean floor south of Iceland. *Earth and Planetary Science Letters*, **13**, 153–160.

Vogt, P.G., Johnson, G.L. & Kristjansson, L. 1980. Morphology and magnetic anomalies north of Iceland. *Journal of Geophysics*, **47**, 67–80.

Weigel, W., Flüh, E.R. & Miller, H. *et al.* 1995. Investigations of the East Greenland continental margin between 70° and 72°N by deep seismic sounding and gravity studies. *Marine Geophysical Researches*, **17**, 167–199.

White, R.S. 1997. Rift–plume interaction in the North Atlantic. *Philosophical Transactions of the Royal Society of London, Series A*, **355**, 319–339.

White, R.S. & McKenzie, D. 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research*, **94**, 7685–7729.

White, R.S., McKenzie, D. & O'Nions, R.K. 1992. Oceanic crustal thickness from seismic measurements and rare earth element inversions. *Journal of Geophysical Research*, **97**, 19683–19715.

Wilks, K.R. & Carter, N.L. 1990. Rheology of some continental lower crustal rocks. *Tectonophysics*, **182**, 57–77.

Wolfe, C.-J., Bjarnason, I.T., Vandecar, J.C. & Salomon, S.C. 1997. Seismic structure of the Iceland mantle plume. *Nature*, **385**, 245–247.

Zhao, D. 2004. Global tomographic images of mantle plumes and subducting slabs: insight into deep Earth dynamics. *Physics of the Earth and Planetary Interiors*, **146**, 3–34.

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