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Rift–plume interaction in the North Atlantic

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Rift-plume interaction in the North Atlantic

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The style of oceanic crustal formation in the North Atlantic is controlled by interaction between the Iceland mantle plume and the lithospheric spreading. There are three main tectonic regimes comprising: (a) oceanic crust formed without fracture zones, with spreading directions varying from orthogonal up to 30° oblique to the ridge axis; (b) oceanic crust with a normal slow-spreading pattern of orthogonal spreading segments separated by fracture zones; and (c) 20–35 km thick crust generated directly above the centre of the mantle plume along the Greenland–Iceland–Færoe Ridge. I show that the main control on the tectonic style is the temperature of the mantle beneath the spreading axis. A mantle temperature increase of as little as 50 °C causes an increase of about 30% in the crustal thickness, and thereby allows the mantle beneath the crust at the ridge axis to remain sufficiently hot that it responds to axial extension in a ductile rather than a brittle fashion. This generates crust without fracture zones and with an axial high rather than a median valley at the spreading centre. Using gravity, magnetic, bathymetric and seismic refraction data I discuss the mantle plume temperatures and flow patterns beneath the North Atlantic since the time of continental breakup, and the response of the crustal generation processes to these mantle temperature variations.

1. Introduction

The volume of melt generated at a mid-ocean ridge spreading centre is highly sensitive to the temperature of the underlying mantle. An increase of as little as 50 °C, just a few per cent of the normal asthenospheric potential temperature of about 1320 °C, causes a 50% increase in the volume of melt that is generated by decompression of the upwelling mantle (McKenzie & Bickle 1988; White *et al.* 1992). Those oceanic spreading ridges that lie above a region of abnormally hot mantle caused by a mantle plume are therefore affected significantly by the increased mantle temperatures.

Since lithospheric plates move relatively fast (typically 20–150 mm yr^{-1}) with respect to the underlying mantle plumes, there are many instances where spreading centres have moved across mantle plumes and have interacted with them. Directly above the rising core of a mantle plume, the result of the interaction is usually the generation of a ridge of igneous crust, often reaching 20–30 km or more in thickness; examples include the Rio-Grande Rise and the Walvis Ridge created by the interaction between the mantle plume now beneath Tristan da Cunha and the South Atlantic spreading centre; and the Chagos–Laccadive Ridge built when the Central Indian Ridge spreading centre lay above the Réunion mantle plume. Since the thermal anomalies in the asthenospheric mantle created by plumes often extend for

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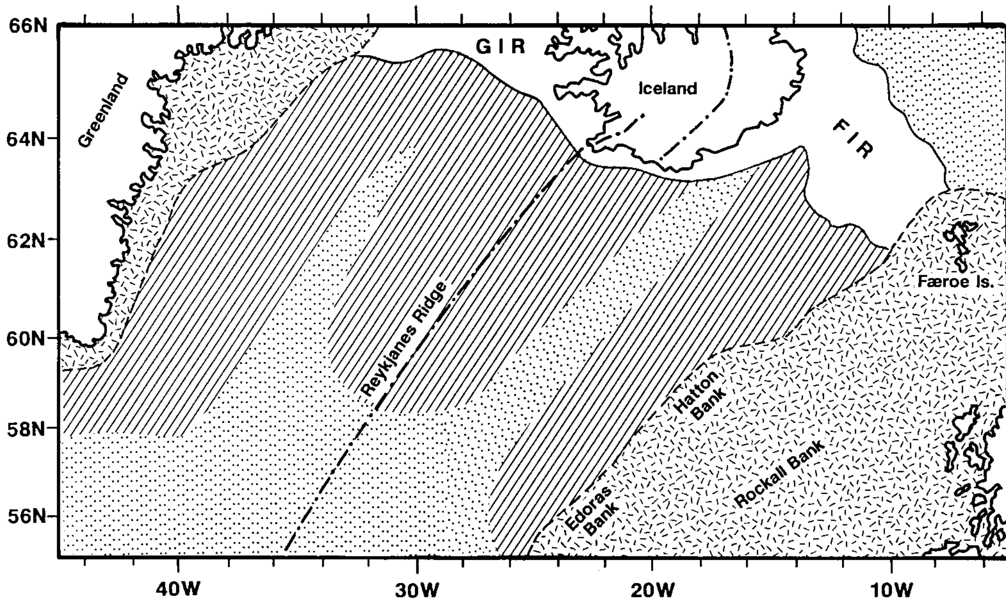

Figure 1. Outline of the main tectonomagmatic areas discussed in this paper. Shaded areas in northwest and southeast corners of the map represent stretched continental crust on the North American and Eurasian plates, respectively. Parallel shading shows oceanic crust devoid of fracture zones and dotted shading is oceanic crust with fracture zones, spreading orthogonally. Blank area is thick igneous crust above the Greenland–Iceland Ridge (GIR) and Færoe–Iceland Ridge (FIR).

distances of the order of 1000 km from the central core, the portions of the spreading axes intersecting these distal regions may also be affected. Typically, however, the temperatures in these distal regions are much lower than in the core of the plume, and the consequences are hard to detect, especially over old crust.

In this paper I discuss the interaction of the Reykjanes Ridge spreading centre in the northern North Atlantic with the Iceland mantle plume. This is a particularly good case study, because the spreading centre crosses directly above the mantle plume and is currently creating oceanic crust on the Reykjanes Ridge with relatively smooth topography and few fracture zones. This means that small fluctuations in the mantle temperature of only a few tens of °C and minor flow rate variations create marked perturbations in the gravity and topographic signatures that can be readily identified and mapped (White *et al.* 1995).

In the following sections I discuss the evidence from geophysical and geochemical data for the spatial and temporal variations in mantle temperature beneath the northern North Atlantic and discuss the effect of these on the crust generated at the Reykjanes Ridge spreading centre.

2. Tectonomagmatic regimes

Interaction between seafloor spreading and the Iceland mantle plume during the Tertiary has produced three distinct tectonomagmatic regimes, which I discuss below in detail. For this discussion, I restrict consideration to the region of the North Atlantic, south of, and including, Iceland. The area north of Iceland has suffered major ridge jumps in the seafloor spreading centre, leaving now extinct spreading centres, such as the Aegir Ridge, and continental fragments such as Jan Mayen. So the tectonics north of Iceland are complicated considerably by these features, obscuring the effects of ridge-plume interaction; to the south of Iceland no major ridge jumps have occurred.

The three main tectonic divisions of the oceanic crust (figure 1) are delineated by the gravity, magnetic and bathymetric fields (figures 2–4). They are distinguished by a first type exhibiting seafloor spreading magnetic anomalies largely unbroken by fracture zone offsets; a second type exhibiting normal (for slow-spreading ridges) ridge-fracture zone geometry with orthogonal spreading; and a third type with over-thickened, initially subaerial crust now found along the Greenland–Iceland–Færoe Ridge.

(*a*) *Crust unbroken by fracture zones*

This type of crust is found in two main areas: in the oldest seafloor generated in the early stages of opening of the North Atlantic; and along the northern two-thirds of the young oceanic crust on the present Reykjanes Ridge spreading axis (diagonal shading on figure 1). The distribution is not entirely symmetric in the North Atlantic. Over the crust formed immediately following continental breakup, it extends more than 1300 km from the plume centre, along the entire northern North Atlantic, whereas on the present spreading axis it extends only 1000 km from the plume centre beneath Iceland (figure 1).

A consistent characteristic of the oceanic crust generated without fracture zones is that it is thicker than normal, reaching 10–11 km thick, compared to 6–7 km for normal oceanic crust (White *et al.* 1992). This is indicative of mantle potential temperatures that were hotter than normal when the crust was generated. A factor which does not appear to correlate with the absence of fracture zones is the obliquity between the spreading direction and the normal to the strike of the spreading axis: over the oldest crust, the spreading direction immediately after continental breakup was normal to the ridge axis, which is the usual configuration for oceanic spreading centres. However, the youngest crust formed in this tectonomagmatic regime exhibits spreading directions up to 30° oblique to the normal from the ridge axis. So the obliquity of spreading does not appear to be a factor in controlling the formation of fracture zones.

On the present spreading axis, the young crust has little sediment cover and V-shaped ridges are prominent in the gravity field (figure 2): these cut across the isochrons marked by seafloor spreading magnetic anomalies (Vogt 1971; White *et al.* 1995), and are themselves split by continued spreading at the ridge axis (Keeton *et al.* 1997). The V-shaped ridges are thought to be caused by relatively small fluctuations in the temperature and flow rate of the mantle plume (White *et al.* 1995). Fine-scale structure is also present in the form of axial volcanic ridges, visible on the youngest, unsedimented crust (Laughton *et al.* 1979; Murton & Parson 1993; Keeton *et al.* 1997).

Over the oldest oceanic crust there are indications in the gravity field (figure 2) of lineations of alternating high and low gravity similar to those that mark the V-shaped ridges on the spreading axis. However, the sediment cover over the oldest crust means that no bathymetric ridges are now visible, and the magnitude of the gravity variations is greatly attenuated. They are therefore not nearly as prominent as are the V-shaped ridges near the present axis, but it is likely that they have a similar cause.

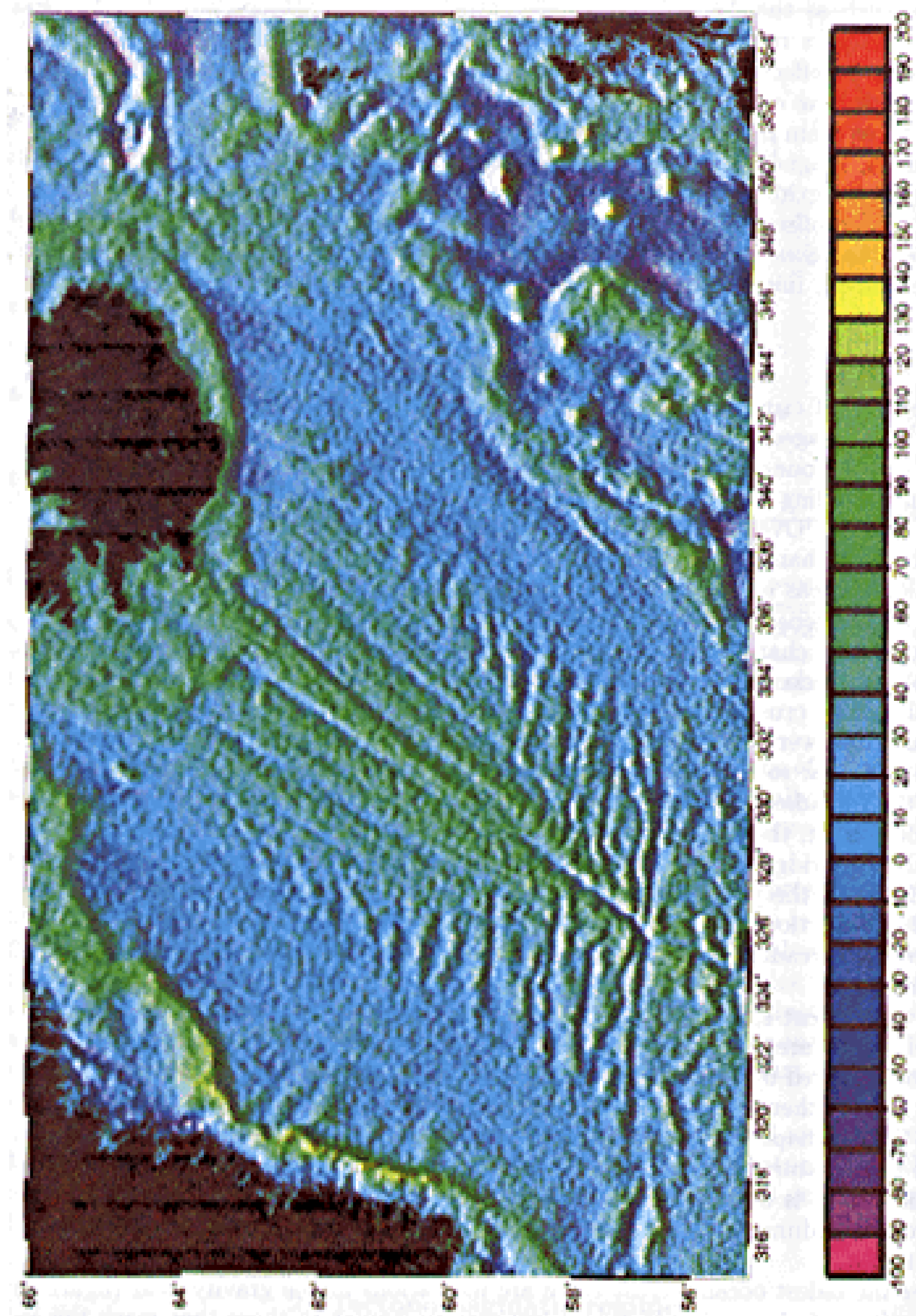

Figure 2. Free-air gravity field derived from Geosat and ERS-1 satellite geoid data (Sandwell & Smith, 1995; version 7.2). Gravity field is illuminated from the north.

(b) *Orthogonal spreading crust with fracture zones*

The second tectonomagmatic type is where normal ridge segments spread in a direction close to orthogonal, with ridge segments separated by fracture zones (dotted ornament on figure 1). This is the normal pattern at slow-spreading ridges observed

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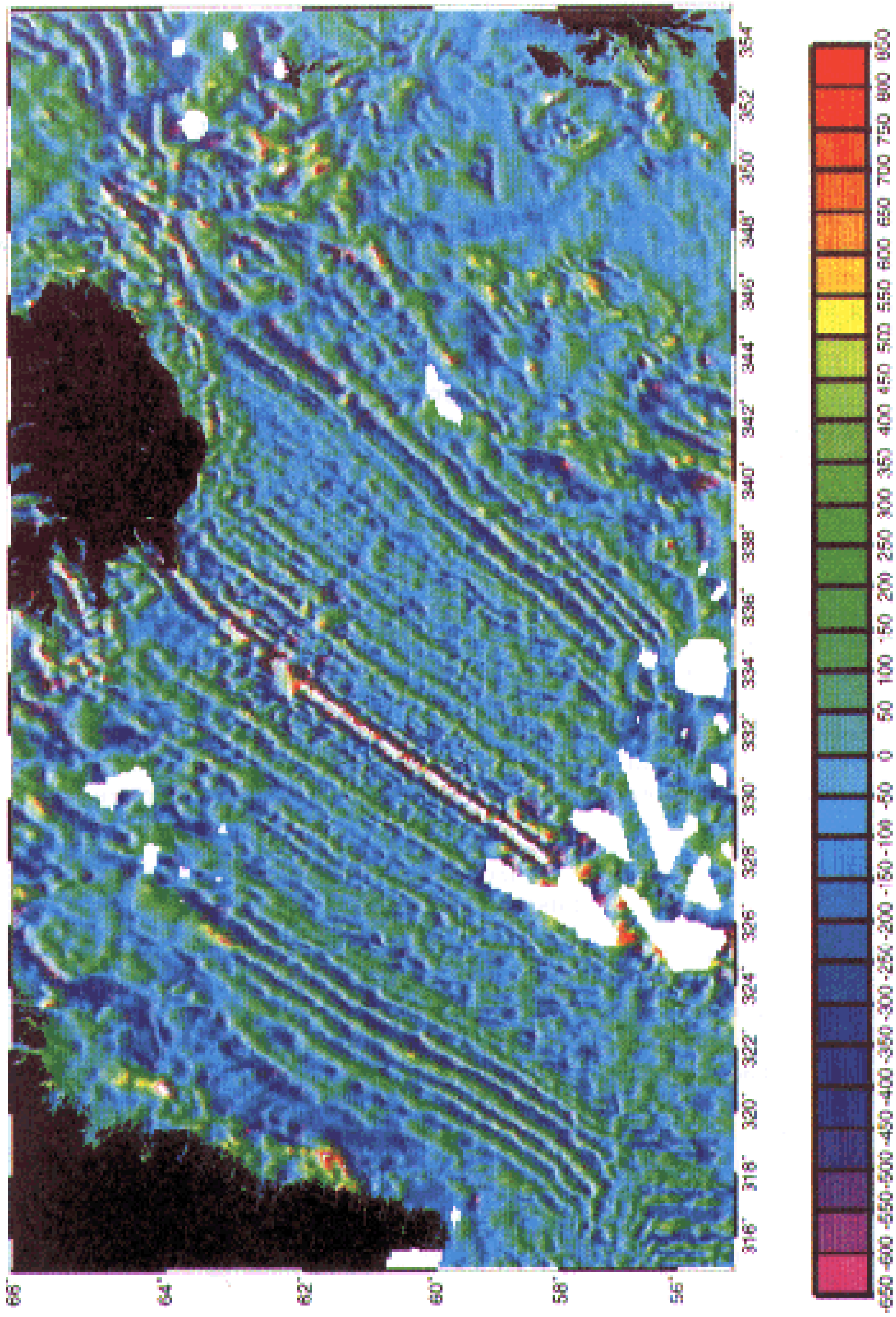

Figure 3. Magnetic anomalies derived from a compilation by Macnab *et al.* (1995). Image is illuminated from the west.

elsewhere in the ocean basins. In the region south of Iceland the change from spreading without fracture zones, to short orthogonal spreading segments terminated by fracture zones, occurs abruptly at magnetic anomaly 19 (42 Ma) on both sides of the ridge axis; this was recognized by Vogt (1971) from sparse ships' tracks, but is confirmed by the modern denser datasets, both of magnetic anomalies (figure 3) and of gravity anomalies (figure 2). The reversion to spreading without fracture zones is time-transgressive, and did not extend as far south as it did on the oldest oceanic crust. There was only a short interval of normal, fracture-zone dominated ridge spreading in the northern part of the area, near Iceland, but a much longer-lived period further south. Indeed, in the southernmost area we are considering here, south of 58° N, the spreading at the present day is still dominated by fracture zones (figure 2). The present day change on the ridge axis, from oblique spreading without fracture zones to orthogonal spreading with fracture zones, occurs about 1000 km from the centre of the Iceland plume at about 58° N, together with a change from a spreading axis dominated by topographic highs and axial volcanic ridges to one marked by a median valley (Keeton *et al.* 1997).

(*c*) *Over-thickened oceanic crust*

The third tectonomagmatic type is the crust formed above the centre of the Iceland mantle plume, which has produced abnormally thick (20–35 km) crust, whose surface was originally subaerial. This type of crust forms the present island of Iceland and the immediately surrounding shallow seafloor (figure 4), together with the bathymetric ridges known as the Greenland–Iceland Ridge and the Færoe–Iceland Ridge (marked GIR and FIR, respectively, on figure 1). Unlike the other tectonic settings, the spreading axis in this regime has suffered multiple ridge jumps: these may be caused by the spreading axis jumping to remain above the hottest upwelling region, as that region migrates with respect to the main North Atlantic spreading axis. At the present day it gives rise to a 150 km offset of the neovolcanic zone in Iceland, joined to the main North Atlantic spreading axis of the Reykjanes Ridge to the south by the South Iceland Transfer Zone, and to the Mohns Ridge in the north by a similar offset with the opposite sense of motion in the Tjörnes Fracture Zone.

3. Mantle temperatures derived from residual basement heights and crustal thickness

The three tectonomagmatic regimes reflect the effect of interaction between rifting and the underlying mantle of different temperatures and flow patterns. For the oceanic crust generated in the North Atlantic (i.e. for the first two tectonic regimes discussed above), I assume that the crust is generated by decompression melting of mantle welling up passively beneath the spreading axis. If this is the case, then the thickness of crust generated is related in a simple manner to the temperature of the mantle (McKenzie & Bickle 1988; White *et al.* 1992): a mantle temperature increase of 50 °C above normal increases the melt thickness and hence the crustal thickness by nearly 50% from 7.0–10.2 km (Bown & White 1994). Yet 50 °C is only a small perturbation on the normal mantle potential temperature of about 1320 °C, and is considerably less than the thermal anomalies of 200–250 °C found in the cores of mantle plumes (Watson & McKenzie 1991; White & McKenzie 1995).

The oceanic crustal thickness, and hence the temperature of the mantle at the time of crustal formation, can be measured directly by wide-angle seismic methods, assuming that the Moho marks the base of the igneous crust. There is now a considerable number of wide-angle seismic experiments in the North Atlantic (table 1), the majority of which have been interpreted using ray-tracing or synthetic seismogram methods. However, most of them are located either above the young oceanic crust on the present spreading axis, or over the oldest oceanic crust close to the continental margin. Few experiments have been done over the intermediate age crust where the tectonic regime reverted for a period to fracture-zone dominated seafloor spreading: an early experiment by Whitmarsh (1971) is the only one I report here.

However, an alternative way of inferring the oceanic crustal thickness is to measure the residual height of the basement. This is the difference between the present day water-loaded basement depth, after backstripping the sediment cover, and the depth it would be expected to have if it had followed a normal oceanic subsidence curve such as that reported by Parsons & Sclater (1977). Provided the present oceanic lithosphere is not dynamically supported by anomalously hot mantle, then the residual heights can be explained solely by the effects of isostacy operating on crust of variable thickness: normal thickness crust would produce a zero residual height anomaly, while crust that was thicker than normal would produce a positive residual height anomaly. In the next two sections I discuss the residual depth anomalies along representative isochrons and flowlines across the North Atlantic.

(*a*) *Residual heights along isochron profiles*

As is apparent from figure 5, which shows residual basement heights along a zero age isochron (solid line) and a 48 Ma isochron (magnetic anomaly 21) south of Iceland, both profiles show similar patterns: above the centre of the mantle plume (0 km on the horizontal distance scale), the crust is 4–5 km higher than normal spreading axes; there is an abrupt decrease in residual height over the next 100–200 km and then a more gentle drop over the next 1000 km; and normal seafloor depths are not reached until more than 1300 km from the plume centre. The similarity of the 0 Ma and 48 Ma curves indicates that the pattern of mantle temperature variation south of Iceland is similar today to what it was shortly after seafloor spreading started.

However, in detail we can draw out differences between the two profiles. At the Icelandic end of the profiles, the backstripped residual height of the Færoe–Iceland Ridge on the 48 Ma profile is broader, flatter-topped and more elevated than that of present day Iceland on the 0 Ma profile. For at least the first 20 Ma of its history, this portion of the Færoe–Iceland Ridge was subaerial and considerable erosion of the uppermost section has occurred. The volcanic edifices found directly above mantle plumes are generally unstable and are eroded rapidly: by analogy it is probable that the 1 km high volcanic 'relief' on the present day Icelandic profile will in due course be eroded to leave a flatter top like that of the Færoe–Iceland Ridge. So the true difference between the present day (solid line, figure 5) and the oldest oceanic (broken line, figure 5) profiles is over 1 km in height, with the present day profile being less high. This is consistent with the smaller present day Icelandic neovolcanic zone crustal thickness of 20–24 km (Bjarnason *et al.* 1993; White *et al.* 1966; Staples *et al.* 1997), compared to the Færoe–Iceland Ridge crustal thickness of about 30 km (Bott & Gunarsson 1980), or the oldest northeastern Icelandic crustal thickness of 35 km (White *et al.* 1996; Staples *et al.* 1997).

This difference in residual height between the youngest and oldest oceanic crust would be enhanced further if the dynamic support due to the abnormally hot mantle beneath the present axis were also to be taken into account. As much as half of the residual height along the present spreading axis can be attributed to dynamic support by hot underlying mantle rather than by crustal thickening (White *et al.* 1995). There is probably less dynamic support of the oldest crust, which lies further from the plume centre, so correction for the effect of dynamic support would decrease

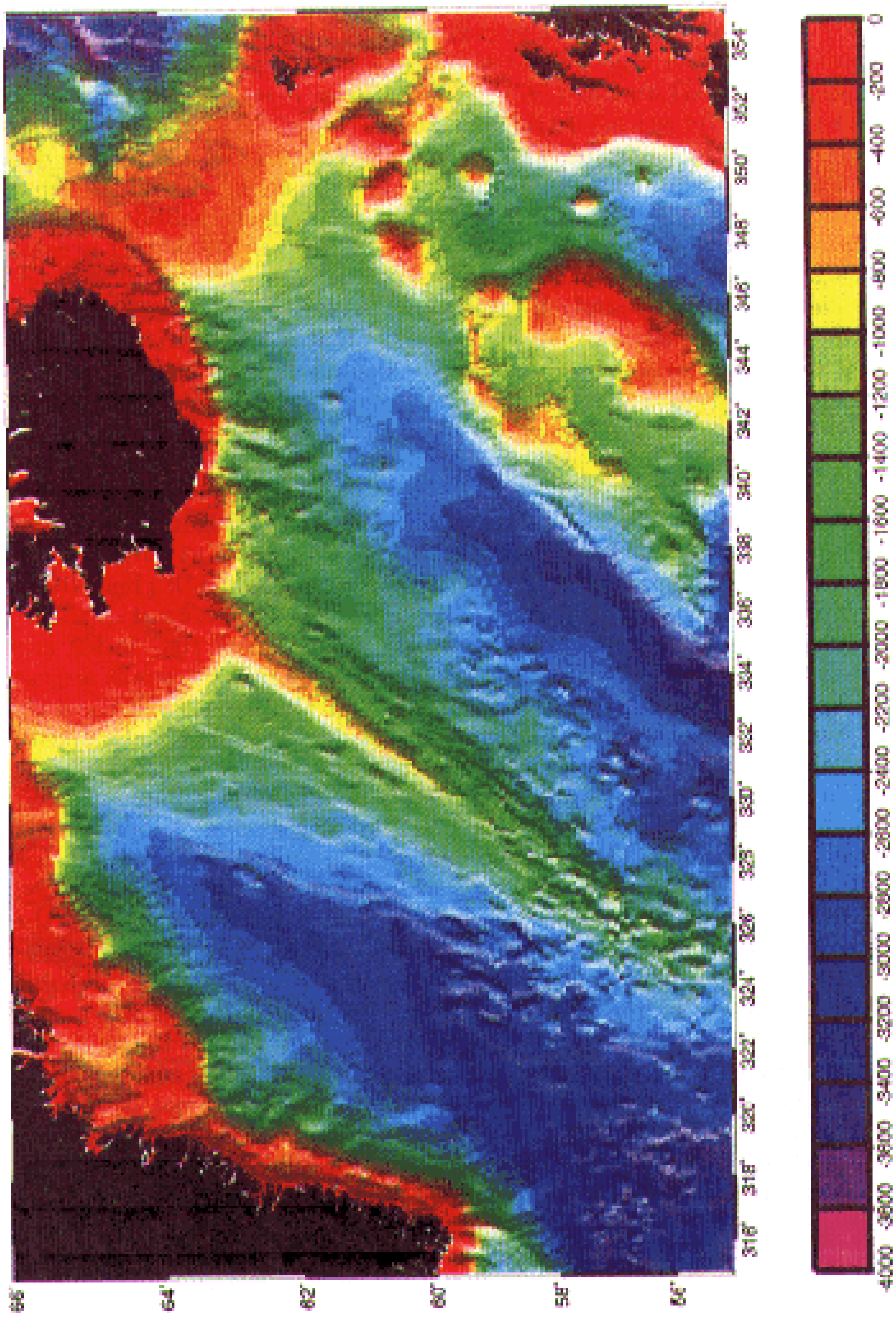

Figure 4. Bathymetry derived from five minute grid (National Geophysical Data Center 1993). Image is illuminated from the west.

the residual height of the young crust more than that of the old crust: the enhanced difference in residual heights would indicate that mantle temperatures were hotter during the earliest phases of seafloor spreading than at the present day.

The last feature that is apparent from the isochron profiles in figure 5 is that

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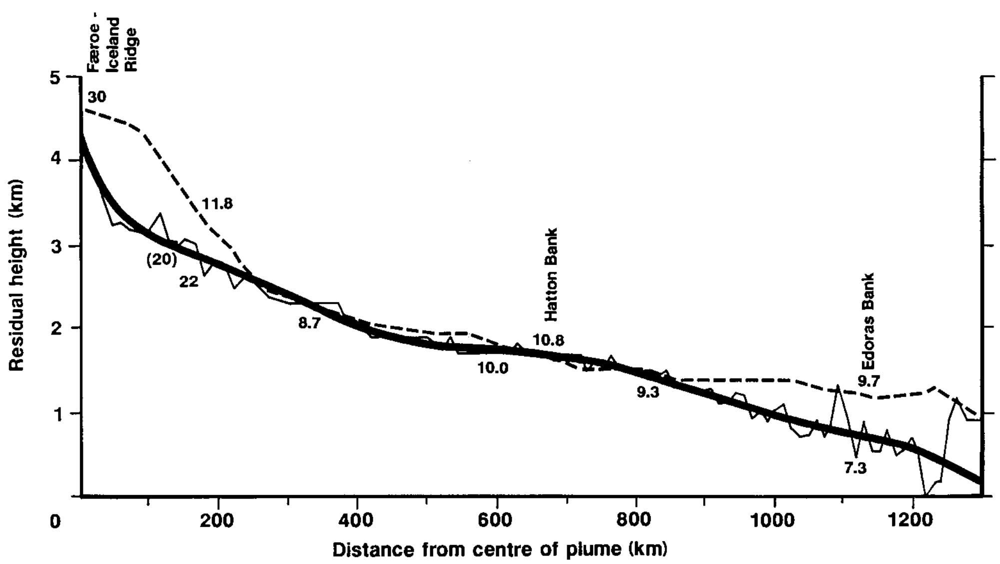

Figure 5. Residual height (i.e. height above 2.5 km below sea-level, the normal water depth at mid-ocean ridge spreading centres) along two isochrons, one along the spreading axis (zero age) and one along seafloor spreading magnetic anomaly 21 (48 Ma) on the European plate. For zero-age crust the fine line shows the actual seafloor depth and the heavier line shows the smoothed average depth. For 48 Ma crust (broken line) the average sediment thickness along the profile has been backstripped assuming Airy isostacy and LeDouran & Parson's (1982) density relationship. Parson & Sclater's (1977) curve was used to remove the effect of the increase of seafloor depth with age due to lithospheric cooling. Crustal thicknesses in km from seismic refraction experiments are shown in parentheses at appropriate distances along the curves: those above the curves refer to the oldest oceanic crust found adjacent to the rifted continental margin, and those below the curves to the zero-age crust on the spreading axis.

at the greatest distances from the plume centre, the residual height of the oldest oceanic crust (broken line) remains greater than that of the spreading axis (solid line), suggesting that mantle temperatures remained abnormally high on the oldest oceanic crust even at distances in excess of 1000 km from the plume centre. Direct measurements of the crustal thickness from wide-angle seismics tell a similar story: the oldest oceanic crust off Edoras Bank is 9.7 km thick (Barton & White 1995), while that found along a flowline on the axis is only 7.3 km thick (Sinha *et al.*, this volume). As I shall subsequently show, this is consistent with a sheet-like pattern to the early thermal anomaly during and immediately following continental breakup.

(*b*) *Residual heights along flowlines*

By constructing residual height profiles of the oceanic basement along flowlines, it is possible to gain some indication of the variation of mantle temperature through time. In figure 6 I illustrate two such flowline profiles, using Srivastava & Tapscott's (1986) poles of rotation. At either end of each flowline there are good crustal thickness determinations from wide-angle seismic experiments. Although the profiles are generated from the regional bathymetry dataset ETOPO-5 (National Geophysical Data Center 1993) and from the average sediment thickness (Ruddiman 1972), and so are not reliable in detail, they do show the general features well. The residual heights along the northernmost profile, about 600 km from the plume centre, remain about 1 km more elevated than along the southern profile which is about 1000 km from the plume centre: the crustal thicknesses (small numbers on figure 6) are also consis-

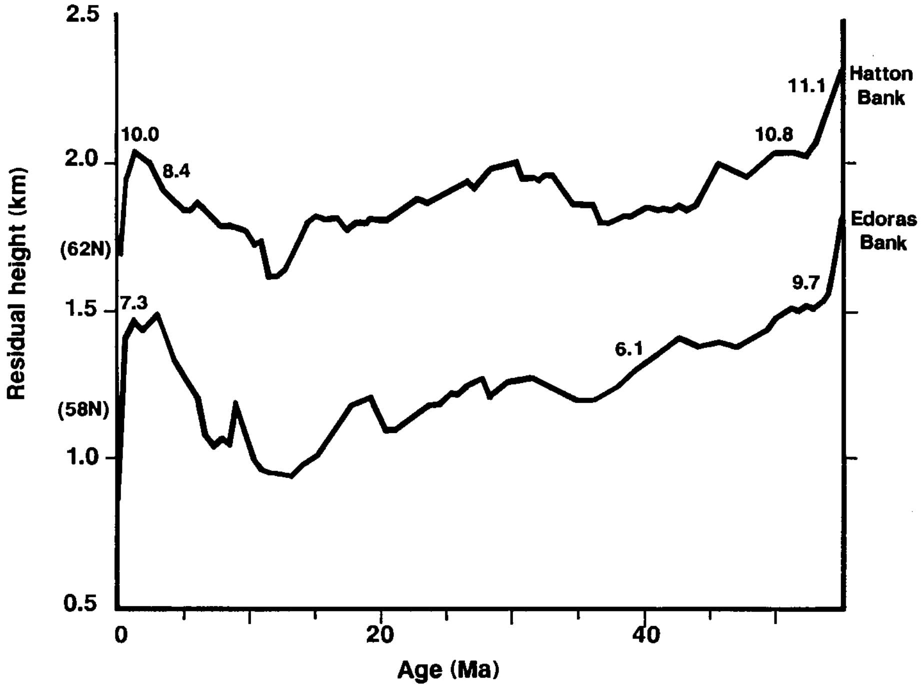

Figure 6. Residual height (i.e. height above 2.5 km below sea-level, the normal water depth at mid-ocean ridge spreading centres), of the top of oceanic basement along two flowlines: one from the spreading axis at 61.7° N to the continental margin at Hatton Bank through the locations of seismic profiles reported by Smallwood *et al.* (1995), Fowler *et al.* (1989) and Morgan *et al.* (1989); and the other from the axis at 57.7° N to the margin at Edoras Bank through seismic profiles reported by Sinha *et al.* (this volume) and Barton & White (1995, 1997). Average sediment thickness variation with age from Ruddiman (1972) was backstripped assuming Airy isostacy and Le Douran & Parson's (1983) density relationship, and increase of seafloor depth with age due to lithospheric cooling was removed using Parsons & Sclater's (1977) curve. Crustal thicknesses in km determined from seismic refraction experiments are shown at appropriate points on the curves.

tently higher beneath the northern profile. This shows that the mantle temperature has consistently remained hotter nearer the plume centre, as would be expected.

Another significant observation is that there was a gradual decrease in residual height following continental breakup and the formation of the oldest oceanic crust at about 54 Ma, with an apparent increase over the past 5 Ma: this, too, is in agreement with crustal thicknesses measured from seismics. If the effect of dynamic support from underlying hot mantle were also to be removed, then the overall decrease from 54 Ma to 5 Ma would be still more marked.

(*c*) *Mantle temperatures from crustal thickness*

We can infer the mantle potential temperature from the crustal thickness determined seismically assuming dry, isentropic melting by passive mantle decompression (McKenzie & Bickle 1988; Bown & White 1994). In figure 7 I show the mantle temperatures determined in this way along three isochrons through the northern North Atlantic. The dotted line marked 'continental margin' (Barton & White 1997) is not the subject of this paper, but is shown for comparison as the temperature distribution during continental breakup immediately preceding generation of the oldest oceanic

Table 1. *North Atlantic oceanic crustal thickness*

location	profile	D^{a} (km)	age (Ma)	crust (km)	$T_{\mathrm{p}}^{\mathrm{b}}$ (°C)	reference
Lofoten	OBS49	−1160	53	7.0	1312	Kodaira *et al.* (1995)
Lofoten	OBS18	−1085	53	10.1	1360	Mjelde *et al.* (1992)
Lofoten	OBS28	−1080	53	8.7	1340	Goldschmidt–Rokita *et al.* (1994)
Lofoten	OBS46	−1080	52	8.2	1332	Kodaira *et al.* (1995)
Lofoten	OBS27	−1080	50	7.3	1317	Goldschmidt-Rokita *et al.* (1994)
Lofoten	OBS19	−1050	53	11.3	1377	Mjelde *et al.* (1992)
Lofoten	OBS44	−1030	52	9.3	1349	Kodaira *et al.* (1995)
Lofoten	OBS20	−1020	53	11.8	1384	Mjelde *et al.* (1992)
Lofoten	OBS21	−990	53	12.3	1391	Mjelde *et al.* (1992)
Lofoten	OBS41	−960	53	7.4	1319	Mutter & Zehnder (1988)
NE Greenland	ESP19	−930	50	7.0	1312	Mutter & Zehnder (1988)
NE Greenland	ESP22	−820	50	12.9	1399	Mutter & Zehnder (1988)
Vøring margin	ESP13	−820	50	12.5	1394	Mutter & Zehnder (1988)
Kolbeinsey Ridge	OBS14/L3	−590	24	8.9	1342	Kodaira *et al.* (1997*b*)
Kolbeinsey Ridge	L2	−580	1	8–9	1329–1344	Kodaira *et al.* (1997*a*)
Møre margin	E45	−480	55	11.7	1383	Olafsson *et al.* (1992)
Møre margin	E46	−480	54	10.4	1365	Olafsson *et al.* (1992)
NE Greenland	82–27	−260	19	9.6	1353	Larsen & Jakobsdóttir (1988)
Færoe–Iceland Ridge	OBS43	−180	43	9.8	1356	Makris *et al.* (1995)

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Table 1. *Cont.*

location	profile	D^{a} (km)	age (Ma)	crust (km)	T_{p}^{b} (°C)	reference
Iceland	FIRE	−120	0	20	1483	Staples *et al.* (1997)
Færoe–Iceland Ridge	NASP	0	50	30	1580	Bott & Gunnarsson (1980)
Færoe–Iceland Ridge	OBS10	55	52	13.7	1409	Makris *et al.* (1995)[c]
SE Greenland	81–20	70	50	11.9	1385	Larsen & Jakobsdóttir (1988)[d]
Færoe–Iceland Ridge	OBS01	90	52	11.8	1384	Makris *et al.* (1995)[c]
Iceland	SIST	150	0–3	20–24	1483–1525	Bjarnason *et al.* (1993)
SE Greenland	82–01	240	52	13.9	1412	Larsen & Jakobsdóttir (1988)[d]
Reykjanes Ridge	BI01	270	12	9.3	1348	Ritzert & Jacoby (1985)
Reykjanes Ridge	BI02	400	15	7.9	1327	Ritzert & Jacoby (1985)
Reykjanes Ridge	1	440	0	11.2	1376	Mochizuki (1995)
Reykjanes Ridge	CAM71	580	0	10.0	1359	Smallwood *et al.* (1995)
Reykjanes Ridge	CAM73	580	4	8.4	1335	Smallwood *et al.* (1995)
Hatton Bank	ESPG	680	53	11.2	1376	Fowler *et al.* (1989)
Hatton Bank	ESPH	680	53	11.1	1374	Spence *et al.* (1989)
Hatton Bank	OBS4	680	50	10.8	1370	Morgan *et al.* (1989)
Iceland Basin	PUBS	740	38	6.1	1297	Whitmarsh (1971)
Reykjanes Ridge	Z	810	9	9.3	1348	Bunch & Kennett (1980)
Reykjanes Ridge	Line 2	1100	0	7.3	1317	Sinha *et al.* (this volume)
Edoras Bank	OBH14	1120	53	9.7	1354	Barton & White (1995)

[a]Distance from the inferred centre of the Iceland mantle plume, taken as the highest point on Iceland for zero-age crust, and the centre of the Greenland–Iceland–Færoe Ridge for older crust. Negative values are to the north of the plume centre, positive values to the south.

[b]Potential temperature of mantle at the time of crustal generation assuming isentropic decompression of dry mantle, using values from Bown & White (1994).

[c]I assume the crust here is oceanic, though Makris *et al.* (1995) assume it is continental.

[d]Moho depth is only poorly constrained from weak signals close to, or at noise level at offsets of 15–30 km.

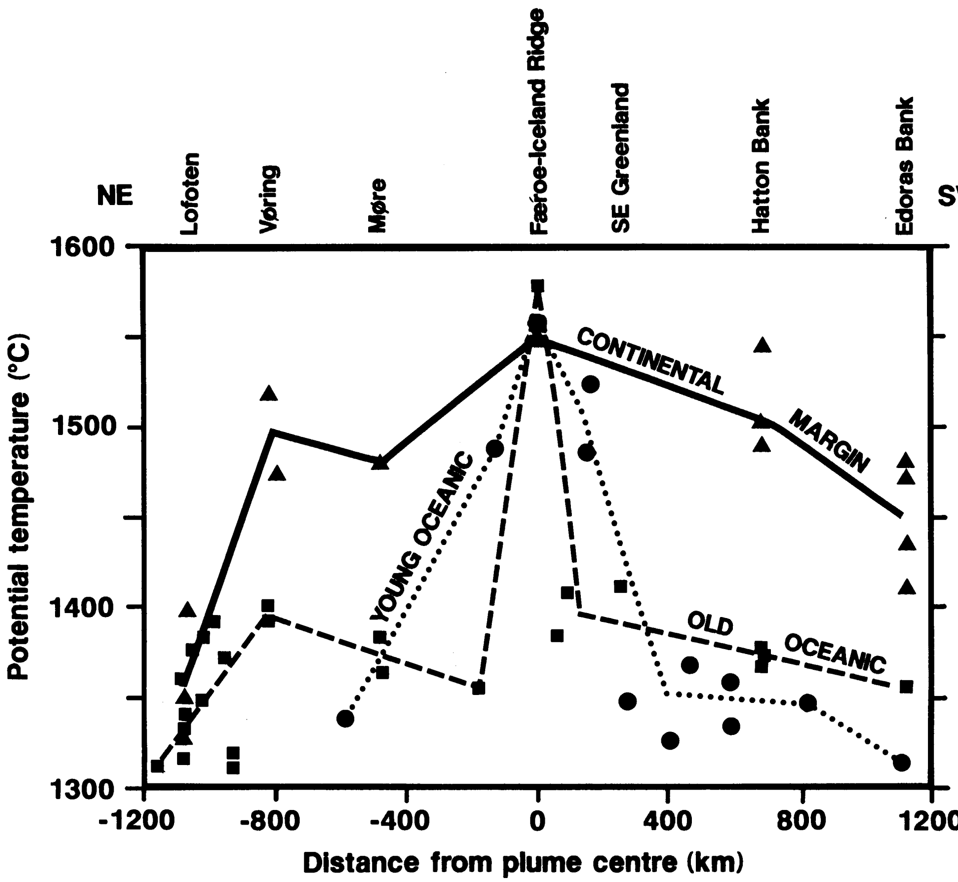

Figure 7. Inferred mantle potential temperature along three isochrons in the northern North Atlantic. 'Young oceanic' isochron is based on all crustal thickness estimates on crust aged 5 Ma or less along the North Atlantic spreading centre and the neovolcanic zones of Iceland (circles and dotted line). 'Old oceanic' isochron uses oceanic crustal thickness measurements over crust aged 50 Ma or more that exhibits clear seafloor spreading magnetic anomalies (squares and broken line). 'Continental margin' isochron, for comparison, is based on igneous crustal thickness in rifted continental margin crust generated at the time of continental breakup (triangles and solid line (Barton & White 1997)). Conversion from igneous crustal thickness to mantle temperature assumes melt generation by isentropic decompression of dry mantle using the relationships of Bown & White (1994). Sources of oceanic crustal thickness determinations are listed in table 1.

crust. It makes two bold assumptions, neither of which is likely to be completely true: these are, first, that decompression melting beneath the rifting margin was passive, without melt enhancement by active convection in the mantle; and, second, that all the melt was retained on the margin and none flowed out of the developing rift. Departure from the first assumption would lead to lower inferred mantle temperatures, while departure from the second would lead to somewhat higher inferred temperatures. For our purposes the important observation is that along a 2000 km-long portion of the North Atlantic continental margin, the melt thickness, and hence the inferred mantle temperature, did not vary significantly, and was considerably hotter than the mantle temperature when the oldest oceanic crust was formed.

The relative uniformity of the high temperatures along the entire margin suggests that the pattern of the thermal anomaly was in the shape of a rising sheet of abnormally hot mantle rather than an axisymmetric plume with a hot rising core only at its centre. Numerical and laboratory convection experiments indicate that boundary layer instabilities that develop into plumes may often originate at depth as radial sheets: these often develop subsequently into an axisymmetric system centred on the junction between the sheets (White & McKenzie 1995). It appears likely that some such system of rising sheets occurred beneath the North Atlantic prior to breakup: one sheet would have extended southward from the region of the present Færoe Islands beneath the western margin of Rockall Bank and southeast Greenland; another northward beneath the Norwegian-northeast Greenland boundary; and a third at a high angle from the Færoe Islands region to the area of Disko Island off western Greenland. This can explain the distribution of high-temperature contemporaneous igneous rocks on the eastern and western Greenland margins and along the North Atlantic margins.

As the mantle plume developed and North Atlantic opening started, the mantle temperatures dropped everywhere along the margins except beneath the Færoe–Iceland Ridge. The oldest oceanic crust (broken line, figure 7) was formed from lower temperature mantle than that which produced the outburst of magmatism on the margins. This may have been due to the mantle upwelling developing into a more axisymmetric system centred on the junction between the rising sheets beneath the Færoe–Iceland Ridge. Nevertheless, the interaction between rifting and the upwelling mantle plume may have kept the upwelling in a broadly sheet-like pattern beneath the early ocean basin. This is evidenced by the broadly constant temperatures inferred beneath the ocean basin southward away from the Færoe–Iceland Ridge and by the vestiges of gravity lineations visible in the free air gravity field over the old oceanic crust and interpreted as minor fluctuations in the temperature of the plume.

The present day temperatures under the spreading axis (solid line, figure 7) are somewhat lower than those inferred for the oldest oceanic crust. The V-shaped ridges, which propagate south from Iceland at apparent rates of 75–150 mm yr^{-1} (compared to the full spreading rate of *ca.* 20 mm yr^{-1}) represent temperature fluctuations of about 30 °C superimposed on this general pattern on a timescale of 3–5 Ma (White *et al.* 1995).

4. The evidence from geochemistry

The geochemistry of the basaltic rocks in the North Atlantic provides strong constraints on the nature and temperature of the mantle from which the melts were generated. I here highlight two features that bear on the mantle plume temperature and circulation. First, the neodymium isotopic content of Icelandic basalts indicate that they consist on average of only 10–15% primitive mantle, with a majority of depleted MORB-source mantle (Condomines *et al.* 1983; Elliot *et al.* 1991; Hemond *et al.* 1993; Meyer *et al.* 1985; O'Nions *et al.* 1977; Zindler *et al.* 1979). This is indicative that there is considerable mixing between a relatively small amount of primitive plume mantle and the surrounding depleted upper mantle.

Second, rare earth element inversions on analyses of basalts along the Greenland and European continental margins suggest that they were formed from mantle with a potential temperature of 1450–1500 °C (Brodie 1995; White & McKenzie 1995), the same as is found from analyses of neovolcanic basalts from the present day core of the plume beneath Iceland (Nicholson & Latin 1992; White & McKenzie 1995). By contrast, basalts from the DSDP holes on the Reykjanes Ridge south of Iceland indicate considerably lower parent mantle temperatures (White *et al.* 1995). All this is consistent with the pattern of mantle temperatures discussed in the previous section during continental breakup and at the present day.

Combination of the melt thicknesses inferred from rare earth element inversions with the crustal thicknesses measured by seismic techniques provides insight into whether the mantle upwelling that caused decompression melting was a passive response to the lithospheric extension, or whether it was a result of forced convection that cycled mantle through the melt region. So, for example, a global review of normal mid-ocean ridge basalts (White *et al.* 1992) showed that estimates of the melt thicknesses from rare earth element inversions are, within error, the same as measurements of melt thickness from seismic crustal thickness determinations. This indicates that mantle upwelling beneath oceanic spreading centres is predominantly a passive response to lithospheric separation. By contrast, rare earth element inversions of basalts from the Hawaiian islands (Watson & McKenzie 1991) show that considerable volumes of melt are produced by mantle being forced through the melting region beneath Hawaii by active convection in the core of the underlying mantle plume, despite there being no lithospheric extension in this mid-plate location.

In the case of Iceland, the rare earth element inversions indicate melt thicknesses of 15–20 km (White *et al.* 1995; White & McKenzie 1995), while the seismic measurements indicate crustal thicknesses of about 20 km spread across the 300 km long neovolcanic zone of Iceland. If the central rising core of the mantle plume beneath Iceland has a diameter of about 100–150 km, as is indicated by the extent of the most highly elevated region, then I conclude that the mantle convection that causes melt generation beneath Iceland is somewhat more active than a purely passive response to plate separation (i.e. there is forced convection analogous to that under, for example, Hawaii), but that the forced convection is not particularly vigorous. It is probable that melt flows laterally at crustal levels from the main locus of melt generation above the core of the plume, so that it becomes distributed along the neovolcanic zones across the width of Iceland. A simple mass calculation suggests that the bulk of the mantle that is convected to shallow levels in the plume beneath Iceland and which becomes partially depleted by melting, eventually becomes incorporated into the lithospheric plates that absorb the cooling mantle and which spread away from the neovolcanic zones in Iceland.

The V-shaped ridges mapped on the Reykjanes Ridge suggest that there is some lateral flow of asthenospheric mantle away from Iceland, but the geochemistry of the basalts on the ridge axis suggest that they have come from mantle that has not been through the melting column beneath Iceland. The relatively small temperature anomalies and high percentage of depleted MORB-source mantle found beneath the Reykjanes Ridge are consistent with the asthenospheric mantle in this region representing a sheath of only slightly hotter than normal mantle that surrounded the central plume core beneath Iceland (White *et al.* 1995).

5. Influence of mantle temperature on oceanic crustal formation

The characteristics of the three main tectonomagmatic regimes in which oceanic crust is generated in the northern North Atlantic are governed by the temperature and flow patterns of the mantle beneath the spreading centres at the time of crustal formation.

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(a) *Spreading axis unbroken by fracture zone offsets*

Oceanic crust unbroken by fracture zones was formed within the entire ocean basin immediately following continental breakup, and again, after a short reversion to a phase of fracture zone dominated crustal formation, is being formed at the present day. However, the present day crust without fracture zones is found only along the northern portion of the Reykjanes Ridge closest to the centre of the Iceland plume. The oldest crust of this type was generated when the spreading direction was orthogonal to the strike of the ridge axis, as is normal for oceanic spreading centres. The present day Reykjanes Ridge, however, is spreading at a high obliquity, of about 30°. So it is clear that the obliquity of the spreading is not the controlling factor as to whether fracture zones are formed.

The consistent characteristic of the crust formed without fracture zones is its thickness: everywhere this type of crust is found in the North Atlantic, the crust is thicker than 8 km, and is on average 10–11 km thick. Reversion to the fracture-zone dominated type of crust that is normal at low spreading rates occurs south of 57° N, where the crustal thickness decreases to a normal value of about 7 km. Away from mantle plumes oceanic crust exhibiting similar characteristics, namely an absence of fracture zones and an axial high rather than a median valley, is found predominantly on fast-spreading ridges such as the East Pacific Rise. However, on the fast-spreading ridges the oceanic crust exhibits normal thicknesses of 6–7 km (White *et al.* 1992). So it cannot be simply the crustal thickness that is the controlling factor in this tectonomagmatic regime, but rather a combination of crustal thickness and spreading rate.

The formation of median valleys (Phipps Morgan *et al.* 1987) and of fracture zones both represent the brittle response of the lithosphere at the spreading axis. If neither a median valley nor fracture zones are present, it suggests that conditions are such that the lithosphere is weak and can respond ductilely to the forces at the spreading centre. The uppermost portion of the oceanic crust, down to a depth of about 2 km below the seafloor, is quenched by hydrothermal circulation. Below the limit of hydrothermal penetration the crust remains hot from igneous intrusions and loses heat mainly by conduction (e.g. Henstock *et al.* 1993). At any given temperature the mantle beneath the crust is considerably stronger than the crust because it has a lower homologous temperature, but the cooling of the mantle is controlled by conductive cooling through the overlying crust.

The strength of the lithosphere beneath the axial region is therefore controlled by the rate of cooling of the lower crust and of the underlying mantle. At normal slow-spreading ridges, with intrusion events occurring, typically, at intervals of 10 000–50 000 years, the crust and underlying mantle near the spreading axis cools sufficiently between igneous injections to allow brittle behaviour, and the formation of a median valley, of fracture zones, and of earthquakes extending to depths of up to 8 km (Toomey *et al.* 1985; Kong *et al.* 1992). On fast-spreading ridges with injections of melt occurring at intervals an order of magnitude smaller, the crust remains hotter and weaker and earthquakes are not only rarer, but extend down to only about 3 km at the axis (Riedesel *et al.* 1982; Orcutt *et al.* 1984).

There is apparently a delicate balance between the spreading rate and the crustal cooling rate. On normal slow-spreading ridges the rate of cooling exceeds the rate of heat input from igneous injections such that faulting can extent down to the stronger mantle near the axis. On fast-spreading ridges the higher rate of igneous injection means that conductive cooling cannot cool the mantle sufficiently for it to behave brittly before the new lithosphere has moved away from the axis.

On the Reykjanes Ridge the increased crustal thickness means that even at slow spreading rates, the mantle beneath the spreading axis remains sufficiently hot to behave ductilely. There are three factors which all tend to keep the mantle hotter. First, the increased crustal thickness means that there is an increase in the frequency of melt injection, assuming that the melt volume per injection event remains the same. Second, the mantle temperature itself is somewhat hotter than it is in areas away from mantle plumes: this is only a small increase, of perhaps 50 °C, but nevertheless it means that the mantle starts off hotter so has further to cool before it becomes brittle. Third, and probably most importantly, the increased crustal thickness provides a thicker insulating layer which means that it takes longer for the underlying mantle to cool. Bell & Buck (1992), using similar arguments, suggest that the lower crust beneath the Reykjanes Ridge remains sufficiently hot to flow ductilely. Chen & Morgan (1990) similarly show that an increase in crustal thickness and in mantle temperature can account for the absence of a median valley on the Reykjanes Ridge. Direct evidence for a relatively thin brittle layer restricted to the upper crust comes from earthquake hypocentral determinations on the ridge axis immediately south of Reykjanes Peninsula which show earthquakes extending down to only 8 km beneath the seafloor, well within the crust (Mochizuki 1995).

(*b*) *Oceanic crust broken by fracture zones*

The normal slow-spreading oceanic crust with abundant fracture zones and orthogonal spreading is found in the North Atlantic in areas where the crustal thickness, and hence the underlying mantle temperature, were normal. This type of crust is found at the present day in the distal regions more than 1000 km from the plume centre, and also throughout the area in the mid-Tertiary. A rather poorly constrained crustal thickness determination over crust formed during the mid-Tertiary phase yields a crustal thickness of 6.3 km (Whitmarsh 1971), which is consistent with the idea discussed above that it is the thermal state of the lithosphere at the spreading axis which controls the tectonic style of the oceanic crust.

(*c*) *Crust created directly above the mantle plume*

The crust of the Greenland–Iceland–Færoe Ridge created directly above the core of the mantle plume is considerably thicker than normal due to the enhanced mantle temperatures and to the forced convection in the plume, which exceeds the mantle upwelling caused solely by plate separation. Its thickness varies from 20–35 km through its history, suggesting variations in the temperature or the mantle flow-rate, or both.

A particular feature of this crust formed directly above the plume centre is that it exhibits multiple jumps of the spreading axis, in a way not seen on the Reykjanes Ridge to the south. These probably occur because the centre of the plume is currently migrating eastward with respect to the centre of the North Atlantic. So the neovolcanic zone on Iceland jumps to keep above the axis of the plume: the most recent eastward jump occurred about 4 Ma ago.

6. History of mantle plume-ridge interaction

There is a close link between the history of the Iceland mantle plume and oceanic crustal formation in the North Atlantic. The continental breakup phase was apparently associated with abnormally hot mantle, probably welling up in a pattern of sheets meeting beneath the central region where Iceland now lies. This sheet-like pattern may have marked the onset of a new plume instability. The new ocean basin broke open along the weak line marking the western edge of the string of Mesozoic basins along the northwest European margin.

The first-formed oceanic crust was 10–11 km thick, reflecting high underlying mantle temperatures beneath the entire oceanic basin. It formed without fracture zones. However, at about the time that the Labrador Sea stopped opening, the mantle temperatures beneath most of the oceanic basin dropped, and oceanic crust formed for a period with a normal slow-spreading pattern of fracture zones and orthogonal spreading. At about the same time that the neovolcanic zone on the Greenland–Færoe Ridge jumped westward toward Greenland there was also renewed uplift in east Greenland. It is possible that an original upwelling sheet of hot mantle beneath the Greenland–Færoe axis developed into a more localized upwelling region under the east Greenland coastal area: this would explain both the westward jump in the neovolcanic zone and the temporary drop in mantle temperatures beneath the North Atlantic, which allowed the formation of normal slow-spreading oceanic crust in this area which lay distant from the new focus of upwelling.

Over the past 25 Ma the centre of the mantle plume has been migrating eastward and the neovolcanic zone has been jumping eastward along the Greenland–Færoe Ridge to keep up with it. Temperatures under the North Atlantic have again increased, producing thick oceanic crust without fracture zones, but mantle temperatures are somewhat cooler now than in the early phase of seafloor spreading following continental breakup. Small fluctuations in the mantle temperature and flow rate are recorded by the V-shaped ridges on the present Reykjanes Ridge, on timescales of 3–5 Ma. Such fluctuations are probably present in all mantle plumes, but can be readily detected here because the spreading axis cuts directly above the plume, and so the crustal thickness directly reflects the mantle temperature.

There is forced convection in the present mantle plume, with the upwelling rate being somewhat faster than the rate that would result from passive upwelling beneath the spreading plates. However, a simple mass balance suggests that the bulk of the mantle plume is absorbed into the thickening lithosphere beneath Iceland. It is likely that the interaction between the North Atlantic lithospheric spreading and the upwelling plume created a sheet-like pattern to the upwelling under the Reykjanes Ridge, which generates the characteristic V-shaped ridges as the asthenospheric mantle moves along the ridge axis and up toward the surface.

It is clear from the history of seafloor spreading in the North Atlantic that on slow-spreading ridges the normal pattern of crustal generation with axial valleys, fracture zones and orthogonal spreading is easily perturbed to a pattern without an axial valley and without fracture zones. There is a delicate thermal balance which controls the temperature of the lithosphere at the spreading axis, and which determines whether there is a ductile or brittle response to the lithospheric extension at the axis. A mantle temperature increase of as little as 50 °C beneath the spreading axis causes a change from a brittle, fracture-zone dominated regime to a ductile spreading regime capable of supporting highly oblique spreading and without fracture zones or an axial valley.

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