RIDGE-PLUME INTERACTION IN THE NORTH ATLANTIC AND ITS INFLUENCE ON CONTINENTAL BREAKUP AND SEAFLOOR SPREADING

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Abstract: Development of the rifted continental margins and subsequent seafloor spreading in the North Atlantic was dominated by interaction between the Iceland mantle plume and the continental and oceanic rifts. There is evidence that at breakup time a thin sheet of particularly hot asthenospheric mantle propagated beneath the lithosphere across a 2500 km diameter region. This event caused transient uplift, massive volcanism and intrusive magmatism, and a rapid transition from continental stretching to seafloor spreading. Subsequently, the initial plume instability developed to an axisymmetric shape, with the ~100 km diameter central core of the Iceland plume generating 30-40 km thick crust along the Greenland-Iceland-Faeroes Ridge. The surrounding 2000 km diameter region received the lateral outflow from the plume, causing regional elevation and the generation of thicker and shallower than normal oceanic crust. We document both long-term (10-20 Ma) and short-term (3-5 Ma) fluctuations in the temperature and/or flow rate of the mantle plume by their prominent effects on the oceanic crust formed south of Iceland. Lateral ridge jumps in the locus of rifting are frequent above the regions of hottest asthenospheric mantle, occurring in both the early history of seafloor spreading, when the mantle was particularly hot, and throughout the generation of the Greenland-Iceland-Faeroes Ridge.

The northern North Atlantic is a classic example of continental breakup above a thermal anomaly in the mantle created by a mantle plume. The most obvious consequence of this was the generation of huge volumes of basaltic magmas by partial melting of the mantle at the time of continental breakup (White & McKenzie 1989). Most of this magma bled up to the crust, where the majority (perhaps 60-80%) was frozen in the lower crust, while the rest was extruded as lava flows and pyroclastic deposits, often after undergoing fractionation in crustal magma chambers. If the extrusion rates are sufficiently high, lava flows may form extensive, sub-horizontal flood basalts extending across the continental hinterland. On rifted continental margins, the extrusive lavas are typically expressed as seaward dipping reflector (sdr) sequences that are readily identifiable on seismic reflection profiles. All these features have been found along more than 2,500 km of the rifted continental margins in the northern North Atlantic, on both the European and the Greenland sides. Furthermore, the mantle plume that gave rise to the thermal anomaly in the mantle is still going strong, lying beneath Iceland at the present day.

The northern North Atlantic is therefore an excellent “natural laboratory” in which to investigate the interplay between rifting and magmatism, because the history of rifting is well recorded from the time of continental breakup in the early Tertiary to the seafloor spreading at the present day. Magnetic anomaly field reversals during the Tertiary were frequent, and south of Iceland there are few large fracture zones in the oceanic crust, so the seafloor spreading magnetic anomalies contain a clear record of the rifting history throughout the ocean basin. There have also been several well-constrained wide-angle seismic surveys of the continental margins on both sides of the Atlantic (e.g. Barton & White 1997b; Smallwood *et al.* 1999; Korenaga *et al*. 2000), often with coincident drilling penetrating the volcanic rocks, so both the structure of the rifted margins and the timing of the igneous activity are well known.

In this paper we examine the interaction between the mantle plume and rifting, first of the continental lithosphere to form the present continental margins, and subsequently of the oceanic lithosphere to form the mid-ocean ridge and the land mass of Iceland.

Continental Breakup

There are two striking features of the magmatism accompanying continental breakup in the North Atlantic region. The first is its relatively short duration (White & McKenzie 1989; Saunders *et al.* 1997). The second is the extremely widespread and almost simultaneous onset of voluminous volcanism across a huge region (Figure 1): extending from Baffin Island and west Greenland in the west, to the northern portion of Norway in the north, and as far as the southern tips of east Greenland and the Rockall Plateau in the south. Since there is no indication of the presence of a mantle plume immediately prior to the widespread onset of volcanism, this provides strong constraints on the initiation of the mantle thermal anomaly under this region.

Throughout the area, the onset of volcanism was consistently at 62-61 Ma (Saunders *et al*. 1997). Frequently, picritic basalts were amongst the first volcanic units erupted, indicative of high temperature mantle sources: perhaps the most striking example of this is on Disko Island, west Greenland, where the most magnesium-rich extrusive rocks have been found, despite being on the periphery of the North Atlantic Igneous Province (NAIP) (Figure 1). Many of these early volcanic units show signs of limited contamination by continental lithosphere, indicative of passage of the magmas through continental lithosphere, and indeed most of them are emplaced onto unequivocal continental crust. There is a general indication of only limited continental rifting during this earliest phase of volcanism (Larsen & Saunders 1998).

Following the first, and widespread magmatic outburst, there appears to have been a hiatus in volcanism lasting up to ~3 m.y., before a second, more sustained period of volcanism during 56-53 Ma which accompanied the start of seafloor spreading in the North Atlantic (Saunders *et al.* 1997). This hiatus is documented off east Greenland (Larsen & Saunders 1998), and in the Faeroe Islands, where thin coal deposits developed between the Lower and Middle Lava formations (Larsen *et al.* 1999). Both the rate of volcanic production and the total melt generation during the second phase of activity were much larger than during the first phase, and the volcanism accompanied a rapid thinning of the continental lithosphere. It was during this second phase that the main series of sdr sequences, up to 150 km wide on both sides of the Atlantic, were generated on the margins, in addition to massive basaltic lava flows which extended into the sedimentary basins adjacent to the rifted margins (e.g. Richardson *et al*. 1999). The total igneous thickness on the rifted margins reached 25-35 km (White & Barton, 1997b; Reid *et al*. 1997), and the sdr sequences were erupted near to, or above sea level.

What can be learnt about the initial mantle thermal anomaly under the North Atlantic region from the distribution and timing of the magmatism?

There are three main indicators of mantle temperature, though each may also be affected by factors other than solely the temperature: (i) the total volume of melt produced; (ii) the uplift and subsidence history that resulted, in part, from the addition of new igneous material to the crust, and in part from dynamic support by the underlying plume (Bown and White 1995, Barton & White 1997b); and (iii) the geophysical, geochemical and petrological characteristics of the igneous rocks themselves (White & McKenzie 1995). The surprising observation from the North Atlantic region is that abnormally hot mantle arrived simultaneously, or at least within the limits of resolution of our measurements of about 1 m.y., across the entire region, at points more than 2000 km apart in the pre-rift reconstruction (Figure1) (White 1988; Saunders *et al.* 1997; Larsen & Saunders 1998; Larsen *et al.* 1999). The main phase of volcanism produced igneous sections along the rifted margins of the North Atlantic that vary little in thickness along a 2000 km length from south Rockall Plateau to northern Norway (Barton & White 1997b), and along the east coast of Greenland from its southern tip to the Greenland-Iceland-Faeroe Ridge. It is also apparent that the oceanic crust that was formed adjacent to the rifted margins within a few million years of continental breakup, everywhere except above the core of the plume (that lay beneath the Greenland-Iceland-Faeroe Ridge), reduced rapidly in thickness compared to the igneous thickness on the rifted margins. Thereafter, the distribution of crustal thickness along the Mid-Atlantic rift remained similar to that found today around the present-day mantle plume centred beneath Iceland, suggesting a broadly similar regional distribution of mantle temperatures since shortly after the onset of seafloor spreading (White *et al*. 1995).

Care must be taken when inferring the mantle temperature directly from the crustal thickness. Where the mantle decompresses passively beneath a rift, and there is no lateral flow of the melt away from the rift, then the crustal thickness provides a direct measure of the mantle temperature (White & McKenzie 1989). Apart from directly above the mantle plume under Iceland, this is the situation that we believe pertains at the present day along the northern North Atlantic spreading centres. However, where there is active convection-driven flow of the mantle through the melting region, as in the narrow (~100 km diameter) central core of the Iceland plume, then even modest temperature anomalies in the mantle can build large crustal thicknesses. The more active the flow, the smaller the temperature anomaly required to generate a given thickness of igneous crust. In present-day Iceland, the crustal thickness of 40 km directly above the mantle plume (Darbyshire *et al.* 2000), is due in part to the excess temperature of the mantle plume and in part to active convection. Geochemical characteristics of the igneous rocks can assist in determining the parent mantle temperature, and even within 150 km of the Iceland plume, geochemical- and seismic-based measurements of the crustal thickness suggest that the mantle decompresses passively beneath the rift and therefore lies outside the central rising core of the mantle plume (White *et al.* 1995; Weir *et al.* 2001).

If the mantle decompression under the nascent rifted margins was passive at the time of continental breakup, then the large igneous thicknesses found along the entire length of the rift are indicative of a thermal anomaly in the asthenospheric mantle of ~150°C (Clift *et al.* 1995). The decrease in crustal thickness after the onset of seafloor spreading would then indicate a rapid drop in the mantle thermal anomaly to ~50°C in the distal areas, though it remained high directly over the plume core (Barton & White 1997b) . These inferred temperatures are probably upper limits: if there was a component of active convection under the rifts at breakup time, then the mantle temperature anomaly would have been smaller. However, if a significant amount of melt flowed laterally away from the rift zones, either on the surface as flows, or at depth as dykes and sills, then our temperature estimates would be on the low side.

While the timing and extent of the volcanism provides control on the magnitude and extent of the abnormally hot mantle underlying the region at breakup time, the uplift and subsidence history provides further information on the dynamic support from the mantle plume, which is proportional to the product of the size of the mantle thermal anomaly and its thickness. Transient dynamic support of a few hundred metres, unaccompanied by any rifting or magmatism, has been reported by Nadin *et al*. (1995) during the early Tertiary in the northern North Sea. This area is well removed from the line of continental breakup, showing that abnormally hot mantle flowed widely beneath the region. Subsidence modelling of data from DSDP holes in the Edoras Bank region of the Rockall margin shows, furthermore, that the abnormally hot asthenospheric mantle in this distal region of the plume anomaly was only a few tens of kilometres thick, and that the thermal anomaly and associated dynamic support decayed rapidly over a period of 5-10 m.y. from ~150°C above normal to only ~50°C above normal shortly after continental breakup (Barton & White 1997b).

From these observations a picture emerges of the extremely rapid propagation of a thin sheet of mantle a few tens of kilometres thick and ~150°C hotter than normal beneath the unrifted continental lithosphere at around 62 Ma. This gave rise to volcanism across a large area, including eastern and western Greenland, the Faeroes, Ireland and the British Tertiary Province. This early phase of volcanism was succeeded by a much larger volume of magmatism as the continental lithosphere stretched and thinned during the process of continental breakup, allowing decompression melting of the hot mantle beneath the rift. The rapid drop of mantle temperature following this phase of volcanism suggests that the widespread sheet of abnormally hot sub-lithospheric mantle intruded prior to continental breakup was absorbed into the melting beneath the rift, and not subsequently replenished (Barton & White 1997; Saunders *et al.* 1997). Thereafter, the pattern of thermal anomalies in the mantle continued in a more axisymmetric shape, with a narrow hot plume beneath Iceland which generated the Greenland-Iceland-Faroes Ridge as the Greenland and European plates moved apart, and a surrounding region of moderately hot material fed laterally from the central plume.

Several different models have been suggested to explain the widespread occurrence of this early Tertiary magmatism. One is that there was a long period of incubation of hot asthenospheric mantle plume material prior to the onset of rifting and volcanism (Kent *et al.* 1992; Saunders *et al.* 1992): however, the simultaneous and widespread onset of volcanism across the entire region and the absence of earlier evidence of uplift or magmatism suggests that such a long period of incubation did not occur in the North Atlantic region (White & McKenzie 1989; Saunders *et al.* 1997).

A second set of models assumes that the onset of volcanism was associated with the arrival of a new mantle plume, which itself had arisen as a boundary layer instability from much deeper in the mantle: such boundary layer instabilities commonly exhibit enhanced temperatures and larger rates of flow in the initial phase, and both these are consistent with the observations (White & McKenzie, 1989). The precise shape of the thermal anomaly is less clear. One possibility to explain the distribution of initial volcanism is that the mantle instability originated as either three or four rising sheets of abnormally hot mantle, which subsequently developed into a more conventional axisymmetric plume centred at the meeting point of the sheets beneath where Iceland now sits. Using this explanation Barton & White (1997b) suggested a tripartite system of hot mantle sheets, with one sheet extending south between Rockall Plateau and southern Greenland, another northeastwards between eastern Greenland and Norway, and a third westward beneath central Greenland to Baffin Island. The first two mantle sheet upwellings lay along the line of subsequent continental breakup, coincident with the locations of maximum igneous thickness. The third sheet, crossing central Greenland, explains the simultaneous, high-temperature melts extruded on both Disko Island, west Greenland, and the east Greenland margin at c. 62 Ma. Although the thick ice cover prevents direct mapping of the geology beneath central Greenland, lineations of marked magnetic (Roest *et al.* 1995) and gravity anomalies along the line of the postulated mantle sheet beneath Greenland have been interpreted by Brozena (1995) as being due to igneous intrusions into the crust. N. J. White (*pers. comm.* 2001) has drawn attention to a possible fourth lineation of mantle melts extending southeastward to the Irish sea (Figure 1): the evidence for this is in the abundance of early Tertiary magmatism along this lineation, together with the evidence for marked uplift which caused c. 2.5 km of denudation in the east Irish sea region between 62-54 Ma (Lewis *et al.* 1992;

Rowley & White 1998). An alternative way to explain the short-lived burst of volcanism in the west Greenland area is that it was caused either by a completely separate mantle plume, or by a blob of hot mantle that detached itself from the main plume beneath the North Atlantic. Lawver & Müller (1994) explained the west Greenland and Atlantic magmatism by postulating that an axisymmetric plume was located beneath central Greenland at the time of breakup, and that only subsequently did it migrate eastward to cross the east Greenland coast at about 35 Ma. We think that this explanation is unlikely, since it cannot explain the continuous formation of the thick crust of the Greenland-Iceland-Faeroes Ridge, which we attribute to a mantle plume that lay close to the rift beneath eastern Greenland at the time of breakup (at the centre of the projection in Figure 1) and continued to sit beneath the North Atlantic ridge axis as the ocean opened, as postulated by White & McKenzie (1989).

On the evidence for an extensive thin horizontal sheet of anomalously hot mantle beneath the lithosphere required to explain the uplift and subsidence patterns, the most likely explanation seems to be that the initial widespread thermal anomaly resulted from the lateral injection just beneath the lithospheric lid of a thin (a few tens of kilometres thick) sheet of abnormally hot asthenospheric mantle from a central plume feeder. The high temperature and low viscosity of such a thin sheet would allow it to be injected rapidly with little conductive heat loss (Larsen *et al.* 1999), and would explain the transient uplift without accompanying volcanism seen, for example, in the northern North Sea, far from the location of eventual breakup (Nadin *et al.* 1995; Barton & White 1997b). Volcanism only occurred where the mantle could decompress, either beneath a pre-existing region of thin lithosphere, or in an area where rifting was actively thinning the lithosphere (Saunders *et al.* 1997). The subsequent rapid cooling of the thin sheet would occur partly in response to decompression and associated partial melting, and partly as a result of conductive heat loss into the adjacent colder mantle that lay both above and below the thin sheet of hot mantle.

Seafloor Spreading

Continent-Ocean Transition

Compared to typical non-volcanic margins, the interval between the beginning of continental stretching and the onset of full seafloor spreading was extremely short in the North Atlantic, lasting perhaps 4-6 m.y., and the width of the continent-ocean transition zone is narrow, reaching only a few tens of kilometres. This compares with the long periods of pre-breakup stretching, perhaps lasting 25 m.y. or more, and the great widths of stretched lithosphere, extending across 150 km or more, found on many non-volcanic margins. The reason is probably because the injection of huge volumes of melt into the stretching continental lithosphere weakened it greatly, allowing rapid thinning and progress to seafloor spreading, and because the added gravitational potential provided by the mantle plume in a region predisposed to extension throughout the Mesozoic may have encouraged continental breakup.

Flowlines

Prior to making reconstructions of the seafloor spreading in the North Atlantic, we calculated flowlines that reproduce the rotations between the Greenland and Eurasian plates (Figure 2). We restrict our discussion primarily to the region south of Iceland (the Irminger and Iceland Basins, Figure 1), because to the north the picture is complicated by major ridge jumps which isolated the continental fragment of Jan Mayen in the mid-Tertiary.

Seafloor spreading magnetic anomalies 20-23 are easy to recognise in the North Atlantic, forming linear, unbroken anomalies with no major offsets south of Iceland (Figure 2c). The most landward anomaly is usually part of anomaly 24, though due to the multiplicity of magnetic reversals between chrons 24 and 25 (Cande & Kent 1995), it is not always possible to be certain which of these, or what combination of them, generated the main magnetic anomaly stripe. The continent-ocean boundary (COB) itself is unlikely to be a precise line, instead representing a continent-ocean transition from highly stretched and intruded continental crust to dominantly new igneous crust. We show in Figure 4 the approximate location of the COB based on wide-angle seismic surveys extrapolated by the gravity maps along the margins. Pre-breakup reconstructions would close to a position somewhat landward of the COB, to account for the stretching of the continental crust prior to breakup. On the western margin of the Rockall Plateau we estimate this pre-breakup stretching to be of the order of 25-35 km from wide-angle seismic experiments.

Several authors have published reconstruction poles between Greenland and Eurasia back to anomaly 24 (e.g. Srivastava & Tapscott 1986; Müller & Roest 1992; Wold 1995).
In Figure 3 we show a comparison of flowlines calculated using different authors' finite rotations. The flowlines calculated using Srivastava & Tapscott's (1986) rotations fit the present-day azimuth of spreading at the Reykjanes Ridge well, but the slow change of azimuth with time means that they are not parallel to the fracture zones formed during the period between chrons 18 and 5 when the ridge was morphologically segmented on a 30-80 km scale. Flowlines calculated using Müller & Roest's (1992) finite rotations also strike obliquely to the fracture zone traces between chrons 18 and 13. The finite rotations of Wold (1995) produce a pronounced kink in the flowline at chron 21 (Figure 3): we believe this may be because he has included in his rotation calculations identifications of magnetic anomalies 7 (from the north) and 20, 21, 22 and 24 (from the south) over the Faeroe-Iceland Ridge. Identification of seafloor spreading anomalies from lavas erupted subaerially, as were those on the Faeroe-Iceland Ridge, is likely to be extremely uncertain. Factors leading to uncertainty include: the long flow lengths of lava flows, and their complex interaction with topography; the subsequent erosion of 1-2 km of the uppermost crust of the Faeroes-Iceland Ridge; and the likelihood of small-scale shifting rift zones and possible multiple rifts such as are observed today in Iceland over the mantle plume in a similar setting to that which generated the Faeroe-Iceland Ridge.
Here we use a combination of Helman's (1989) North America-Eurasia rotations and Roest & Srivastava's (1989) North America-Greenland rotations to obtain finite rotation poles for Greenland-Eurasia (Table 1). Helman's (1989) rotations are based on magnetic anomalies both in the region south of Iceland that we are investigating, and in the Atlantic south of the Charlie Gibbs fracture zones: they give synthetic fracture zone traces (flowlines) that match well the trends of the Charlie Gibbs fracture zones. Seafloor spreading at the Ran Ridge in the Labrador Sea ceased prior to anomaly 13 time (Kristofferson & Talwani 1977). Detailed study of the final spreading episode of the Ran Ridge suggests that the spreading slowed between chrons 21 and 13 (47-33 Ma) (Louden *et al.* 1996). We assume that the slowing occurred linearly between chrons 21 and 13. The resultant rotations produce flowlines that match the fracture zone traces well (Figure 2), with a fairly smooth angular change as spreading slowed down and stopped in the Labrador Sea (solid line, Figure 3).

Reconstructions

Finite rotations are used here to reconstruct the opening of the North Atlantic south of Iceland from shortly after breakup time to the present day, rotating some of the most prominent seafloor spreading magnetic anomalies to illustrate major changes in the style of seafloor spreading and ridge segmentation (Figures 4a-e). The oceanic crust formed during seafloor spreading falls into four main regimes. These changes are probably directly related to changes in the flow pattern or temperature of the Iceland mantle plume.

The very oldest oceanic crust comprises a zone of seaward dipping reflectors, up to 150 km wide at its greatest width, generated between breakup time and chron 24 (i.e. between approximately 56-53 Ma). Because they were extruded sub-aerially, clear seafloor spreading magnetic anomalies were not always developed, although fine lineations perpendicular to the spreading direction have been recognised off the east Greenland margin (Larsen & Saunders 1998). This zone is bounded at the seaward end by the prominent magnetic anomaly 24. Although an oceanward zone of sdrs is developed on both sides of the Atlantic basin (e.g. Larsen & Jakobsdóttir 1988; Mutter & Zehnder 1988; Barton & White 1997a), there appears to be a gross asymmetry, with anomaly 24 lying some 100-150 km off the COB on the Greenland margin, but lying close to the COB on the Rockall Plateau margin (Figure 4a). This asymmetry was probably removed by one or more minor eastward ridge jumps prior to anomaly 24, after which the seafloor spreading in the bulk of the region was broadly symmetric on both sides of the spreading axis. We return in a later section to discuss the rate of seafloor spreading during formation of this oldest oceanic crust.

The second main phase of seafloor spreading occurred between chrons 24 and 21 (53-46 Ma). This generated prominent seafloor spreading anomalies, with the direction of spreading perpendicular to the spreading axis (see flowline on Figure 2c). The spreading axis was unbroken by any fracture zones other than a 50 km dextral offset between the northern and southern parts of the ridge axis south of Iceland (Figure 4b). Oceanic crust was somewhat thicker than normal (9-11 km, compared to ~6 km for normal ocean basins), indicative of asthenospheric mantle temperatures in the region being some 50ºC above normal (White 1997). In tectonic style, this interval of spreading most closely resembles the fast-spreading ridges such as the East Pacific Rise at the present day. We suggest that the similarity arises because the enhanced melt production and thicker than normal crust produced in the North Atlantic produces a rheologically weak spreading axis. On the East Pacific Rise, although the crust is thinner, the frequent crustal melt injection episodes produced by the fast spreading rate maintain the spreading axis at a hotter temperature, on average, than that of a normal slow-spreading ridge. So either fast spreading or enhanced melt production on a slow-spreading ridge may have a similar effect of producing a weak spreading axis. Between anomalies 22 and 21 the dextral offset in the central part of the spreading axis was removed by a westward ridge jump, producing an unbroken, linear axis at anomaly 21 time (see Figure 4c). The ridge jump has left two sections of anomaly 24 on the eastern side of the basin south of the Faroe-Iceland Ridge (Voppel *et al*. 1979). The extinct ridge axis (dashed line on Figure 2a and arrowed on Figure 2b), has been previously recognised in gravity anomaly data (Roest *et al*. 1995).

A major change of spreading style accompanied a marked change in spreading direction after chron 20 (43 Ma). Reorientation of the azimuths of the spreading axes produced a third phase of seafloor spreading style, with a zone of short (30-80 km long) spreading segments offset by small transform faults. The spreading segments strike approximately perpendicular to the new spreading direction, and the transform faults lay approximately parallel to the spreading direction, although like small-offset transforms found elsewhere, locally they may deviate from it considerably. This change in style can be seen in both the gravity and magnetic maps in Figure 2, and can also be seen in the sediment distribution mapped using seismic reflection profiles (Ruddiman 1972). This style of spreading, typical of slow-spreading ridges formed away from the influence of mantle plumes, occurred diachronously in the North Atlantic, as shown by the region outlined by broken white lines in Figures 2b,c. The reduced residual depth anomalies in this region suggests that the mantle was less hot than during the preceding and succeeding intervals (White 1997), which is consistent with reversion to normal seafloor spreading. The diachronous onset is consistent with the mantle thermal anomaly generated by the Iceland plume withdrawing slowly northward, associated with a cooling of the plume material fed southward from the plume core.

The fourth and final change of spreading style in the North Atlantic saw a gradual reversion to an unsegmented, but now obliquely spreading axis on the Reykjanes Ridge. This propagated from the north, starting at approximately chron 9 (27 Ma), (Figures 4c-e). We suggest that this is indicative of somewhat higher temperature mantle feeding southward from the plume core beneath Iceland, allowing the development of abnormally thick crust in the northern part of the Reykjanes Ridge. South of 57ºN on the Reykjanes Ridge, the crustal thickness drops toward a more normal 7 km, and a median valley reappears (White 1997).

These gross changes in seafloor spreading style on timescales of 10-20 Ma are probably related to large-scale changes either in the temperature and/or the flow rate of the mantle plume, or in the relationship between the narrow rising core of the mantle plume and the spreading axis. As we show later, superimposed on these major changes are shorter term fluctuations in the plume on timescales of 1-3 m.y., which generated prominent 'V-shaped' ridges on the Reykjanes Ridge (Vogt 1971).

Ridge Jumps

We have already discussed the likelihood of small ridge jumps during the initial few million years of seafloor spreading. However, in the Irminger and Iceland basins there is little indication of resolvable ridge jumps thereafter, apart from the small changes in the spreading axis that must have occurred as the spreading regime reoriented from linear to segmented and back again.

There is a long history of major ridge jumps in the region close to the core of the Iceland plume, in the Greenland-Iceland-Faeroes Ridge itself. This was produced above the region of hottest mantle, with temperature anomalies of perhaps 150ºC above normal. So it may be no coincidence that the most frequent ridge jumps elsewhere in the Iceland and Irminger basins are recorded from the first few million years of seafloor spreading, when we infer that the mantle beneath the rifting continental margins and the newly-formed oceanic crust was hotter than subsequently (Barton & White 1997b).

The best documented ridge jumps on Iceland are those which have moved the rift axes eastward so as to keep them centred approximately above the core of the mantle plume, as the plume has drifted eastward with respect to the spreading axis. Two major eastward ridge jumps have been documented in northern Iceland, one at ~16 Ma from Vestfirdir to the Skagi Peninsula, and another during 7-3 Ma from the Skagi Peninsula to the present Northern Volcanic Zone (NVZ) (Saemundsson 1974, 1979; Helgason 1985; Hardarson *et al*. 1997) (see Figure 2a for locations). Similar ridge jumps toward the east have occurred in south Iceland (Krisjánsson & Jónsson 1998), producing at the present day a ~150 km eastward offset of the rift zones on Iceland compared to the offshore spreading centres of Reykjanes Ridge to the south and Kolbeinsey Ridge to the north (Figure 2a). The ridge jumps do not occur as instantaneous transfers of spreading from the old rift axis to the new one, but rather occur over a period (estimated as ~ 4 m.y. for the most recent ridge jumps of the NVZ), during which spreading is distributed across both the old and the new rift zones.

There are also several other more tentative identifications of ridge jumps on the portions of the Greenland-Iceland-Faeroes Ridge that are now underwater. Smallwood *et al.* (1999) have postulated at least two westward ridge jumps from the Faeroes-Iceland Ridge, one of which is marked by a change in basement structure, and the other (at ~40 Ma) at the position of the abrupt change in crustal thickness that lies beneath the present Icelandic shelf edge (Figure 5). On the western side of Iceland, Larsen & Jakobsdóttir (1988) have identified a ridge jump on the Greenland-Iceland Ridge from an angular unconformity in the lavas.

This multiplicity of ridge jumps directly above the track of the core of the mantle plume is not mirrored in the regions to the north and south of the plume. This means that the thickened crust of the Greenland-Iceland-Faeroes Ridge which formed above the plume core, is often bounded by transform faults of variable offset (Nunns 1983). At the present day, for example, the Tjörnes Fracture Zone off northern Iceland connects the Northern Volcanic Zone to the Kolbeinsey Ridge. Another consequence of the frequent ridge jumps is that they may cause abrupt changes in crustal thickness along the plume track. Where the rift zone lies directly above the narrow plume core, igneous crust of up to 40 km thickness may be generated, as is currently the case in Iceland (Darbyshire *et al.* 1998). Thus, the crustal thickness changes along the Greenland-Iceland-Faeroes Ridge probably reflect both variations in the precise location of the rift axis with respect to the narrow core of the mantle plume, as well as long-term variations in the temperature or flow rate of the plume which, as discussed in the previous section, also affect the oceanic crust formed much further from the centre.

Short Term Plume Fluctuations

A characteristic feature of the oceanic crust generated south of Iceland is that it exhibits prominent ‘V-shaped’ ridges in the bathymetry and gravity signatures (Figure 2). These were first recognised by Vogt (1971). They propagate away from the plume core with apparent velocities of 75-150 mm/a, some 4-8 times faster than the full spreading rate. Geochemical and seismic data suggest that they are generated by crustal thickness variations of 1-2 km (White *et al.* 1995, Smallwood & White 1998, Weir *et al.* 2001), equivalent to mantle temperature fluctuations of ~30ºC, on timescales of 3-5 Ma. Similar gravity lineations can also be discerned in the oceanic crust formed soon after breakup, although the thicker sediment cover on the older crust tends to reduce their prominence.

Indirect effects of the short-term temperature fluctuations in the mantle can also be seen in two other ways. First, around breakup time in the Palaeocene, the mantle temperature fluctuations produced rapid changes in uplift and subsidence in the regions bordering the rift zone, which have been used to explain pulses of sedimentation recorded particularly in fan deposits (White & Lovell 1997). Second, during the Neogene, regional variations in uplift associated with changes in the buoyancy flux of the Iceland plume have been invoked to explain changes in the amount of Northern Component water fed southward into the ocean basin south of Iceland (Wright & Miller 1996): the Greenland-Iceland-Faeroes Ridge acted as a barrier to this oceanographic flow during periods of uplift associated with high buoyancy flux from the Iceland plume, but was lowered and thus allowed passage of the Northern Component water during periods of reduced mantle buoyancy flux.

It therefore seems that small-scale, relatively short-term fluctuations in the mantle temperature have been a consistent feature of the Iceland plume throughout its history. It is likely that similar fluctuations occur in all mantle plumes, but that they can be detected well in the North Atlantic because the plume has lain for a long period directly beneath the seafloor spreading centre: in this configuration the passive mantle decompression that occurs in the spreading axis is an extremely sensitive measure of small temperature changes.

Seafloor Spreading Rates

There have been suggestions that not only was the duration of the breakup stage very short, but that the initial rate of seafloor spreading was extremely high, reaching 88 mm/a full rate for a period of 3 m.y. immediately following breakup (from ca 55.3-52.3 Ma), before dropping back to the full spreading rate of ~22 mm/a found regionally in the northern North Atlantic for the interval between chrons 23n and 20n (Larsen & Saunders 1998). This short-lived, extremely high spreading rate episode is determined only from an aeromagnetic survey off the east Greenland margin, based on lineations seen in the aeromagnetic map and on radiometric age determinations from basalts drilled in ODP sites 915, 917 and 918.

We discuss here an alternative explanation of the magnetic and other data that suggest that spreading rates immediately after breakup reached only 25-30 mm/a full rate, decreasing only slightly after chron 21 (47 Ma) to an average of ~20 mm/a.

We calculate half-spreading rates from two magnetic anomaly profiles shot along flowlines which extend from the continental margin to the present seafloor spreading axis in the Iceland Basin. Profile A is from *RRS Charles Darwin* cruise 70 and Profile B from *RRS Charles Darwin* cruise 87 (for location see Figure 2a). We also use identifications from the regional magnetic anomaly maps (Figure 2c) (MacNab *et al*. 1995), and from a detailed aeromagnetic survey in the Irminger Basin (Mercuriev *et al*. 1994).

Anomaly identifications were aided by first constructing a synthetic magnetic anomaly profile, including an appropriate skewness for the magnetic latitude at which the crust was generated (Figure 6). Our identifications of the anomaly peaks are shown above profiles A and B in Figure 6. Note that the anomaly picks on the oldest oceanic crust closest to the COB have some uncertainty, due to them being close to the end of the seafloor spreading anomalies, and to the possible 'smearing' by lateral lava flow in the seaward dipping reflectors, together with possible diffuse spreading or minor ridge jumps in the earliest stage of seafloor spreading. Spreading rates from the two flowline magnetic anomaly profiles are compared in Figure 7 with the rates calculated at the same latitudes from the Greenland-Eurasian rotation poles used to make the reconstructions in Figure 3.

Half-spreading rates along our magnetic anomaly flowlines stabilise after anomaly 21 (47 Ma) to ~ 10 mm/a. Similar rates are found from both sides of the ocean, showing that spreading about the axis was symmetric. However, during the initial period of seafloor spreading the half-rates were appreciably higher. Over the period between anomalies 24 and 21, we deduce half-rates of about 15 mm/a from our magnetic profiles (Figure 7). We suggest that during the period from continental breakup at about 55 Ma to anomaly 23n (52 Ma), the spreading rate was only half of the rate of 88 mm/a suggested by Larsen & Saunders (1998), because they failed to recognise a ridge jump that occurred in the middle of generating the seafloor magnetic lineations that they report from off the east Greenland margin. Such a ridge jump is necessary to explain the production of ~100 km of oceanic crust between the Greenland COB and the end of the flowlines at anomaly 24 time off the Greenland margin, since on the Eurasian side the end of the flowlines lie close to the Eurasian COB (Figure 2). With the identification of this ridge jump all along the east Greenland margin, the observed half-spreading rate during this early interval, using the same ages as Larsen & Saunders (1998), drops to 22 mm/a. This is just a little higher than the half rate of 15 mm/a we find for the period immediately succeeding it. As we noted in an earlier section, ridge jumps apparently occur more readily above the hottest mantle, such as beneath Iceland at the present day, so the ridge jumps off the Greenland margin are consistent with our observation that transient abnormally hot mantle lay beneath the line of continental breakup in the early Tertiary.

There is also some evidence in the seismic reflection profiles off Greenland (Larsen *et al.* 1998) for this ridge jump during anomaly 24r. In the central part of the 100 km portion of oceanic crust seaward of the Greenland COB containing seaward dipping reflectors is a zone of diffuse reflectivity which Larsen *et al.* (1998) interpret as a period when the seafloor spreading axis may have been submarine. They suggest that the remainder of the seaward dipping reflector sequence was generated with the seafloor at or above sea level. Consequently, this central diffuse zone may correspond to our postulated ridge jump, since the dying rift axis would sink below sea level as the magma productivity decreased. The new rift axis further to the east would then produce more voluminous lavas and seaward dipping reflectors that would flow across and overwrite the reflections from the older oceanic crust beneath them formed prior to the ridge jump. If the ridge jump did not occur as an instantaneous transfer from the old to the new rift axis, but rather went through an interval of simultaneous spreading as the old rift died down and the new one started up, there is a high likelihood of generating a diffuse zone of confused reflectivity in the region of the now extinct rift axis. Such a zone of confused reflectivity is observed off east Greenland.

Discussion

Interaction between the Iceland mantle plume and rifting, first of continental lithosphere, and subsequently of oceanic lithosphere, in the northern North Atlantic region allows us to investigate the long-term behaviour of a mantle plume.

At the time of initiation of the Iceland plume at ~62 Ma, the asthenospheric mantle fed by the plume was much more extensive and hotter than the subsequent plume flow that characterised its behaviour throughout the rest of the Tertiary. This is a feature of boundary layer instabilities, with both the flow rate and the temperature of the initial instability being much greater than the subsequent flow. However, from the existing data we cannot determine whether the plume instability arose from the upper/lower mantle boundary or from the core/mantle boundary.

The initial anomalously hot plume material spread widely across the region: its geometry may have comprised rising sheets of hot mantle, or separate blobs. However, within a relatively short period of less than 10 m.y., the initial disturbance had collapsed to the axisymmetric shape typical of the present day configuration. This has a narrow (~ 100 km diameter) central core of mantle probably ~100-150ºC hotter than normal, with a wide surrounding region of plume-fed asthenospheric mantle that is still hotter than normal mantle, but only by a few tens of degrees Celsius.

Several temporal scales of fluctuation can be recognised in the Iceland plume by their effects on the region. Long term fluctuations on the order of tens of millions of years produced regional changes in the structure of the oceanic crust. Shorter term fluctuations of temperature and flow-rate on 3-5 m.y. timescales can be recognised throughout the history of the Iceland plume and are probably characteristic of all such mantle plumes.

Where rifting has occurred directly above the mantle plume, ridge jumps are a common feature as the active rift migrates in an attempt to remain above the hottest upwelling mantle. This effect has been documented along extensive portions of the continental margin during continental breakup and during the early seafloor spreading history. Subsequently, it has been restricted to the crust formed directly above the upwelling plume core (which has produced the thickened crust of the Greenland-Iceland-Faeroes Ridge). It is presumably a result of the higher gravitational potential and weaker lithosphere found directly above the plume core, which makes rifting in that location easiest. The prevalence of ridge jumps during the earliest phase of seafloor spreading is consistent with the widespread extent of abnormally hot mantle during and immediately following continental breakup.

Acknowledgments

The reconstructions in this contribution were made using the ATLAS computer program developed by Cambridge Paleomap Services. We are grateful to those who sailed on *RRS Charles Darwin* legs 70 and 87 for collecting the magnetic anomaly data. We thank Andy Saunders for his comments on an earlier draft of this paper, which led to significant improvement. The opinions and interpretations expressed herein are those of the authors and not necessarily those of Amerada Hess Ltd. Department of Earth Sciences, Cambridge contribution number 6382.

References

Barton, A. J. & White, R. S. 1997a. Volcanism on the Rockall continental margin. Journal of the Geological Society, London, **154**, 531-536.

Barton, A. J. & White, R. S. 1997b. Crustal structure of the Edoras Bank continental margin and mantle thermal anomalies beneath the North Atlantic. Journal of Geophysical Research, **102,** 3109-3129.

Bown, J. W. & White, R. S. 1995. Effect of finite extension rate on melt generation at rifted continental margins. Journal of Geophysical Research, **100**, 18,011-18,029.

Brozena, J. M. 1995. Kinematic GPS and aerogeophysical measurement: gravity, topography and magnetics. PhD dissertation, University of Cambridge, 185pp. and maps.

Cande, S. C. & Kent, D. V. 1995. Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic. Journal of Geophysical Research, **100**, 6093-6095.

Clift, P. D., Turner, J. & Ocean Drilling program Leg 152 Scientific Party 1995. Dynamic support by the Icelandic plume and vertical tectonics of the northeast Atlantic continental margins. Journal of Geophysical Research, **100**, 24,473-24,486.

Darbyshire, F. A., Bjarnason, I. Th., White, R. S. & Flóvenz, Ó. G. 1998. Crustal structure above the Iceland mantle plume imaged by the ICEMELT refraction profile. Geophysical Journal International, **135**, 1131-1149.

Darbyshire, F. A., White, R. S. & Priestley, K. P. 2000. Structure of the crust and uppermost mantle of Iceland from a combined seismic and gravity study. Earth and Planetary Science Letters, **181,** 409-428.

Hardarson, B. S., Fitton, J. G., Ellam, R. M. & Pringle, M. S. 1997. Rift relocation - a geochemical and geochronological investigation of a palaeo-rift in NW Iceland. Earth and Planetary Science Letters, **153**, 181-195.

Helgason, J. 1985. Shifts of the plate boundary in Iceland: some aspects of Tertiary volcanism. Journal of Geophysical Research, **90**, 10,084-10,092.

Helman, M. 1989. Tectonics of the western Mediterranean. D.Phil dissertation (unpublished), Oxford University.

Kent, R. W., Storey, M. & Saunders, A. D. 1992. Large igneous provinces: sites of plume impact or plume incubation? Geology, **22,** 891-894.

Korenaga, J., Holbrook, W. S., Kent, G. M., Kelemen, P. B., Detrick, R. S., Larsen, H.-C., Hopper, J. R. & Dahl-Jensen, T. 2000. Crustal structure of the southeast Greenland margin from joint refraction and reflection seismic tomography. Journal of Geophysical Research, **105,** 21,591-21,614.

Krisjánsson, L. & Jónsson, G. 1998. Aeromagnetic results and the presence of an extinct rift zone in western Iceland. Journal of Geodynamics, **25**, 99-108.

Larsen, H. C. & Jakobsdóttir, S. 1988. Distribution, crustal properties and significance of seaward dipping subbasement reflectors off East Greenland. *In:* Morton, A. C. & Parson, L. M. (eds), *Early Tertiary Volcanism and the Opening of the NE Atlantic*. Geological Society of London Special Publication No. **39**, 95-114.

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**, College Station, Texas (Ocean Drilling Program), pp. 503-533.

Larsen, H. C., Dahl-Jensen, T. & Hopper, J. R.1998. Crustal structure along the leg 152 drilling transect. In: Proceedings of the Ocean Drilling Program, Scientific Results, **152**, 463-474.

Larsen, L. M. Waagstein, R., Pedersen, A. K. & Storey, M. 1999. Trans-Atlantic correlation of the Palaeogene volcanic successions in the Faeroe Islands and East Greenland. Journal of the Geological Society, London, **156**, 1081-1095.

Larsen, T. B., Yuen, D. A. & Storey, M. 1999. Ultrafast mantle plumes and implications for flood basalt volcanism in the northern Atlantic region. Tectonophysics, **311,** 31-43.

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

Lewis, C. L. E., Green, P. F., Carter, A. & Hurford, A. J. 1992. Elevated K/T palaeotemperatures throughout northwest England: three kilometres of Tertiary erosion? Earth and Planetary science Letters, **112,** 131-145.

Louden, K. E., Osler, J. C., Srivastava, S. P. & Keen, C. E. 1996. Formation of oceanic crust at slow spreading rates: New constraints from an extinct spreading centre in the Labrador Sea. Geology, **24**, 771-774.

Macnab, R., Verhoef, J., Roest, W. & Arkani-Hamed, J. 1995. New database documents the magnetic character of the Arctic and North Atlantic. EOS, Transactions, American Geophysical Union, **76**, 449 and 458.

Mercuriev, S. A., Sotchevanova, N. A., Macnab, R., Lévesque, S. & Oakey, G. 1994. Evidence for a propagating rift in Irminger Basin near Reykjanes Ridge: detailed magnetic and bathymetric investigations (abstract). EOS, Transactions, American Geophysical Union, **75**, suppl. no. 16, 131.

Müller, R. D. & Roest, W. R. 1992. Fracture zones in the North Atlantic from combined Geosat and Seasat data. Journal of Geophysical Research, **97**, 3337-3350.

Mutter, J. C. & Zehnder, C. M. 1988. Deep crustal structure and magmatic processes: the inception of seafloor spreading in the Norwegian-Greenland sea. *In:* Morton, A. C. & Parson, L. M. (eds), *Early Tertiary Volcanism and the Opening of the NE Atlantic*.

Geological Society of London Special Publication No. **39**, 35-48.Nadin, P. A., Kusznir, N. I. & Toth, J. 1995. Transient regional uplift in the Early Tertiary of the northern North Sea and the development of the Iceland plume. Journal of the Geological Society, London, **152**, 935-948.

Nunns, A.G. Plate tectonic evolution of the Greenland-Scotland Ridge and surrounding regions. 1983. In: Bott, M. H. P., Saxov, S., Talwani, M. & J. Thiede, J. (eds) Structure and Development of the Greenland-Scotland Ridge. Plenum Press, New York, 11-30.

Reid, I. D., Dahl-Jensen, T., Holbrook, W. S., Larsen, H. C., Kelemen, P. H., Hopper, P. R., Korenaga, J., Detrick, R. & Kent, G. 1997. 32-38 km thick mafic igneous crust beneath the Greenland-Iceland Ridge. EOS, Transactions, American Geophysical Union, **78**, 656.

Richardson, K. R., Smallwood, J. R., White, R. S., Snyder, D. & Maguire, P. K. H. 1998. Crustal structure beneath the Faroe Islands and the Faroe-Iceland Ridge. Tectonophysics, **300**, 159-180.

Richardson, K. R., White, R. S., England, R. W. & Fruehn, J. 1999. Crustal structure east of the Faroe Islands. Petroleum Geoscience, **5**, 161-172.

Roest, W. R. & Srivastava, S. P. 1989. Sea-floor spreading in the Labrador Sea: A new reconstruction. Geology, **17**, 1000-1003.

Roest, W. R., Verhoef, J. & Macnab, R. 1995. Regional tectonics of the Arctic and North Atlantic oceans and adjacent land areas, as seen in new maps of compiled magnetic observations (abstract). EOS, Transactions, American Geophysical Union, **76**, 46 suppl., 74.

Rowley, E. & White, N. 1988. Inverse modelling of extension and denudation in the east Irish sea and surrounding areas. earth and Planetary Science Letters, **161,** 57-71.

Ruddiman, W. F. 1972. Sediment redistribution on the Reykjanes Ridge: Seismic evidence. Geological Society of America Bulletin, **83**, 2039-2062.

Saemundsson, K., 1974. Evolution of the axial rifting zone in northern Iceland and the Tjörnes Fracture Zone. Geological Society of America Bulletin, **85**, 495-504.

Saemundsson, K., 1979. Outline of the geology of Iceland. Jökull, **2**, 7-28.

Sandwell, D. T. & Smith, W. H. F. 1997. Marine gravity anomaly from Geosat and ERS 1 satellite altimetry. Journal of Geophysical Research, **102**, 10,039-10,054.

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

Saunders, A. D., Storey, M., Kent, R. W. & Norry, M. J. 1992. Consequences of plume-lithosphere interactions. *In:* Storey, B. C., Alabaster, T. & Pankhurst, R. J. (eds), *Magmatism and the Causes of Continental Break-up*. Geological Society of London Special Publication No. **68**, 41-60.

Smallwood, J. R. & White, R. S. 1998. Crustal accretion at the Reykjanes Ridge, 61-62ºN. Journal of Geophysical Research, **103**, 5185-5201.

Smallwood, J. R., Staples, R. K., Richardson, K. R., White, R. S. & FIRE Working Group. 1999. Crust generated above the Iceland mantle plume: from continental rift to oceanic spreading center. Journal of Geophysical Research, **104**, 22,885-22,902.

Smallwood, J. R., White, R. S. & Staples, R. K. 1998. Deep crustal reflectors under Reydarfjördur, eastern Iceland: Crustal accretion above the Iceland mantle plume. Geophysical Journal International, **134**, 277-290.

Shackleton, N. J., Hall, M. A.,Raffi, I., Tauxe, L. & Zachos, J. 2000. Astronomical calibration age for the Oligocene-Miocene boundary. Geology, **28,** 447-450.

Srivastava, S. P. & Tapscott, C. R. 1986. Plate kinematics of the North Atlantic. In: Vogt, P. R. & Tucholke, B. E. (eds) The Geology of North America, M, The Western North Atlantic Region. Geological Society of America, 379-404.

Staples, R. K., White, R. S., Brandsdóttir, B., Menke, W., Maguire, P. K. H. & McBride, J. H. 1997. Färoe-Iceland Ridge Experiment, 1. Crustal structure of northeastern Iceland. Journal of Geophysical Research, **102**, 7849-7866.

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

Voppel, D., Srivastava, S. P. & Fleischer, U. 1979. Detailed magnetic measurements south of the Iceland-Faroe Ridge. Deutschland Hydrograph Zeit, **32**, 154-172.

Weir, N. R. W., White, R. S., Brandsdóttir, B., Einarsson, P., Shimamura, H. & Shiobara, H. 2001. Crustal structure of the northern Reykjanes Ridge and Reykjanes peninsula, south-west Iceland. Journal of Geophysical Research, **106,** 6347-6368.

White, N. J. & Lovell, B. 1997. Measuring the pulse of a plume with the sedimentary record. Nature, **387**, 888-891.

White, R. S. 1992. Crustal structure and magmatism of North Atlantic continental margins. Journal of the Geological Society, **149**, 841-854.

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

White, R. & 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. 1995. Mantle plumes and flood basalts. Journal of Geophysical Research, **100**, 17,543-17,585.

White, R. S., Bown, J. W. & Smallwood, J. R. 1995. The temperature of the Iceland plume and origin of outward propagating V-shaped ridges. Journal of the Geological Society, London, **152**, 1039-1045.

Wold, C. N. 1995. Palaeobathymetric reconstruction on a gridded database: the northern North Atlantic and southern Greenland-Iceland-Norwegian sea. In: Scrutton, R. A., Stoker, M. S., Shimield, G. B & Tudhope, A.W. (eds) The Tectonics, Sedimentation and Palaeoceanography of the North Atlantic Region. Geological Society, London, Special Publications, **90**, 270-302.

Wright, J. D. & Miller, K. G. 1996. Control of North Atlantic deep water circulation by the Greenland-Scotland Ridge. Palaeoceanography, **11**, 157-170.

FIGURE CAPTIONS

Figure 1. Reconstruction of the northern North Atlantic at 55 Ma, shortly after the onset of seafloor spreading (after White 1992). Shaded area shows known extent of lava flows and sills emplaced during continental breakup with dykes shown as thin lines. Note that the extent of any dykes emplaced beneath mainland Greenland is unknown due to the ice cover, although an approximately east-west track of circular magnetic anomalies across central Greenland (Roest *et al.* 1995; Brozena 1995) may represent a line of igneous centres. The configuration of a possible set of rising hot sheets of asthenospheric mantle at the time of breakup is outlined by heavy broken lines. The reconstruction uses an equal area Lambert stereographic projection centred on the core of the mantle plume, and encompasses an area with a radius of 1500 km.

Figure 2. Bathymetry, free air gravity anomaly and magnetic anomaly in the northeast Atlantic. Panels b and c include flowlines back to anomaly 24 time centred on the present spreading axis. (a) Bathymetry, 1000 m interval contours. Location of FIRE profile shown in Figure 5, and magnetic anomaly flowline profiles recorded by Charles Darwin 70 (Profile A) and Charles Darwin 87 (Profile B) shown in Figure 6 are marked. Present active rift axes are shown in red, and selected extinct rifts are shown by dashed red lines: S, Skagi; V, Vestfirdir; NVZ, EVZ, WVZ, northern, eastern and western volcanic zones; KR, Kolbeinsey Ridge; RR, Reykjanes Ridge; GIR, Greenland-Iceland Ridge; FIR, Faeroe-Iceland Ridge; IrB, Irminger Basin; IB, Iceland Basin; LS, Labrador Sea; ICE, Iceland; GB, Great Britain; RP, Rockall Plateau; FI, Faeroe Islands. (b) Free air gravity anomaly (Sandwell & Smith, 1997). Zone of ephemeral fracture zones outlined by broken white line. (c) Magnetic anomaly (Macnab *et al*. 1995).

Figure 3. Flowlines calculated for the Iceland Basin from a point on the Reykjanes Ridge, using a variety of published Greenland-Eurasia reconstruction poles. Points corresponding to magnetic chrons 13, 21 and 24 are shown as circles, squares and triangles respectively. Previously published reconstruction poles (Srivastava & Tapscott 1986, Muller & Roest 1992, Wold 1995) give flowlines that strike oblique to the fracture zones from chron 13, or later. The preferred flowline from this study is shown as a solid line (see text for details).

Figure 4. Reconstruction of seafloor spreading in the Irminger and Iceland Basins, with present land-masses shown shaded, based on finite rotations with Eurasian plate held fixed. (a) Shortly after the onset of seafloor spreading, at 54 Ma. The landward-most magnetic lineations (anomaly 24) in the Irminger and Iceland Basins are superposed. The approximate continent-ocean boundaries (COB) on the conjugate rifted margins are shown by broken lines: the region between the COB and the magnetic lineations are occupied by sdr sequences. The ridge crest was offset about 50 km dextrally. (b) Early unsegmented axis, chron 22 (49 Ma). Between chron 22 and chron 21 a ridge jump in the northern part of the ocean removed the dextral offset to give a continuous ridge axis by chron 21. (c) Transition from segmented to continuous ridge axis, chron 9 (27.5 Ma). The 30-80 km ridge segmentation formed around 40 Ma is now being combined into an unsegmented axis propagating from the north, with associated oblique spreading. (d) Chron 6 (20 Ma). Continued southward propagation of the unsegmented ridge axis has removed all but the southernmost ridge segmentation. (e) Present day, with oblique spreading along a continuous 750 km length of ridge crest with no major offsets. Fine line shows 2000 m contour, and broken lines show approximate location of COB.

Figure 5. Interpretation of crustal structure from seismic profile along FIRE line on the Faeroes-Iceland Ridge (see Figure 2a for location and Smallwood *et al.* 1999 for details). Note the abrupt crustal thickness changes associated with ridge jumps, particularly near the Iceland shelf edge and the present Northern Volcanic Zone at Krafla. COB marks continent-ocean boundary offshore the Faeroe Islands. Inferred region of underplating below the Faroes continental block is from Richardson *et al.* (1998). Zero on the distance scale is at the present rift axis under Krafla.

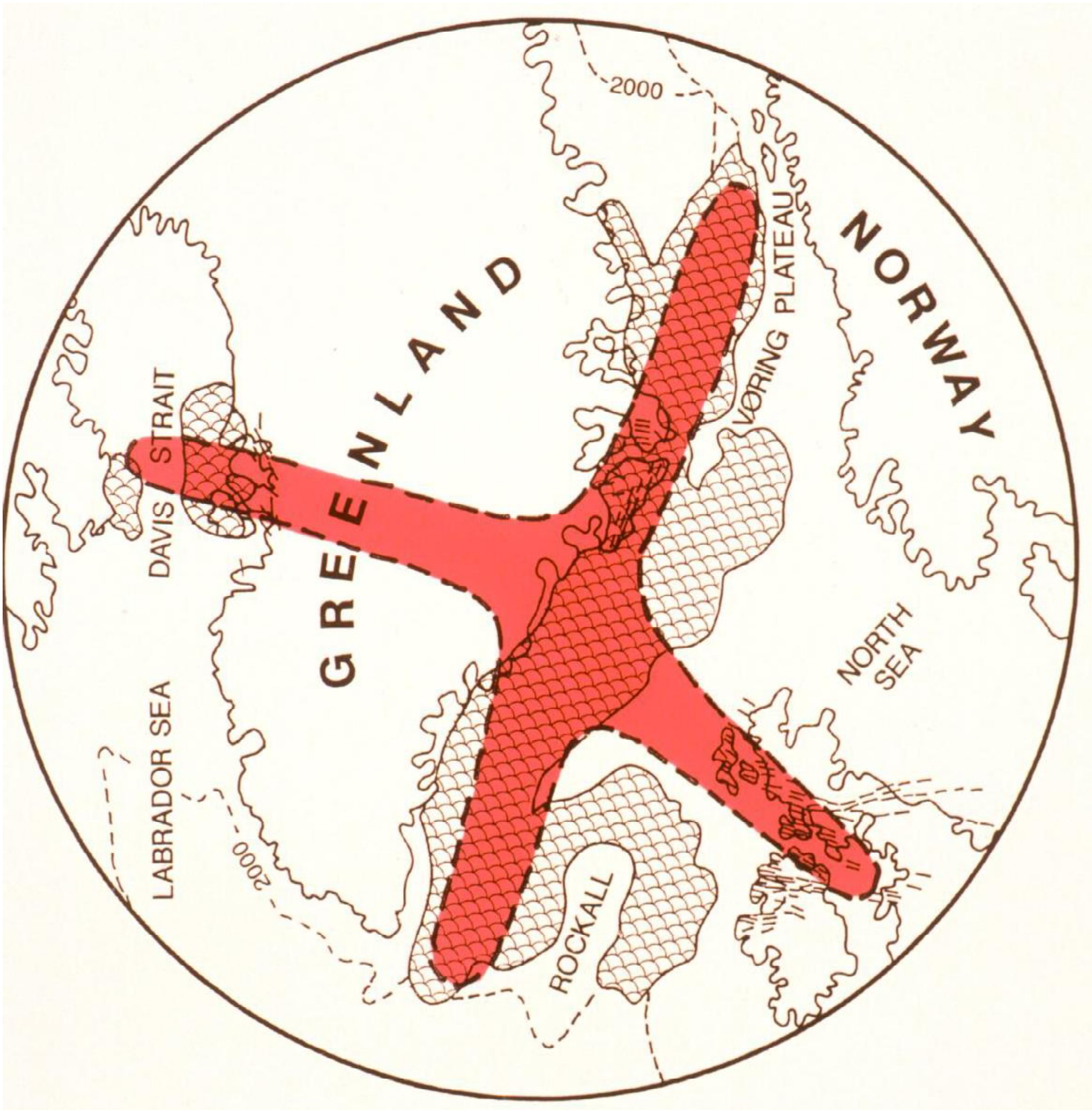
Figure 6. Magnetic anomaly flowline profiles. Top curve shows synthetic for a constant spreading half-rate of 10 mm/a using the reversal timescale from Cande & Kent (1995), with a constant thickness and uniform magnetization source layer and with a skewness of 42° to allow for the latitude at which the anomalies were formed. Bottom two curves show magnetic anomalies across the Iceland Basin along profiles A and B; see Figure 2a for locations. Anomaly identifications are annotated above each profile.

Figure 7. Distance along flowline versus magnetic chron age (Cande & Kent 1995 timescale modified by Shackleton *et al.* 2000) for two flowlines across the Iceland Basin. In both plots the distance is reduced at 10 mm/a, and the gradient gives spreading rate (see key). (a) Calculated rates from our rotation poles (see text for details); (b) Observed rates from anomaly identifications on profiles A and B. Tentative anomaly identifications are shown with open symbols. The lines show a 3-point smoothing filter fit to the data points. Estimated error on points is ±2 km. The thick dashed line indicates the period when the spreading centre was segmented for the profile A (below) and profile B (above).

Table 1. *Greenland-Eurasia Closure Poles*

	${}^{0}_{NAm}ROT^{t}_{Grn}$			${}^{0}_{Eur}ROT^{t}_{NAm}$			${}^{0}_{Eur}ROT^{t}_{Grn}$		
Chron	Lat, °N	Lon, °E	Angle,°	Lat, °N	Lon, °E	Angle,°	Lat, °N	Lon, °E	Angle,°
5A				68.00	137.00	3.01	68.00	137.00	3.01
8				66.85	135.46	5.97	66.85	135.46	5.97
13				62.50	138.50	7.30	62.50	138.50	7.30
17	62.80	-91.95	-0.21	62.50	138.50	7.83	61.41	137.50	7.69
18	62.80	-91.95	-0.48	62.50	141.50	8.39	60.17	139.32	8.08
20	62.80	-91.95	-1.33	62.50	141.50	9.29	56.25	136.41	8.47
21	62.80	-91.95	-2.61	55.70	143.25	9.81	42.20	135.83	8.54
24	55.86	-104.55	-4.44	61.10	144.90	12.04	43.17	128.30	9.87

Greenland-Eurasia closure poles, from the Eurasia-North America poles of Helman (1989) and the Greenland-North America poles of Roest & Srivastava (1989), assuming a linear slowing of spreading in the Labrador Sea from anomaly 21 to anomaly 13 time. Subscripts NAm, Grn & Eur refer to North American, Greenland and Eurasian plates respectively.



GREENLAND
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VØRING PLATEAU
DAVIS STRAIT
LABRADOR SEA
NORTH SEA
ROCKALL
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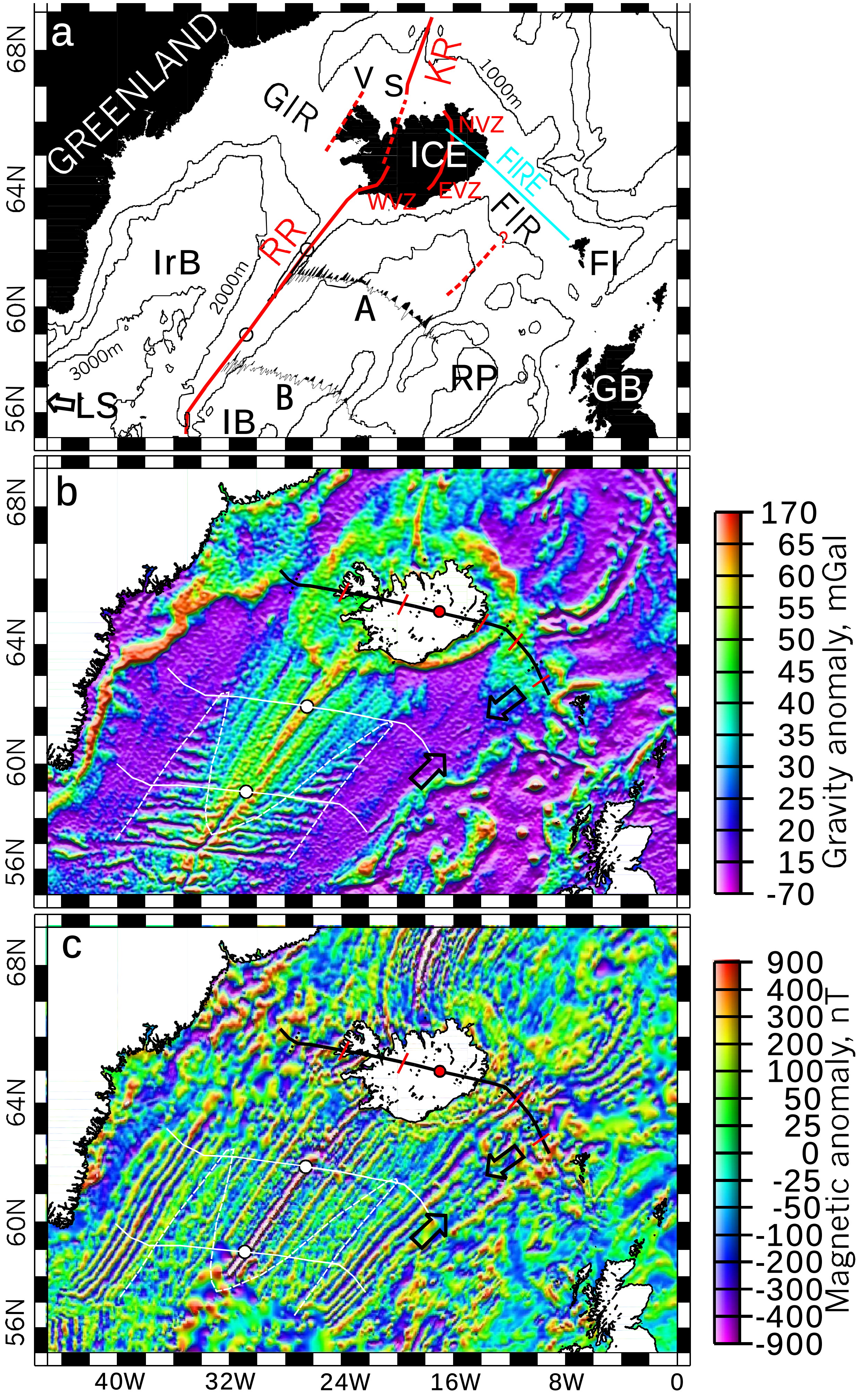
Figure 1



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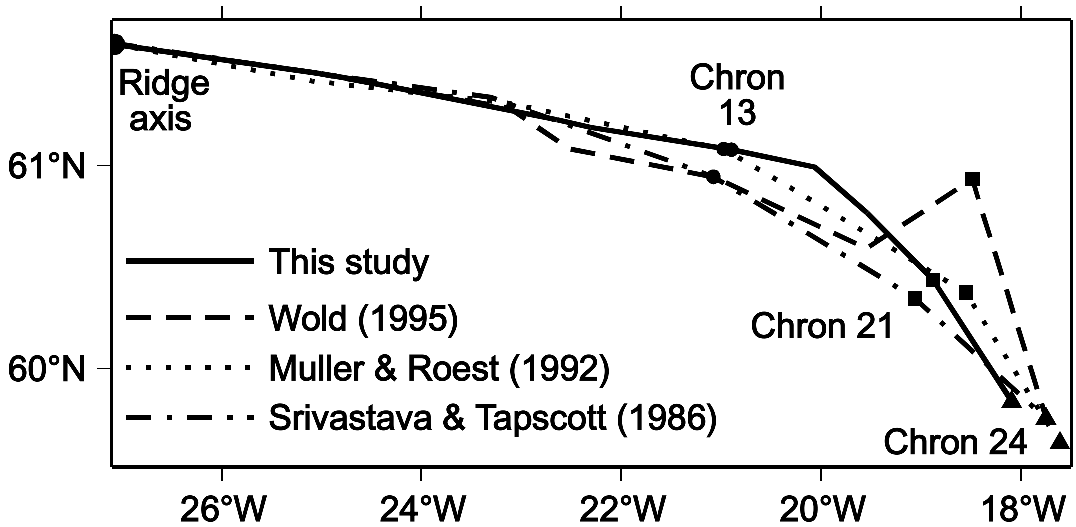
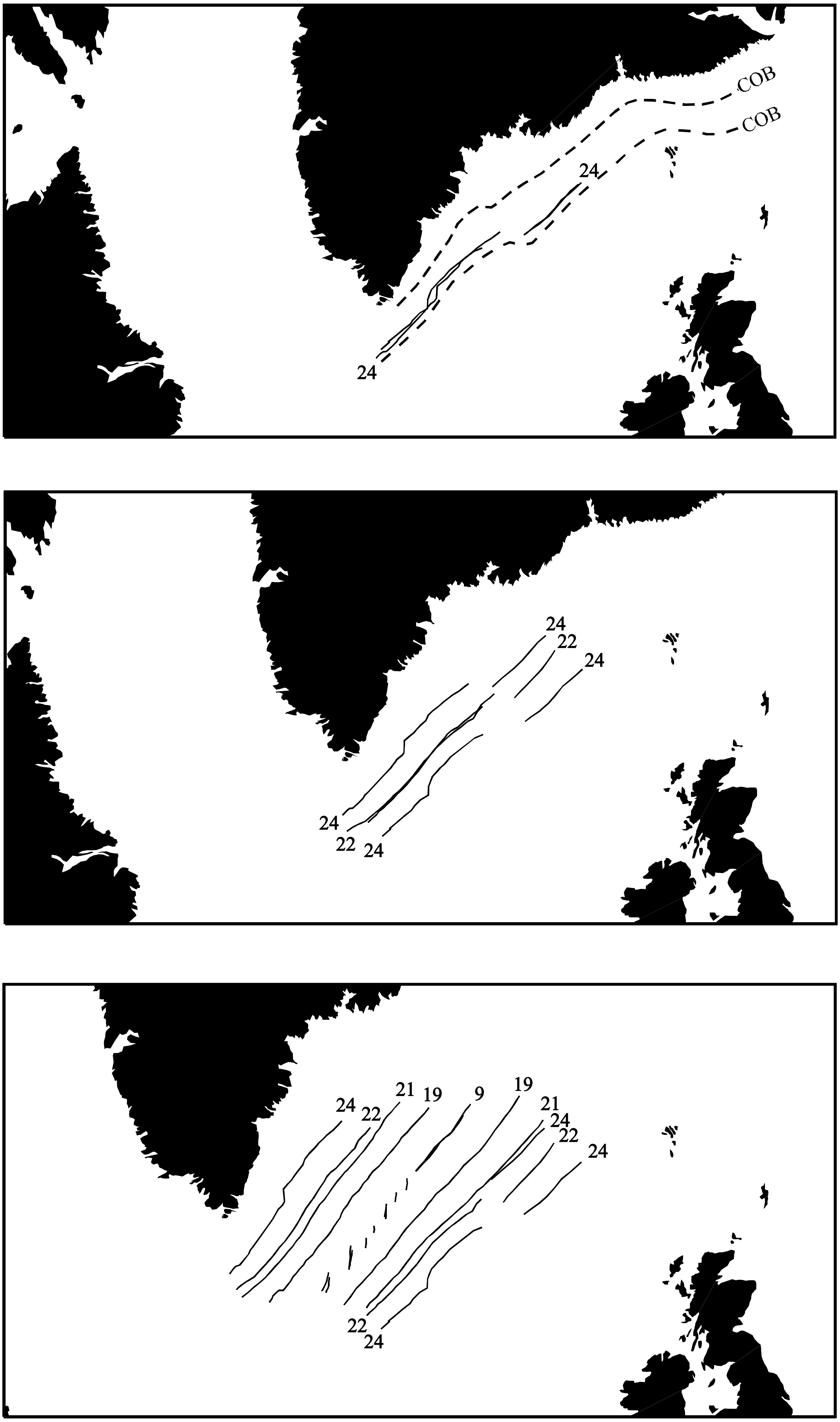
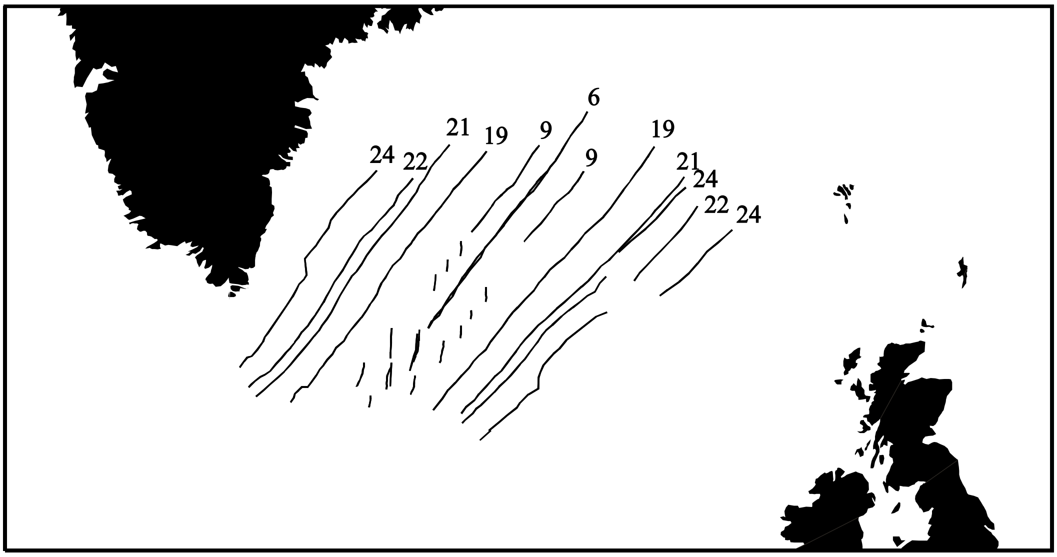
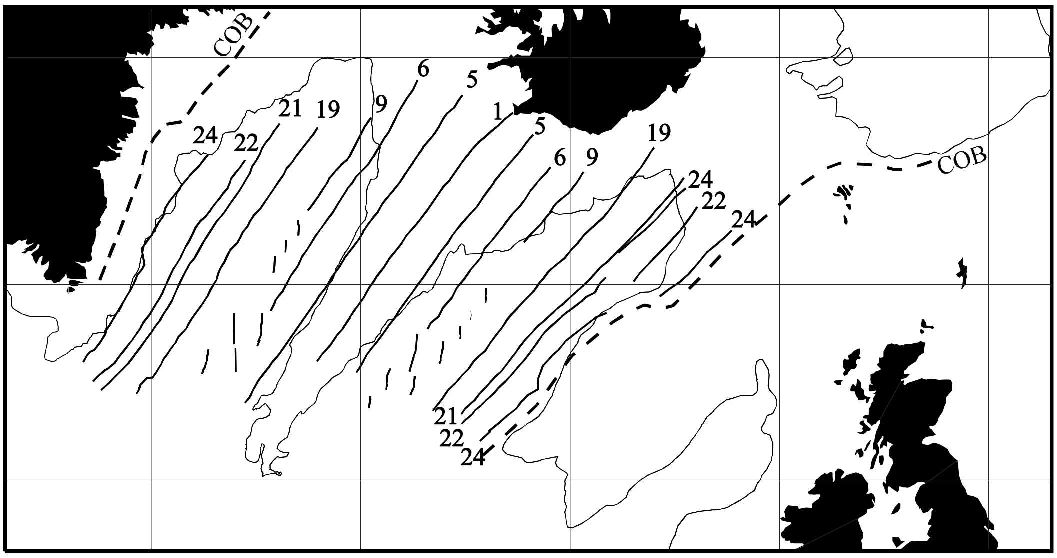


Smallwood & White Figure 3

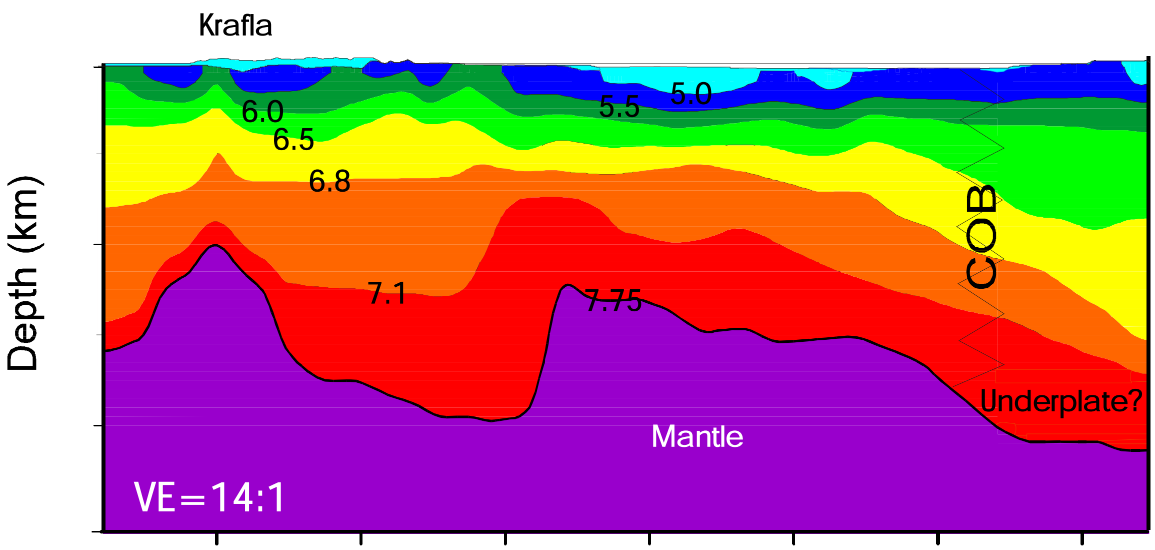


Smallwood & White Figure 4 (part 1)

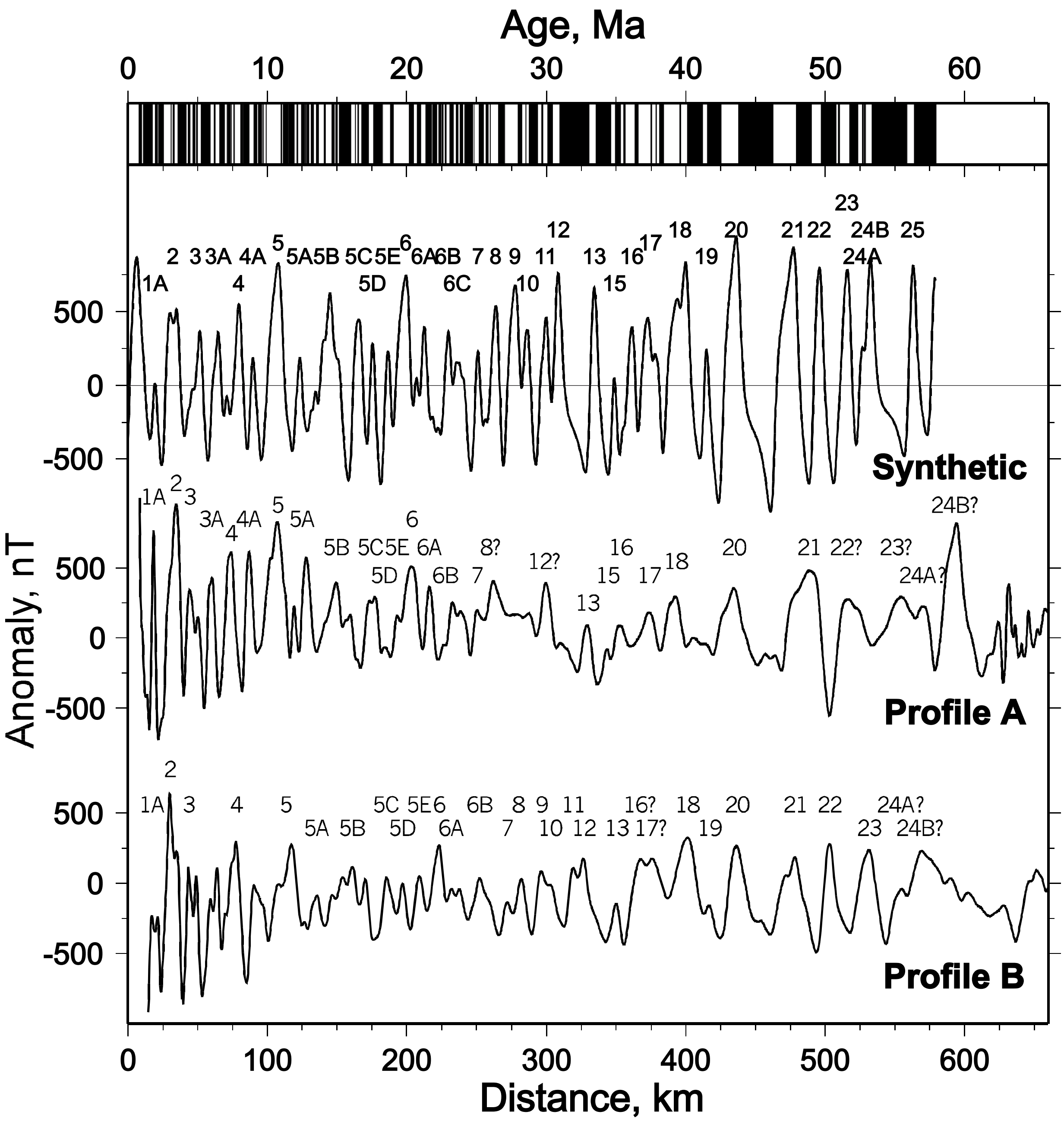




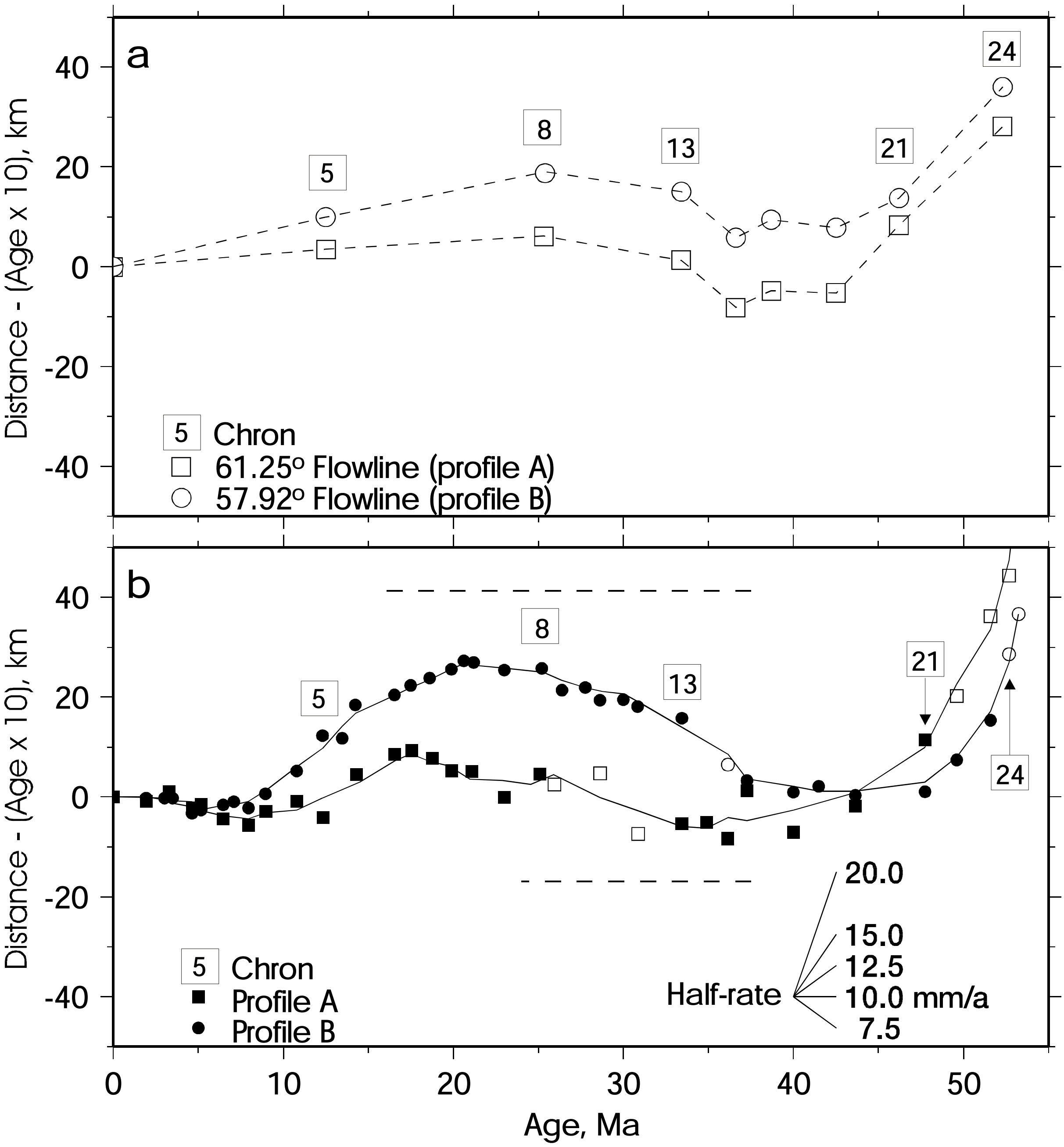
Smallwood & White Figure 4 (part 2)



Smallwood & White Figure 5



Smallwood & White Figure 6



Smallwood & White Fig 7