Available online at www.sciencedirect.com
SCIENCE DIRECT®
Journal of Volcanology and Geothermal Research 141 (2005) 1–22
Journal of volcanology and geothermal research
www.elsevier.com/locate/jvolgeores

A cool model for the Iceland hotspot

G.R. Foulger[a,*], Don L. Anderson[b]

[a]Department of Earth Sciences, University of Durham, Science Laboratories, South Rd., Durham DH1 3LE, UK
[b]California Institute of Technology, Seismological Laboratory, MC 252-21, Pasadena, CA 91125, USA

Received 23 July 2003; accepted 13 October 2004

Abstract

Several primary features of the Iceland region require a posteriori adaptions of the classical plume hypothesis to explain them, which erodes confidence in this model. These include the lack of a time-progressive volcanic track and the paucity of evidence for a seismic anomaly in the lower mantle. Diverse studies suggest a mantle potential temperature anomaly beneath the region of no more than 50–100 K, which is probably insufficient for a thermally buoyant plume. We suggest an alternative model that attributes the enhanced magmatism in the Iceland region to high local mantle fertility from subducted Iapetus oceanic crust trapped in the Laurasian continental mantle lithosphere within the collision zone associated with the Caledonian suture. This crust is recycled into the melt zone locally beneath the mid-Atlantic ridge where isentropic upwelling of eclogitized crust or a crust–peridotite mixture produces excess melt. The production of anomalously large volumes of melt on this part of the spreading ridge has built a zone of thick seismic crust that traverses the whole north Atlantic. A weak, downward continuation of the seismic low-velocity zone into the transition zone between the Charlie Gibbs and Jan Mayen fracture zones may correspond to a component of partial melt, too low to be extractable, that indicates the depth extent of enhanced fusibility or volatiles resulting from the recycled crust. The Iceland region separates two contrasting oceanic tectonic regions to its north and south that may behave independently to some degree. Perhaps as a result of this, it has persistently been characterized by complex and unstable tectonics involving spreading about a parallel pair of ridges, intervening microplates, ridge migrations, and local variations in the spreading direction. These tectonic complexities can explain a number of primary features observed on land in Iceland. A captured microplate that may contain oceanic crust up to ~30 m.y. old underlies central Iceland submerged beneath younger lavas. This may account for local thickening of the seismic crust to ~40 km there. Eastward-widening, fan-shaped extension about a west–east zone that traverses central Iceland culminates in northwest Vatnajokull and may cause the enhanced volcanism there that is traditionally assumed to indicate the center of a plume. The general locus of spreading has not migrated east as is often suggested in support of an eastward-migrating plume model. The model suggested here attributes the Iceland melting anomaly to structures and processes related to plate tectonics that are sourced in the shallow upper mantle. Similar models may explain other "hotspots." Such models suggest a simplifying view of mantle convection since they require

* Corresponding author.
E-mail address: g.r.foulger@durham.ac.uk (G.R. Foulger).

0377-0273/$ - see front matter
doi:10.1016/j.jvolgeores.2004.10.007

only a single mode of convection, that associated with plate tectonics, and not an additional second independent mode associated with deep mantle plumes.

Keywords: Iceland; plume; hotspot; plate tectonics; upper mantle

1. Introduction

The concept that the Iceland volcanic province results from a thermal plume rising from deep within the mantle is almost universally employed as an a priori assumption and a framework within which new data are interpreted. Nevertheless, several first-order observations at Iceland are unpredicted by the original classic plume model of Morgan (1971). This has not, however, led to abandonment of the model, which has instead been adapted to fit by the addition of a posteriori model supplements to account for individual observations. In this way, contemporary plume models are able to account for essentially any observation or the lack of expected observations at Iceland and other hotspots. At individual hotspots, such as Iceland, multiple mutually exclusive plume models may be extant, each devised to explain a particular data set.

These matters notwithstanding, it is not the objective of this paper to attempt to disprove the contemporary plume model at Iceland. We recognise that this is essentially impossible in view of the model flexibility. Nor does it fall within the remit of this paper to characterize the general plume model or reiterate the merits or shortcomings of individual variants since that is done in numerous other books and papers (e.g., see Davies, 1999 wherein a full treatment of the hypothesis may be found). Instead, we point out that the plume hypothesis is not proven, and that it should be subjected to critical scientific testing rather than accepted a priori without question. We take the approach of devising an alternative competing hypothesis in the hope that it may explain the observations at least as well if not better.

We present a set of ideas that may provide a starting point for such an alternative hypothesis in the case of the Iceland hotspot. The ideas outlined herein have not benefited from decades of numerical modeling as the plume hypothesis has done and are thus of necessity still immature. Nevertheless, the model proposed is consistent with many of the primary geological and geophysical characteristics at Iceland, and we offer them in the hope of stimulating criticism and further scientific testing.

Some aspects of Iceland and the north Atlantic volcanic province that are not predicted by the classical plume hypothesis are as follows:

(1) At the time of breakup of the Laurasian supercontinent (Fig. 1), Tertiary volcanism occurred in localized areas, e.g., in west and east Greenland and northern Britain (Chalmers et al., 1995). This has been attributed to lateral flow for up to 1000 km from a plume that impinged on the base of the lithosphere beneath Greenland (e.g., see Vink, 1984). Long-range lateral flow in plume heads has been often hypothesised and numerically modeled (e.g., see Ribe and Delattre, 1998), but in the north Atlantic, it is apparently required to have been restricted to a few narrowly confined directions only. Following continental breakup, melt is required to have been channeled only to a small restricted section of the mid-Atlantic ridge (MAR) at the latitude of Iceland (Fig. 2; Vink, 1984).

(2) For a mantle plume currently beneath Iceland to have been fixed relative to other Atlantic and Indian ocean hotspots, it is predicted to have migrated southeastwards at a rate of ~2 cm/a relative to the North American plate, from an original location beneath central Greenland at ~60 Ma (Lawver and Muller, 1994). No corresponding age progression of volcanism across Greenland or diachronous volcanism on the sea floor west of Iceland is associated with such a trajectory. On the contrary, since the opening of the north Atlantic at ~54 Ma, the locus of melt extraction has been centered at the MAR. This is clearly indicated by the symmetric bands of thickened crust that lie to the west and east of the MAR (the Greenland–Iceland and Iceland–Faeroes ridges) across which magnetic isochrons from the ocean basins to the north and south are continuous (Lundin and Doré, in press).

(3) An alternative to a plume fixed relative to other Atlantic and Indian hotspots is one that instead

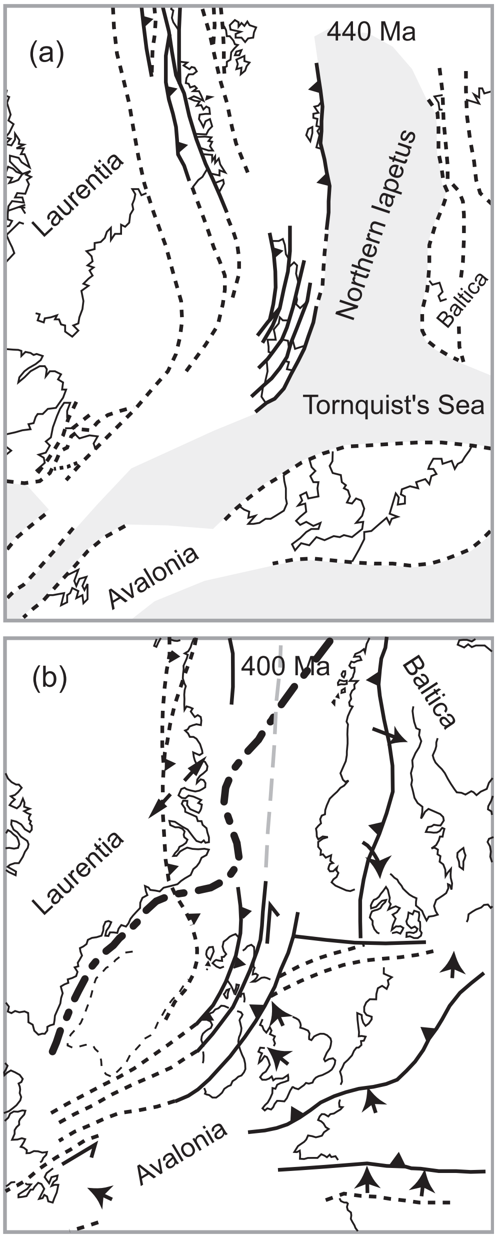


Fig. 1. Caledonian collision zone associated with the closure of the Iapetus ocean at (a) 440 and (b) 400 Ma by convergence of Laurentia, Baltica, and Avalonia. Arrows—convergence directions; thick lines—faults and orogenic fronts. Black triangles indicate sense of thrust faults. Gray dashed line—inferred position of the Caledonian suture. Slabs were subducted beneath Greenland, Baltica, and Britain. Bold dashed line indicates position of MAR that formed at ~54 Ma (after Soper et al., 1992; Skogseid et al., 2000; Roberts, 2003).

migrated northwest by ~500 km during the last ~50 m.y., thus, it has been persistently located beneath the MAR since the opening of the Atlantic (Lundin and Doré, in press). This not only requires the coincidental colocation of a ridge and a plume but a plume that migrates with respect to other hotspots at exactly the right amount to remain on the ridge.

(4) Iceland lies where the MAR crosses the western frontal thrust of the ~400 Ma Caledonian collision zone where this thrust passes from Greenland to Britain. This is a coincidence in the plume hypothesis.

(5) There is little evidence in the Iceland region for plume-like temperature anomalies, which are typically expected to be at least 200 K (see Vinnik et al., in press for a review). Seismic anomalies in the mantle indicate the presence of partial melt, and if this is taken into account, the temperature anomalies required by tomography images are much less than 200 K and may be zero (Wolfe et al., 1997; Foulger et al., 2000, 2001; Vinnik et al., in press). Olivine-glass thermometry, the absence of picritic glass, high-MgO glass geothermometry and major element systematics of Icelandic MORB suggest temperatures little different from those on normal ridges (Breddam, 2002; Gudfinnsson et al., 2003; Presnall and Gudfinnsson, in press). Bathymetric modeling, the subsidence of oceanic crust, and uplift of the Hebrides shelf are consistent with moderate temperature anomalies of 50–100 K only (Ribe et al., 1995; Clift, 1997; Clift et al., 1998). Heat flow, although a weak indicator of mantle temperature, shows no anomaly west of the Reykjanes ridge but a small positive anomaly on the European plate to the east (Stein and Stein, 2003; DeLaughter et al., in press). This is the opposite of what is predicted for an eastward-migrating plume.

The presence of picrites erupted at the time of continental breakup in west and east Greenland is the strongest evidence available for elevated temperature associated with the North Atlantic volcanic province (Larsen and Pedersen, 2000). However, this may be due to initial breakup drawing magma from much greater depths than later processes. In any case, these observations suggest that the highest temperatures were associated with the earliest manifestations of volcanism, the reverse of what is predicted by plume models that suggest that early cool plume-head temperatures are followed by eruption of hotter plume-stem magmas (see also Tegner et al., 1998).

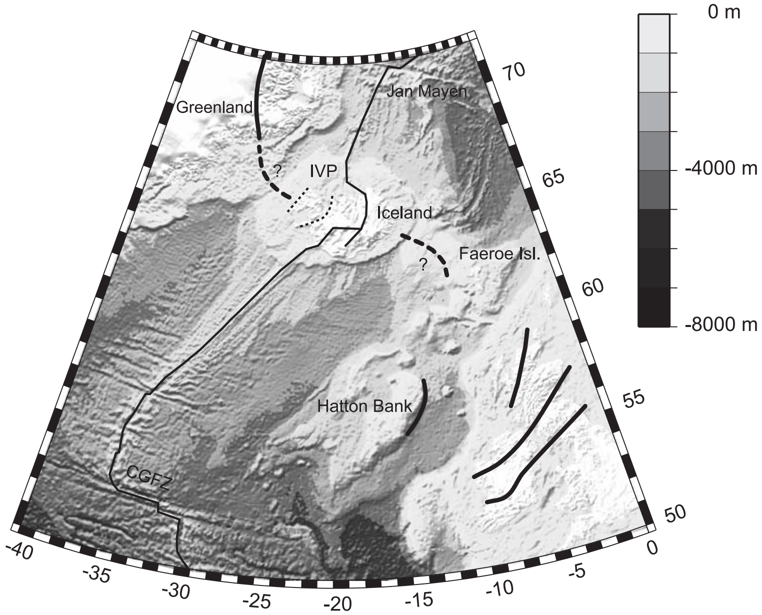


Fig. 2. Present-day bathymetry of the north Atlantic showing the Greenland–Iceland–Faeroe bathymetric ridge that is underlain by crust ~30 km thick. Other shallow areas are blocks of stretched continental crust. Thin black line—MAR; thin dashed black lines—extinct ridges; thick lines—faults of the Caledonian collision zone (Soper et al., 1992); thick dashed line—inferred trend of the western frontal thrust of the Caledonian collision zone crossing the Atlantic Ocean (Bott, 1987), IVP—Iceland volcanic plateau, CGFZ—Charlie Gibbs fracture zone.

(6) Seismic tomography finds no evidence that is repeatable between independent studies for the deep downward continuous low-wave-speed anomaly beneath Iceland that is expected for a thermal plume (see Foulger et al., 2001 for a detailed review). Both teleseismic tomography and whole-mantle tomography show that the strong low-wave-speed anomaly observed in the shallow mantle there does not extend deeper than the mantle transition zone (Ritsema et al., 1999; Foulger et al., 2000, 2001; Montelli et al., 2004). The bottoming of the anomaly in the transition zone is not an artifact of depth-dependent reductions in resolution or sensitivity of seismic wave speed to temperature since both are continuous at transition zone depths. Serious degradation in detection ability only begins at mid-lower mantle depths (Karato, 1993; Ritsema and Allen, 2003).

An early study that reported a low-wave-speed anomaly traversing the entire mantle and interpreted it as an Icelandic plume (Bijwaard and Spakman, 1999) lacks the required resolution in the lower mantle beneath the north Atlantic and imaged similar anomalies beneath Canada and Scandinavia where plumes are not proposed (Foulger et al., 2001). Later studies failed to confirm this deep anomaly (Ritsema et al., 1999; Megnin and Romanowicz, 2000; Montelli et al., 2004). All studies agree, however, that the upper mantle anomaly is an order of magnitude stronger than any anomaly in the lower mantle beneath the Iceland region.

The strong upper mantle seismic anomaly imaged by all tomography studies could be explained by a suite of physical models ranging from a temperature anomaly of ~200 K in the absence of partial melt to up to a maximum of ~0.5–1% partial melt in the absence of a temperature anomaly. Regardless of its interpretation, however, it is clear that repeatable evidence for downward continuation of the anomaly into the lower mantle, a primary requirement of the classical plume model, is lacking.

A holistic appraisal of the evidence available from the Iceland melt anomaly suggests that, if the assumption of a plume is dropped, an upper mantle, moderate-temperature model is natural, reasonable, and not without support. The primary observations that require explanation are the local production of up to three times the amount of melt produced on the Reykjanes and Kolbeinsey ridges to the north and south (Foulger et al., 2003), ocean island basalt (OIB) type geochemistry in some basalts (Schilling et al., 1983; Fitton et al., 2003), local tectonic complexity (Foulger, 2003), and a seismic anomaly that extends throughout the upper mantle (Ritsema et al., 1999; Foulger et al., 2001). The colocation of the Greenland–Iceland–Faeroe ridge with the line of the western frontal thrust of the Caledonian suture suggests a causative relationship. In this paper, we explore the possibility that the volcanic anomaly results from remelting recycled subducted Caledonian oceanic crust locally at the MAR. Experimental studies of petrological phase and melting relations suggest that it cannot currently be ruled out that such a process can generate the exceptionally large volume of melt observed at relatively normal mantle temperatures. Many aspects of the petrology and geochemistry of Icelandic lavas are consistent with a source in recycled subducted oceanic crust (Chauvel and Hemond, 2000), and further modeling of that is presented in a companion paper (Foulger et al., 2004). Local tectonic complexity has characterized the Iceland region since the initial opening of the north Atlantic (Bott, 1985) and may have been encouraged by both inherited regional heterogeneous stress related to the Caledonian suture and local effects caused by the very thick crust. The upper mantle anomaly is clearly an extensive feature that characterizes much of the Atlantic north of the Charlie Gibbs fracture zone and is not confined to Iceland alone.

This model is supported by other work which argues for a compositional rather than a thermal origin for "hotspots" (e.g., see Green et al., 2001; Presnall, 2003). It reverses the traditional cause-and-effect chain of reasoning and considers the magmatic anomaly in the Iceland region to be the result of plate tectonic processes–continental breakup, mantle inhomogeneity, and complex, unstable tectonics and not the cause of these features via an independent high-temperature deep-mantle-plume mode of convection.

2. Tectonic evolution of the North Atlantic

The north Atlantic formed in a region with a complex geological history. Prior to ~500 Ma, the continents of Laurentia and Baltica were separated by the Iapetus ocean, and these were separated from the continent of Avalonia to the south by Tornquist's Sea (Fig. 1a; e.g., see Soper et al., 1992). Between ~420 and ~410 Ma, the last stages of closure consumed these oceans and formed the Caledonian suture, which was associated with a broad zone of compression with branches in east Greenland, Scandinavia, Britain, Europe, and Newfoundland (Fig. 1b). The westernmost frontal thrust ran down what is now the east coast of Greenland to merge with a complex array of northerly trending left-lateral shear zones that now dissects Britain. Subduction beneath this frontal thrust was to the southwest (e.g., see Barker and Gayer, 1984).

The colliding cratons are thought to have been ~150–200 km thick (Polet and Anderson, 1995), and at least this length of latest subducting slab may thus have been trapped at a high angle in the lithosphere beneath the collision belt. The latest crust to subduct may have been relatively young. Such crust is buoyant because it has had little time to cool. Crust younger than ~50 m.y. is predicted to become neutrally buoyant at shallow depth in the mantle (Oxburgh and Parmentier, 1977) despite the density increase that accompanies transformation to eclogite at ~50 km depth (Oxburgh and Parmentier, 1977). Part of the mantle lithosphere beneath the sutured supercontinent may thus have been refertilized (that is, its ability to produce basalt increased; Yaxley and Green, 1998) by the addition of trapped, eclogitized subducted oceanic crust. The thickness of trapped crust potentially underlying any point on the surface (and thus, the volume of magma potentially extractable if it remelts) would depend on the dip of the slab. The vertical thickness of crust in an inclined slab is twice that of the original horizontal section if the slab dips at 60°, and a dip of 70° triples the thickness. Such angles are within the range of dips of modern subducting slabs. Field evidence for abundant eclogites in the Caledonian collision zone can be found, for example, in exhumed rocks in the Western Gneiss region in Norway, where unroofed eclogites are exposed. Such observations suggest that crust subducting latest during continental collision may not be completely lost through subduction, eclogitization, and sinking through the asthenospheric mantle.

There are few constraints on the form in which trapped eclogite might be retained beneath a collision zone. In the ~350 m.y. between the final collision and the opening of the north Atlantic, refertilization of an extensive volume of lithosphere might have occurred through progressive remelting, ascent, and resorbtion into the mantle and lower crustal parts of the continental lithosphere. Conversely, some may have been retained in an essentially little changed state.

The cratons separated once again when the MAR formed at ~54 Ma. To the north, the new ridge formed along the collision zone, but to the south, it formed a new split within the Laurasian supercontinent (Fig. 1b). At the boundary between these two contrasting regions, the new MAR crossed the east Greenland–Britain frontal thrust. This crossing was subsequently the locus of chronic high melt productivity where the Greenland–Iceland–Faeroe ridge of thick crust formed (Fig. 2; e.g., see Bott, 1985).

That ridges migrate laterally with respect to the mantle beneath is evidenced by the growth and shrinking of the various plates on Earth's surface, which requires ridges to move with respect to one other. The most plausible reference frame fixed to the deeper mantle might be Antarctica fixed, since the Antarctic plate has almost no subduction zone along its boundary. Relative to Antarctica (and thus plausibly relative to the deeper mantle), the MAR in the north Atlantic migrates west at ~1 cm/a. In this reference frame, in the ~54 m.y. since the north Atlantic opened, the MAR would have migrated west by ~540 km. Because the Greenland–Britain frontal thrust was approximately orthogonal to the new MAR, this lateral migration may not have transported the spreading ridge away from the associated fertility zone early on, as occurred to the north and south where the MAR formed along the collision zone. Instead, this part of the MAR may have migrated along the strike of a zone of mantle refertilized by subducted Caledonian crust. This model suggests that the ongoing tapping of a compositional fertility heterogeneity may explain the steady-state production, in the Iceland region, of igneous crust that is up to ~30 km thick, three times thicker than along the Kolbeinsey and Reykjanes ridges to the north and south (Foulger et al., 2003).

How might subducted slab material trapped during suturing be recycled back into the asthenosphere and tapped at a spreading ridge? It is difficult to see how slabs subducted deep into the asthenosphere at the time of continental collision could be tapped at the latitude of present-day Iceland because Laurasia migrated north by ~60° subsequent to the closure of the Caledonian suture. Much of the Caledonian slab material subducted into the asthenosphere may thus now be distant from the surface location of the suture zone. The source of recycled crust might then originate mostly from the Laurasian continental lithosphere. Catastrophic delamination has been suggested as a process that may recycle continental mantle lithosphere into the asthenosphere and trigger the eruption of flood basalts (Tanton and Hager, 2000). The pattern of volcanism in the Iceland region is one of continued enhanced activity and suggests that erosion of the subcontinental lithosphere and recycling back into the asthenosphere might be a continuous process that accompanies westward migration of Greenland and the MAR behind it.

There appears to be little mixing of the source of Icelandic basalts with the regional north Atlantic asthenosphere. Fitton et al. (2003) argue from trace element and radiogenic isotope data that the source of Icelandic basalts does not contain a component of ambient north Atlantic MORB source. Furthermore, much of the petrology of Icelandic rocks can be explained by remelting various parts of a complete section of oceanic crust (Chauvel and Hemond, 2000; Natland, 2003; Foulger et al., 2004). These observations suggest that whatever the mechanism of supply of source material to the melting zone beneath Iceland, it does not involve efficient homogenization with the regional mantle.

Supporting evidence for recycled crust of Caledonian age in Icelandic basalts comes from the calculated compositions of parental melts, trace and rare earth elements (REE) and radiogenic isotope ratios (Chauvel and Hemond, 2000; Korenaga and Kelemen, 2000; Breddam, 2002; McKenzie et al., 2004). Subducted slabs include sediments, altered basaltic upper crust, gabbroic lower crust, and depleted lithospheric mantle. Continental debris may also be included because of the proximity of the latest subducting slab to the colliding cratons. The crustal part may include enriched mid-ocean ridge basalts (E-MORB) and alkalic olivine basalt such as those occurring on and near spreading ridges today. This variety of source material combined with processes associated with remelting the Icelandic crust (Oskarsson et al., 1982) is likely to be able to account for the petrological and geochemical variability of Icelandic basalts, in addition to their exceptionally large volume (Foulger et al., 2004).

Compositional heterogeneity is the preferred explanation for the large melt volumes in the absence of very high temperatures (Ribe et al., 1995; Korenaga and Kelemen, 2000; Von Herzen, 2001; Breddam, 2002; Stein and Stein, 2003). High volatile content can also decrease the melting point and lengthen the melting column although this increases crustal thickness only slightly (Presnall et al., 2002; Asimow and Langmuir, 2003). Volatiles may, however, be able to account for the unusually deep regional low-seismic wave speed anomaly that pervades the mantle beneath the north Atlantic.

3. Present regional setting

Today, a major part of the north Atlantic ocean is characterized by global-scale bathymetry and gravity anomalies. The geoid anomaly is some 60 m in amplitude, extends from ~40–80°N, and encompasses the whole breadth of the north Atlantic and major parts of Europe and north Africa (Marquart, 1991; Lundin and Dore, in press). The bathymetric anomaly occupies the entire ~1000-km broad part of the Atlantic ocean between the Charlie Gibbs and Jan Mayen fracture zones and shallows the ocean floor by 2–3 km compared with normal depths for oceanic crust of equivalent age. This region is underlain by a low-wave-speed seismic tomography anomaly that fills the upper mantle (Ritsema et al., 1999; Megnin and Romanowicz, 2000). The bathymetric and seismic anomalies are consistent with a regional temperature anomaly of up to 50–100 °C (e.g., see Ribe et al., 1995; Foulger et al., 2001). Such an anomaly is arguably within the normal range of variation for the mantle and might have resulted from continental insulation by the Laurasian supercontinent prior to the opening of the north Atlantic (e.g., see Klein and Langmuir, 1987; Anderson, 1994b).

The topographic and seismic anomalies, although not the geoid high, culminate at Iceland, which is also the center of a geochemical anomaly which extends south down the Reykjanes ridge as far as ~62.5°N. Variable rates of magmatism on the Reykjanes ridge have formed southward-propagating, diachronous V-shaped topographic ridges on its flanks (Vogt, 1971). These have been variously attributed to the migration of temperature pulses southward along the ridge or correlated with rift reorganizations in Iceland (Hardarson et al., 1997). To the north of Iceland, discontinuities in structure and geochemistry occur across the 120-km-long Tjornes fracture zone, and diachronous topographic ridges are present but less clear than on the Reykjanes ridge.

Iceland itself is a ~350×500 km subaerial portion of the aseismic Greenland–Iceland–Faeroe ridge. Most of the currently submarine part of this ridge is thought to have been originally emplaced subaerially and subsequently cooled and subsided below sea level (Nilsen, 1978). It is not clear to what extent the subaerial nature of Iceland itself is responsible for the many anomalies that appear to peak there. Seismic experiments and petrological sampling can be conducted in Iceland with much greater facility than is possible on submarine parts of the MAR, crustal accretion processes are modified (Palmason, 1980), petrological diversity is increased (Oskarsson et al., 1982), isotopic signatures are modified (Taylor, 1968), and glaciations influence the geology radically (e.g., see Hardarson and Fitton, 1991). It is possible that a major global-scale anomaly is responsible for the shallowing of the sea floor of the north Atlantic and the consequential stand above sea level of this portion of the MAR, and that this makes comparisons of many geological and geophysical data with those from the adjoining submarine ridge portions misleading.

4. The volumetric melt anomaly

How much excess melt is there in the Iceland region? The seismic crustal thickness beneath the Greenland–Iceland–Faeroe ridge is typically ~30±5 km, contrasting sharply with the ~10 km seismic thickness of north Atlantic oceanic crust elsewhere (Foulger et al., 2003; Fig. 3a). Tectonic complexities have resulted in local variations in crustal thickness beneath Iceland itself, including somewhat thinner crust beneath western Iceland than eastern Iceland and a block up to ~40 km thick beneath central Iceland (Fig. 3b). The complex tectonic history of Iceland requires the capture of a substantial volume of old oceanic and/or continental crust, which may account for the local thickening of the seismic crust from ~30 to ~40 km beneath central Iceland (Sections 5 and 7; Amundsen et al., 2002; Foulger, submitted for publication). This suggests that continuous processes at the spreading zone in Iceland currently produce a seismic crust ~30 km thick.

Crustal thickness determined seismically is often used as a proxy for the amount of melt formed at mid-ocean ridges although retention of melt in the uppermost mantle may render it a slight underestimate (Cannat, 1996). In the case of the Icelandic crust, an additional complexity is the mismatch

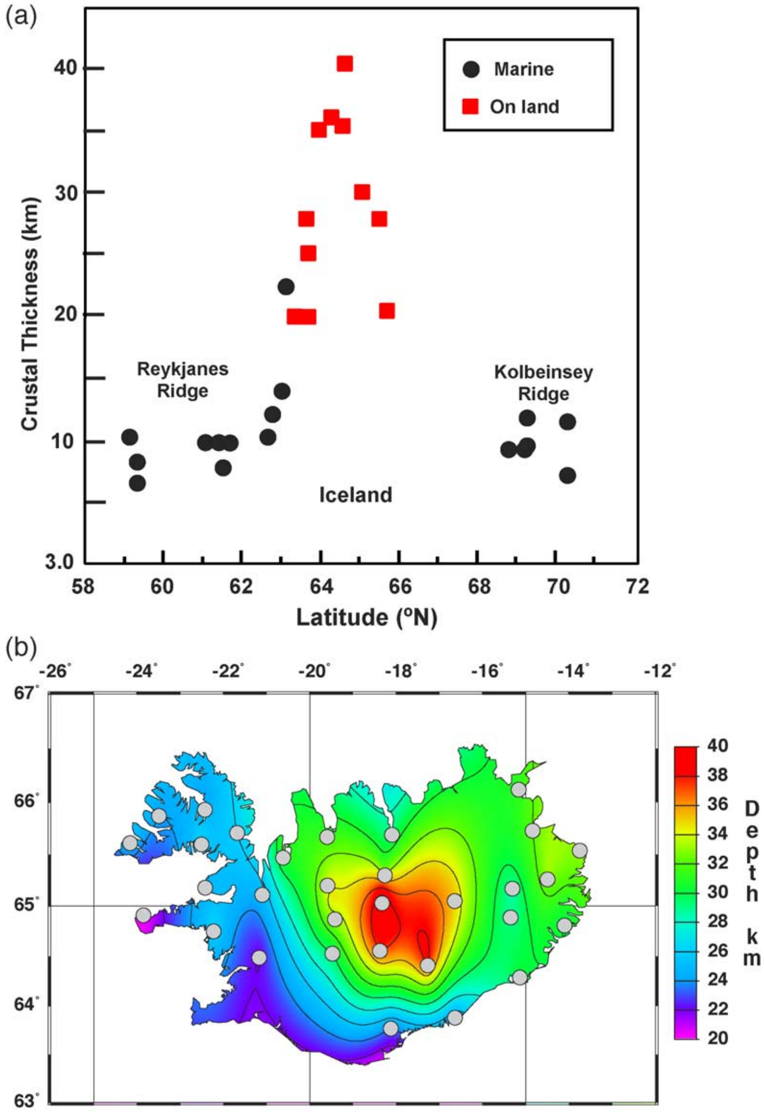


Fig. 3. (a) Crustal thickness vs. latitude from a compilation of seismic experiments in Iceland and the north Atlantic (Whitmarsh, 1975; Bunch and Kennett, 1980; Goldflam et al., 1980; Ritzert and Jacoby, 1985; Whitmarsh and Calvert, 1986; Larsen and Jakobsdottir, 1988; Mutter and Zehnder, 1988; Smallwood et al., 1995; Weigel et al., 1995; Kodaira et al., 1997, 1998; Navin et al., 1998; Smallwood and White, 1998; Du and Foulger, 1999, 2001; Holbrook et al., 2001; Weir et al., 2001; Du et al., 2002; Foulger et al., 2003). (b) Contour map showing the thickness of the crust defined as the depth to the V_s=4.2 km/s horizon from receiver functions (from Foulger et al., 2003). Grey dots show positions of stations where crustal thickness was determined.

between the density contrast expected between oceanic lower crust and mantle (~300 kg/m^3, assuming an olivine gabbro lower crust and peridotite upper mantle) and that indicated by isostasy for average Icelandic lower crust and mantle (~90 kg/m^3; Menke, 1999). On density grounds, it is unsafe to assume that Icelandic-type crust is simply scaled-up oceanic crust with the same petrology.

One possibility is that some of the ~23±5-km-thick Icelandic lower crust may comprise entrained mantle rocks. The small density contrast with the mantle would encourage such entrainment in upwardly mobile melt although the absence of any mantle xenoliths in Iceland and seismic wave speeds for the lower crust that are typical of gabbro provide no supporting evidence (Foulger et al., 2003). Nevertheless, such a model gives a lower bound for the melt thickness. If the entire density-contrast anomaly is explained as the lower crust comprising a mixture of average olivine gabbros and depleted harzburgite, then it follows that only ~30% of the lower crust is olivine gabbro, and only ~50% or ~15 km of the total crust represents melt. If, on the other hand, the anomalously high lower crustal densities result from high-density cumulates, e.g., oxide gabbro (Foulger et al., in press), then the entire lower crust may represent melt. These considerations suggest lower and upper bounds for the maximum melt thickness of ~15 and ~30 km or 1.5–3 times that at the immediately adjacent Reykjanes and Kolbeinsey ridges.

The amount of "active upwelling" required to explain the melting anomaly, assuming a similar peridotite source under both Iceland and the oceanic ridges, has been estimated using a variety of methods (e.g., see Ito et al., 1999; Holbrook et al., 2001), but it is unclear what process could be driving such upwelling in the absence of a major thermal anomaly imparting buoyancy. In the absence of a major temperature anomaly, a more fusible mantle composition is the primary candidate explanation for excess melt. As discussed above, this might be provided by subducted slabs, including eclogitized crust, trapped beneath the Caledonian collision belt and recycled into the melt zone beneath this part of the MAR.

The volume of melt that may be extracted from subducted oceanic crust, peridotite, and crust–peridotite mixtures has been investigated quantitatively by Yaxley (2000), who studied experimentally the melting relations of mixtures of homogeneous fertile peridotite and average oceanic crust. The addition of basalt to peridotite reduces both the solidus and the liquidus (Fig. 4a). This results both in melting beginning at a lower temperature in the basalt–peridotite mixture and in higher melt fractions at a given temperature. Fig. 4b shows that under the experimental conditions, the addition of a few tens of percent of basalt to peridotite can enhance the melt volume by up to a factor of 4. In this way, at normal mid-ocean ridge temperatures, significantly more melt is predicted for a crust–peridotite mixture than for a peridotite mantle. A few tens of percent of basalt in a peridotite mantle might thus account for the great melt thickness formed at Iceland.

Heat-balance questions are raised by such a model, which requires large amounts of heat to produce the required melt (Foulger et al., 2004). This heat must be either advected or conducted into the melt zone, or high temperature is required for which there is scant evidence. Plume models also appeal to the melting of eclogite to account for the volumes of basalt observed at hotspots and large igneous provinces and have difficulties accounting for the source of heat since numerical modelling shows that insufficient melt can be generated in upwelling peridotite at reasonable temperatures (e.g., see Cordery et al., 1997).

5. Tectonic evolution of the Iceland region

A number of local features in Iceland are commonly quoted as being consistent with a plume beneath central Iceland. These include eastward migration of the locus of spreading in Iceland thought to be consistent with an eastward-migrating plume and a maximum in crustal thickness and volcanic activity beneath central Iceland. These features are, however, just two aspects of a much more complex tectonic history than is commonly appreciated.

Following the opening of the north Atlantic at ~54 Ma, the spreading history of the ocean basin that formed within the Caledonian collision zone to the north of present-day Iceland contrasted with that which formed within the Laurentian continent to the south (Fig. 5). The zone where the Greenland–Iceland–Faeroe ridge later formed separated these two contrasting tectonic regions and was persistently

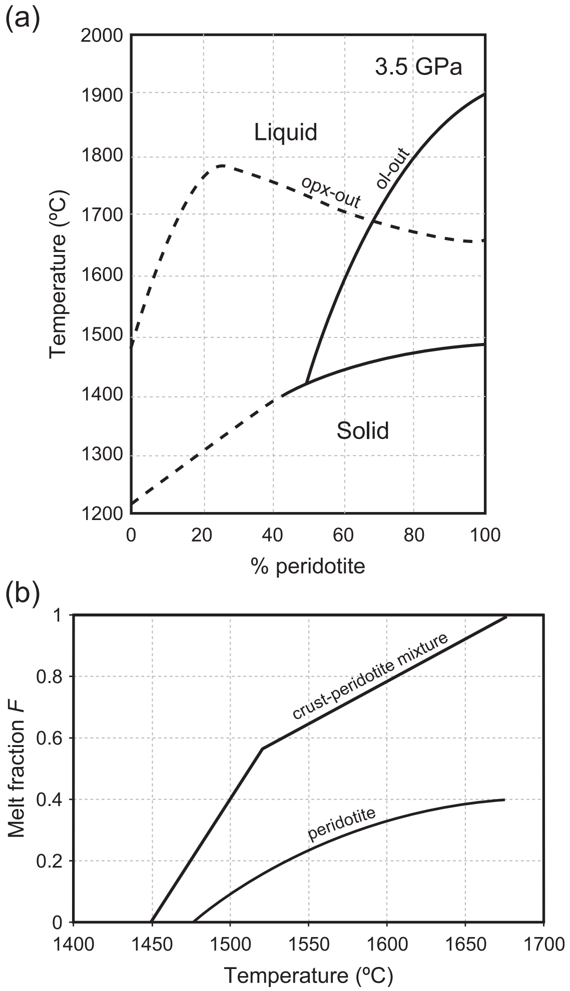


Fig. 4. (a) Solidus and liquidus for fertile peridotite containing varying percentages of average altered oceanic crust. opx—Orthopyroxene, ol—olivine (adapted from Yaxley, 2000). (b) Relationship between melt fraction F and temperature for fertile peridotite and a mixture of 30% average altered oceanic crust and 70% fertile peridotite. The peridotite line is the parameterisation of McKenzie and Bickle (1988) for normal fertile peridotite, and the crust–peridotite line is an approximate estimate for the bulk composition corresponding to the liquidus minimum of panel (a). The higher average dF/dT and lower solidus temperature of the mixture results in enhanced melt productivity at a given temperature (derived from data in Yaxley, 2000).

unstable, featuring paired spreading ridges and intervening microplates.

At the time of opening, a ~100-km-long, right-stepping transform fault, the Faeroe transform fault (FTF), formed where the new spreading ridge crossed the western frontal thrust of the Caledonian collision zone (Bott, 1985; Fig. 5a). Along much of the new margin, rifting caused thick Archean lithosphere and new oceanic crust to be juxtaposed. Vigorous magmatism building a volcanic margin with crust up to ~25 km thick occurred along a ~2000-km-long portion of the new rift including the margins of Greenland (Boutilier and Keen, 1999). This may have resulted from EDGE convection, which is driven by lateral temperature gradients where thick lithosphere meets hotter mantle (Anderson, 1994a, 1995; Boutilier and Keen, 1999; Korenaga and Kelemen, 2000). Increased mantle fertility from Caledonian age subducted crust in the shallow upper mantle may also have enhanced melt production. Initial vigorous magmatism dwindled to more typical oceanic rates along most of the margin after a few m.y. However, the magmatic rate remained persistently high at the latitude of present-day Iceland and built the ridge of crust ~30±5 km thick that now traverses the entire north Atlantic. Tectonic complexities rafted several continental blocks into the ocean, including the Faeroe block and the Jan Mayen microcontinent (JMM), part of which might presently be trapped beneath Iceland (Amundsen et al., 2002).

Spreading proceeded relatively simply for the first ~10 m.y. following opening, but at ~44 Ma, a major reorganization occurred north of the FTF. A second spreading center, the Kolbeinsey ridge, developed within the Greenland craton, renewing volcanism there (Fig. 5b). Complimentary fan-shaped spreading then occurred along both the original Aegir ridge and the Kolbeinsey ridge during the period 44–26 Ma, causing ~30° of counterclockwise rotation of the intervening captured continental JMM (Fig. 5c). This resulted in up to ~60 km of transtensional opening across the FTF, which corresponds to opening at up to ~15% of the local full-spreading rate of 1.9 cm/a (DeMets et al., 1994). The onset of this phase of extension coincides with a time when the magmatic production rate increased greatly, and formation of the ~250-km-wide Iceland–Faeroe ridge of thickened crust gave way to formation of the Iceland volcanic plateau, which is up to ~500 km in north–south extent (Fig. 2). This increased magmatism may have been permitted by the transtensional opening across the FTF.

The region continued to be characterized by relatively simple spreading about the Reykjanes ridge

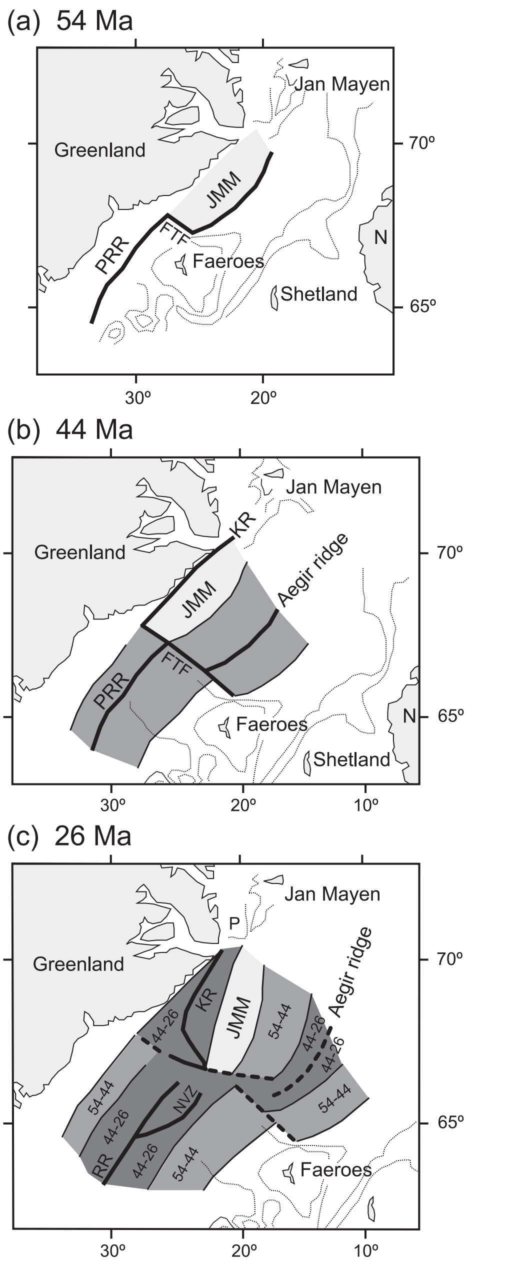
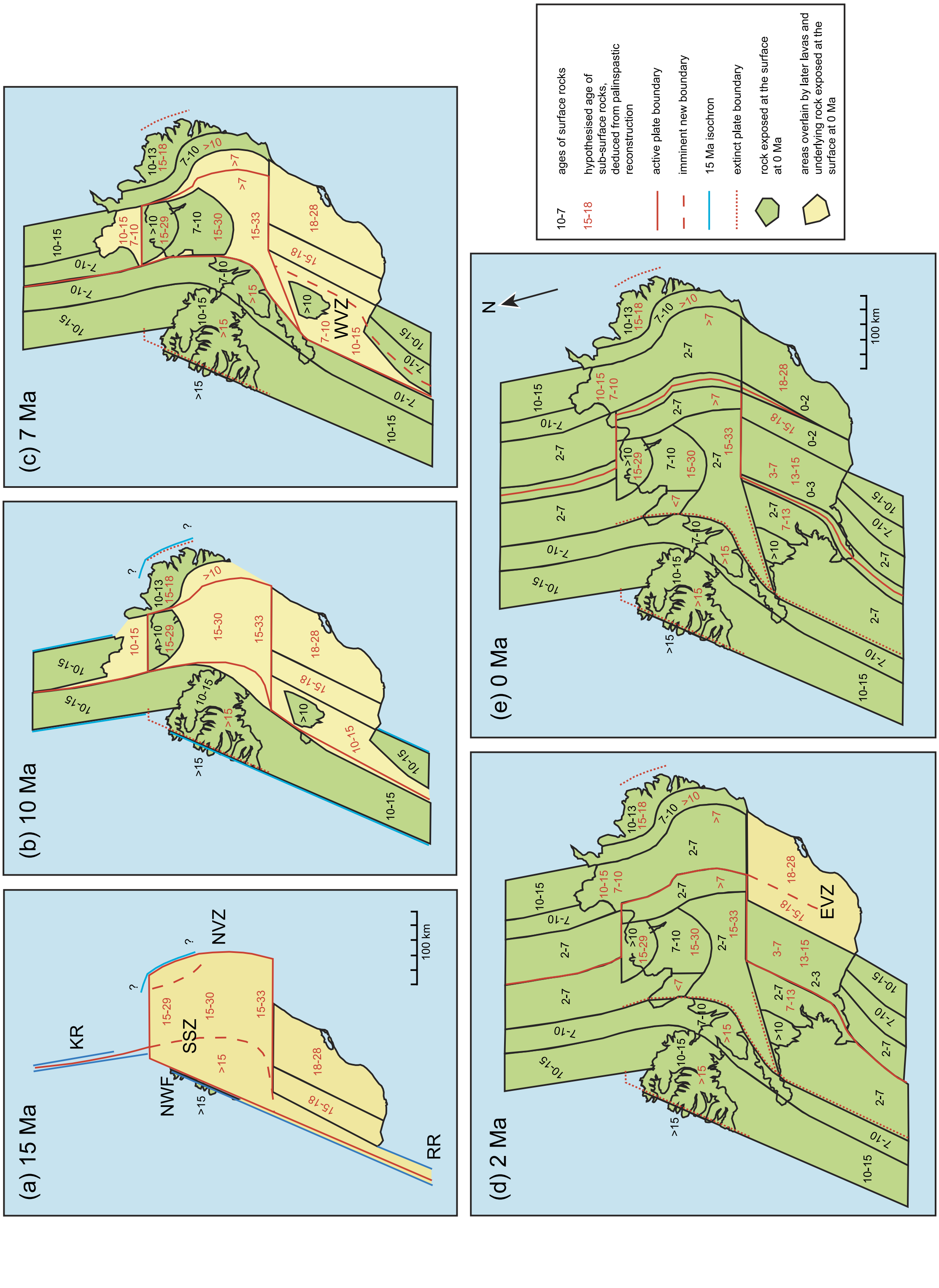


Fig. 5. Tectonic evolution of the Iceland region at 54, 44, and 26 Ma in Mercator projection (from Bott, 1985). Light gray—continental crust, mid gray—oceanic crust 54–44 Ma, dark gray—oceanic crust 44–26 Ma, heavy solid lines—active plate boundaries, heavy dashed lines—extinct plate boundaries, thin lines—bathymetric contours, FTF—Faeroe transform fault, KR, PRR, RR—Kolbeinsey, proto-Reykjanes, Reykjanes ridges, NVZ—Northern Volcanic Zone, JMM—Jan Mayen microcontinent, N—Norway, P—pole of rotation of JMM 44—26 Ma, numbers—sea floor age in Ma.

to the south. In the neighborhood of the developing Greenland–Faeroe ridge, however, spreading proceeded about a parallel pair of centers that progressively migrated south through the repeated extinction of old rifts and the opening of new. At ~26 Ma, the Aegir ridge became extinct, spreading became confined to the Kolbeinsey ridge, and a second parallel spreading center formed to the immediate south (Fig. 5c). A block of crust that may have been all oceanic or may have included a fragment of the JMM was captured between the new ridge pair (Foulger, submitted for publication). The eastern center later developed into the currently active Northern Volcanic Zone (NVZ; c.f. Figs. 5c and 6a). Continuity of the lava succession now exposed in eastern Iceland shows that the eastern center has been long lived. Apart from two possible minor westward migrations, it has remained approximately fixed with respect to the Kolbeinsey ridge since at least 15 Ma and perhaps since ~26 Ma (Saemundsson, 1979; Bott, 1985; Jancin et al., 1985).

As a consequence of the fixed spatial relationship of the eastern spreading center with the Kolbeinsey ridge, the western center was progressively transported west relative to the Kolbeinsey ridge with the north American plate. This center experienced at least two extinctions accompanied by the opening of new rifts further east that were more colinear with the Kolbeinsey ridge. At ~15 Ma, the western rift became extinct, and a new rift opened ~80 km further east (Fig. 6a). The extinct rift currently lies offshore of NW Iceland, and its products are recognised below an unconformity in the extreme northwest. At ~7 Ma, the new rift in turn became largely extinct, and the presently active Western Volcanic Zone (WVZ) formed (Fig. 6c). Extension across a pair of parallel spreading centers has occurred in south Iceland since ~2 Ma when the Eastern Volcanic Zone (EVZ) formed (Saemundsson, 1979) (Fig. 6d).

The continually evolving parallel-pair spreading center configuration has created a jigsaw of ephemeral microplates. These include the JMM (Fig. 5) and two microplates in Iceland. One lies between the northern pair of parallel spreading centers (the Trollaskagi microplate), and one lies between the southern pair (the Hreppar microplate; Figs. 6 and 7). The palinspastic reconstruction (Fig. 6) shows that crust older than that exposed at the surface has been captured beneath central and southeast Iceland as a


(a) 15 Ma
(b) 10 Ma
(c) 7 Ma
(d) 2 Ma
(e) 0 Ma
KR
NWF
SSZ
NVZ
RR
WVZ
EVZ
N
100 km
10-7 ages of surface rocks
15-18 hypothesised age of sub-surface rocks, deduced from palinspastic reconstruction
active plate boundary
imminent new boundary
15 Ma isochron
extinct plate boundary
rock exposed at the surface at 0 Ma
areas overlain by later lavas and underlying rock exposed at the surface at 0 Ma

consequence of the complex history of spreading. If entirely oceanic, this captured crust must be up to ~30 m.y. old (Foulger, submitted for publication). Recent ancient ages determined for zircons from basalts in southeast Iceland suggest the presence of continental crust beneath that area, suggesting that the JMM may have continued further south than shown in Fig. 5c (Amundsen et al., 2002). If so, a sliver of continental crust might have been captured between the parallel pair of spreading centers that formed at ~26 Ma and currently underlie central and southeastern Iceland. The presence of such a sliver may have influenced the formation of the parallel spreading center pair.

The palinspastic reconstruction of the evolution of Iceland shown in Fig. 6 is the simplest possible interpretation of the data available. It is based on the ages of surface lavas and the assumption that spreading in Iceland is shared between the parallel ridges but the total rate is the same as that along the adjacent oceanic ridges. Lavas that flowed subaerially may have flowed tens of kilometers from their eruption site. No attempt has been made to correct for this error source, which would tend to result in errors in the deduced locations of spreading centers in the past. The width and age of the old, captured crust is independent of this, however, and is dependent only on distance in the spreading direction. The maximum width of Iceland in the spreading direction is ~550 km, compared with the 285 km that corresponds to 15 m.y. of extension at the 1.9 cm/a local average spreading rate. The conclusion that an expanse of some 265 km of older, captured crust underlies Iceland is thus inescapable.

6. Local variations in spreading direction

The tectonic disequilibrium in the Iceland region throughout its history must have been accompanied by local minor variations in the direction of plate motion associated with the propagating spreading axes and intervening microplates. Four sets of observations suggest that the local spreading direction rotates clockwise from north to south (Fig. 7):

(1) The trend of dikes in the NVZ is ~10°N suggesting a local spreading direction in north Iceland of ~100°, close to the global plate direction of ~105° (DeMets et al., 1994). In south Iceland, dikes in the EVZ and WVZ trend at ~45° and ~35°, respectively, suggesting clockwise rotation of the spreading direction by 25–35°.

(2) GPS measurements show that the direction of motion at the east coast rotates clockwise by ~45° from northeast to southeast Iceland (Hofton and Foulger, 1996; Voelksen, 2000).

(3) The tension axes of focal mechanisms of large strike–slip earthquakes in the fracture zones in north and south Iceland are orientated ~60° more southerly in south Iceland than in north Iceland (Bjarnason and Einarsson, 1991; Einarsson, 1991).

(4) Plio-Pleistocene and Tertiary dikes in north and south Iceland trend typically at 0–10°and 40–50°, respectively, suggesting that a ~30–50° difference in local direction of motion has persisted since at least ~15 Ma (Saemundsson, 1979).

Long-term large-scale motion between the North American and Eurasian plates is coherent, and the variations in direction of motion are apparently local to the Iceland region only. Such local variations in the direction of surface motion may be possible in the context of large-scale coherent plate motion through rapid post-tectonic stress redistribution in the low-viscosity Icelandic lower crust (Foulger et al., 1992; Hofton and Foulger, 1996).

A more southerly direction of motion in south Iceland predicts transtension across a zone trending west–east through central Iceland. Such transtension

Fig. 6. Tectonic evolution of Iceland at 15, 10, 7, 2, and 0 Ma. Black lines indicate the boundaries of modeled blocks. The oldest rocks exposed in Iceland are 17 m.y. old. Blue—unmodeled oceanic areas, yellow—areas that are part of present-day Iceland but are now covered with later lavas, green—rock currently exposed at the surface, solid, dashed, and dotted red lines—active, imminent, and extinct plate boundaries, blue lines—inferred position of the 15 Ma isochron, red and black numbers—ages of rock in m.y., red—ages of rocks currently covered with younger lavas, black—ages of surface rocks. Where ages are shown in both black and red, older rock underlies younger surface rock. The extent of rock in a given age range offshore is inferred assuming a spreading rate of 1.9 cm/a about the KR and the RR. On land, the possible age range of rocks in a given block is deduced assuming spreading is equally distributed between parallel spreading center pairs where these exist. The ages of rocks on land are taken from the literature (Saemundsson, 1979; McDougall et al., 1984; Hardarson et al., 1997; Duncan and Helgason, 1998; Udagawa et al., 1999; Helgason and Duncan, 2001). The positions of the extinct Northwest Fjords (NWF) and Snaefellsnes–Skagi zone (SSZ) rifts are taken from Saemundsson (1979). Other extinct rifts are required by space considerations given the ages of surface rocks.

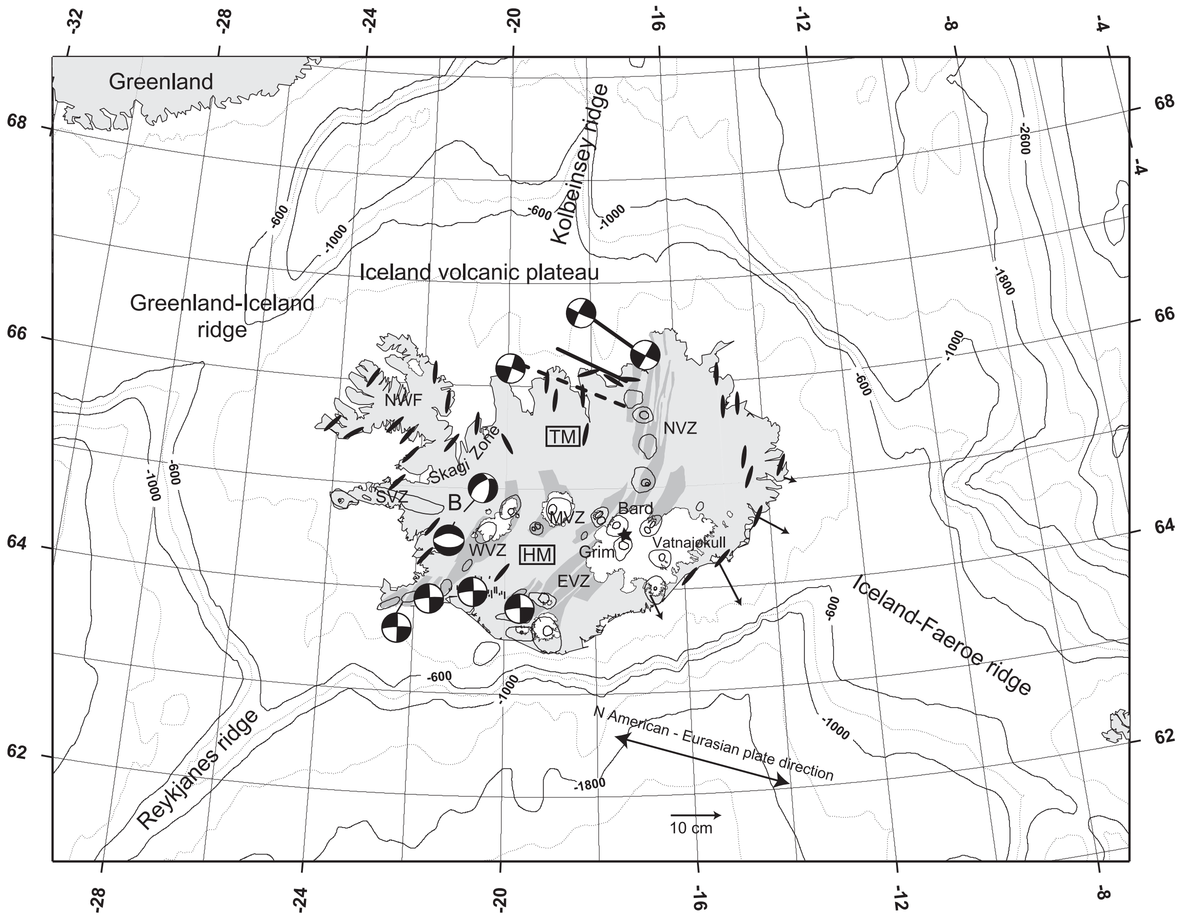


Fig. 7. Bathymetry of the Greenland–Iceland–Faeroe ridge, along with present-day tectonics of Iceland, in general stereographic projection. SVZ, WVZ, EVZ, MVZ, NVZ—Snaefellsnes, Western, Eastern, Middle, and Northern Volcanic Zones, NWF—Northwest Fjords area. TM, HM—the northern (Trollaskagi) and southern (Hreppar) microplates, B—Borgarfjordur area. Thick lines in north—faults of the Tjornes Fracture Zone (TFZ), short thin lines in south—faults of the South Iceland Seismic Zone, dark grey zones—fissure/dike swarms of presently active spreading segments, white—icecaps, black outlines—active central volcanoes/calderas, black elongate ellipses—Plio-Pleistocene and Tertiary dike swarms (Saemundsson, 1979), arrows—motion relative to points in the NVZ measured using GPS 1987–1992 (Hofton and Foulger, 1996). Representative focal mechanisms of earthquakes are shown in lower hemisphere stereographic projections with black compressional quadrants (Bjarnason and Einarsson, 1991; Einarsson, 1991). The long-lived composite transtensional zone that traverses central Iceland today comprises the SVZ, the Borgarfjordur area of normal-faulting earthquakes in west Iceland, the MVZ, and the confluence of this with the NVZ and EVZ in northwest Vatnajokull. Bard—Bardarbunga, Grim—Grimsvotn, star—Gjalp. Volcanism and the size and number of major central volcanoes increase to the east. Volcanism dwindles beneath eastern Vatnajokull and is extinct at the southeast coast of Iceland (Jóhannesson and Saemundsson, 1998).

encourages normal faulting, diking, and volcanism. Iceland is traversed from west to east by a zone of volcanism and normal-faulting earthquakes (Fig. 7). From west to east, this zone includes the Snaefellsnes Volcanic Zone (SVZ), a reactivated older rift with a low level of volcanism, and the Borgarfjordur area immediately east of this, where normal-faulting earthquakes, some with easterly orientated fault planes, have occurred (Einarsson et al., 1977). Volcanic production increases eastward along the Middle Volcanic Zone (MVZ) and culminates at a cluster of major volcanoes beneath the Vatnajokull icecap that includes Bardarbunga, Grimsvotn, and the recently active Gjalp volcano (Gudmundsson et al., 1997). This area is a triple junction formed by the confluence of the NVZ, EVZ, and the MVZ and represents the topographic and volcanic culmination of Iceland (Foulger et al., 2003). Beyond it, volcanism dwindles rapidly beneath eastern Vatnajokull. Extinct Tertiary volcanoes lie in the southeast coastal zone (Saemundsson, 1979; Jóhannesson and Saemundsson, 1998) and show that major magmatism, suggesting local dilation, has been a long-term characteristic of the region presently beneath northwest Vatnajokull but does not extend far into the Eurasian plate.

Both the observed local variation in the direction of crustal motion and the nature of the east–west Snaefellsnes–Vatnajokull zone are consistent with transtension increasing from west to east and culminating beneath the MVZ and northwestern Vatnajokull. The palinspastic reconstruction suggests that this zone may be the modern expression of a complex of colinear microplate boundary elements that has developed since ~26 Ma. It may include reactivated older spreading centers, transform faults, and fracture zones (c.f. Figs. 6 and 7). It is orientated at ~95°, which is ~10° more northerly than the plate direction. This suggests that up to a few kilometers of north–south opening may have occurred at its easternmost end beneath northwest Vatnajokull since the development of the EVZ at ~2 Ma. Such a model can explain the increase in volcanism from west to east across central Iceland and its culmination beneath northwest Vatnajokull. The orientations of Plio-Pleistocene and Tertiary dikes indicate that total microplate rotations in Iceland have been small, probably <~10°. Major seismic activity in south Iceland and volcanic activity at the southern end of the EVZ (Saemundsson, 1979; Einarsson, 1991) suggest that a new, easterly orientated transtensional shear zone may be currently developing along the southern edge of the southern (Hreppar) microplate.

7. Discussion

Interpretations of data from the Icelandic volcanic province usually assume that the source of excess melt is localized beneath central Iceland, specifically northwest Vatnajokull, and that melt flows laterally from there, perhaps even as far as the Charlie Gibbs fracture zone, to form the north Atlantic bathymetric swell. This simple model does not account well for many observations, however, and the extensive regional-scale anomalies may well result from dispersed structure and processes rather than radial flow from a localized source beneath Iceland.

Crustal thickness beneath the entire north Atlantic ocean is ~10 km, somewhat greater than the ~7 km global average (Mutter and Mutter, 1993; Fig. 3a). The whole area has been influenced by the Caledonian orogenic belt, which extends from northern Scandinavia and Greenland to Newfoundland and central France (Fig. 1). Several subduction zones, some of which experienced reversals in polarity, were involved (Dewey and Shackleton, 1984), and late stage subducted slabs may have refertilized much of the upper mantle throughout the region.

A model that attributes the local volumetric melt anomaly in the Iceland region to remelting of trapped Caledonian oceanic crust can account for several primary observations. The recycling of subducted crust has been proposed to explain ocean island basalt geochemistry in general (e.g., see Hofmann and White, 1980), and eclogite is proposed to explain the great melt volumes at large igneous provinces that cannot be explained by peridotitic plumes at realistic temperatures (e.g., see Cordery et al., 1997). Recycling to the core–mantle boundary, the thermal boundary layer from which classical plumes are thought rise, is the conventional model. However, no appropriate lower mantle seismic anomaly is detected beneath Iceland, and youthful slabs are expected to equilibrate at shallow depth (Oxburgh and Parmentier, 1977). Recycled crustal components of Caledonian age have been suggested from lavas of the Iceland volcanic province (e.g., see Korenaga and Kelemen, 2000; McKenzie et al., 2004). The crustal thickness anomaly is sharply bounded by the Icelandic shelf edge (Figs. 2, 3a and 7). It does not reduce quasi-exponentially between Iceland and the Charlie Gibbs fracture zone, as has sometimes been suggested (e.g., see Jones et al., 2002). This, along with discontinuities in geochemistry observed across transform zones in the region, suggests shallow local control of melt extraction rather than deep broad-scale control.

The possibility that the anomalous melt volume may be derived simply from a relatively shallow fertility anomaly raises some interesting questions:

(a) The question of heat. Whereas peridotitic systems have been studied exhaustively (e.g., see Asimow et al., 2001), details of how eclogite or eclogite–peridotite mixtures melt under conditions of passive adiabatic upwelling are less well understood, and at present, the volumes of melt expected cannot be calculated to a meaningful degree of certainty (P. Asimow and D.C. Presnall, personal communications, 2003). What is known is that basalt- or eclogite-rich assemblages can be completely molten at temperatures below the peridotite solidus, and that they will start to melt at greater depths. For a given amount of melt, the thermal requirements of each are about the same except that less specific heat is required to melt the eclogite assemblage since it does not have to be raised to such a high temperature before melting begins. Detailed thermal information that is not yet available is required before heat-balance calculations can be made that reveal whether the full melt thickness at Iceland can be provided by normal, passive ridge-like upwelling of eclogite-rich mantle alone.

(b) The question of homogenization. The degree to which subducted slabs homogenize with the surrounding host mantle is unknown. The results of Yaxley (2000) correspond to homogenized basalt–peridotite mixtures not intact slabs. A progressively warming slab of eclogitized oceanic crust embedded in the upper mantle could blend with surrounding mantle by the resorption of eclogitic melt into subsolidus peridotite before it reached the surface. Such a process has been shown experimentally to occur and would suppress the extraction of early siliceous melts since these are highly reactive with peridotite (Yaxley and Green, 1998). Solid-state convection at upper mantle Rayleigh numbers is unlikely to be chaotic or turbulent and is therefore not an efficient homogenizer (Meibom and Anderson, 2003).

Partial melting of efficiently homogenized subducted, eclogitized crust and peridotite may be viewed as one end-member scenario, and the remelting of largely intact eclogite slabs may be viewed as another. Extensive melting of such slabs and ponding and homogenization of the melt can explain the detailed petrology and geochemistry of Icelandic lavas, as discussed in a companion paper (Foulger et al., in press).

The MAR in general is underlain by a seismically anomalous layer known as the “low-velocity zone” that extends down to ~200 km depth. The region between the Charlie Gibbs and Jan Mayen fracture zones is unusual in that there the low-velocity zone extends weakly down into the transition zone (Montagner and Ritsema, 2001). This anomaly roughly correlates spatially with the bathymetric shallowing that results in Iceland being subaerial. The low-velocity zone is thought to represent partial melt which forms as a result of depression of the lherzolite solidus below the temperature of the mantle by CO_2 (Presnall et al., 2002; Presnall and Gudfinnsson, in press). Up to ~1% of partial melt could explain the seismic anomalies observed (Vinnik et al., in press). An increase in the depth extent of this phenomenon could be caused both by a more fusible composition resulting from the presence of recycled crust, which is a carrier of CO_2, or a moderate temperature anomaly as is thought to occur in this region or a combination of both. The strength of the deeper seismic anomaly could be explained by a temperature anomaly of 50–100 K and ~0.1% of partial melt, an amount that is not thought to be extractable (McKenzie, 1989). Melt may thus exist in this region but not contribute to shallow magmatism at Iceland and along the neighboring ridges.

The unstable tectonics that have persistently characterized the Iceland volcanic province can explain several significant features. The development of a parallel pair of spreading ridges at ~26 Ma captured an intervening crustal block which was subsequently loaded with additional surface lavas, perhaps also thickened by intrusions, and remains beneath central Iceland at the present day. It correlates with the locus of exceptionally thick (>30 km) crust beneath central Iceland (Fig. 3b) and corresponds approximately to this feature in volume (Foulger, submitted for publication). The region of crust >30 km in thickness is usually interpreted as the center of a plume beneath Iceland, but it may thus instead represent a thickened, subsided, captured microplate. This explanation fits the observations better than one involving high melt productivity over a plume stem since that predicts a laterally extensive band of equally thick crust extending to the northwest and southeast parallel to the predicted plume trajectory, which is not observed (Foulger et al., 2003).

Eastward migration of spreading in Iceland is often cited as evidence in support of an eastward-migrating plume. The palinspastic reconstruction (Fig. 6) shows, however, that such a model is incomplete and misleading. The regular progression of basalts extending from the NVZ to the east coast shows that spreading has proceeded about an eastern center since at least ~15 Ma (Saemundsson, 1979; Bott, 1985; Jancin et al., 1985), and marine observations suggest that this situation has probably been stable since ~26 Ma (Bott, 1985). Given the fixed location of this rift relative to the Kolbeinsey ridge, the western rift has thus tended to be rafted west with the north American plate. To remain in proximity with the MAR melting zone, it has had to repeatedly relocate to the east. These relocations do not represent an eastward migration of the entire melt extraction locus in the region, which has remained generally fixed relative to the Kolbeinsey ridge probably since ~26 Ma. Indeed, southerly ridge propagation is a more prominent characteristic of the Iceland region than easterly migration. The broad distribution of spreading qualifies the Iceland region as a diffuse plate boundary (Zatman et al., 2001).

It is unclear whether the western spreading center relocated further east in single episodes ("jumps") or by gradual propagation of new rift branches from the MAR, behavior similar to propagating centers on the East Pacific rise. In view of the shortness of the newer ridges, which are of the order of the length of one or two spreading centers, this may be an ill-posed distinction. However, the net effect seems to have been similar to what occurs on the East Pacific rise, which is to shorten transform offsets, increase the linearity of the active ridge, and raft abandoned ridges away with the plate.

The tectonic history can also account for variations in crustal thickness within Iceland. The crust beneath western Iceland is thinner than beneath eastern Iceland (Fig. 3b). This observation is enigmatic in the context of the southeastward-migrating plume model since that model predicts thicker crust beneath northwestern Iceland in the wake of the plume than beneath eastern Iceland, where the plume has yet to arrive (Foulger et al., 2003). If the eastern spreading center has persistently been dominant compared with the western center, which has experienced rift extinctions and migrations, magmatism at the western center may have been subsidiary to that of the eastern center. This would result in thinner crust being formed in the west.

The unusual jigsaw tectonics of the Iceland region may result from several factors. Structure in the Caledonian collision zone may have guided the nature of original continental breakup. The region separates the ocean basins to the north and south that have persistently exhibited contrasting spreading styles and appear to behave tectonically independently to some degree. At least two continental microplates, the JMM and the Faeroe microplate, have affected spreading locally, and recent evidence suggests that a sliver of continental crust may underlie Iceland itself (Amundsen et al., 2002). The geochemistry of Icelandic lavas suggests that this may be small in volume, but it may nevertheless have influenced surface tectonics in the region. In addition, the great crustal thickness itself has probably affected surface tectonics. Melt forming in the mantle must be transported upward through at least 30 km of crust before reaching the surface rather than the 7 km typical of mid-ocean ridges, a factor that in itself is likely to give rise to a broader region of surface volcanism. The lower crust itself is ductile, and despite the fact that seismic measurements find no evidence for melt there, it is unclear whether it is also part of the melting zone in some sense.

The coupled complex jigsaw tectonics and high magmatic rate may provide mutual positive feedback. Where spreading is shared between parallel rifts, the spreading rate across each is reduced to the level of an ultraslow spreading ridge. Low spreading rates and high eruption rates maximize the downward advection of volatiles by progressive subsidence of surface lavas (Palmason, 1980). Transtension across easterly orientated zones bounding the ephemeral microplates allows additional leakage of melt and contributes locally to the building of central volcano edifices that rise up to an additional kilometer above the general elevation of the present-day subaerial basalt shield, which is ~1 km above sea level.

The model proposed here offers an alternative to the plume model, which attributes Iceland to a narrow cylindrical column of hot mantle rising from the deep mantle (Morgan, 1971). It can account for most geological and geophysical observations from the Iceland volcanic province. It requires further testing, but in its present speculative state, it nevertheless requires fewer forced explanations and fewer contradictions than a plume model. It reverses the traditional view of the origin of melting at "hotspots." We suggest that melt extraction is controlled by lithospheric extension and variations in mantle composition that are related to plate tectonic structures and processes rather than by hot upwellings from great depth driven by a second mode of convection independent of plate tectonics.

Many of the key characteristics of the Iceland volcanic province, e.g., the large melt volume, ocean–island basalt geochemistry and the paucity of evidence for high source temperatures are similar to those at other “hotspots” and large igneous provinces. Subducted slabs recycled in the shallow upper mantle, perhaps linked to delamination of continental lithosphere, may be the source of voluminous ocean–island basalt and compositional variations in MORB along the ridges in general. Similar models may thus provide candidate alternative hypotheses for the genesis of other “hotspots.”

8. Conclusions

(1) Several primary observations at the Iceland volcanic province, e.g., the paucity of evidence for a time-progressive volcanic track, high temperatures or a seismic anomaly in the lower mantle, require special a posteriori adaptions of the classical plume hypothesis, which erode confidence in the model.

(2) An alternative model suggests that the Iceland melt extraction anomaly is related to local enhanced fertility resulting from recycled subducted Caledonian crust in the upper mantle beneath the portion of the MAR that crossed the frontal thrust of the Caledonian suture. This has resulted in intense magmatism along a 250–500-km section of the MAR, which has built a ridge of thick crust traversing the whole north Atlantic. In addition to accounting for the melt volume, this model can explain the OIB geochemistry, the lack of a lower mantle seismic anomaly, and Caledonian ages determined for recycled material in Icelandic basalts.

(3) The low-wave-speed seismic anomaly that extends down to the mantle transition zone beneath the region between the Charlie Gibbs and Jan Mayen fracture zones may be explained by the presence of traces of partial melt caused by a combination of enhanced fusibility and a moderate regional temperature anomaly.

(4) The Iceland region has persistently exhibited complex leaky jigsaw tectonics involving a southward-propagating style of parallel-pair spreading, intervening microplates and local variations in the spreading direction. The overall locus of spreading has not migrated east during the last ~26 m.y. as is often claimed in support of an eastward-migrating plume model.

(5) The tectonic complexities at Iceland can explain the variations in crustal thickness there. Subsidiary spreading along the westernmost ridges has resulted in thinner crust beneath west Iceland than east Iceland. Local crustal thickening of up to ~40 km beneath central Iceland may represent a captured microplate submerged beneath younger lavas. If this is entirely oceanic, it must contain crust up to ~30 m.y. old.

(6) Eastward-increasing, fan-shaped transtension may be occurring about a west–east zone traversing central Iceland from the Snaefellsnes peninsula to Vatnajokull. Enhanced volcanism beneath the MVZ and northwest Vatnajokull, traditionally assumed to represent the location of a mantle plume may instead represent the culmination of extension at the eastern end of this zone.

(7) The model described here offers an alternative explanation for the observations at Iceland that is based on shallow structures and processes related to plate tectonics. Similar interpretations may be viable for other “hotspots.” Such a model represents a simplifying view of mantle convection since it appeals to only one mode of convection—that associated with plate tectonics—and does not require additionally a second independent mode associated with deep mantle plumes.

Acknowledgement

We are grateful for discussions with Jim Natland, Dean Presnall, Joann Stock, Hugh Taylor, Peter Wyllie, Paul Asimow, Brian Wernicke, Bruce Julian, Seth Stein, Carol Stein, Warren Hamilton, Jerry Winterer, and Anders Meibom. Detailed reviews by Geoffrey Davies and an anonymous reviewer, who do not necessarily ascribe to the ideas described herein, improved the paper. The work was supported by NERC grant GR3/10727 and a Sir James Knott Foundation Fellowship held by GRF.

References

Amundsen, H.E.F., Schaltegger, U., Jamtveit, B., Griffin, W.L., Podladchikov, Y.Y., Torsvik, T., Gronvold, K., 2002. Reading the LIPs of Iceland and Mauritius. In: Jamtveit, B., Amundsen, H.E.F. (Eds.), 15th Kongsberg Seminar, Kongsberg, Norway.

Anderson, D.L., 1994a. The sublithospheric mantle as the source of continental flood basalts; the case against continental lithosphere and plume head reservoirs. Earth Planet. Sci. Lett. 123, 269–280.

Anderson, D.L., 1994b. Superplumes or supercontinents. Geology 22, 39–42.

Anderson, D.L., 1995. Lithosphere, asthenosphere and perisphere. Rev. Geophys. 33, 125–149.

Asimow, P.D., Langmuir, C.H., 2003. The importance of water to oceanic mantle melting regimes. Nature 421, 815–820.

Asimow, P.D., Hirschmann, M.M., Stolper, E., 2001. Calculation of peridotite partial melting from thermodynamic models of minerals and melts: IV. Adiabatic decompression and the composition and mean properties of mid-ocean ridge basalts. J. Pet. 42, 963–998.

Barker, A.J., Gayer, R.A., 1984. Caledonide–Appalachian tectonic analysis and evolution of related oceans. In: Geyer, R.A. (Ed.), The Tectonic Evolution of the Caledonide–Appalachian Orogen. International Monograph Series on Interdisciplinary Earth Science Research and Applications. Friedrich Vieweg 7 Sohn, Braunschweig V, pp. 126–165.

Bijwaard, H., Spakman, W., 1999. Tomographic evidence for a narrow whole mantle plume below Iceland. Earth Planet. Sci. Lett. 166, 121–126.

Bjarnason, I., Einarsson, P., 1991. Source mechanism of the 1987 Vatnafjoll earthquake in south Iceland. J. Geophys. Res. 96, 4313–4324.

Bott, M.H.P., 1985. Plate tectonic evolution of the Icelandic transverse ridge and adjacent regions. J. Geophys. Res. 90, 9953–9960.

Bott, M.H.P., 1987. The continental margin of central east Greenland in relation to north Atlantic plate tectonic evolution. J. Geol. Soc. (Lond.) 144, 561–568.

Boutilier, R.R., Keen, C.E., 1999. Small-scale convection and divergent plate boundaries. J. Geophys. Res. 104, 7389–7403.

Breddam, K., 2002. Kistufell: primitive melt from the Iceland mantle plume. J. Pet. 43, 345–373.

Bunch, A.W.H., Kennett, B.L.N., 1980. The crustal structure of the Reykjanes ridge at 59°30’N. Geophys. J. R. Astron. Soc. 61, 141–166.

Cannat, M., 1996. How thick is the magmatic crust at slow spreading oceanic ridges? J. Geophys. Res. 101, 2847–2857.

Chalmers, J.A., Larsen, L.M., Pederson, A.K., 1995. Widespread volcanism around the north Atlantic and Labrador Sea: evidence for a large, hot early plume head. J. Geol. Soc. (Lond.) 152, 965–969.

Chauvel, C., Hemond, C., 2000. Melting of a complete section of recycled oceanic crust: Trace element and Pb isotopic evidence from Iceland. Geochem. Geophys. Geosyst. 1, 1999GC000002.

Clift, P.D., 1997. Temperature anomalies under the northeast Atlantic rifted volcanic margins. Earth Planet. Sci. Lett. 146, 195–211.

Clift, P.D., Carter, A., Hurford, A.J., 1998. The erosional and uplift history of NE Atlantic passive margins: constraints on a passing plume. J. Geol. Soc. (Lond.) 155, 787–800.

Cordery, M.J., Davies, G.F., Campbell, I.H., 1997. Genesis of flood basalts from eclogite-bearing mantle plumes. J. Geophys. Res. 102, 20179–20197.

Davies, G.F., 1999. Dynamic Earth. Cambridge University Press, Cambridge, 458+xi pp.

DeLaughter, J., Stein, C.A., Stein, S., 2004. Hotspots: a view from the swells. In: Foulger, G.R., Anderson, D.L., Natland, J.H., Presnall, D.C. (Eds.), Plates, Plumes and Paradigms, Geological Society of America, Special Paper (in press).

DeMets, C., Gordon, R.G., Argus, D.F., Stein, S., 1994. Effect of recent revisions to the geomagnetic reversal time scale on estimate of current plate motions. Geophys. Res. Lett. 21, 2191–2194.

Dewey, J.F., Shackleton, R.M., 1984. A model for the evolution of the Grampian tract in the early Caledonides and Appalachians. Nature 312, 115–121.

Du, Z., Foulger, G.R., 1999. The crustal structure beneath the Northwest Fjords, Iceland, from receiver functions and surface waves. Geophys. J. Int. 139, 419–432.

Du, Z.J., Foulger, G.R., 2001. Variation in the crustal structure across central Iceland. Geophys. J. Int. 145, 246–264.

Du, Z., Foulger, G.R., Julian, B.R., Allen, R.M., Nolet, G., Morgan, W.J., Bergsson, B.H., Erlendsson, P., Jakobsdottir, S., Ragnarsson, S., Stefansson, R., Vogfjord, K., 2002. Crustal structure beneath western and eastern Iceland from surface waves and receiver functions. Geophys. J. Int. 149, 349–363.

Duncan, R.A., Helgason, J., 1998. Precise dating of the Holmatindur cooling event in eastern Iceland: evidence for mid-Miocene bipolar glaciation. J. Geophys. Res. 103, 12397–12404.

Einarsson, P., 1991. Earthquakes and present-day tectonism in Iceland. Tectonophysics 189, 261–279.

Einarsson, P., Klein, F.W., Bjornsson, S., 1977. The Borgarfjordur earthquakes of 1974 in west Iceland. Bull. Seismol. Soc. Am. 67, 187–208.

Fitton, J.G., Saunders, A.D., Kempton, P.D., Hardarson, B.S., 2003. Does depleted mantle form an intrinsic part of the Iceland plume? Geochem. Geophys. Geosyst. 4, 1032.

Foulger, G.R., 2003. On the apparent eastward migration of the spreading ridge in Iceland. In: Foulger, G.R., Natland, J.H., Anderson, D.L. (Eds.), Proceedings of the Plume: VI. Beyond the Plume Hypothesis Conference, Geological Society of America, Hveragerdi, Iceland, 2003.

Foulger, G.R., 2004. Old crust underlies Iceland. Geophys. J. Int. (submitted for publication).

Foulger, G.R., Jahn, C.-H., Seeber, G., Einarsson, P., Julian, B.R., Heki, K., 1992. Post-rifting stress relaxation at the divergent plate boundary in Iceland. Nature 358, 488–490.

Foulger, G.R., Pritchard, M.J., Julian, B.R., Evans, J.R., Allen, R.M., Nolet, G., Morgan, W.J., Bergsson, B.H., Erlendsson, P., Jakobsdottir, S., Ragnarsson, S., Stefansson, R., Vogfjord, K., 2000. The seismic anomaly beneath Iceland extends down to the mantle transition zone and no deeper. Geophys. J. Int. 142, F1–F5.

Foulger, G.R., Pritchard, M.J., Julian, B.R., Evans, J.R., Allen, R.M., Nolet, G., Morgan, W.J., Bergsson, B.H., Erlendsson, P., Jakobsdottir, S., Ragnarsson, S., Stefansson, R., Vogfjord, K., 2001. Seismic tomography shows that upwelling beneath Iceland is confined to the upper mantle. Geophys. J. Int. 146, 504–530.

Foulger, G.R., Du, Z., Julian, B.R., 2003. Icelandic-type crust. Geophys. J. Int. 155, 567–590.

Foulger, G.R., Natland, J.H., Anderson, D.L., 2004. A source for Icelandic magmas in remelted Iapetus crust. J. Volcanol. Geotherm. Res. (doi:10.1016/j.volgeores.2004.10.006).

Goldflam, P., Weigel, W., Loncarevic, B.D., 1980. Seismic structure along RRISP-Profile I on the southeast flank of the Reykjanes ridge. J. Geophys. 47, 250–260.

Green, D.H., Falloon, T.J., Eggins, S.M., Yaxley, G.M., 2001. Primary magmas and mantle temperatures. Eur. J. Mineral. 13, 437–451.

Gudfinnsson, G., Presnall, D.C., Oskarsson, N., 2003. Contrasting origins of the most magnesian glasses from Iceland and Hawaii. In: Foulger, G.R., Natland, J.H., Anderson, D.L. (Eds.), Proceedings of the Plume: IV. Beyond the Plume Hypothesis Conference, Geological Society of America, Hveragerdi, Iceland, 2003.

Gudmundsson, MT., Sigmundsson, F., Bjornsson, H., 1997. Ice–volcano interaction of the 1996 Gjalp subglacial eruption, Vatnajokull, Iceland. Nature 389, 954–957.

Hardarson, B.S., Fitton, J.G., 1991. Increased mantle melting beneath Snaefellsjokull volcano during late Pleistocene deglaciation. Nature 353, 62–64.

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 northwest Iceland. Earth Planet. Sci. Lett. 153, 181–196.

Helgason, J., Duncan, R.A., 2001. Glacial–interglacial history of the Skaftafell region, southeast Iceland, 0–5 Ma. Geology 29, 179–182.

Hofmann, A.W., White, W.M., 1980. The role of subducted oceanic crust in mantle evolution. Carnegie Inst. Wash. Yearb. 79, 477–483.

Hofton, M.A., Foulger, G.R., 1996. Post-rifting anelastic deformation around the spreading plate boundary, north Iceland: 1. Modeling of the 1987–1992 deformation field using a viscoelastic Earth structure. J. Geophys. Res. 101, 25403–25421.

Holbrook, W.S., Larsen, H.C., Korenaga, J., Dahl-Jensen, T., Reid, I.D., Kelemen, P.B., Hopper, J.R., Kent, G.M., Lizarralde, D., Bernstein, S., Detrick, R.S., 2001. Mantle thermal structure and active upwelling during continental breakup in the north Atlantic. Earth Planet. Sci. Lett. 190, 251–266.

Ito, G., Shen, Y., Hirth, G., Wolfe, C.J., 1999. Mantle flow, melting and dehydration of the Iceland mantle plume. Earth Planet. Sci. Lett. 165, 81–96.

Jancin, M., Young, K.D., Voight, B., Aronson, J.L., Saemundsson, K., 1985. Stratigraphy and K/Ar ages across the west flank of the northeast Iceland axial rift zone, in relation to the 7 Ma volcano–tectonic reorganization of Iceland. J. Geophys. Res. 90, 9961–9985.

Jóhannesson, H., Saemundsson, K., 1998. Jardfraedikort af Islandi. Hoggun (Geological Map of Iceland. Tectonics). Náttúrfraedistofunu Islands (Icelandic Institute of Natural History), Reykjavík.

Jones, S.M., White, N., Maclennan, J., 2002. V-shaped ridges around Iceland; implications for spatial and temporal patterns of mantle convection. Geochem. Geophys. 3.

Karato, S., 1993. Importance of anelasticity in the interpretation of seismic tomography. Geophys. Res. Lett. 20, 1623–1626.

Klein, E.M., Langmuir, C.H., 1987. Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness. J. Geophys. Res. 92, 8089–8115.

Kodaira, S., Mjelde, R., Gunnarsson, K., Shiobara, H., Shimamura, H., 1997. Crustal structure of the Kolbeinsey ridge, north Atlantic, obtained by use of ocean bottom seismographs. J. Geophys. Res. 102, 3131–3151.

Kodaira, S., Mjelde, R., Gunnarsson, K., Shiobara, H., Shimamura, H., 1998. Evolution of oceanic crust on the Kolbeinsey Ridge, north of Iceland, over the past 22 Myr. Terra Nova 10, 27–31.

Korenaga, J., Kelemen, P.B., 2000. Major element heterogeneity in the mantle source of the north Atlantic igneous province. Earth Planet. Sci. Lett. 184, 251–268.

Larsen, H.C., Jakobsdottir, S., 1988. Distribution, crustal properties and significance of seawards-dipping sub-basement reflectors off E Greenland. In: Morton, A., Parson, L.M. (Eds.), Early Tertiary volcanism and the opening of the NE Atlantic. Special Publication Geological Society, 39, 95–114.

Larsen, L.M., Pedersen, A.K., 2000. Processes in high-Mg, high-T magmas: evidence from olivine, chromite and glass in palaeogene picrites from West Greenland. J. Pet. 41, 1071–1098.

Lawver, LA., Muller, R.D., 1994. Iceland hotspot track. Geology 22, 311–314.

Lundin, E., Dore, T., 2004. The fixity of the Iceland "hotspot" on the Mid-Atlantic Ridge: observational evidence, mechanisms and implications for Atlantic volcanic margins. In: Foulger, G.R., Natland, J.H., Presnall, D.C., Anderson, D.L. (Eds.), Plates, Plumes and Paradigms. Special Paper Geological Society of America (in press).

Lundin, E.R., Doré, A.G., 2004. NE-Atlantic break-up: a re-examination of the Iceland mantle plume model and the Atlantic–Arctic linkage. In: Doré, A.G., Vining, B.A. (Eds.), North-West European Petroleum Geology and Global Perspectives: 6th Conference. Geological Society, London (in press).

Marquart, G., 1991. Interpretations of geoid anomalies around the Iceland hotspot. Geophys. J. Int. 106, 149–160.

McDougall, I., Kristjansson, L., Saemundsson, K., 1984. Magnetostratigraphy and geochronology of northwest Iceland. J. Geophys. Res. 89, 7029–7060.

McKenzie, D., 1989. Some remarks on the movement of small melt fractions in the mantle. Earth Planet. Sci. Lett. 95, 53–72.

McKenzie, D., Bickle, J., 1988. The volume and composition of melt generated by extension of the lithosphere. J. Pet. 29, 625–679.

McKenzie, D., Stracke, A., Blichert-Toft, J., Albarede, F., Gronvold, K., O'Nions, R.K., 2004. Source enrichment processes responsible for isotopic anomalies in oceanic island basalts. Geochim. Cosmochim. Acta 68, 2699–2724.

Megnin, C., Romanowicz, B., 2000. The 3D velocity structure of the mantle from the inversion of body, surface, and higher mode waveforms. Geophys. J. Int. 143, 709–728.

Meibom, A., Anderson, D.L., 2003. The statistical upper mantle assemblage. Earth Planet. Sci. Lett. 217, 123–139.

Menke, W., 1999. Crustal isostasy indicates anomalous densities beneath Iceland. Geophys. Res. Lett. 26, 1215–1218.

Montagner, J.P., Ritsema, J., 2001. Interactions between ridges and plumes. Science 294, 1472–1473.

Montelli, R., Nolet, G., Dahlen, F.A., Masters, G., Engdahl, R.E., Hung, S.-H., 2004. Finite frequency tomography reveals a variety of plumes in the mantle. Science 303, 338–343.

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

Mutter, C.Z., Mutter, J.C., 1993. Variations in thickness of layer 3 dominate oceanic crustal structure. Earth Planet. Sci. Lett. 117, 295–317.

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. Special Publications Geological Society, 39, 35–48.

Natland, J., 2003. What's going on at Iceland?. In: Foulger, G.R., Natland, J.H., Anderson, D.L. (Eds.), Proceedings of the Plume: IV. Beyond the Plume Hypothesis Conference. Geological Society of America, Hveragerdi, Iceland.

Navin, D.A., Peirce, C., Sinha, M.C., 1998. The RAMESSES experiment: II. Evidence for accumulated melt beneath a slow spreading ridge from wide-angle refraction and multichannel reflection seismic profiles. Geophys. J. Int. 135, 746–772.

Nilsen, T.H., 1978. Lower Tertiary laterite on the Iceland–Faeroe ridge and the Thulean land bridge. Nature 274, 786–788.

Oskarsson, N., Sigvaldason, G.E., Steinthorsson, S., 1982. A dynamic model of rift zone petrogenesis and the regional petrology of Iceland. J. Pet. 23, 28–74.

Oxburgh, E.R., Parmentier, E.M., 1977. Compositional and density stratification in oceanic lithosphere—causes and consequences. J. Geol. Soc. (Lond.) 133, 343–355.

Palmason, G., 1980. A continuum model of crustal generation in Iceland; kinematic aspects. J. Geophys. 47, 7–18.

Polet, J., Anderson, D., 1995. Depth extent of cratons as inferred from tomographic studies. Geology 23, 205–208.

Presnall, D.C., 2003. Low temperature origin of oceanic ridges and associated hot spots. EOS Trans. AGU, Abstract V31F-01.

Presnall, D.C., Gudfinnsson, G.H., 2004. Potential temperatures at ridges and associated "hot spots": low velocity zone constraints. In: Foulger, G.R., Natland, J.H., Presnall, D.C., Anderson, D.L. (Eds.), Plates, Plumes and Paradigms. Special Paper Geological Society of America (in press).

Presnall, D.C., Gudfinnsson, G.H., Walter, M.J., 2002. Generation of mid-ocean ridge basalts at pressures from 1 to 7 GPa. Geochim. Cosmochim. Acta 66, 2073–2090.

Ribe, N.M., Delattre, W.L., 1998. The dynamics of plume–ridge interaction: 3. The effects of ridge migration. Geophys. J. Int. 133, 511–518.

Ribe, N.M., Christensen, U.R., Theissing, J., 1995. The dynamics of plume–ridge interaction: 1. Ridge-centered plumes. Earth Planet. Sci. Lett. 134, 155–168.

Ritsema, J., Allen, R.M., 2003. The elusive mantle plume. Earth Planet. Sci. Lett. 207, 1–12.

Ritsema, J., van Heijst, H.J., Woodhouse, J.H., 1999. Complex shear wave velocity structure imaged beneath Africa and Iceland. Science 286, 1925–1928.

Ritzert, M., Jacoby, W.R., 1985. On the lithospheric seismic structure of Reykjanes ridge at 62.5°N. J. Geophys. Res. 90, 10117–10128.

Roberts, D., 2003. The Scandanavian Caledonides: event chronology, palaeogeographic settings and likely, modern analogues. Tectonophysics 365, 283–299.

Saemundsson, K., 1979. Outline of the geology of Iceland. Jokull 29, 7–28.

Schilling, J.-G., Zajac, M., Evans, R., Johnston, T., White, W., Devine, J.D., Kingsley, R., 1983. Petrologic and geochemical variations along the mid-Atlantic ridge from 29°N to 73°N. Am. J. Sci. 283, 510–586.

Skogseid, J., Planke, S., Faleide, J.I., Pedersen, T., Eldholm, O., Neverdal, F., 2000. NE Atlantic continental rifting and volcanic margin formation. In: Nottvedt, A. (Ed.), Dynamics of the Norwegian Margin, Spec. Publ.-Geol. Soc. Lond. 167, 295–326.

Smallwood, J.R., White, R.S., 1998. Crustal accretion at the Reykjanes ridge, 61°–62° N. J. Geophys. Res. 103, 5185–5201.

Smallwood, J.R., White, R.S., Minshull, T.A., 1995. Sea-floor spreading in the presence of the Iceland plume: the structure of the Reykjanes ridge at 61°40'N. J. Geol. Soc. (Lond.) 152, 1023–1029.

Soper, N.J., Strachan, R.A., Holdsworth, R.E., Gayer, R.A., Greiling, R.O., 1992. Sinistral transpression and the Silurian closure of Iapetus. J. Geol. Soc. (Lond.) 149, 871–880.

Stein, C., Stein, S., 2003. Sea floor heat flow near Iceland and implications for a mantle plume. Astron. Geophys. 44, 1.8–1.10.

Tanton, L.T.E., Hager, B.H., 2000. Melt intrusion as a trigger for lithospheric foundering and the eruption of the Siberian flood basalts. Geophys. Res. Lett. 27, 3937–3940.

Taylor Jr, H.P., 1968. The oxygen isotope geochemistry of igneous rocks. Contrib. Mineral. Petrol. 19, 1–71.

Tegner, C., Lesher, C.E., Larsen, L.M., Watt, W.S., 1998. Evidence from the rare-earth-element record for mantle melting for cooling of the Tertiary Iceland plume. Nature 395, 591–594.

Udagawa, S., Kitagawa, H., Gudmundsson, A., Hiroi, O., Koyaguchi, T., Tanaka, H., Kristjansson, L., Kono, M., 1999. Age and magnetism of lavas in Jokuldalur area, Eastern Iceland: Gilsa event revisited. Phys. Earth Planet. Inter. 115, 147–171.

Vink, G.E., 1984. A hotspot model for Iceland and the Voring plateau. J. Geophys. Res. 89, 9949–9959.

Vinnik, L.P., Du, Z., Foulger, G.R., 2004. Seismic boundaries in the mantle beneath Iceland: a new constraint on temperature. Geophys. J. Int. (in press).

Voelksen, C., 2000. Die Nutzung von GPS für die Deformationsanalyse in regionalen Netzen am Bespiel Islands. PhD thesis, Hannover, Hannover.

Vogt, P.R., 1971. Asthenosphere motion recorded by the ocean floor south of Iceland. Earth Planet. Sci. Lett. 13, 153–160.

von Herzen, R.P., 2001. Summary of geothermal data around the Iceland hot spot. EOS Trans. AGU 82, F1184.

Weigel, W., Flueh, E.R., Miller, H., Butzke, A., Dehghani, G.A., Gebhardt, V., Harder, I., Hepper, J., Jokat, W., Klaeschen, D., Kreymann, S., Schuessler, S., Zhao, Z., 1995. Investigations of the East Greenland continental margin between 70° and 72° N by deep seismic sounding and gravity studies. Mar. Geophys. Res. 17, 167–199.

Weir, N.R.W., White, R.S., Brandsdottir, B., Einarsson, P., Shimamura, H., Shiobara, H., 2001. Crustal structure of the northern Reykjanes ridge and Reykjanes peninsula, southwest Iceland. J. Geophys. Res. 106, 6347–6368.

Whitmarsh, R.B., 1975. Axial intrusion zone beneath the median valley of the mid-Atlantic ridge at 37° N detected by explosion seismology. Geophys. J. R. Astron. Soc. 42, 189–215.

Whitmarsh, R.B., Calvert, A.J., 1986. Crustal structure of Atlantic fracture zones: I. The Charlie–Gibbs fracture zone. Geophys. J. R. Astron. Soc. 85, 107–138.

Wolfe, C.J., Bjarnason, I.T., VanDecar, J.C., Solomon, S.C., 1997. Seismic structure of the Iceland mantle plume. Nature 385, 245–247.

Yaxley, G.M., 2000. Experimental study of the phase and melting relations of homogeneous basalt + peridotite mixtures and implication for the petrogenesis of flood basalts. Contrib. Mineral. Petrol. 139, 326–338.

Yaxley, G.M., Green, D.H., 1998. Reactions between eclogite and peridotite: mantle refertilisation by subduction of oceanic crust. Schweiz. Mineral. Petrogr. Mitt. 78, 243–255.

Zatman, S., Gordon, R.G., Richards, M.A., 2001. Analytic models for the dynamics of diffuse oceanic plate boundaries. Geophys. J. Int. 145, 145–156.