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Genesis of the Iceland melt anomaly by plate tectonic processes

Gillian R. Foulger*
Department of Earth Sciences, University of Durham, Durham DH1 3LE, UK
James H. Natland*
Rosenstiel School of Marine and Atmospheric Science, University of Miami,
4600 Rickenbacker Causeway, Miami, Florida 33149, USA
Don L. Anderson*
California Institute of Technology, Seismological Laboratory 252-21, Pasadena, California 91125, USA

ABSTRACT

Iceland is the best studied, large-volume, active volcanic region in the world. It features the largest subaerial exposure of any hotspot at a spreading ridge, and it is conventionally attributed to a thermal plume in the mantle. However, whereas the apparently large melt productivity and low-wavespeed mantle seismic anomaly are consistent with this attribution, at any more detailed level, the observations are poorly predicted by the plume hypothesis. There is no time-progressive volcanic track, the melt anomaly having been persistently centered on the Mid-Atlantic Ridge. Spatial variations in crustal structure are inconsistent with the southeastward migration that is required of a plume fixed with respect to other Indo-Atlantic hotspots. The mantle seismic anomaly weakens with depth and does not extend into the lower mantle. Estimates of excess temperature using a broad range of methods are inconsistent with a mantle potential temperature anomaly greater than a few tens of K. Much of the lava erupted in Iceland has geochemistry little different from normal mid-ocean ridge basalt, and the detailed spatial geochemical pattern bears little resemblance to what is predicted for a plume beneath central Iceland. We propose an alternative model in an attempt to explain the observations at Iceland with fewer difficulties. Our model involves only shallow plate tectonic processes and attributes the large melt volume to the remelting of subducted oceanic crust trapped in the Caledonian suture in the form of eclogite or mantle peridotite fertilized by resorbed eclogite. Delaminated continental mantle lithosphere may also be involved. Such a source can produce several times more melt than pure peridotite without the need for high temperatures. The longevity of anomalous volcanism at the Mid-Atlantic Ridge at the latitude of Iceland is attributed to its location on a Caledonian structure that runs transversely across the north Atlantic. Many aspects of the geochemistry of Icelandic lavas fit this model, which also provides an explanation for the high maximum helium isotope ratios observed there. The "depleted plume component" may be derived from abyssal olivine gabbro cumulates and the "enriched plume component" from recycled enriched material that forms part of the crustal section of subducted slabs. Such a model for the Iceland melting anomaly raises new questions concerning how much thermal energy can be generated by isentropic upwelling of eclogite at a ridge, the location of the homogenizing

*E-mails: g.r.foulger@durham.ac.uk; jnatland@msn.com; dla@gps.caltech.edu.

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reservoirs required, and the mechanism by which fertile material is incorporated into the asthenosphere beneath new oceans. Most fundamentally, if validated, such a model can explain the generic observations associated with hotspots as shallow processes associated with plate tectonics, and thus raises the question of whether thermal plumes are required in general in the Earth.

Keywords: Iceland, hotspot, plume, tectonics, volcanism

INTRODUCTION

Iceland is arguably the best-studied large-volume volcanic anomaly in the world. It features the largest subaerial exposure of any portion of the global spreading plate boundary and is considered to be the type example of a ridge-centered hotspot. Its structure, geology, geophysics, and tectonics have been described in detail in many papers (e.g., Saemundsson, 1979; Björnsson, 1985; Foulger et al., 2001, 2003). In brief, Iceland comprises a basaltic plateau some 450 × 350 km in size, centered on the Mid-Atlantic Ridge (MAR) (Fig. 1). Some 350 km of spreading ridge are exposed on land, including ~35 en echelon spreading segments, most containing a central volcano. Large intraplate volcanoes and volcanic systems also occur. Iceland is flanked by the Greenland-Iceland and the Iceland-Faeroe aseismic ridges, and all three regions are underlain by crust with a seismic thickness (i.e., the thickness of the layer in which the seismic wavespeed is characteristic of crustal rocks) of ~30 km (Bott and Gunnarsson, 1980; Staples et al., 1997; Holbrook et al., 2001; Foulger et al., 2003). Much of the north Atlantic is underlain by a low-wavespeed mantle seismic anomaly similar to that beneath other parts of the MAR in the upper ~200 km, but with, additionally, a weak extension that continues down into the transition zone (Montagner and Ritsema, 2001; Lundin and Doré, this volume).

Iceland and its flanking aseismic ridges are most popularly attributed to a mantle plume that is postulated to have impinged on the base of the lithosphere beneath central Greenland at ca. 62 Ma (e.g., Morgan, 1971, 1972; White and McKenzie, 1995). In this paper, we criticize that model and propose an alternative. A necessary prerequisite is to define what is meant by a plume, and this presents the first challenge. The original, classical plume hypothesis proposed the existence of thermally buoyant diapirs rising from the deep mantle (Morgan, 1971). A deep origin was postulated to explain the apparent relative fixity of hotspots, which would not occur if the melt sources originated in the shallow convecting mantle and thus moved relative to one another. It is generally accepted that thermal plumes must rise from a thermal boundary layer, and the only such layer known to occur in the deep mantle is near the core-mantle boundary.

Subsequent to the original proposal, however, the hypothesis was adapted in many ways to fit observations that were either unpredicted by the original model or were apparently contrary to its predictions. These adaptations led to a loss of clarity regarding what characteristics are required by a plume, and to usage of the term by different authors to mean different, sometimes mutually exclusive, things (see www.mantleplumes.org/DefinitionOfAPlume.html for diverse definitions). For example, ocean-island basalt (OIB) geochemistry is empirically associated with hypothesized plumes (e.g., Schilling, 1973b), and is not required by a thermal plume. Nevertheless, OIB is now generally assumed to have a plume origin, even if lateral flow, perhaps for thousands of km from the nearest presumed plume, must be invoked. The term *plume* may be used to refer to features that are confined to the lithosphere (e.g., Courtillot et al., 2003), or to correspond simply to convective upwellings of unspecified origin and genesis (McKenzie et al., 2004). Neither of these models can explain the apparent relative fixity of hotspots that was one of the primary observations that the plume hypothesis was invoked to explain. Nevertheless, the use of the term *plume* for these, and many other envisaged advective phenomena in the mantle, makes testing of the theory difficult, because under these circumstances it is, in practice, ill defined. We distinguish here between the original, classical plume hypothesis, which was precisely defined by Morgan (1971), and the contemporary plume hypothesis, which is flexible and can, in practice, explain almost any observation.

We argue that the classical plume hypothesis is not well supported at Iceland. Evidence popularly considered to be consistent with this model includes the north Atlantic volcanic margin that formed at the time of breakup (e.g., Boutilier and Keen, 1999); the large seismic crustal thickness, which is interpreted to result from extensive partial melting requiring high temperature; the unusually great depth extent of the mantle seismic anomaly; high maximum $^3He/^4He$ isotope ratios; and OIB geochemistry (e.g., Schilling, 1973a; Hilton et al., 1999; Darbyshire et al., 2000; Ritsema and Allen, 2003).

In this paper, we do not address the contemporary plume hypothesis because its flexibility, in practice, makes it effectively invulnerable. For example, a time-progressive volcanic track is predicted for a classical plume, but the lack of one in the Iceland region is explained by "mantle wind," "hotspot mobility," or lateral flow from a plume that does not manifest at its true location (e.g., Vink, 1984). Depleted components in Icelandic magmas have been attributed to a "depleted plume," a concept invented purely to account for this observation. Both thick and thin crustal models have been cited in support of the plume hypothesis at Iceland, the controversial "lower crust" being variously interpreted as hot upper mantle or a thick layer of gabbro cumulates

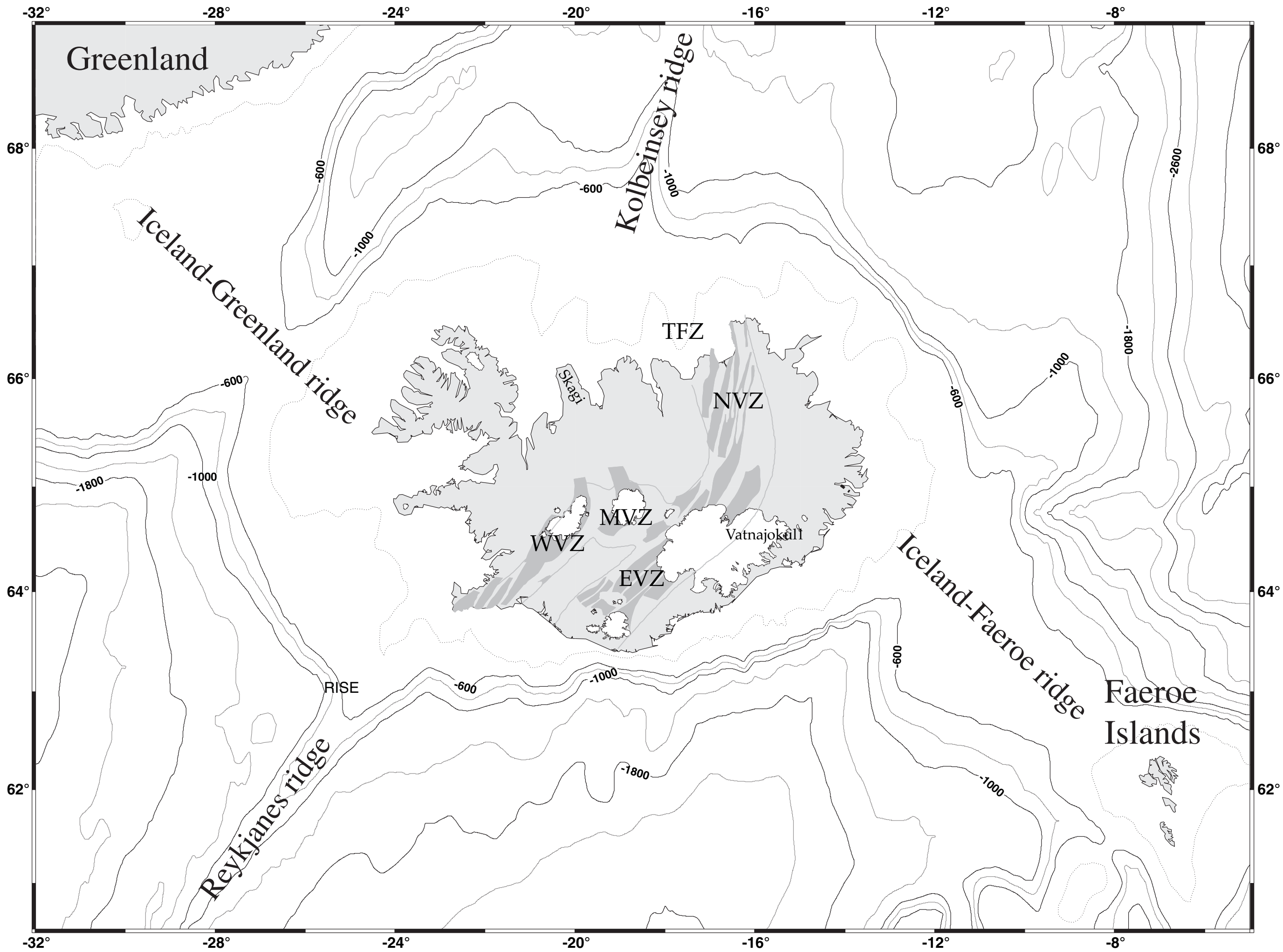

Figure 1. Map of the Icelandic transverse ridge showing bathymetric contours and tectonic features in Iceland. The neovolcanic zones are outlined. Spreading segments (volcanic systems) are shown in dark gray and glaciers in white on land. EVZ—Eastern Volcanic Zone; MVZ—Middle Volcanic Zone; NVZ—Northern Volcanic Zone; TFZ—Tjörnes Fracture Zone; WVZ—Western Volcanic Zone.

(Bjarnason et al., 1994; Björnsson et al., this volume). Seismic tomographic images showing the mantle anomaly to be truncated at the base of the upper mantle (Montagner and Ritsema, 2001; Montelli et al., 2004) have been explained either as the presumed lower-mantle portion being too narrow to be resolved or as a plume rising from the base of the upper mantle, even though it is not expected that this horizon is a thermal boundary layer.

Some of the most remarkable anomalies in the north Atlantic region that require explanation include the shallow regional bathymetry, which peaks at Iceland, causing it to be subaerial; the very thick band of seismic crust that traverses the entire Atlantic from Greenland to the Faeroe Islands (Foulger et al., 2003); and the mantle anomaly that extends down into the transition zone (Ritsema et al., 1999; Foulger et al., 2001). The geochemical anomaly includes some of the highest noncosmogenic $^3He/^4He$ ratios on Earth; steep rare-earth-element (REE) patterns; isotope ratios indicative of long-term source enrichment in some magmas and depletion in others; and diverse petrology, including picrites and a wide range of both alkaline and tholeiitic rocks. In this paper, we present a critical review of the tectonics, geophysics, and geochemistry of the region and propose an alternative model that involves shallow, plate-tectonic processes only and does not appeal to a bottom-heated, thermally buoyant mantle plume. We draw on work published in several recent papers (Foulger et al., 2000, 2001, 2003, 2005; Foulger, 2002, 2004, unpubl. data; Z. Du, L.P. Vinnik, and G.R. Foulger, 2004, unpubl. data.; Foulger and Anderson, 2005). Our model can explain the anomalies in the Iceland region while avoiding much of the special pleading necessary in the plume model. Nevertheless, it presents its own new challenges and problems that require further work.

VOLCANIC HISTORY OF THE NORTH ATLANTIC REGION

Magmatism in the north Atlantic began at ca. 62 Ma and produced volcanic rocks distributed over a ~2,000-km-broad zone encompassing Baffin Island, west Greenland, and northern Britain (e.g., Lundin and Doré, this volume). It accompanied continental breakup and the onset of sea-floor spreading in the Labrador Sea. An important question is that of cause and effect. Is the magmatism a transient result of breakup of a former supercontinent, or was the rifting caused by active upwelling? At ca. 54 Ma, spreading in the Labrador Sea ceased and volcanism and seafloor spreading transferred to what is now the north Atlantic, where volcanic margins ~25 km thick developed (Keen and Potter, 1995; Boutilier and Keen, 1999). After a few m.y., volcanism dwindled along most of the volcanic margin, but it persisted until the present day along a 100- to 350-km-long section of the MAR centered at ~65° N. Thus a band of crust with a seismic thickness of ~30 km developed that traverses the entire north Atlantic. The early widespread volcanism is traditionally attributed to a plume head, which is postulated to have caused continental breakup, and the later localized volcanism to a plume "tail."

The plume model has problems in explaining the detailed volcanic distribution. First, the volcanism that accompanied continental breakup, early in the Labrador Sea and later in the north Atlantic, forms linear arrays along the margins rather than having a circular region of influence (e.g., Chalmers et al., 1995). A rifting model thus explains the distribution of early volcanism better than does radial flow from a localized source. Second, there is no evidence for the time progression of volcanism predicted for a plume fixed relative to other Indo-Atlantic hotspots. Since ca. 62 Ma, such a plume beneath the north Atlantic is predicted to have migrated southeastward from a location beneath central Greenland at ~2 cm/yr with respect to the North American plate, to currently underlie central Iceland (Fig. 2) (Lawver and Muller, 1994). No such track is observed. Volcanism has been focused at the MAR since the opening of the north Atlantic at ca. 54 Ma, as may be deduced from the symmetric ridges of thick crust that flank Iceland (Lundin and Doré, 2003, this volume) (Figs. 1 and 2).

The migration of spreading on land in Iceland, where rocks up to ca. 15 m.y. old are exposed, is frequently cited as evidence for a southeasterly migrating plume. Spreading axes in western Iceland have twice become extinct, and spreading has been transferred to axes farther east. Spreading currently occurs along the Northern and Eastern Volcanic zones (NVZ and EVZ) in eastern Iceland (see Fig. 1). However, the ages of surface lavas (Saemundsson, 1979) show that the present NVZ did not develop recently in response to extinction of an axis farther west. On the contrary, its predecessors have been active ever since the present-day landmass formed, during which time the spatial relationship

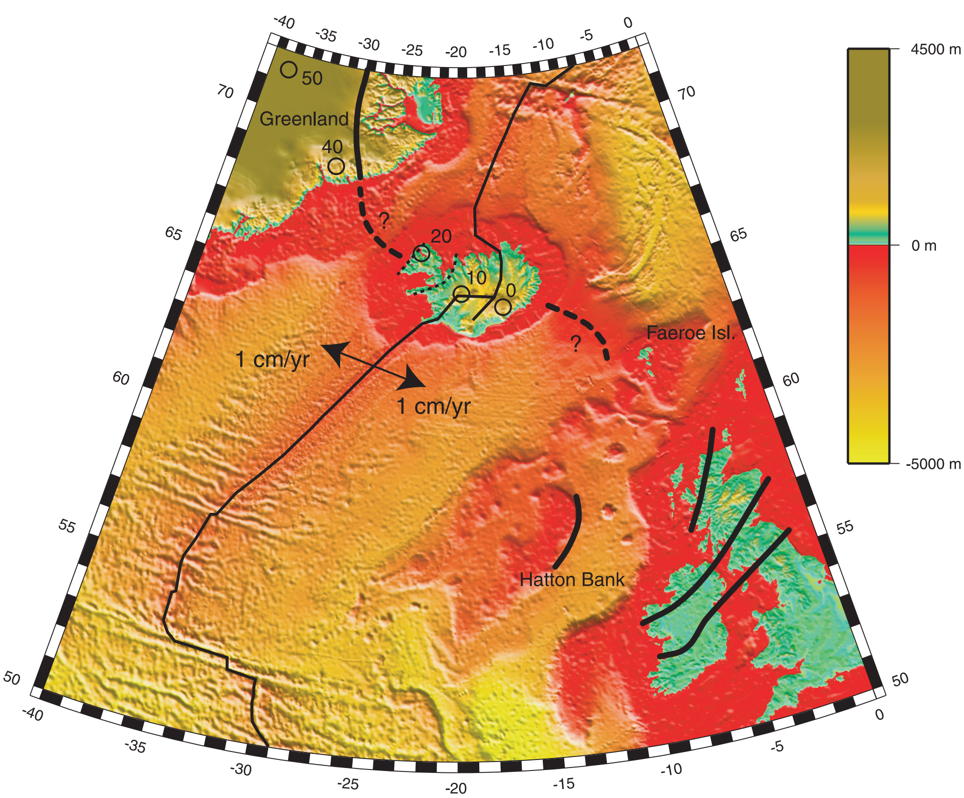

Figure 2. Bathymetry of the north Atlantic region. The shallow bathymetric ridge that crosses the Atlantic ocean from Greenland to the Faeroe Islands and marks the location of thick crust can be seen clearly. Other shallow bathymetric areas (e.g., the Hatton Bank) are blocks of stretched, thinned continental crust. The thin black line indicates the Mid-Atlantic Ridge. Thin dashed lines indicate the locations of extinct ridges in Iceland. Thick lines indicate faults of the Caledonian suture (Soper et al., 1992). The thick dashed line indicates the inferred overall trend of the western frontal thrust where it crosses the Atlantic Ocean (Bott, 1987). Circles indicate the hypothesized locations of an Icelandic mantle plume at the times indicated, which are in m.y. (Lawver and Muller, 1994). Adapted from Foulger and Anderson (2005).

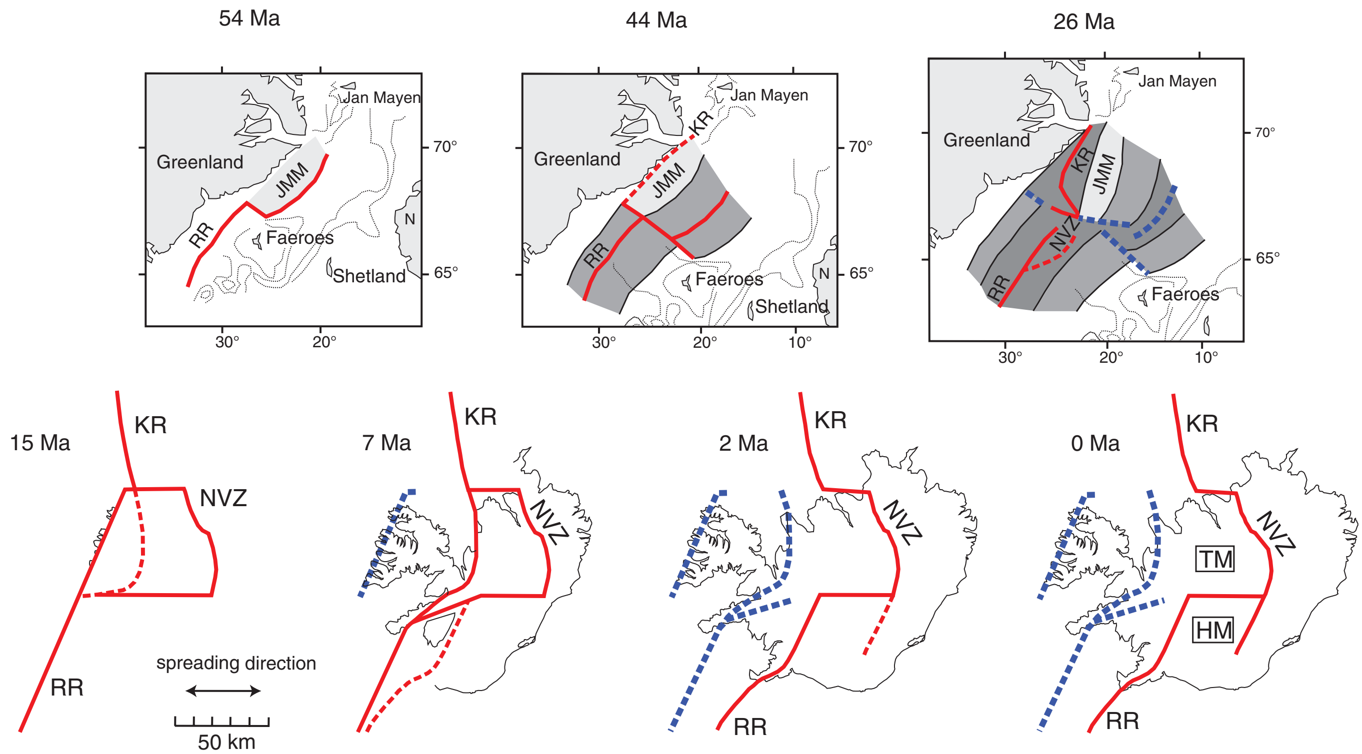

Figure 3. Tectonic evolution of the Iceland region during the past 54 m.y. Light gray depicts continental crust; medium gray shows seafloor that formed 54–44 Ma; dark gray depicts seafloor that formed 44–26 Ma; red lines show currently active plate boundaries; dashed red lines show imminent plate boundaries; dashed blue lines mark extinct plate boundaries; thin lines show bathymetric contours. HM—Hreppar microplate; JMM—Jan Mayen microcontinent; KR—Kolbeinsey Ridge; N—Norway; NVZ—Northern Volcanic Zone; RR—Reykjanes Ridge; TM—Trollaskagi microplate. The top three panels are redrawn from Bott (1985). Adapted from Foulger (2002).

of spreading in northeast Iceland with the Kolbeinsey Ridge remained approximately constant (Foulger, 2002, 2003b, unpubl. data; Foulger and Anderson, 2005).

The Iceland region has been spreading about a parallel pair of ridges for much of the past 17 m.y. and perhaps the past 26 m.y. (Fig. 3) (Bott, 1985; Foulger and Anderson, 2005). In addition to spreading about the dominant axis in northeastern Iceland (represented by the present-day NVZ), spreading also occurred along subsidiary ridges in the North American plate in western Iceland (Figs. 2 and 3). The repeated extinctions of these ridges may be explained by their continual westerly transport with the North American plate, which progressively distanced them from the magma source associated with the Kolbeinesey-Reykjanes axis. The ridges in western Iceland thus repeatedly succumbed to transport off axis and were replaced by new ridges farther east that were more nearly collinear with the marine MAR. Persistent simultaneous spreading about more than one ridge in Iceland is also supported by geochemistry. Dated samples with ages of 7–2 Ma show that in eastern Iceland, La/Sm and $^{87}Sr/^{86}Sr$ ratios decreased through time but in western Iceland, they increased, suggesting that the same mantle was not tapped in the two areas (Schilling et al., 1982). Like the Easter microplate, the Iceland region is a "diffuse oceanic spreading plate boundary" containing two triple junctions (Zatman et al., 2001). Why such plate boundary configurations develop at a few places on Earth is not understood.

Spreading about a parallel pair of ridges is tectonically unstable and expected to result in complex "leaky microplate" tectonics (Foulger and Anderson, 2005). Several enigmatic observations in Iceland may be interpreted in this framework (Fig. 4). These include the variable rates and directions of extension indicated by Global Positioning System (GPS) surveying, the *P*- and *T*-axes of earthquake focal mechanisms, the trends of presently active volcanic zones, and the regionally variable orientations of Tertiary dikes (Saemundsson, 1979; Einarsson, 1991; Hofton and Foulger, 1996; Foulger and Anderson, 2005). Locally variable directions of motion are expected to give rise to spatially variable extension across transverse zones and variations in magma production rates. These variations may explain the increase in volcanic productivity along an easterly oriented zone traversing central Iceland from the Snaefellsnes Volcanic Zone to Vatnajokull (Fig. 4). That zone may represent a long-lived line of weakness that is a composite of various plate-boundary elements. A palinspastic reconstruction of the development of Iceland predicts the existence of a captured oceanic microplate beneath central Iceland, with crust as old as

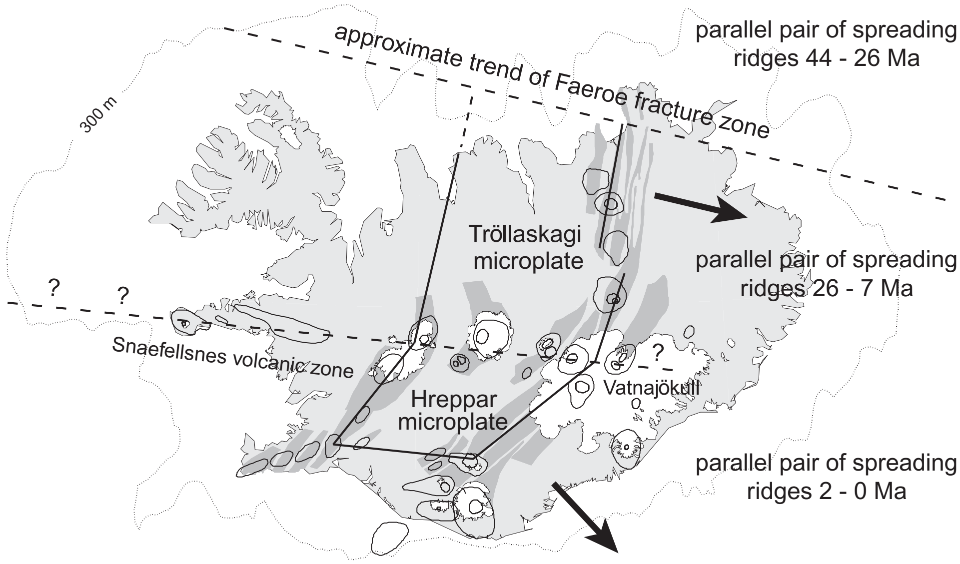

Figure 4. Schematic diagram showing a simplified tectonic map of Iceland. Parallel pair spreading has migrated south during the period 44 Ma to present. The Faeroe fracture zone formed when the north Atlantic opened initially at ca. 54 Ma (upper dashed line). A second transverse zone that may represent a long-lived composite zone of various plate-boundary elements extends from the Snaefellsnes volcanic zone across central Iceland and into Vatnajökull (lower dashed line). Solid black lines suggest locations of other boundaries of the Tröllaskagi and Hreppar microplates. The evolution of these microplates may be understood with reference to Figure 3. Arrows show local directions of motion deduced from Global Positioning System surveying, earthquake focal mechanisms, the trends of presently active volcanic zones, and the orientations of Tertiary dikes (for details, see Foulger and Anderson, 2005).

ca. 30 Ma, that has been buried by later eruptives (Fig. 3) (Foulger, unpubl. data; Foulger and Anderson, 2005). Kinematically, this microplate could have trapped a sliver of continental crust comprising a southerly extension of the Jan Mayen microcontinent, which currently lies offshore to the northeast of Iceland (Fig. 3; Amundsen et al., 2002).

CRUSTAL STRUCTURE

The seismic structure of the crust beneath Iceland and the MAR is shown in Figure 5. The thickness deduced from seismology pivots on whether Layer 4 is interpreted as upper mantle or lower crust (see Björnsson et al., this volume, for a critical review). Layer 4 is taken here to correspond to the layer in which the shear wave speed V_s = 3.7–4.2 km/s and the compressional wave speed V_p = 6.6–7.5 km/s, assuming V_p/V_s = 1.78, after Foulger et al. (2003). Other authors use slightly different assignations, but this does not affect the first-order results. If Layer 4 is interpreted as upper mantle, then Figure 5A represents a map of crustal thickness, which is typically 7–10 km beneath Iceland. If Layer 4 is interpreted as lower crust, then Figure 5B represents crustal thickness, which is 20–40 km, and typically ~30 km. Both models have been cited to support the plume hypothesis. One model postulates that Layer 4 is mantle peridotite that has an unusually low seismic wavespeed because it is hot and partially molten (the thin, hot crustal model). The other postulates that Layer 4 is gabbroic and represents the anomalously large volume of melt expected to be produced by a plume (the thick, cold model).

Both interpretations have unresolved disagreements with other observations and difficulties with the tectonic models implied. In oceanic regions, the seismic crustal thickness is presumed to be an approximate proxy for melt thickness, although analogies with ophiolites and arguments for melt retention in the mantle suggest that it may be an underestimate (Christensen and Smewing, 1981; Cannat, 1996).

The thin, hot crustal model requires substantial percentages of partial melt in Layer 4 that are not detected by seismic attenuation or V_p/V_s measurements (Menke and Levin, 1994) but

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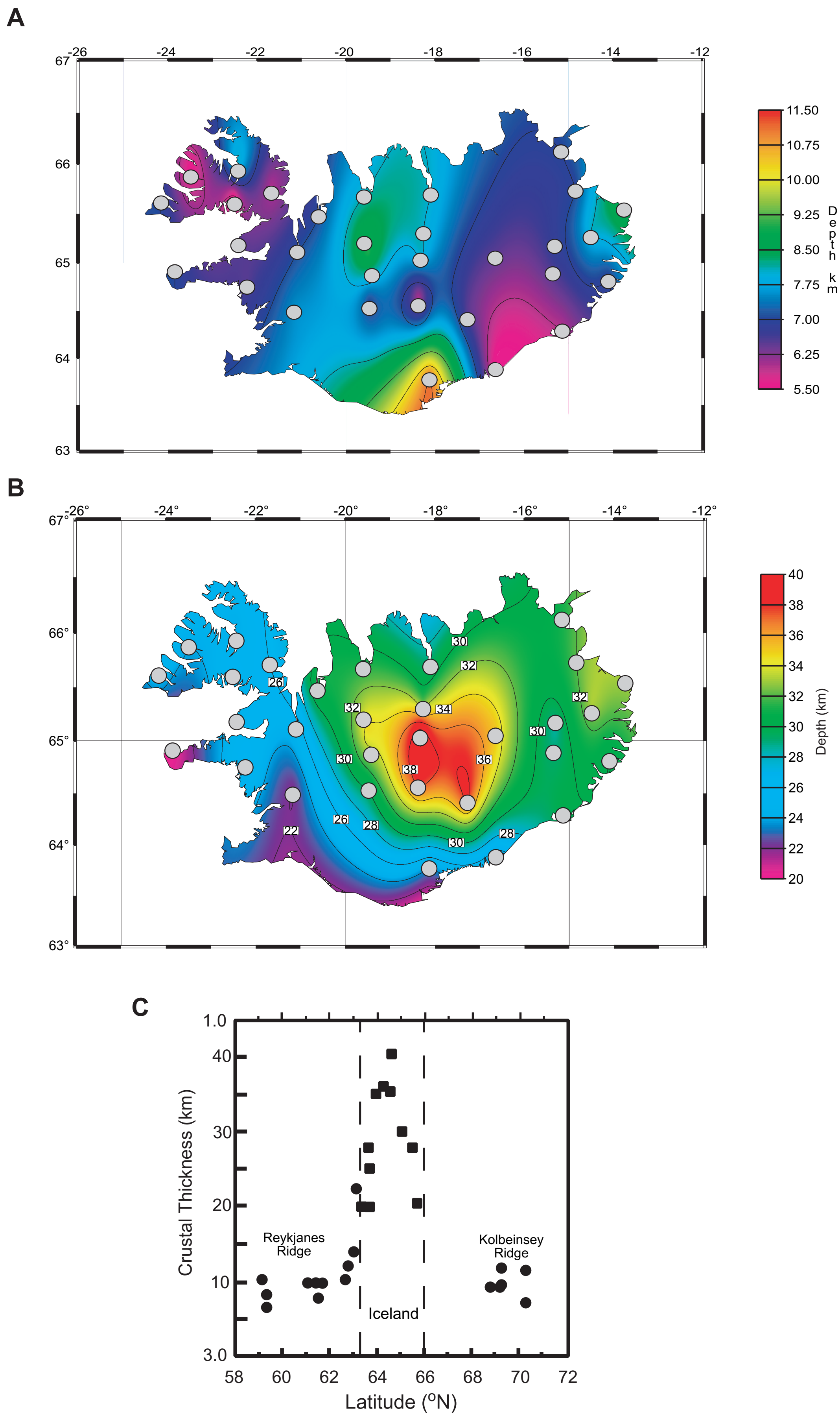
Figure 5. (A) Contour map showing the depth to the base of the upper crust (defined as the depth to the V_s = 3.7 km/s horizon, and approximately equivalent to the top of Layer 4), from seismic receiver function results (from Foulger et al., 2003). Gray circles indicate seismic stations. (B) Contour map showing the depth to the base of the lower crust (defined as the depth to the V_s = 4.2 km/s horizon). (C) Crustal thickness versus latitude, from a compilation of seismic experiments in Iceland and the north Atlantic. Circles indicate oceanic measurements and squares indicate measurements made on land (adapted from Foulger et al., 2004).



A
-26
-24
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67
66
65
64
63
11.50
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8.50
7.75
7.00
6.25
5.50
Depth km
B
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67°
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64°
63°
40
38
36
34
32
30
28
26
24
22
20
Depth (km)
C
1.0
40
30
20
10
3.0
Crustal Thickness (km)
Reykjanes Ridge
Iceland
Kolbeinsey Ridge
58
60
62
64
66
68
70
72
Latitude (°N)

which provide an explanation for the extensive low-resistivity layer detected beneath Iceland by magnetotelluric measurements (Björnsson et al., this volume). What the relationship would be with the Greenland-Iceland and Iceland-Faeroe ridges, which both have thick seismic crust but are not expected to be hot, is unclear. Unlike Iceland, the Iceland-Faeroe Ridge is underlain by a clear seismic Moho and is thus less ambiguously associated with ~30-km-thick crust (Bott and Gunnarsson, 1980; Staples et al., 1997). The thin, hot crustal model might thus suggest that Iceland is not simply a broader continuation of the Iceland-Faeroe Ridge but an entirely different structure. If correct, the thin, hot crustal model implies that the crust beneath Iceland is no thicker than that beneath the surrounding ocean basins (Fig. 5C) and a magmatically productive plume is not supported.

The thick, cold crustal model also poses problems. First, isostatic calculations require that the density of the ~20-km-thick "lower crust" (Layer 4) is only ~90 kg/m^3 less than the average uppermost mantle density of ~3300 kg/m^3 (Menke, 1999; Gudmundsson, 2003). For a gabbroic lower crust, a density contrast of 250–300 kg/m^3 would be expected. If Layer 4 had normal gabbroic densities, Iceland would have an elevation of ~4 km, very different from its actual ~1-km elevation. A possibility that has not been fully explored is that the lower crust may contain oxide gabbro, which might combine gabbrolike seismic wavespeeds with a relatively high density. A downward gradation from a mostly gabbro composition to one that is mostly peridotite with subsidiary retained crystallized melt could explain the density observations and would imply that only about half of Layer 4 represents melt. This structure does not, however, explain the low seismic wavespeeds, which are typical of gabbro. Having said this, a mixture of crystallized melt and mantle residuum that fits the density observations would imply that approximately twice the amount of melt produced at the adjacent MAR is produced at Iceland (Foulger et al., 2003). A second problem raised by the thick, cold crustal model is that it requires Layer 4 to be below the gabbro solidus and thus relatively cool—cooler than the Earth at similar depths in the mantle beneath the East Pacific Rise (Menke and Levin, 1994). It also offers no explanation for the low-resistivity layer detected by magnetotelluric measurements. The thick, cold crustal model is more consistent with the plume hypothesis than is the thin-crust model, as it implies that up to three times as much melt is produced at Iceland than along the neighboring MAR.

For the thick, cold crustal model, how well do the apparent variations in thickness across Iceland fit the predictions of the plume model? A maximum crustal thickness of ~40 km is suggested beneath central Iceland, thinning to as little as ~20 km toward the coast in the west and south (Fig. 5B). For a plume migrating from northwestern to southeastern Iceland, a trail of thick crust would be expected in its wake. On the contrary, the crust is thinner beneath western Iceland than beneath eastern Iceland. Offshore, the crust is equally thick beneath the Iceland-Faeroe Ridge, ahead of the present location of the postulated plume, as it is beneath the Greenland-Iceland Ridge (Bott and Gunnarsson, 1980; Staples et al., 1997; Holbrook et al., 2001), which lies behind. If a plume were persistently ridge-centered (and thus not fixed with respect to other Indo-Atlantic hotspots), a symmetric, 40-km-thick belt of crust traversing Iceland from east to west would be expected. This is not observed either.

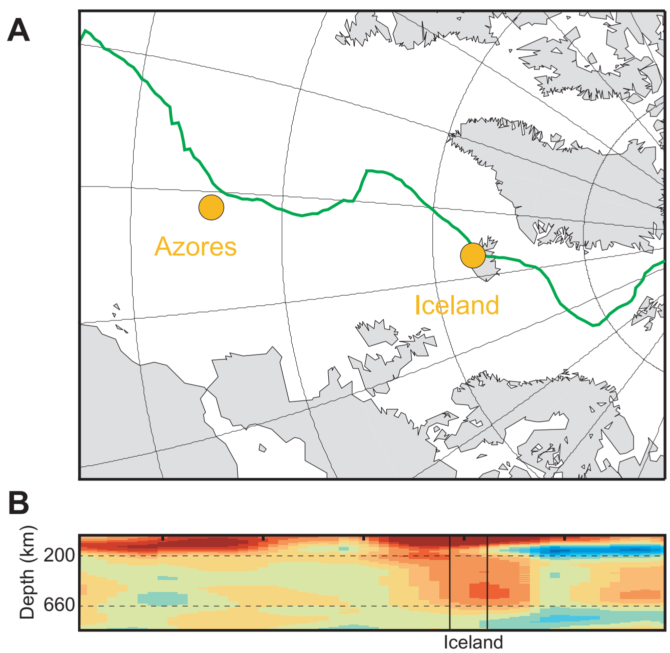
If the thin, hot crustal model is correct, it would imply that there is no significant volumetric melt anomaly at Iceland. The geological and geophysical differences between Iceland and the neighboring spreading ridges to the north and south would then result solely from the subaerial eruptive environment of Iceland, which is a local consequence of the regional bathymetric anomaly. If the thick, cold crustal model is correct, a nonthermal model would require a mantle source with enhanced fusibility, a possibility that we discuss below. Explanations for the variations in crustal thickness would then be expected in the complex history of spreading in the region.

MANTLE STRUCTURE

The objective of most experiments studying the structure of the mantle beneath the Iceland region has been to assess the dimensions of the presumed mantle plume. The depth of the anomaly is then critical. The classical plume model predicts that structures extend throughout both the upper and lower mantles, a feature that was invoked to explain the relative fixity of some hotspots (Morgan, 1971).

Whole-mantle tomography shows that the north Atlantic between the Charlie Gibbs and Jan Mayen fracture zones is underlain by a low-wavespeed anomaly that all studies with good resolution agree extends down only to the mantle transition zone (Fig. 6) (Ritsema et al., 1999; Foulger et al., 2000, 2001; Megnin and Romanowicz, 2000; Montelli et al., 2004; Pilidou et al., 2004; Ritsema, this volume). At depths greater than ~200 km, the low-wavespeed anomaly is elongate in a direction parallel to the MAR, taking on the shape of a vast dike. This morphological change is discernable in whole-mantle tomography, teleseismic tomography images, and surface-wave studies. The truncation of the anomaly at the base of the upper mantle is not consistent with what is expected for a classical plume. Furthermore, the nonaxisymmetric morphology of the deeper part of the anomaly and its elongation parallel to the continental margins and the MAR suggest rather a relationship with the regional morphology and tectonics of the north Atlantic ocean basin (Foulger et al., 2000, 2001).

A tomographic cross-section illustrating a continuous, low-wavespeed body extending from the surface to the core-mantle boundary beneath Iceland was produced by Bijwaard and Spakman (1999) (Fig. 6E). This image of an apparently plumelike body traversing the whole mantle was achieved by two means: (1) The color scale was saturated at an anomaly strength of $\Delta V_p = 0.5\%$, only ~10% of the maximum anomaly strength in the upper mantle. Such a procedure imparts the visual impression of continuity between strong anomalies in the upper mantle and weak anomalies in the lower mantle. (2) The line of section



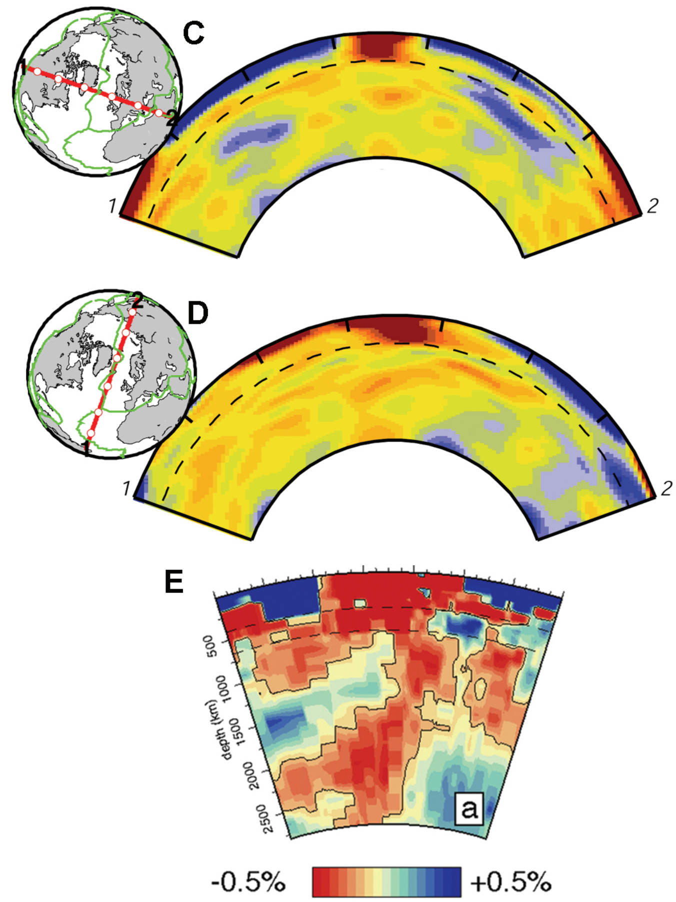


Figure 6. (A) Map of the north Atlantic showing land masses, the Mid-Atlantic Ridge (green line) and the location of the Iceland and Azores hotspots (yellow dots). (B) Whole-mantle tomographic cross-section running along the ridge. Tick marks are spaced at intervals of 1000 km. Black vertical lines indicate the extent of the Icelandic landmass. The color scale is saturated at $\Delta V_s = 2.5\%$ (adapted from Montagner and Ritsema, 2001). (C) East-west cross-section through the entire mantle passing through Iceland. Inset shows the line of the section (red dashed line). Green lines show plate boundaries. The color scale is saturated at $\Delta V_s = 3\%$ (from the model of Ritsema et al., 1999). (D) Cross-section through the entire mantle passing through Icelend, with the line of the section along the ridge (red dashed line in inset). (E) Whole-mantle cross-section along a similar line as (C) (from the model of Bijwaard and Spakman 1999). The color scale is saturated at $\Delta V_p = 0.5\%$.

was truncated to remove similar weak downward-continuous anomalies beneath the Canadian shield and Scandinavia, where plumes are not expected. The study of Bijwaard and Spakman (1999) did not have adequate resolution to detect bodies in the lower mantle beneath Iceland of strength similar to the upper-mantle anomaly (Foulger et al., 2001; R. van der Hilst, 2001, personal commun.), and the weak lower-mantle anomalies detected and proposed to represent an Icelandic plume extending down to the core-mantle boundary are not confirmed by other studies (e.g., compare Fig. 6C and E).

Topography of the discontinuities bounding the transition zone also bears on the depth extent of the mantle anomaly beneath Iceland. A hot conduit extending from the lower mantle up into the upper mantle, traversing the transition zone, is predicted to warp the 410-km discontinuity down and the 660-km discontinuity as a result of the different signs of the Clapeyron slopes associated with these mineralogical phase transitions (Bina and Helffrich, 1994). The topography of the transition-zone discontinuities beneath Iceland has been investigated in very high-quality receiver-function analyses involving large combined datasets from three separate broadband seismic networks (Shen et al., 1996, 1998, 2002; Z. Du, L.P. Vinnik, and G.R. Foulger, 2004, unpubl. data). Only one of the effects predicted by the plume hypothesis occurs, namely, downwarping of the 410-km discontinuity.

All studies of the transition zone agree that it has a normal thickness of ~250 km beneath the whole island except for the south-central part, where it is 15–20 km thinner. This thinning was interpreted by Shen et al. (1996, 1998, 2002) as resulting from deflections on both the 410- and the 660-km discontinuities of a kind consistent with elevated temperature, and supporting a plume with a temperature anomaly of ~150 K rising from the lower mantle. The studies of Shen et al. (1996, 1998, 2002) have weaker control on the topographies of the separate discontinuities than on the total transition zone thickness. For this reason, the problem was revisited by Z. Du, L.P. Vinnik, and G.R. Foulger (2005, unpubl. data), who studied the discontinuities separately. They used stacks of up to approximately two hundred receiver functions from seismic waves rising obliquely beneath Iceland and penetrating the two discontinuities separately inside and outside south-central Iceland. The results showed that the thinning of the transition zone is due to depression of the

410-km discontinuity, but that the 660-km discontinuity is flat within the resolution possible. Receiver function studies of the transition zone have a lateral spatial resolution of ~200 km and are able to resolve deflections on the 660-km discontinuity of ~5 km. This magnitude of deflection corresponds to a temperature anomaly of ~50 K if deflections are interpreted solely in terms of temperature, or zero if reasonable compositional variations are allowed (Presnall, 1995).

Although topography on the transition-zone discontinuities is commonly interpreted in terms of temperature alone, composition (e.g., FeO, H_2O contents) and mineralogy (e.g., garnet, clinopyroxene contents) can also cause them to vary in shape, depth, and thickness. Normal variations in Mg# of mantle peridotite from ~88 to ~92 could account for up to half the observed topography on the 410-km discontinuity if the mantle beneath south-central Iceland were more magnesium-rich (i.e., more depleted) than the surrounding mantle (Presnall, 1995). Such a compositional variation might be expected if melt is mined from such great depths beneath Iceland. The observed depression of the 410-km discontinuity could thus be interpreted either as a temperature anomaly of ~150 K or as one of ~75 K combined with a depleted mantle composition. A temperature anomaly of a few tens of K cannot be ruled out at a depth of 660 km, but there is no evidence for one.

The strengths of seismic wavespeed anomalies beneath Iceland are commonly interpreted in terms of temperature alone. Estimates for the temperature derivatives in the mantle suggest that an anomaly of ~100 K would depress seismic wave speed by roughly V_p ~1% and V_s ~2% in the shallow mantle (Goes et al., 2000). Teleseismic tomography in Iceland suggests that the anomaly is strongest beneath central Iceland, where its strength relative to coastal regions is up to ~5% in V_s in the upper ~200 km and 1–2% below this (Wolfe et al., 1997; Foulger et al., 2000, 2001). P and S receiver-function studies show that teleseismic tomography underestimates anomaly strengths by about a factor of two and so the true anomaly strength may be up to V_s ~10% (Z. Du, L.P. Vinnik, and G.R. Foulger, 2005, unpubl. data; Vinnik et al., 2005). This strength would correspond to a temperature anomaly of up to ~500 K, an unrealistically high value that is not supported by other work (Table 1). The inescapable conclusion is that at least some of the anomaly is due to partial melt, which can depress V_s by up to ~8% per % melt (Goes et al., 2000). The strongest part of the shallow anomaly could thus be interpreted as <1% of partial melt, and much less in the deeper parts, which may not be extractable. The observations are also consistent with no temperature anomaly at all if the mantle has an unusually low solidus. The anomaly weakens with depth such that if interpreted solely in terms of temperature (Karato, 1993), it corresponds to a body that is cooler toward its base than at its top.

In summary, the structure of the mantle beneath Iceland supports the plume hypothesis insofar as a low-wavespeed anomaly exists there. However, anything beyond this first-order observation involves significant difficulties with the plume interpretation. The low-wavespeed anomaly fills a large region of the north Atlantic, and is not confined to the mantle immediately beneath Iceland. Multiple high-quality seismic experiments using a variety of methods find no evidence that it extends into the lower mantle, but instead show that it weakens downward and terminates in the transition zone (Ritsema et al., 1999; Foulger et al., 2000, 2001; Montelli et al., 2004; Ritsema, this volume). It is not axisymmetric but its morphology mirrors that of the north Atlantic continental margins and the MAR. Estimates of temperature anomaly based on the topography of the transition-zone discontinuities and the magnitudes of seismic wavespeed anomalies permit no more than ~150 K at transition-zone depths, require no more than a few tens of K in the presence of reasonable compositional variations, and require no temperature anomaly at all at the bottom of the transition zone. At shallow depths, wavespeed anomalies are so strong that partial melt is required, which renders inescapable the fact that, in the extreme, all of the anomaly could be explained as partial melt.

TEMPERATURE AND HEAT

How Large a Temperature Anomaly Is Required?

One of the few characteristics of thermal plumes that cannot be negotiated is that of high temperature. To rise through thermal buoyancy, an anomaly of at least 200–300 K is required, even for the weakest upper-mantle plume (Courtney and White, 1986; Sleep, 1990, 2004). Deep thermal boundary layers have temperature contrasts of at least 1000 K and in plume theory, these thermal anomalies must be transported quickly into the upper mantle. Absolute mantle temperature and temperature anomalies in the north Atlantic have been explored using seismology, petrology, heat flow, bathymetry, uplift, and subsidence of the crust (Table 1 and Fig. 7).

Seismology

Crustal seismology, including both attenuation and V_p/V_s, indicates that the crust is relatively cool, or less than ~900° C down to its base at 30–40 km (Menke and Levin, 1994; see also Foulger et al., 2003 for a review). This is cooler than at equivalent depths beneath the East Pacific Rise. In the mantle, seismic results yield extremal estimates of 0–500 K for the temperature anomaly in the upper ~200 km, depending on whether partial melt is invoked (Foulger et al., 2001; Vinnik et al., 2005). The very high estimates are unreasonable and at odds with all other results, so partial melt is required to explain those observations (Vinnik et al., 2005). The data suggest that any temperature anomaly beneath Iceland decreases with depth, and that temperature anomalies are at most a few tens of K at the bottom of the transition zone.

Petrology and Olivine-Glass MgO-FeO Partitioning

Petrological estimates of the temperature and temperature anomaly at Iceland have been made using several geothermo-

TABLE 1. TEMPERATURE ESTIMATES IN THE ICELAND REGION

Method	Potential temperature (°C) or temperature-anomaly (K) estimate	Depth (km)	Reference
Seismology			
Seismic attentuation and V_p/V_s	Less than ~900° C in Layer 4 (if assigned to lower crust)	20–40	Menke and Levin (1994); Menke et al. (1995, 1996)
S receiver functions	~50 K (relative to 1350° C adiabat)	80–130	Vinnik et al. (2004)
Global and teleseismic tomography	Up to ~200 K or 0 K and ~0.5% partial melt (relative to mid-ocean ridges)	Less than ~200	See Foulger et al. (2001) for a review
Absolute travel-time delays	Up to 500 K or 0 K and ~1% partial melt (relative to average Earth—IASP91)	Less than ~400	Du et al. (unpubl. data)
Global and teleseismic tomography	~100 K or 0 K and ~0.15% partial melt (relative to mid-ocean ridges)	200–400	See Foulger et al. (2001) for a review
P receiver functions	~150 K or 75 K and a compositional anomaly of ~4 in Mg# (relative to average Earth—IASP91)	~410	Shen et al. (2002); Du et al. (unpubl. data)
P receiver functions	0 K (relative to average Earth—IASP91)	~660	Du et al. (unpubl. data)
Petrology			
Olivine-glass thermometry	1270° C (~0 K relative to mid-ocean ridges)	~50	Breddam (2002)
Melt inclusions	1300 ± 26° C (100 K, relative to an assumed primitive MORB source of 1230° C)	~50	Foulger et al. (this paper)
CMASNF geothermometer and high-MgO glasses	1240–1260° C (~0 K relative to mid-ocean ridges)	~50	Gudfinnsson et al. (2003)
Major element systematics of Icelandic MORB	0 K (relative to mid-ocean ridges)	~50	Gudfinnsson et al. (2003)
Other			
Bathymetry of the north Atlantic	~70 K (relative to "background")	Shallow upper mantle	Ribe et al. (1995)
Subsidence of ocean crust	50–100 K (relative to "background")	Shallow upper mantle	Clift (1997, this volume)
Uplift of Hebrides shelf	100 K (relative to "background")	Shallow upper mantle	Clift et al. (1998)
Heat flow	<200 K (relative to global average)	Shallow upper mantle	Stein and Stein (2003); DeLaughter et al. (this volume)

Notes: CMASNF—CaO-MgO-Al_2O_3-SiO_2-Na_2O-FeO system; IASP91—International Association of Seismology and Physics 1991; MORB—mid-ocean ridge basalt.

metric approaches. Estimates of the average potential temperature of the mantle beneath (the temperature that mantle would have if it ascended adiabatically from its source to the Earth's surface without melting) range from ~1240 to 1600° C, depending on the methodology used, the location, and estimates of eruption temperature using olivine-liquidus relationships (e.g., Anderson, 2000; Herzberg and O'Hara, 2002; Presnall et al., 2002). Potential temperatures for mantle beneath the early north Atlantic Igneous Province based on picrites at Baffin Bay, western Greenland, are at the high end of this range (Larsen and Pedersen, 2000). In investigating the temperature anomaly at an individual location, such as Iceland, it is thus most meaningful to consider differences in temperature compared with some measure of the average mantle estimated using the same methodology. In the case of petrology, comparisons with estimates for "normal" mid-ocean ridges (MORs) are possible.

The bulk, or major-oxide, composition of Icelandic basalts overlaps that of normal mid-ocean ridge basalt (N-MORB), and temperatures similar to those of most N-MORB are indicated by petrological geothermometers. This observation extends to estimates of the eruptive temperatures of primitive lavas from central Iceland, where the crust is thickest and the center of a plume is commonly assumed to lie (Breddam, 2002). The maximum temperature estimated for those glasses, 1240° C, is corroborated by major-element systematics (Presnall et al., 2002) and application of the CaO-MgO-Al_2O_3-SiO_2-Na_2O-FeO geothermometer to high-MgO glasses (Gudfinnsson et al., 2003).

An exception to this general picture may lie in temperature estimates based on olivine-glass MgO-FeO partitioning. Thus both Larsen and Pedersen (2000) and Herzberg and O'Hara (2002) infer melt temperatures >1400° C for calculated parental picritic liquids estimated to be in equilibrium with olivine of composition Fo91 found in Paleocene picrites from Baffin Bay, western Greenland, where the Iceland plume is considered to

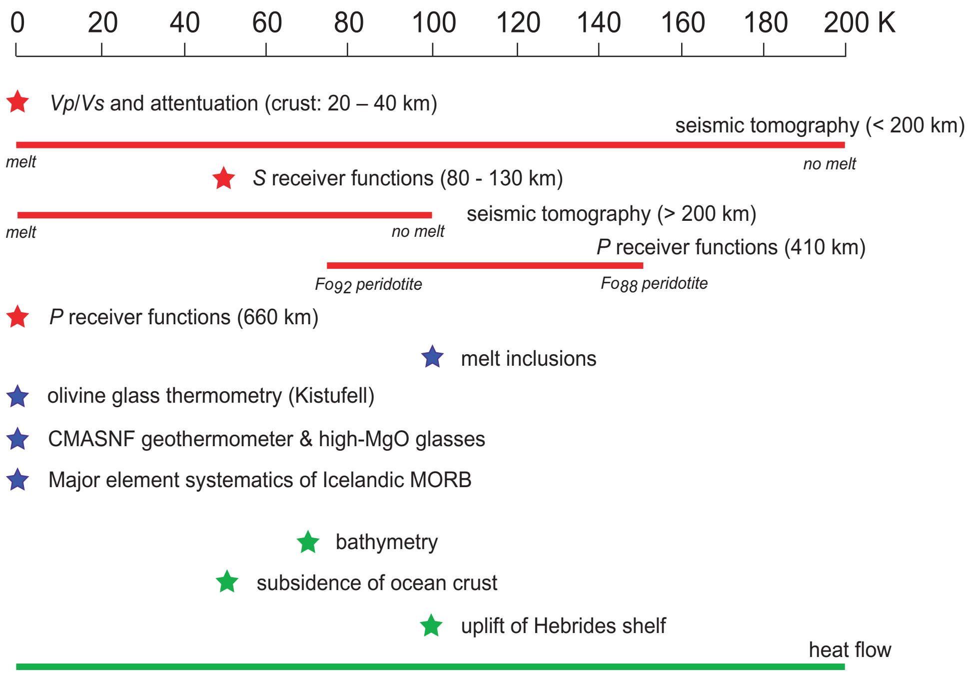


Figure 7. Estimates of temperature anomalies for the Iceland region. Bars correspond to temperature ranges and stars to individual temperature anomaly estimates. Temperature anomalies are relative to the average Earth, the north Atlantic away from Iceland, or "normal" mid-ocean ridges, depending on the study. CMASNF: CaO-MgO-Al_2O_3-SiO_2-Na_2O-FeO system. See Table 1 for sources and details.

have first impacted the lithosphere. This value indicates a potential temperature anomaly of ~180 K relative to MORs. The procedure of Herzberg and O'Hara (2002) is based on the assumption that the liquids hosting such forsteritic olivine and the olivine itself lie along a common olivine–controlled liquid line of descent. If the assumption is valid, then their procedure is simply to add incremental amounts of successively more magnesian olivine to successively estimated liquid compositions until a liquid in equilibrium with olivine of composition Fo90–Fo91 is obtained. Thus for picrites from Baffin Bay, Hawaii, and Gorgona, parental liquids are estimated to contain 16–21% MgO.

However, we question whether a common liquid line of descent has been, or can be, demonstrated for Baffin picrites, as those are all rocks that have accumulated olivine, and their host glasses have only 7.1–8.5% MgO (Larsen and Pedersen, 2000). Similarly, picritic glass is not found in Iceland, the most magnesian glass thus far discovered having only 10.6% MgO (Breddam, 2002). Careful study of picrite mineralogy often reveals that picrites are hybrids between some differentiated (i.e., relatively iron-rich) magma and an array of scavenged, porphyritic, primitive magma types incorporated at depth that crystallized two, three, or even more populations of olivine and associated Cr-spinel. This composition has been shown for eastern Pacific MORB (Natland, 1989), for tholeiites of the Juan Fernandez Islands (Natland, 2003b), and for Hawaiian tholeiites (Clague et al., 1995). It is also evident from ranges in olivine phenocryst compositions in Baffin picrites (i.e., Fo84.5–Fo92.8 in one sample; Larsen and Pedersen, 2000), olivine in a single Iceland flow at Borgarhraun (Fo80–Fo92.5; Maclennan et al., 2003b), and spinel compositions in primitive basalts from the Theistareykir region of Iceland (Mg# = 0.8–0.5 in phenocryst cores, to 0.3 in rims, of single samples; Sigurdsson et al., 2000).

Typical Icelandic basalts follow a general iron-enrichment trend of differentiation (Fig. 8), one similar to MORB in being controlled by polyphase crystallization of plagioclase, clinopyroxene, and olivine in approximately decreasing order of importance. Picritic liquids are argued to follow a simple olivine-controlled line of descent in which iron-enrichment is barely evident or does not occur at all. Using whole-rock compositions, such trends are overlapped strongly by compositions resulting from olivine accumulation (e.g., Theistareykir basalts with 10–25% MgO content in Fig. 8). However, the presence in picrites of olivine less forsteritic than about Fo85 and spinel with Mg# less than ~0.6 is a fair indication that they are hybrid lavas, and that mixing involved at least one relatively iron-rich end-member that crystallized along a polyphase cotectic with strong iron enrichment. Given the complexities of conduit and rift systems, and that liquids plus crystals are combined (i.e., they are magmas), mixing involving only one differentiated magma and one primitive magma is unlikely. From mineral data, a differentiated aggregate of magmas usually mixes with another primitive aggregate of magmas, and such mechanical processes as flowage differentiation and mineral sorting then occur to produce a picrite. To the extent that the differentiated aggregate contributes to a bulk liquid composition, the resulting hybrid will have higher FeO at a given MgO content than any uncontaminated olivine-controlled primitive aggregate with which it is mixed. Adding olivine incrementally to the hybrid composition using the procedure of Herzberg and O'Hara (2002) thus almost always will result in artificially high estimates of MgO content of liquids

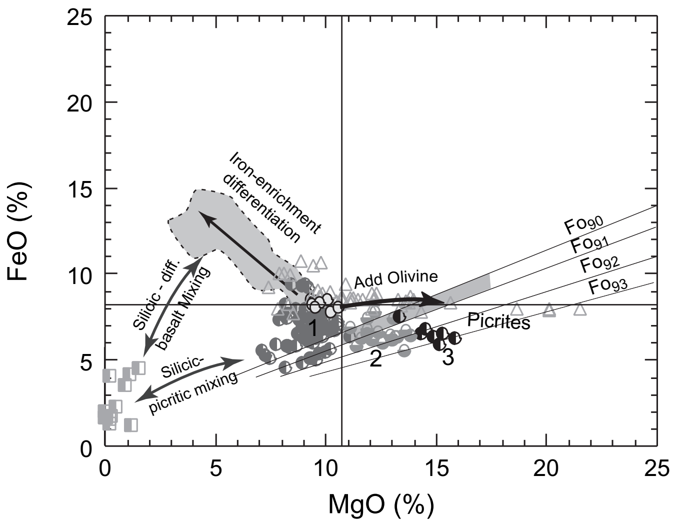


Figure 8. MgO versus FeO for Iceland glass and whole-rock compositions. FeO = 0.9 × FeOT, where FeOT is the total iron present as FeO, as measured by electron-probe microanalysis. The coefficient assumes $Fe^{2+}/[Fe^{2+} + Fe^{3+}] = 0.9$, an average ratio in mid-ocean ridge basalt (Christie et al., 1986). Basalt glasses are shown in the light gray field with a dashed border, with an arrow indicating direction of iron-enrichment differentiation (data from Sigurdsson and Sparks, 1981; Meyer et al., 1985; and Trønnes, 1990). Primitive glasses from Kistufell, black-rimmed gray-filled open circles (Breddam, 2002). Glass inclusions (Hansteen, 1991; Gurenko and Chaussidon, 1995; Sigurdsson et al., 2000; Slater et al., 2001) are divided into three groups (see text): Group 1, medium-gray left-half-filled circles; Group 2, light-gray bottom-half-filled circles; Group 3, black left-half-filled circles. Whole-rock analyses of basalts from Theistareykir are shown as open gray triangles (Stracke et al., 2003a). Arrow among those symbols indicates the effect of addition of olivine to Icelandic olivine tholeiite, as in the Theistareykir range of compositions, out to high MgO content. Rhyolites are depicted by half-filled gray squares (Gunnarsson et al., 1998). Isopleths of constant olivine composition, shaded between Fo90 and Fo91, are from Herzberg and O'Hara (2002). The horizontal and vertical lines intersect at the composition of the most magnesian Iceland basalt glass from Kistufell, central Iceland. Possible mixing trends between basaltic liquids and silicic melts, in the mantle or in the crust, are indicated by double-headed arrows.

in equilibrium with olivine of Fo90–Fo91 composition (arrow labeled "add olivine" in Fig. 8), and therefore in estimates of both crystallization and potential temperatures that are too high.

The primitive magma aggregate adds further complexity. Most Iceland picrites, in fact, have olivine (Fo86–Fo92), ubiquitous Cr-spinel, at least some plagioclase (to An90), but in some cases, they also contain clinopyroxene phenocrysts (Hansteen, 1991; Gurenko and Chaussidon, 1995; Breddam et al., 2000; Slater et al., 2001; Maclennan et al., 2003a). In some cases, the minerals are intergrown. However, olivine is so much more abundant than the other minerals that they are far from being present in cotectic proportions; thus the general mechanism of olivine concentration must be mechanical, such as density sorting during gravitational settling and flowage differentiation. Hansteen (1991) infers that the mineral assemblage of picrites from the Reykjanes Peninsula is scavenged from primitive gabbroic and pyroxenitic assemblages in the lower Icelandic crust. When full descriptions of Icelandic picrites and the compositions of all the minerals they contain are provided, then it is clear that many of the phenocrysts in them have no relationship to the minerals that actually would be in equilibrium with obviously hybrid host liquids. Indeed, because of the common presence of plagioclase and clinopyroxene phenocrysts, no Icelandic tholeiite, not even a picrite, can yet be said to belong to a strictly olivine-controlled liquid line of descent. The procedure of Herzberg and O'Hara (2002) therefore cannot be applied to these rocks. These mineralogical attributes are in common with many MORB (e.g., Natland et al., 1983), as is the restriction of glass compositions to nonpicritic MgO contents (Breddam et al., 2000; Gudfinnsson et al., 2003).

What of glass inclusions in minerals in the picrites? The studies of Hansteen (1991), Gurenko and Chaussidon (1995), Sigurdsson et al (2000), and Slater et al. (2001) allow distinction of three different groups of inclusions: (1) those with 7–11% MgO and 14.5–18.5% Al_2O_3; (2) those with 10–14% MgO, 13–15% Al_2O_3, and CaO/Al_2O_3 = 1.1–1.5; and (3) those with 11–15% MgO, <13% Al_2O_3, and CaO/Al_2O_3 = 0.9–1.1. In Figure 8, Group 1 compositions lie just below the high-MgO end of the multiphase cotectic trend of iron-enrichment differentiation of Icelandic glasses found in pillows and hyaloclastites. Groups 2 and 3 plot to the right of a vertical line and below a horizontal line drawn through the data point for the most magnesian glass from Kistufell. Thus we see that all the inclusions have lower FeO content than do the glass rims of picritic pillow lavas, including the most magnesian Kistufell glasses.

Accumulation of olivine in basalts from Theistareykir produces a trend of increasing MgO content that systematically has 1–2% higher FeO content than that of Group 2 or Group 3 inclusions, including those in the same basalts (Slater et al., 2001). This trend is, as discussed previously, the consequence of mixing between primitive and differentiated magma aggregates followed by olivine accumulation. The inclusions occur not only in olivine (Fo87–Fo92), but also in calcic plagioclase (An84.8–An89.2), clinopyroxene, and a wide range of compositions of Cr-spinel. The inclusions are not related by olivine controlled differentiation. Based on the high CaO/Al_2O_3 ratios yet different Al_2O_3 contents of Groups 2 and 3 melt inclusions in Cr-spinel, Sigurdsson et al. (2000) concluded that they are not partial melts of lherzolite but instead derive from partial melting of pyroxenite and wehrlite, respectively. We provide a new interpretation below.

From Figure 8, Group 2 and 3 inclusions with 10–14% MgO could crystallize olivine ranging in composition from Fo90 to Fo91 at temperatures <1300° C, using the glass geothermometer of Beattie (1993). There is no need to add olivine incrementally to any of them to compute a parental liquid. Glass inclusions that are more magnesian than this appear in Figure 8 to be in equilibrium with olivine more forsteritic than Fo92, which is rare in Iceland and more forsteritic than most olivine hosting those inclusions (i.e., Slater et al., 2001). Indeed, Figure 8 in the main confirms that most inclusions are in equilibrium with Fo89–Fo92 olivine—that which is found in some picrites (Maclennan et al., 2003a). Nevertheless, melt inclusions tend readily to reequilibrate with surrounding olivine, in the process becoming more magnesian (Gaetani and Watson, 2002), and this may have happened to some, particularly the Group 3, Icelandic inclusions. Others may have MgO contents too high because of partial dissolution of adjacent minerals during heating-stage reequilibration, particularly those in Cr-spinel that contain as much as 1.1% Cr_2O_3 (Sigurdsson et al., 2000). Therefore, values of 14–16% MgO content for these inclusions (which appear in Fig. 8 to be in equilibrium with olivine more forsteritic than Fo92) are probably spurious; no confidence can be placed in temperatures estimated for them by the raw application of geothermometry. We conclude that the highest likely liquidus temperature for Icelandic melt inclusions at 1 atmosphere, using the geothermometer of Beattie (1993), is ~1300° C. This is in good accord with estimates of crystallization temperatures (1260–1280° C) of olivine and clinopyroxene phenocrysts and of these minerals found in portions of cognate xenoliths in basalts from Theistareykir (Maclennan et al., 2003a). The maximum potential temperature anomaly for Iceland relative to primitive MORB (1230° C) is then ~100 K.

Bathymetry, Vertical Motions, and Heat Flow

Other methods that have been applied to estimate temperature include modeling the bathymetry of the north Atlantic, assuming it represents lateral flow from a plume upwelling beneath Iceland (Ribe et al., 1995), and modeling uplift and subsidence of the ocean margins at the time of continental breakup, using data from sedimentary sequences and assuming a thermal source for the vertical motions (Clift, 1997; Clift et al., 1998; Clift, this volume). These analyses suggest temperature anomalies of 50–100 K. Heat flow measurements from the ocean floor north and south of Iceland show no significant anomaly compared with global average models (Stein and Stein, 2003; DeLaughter et al., this volume), although given the large errors in these data, a temperature anomaly of up to ~200 K in the Iceland region probably cannot be ruled out (C. Stein, 2004, personal commun.). The heat flow is lower beneath the north Atlantic west of the MAR than east of it, however, the opposite of that expected for a southeastward-migrating plume.

Summary

Most estimates of the temperature anomaly beneath Iceland fall in the range 0–100 K, with extremal seismic estimates as high as ~200 K. The highest values are marginally sufficient for a weak, shallow, thermally buoyant plume, and inconsistent with the more numerous moderate estimates. The apparent reduction in the temperature anomaly with increasing depth in the mantle beneath Iceland is inconsistent with a plume, and the mechanism by which a shallow, cool, bottom-heated upper-mantle plume might form in the absence of any known thermal boundary layer is unclear.

PETROLOGY AND GEOCHEMISTRY

Basalts from the spreading ridges in the north Atlantic have been extensively sampled along the axis of the MAR and to a considerable extent across the basin into Greenland, Great Britain, and the Norwegian margin. Geochemical measurements show that the north Atlantic is occupied by a broad compositional excursion involving enrichment in incompatible elements and radiogenic isotope ratios, including the highest noncosmogenic maximum $^3He/^4He$ ratios observed anywhere on Earth (Stuart et al., 2003). On the ridge itself, the geochemical anomaly culminates in the Iceland region. Plume theory was adapted early on to account for such geochemical signatures (Schilling, 1973a).

The general assumption that high temperatures exist at Iceland in part arises from applying to Icelandic basalts MORB-based melt-column models that assume a homogeneous peridotite source, judge that more must be processed to provide the large volumes of melt observed, and conclude that more heat is required (Klein and Langmuir, 1987a; McKenzie and Bickle, 1988; Langmuir et al., 1992; Maclennan et al., 2001b). However, the index of partial melting extent of Klein and Langmuir (1987b), Na8, which is Na_2O-corrected for fractionation to a near-parental MgO content of 8%, is higher and has greater variance in basalt glasses, with 5–10.6% MgO content on Iceland than it does on the adjacent Reykjanes and Kolbeinsey ridges (Natland, 2003a; Foulger et al., 2005). It thus should represent a lesser rather than a greater extent of partial melting. Suggestions that the Icelandic source is not homogeneous, and that it includes wehrlite or pyroxenite (Gurenko and Chaussidon, 1995; Sigurdsson et al., 2000) or recycled ocean crust (Chauvel and Hémond, 2000; Breddam, 2002) may explain this variance, although if such material is volumetrically important, it will likely disqualify MORB-based melting models. The lower melting temperatures of such material (Hirschmann and Stolper, 1996; Hirschmann et al., 2003; Pertermann and Hirschmann, 2003), or of peridotite reequilibrated with partial melt derived from eclogite (Yaxley, 2000), would clearly mitigate the need for a high-temperature mantle beneath Iceland and thus render arguments for a hot or deeply-sourced plume unnecessary.

The plume proposed to explain the geochemistry of the Iceland region has become progressively more complicated as isotopic and trace-element studies have proliferated. Increasing enrichment along the Reykjanes Ridge approaching Iceland from the south was originally interpreted in terms of mixing between two simple sources—depleted mantle similar to that beneath most MORs and enriched mantle in a plume beneath Iceland (Hart et al., 1973; Schilling, 1973a). With the addition of isotope observations from Iceland, the number of components rose to three (Hanan and Schilling, 1997). Recently, as many as four components have been invoked to explain the Sr-Nd-Pb-Hf isotopic variations (Kempton et al., 2000). The location of these sources in the mantle—whether within the assumed plume or outside it, and whether shallow or deep—are disputed. Whether even more "components" are required is still under investigation (G. Fitton, 2004, personal commun.). Controversy exists over the source of a distinctive depleted component (Thirlwall et al., 1994; Hards et al., 1995; Kerr et al., 1995; Thirlwall, 1995). Is it a variety of peridotite entrained in a plume (a "depleted plume component"; Fitton et al., 1997, 2003; Kempton et al., 2000), the gabbroic portion of ancient recycled ocean crust entrained in a plume (Breddam et al., 2000; Chauvel and Hémond, 2000), or does it result from unusually extensive partial melting of common depleted peridotite MORB source (Hanan et al., 2000; Stracke et al., 2003a)?

Differentiated Icelandic basalts are systematically more enriched than are the depleted olivine tholeiites and picrites. Indeed, picrites and komatiites worldwide tend to have depleted compositions (Anderson, 1994a). That these depleted lavas are the highest-temperature eruptives on Iceland (Breddam et al., 2000) is undisputed. Thus the high-temperature core of any assumed plume must paradoxically comprise partly or mainly the most depleted material sampled from the North Atlantic mantle. But why then are the differentiated basalts systematically more enriched? This would not be the case if they were related to the depleted olivine tholeiites and picrites along a common liquid line of descent controlled by shallow crystallization differentiation. This difficulty was first highlighted by O'Hara (1973) in discussion of the work by Schilling (1973a) on Reykjanes Ridge basalts.

Schilling (1973b) responded that the extent of differentiation was irrelevant because the sources of the basalts are spatially separated beneath the Reykjanes Ridge. However, it was subsequently shown that in Iceland, relatively enriched differentiates and depleted olivine tholeiites and picrites erupt over short periods from the same volcanic centers (e.g., Theistareykir; Stracke et al., 2003a). Enriched and depleted magnesian melt inclusions are even found in single spinel crystals in picrites (Sigurdsson et al., 2000). The proposed geochemical explanation is that the enriched and depleted materials represent sequential extraction of smaller to greater partial melts from within the plume during its ascent (Fitton et al., 1997; Maclennan et al., 2001b; Stracke et al., 2003a). We agree with part of this assessment, but below propose a different explanation involving fractional melting of eclogite that does not require a plume.

It is widely assumed that among geochemical signals, only high $^3He/^4He$ ratios indicate an unambiguous lower-mantle component (Farley and Neroda, 1998; Graham, 2000). Thus Courtillot et al. (2003) used $^3He/^4He$ as one of five plume indicators, but concluded that some high-value $^3He/^4He$ hotspots are not underlain by deeply sourced plumes, thus implicitly negating one of their own assumed plume indicators (Anderson, this volume). Although many high $^3He/^4He$ values are observed in Iceland, the scatter is large and the values are lower in central Iceland than on the more distant and indirectly linked Reykjanes Peninsula. The early determination that $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ are uncorrelated with $^3He/^4He$ (Condomines et al., 1983) has not changed with the acquisition of much new data, although supporting data for major oxides; trace elements; REEs; and Sr, Nd, and Pb isotopes on the same samples are disappointingly inadequate for the necessary comparisons. The $^3He/^4He$ values for nine Icelandic picrites (MgO > 12%) range from a MORB-like 8 R_A (where R_A is the atmospheric $^3He/^4He$ ratio of 1.38×10^{-6}) to 37.5 R_A (Hilton et al., 1999). For basalts with MgO content of 5.5–12%, $^3He/^4He$ values range from 5 to 28 R_A. There is thus no evidence for a relationship between high $^3He/^4He$ ratio, degree of depletion or enrichment, and therefore source temperature. In summary, He isotopes do not correlate with other indices that have been ascribed to apparently enriched deep mantle material, and consequently there is a decoupling between He isotopes and other geochemical indicators. Either He isotopes cannot be unambiguously attributed to deep mantle material or the apparent deep mantle signature (high He isotopes) is imparted in such a way that other isotopes are not influenced.

The highest noncosmogenic value of $^3He/^4He$ on the Earth, 49.5 R_A, is found in a Paleocene picrite from Baffin Island, offshore of west Greenland (Stuart et al., 2003), where a plume is argued to have first impacted (Lawver and Muller, 1994). The lava erupted was associated with the opening of the Labrador Sea. The high $^3He/^4He$ values correlate with the degree of depletion. Stuart et al. (2003) suggest that a depleted upper-mantle source was infiltrated by a potent high-$^3He/[U+Th]$ contaminant from the lower mantle that significantly altered only the $^3He/^4He$ characteristics of the lavas. This suggestion implies that the "depleted plume component" arises from the upper mantle, but again, it does not explain why this component should apparently be associated with the highest temperatures (Holm et al., 1993; Graham et al., 1998; Larsen and Pedersen, 2000).

Recent geochemical models for Iceland consider the enriched and depleted components to be mixed together in a high-temperature plume (e.g., Kempton et al., 2000; Maclennan et al., 2001a) and sequentially sampled during progressive partial melting as the material ascends. The high $^3He/^4He$ value is assumed to come from material long isolated in the lower mantle (e.g., Hanan and Graham, 1996). A variant model suggests that the ascending plume carrying the enriched, depleted, and vapor components acquired a sheath from a two-component, shallow-mantle layer (Kempton et al., 2000). Like the original two-component model suggested by Schilling (1973a), the sheath model predicts radial geochemical symmetry, which, on a spreading ridge, will produce a bilaterally symmetric pattern (Kent et al., 1992; Fitton et al., 1997; Kempton et al., 2000). Such symmetry would still occur in the presence of lateral asthenosphere flow from central Iceland (e.g., Ribe et al., 1995).

The geochemical pattern is not symmetric, however. Pb isotope ratios do not peak at the proposed plume center in southeast

Iceland. Some basalt with highly unradiogenic Pb values is found in central Iceland (Chauvel and Hémond, 2000). Immediately north of the proposed plume center, in the NVZ and along the Kolbeinsey Ridge, Pb is relatively unradiogenic ($^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ ratios are low), suggesting depleted mantle sources for these regions (Mertz et al., 1991; Thirlwall et al., 2004). In fact, Pb isotope values correlate with rock type and not location (Chauvel and Hémond, 2000). Rocks enriched in $^{238}U/^{204}Pb$ comprise only the more alkaline and generally off-axis basalts in Iceland that represent a small percentage of all the lavas, whereas the majority of Icelandic lavas in the central rift are tholeiites that are as low as MORB in $^{206}Pb/^{204}Pb$ (Chauvel and Hémond, 2000).

The chondrite-normalized ratio [La/Sm] enrichment factor (e.f.) has both much higher and lower values in Iceland than on the nearshore portions of the Reykjanes and Kolbeinsey ridges. It is low in picrites in Iceland and higher in differentiates. Along with isotope ratios and the ratios of incompatible trace elements, [La/Sm]e.f. correlates with the extent of differentiation of the host rock. Changes in these ratios are not expected for iron-enrichment crystallization differentiation (Schilling, 1973a), which is generally thought to be the dominant process in Iceland. If the interior of a plume is its most enriched part, [La/Sm]e.f. should increase toward it. However, whereas [La/Sm]e.f. does increase toward Iceland along the Reykjanes Ridge, it has a wide spread of values, including both high and low values within Iceland, and increases once more northward along the Kolbeinsey Ridge (Mertz et al., 1991). Kolbeinsey Ridge basalts also have a depleted Pb isotopic signature.

Geochemical discontinuities of different characters thus occur at the junctures of Iceland and the Reykjanes and Kolbeinsey ridges. These discontinuities are manifest in both Na_8 and the isotopic and trace-element indicators of an enriched mantle source. These observations cast doubt on whether the shallow elevations of Reykjanes and Kolbeinsey ridges and their chevron bathymetric ridges (Vogt et al., 1980; Jones et al., 2002) are formed by the outward flow of pulses of hot plume material from a plume beneath southeast Iceland (Mertz et al., 1991) as suggested, for example, in some geophysical models (Ribe et al., 1995; Yale and Morgan, 1998). Even if this occurs along the Reykjanes Ridge, the extent of the proposed lateral flow is unclear, as different geochemical "plume tracers" disagree. For example, [La/Sm]e.f. and Pb isotopes increase toward Iceland, whereas there is an abrupt step-increase in $^{87}Sr/^{86}Sr$ (Hart et al., 1973). Elevated $^{87}Sr/^{86}Sr$ and Nb extend only as far south as 61° N (e.g., Fitton et al., 1997), whereas $^{3}He/^{4}He$ values are elevated as far south as the Charlie Gibbs Fracture Zone at 53° N (Poreda et al., 1986).

Icelandic petrology and geochemistry are complicated by subaerial and crustal processes, which are minor along MORs. Rhyolite is a significant component of the Icelandic crust, especially at central volcanoes (e.g., Walker, 1963; Carmichael, 1964), and Iceland consequently is a classic locality for the study of magma mixing between basalt and rhyolite in zoned magma chambers (Yoder, 1973; Sigurdsson and Sparks, 1981; Blake, 1984; Gunnarsson et al., 1998). The recycling of subsided, surface-erupted lavas; residence in shallow magma chambers; and perhaps long transit distances through a thickened crust increase differentiation and alter isotope ratios (e.g., Oskarsson et al., 1982). Glacial unloading allows eruption of less differentiated and more depleted lavas, perhaps because they experienced shorter crustal residence times (Gee et al., 1998). These factors show that local processes may influence petrological and geochemical attributes that are commonly interpreted in terms of a plume model. It may be difficult to separate the effects of shallow differentiation, assimilation, and mixing from melting processes in the mantle.

The issues discussed above illustrate that, if interpreted in terms of a plume model, the geochemistry of Iceland requires the addition of considerable complexity, and the resulting models are inconsistent with geophysical plume models postulated to explain the surrounding seafloor elevation. Furthermore, general systematic geochemical relationships have not emerged from the huge dataset that now exists. Similar results from other proposed plume locations (e.g., Tristan da Cuhna) have led to modification of the hypothesis in the proposal that plumes are geochemically heterogeneous and that their structure cannot be studied using local geochemical variations. A complex plume showing no systematic geochemical spatial or temporal trends might also be the only possible fit for Iceland. Nevertheless, geochemical data still continue to be interpreted in terms of a plume framework on an opportunistic basis where individual studies permit (e.g., Breddam et al., 2000).

PLATE TECTONIC MODEL FOR ICELAND

What Needs to Be Explained?

It is helpful to reassess what are basic observations and what merely assumptions or nonunique deductions. The primary anomalies in the Iceland region are:

1. Seismic crustal thickness of up to 40 km, compared with 10 km along the Reykjanes and Kolbeinsey ridges. This thickness implies the possible production of several times more melt at the MAR between ~63°30′ N and ~66°30′ N than beneath the neighboring ridges;
2. Temperatures that are probably at most only mildly elevated relative to those on other MORs;
3. A ~2000-km-wide mantle seismic anomaly, centered on Iceland, with a weak downward extension into the transition zone, in contrast to the ~200-km depth of low wavespeeds beneath neighboring marine parts of the spreading ridge system; and
4. A geochemical anomaly that extends southward along the Reykjanes Ridge, occupying a total of ~600 km of the spread-

ing plate boundary. It abruptly assumes higher variance in almost all parameters at Iceland, but is absent to the north (Korenaga and Kelemen, 2000).

Melt Volume

To produce anomalously large volumes of melt requires high temperature, an unusually fusible source composition, a process that drives excess mantle through the melting zone, or a combination of these conditions. Such processes as lithospheric delamination or melt ponding might explain ephemeral, large-volume melt production (e.g., in large igneous provinces; Tanton and Hager, 2000; Elkins-Tanton, this volume) but in the Iceland region, if the thick-crust model is correct, anomalous melt production has been essentially steady-state since the north Atlantic opened ca. 54 Ma. Induction of volcanism by lithospheric delamination or edge-driven gyres and eddies (EDGE) convection (King and Anderson, 1998) may have contributed to the formation of the north Atlantic Igneous Province at the time of continental breakup (Boutilier and Keen, 1999).

Almost all studies of temperature beneath the Iceland region either require or permit an anomaly of no more than a few tens of K and maximally, ~100 K. Such a small anomaly can account for only a little of the excess melt. It thus seems inescapable that much of the melt anomaly must be attributed to an unusually fusible source composition; this explanation is supported by the ample geochemical evidence for a compositionally anomalous source.

The suggestion that oceanic crust is recycled at Iceland (e.g., Fitton et al., 1997; Chauvel and Hémond, 2000; Breddam, 2002), and is the source of the "depleted plume component" of Kempton et al. (2000) has prompted us to consider whether eclogite might be present in the source in considerable amounts. Both Chauvel and Hémond (2000) and Breddam (2002) suggest that oceanic crust is entrained in a plume source along with a high value of $^3He/^4He$ from the lower mantle. Chauvel and Hémond (2000) further suggest that differentiated Icelandic tholeiites are derived from partial-to-large-scale melting of the basaltic portion of the ocean crust (originally extrusives and sheeted dikes), and that olivine tholeiites and picrites are derived from the gabbro cumulate portion. They thus suggest that Iceland is derived from partial melting of a complete section of ocean crust. Their model requires melting of a harzburgitic component to explain the high Ni content of the picrites.

Iceland and the north Atlantic Igneous Province formed in the Caledonian suture, which was created at ca. 400 Ma, when what are now Greenland and Scandinavia collided as the Iapetus Ocean closed (Fig. 9) (Soper et al., 1992; Lundin and Doré, this volume). When the Eurasian supercontinent broke up again at ca. 54 Ma, the new MAR ran along the suture for much of its length. At the present location of Iceland, however, it crossed the western Caledonian frontal thrust, where the latter runs from present-day Greenland into Britain. The Caledonian suture is the site of earlier subduction and, if slabs are retained in the shallow upper mantle, is expected to be abundant there.

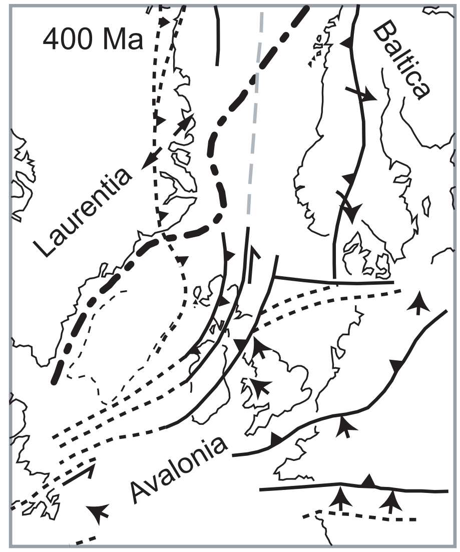


Figure 9. Closure of the Iapetus Ocean at 400 Ma, by convergence of Laurentia, Baltica, and Avalonia. Arrows indicate convergence directions; black lines show faults and orogenic fronts. Black triangles indicate sense of thrust faults. Gray dashed line shows inferred position of the Caledonian suture. Slabs were subducted beneath Greenland, Baltica, and Britain (after Soper et al., 1992). Bold dashed line indicates position of the Mid-Atlantic Ridge that formed at ca. 54 Ma. Lighter dashed lines indicate postulated extensions of the thrusts mapped. Adapted from Foulger and Anderson (2005).

Eclogite and eclogite-peridotite mixtures have lower liquidi and solidi and, in contrast to previous conclusions (e.g., Yoder and Tilley, 1962; Ito and Kennedy, 1974), have melting intervals as wide or wider than that for peridotite, displaced overall to lower temperatures (Fig. 10A) (Yaxley, 2000). The eclogite liquidus is ~180 K above the lherzolite dry solidus at all pressures and close to the temperature at which N-MORB is generated by ~13% melting of lherzolite at a potential temperature of 1280° C (McKenzie and Bickle, 1988). Thus eclogite may melt almost completely at temperatures similar to those required to produce N-MORB. In the case of eclogite-peridotite mixtures, at a given temperature, up to several times the amount of melt is expected than from the same volume of pure peridotite (Fig. 10B). Thus it may be possible to explain the large volume of melt at Iceland by passive isentropic upwelling the same as elsewhere along the MAR, but occurring where the mantle is largely eclogite or fertilized by eclogite from ancient subduction.

The gabbroic portion of oceanic crust spans a wide range of lithologies, including troctolite, olivine gabbro, oxide ferrogabbro, silicic trondhjemite, and tonalite. These lithologies are mixed by complex deformational processes in the gabbroic section of typical ocean crust (Natland and Dick, 2002). At low pressure, in the ocean crust itself, the melting interval is probably

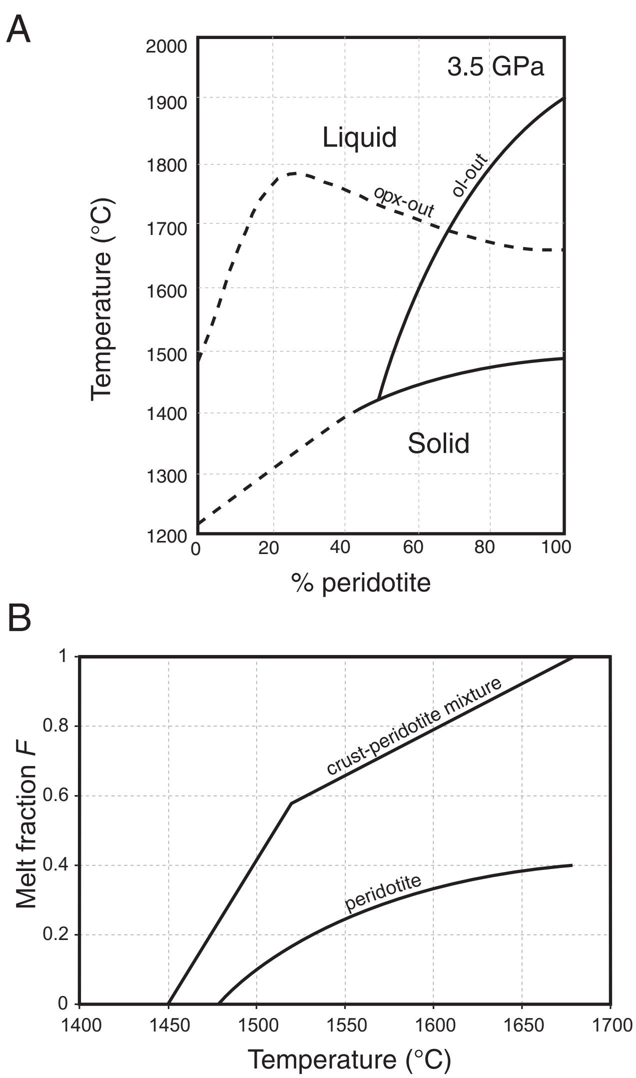


Figure 10. (A) Solidus and liquidus for fertile peridotite containing varying percentages of average altered oceanic crust. opx, orthopyroxene; ol, olivine (adapted from Yaxley, 2000). (B) Relationship between melt fraction F and temperature for fertile peridotite and a mixture of 30% average altered oceanic crust and 70% fertile peridotite. The peridotite line is the parameterization 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 (A). The higher average dF/dT and lower solidus temperature for the mixture results in enhanced melt productivity at a given temperature (derived from data in Yaxley, 2000; adapted from Foulger and Anderson, 2005).

~500 K (Natland et al., 1991; Koepke et al., 2004) and the range cannot be substantially less for the full range of gabbros and basalts transformed to eclogite facies at pressures above ~1.6 GPa. Near-solidus melts of ocean-crust assemblages spanning the gabbro-garnet/granulite-eclogite transition are initially rhyodacitic to ferro-andesitic at pressures up to ~6 GPa. More extended melting produces tholeiitic ferrobasalt and finally olivine tholeiite at ~70% partial melting (Ito and Kennedy, 1974; Yasuda et al., 1994; Pertermann and Hirschmann, 2003). The absence of a nephelinite-series sequence at Iceland and the inference of a tholeiitic partial melting sequence (e.g., Stracke et al., 2003a) may thus be explained if eclogite is dominant in the source and is the major facies undergoing fractional partial melting.

Geochemistry

The presence of remelted crust of Caledonian age in the basalts of east Greenland, Iceland, and Britain has been proposed on the basis of calculated compositions of parental melts, trace elements, REEs, and radiogenic isotope ratios (Fitton et al., 1997; Chauvel and Hémond, 2000; Korenaga and Kelemen, 2000; Breddam, 2002). Chauvel and Hémond (2000) suggest that the ocean-crust source of Icelandic basalts corresponds to the bulk compositions of (1) extrusive basalts and dikes and (2) gabbro cumulates. Although this simple twofold characterization of ocean crust is generally correct, drilled sections show that the ocean crust is considerably more complicated in detail (e.g., Natland and Dick, 2002). At slowly spreading ridges, tectonic processes form core complexes and high transverse ridges impose complex patterns of shear deformation on gabbros crystallizing beneath rift valleys. These add to the normal processes of compaction and crystal growth to drive intercumulus melts into fractures and deformation channels, where their differentiation and further expulsion of intercumulus liquid continues. The result is a matrix of primitive olivine gabbro that is an almost ideal adcumulate (i.e., an aggregation of pure cumulus minerals with virtually no intercumulus liquid), but which is crosscut by narrow seams of oxide gabbro. Many of these are cored by veins of silicic differentiates that also supply networks of silicic veins extending outward into olivine gabbros. In addition, the rocks are variably altered hydrothermally.

Thus a complex assemblage of rocks experiences partial melting at elevated pressure rather than any single composition of eclogite, such as those studied experimentally. Although we know little in detail about how such a section will look upon transformation to the eclogite facies and long residence in the mantle, Alpine eclogites derived from Tethyan Ocean crust of Jurassic age show that all of the principal ocean-crust gabbroic assemblages remained essentially intact chemically from initial transformation to eclogite in subduction zones until subsequent exhumation in the modern Alps (Mottana and Bocchio, 1975; Cortesogno et al., 1977; Evans et al., 1981; Ernst et al., 1983; Chalot-Prat, this volume). Oxide gabbros in the eclogite facies thus can contain up to 20% rutile (after ilmenite and magnetite), and most are at least slightly hydrous and contain glaucophane. In our model, we propose that ocean crust in the upper few hundred km of a subduction zone became trapped in a suture resulting from closure of an ocean basin, thus never descended to great depths in the mantle, and gradually reheated until continental rifting opened the north Atlantic in the Paleocene. In this situation, the chemical integrity of the initial gabbroic protolith may have been retained.

Partition coefficients for eclogite have been obtained on only a subset of their full range in composition and mineralogy (e.g., Stracke et al., 1999). Most eclogites studied experimentally (e.g., Ito and Kennedy, 1974; Yasuda et al., 1994; Pertermann and Hirschmann, 2003) also represent only a fraction of the known range in eclogite compositions, having generally been selected for their similarity to both primitive (high MgO content) and depleted N-MORB, but actually usually resembling an olivine gabbro cumulate. Therefore geochemistry is far from revealing what a full range in eclogite compositions might produce during fractional melting, and in defining the stability relations and effects on partition coefficients of accessory phases, such as high-pressure amphibole, mica, apatite, and zircon, not merely rutile. Inferences from trace-element geochemistry suggest that for bimineralic eclogite, inferred partial melt extracts represented by primitive basalts and melt inclusions behave such that the original trace-element concentrations and ratios of the gabbroic protolith are retained. Thus high values of Y/Zr, Zr/REE (e.g., Chauvel and Hémond, 2000; Breddam, 2002; Foulger et al., 2005) and Sr/REE (Sobolev et al., 2000) are good indications of an original cumulus olivine gabbro protolith to the eclogite (Fig. 11) (Foulger et al., 2005). This protolith provides the more depleted component, not the enriched component inferred for Hawaiian and Icelandic basalts. What might derive from partial melting of differentiated eclogite (equivalent to disseminated-oxide, oxide-rich gabbro, tonalite, and trondhjemite) is less certain.

On the assumption of geochemical integrity during the transition to eclogite, then, the first melts that form in eclogite probably come mainly from what originally was oxide gabbro (now rutile eclogite) and granitic seams and veins (now quartz eclogite). These melts will be silicic, yet with high iron content and enriched in incompatible elements, such as Rb, Nb, and light REEs (e.g., Natland and Dick, 2002). We propose that these melts, which would likely comprise early fractional melts extracted from eclogite, are a major contributor to eruptive ferrobasalts and probably ferroandesites on Iceland. Others clearly are derived from shallow crystallization differentiation of primitive basalt in the Icelandic crust, but the spectrum of Icelandic parental compositions probably includes basalts of strongly differentiated composition (with ~8–5% MgO content). It thus ranges from primitive and depleted picrite at least to ferrobasalt. In other words, many ferrobasalt liquids derived from differentiated eclogite cross the mantle. This proposal is supported by experimental petrology to the extent that both andesitic and moderately iron-rich tholeiitic liquids with low-to-moderate MgO contents have been produced in eclogite melting experiments from 0.5 to 8 GPa (Ito and Kennedy, 1974; Yasuda et al., 1994; Pertermann and Hirschmann, 2003). In the range of 1–3 GPa, solidus temperatures of eclogite are ~200 K lower than those of dry lherzolite. For the compositions studied, the extent of melting of eclogite at the dry lherzolite solidus is ~70%, and the liquids produced at that extreme are basaltic and similar to primitive Icelandic tholeiite.

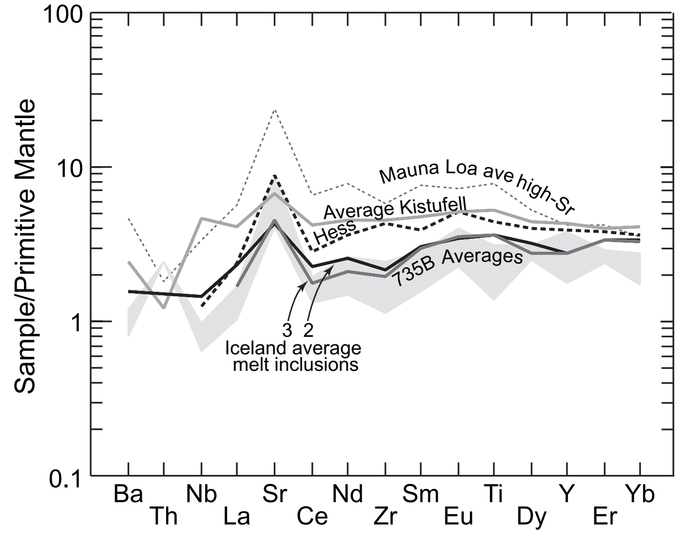


Figure 11. Spider diagram for trace elements comparing compositions of average Iceland melt inclusions and primitive basalt from Kistufell, central Iceland, with abyssal gabbros. Data are normalized to the primitive mantle of Hofmann (1988). Following the argument of Sobolev et al. (2000) for glass inclusions from Mauna Loa, Hawaii, positive Sr anomalies in Icelandic glass inclusions and primitive Kistufell basalt indicate the original presence of cumulus plagioclase in the source protolith, once abyssal gabbro and now eclogite. Glass inclusion averages are for Groups 2 and 3 shown in Figure 8, using data from Gurenko and Chaussidon (1995), Sigurdsson et al. (2000), and Slater et al. (2001). The "Average Kistufell" line is a pattern from average primitive olivine tholeiites from Kistufell, central Iceland (Breddam, 2002). The shaded band extends between averages of troctolite (lower bound) and olivine gabbro (upper bound) from ODP Hole 735B, Southwest Indian Ridge, using data compiled from several sources by Natland and Dick (2002), whose chemical identification of lithologies was used. Analyzed gabbro samples containing silicic veinlets were screened from the dataset by restricting it only to those with chondrite-normalized $[La/Sm]_N < 1$. The bold dashed line is a pattern for average olivine gabbro from Hess Deep, eastern equatorial Pacific (Pedersen et al., 1996; J. Natland, unpublished data). The gray dashed line is a pattern for average high-Sr melt inclusions in olivine from Mauna Loa, Hawaii, from Sobolev et al. (2000).

The eclogite hypothesis might be criticized because the implied extents of melting are greater than those usually considered likely for a peridotitic source. Certainly the processes of radioactive or conductive heating are too slow to keep up with the melt extraction processes. However, Iceland was first built on thin lithosphere adjacent to two thick cratons. In the process of opening the ocean, material from depths of ~100–200 km was brought up to depths of ~50–60 km. Thus adiabatic ascent and reduction of pressure is the most likely mechanism for extensive melting. A few tens of km of ascent may be able to bring eclogite from below its melting point to the 70% melt condition, and even if fractional melts drain out during ascent, they could be subsequently joined by later melts. Abyssal peridotite entrained with the ascending eclogite would also begin to melt, and its melt aliquots would contribute to the melts from eclogite during full aggregation and mixing of the magma strains, perhaps contributing to the high Ni concentrations of picritic liquids (Chauvel and Hémond, 2000).

In Iceland the degree of isotopic and trace-element enrichment correlates with the extent of iron-enrichment differentiation of tholeiitic basalts. In fact, rhyolites are the most enriched eruptives on Iceland. Isotopically they resemble the common mantle component focal zone ("FOZO") of Hart et al. (1992) (Fig. 12). The details of this correlation extend even to subsets of primitive basalts. Maclennan et al. (2003a,b), Slater et al. (2001) and Stracke et al. (2003a,b) infer that the isotopic and trace-element variability of primitive basalts from Theistareykir result from neither crystallization differentiation nor crustal assimilation, but from the mantle source. The character of the geochemical enrichment, correlating with extent of iron enrichment and decrease in MgO content, is a portion of the more extended correlation for the rest of Iceland that reaches to ferrobasalt and even more differentiated compositions. It is substantially different from isotopic/trace-element correlations in rocks of generally basaltic composition at midplate volcanoes like Hawaii. There, tholeiites, alkalic basalts, basanites, and olivine nephelinites are inferred to represent successively smaller degrees of partial melting. However, the primitive members of all of these mafic lavas have similar MgO content, indicating that they are little evolved from parental liquids produced by partial melting from peridotite containing magnesian olivine. In Iceland, in contrast, the most enriched basalts have MgO content too low to have been in equilibrium with olivine in peridotite, and their corresponding enriched but primitive, high-MgO liquid compositions are not present.

Formerly, iron-enrichment in Iceland was inferred to result from shallow crystallization differentiation (e.g., Walker, 1963; Carmichael, 1964). More recently it has been attributed to the mantle and to differences in the extent and depth of partial melting (Slater et al., 2001; Maclennan et al., 2002, 2003b; Stracke et al., 2003a,b). However, the source then cannot be peridotite. A correlation with the extent of iron enrichment differentiation during partial melting in general means that MgO in the liquid—and we would include basalts with as little as 5–7% MgO content—is not buffered by olivine in the melt. Peridotite therefore cannot be the principal source. We suggest that many ferrobasalts are direct partial melts derived from a source that is already substantially iron-rich—namely, oxide gabbro converted to eclogite—and that such liquids must cross the crust-mantle boundary. If the source cannot be peridotite, then the best, and really only, alternative candidate for low-MgO basalts is eclogite.

We thus propose that Icelandic basalts derive from a petrologically variable and mainly eclogitic source on which fractional melting acts first to extract the smallest partial melts that are either silicic, very iron-rich, or both, from abyssal gabbro. More extended partial melting produces primitive basalt. Oxide ferrogabbros represent ~20% of the 1508-m section of gabbro drilled at Hole 735B (Dick et al., 2000), and if such material underlies Iceland in the eclogite facies, the equivalents of these would probably be the first tapped at relatively low temperatures by fractional extraction from the mantle. Blending of these melts with other potential enriched components in the ocean crust—namely, magmatic amphibole, felsic veins with minerals (e.g., zircon), and enriched MORB (E-MORB) would ensure that initial fractional melts are on the average more enriched geochemically than later fractional melts generated at higher T, which would derive mainly from olivine gabbro adcumulates. This scenario is speculative, because how low-pressure accessory phases transform under eclogite-facies conditions is unknown. Nevertheless the more extensive fractional melts would look more like abyssal olivine gabbro cumulates, as we see in the geochemistry of the primitive tholeiites, picrites, and melt inclusions in Iceland.

McKenzie et al. (2004) recently proposed that the enriched component in Iceland derives from an ancient OIB seamount complex that formed atop the now-eclogitic ocean crust that is the principal source of the most depleted primitive basalts. McKenzie et al. (2004) do not speculate on the existence of accessory phases, their partitioning, or their stability during partial melting, but only note the geochemical similarities. Recycled seamounts might be involved, but enriched abyssal tholeiites (E-MORB) with flat-to-enriched REE patterns already comprise 6% of basalts dredged along the East Pacific Rise, and may contribute to the enriched component in Iceland (Foulger et al., 2005). Iceland also has two groups of enriched rhyolites with different Sr-Nd-Pb isotopes (Fig. 12). Fragments of ancient continental crust and its associated upper mantle might then also contribute to the melt source (Amundsen et al., 2002). The basic picture is, however, as we have described it. It is currently unclear whether the entire enriched component is derived from a minor component of E-MORB in "normal" oceanic crust; a seamount; aged components in abyssal gabbro that experienced radiogenic ingrowth of, for example, Sr isotopes; altered ocean crust; or fragments of old subcontinental mantle and crust marooned in the ocean from continental rifting. All of these scenarios could provide melt strains at relatively low temperatures during early stages of fractional melting. It is clear, however, that the "eclogite signal" in Icelandic picrites resembles abyssal gabbro, as previously discussed by Chauvel and Hémond (2000) and Breddam (2002).

An important attribute of primitive Icelandic melt inclusions is their crossing REE patterns and diverse yet low oxygen isotope levels, even in single crystals of olivine (Gurenko and Chaussidon, 1995; Slater et al., 2001; Maclennan et al., 2003a,b). Crossing REE patterns in lava sequences or melt inclusions are usually interpreted as indicating fractional melting. However, among the gabbros of Hole 735B, our suggested model protolith, olivine gabbro, oxide gabbro, and tonalite-trondhjemite veins were interleaved and cross-intruded in complicated ways by deformation beneath a rift valley while the rocks were still partially molten (Natland and Dick, 2001, 2002). Many relatively primitive gabbros were infiltrated along grain boundaries by silicic melt resulting in, for example, variable La/Sm ratios in

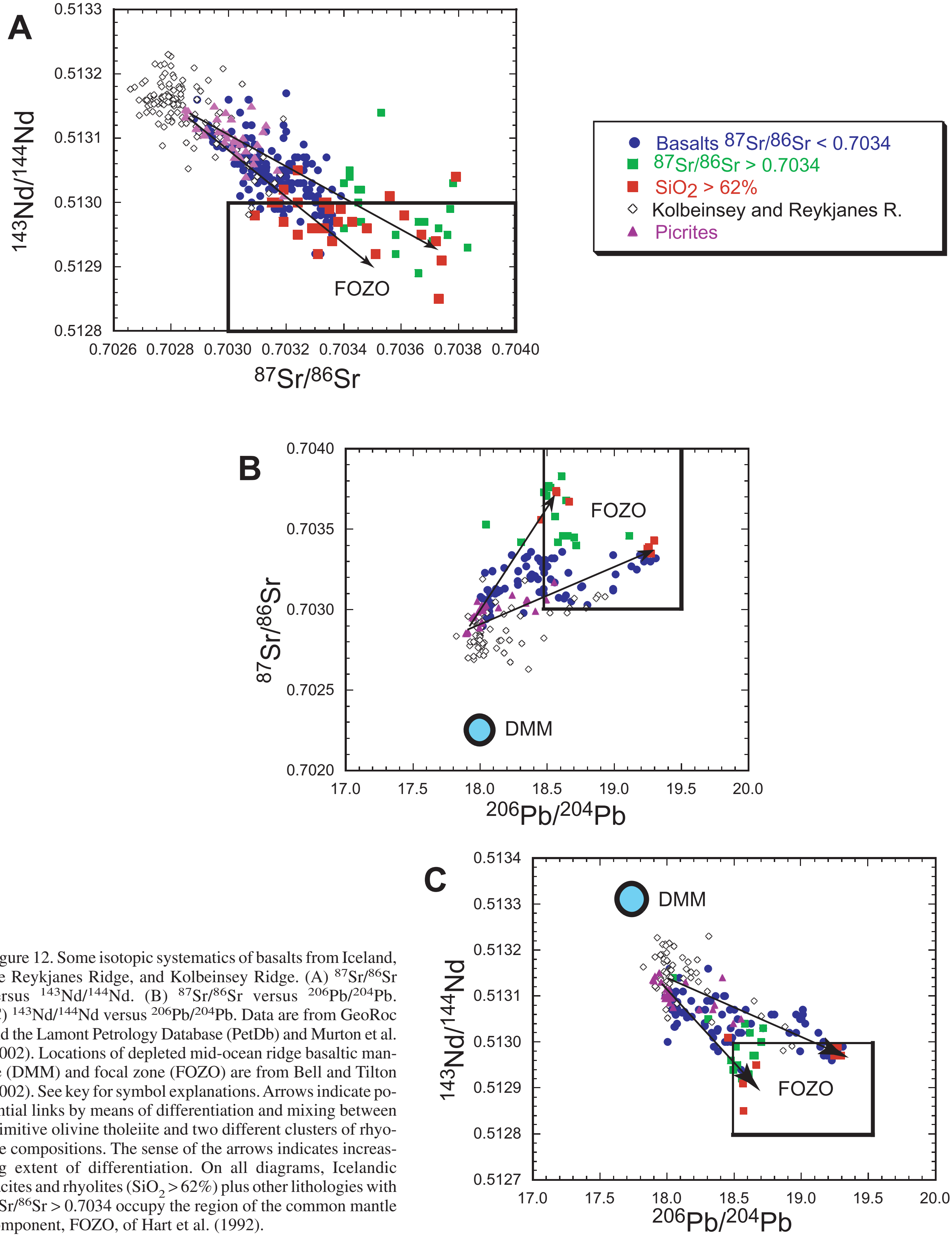


Figure 12. Some isotopic systematics of basalts from Iceland, the Reykjanes Ridge, and Kolbeinsey Ridge. (A) $^{87}Sr/^{86}Sr$ versus $^{143}Nd/^{144}Nd$. (B) $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$. (C) $^{143}Nd/^{144}Nd$ versus $^{206}Pb/^{204}Pb$. Data are from GeoRoc and the Lamont Petrology Database (PetDb) and Murton et al. (2002). Locations of depleted mid-ocean ridge basaltic mantle (DMM) and focal zone (FOZO) are from Bell and Tilton (2002). See key for symbol explanations. Arrows indicate potential links by means of differentiation and mixing between primitive olivine tholeiite and two different clusters of rhyolite compositions. The sense of the arrows indicates increasing extent of differentiation. On all diagrams, Icelandic dacites and rhyolites ($SiO_2 > 62\%$) plus other lithologies with $^{87}Sr/^{86}Sr > 0.7034$ occupy the region of the common mantle component, FOZO, of Hart et al. (1992).

the host gabbro. Enriched and depleted REE patterns thus were measured on physically and petrographically similar olivine gabbro samples separated by only a few tens of cm in the core. The same variability among primitive Icelandic melt inclusions that are inferred from other geochemical attributes to resemble abyssal olivine gabbro thus may indicate not fractional melting but incomplete mixing of melt strains derived simultaneously from nearby enriched and depleted eclogitized olivine gabbro. If other evidence is persuasive that melting was in the eclogite facies, then the variably enriched melt inclusions suggest that local source heterogeneity, probably including isotopes, survived transformation to the high-pressure facies. Wherever host olivine, pyroxene, and plagioclase phenocrysts crystallized in the Icelandic crust or upper mantle, in our model, they reflect the local diversity of melt strains derived from partial melting of the abyssal gabbro protolith. This diversity includes strong local variability of oxygen isotopes, which at Hole 735B is centered on narrow veins filled with secondary amphibole and associated hydrothermal minerals (c.f. Stakes et al., 1991; Hart et al., 1999).

In Figure 8, Group 1 melt inclusions most closely resemble the most primitive Icelandic basalt glasses, except for their lower FeO content. They are similar in other respects as well, but range to lower Na_2O and TiO_2 contents and higher CaO/Al_2O_3. Foulger et al. (2005) note the similarity in bulk composition of primitive Icelandic basalt to average olivine gabbro from Hole 735B. Group 1 melt inclusions also resemble olivine gabbro, but are more inclusive of Hole 735B compositions that have more calcic plagioclase than average, and a lower percentage of intercumulus melt resulting from nearly ideal adcumulus development, in the sense of Natland et al. (1991) and Natland and Dick (2002). Group 2 melt inclusions are similar in major-element content to pyroxene-rich gabbro cumulates, or pyroxenites of Hole 735B. Group 3 melt inclusions are similar to olivine-plagioclase adcumulates, or troctolites. Sigurdsson et al. (2000) infer pyroxenitic and wehrlitic precursors of the mantle sources of the latter two groups of inclusions. We suggest instead that varieties of primitive eclogite, transformed from plagioclase pyroxenite and troctolite, were involved. This is in accord with the suggestion from crossing REE patterns among melt inclusions in single phenocrysts that local melt strains are produced during partial melting of a rock complex that originally was part of the lower ocean crust.

The influence during partial melting of lithological diversity of the mantle and especially of potential masses of subducted ocean crust embedded in that mantle cannot be too strongly stressed. In the gabbroic portion of ocean crust, the diversity is particularly extreme because of what Bowen (1920) called "differentiation by deformation." He pointed to what today would be described as formation of adcumulates "of extreme purity" in rock masses that were deformed as they completed their crystallization (Natland and Dick, 2001). Thus in such rocks, late-stage melt between cumulus minerals is not trapped, as standard cumulus theory states. Instead, interstitial liquid is so effectively expelled from the compacting and deforming partly molten rock mass that virtually none is left. This mechanism has profound consequences on the highly incompatible trace elements, reducing for example, the U and Th concentrations in olivine gabbros and troctolites of ODP Hole 735B by two orders of magnitude below concentrations in the corresponding basaltic liquids from which the cumulus minerals crystallized (Natland and Dick, 2002). Compared with the liquids, such cumulates can be described as ultradepleted, and it is the ultradepleted signal (low Ti, Zr, Y, and REE values in general; high Y/Zr, high Zr/REE, very high Sr/REE values, etc.) that appears in Icelandic picrite. However, in the abyssal gabbros of Hole 735B, all the incompatible elements, including U and Th, are strongly concentrated by an order of magnitude more than in basaltic liquids, in seams of oxide gabbro and granitic veinlets that crosscut olivine gabbros and troctolites in hundreds of places all along the core. What was squeezed out of one part of the rock became highly concentrated in another, but only later, at the much lower temperatures of the later liquid line of descent (Natland et al., 1991). This process is another example of what Bowen (1920) named "differentiation by deformation"; namely, juxtaposition of rocks representing profoundly different stages of differentiation at abrupt contacts.

Stracke et al. (2003b) argue that partitioning of U and Th during melting of bimineralic eclogite is insufficient to achieve the extent of U-Th disequilibrium observed in Icelandic basalts, including picrites, and thus conclude that eclogite cannot be abundant in the source. However, in many silicic veinlets in gabbros of Hole 735B, the concentrations of U and Th are orders of magnitude higher, and the ratio Th/U is several times higher, than in ultradepleted olivine gabbro (e.g., Niu et al., 2002). Thus the partitioning of U and Th into subsequent fractional melts from the total rock mass—olivine gabbros plus intimately juxtaposed seams and veins of more differentiated material—will be dominated by the original distribution of the two elements in minor granitic veins and seams of oxide gabbros, and very likely their partitioning into accessory minerals, such as mica, amphibole, apatite, and zircon. These will swamp any signal from ultradepleted eclogite, including that provided by melt strains derived only a few tens of cm away. Therefore no limitation on the total proportion of eclogite involved during partial melting of Icelandic picrite can be calculated based on the erroneous assumption that ultradepleted eclogite is the only eclogite-facies lithology present in the source (e.g., Stracke et al., 2003b). Intimate juxtaposition of enriched and ultradepleted material in the source on the scale of cm to m may obviate the necessity for models of extreme fractional melting to produce the strong contrasts in enriched and depleted lava compositions that erupt in close proximity and over short periods along the Icelandic rift systems, and which are even evident in compositions of melt inclusions within single mineral grains. Geometrical models invoking wide separation of low-temperature enriched and high-temperature depleted components in the melt source (e.g., that of Stracke et al., 2003b), are contradicted by the facies distribution that is likely in eclogite derived from an ocean-crust protolith beneath Iceland.

A persistent problem of Icelandic petrogenesis has been the fairly high volume of silicic eruptives (i.e., breccias and flows of rhyolite, dacite, and andesite) at Icelandic central volcanoes, and the potential that basalt at such localities has been mixed or contaminated with them (e.g., Walker, 1963; Carmichael, 1964; Yoder, 1973; Sigurdsson and Sparks, 1981; Gunnarsson et al., 1998). Thus many petrologists and geochemists concerned with mantle processes beneath Iceland (Hanan and Schilling, 1997; Slater et al., 2001; Maclennan et al., 2003a,b; Stracke et al., 2003a,b) have excluded silicic compositions from consideration and focused on magnesian basalts found in rift zones away from central volcanoes, to screen out possible effects of crustal contamination, such as assimilation and amphibolite melting. Nevertheless all lithologies will be present in any mass of ocean crust underlying Iceland, however it arrived there. Monolithic bimineralic eclogite is not exclusively present. In fact, silicic and mafic lithologies are so intimately interspersed in abyssal gabbro protoliths that immediate mixing at the source cannot be avoided. Silicic liquids are also among the initial partial melts produced in experiments on what we have termed *ultradepleted eclogite*—olivine gabbro in the eclogite facies (Yasuda et al., 1994; Pertermann and Hirschmann, 2003). In our view, then, both the general petrology and geochemistry of Icelandic basalts strongly implicate the mixing of one or more enriched silicic magma components (Fig. 12) into more primitive melt extracts both near the melt source and in the crust. To establish the details of mixing at high pressure and to distinguish this from processes in the crust require further work both experimentally and geochemically, but melting models involving only one or two lithologies of eclogite and/or peridotite are not likely to be correct. Instead, the source rocks are predifferentiated and wildly variable in composition on a very fine scale. It is specious to argue that very long residence of such material recycled into the mantle will rehomogenize the rocks completely when we can already see the consequences of such heterogeneity in the diversity and crystallization histories of primitive Icelandic tholeiite. We need to consider partial melting in that framework.

Our model requires melts to move from their sources to crustal magma chambers with little interaction with the crust through which they pass, and to retain correlations between different isotopes, trace elements, and major elements through to eruption on the surface. This is most readily explained if the enriched component is small in scale and volume. Oxide gabbros transformed to eclogite are the largest potential contributor to the enriched component and may make up 15–20% of the lower oceanic crust. Other components (e.g., E-MORB, seamount basalt, silicic material) are smaller fractions but potentially more potent geochemically. All these materials have a lower solidus temperature than the depleted component. In this model, however, it is unnecessary to postulate different source depths for the enriched and depleted components (e.g., Stracke et al., 2003a), or derivation from different portions of a zoned plume (e.g., Kempton et al., 2000).

Stuart et al. (2003) concluded that helium comprises essentially a pure constituent in mixing arrays of magmas from isotopically different sources in the north Atlantic, and arises from a helium-rich, lower-mantle source. It apparently influences no other chemical species. In picrites from Baffin Island, $^{3}He/^{4}He$ correlates positively with degree of depletion, although in Iceland, it is variable and has no systematic relationship with the radiogenic isotopes of Sr, Nd, or Pb. The observations from both Baffin Island and Iceland can be explained if helium is physically separated from U + Th and old, high-valued $^{3}He/^{4}He$ is preserved from ^{4}He ingrowth in a helium-poor host (Anderson, 1998a,b; Anderson et al., 2004; Meibom and Anderson, 2004; see also Meibom et al., this volume). This separation could be achieved if helium is isolated in gas bubbles trapped in a low-U + Th mineral, such as olivine, and the older the entrapment, the higher the $^{3}He/^{4}He$ value will be. Thus extraction of helium from olivine long ago concentrated by crystallization differentiation into, for example, dunite cumulates, could account for the high $^{3}He/^{4}He$ values observed. These rocks thus function as time capsules for preservation of old, high $^{3}He/^{4}He$ values.

Volatile exsolution and capture in olivine is necessarily a shallow process occurring in volcanic conduits and rift systems, where vapor nucleates as bubbles on mineral surfaces and is then trapped by skeletal growth of the mineral around the bubbles (Natland, 2003b). If erupted magma is derived from a very degassed source, such as recycled ocean crust transformed to eclogite, it may then take on the $^{3}He/^{4}He$ characteristics of even very small amounts of volatiles that it extracts during ascent to the surface.

Continental lithosphere might also significantly influence the geochemistry of Iceland. Palinspastic reconstructions permit the capture beneath central Iceland of a southerly extension of the Jan Mayen microcontinent that split off from Greenland at ca. 44 Ma, and central and southeast Iceland are the most likely regions where it might lie today (Amundsen et al., 2002; Foulger, 2003b; Foulger and Anderson, 2005). The picrites of Baffin Island and Skye erupted through ancient continental lithosphere ranging in age from 3.7 to 2.5 Ga (Bernstein et al., 1998). Upper cratonic lithosphere may also contain high values of $^{3}He/^{4}He$ olivine-rich cumulates. Subcontinental lithosphere may also have delaminated and been cycled into the upper mantle beneath the new ocean basin when the north Atlantic first formed. The elevated $^{3}He/^{4}He$ ratios on the northern Reykjanes Ridge suggest that this material is not solely confined to Iceland.

It is unnecessary to invoke a helium-rich reservoir in the lower mantle solely to account for the high $^{3}He/^{4}He$ values observed in either Greenland or Iceland. We suggest, instead, derivation from a helium-poor, shallow upper-mantle source from which helium was sequestered from U + Th. The best evidence for this derivation is the occurrence of depleted picrites with extraordinarily high $^3He/^4He$ values in western Greenland. High $^3He/^4He$ values must be found in old material, but such material does not necessarily come from great depths in the Earth.

Physical Models for Shallow Recycling of Slabs

The opening and closing stages of oceans and the formation of collisional sutures are radically different processes from steady-state plate tectonics. The final stages of ridge-trench collision introduce sediments; water; back-arc basins; and young, thin, hot oceanic crust into the shallow mantle. Such lithosphere is buoyant, and thermal modeling suggests that if younger than ca. 50 Ma, it cannot sink deeper than a few hundred km (Oxburgh and Parmentier, 1977). At a half-spreading rate of ~1 cm/yr, this would amount to ~500 km of plate. A length of late-subducting lithosphere equivalent to the thickness of the colliding cratons, or ~200 km, could thus be trapped in the continental lithosphere. The remainder, perhaps up to several hundred km in length, might reach neutral buoyancy in the asthenosphere beneath the sutured cratons.

When continental breakup occurs along old sutures, magmatism may be enhanced by mantle made unusually fertile by eclogitized subducted oceanic crust trapped in the rifting lithosphere, which may contribute to the formation of volcanic margins (see also Lundin and Doré, this volume). Enhanced magmatism may continue longer than the initial breakup stage if subducted material, or continental mantle lithosphere delaminated into the asthenosphere, continues to be recycled into the melt extraction zone beneath the ridge, despite lateral ridge migration (see also Vogt and Jung, this volume). Lateral ridge migration with respect to underlying mantle must occur because globally ridges migrate with respect to one another as plates shrink and grow. A model involving a transverse belt of fertility may also explain magmatism in the Tristan da Cunha region in the south Atlantic (Smith and Lewis, 1999; Fairhead and Wilson, this volume; see also Vogt and Jung, this volume).

The timescale and extent to which subducted crust trapped at shallow levels in the mantle rehomogenizes with its peridotite host are not known. The retention of essentially pristine blocks of crust with dimensions of the order of km and complete homogenization with mantle peridotite represent end-member scenarios (Meibom and Anderson, 2004). The solidus for the full suite of abyssal gabbro transformed to eclogite facies is much lower than that for dry peridotite. Liquid compositions from high-pressure melting experiments on garnet granulite and eclogite (Ito and Kennedy, 1974; Yasuda et al., 1994; Pertermann and Hirschmann, 2003) are similar to the compositions of both natural gabbros and Icelandic tholeiites and suggest that 60–80% melting of the original bulk gabbroic assemblage is required to reproduce the compositions of Icelandic tholeiites.

Melt extraction from partially molten rock is thought to begin at degrees of melting <1%. Consequently, progressively extracted melt increments must pond and rehomogenize in some reservoir prior to eruption. Such a process is also required be-

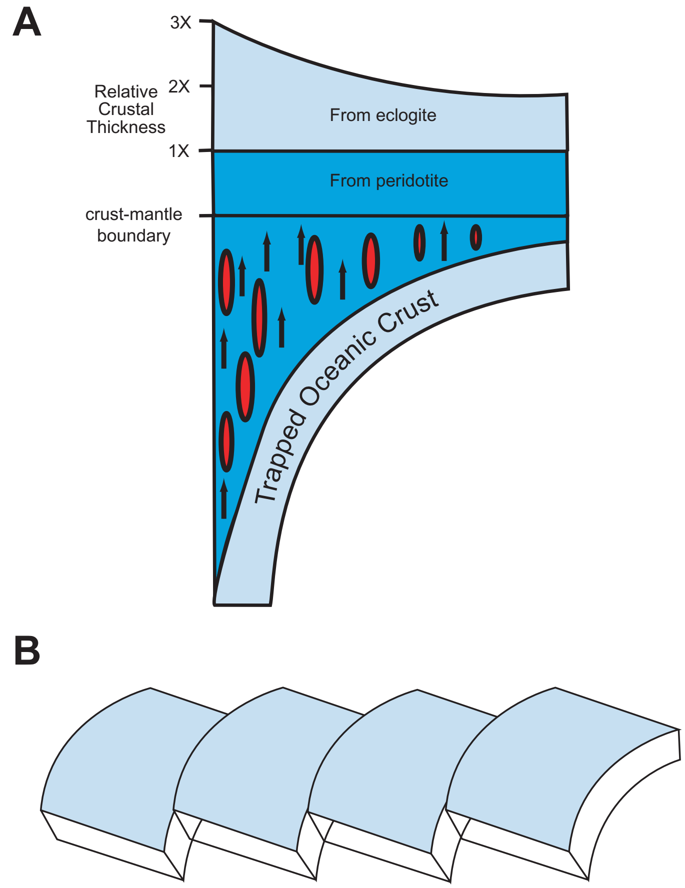


Figure 13. Schematic diagram illustrating how anomalously large amounts of melt may be obtained from remelting a subducted crustal slab of normal thickness. (A) The slab is emplaced at a high angle in the mantle, and in this particular example, two thicknesses of melt might be derived from remelting eclogite in trapped oceanic crust, and one thickness is derived from melting mantle peridotite, yielding triple the amount of melt normally observed at mid-ocean ridges. Red ellipses schematically signify rising melt (from Foulger et al., 2005). (B) Slab material may be thickened by imbrication.

neath normal spreading ridges, as MORB is thought to be formed by up to ~20% partial melting of peridotite integrated over the melt column.

To consistently produce two to three times as much melt as on the Reykjanes Ridge (i.e., 20–30 km), more than one "normal thickness" of subducted oceanic crust is required. This could be available if subducted slabs are emplaced at a steep angle in the mantle, or imbricated (Fig. 13). In the case of eclogite dispersed in a host of peridotite, the amount of fusible material available would depend on the degree of refertilization that took place and the depth extent of the source region.

How Well Do Plate Tectonic Processes Fit the Observations?

Plate tectonic processes do not require high, localized mantle potential temperatures and, in this respect, the model is consistent with the evidence for moderate temperatures in the Iceland region. In general, mantle temperature is expected to vary spatially in the Earth as a result of processes at spreading ridges and subduction zones and because of the variable insulating effects of oceanic and continental lithosphere (e.g., Anderson, 1994b). The entire Atlantic between the Charlie Gibbs and Jan Mayen fracture zones is topographically elevated and underlain by oceanic crust ~10 km thick—that is, 40% thicker than the global average of 7 km (Mutter and Mutter, 1993). This increased thickness is in keeping with a moderate temperature anomaly of regional extent.

The model we propose removes the requirement in the plume hypothesis for an eastward-younging, time-progressive volcanic track extending from central Greenland to southeast Iceland, for which there is no evidence (Lundin and Doré, this volume). The variation in seismic crustal structure, with a 40-km-thick block beneath central Iceland and thinner crust in the west than in the east, is predicted by the complex tectonics of this diffuse spreading plate boundary section and greater magmatic productivity along the dominant, eastern spreading ridge than along the subsidiary western one (Foulger and Anderson, 2005).

Eurasia drifted through several tens of degrees of latitude between 400 and 54 Ma. We expect that the continental lithosphere, which tomography suggests is ~200 km thick (Polet and Anderson, 1995), moved relative to the deeper asthenosphere during this migration. This motion suggests that the subducted material currently being remelted at Iceland was mostly trapped in the lithosphere and may have been recycled into the asthenosphere beneath the north Atlantic by lithospheric delamination. Alternatively, the young, buoyant, last-subducting lithosphere adjacent to the converging cratons may have been plated onto the bottom of the lithosphere and transported with it (Meibom and Anderson, 2004).

The seismic anomaly in the mantle beneath Iceland is ~1000 km in diameter and fills the whole ocean between the Charlie Gibbs and Jan Mayen fracture zones (see Fig. 6) (Ritsema et al., 1999). Whole-mantle tomography only has a resolution of ~1000 km, but the observations indicate that the body is significantly larger than this resolution limit (Ritsema, this volume). It extends at least down to the top of the transition zone at 410 km, but apparently does not penetrate the 660-km discontinuity and continue into the lower mantle (Z. Du, L.P. Vinnik, and G.R. Foulger, 2004, unpubl. data). The anomaly is thus much wider than it is tall, a fact that is obscured in published cross-sections with vertical exaggerations that may exceed 10:1 (e.g., Montagner and Ritsema, 2001). It is elongated north-south as is the north Atlantic itself (Foulger et al., 2000, 2001). The observations suggest that the anomaly is strongest beneath central Iceland, but the data available from on land in Iceland are of an entirely different quality and quantity from those available anywhere else in the north Atlantic region, and so meaningful comparisons are difficult.

In terms of anomaly strength, the body is bipartite. Anomalies are estimated variously to be up to 10% in V_s in some parts of the upper 200 km (Z. Du, L.P. Vinnik, and G.R. Foulger, 2004, unpubl. data.; Vinnik et al., 2005), whereas at greater depth, they are only a fraction of this strength. For the upper 200 km, the strength of the anomaly is so large that melt is required. End-member interpretations suggest either excess temperatures of several hundred K, which are unreasonably high and for which there is no independent support, or melt fractions of up to ~1% (Goes et al., 2000). At depths >200 km, a temperature excess of up to ~100 K or a retained melt fraction of up to ~0.1% would fit the observations. It is not possible to distinguish between these two candidate interpretations for the deeper anomaly on the basis of seismic wavespeed tomography alone, nor is it possible to rule out a contribution from compositional variation. It is clear, however, that the seismic anomaly weakens with depth, and even though the sensitivity of wavespeed to temperature decreases with depth, the implied temperature anomalies reduce downward. The anomaly peters out in the transition zone.

In whole-mantle tomography images, the strong anomaly in the upper ~200 km is little different from what is observed elsewhere beneath MORs (e.g., Ritsema and Allen, 2003; Pilidou et al., 2004). The unusual feature of mantle structure in the Iceland region is the weaker anomaly that extends from 200 km into the transition zone. The entire anomaly correlates with the regional shallow bathymetric anomaly of the north Atlantic. At the same time, the bathymetry does not correlate well with the geochemistry, which brings into question the link between the deeper mantle anomaly and surface volcanism. The deeper anomaly may represent a region of upper mantle that is a few tens of degrees warmer than surrounding mantle, or one that contains a fraction of a percent of partial melt, or a combination of both. A more fusible composition, perhaps involving eclogite, and/or excess volatiles (e.g., H_2O, CO_2) could account for such partial melt, as has been suggested as an explanation for the asthenospheric low-velocity zone and also low-velocity zones seen deeper in the upper mantle, extending down into the transition zone (Presnall, 2003; Presnall and Gudfinnsson, this volume). Iceland itself comprises merely the subaerial tip of the regional bathymetric anomaly, and is at most only a minor topographic anomaly in this context. It is not centered on the north Atlantic geoid anomaly, which is a feature of global proportions that also covers much of Greenland and Europe (Lundin and Doré, this volume). These large-scale features probably influence processes in the Iceland region, but they do not fit well into the model whereby they are all caused by a narrow conduit delivering melt locally beneath central Iceland.

A source for OIB involving recycled subducted crust was suggested early and is widely accepted as fitting the observed geochemistry well at Iceland and other hotspots (Hofmann and White, 1982). At Iceland, both Caledonian and older ages have been proposed for the recycled crust (e.g., Korenaga and Kelemen, 2000; McKenzie et al., 2004). Plate tectonic considerations suggest that such material is distributed throughout the upper mantle beneath Iceland and the adjacent aseismic ridges. Its distribution is mapped by the geochemistry of surface lavas.

There is no need to appeal to lateral flow from a localized source to explain the spatial distribution of geochemical anomalies, and the model does not require a radial pattern. If residual structure has survived in the Caledonian suture, broad-scale north-south bilateral symmetry would be expected, superimposed on relatively disordered smaller-scale geochemical variations related to the diverse lithologies of the source. The north-south geochemical asymmetry across Iceland apparent, for example, in the contrast in $^{87}Sr/^{86}Sr$, $^{3}He/^{4}He$, and Pb isotope ratios on the Reykjanes and Kolbeinsey ridges, may be explained if a vestige of southerly dipping slab structure is retained in the source. A relatively shallow source would also explain well the geochemical discontinuity across the short Tjörnes Fracture Zone north of Iceland (see Fig. 1).

SUMMARY

Models based on plate tectonic and shallow processes do not have many of the problems of the bottom-heated, thermal plume hypothesis at Iceland but nevertheless pose several questions and problems of their own, including:

- Can thermal equilibrium and isentropic upwelling of eclogite or a peridotite-eclogite mixture provide enough energy to supply the latent heat necessary to produce the large melt volumes observed? At present the required entropy values for the individual minerals are known to insufficient accuracy for reliable calculations of melt productivity at sufficiently high pressures (Presnall et al., 2002; D. Presnall and P. Asimow, 2003, personal commun.).
- Can large melt fractions, perhaps up to 60–80%, be ponded and homogenized above the melting column prior to eruption, and at what depth might this occur?
- At what depth do subducted slabs of various ages reside? If drifting continents are decoupled at the base of the lithospheric mantle from deeper parts of the upper mantle and the source of melt at Iceland is Caledonian-age eclogite, then this eclogite may comprise slab material retained in the upper ~200 km of the Earth. Access to Caledonian-age eclogite at greater depth would require that the drifting continents transported parts of the asthensospheric mantle with them. To derive intra-oceanic melting anomalies from eclogite in the lithospheric mantle, the latter must be incorporated into the asthenosphere beneath new oceans. How does this occur?

A scenario such as we describe here, which is radically different from the conventional plume model, may require rethinking of wider-ranging assumptions than simply the existence of a hot plume beneath Iceland. Perhaps the most intriguing question raised is whether plumes exist at all in the Earth. (Extensive resources on the subject of the origin of hotspots and the existence or otherwise of plumes may be accessed at http://www.mantleplumes.org.) Hotspots are variable in character and if the plume hypothesis is abandoned, explanations for observations at other locations are required. Yellowstone, for example, lacks a seismic anomaly deeper than ~150 km and does not feature thickened crust. However, it exhibits a time-progressive track of silicic calderas with an orientation that fits fixed-hotspot reference frames and $^{3}He/^{4}He$ isotope ratios are as high as 16 R_A, which would conventionally be considered to be unequivocal evidence for material from the lower mantle (Christiansen et al., 2002). Hawaii has many major features that are unexplained by a plume model; for example, along-chain order-of-magnitude variations in magmatic rate (Bargar and Jackson, 1974), a southward drift of some 800 km of the hotspot relative to the Earth's paleomagnetic pole during the formation of the Emperor chain (Tarduno and Cottrell, 1997), and the absence of seismic anomalies in the mantle beneath (Wolfe et al., 2002; see also Foulger, 2003a; Julian and Foulger, 2003). However, the Hawaiian chain is not forming along any known suture zone, and igneous geothermometers suggest that the mantle there is the hottest anywhere on the Earth where lavas are currently being erupted (e.g., Gudfinnsson and Presnall, 2002), although whether they are hotter than MORs is disputed (Green and Falloon, this volume).

For many years, the classical plume model has been adapted to fit unexpected results, either by relaxing the original definition (e.g., the requirement for relative fixity) by ad hoc additions to the basic model (e.g., multiple geochemical components), or by proposing that predicted features not found are fundamentally unobservable (e.g., that plume conduits in the lower mantle are too narrow to be detected). In this flexible, contemporary form, the hypothesis cannot be disproved. We therefore do not contend that the arguments presented here disprove the contemporary plume hypothesis at Iceland. We do argue, however, that there is little support for the classical plume model and that other explanations are possible, reasonable, and in many ways simpler. The contemporary plume model has little predictive power or capability to increase our fundamental understanding of the processes working at Iceland. If other explanations of the observed phenomena are not pursued, advances in our knowledge about this volcanic province will be largely limited to accumulating and documenting new observations. The challenge of earth science at Iceland is then reduced to designing plume variations. In the present paper, we set forth a new model that we hope will be tested and its own predictive capabilities assessed at Iceland and elsewhere.

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REFERENCES CITED

Amundsen, H.E.F., Schaltegger, U., Jamtveit, B., Griffin, W.L., Podladchikov, Y.Y., Torsvik, T., and Gronvold, K., 2002, Reading the LIPs of Iceland and Mauritius, in Jamtveit, B., and Amundsen, H.E.F., eds., 15th Kongsberg Seminar, 8–10 May, Kongsberg, Norway.

Anderson, D.L., 1994a, Komatiites and picrites: Evidence that the “plume” source is depleted: Earth and Planetary Science Letters, v. 128, p. 303–311, doi: 10.1016/0012-821X(94)90152-X.

Anderson, D.L., 1994b, Superplumes or supercontinents: Geology, v. 22, p. 39–42, doi: 10.1130/0091-7613(1994)022<1054:SOSCAR>2.3.CO;2.

Anderson, D.L., 1998a, The helium paradoxes: Proceedings of the National Academy of Sciences, USA, v. 95, p. 4822–4827, doi: 10.1073/pnas.95.9.4822.

Anderson, D.L., 1998b, A model to explain the various paradoxes associated with mantle noble gas geochemistry: Proceedings of the National Academy of Sciences, USA, v. 95, p. 9087–9092, doi: 10.1073/pnas.95.16.9087.

Anderson, D.L., 2000, The thermal state of the upper mantle; no role for mantle plumes: Geophysical Research Letters, v. 27, p. 3623–3626, doi: 10.1029/2000GL011533.

Anderson, D.L., Foulger, G.R., and Meibom, A., 2004, Helium: Fundamental models, http://www.mantleplumes.org/HeliumFundamentals.html.

Bargar, K.E., and Jackson, E.D., 1974, Calculated volumes of individual shield volcanoes along the Hawaiian-Emperor chain: U.S. Geological Survey Journal of Research, v. 2, p. 545–550.

Beattie, P., 1993, Uranium thorium disequilibria and partitioning on melting of garnet peridotite: Nature, v. 363, p. 63–65, doi: 10.1038/363063a0.

Bell, K., and Tilton, G.R., 2002, Probing the mantle: The story from carbonatites: Eos (Transactions, American Geophysical Union), v. 83, p. 276–277.

Bernstein, S., Kelemen, P.B., and Brooks, C.K., 1998, Depleted spinel harzburgite xenoliths in Tertiary dykes from east Greenland: Restites from high degree melting: Earth and Planetary Science Letters, v. 154, p. 221–235, doi: 10.1016/S0012-821X(97)00175-1.

Bijwaard, H., and Spakman, W., 1999, Tomographic evidence for a narrow whole mantle plume below Iceland: Earth and Planetary Science Letters, v. 166, p. 121–126, doi: 10.1016/S0012-821X(99)00004-7.

Bina, C., and Helffrich, G., 1994, Phase transition Clapyron slopes and transition zone seismic discontinuity topography: Journal of Geophysical Research, v. 99, p. 15,853–15,860, doi: 10.1029/94JB00462.

Bjarnason, I.T., Menke, W., and Flovenz, O.G., 1994, Tomographic image of the Mid-Atlantic Plate boundary in southwestern Ireland—Reply: Journal of Geophysical Research, v. 99, p. 17,915–17,917, doi: 10.1029/94JB01631.

Björnsson, A., 1985, Dynamics of crustal rifting in NE Iceland: Journal of Geophysical Research, v. 90, p. 10,151–10,162.

Blake, S., 1984, Magma mixing and hybridization processes at the alkalic, silicic, Torfajokull central volcano triggered by tholeiitic Veidivötn fissuring: South Iceland: Journal of Volcanology and Geothermal Research, v. 22, p. 1–31, doi: 10.1016/0377-0273(84)90033-7.

Bott, M.H.P., 1985, Plate tectonic evolution of the Icelandic transverse ridge and adjacent regions: Journal of Geophysical Research, v. 90, p. 9953–9960.

Bott, M.H.P., 1987, The continental margin of central east Greenland in relation to north Atlantic plate tectonic evolution: Journal of the Geological Society of London, v. 144, p. 561–568.

Bott, M.H.P., and Gunnarsson, K., 1980, Crustal structure of the Iceland-Faeroe Ridge: Journal of Geophysics, v. 47, p. 221–227.

Boutilier, R.R., and Keen, C.E., 1999, Small-scale convection and divergent plate boundaries: Journal of Geophysical Research, v. 104, p. 7389–7403, doi: 10.1029/1998JB900076.

Bowen, N.L., 1920, Differentiation by deformation: Proceedings of the National Academy of Sciences, USA, v. 6, p. 159–162.

Breddam, K., 2002, Kistufell: Primitive melt from the Iceland mantle plume: Journal of Petrology, v. 43, p. 345–373, doi: 10.1093/petrology/43.2.345.

Breddam, K., Kurz, M.D., and Storey, M., 2000, Mapping out the conduit of the Iceland mantle plume with helium isotopes: Earth and Planetary Science Letters, v. 176, p. 45–55, doi: 10.1016/S0012-821X(99)00313-1.

Cannat, M., 1996, How thick is the magmatic crust at slow spreading oceanic ridges? Journal of Geophysical Research, v. 101, p. 2847–2857, doi: 10.1029/95JB03116.

Carmichael, I.S.E., 1964, The petrology of Thingmuli, a Tertiary volcano in eastern Iceland: Journal of Petrology, v. 5, p. 435–460.

Chalmers, J.A., Larsen, L.M., and Pederson, A.K., 1995, Widespread volcanism around the north Atlantic and Labrador Sea: Evidence for a large, hot early plume head: Journal of the Geological Society of London, v. 152, p. 965–969.

Chauvel, C., and Hémond, C., 2000, Melting of a complete section of recycled oceanic crust: Trace element and Pb isotopic evidence from Iceland: Geochemistry, Geophysics, Geosystems, v. 1, 10.1029/1999GC000002.

Christensen, N.I., and Smewing, J.D., 1981, Geology and seismic structure of the northern section of the Oman ophiolite: Journal of Geophysical Research, v. 86, p. 2545–2555.

Christiansen, R.L., Foulger, G.R., and Evans, J.R., 2002, Upper mantle origin of the Yellowstone hotspot: Bulletin of the Geological Society of America, v. 114, p. 1245–1256, doi: 10.1130/0016-7606(2002)114<1245:UMOOTY>2.0.CO;2.

Christie, D.M., Carmichael, S.E., and Langmuir, C.H., 1986, Oxidation states of mid-ocean ridge basalt glasses: Earth and Planetary Science Letters, v. 79, p. 397–411, doi: 10.1016/0012-821X(86)90195-0.

Clague, D.A., Moore, J.G., Dixon, J.E., and Friesen, W.B., 1995, Petrology of submarine lavas from Kilauea’s Puna Ridge, Hawaii: Journal of Petrology, v. 36, p. 299–349.

Clift, P.D., 1997, Temperature anomalies under the northeast Atlantic rifted volcanic margins: Earth and Planetary Science Letters, v. 146, p. 195–211, doi: 10.1016/S0012-821X(96)00228-2.

Clift, P.D., Carter, A., and Hurford, A.J., 1998, The erosional and uplift history of NE Atlantic passive margins: Constraints on a passing plume: Journal of the Geological Society of London, v. 155, p. 787–800.

Condomines, M., Gronvold, K., Hooker, P.J., Muehlenbacks, K., O’Nions, R.K., Oskarsson, N., and Oxburgh, E.R., 1983, Helium, oxygen, strontium and neodymium isotopic relationships in Icelandic volcanics: Earth and Planetary Science Letters, v. 66, p. 125–136, doi: 10.1016/0012-821X(83)90131-0.

Cortesogno, L., Ernst, W.G., Galli, M., Messiga, B., Pedemonte, G.M., and Piccardo, G.B., 1977, Chemical petrology of eclogitic lenses in serpentinite, Gruppo di Voltri, Ligurian Alps: Journal of Petrology, v. 85, p. 255–277.

Courtillot, V., Davaillie, A., Besse, J., and Stock, J., 2003, Three distinct types of hotspots in the Earth’s mantle: Earth and Planetary Science Letters, v. 205, p. 295–308, doi: 10.1016/S0012-821X(02)01048-8.

Courtney, R.C., and White, R.S., 1986, Anomalous heat flow and geoid across the Cape Verde Rise: Evidence for dynamic support from a thermal plume in the mantle: Geophysical Journal of the Royal Astronomical Society, v. 87, p. 815–867, microfiche GJ 87/1.

Darbyshire, F.A., White, R.S., and Priestley, K.F., 2000, Structure of the crust and uppermost mantle of Iceland from a combined seismic and gravity study: Earth and Planetary Science Letters, v. 181, p. 409–428, doi: 10.1016/S0012-821X(00)00206-5.

Dick, H.J.B., Natland, J.H., Alt, J.C., Bach, W., Bideau, D., Gee, J.S., Haggas, S., Hertogen, J.G.H., Hirth, G., Holm, P.M., Ildefonse, B., Iturrino, G.J., John, B.E., Kelley, D.S., Kikawa, E., Kingdon, A., LeRoux, P.J., Maeda, J., Meyer, P.S., Miller, D.J., Naslund, H.R., Niu, Y., Robinson, P.T., Snow, J., Stephen, R.A., Trimby, P.W., Worm, H.-U., and Yoshinobu, A., 2000. A long in situ section of the lower ocean crust: Results of ODP Leg 176 drilling at the Southwest Indian Ridge: Earth and Planetary Science Letters, v. 179, p. 31–51, doi: 10.1016/S0012-821X(00)00102-3.

Einarsson, P., 1991, Earthquakes and present-day tectonism in Iceland: Tectonophysics, v. 189, p. 261–279, doi: 10.1016/0040-1951(91)90501-I.

Ernst, W.G., Rambaldi, E., and Piccardo, G.B., 1983, Trace element geochemistry of iron + titanium–rich eclogitic rocks, Gruppo di Voltri, Western Liguria: Journal of Geology, v. 91, p. 413–425.

Evans, B.W., Trommsdorff, V., and Goles, G.G., 1981, Geochemistry of high-grade eclogites and metarodingites from the Central Alps: Contributions to Mineralogy and Petrology, v. 76, p. 301–311, doi: 10.1007/BF00375457.

Farley, K.A., and Neroda, E., 1998, Noble gases in the Earth's mantle: Ann. Rev. Earth Planet. Sci. v. 26, p. 189–218, doi: 10.1146/annurev.earth.26.1.189.

Fitton, J.G., Saunders, A.D., Norry, M.J., Hardarson, B.S., and Taylor, R.N., 1997, Thermal and chemical structure of the Iceland plume: Earth and Planetary Science Letters, v. 153, p. 197–208, doi: 10.1016/S0012-821X(97)00170-2.

Fitton, J.G., Saunders, A.D., Kempton, P.D., and Hardarson, B.S., 2003, Does depleted mantle form an intrinsic part of the Iceland plume?: Geochemistry, Geophysics, Geosystems, v. 4, 1032, doi: 10.1029/2002GC000424.

Foulger, G.R., 2002, Plumes, or plate tectonic processes?: Astronomy and Geophysics, v. 43, p. 6.19–6.23.

Foulger, G.R., 2003a, The Emperor and Hawaiian volcanic chains, *in* Foulger, G.R., et al., eds., Penrose Conference Plume IV: Beyond the plume hypothesis, 25–29 August, Hveragerdi, Iceland: Boulder, Colorado, Geological Society of America.

Foulger, G.R., 2003b, On the apparent eastward migration of the spreading ridge in Iceland, *in* Foulger, G.R., et al., eds., Penrose Conference Plume IV: Beyond the plume hypothesis, 25–29 August, Hveragerdi, Iceland: Boulder, Colorado, Geological Society of America.

Foulger, G.R., and Anderson, D.L., 2005, A cool model for the Iceland hot spot: Journal of Volcanology and Geothermal Research, v. 141, p. 1–22.

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., and Vogfjord, K., 2000, The seismic anomaly beneath Iceland extends down to the mantle transition zone and no deeper: Geophysical Journal International, v. 142, p. F1–F5, doi: 10.1046/j.1365-246x.2000.00245.x.

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., and Vogfjord, K., 2001, Seismic tomography shows that upwelling beneath Iceland is confined to the upper mantle: Geophysical Journal International, v. 146, p. 504–530, doi: 10.1046/j.0956-540x.2001.01470.x.

Foulger, G.R., Du, Z., and Julian, B.R., 2003, Icelandic-type crust: Geophysical Journal International, v. 155, p. 567–590, doi: 10.1046/j.1365-246X.2003.02056.x.

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

Gaetani, G.A., and Watson, E.B., 2002, Modeling the major-element evolution of olivine-hosted melt inclusions: Chemical Geology, v. 183, p. 25–41, doi: 10.1016/S0009-2541(01)00370-9.

Gee, M.A.M., Taylor, R.N., Thirlwall, M.F., and Murton, B.J., 1998, Glacioisostacy controls chemical and isotopic characteristics of tholeiites from the Reykjanes Peninsula, SW Iceland: Earth and Planetary Science Letters, v. 164, p. 1–5, doi: 10.1016/S0012-821X(98)00246-5.

Goes, S., Govers, R., and Vacher, P., 2000, Shallow mantle temperatures under Europe from P and S wave tomography: Journal of Geophysical Research, v. 105, p. 11,153–11,169, doi: 10.1029/1999JB900300.

Graham, D.W., 2000, Noble gas isotope geochemistry of mid-ocean ridge and ocean island basalts: Characterization of mantle source reservoirs, *in* Porcelli, D., et al., eds., Noble gases in geochemistry and cosmochemistry, Volume 47, Reviews of Minerology and Geochemisty, p. 247–317.

Graham, D.W., Larsen, L.M., Hanan, B.B., Storey, M., Pedersen, A.K., and Lupton, J.E., 1998, Helium isotope composition of the early Iceland mantle plume inferred from Tertiary picrites of west Greenland: Earth and Planetary Science Letters, v. 160, p. 241–255, doi: 10.1016/S0012-821X(98)00083-1.

Gudfinnsson, G.H., and Presnall, D.C., 2002, The minimum potential temperature of the Hawaiian mantle is about 1420° C: Eos (Transactions, American Geophysical Union), Fall Meeting Suppl. Abstract, v. 83, V52D-06.

Gudfinnsson, G.H., Presnall, D.C., and Oskarsson, N., 2003, Contrasting origins of the most magnesian glasses from Iceland and Hawaii, *in* Foulger, G.R., et al., eds., Penrose Conference Plume IV: Beyond the plume hypothesis, 25–29 August, Hveragerdi, Iceland: Boulder, Colorado, Geological Society of America.

Gudmundsson, O., 2003, The dense root of the Iceland crust: Earth and Planetary Science Letters, v. 206, p. 427–440, doi: 10.1016/S0012-821X(02)01110-X.

Gunnarsson, B., Marsh, B.D., and Taylor, H.P., 1998, Generation of Icelandic rhyolites: Silicic lavas from the Torfajokull central volcano: Journal of Volcanology and Geothermal Research, v. 83, p. 1–45, doi: 10.1016/S0377-0273(98)00017-1.

Gurenko, A.A., and Chaussidon, M., 1995, Enriched and depleted primitive melts included in olivine from Icelandic tholeiites: Origin by continuous melting of a single mantle column: Geochimica et Cosmochimica Acta, v. 59, p. 2905–2917, doi: 10.1016/0016-7037(95)00184-0.

Hanan, B.B., and Graham, D.W., 1996, Lead and helium isotope evidence from oceanic basalts for a common deep source of mantle plumes: Science, v. 272, p. 991–995.

Hanan, B.B., and Schilling, J.-G., 1997, The dynamic evolution of the Iceland mantle plume: The lead isotope perspective: Earth and Planetary Science Letters, v. 151, p. 43–60, doi: 10.1016/S0012-821X(97)00105-2.

Hanan, B.B., Blichert Toft, J., Kingsley, R., and Schilling, J.-G., 2000, Depleted Iceland mantle plume geochemical signature: Artifact of multicomponent mixing?: Geochemistry, Geophysics, Geosystems, v. 1, doi: 10.1029/1999GC000009.

Hansteen, T.H., 1991, Multistage evolution of the picritic Maelifell rocks, SW Iceland—Constraints from mineralogy and inclusions of glass and fluid in olivine: Contributions to Mineralogy and Petrology, v. 109, p. 225–239, doi: 10.1007/BF00306481.

Hards, V.L., Kempton, P.D., and Thompson, R.N., 1995, The heterogeneous Iceland plume—New insights from the alkaline basalts of the Snaefell volcanic center: Journal of the Geological Society of London, v. 152, p. 1003–1009.

Hart, S.R., Schilling, J.-G., and Powell, J.L., 1973, Basalts from Iceland and along the Reykjanes Ridge: Sr isotope geochemistry: Nature, v. 246, p. 104–107.

Hart, S.R., Hauri, E.H., Oschmann, L.A., and Whitehead, J.A., 1992, Mantle plumes and entrainment: Isotopic evidence: Science, v. 256, p. 517–520.

Hart, S.R., Blusztajn, J., Dick, H.J.B., Meyer, P.S., and Muehlenbachs, K., 1999, The fingerprint of seawater circulation in a 500-meter section of ocean crust gabbros: Geochimica et Cosmochimica Acta, v. 63, p. 4059–4080, doi: 10.1016/S0016-7037(99)00309-9.

Herzberg, C., and O'Hara, M.J., 2002, Plume-associated ultramafic magmas of Phanerozoic age: Journal of Petrology, v. 43, p. 1857–1883, doi: 10.1093/petrology/43.10.1857.

Hilton, D.R., Gronvold, K., Macpherson, C.G., and Castillo, P.R., 1999, Extreme $^3He/^4He$ ratios in northwest Iceland: Constraining the common component in mantle plumes: Earth and Planetary Science Letters, v. 173, p. 53–60, doi: 10.1016/S0012-821X(99)00215-0.

Hirschmann, M.M., and Stolper, E.M., 1996, A possible role for garnet pyroxenite in the origin of the "garnet signature" in MORB: Contributions to Mineralogy and Petrology, v. 124, p. 185–208, doi: 10.1007/s004100050184.

Hirschmann, M.M., Kogiso, T., Baker, M.B., and Stolper, E.M., 2003, Alkalic magmas generated by partial melting of garnet pyroxenite: Geology, v. 31, p. 481–484, doi: 10.1130/0091-7613(2003)031<0481:AMGBPM>2.0.CO;2.

Hofmann, A.W., 1988, Chemical differentiation of the Earth: The relationship between mantle, continental crust, and oceanic crust: Earth and Planetary Science Letters, v. 90, p. 297–314, doi: 10.1016/0012-821X(88)90132-X.

Hofmann, A.W., and White, W.M., 1982, Mantle plumes from ancient oceanic crust: Earth and Planetary Science Letters, v. 57, p. 421–436, doi: 10.1016/0012-821X(82)90161-3.

Hofton, M.A., and 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: Journal of Geophysical Research, v. 101, p. 25,403–25,421, doi: 10.1029/96JB02466.

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., and Detrick, R.S., 2001, Mantle thermal structure and active upwelling during continental breakup in the north Atlantic: Earth and Planetary Science Letters, v. 190, p. 251–266, doi: 10.1016/S0012-821X(01)00392-2.

Holm, P.M., Gill, R.C.O., Pedersen, A.K., Larsen, J.G., Hald, N., Nielsen, T.F.D., and Thirlwall, M.F., 1993, The Tertiary picrites of West Greenland: Contributions from "Icelandic" and other sources: Earth and Planetary Science Letters, v. 115, p. 227–244, doi: 10.1016/0012-821X(93)90224-W.

Ito, K., and Kennedy, G.C., 1974, The composition of liquids formed by partial melting of eclogites at high temperatures and pressures: Journal of Geology, v. 82, p. 383–392.

Jones, S.M., White, N., and Maclennan, J., 2002, V-shaped ridges around Iceland; implications for spatial and temporal patterns of mantle convection: Geochemistry, Geophysics, Geosystems, v. 3, doi: 10.1029/2002GC000361.

Julian, B.R., and Foulger, G.R., 2003, What can seismology say about hot spots? *in* Foulger, G.R., et al., eds., Penrose Conference Plume IV: Beyond the plume hypothesis, 25–29 August, Hveragerdi, Iceland: Boulder, Colorado, Geological Society of America.

Karato, S., 1993, Importance of anelasticity in the interpretation of seismic tomography: Geophysical Research Letters, v. 20, p. 1623–1626.

Keen, C.E., and Potter, D.P., 1995, The transition from a volcanic to a nonvolcanic rifted margin off eastern Canada: Tectonics, v. 14, p. 359–371, doi: 10.1029/94TC03090.

Kempton, P.D., Fitton, J.G., Saunders, A.D., Nowell, G.M., Taylor, R.N., Hardarson, B.S., and Pearson, G., 2000, The Iceland plume in space and time: A Sr-Nd-Pb-Hf study of the north Atlantic rifted margin: Earth and Planetary Science Letters, v. 177, p. 255–271, doi: 10.1016/S0012-821X(00)00047-9.

Kent, R.W., Storey, M., and Saunders, A.D., 1992, Large igneous provinces—Sites of plume impact or plume incubation: Geology, v. 20, p. 891–894, doi: 10.1130/0091-7613(1992)020<0891:LIPSOP>2.3.CO;2.

Kerr, A.C., Saunders, A.D., Tarney, J., Berry, N.H., and Hards, V.L., 1995, Depleted mantle-plume geochemical signatures—No paradox for plume theories: Geology, v. 23, p. 843–846, doi: 10.1130/0091-7613(1995)023<0843:DMPGSN>2.3.CO;2.

King, S.D., and Anderson, D.L., 1998, Edge-driven convection: Earth and Planetary Science Letters, v. 160, p. 289–296, doi: 10.1016/S0012-821X(98)00089-2.

Klein, E.M., and Langmuir, C.H., 1987a, Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness: Journal of Geophysical Research, v. 92, p. 8089–8115.

Klein, E.M., and Langmuir, C.H., 1987b, Ocean ridge basalt chemistry, axial depth, crustal thickness and temperature variations in the mantle: Journal of Geophysical Research, v. 92, p. 8089–8115.

Koepke, J., Feign, S.T., Snow, J., and Freise, M., 2004, Petrogenesis of oceanic plagiogranites by partial melting of gabbros: An experimental study: Contributions to Mineralogy and Petrology, v. 146, p. 414–432, doi: 10.1007/s00410-003-0511-9.

Korenaga, J., and Kelemen, P.B., 2000, Major element heterogeneity in the mantle source of the north Atlantic igneous province: Earth and Planetary Science Letters, v. 184, p. 251–268, doi: 10.1016/S0012-821X(00)00308-3.

Langmuir, C.H., Klein, E.M., and Plank, T., 1992, Petrological constraints on melt formation and migration beneath mid-ocean ridges, *in* Phipps Morgan, J., et al., eds., Mantle flow and melt generation at mid-ocean ridges, Geophysical Monograph 71: Washington, D.C., American Geophysical Union, p. 183–280.

Larsen, L.M., and Pedersen, A.K., 2000, Processes in high-Mg, high-T magmas: Evidence from olivine, chromite and glass in palaeogene picrites from west Greenland: Journal of Petrology, v. 41, p. 1071–1098, doi: 10.1093/petrology/41.7.1071.

Lawver, L.A., and Muller, R.D., 1994, Iceland hotspot track: Geology, v. 22, p. 311–314, doi: 10.1130/0091-7613(1994)022<0311:IHT>2.3.CO;2.

Lundin, E., and Doré, A., 2003, The Iceland "anomaly"—An outcome of plate tectonics, http://www.mantleplumes.org/Iceland2.html.

Maclennan, J., McKenzie, D., and Gronvold, K., 2001a, Plume-driven upwelling under central Iceland: Earth and Planetary Science Letters, v. 194, p. 67–82, doi: 10.1016/S0012-821X(01)00553-2.

Maclennan, J., McKenzie, D., Gronvold, K., and Slater, L., 2001b, Crustal accretion under northern Iceland: Earth and Planetary Science Letters, v. 191, p. 295–310, doi: 10.1016/S0012-821X(01)00420-4.

Maclennan, J., Jull, M., McKenzie, D., Slater, L., and Gronvold, K., 2002, The link between volcanism and deglaciation on Iceland: Geochemistry, Geophysics, Geosystems, v. 3, doi: 10.1029/2001GC000282.

Maclennan, J., McKenzie, D., Grönvold, K., Shjimizu, N., Eiler, J.M., and Kitchen, N., 2003a, Melt mixing and crystallization under Theistareykir, northeast Iceland: Geochemistry, Geophysics, Geosystems, v. 4, doi:10.1029/2003GC000558.

Maclennan, J., McKenzie, D., Hilton, F., Grönvold, K., and Shimizu, N., 2003b, Geochemical variability in a single flow from northern Iceland: Journal of Geophysical Research, v. 3, doi: 10.1029/2001GC000384.

McKenzie, D., and Bickle, J., 1988, The volume and composition of melt generated by extension of the lithosphere: Journal of Petrology, v. 29, p. 625–679.

McKenzie, D., Stracke, A., Blichert-Toft, J., Albarede, F., Gronvold, K., and O'Nions, R.K., 2004, Source enrichment processes responsible for isotopic anomalies in oceanic island basalts: Geochimica et Cosmochimica Acta, v. 68, p. 2699–2724, doi: 10.1016/j.gca.2003.10.029.

Megnin, C., and Romanowicz, B., 2000, The 3D velocity structure of the mantle from the inversion of body, surface, and higher mode waveforms: Geophysical Journal International, v. 143, p. 709–728, doi: 10.1046/j.1365-246X.2000.00298.x.

Meibom, A., and Anderson, D.L., 2004, The statistical upper mantle assemblage: Earth and Planetary Science Letters, v. 217, p. 123–139, doi: 10.1016/S0012-821X(03)00573-9.

Menke, W., 1999, Crustal isostasy indicates anomalous densities beneath Iceland: Geophysical Research Letters, v. 26, p. 1215–1218, doi: 10.1029/1999GL900202.

Menke, W., and Levin, V., 1994, Cold crust in a hot spot: Geophysical Research Letters, v. 21, p. 1967–1970, doi: 10.1029/94GL01896.

Menke, W., Levin, V., and Sethi, R., 1995, Seismic attenuation in the crust at the Mid-Atlantic Plate boundary in south-west Iceland: Geophysical Journal International, v. 122, p. 175–182.

Menke, W., Brandsdottir, B., Einarsson, P., and Bjarnason, I.T., 1996, Reinterpretation of the RRISP-77 Iceland shear-wave profiles: Geophysical Journal International, v. 126, p. 166–172.

Mertz, D.F., Devey, C.W., Todt, W., Stoffers, P., and Hofmann, A.W., 1991, Sr-Nd-Pb isotope evidence against plume-asthenosphere mixing north of Iceland: Earth and Planetary Science Letters, v. 107, p. 243–255, doi: 10.1016/0012-821X(91)90074-R.

Meyer, P.S., Sigurdsson, H., and Schilling, J.-G., 1985, Petrological and geochemical variations along Iceland's neovolcanic zones: Journal of Geophysical Research, v. 90, p. 10,043–10,072.

Montagner, J.P., and Ritsema, J., 2001, Interactions between ridges and plumes: Science, v. 294, p. 1472–1473, doi: 10.1126/science.1067486.

Montelli, R., Nolet, G., Dahlen, F.A., Masters, G., Engdahl, R.E., and Hung, S.-H., 2004, Finite frequency tomography reveals a variety of plumes in the mantle: Science, v. 303, p. 338–343, doi: 10.1126/science.1092485.

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

Morgan, W.J., 1972, Deep mantle convection plumes and plate motions: Bulletin of the American Association of Petroleum Geologists, v. 56, p. 203–213.

Mottana, A., and Bocchio, R., 1975, Superferric eclogites of the Voltri Group (Pennidic Belt, Apenines): Contributions to Mineralogy and Petrology, v. 49, p. 201–210, doi: 10.1007/BF00376588.

Murton, B.J., Taylor, R.N., and Thirlwall, M.F., 2002, Plume-ridge interaction: A geochemical perspective from Reykjanes Ridge: Journal of Petrology, v. 43, p. 1987–2012, doi: 10.1093/petrology/43.11.1987.

Mutter, C.Z., and Mutter, J.C., 1993, Variations in thickness of layer 3 dominate oceanic crustal structure: Earth and Planetary Science Letters, v. 117, p. 295–317, doi: 10.1016/0012-821X(93)90134-U.

Natland, J.H., 1989, Partial melting of a lithologically heterogeneous mantle: Inferences from crystallization histories of magnesian abyssal tholeiites from the Siqueiros Fracture Zone, *in* Saunders, A.D., and Norry, M., eds., Magmatism in the ocean basins, Volume 42: London, Geological Society of London Special Publication, p. 41–77.

Natland, J.H., 2003a, What's going on at Iceland? *in* Foulger, G.R., et al., eds., Penrose Conference Plume IV: Beyond the plume hypothesis, 25–29 August, Hveragerdi, Iceland: Boulder, Colorado, Geological Society of America.

Natland, J.H., 2003b, Capture of mantle helium by growing olivine phenocrysts in picritic basalts from the Juan Fernandez Islands: SE Pacific: Journal of Petrology, v. 44, p. 421–456.

Natland, J.H., and Dick, H.J.B., 2001, Formation of the lower ocean crust and the crystallization of gabbroic cumulates at a very slowly spreading ridge: Journal of Volcanology and Geothermal Research, v. 110, p. 191–233, doi: 10.1016/S0377-0273(01)00211-6.

Natland, J.H., and Dick, H.J.B., 2002, Stratigraphy and composition of gabbros drilled at ODP Hole 735B, Southwest Indian Ridge: A synthesis, *in* Natland, J.H., et al., eds., Proceedings of the Ocean Drilling Program, Scientific Results, Volume 176: College Station, Texas, Ocean Drilling Program, p. 1–69.

Natland, J.H., Adamson, A.C., Laverne, C., Melson, W.G., and O'Hearn, T., 1983, A compositionally nearly steady-state magma chamber at the Costa Rica Rift: Evidence from basalt glass and mineral data DSDP Sites 501, 504, and 505, *in* Cann, J.R., et al., eds., Initial Reports DSDP, Volume 69: Washington, D.C., U.S. Government Printing Office, p. 1065–1077.

Natland, J.H., Meyer, P.S., Dick, H.J.B., and Bloomer, S.H., 1991, Magmatic oxides and sulfides in gabbroic rocks from ODP Hole 735B and the later development of the liquid line of descent, *in* Von Herzen, R.P., and Robinson, P.T.E.A., eds., Proceedings of the Ocean Drilling Program, Scientific Results, Volume 118: College Station, Texas, Ocean Drilling Program, p. 75–111.

Niu, Y., Regelous, M., Wendt, I.J., Batiza, R., and O'Hara, M.J., 2002, Geochemistry of near-EPR seamounts: Importance of source vs. process and the origin of enriched mantle component: Earth and Planetary Science Letters, v. 199, p. 327–345, doi: 10.1016/S0012-821X(02)00591-5.

O'Hara, M.J., 1973, Non-primary magmas and dubious mantle plume beneath Iceland: Nature, v. 243, p. 507–508.

Oskarsson, N., Sigvaldason, G.E., and Steinthorsson, S., 1982, A dynamic model of rift zone petrogenesis and the regional petrology of Iceland: Journal of Petrology, v. 23, p. 28–74.

Oxburgh, E.R., and Parmentier, E.M., 1977, Compositional and density stratification in oceanic lithosphere—Causes and consequences: Journal of the Geological Society of London, v. 133, p. 343–355.

Pedersen, R.B., Malpas, J., and Falloon, T.J., 1996, Petrology and geochemistry of gabbroic and related rocks from Site 894, Hess Deep, *in* Mevel, C., et al., eds., Proceedings of the Ocean Drilling Program, Scientific Results, Volume 147: College Station, Texas, Ocean Drilling Program, p. 3–19.

Pertermann, M., and Hirschmann, M.M., 2003, Anhydrous partial melting experiments on MORB-like eclogite: Phase relations, phase compositions and mineral melt–partitioning of major elements at 2–3 GPa: Journal of Petrology, v. 44, p. 2173–2201, doi: 10.1093/petrology/egg074.

Pilidou, S., Priestly, K., Gudmundsson, O., and Debayle, E., 2004, Upper mantle S-wave speed heterogeneity and anisotropy beneath the north Atlantic from regional surface wave tomography: Iceland and Azores plumes: Geophysical Journal International, v. 159, p. 1057–1076.

Polet, J., and Anderson, D., 1995, Depth extent of cratons as inferred from tomographic studies: Geology, v. 23, p. 205–208, doi: 10.1130/0091-7613(1995)023<0205:DEOCAI>2.3.CO;2.

Poreda, R., Schilling, J.-G., and Craig, H., 1986, Helium and hydrogen isotopes in ocean-ridge basalts north and south of Iceland: Earth and Planetary Science Letters, v. 78, p. 1–17, doi: 10.1016/0012-821X(86)90168-8.

Presnall, D.C., 1995, Phase diagrams of Earth-forming minerals, *in* Mineral physics and crystallography: A handbook of physical constants: Washington, D.C., American Geophysical Union, p. 248–268.

Presnall, D.C., 2003, Low temperature origin of oceanic ridges and associated hot spots: Eos (Transactions, American Geophysical Union), v. 84, Abstracts, p. V31F–01.

Presnall, D.C., Gudfinnsson, G.H., and Walter, M.J., 2002, Generation of mid-ocean ridge basalts at pressures from 1 to 7 GPa: Geochimica et Cosmochimica Acta, v. 66, p. 2073–2090, doi: 10.1016/S0016-7037(02)00890-6.

Ribe, N.M., Christensen, U.R., and Theissing, J., 1995, The dynamics of plume-ridge interaction, 1: Ridge-centered plumes: Earth and Planetary Science Letters, v. 134, p. 155–168, doi: 10.1016/0012-821X(95)00116-T.

Ritsema, J., and Allen, R.M., 2003, The elusive mantle plume: Earth and Planetary Science Letters, v. 207, p. 1–12, doi: 10.1016/S0012-821X(02)01093-2.

Ritsema, J., van Heijst, H.J., and Woodhouse, J.H., 1999, Complex shear wave velocity structure imaged beneath Africa and Iceland: Science, v. 286, p. 1925–1928, doi: 10.1126/science.286.5446.1925.

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

Schilling, J.-G., 1973a, Iceland mantle plume: Geochemical study of Reykjanes Ridge: Nature, v. 242, p. 565–571.

Schilling, J.-G., 1973b, Iceland mantle plume: Nature, v. 243, p. 141–143.

Schilling, J.-G., Meyer, P.S., and Kingsley, R.H., 1982, Evolution of the Iceland hotspot: Nature, v. 296, p. 313–320, doi: 10.1038/296313a0.

Shen, Y., Solomon, S.C., Bjarnason, I.T., and Purdy, G.M., 1996, Hot mantle transition zone beneath Iceland and the adjacent Mid-Atlantic Ridge inferred from P-to-S conversions at the 410- and 660-km discontinuities: Geophysical Research Letters, v. 23, p. 3527–3530, doi: 10.1029/96GL03371.

Shen, Y., Solomon, S.C., Bjarnason, I.T., and Wolfe, C.J., 1998, Seismic evidence for a lower-mantle origin of the Iceland plume: Nature, v. 395, p. 62–65, doi: 10.1038/25714.

Shen, Y., Solomon, S.C., Bjarnason, I.T., Nolet, G., Morgan, W.J., Allen, R.M., Vogfjord, K., Jakobsdottir, S., Stefansson, R., Julian, B.R., and Foulger, G.R., 2002, Seismic evidence for a tilted mantle plume and north-south mantle flow beneath Iceland: Earth and Planetary Science Letters, v. 197, p. 261–272, doi: 10.1016/S0012-821X(02)00494-6.

Sigurdsson, H., and Sparks, R.S.J., 1981, Petrology of rhyolitic and mixed magma ejecta from the 1875 eruption of Askja, Iceland: Journal of Petrology, v. 22, p. 41–84.

Sigurdsson, I.A., Steinthorsson, S., and Gronvold, K., 2000, Calcium-rich melt inclusions in Cr-spinels from Borgarhraun, northern Iceland: Earth and Planetary Science Letters, v. 183, p. 15–26, doi: 10.1016/S0012-821X(00)00269-7.

Slater, L., McKenzie, D., Gronvold, K., and Shimizu, N., 2001, Melt generation and movement beneath Theistareykir, NE Iceland: Journal of Petrology, v. 42, p. 321–354, doi: 10.1093/petrology/42.2.321.

Sleep, N.H., 1990, Hotspots and mantle plumes: Some phenomenology: Journal of Geophysical Research, v. 95, p. 6715–6736.

Sleep, N.H., 2004, Thermal haloes around plume tails: Geophysical Journal International, v. 156, p. 359–362, doi: 10.1111/j.1365-246X.2003.02176.x.

Smith, A.D., and Lewis, C., 1999, The planet beyond the plume hypothesis: Earth Science Reviews, v. 48, p. 135–182, doi: 10.1016/S0012-8252(99)00049-5.

Sobolev, A.V., Hofmann, A.W., and Nikogosian, I.K., 2000, Recycled oceanic crust observed in "ghost plagioclase" within the source of Mauna Loa lavas: Nature, v. 404, p. 986–990, doi: 10.1038/35010098.

Soper, N.J., Strachan, R.A., Holdsworth, R.E., Gayer, R.A., and Greiling, R.O., 1992, Sinistral transpression and the Silurian closure of Iapetus: Journal of the Geological Society of London, v. 149, p. 871–880.

Stakes, D., Mével, C., Cannat, M., and Chaput, T., 1991, Metamorphic stratigraphy of Hole 735B, *in* Von Herzen, et al., eds., Proceedings of the Ocean

Drilling Program, Scientific Results, Volume 118: College Station, Texas, Ocean Drilling Program, p. 254–280.

Staples, R.K., White, R.S., Brandsdottir, B., Menke, W., Maguire, P.K.H., and McBride, J.H., 1997, Faeroe-Iceland ridge experiment 1. Crustal structure of northeastern Iceland: Journal of Geophysical Research, v. 102, p. 7849–7866, doi: 10.1029/96JB03911.

Stein, C., and Stein, S., 2003, Sea floor heat flow near Iceland and implications for a mantle plume: Astronomy and Geophysics, v. 44, p. 1.8–1.10.

Stracke, A., Salters, V.J.M., and Sims, K.W.W., 1999, Assessing the presence of garnet-pyroxenite in the mantle sources of basalts through combined hafnium-neodymium-thorium isotope systematics: Geochemistry, Geophysics, Geosystems, v. 1, doi: 10.1029/1999GC000013.

Stracke, A., Zindler, A., Salters, V.J.M., McKenzie, D., Blichert-Toft, J., Albarede, F., and Gronvold, K., 2003a, Theistareykir revisited: Geochemistry, Geophysics, Geosystems, v. 4, doi: 10.1029/2001GC000201.

Stracke, A., Zindler, A., Salters, V.J.M., McKenzie, D., and Grönvold, K., 2003b, The dynamics of melting beneath Theistareykir, northern Iceland: Geochemistry Geophysics Geosystems, v. 4, doi: 10.1029/2002GC000347.

Stuart, F.M., Lass-Evans, S., Fitton, J.G., and Ellam, R.M., 2003, Extreme $^3He/^4He$ in picritic basalts from Baffin Island: The role of a mixed reservoir in mantle plumes: Nature, v. 424, p. 57–59, doi: 10.1038/nature01711.

Tanton, L.T.E., and Hager, B.H., 2000, Melt intrusion as a trigger for lithospheric foundering and the eruption of the Siberian flood basalts: Geophysical Research Letters, v. 27, p. 3937–3940, doi: 10.1029/2000GL011751.

Tarduno, J.A., and Cottrell, R.D., 1997, Paleomagnetic evidence for motion of the Hawaiian hotspot during formation of the Emperor seamounts: Earth and Planetary Science Letters, v. 153, p. 171–180, doi: 10.1016/S0012-821X(97)00169-6.

Thirlwall, M.F., 1995, Generation of the Pb isotopic characteristics of the Iceland plume: Journal of the Geological Society of London, v. 152, p. 991–996.

Thirlwall, M.F., Upton, B.G.J., and Jenkins, C., 1994, Interaction between continental lithosphere and the Iceland plume Sr-Nd-Pb isotope geochemistry of Tertiary basalts, NE Greenland: Journal of Petrology, v. 35, p. 839–879.

Thirlwall, M.F., Gee, M.A.M., Taylor, R.N., and Murton, M.J., 2004, Mantle components in Iceland and adjacent ridges investigated using double-spike Pb isotope ratios: Geochemica et Cosmochemica Acta, v. 68, p. 361–386, doi: 10.1016/S0016-7037(03)00424-1.

Trønnes, R.G., 1990, Basaltic melt evolution of the Hengill volcanic system, SW Iceland, and evidence for clinopyroxene assimilation in primitive tholeiitic magmas: Journal of Geophysical Research, v. 95, p. 15,893–15,910.

Vink, G.E., 1984, A hotspot model for Iceland and the Voring Plateau: Journal of Geophysical Research, v. 89, p. 9949–9959.

Vinnik, L.P., Du, Z., and Foulger, G.R., 2005, Seismic boundaries in the mantle beneath Iceland: A new constraint on temperature: Geophysical Journal International, v. 160, p. 533–538.

Vogt, P.R., Johnson, G.L., and Kristjansson, L., 1980, Morphology and magnetic anomalies north of Iceland: Journal of Geophysics—Zeitschrift für Geophysik, v. 47, p. 67–80.

Walker, G.P.L., 1963, The Breiddalur central volcano, eastern Iceland: Journal of the Geological Society of London, v. 119, p. 29–63.

White, R.S., and McKenzie, D.P., 1995, Mantle plumes and flood basalts: Journal of Geophysical Research, v. 100, p. 17,543–17,585, doi: 10.1029/95JB01585.

Wolfe, C.J., Bjarnason, I.T., VanDecar, J.C., and Solomon, S.C., 1997, Seismic structure of the Iceland mantle plume: Nature, v. 385, p. 245–247, doi: 10.1038/385245a0.

Wolfe, C.J., Solomon, S.C., Silver, P.G., VanDecar, J.C., and Russo, R.M., 2002, Inversion of body-wave delay times for mantle structure beneath the Hawaiian Islands: Results from the PELENET experiment: Earth and Planetary Science Letters, v. 198, p. 129–145, doi: 10.1016/S0012-821X(02)00493-4.

Yale, M.M., and Morgan, J.P., 1998, Asthenosphere flow model of hotspot-ridge interactions: A comparison of Iceland and Kerguelen: Earth and Planetary Science Letters, v. 161, p. 45–56, doi: 10.1016/S0012-821X(98)00136-8.

Yasuda, A., Fujii, T., and Kurita, K., 1994, Melting phase relations of an anhydrous mid-ocean ridge basalt from 3 to 20 Gpa: Implications for the behavior of subducted oceanic crust in the mantle: Journal of Geophysical Research, v. 99, p. 9401–9414, doi: 10.1029/93JB03205.

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: Contributions to Mineralogy and Petrology, v. 139, p. 326–338, doi: 10.1007/s004100000134.

Yoder, H.S., Jr., 1973, Contemporaneous basaltic and rhyolitic magmas: American Mineralogist, v. 58, p. 153–171.

Yoder, H.S., Jr., and Tilley, C.E., 1962, Origin of basalt magmas: an experimental study of natural and synthetic rock systems: Journal of Petrology, v. 3, p. 342–532.

Zatman, S., Gordon, R.G., and Richards, M.A., 2001, Analytic models for the dynamics of diffuse oceanic plate boundaries: Geophysical Journal International, v. 145, p. 145–156.

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