

*Chapter 1: “The “plate” model for the genesis of melting anomalies” (G.R. Foulger) in  
SPE430: Plates, Plumes, and Planetary Processes (G.R. Foulger and D.M. Jurdy)*

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## *The “plate” model for the genesis of melting anomalies*

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### ABSTRACT

The plate tectonic processes, or “plate,” model for the genesis of melting anomalies (“hotspots”) attributes them to shallow-sourced phenomena related to plate tectonics. It postulates that volcanism occurs where the lithosphere is in extension, and that the volume of melt produced is related primarily to the fertility of the source material tapped. This model is supported in general by the observation that most present-day “hotspots” erupt either on or near spreading ridges or in continental rift zones and intraplate regions observed or predicted to be extending. Ocean island basalt-like geochemistry is evidence for source fertility at productive melting anomalies. Plate tectonics involves a rich diversity of processes, and as a result, the plate model is in harmony with many characteristics of the global melting-anomaly constellation that have tended to be underemphasized. The melting anomalies that have been classified as “hotspots” and “hotspot tracks” exhibit extreme variability. This variability suggests that a “one size fits all” model to explain them, such as the classical plume model, is inappropriate, and that local context is important. Associated vertical motion may comprise pre-, peri-, or post-emplacment uplift or subsidence. The total volume erupted ranges from trivial in the case of minor seamount chains to  $\sim 10^8$  km<sup>3</sup> for the proposed composite Ontong Java–Manihiki–Hikurangi plateau. Time progressions along chains may be extremely regular or absent. Several avenues of testing of the hypothesis are being explored and are stimulating an unprecedented and healthy degree of critical debate regarding the results. Determining seismologically the physical conditions beneath melting anomalies is challenging because of problems of resolution and interpretation of velocity anomalies in terms of medium properties. Petrological approaches to determining source temperature and composition are controversial and still under development. Modeling the heat budget in large igneous provinces requires knowledge of the volume and time scale of emplacement, which is often poorly known. Although ocean island basalt–type geochemistry is generally agreed to be derived from recycled near-surface materials, the specifics are still disputed. Examples are discussed from the Atlantic and Pacific oceans, which show much commonality. Each ocean hosts a single, currently forming, major tholeiitic province (Iceland and Hawaii). Both of these comprise large igneous provinces that are forming late in the sequences of associated volcanism rather than at their beginnings. Each ocean contains several melting anomalies on or near spreading ridges, both time- and non-time-progressive linear volcanic chains of various lengths, and regions of scattered volcanism several hundred kilometers broad. Many continental large igneous provinces lie on the edges of continents and clearly formed in association with continental breakup. Other volcanism is associated with extension in rift valleys, back-arc regions, or above sites

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**of slab tearing or break-off. Specific plate models have been developed for some melting anomalies, but others still await detailed application of the theory. The subject is currently developing rapidly and poses a rich array of crucial but challenging questions that need to be addressed.**

**Keywords:** plate, plume, hotspot, Iceland, Hawaii

## INTRODUCTION

It is common to hear, when debates are held regarding the origin of melting anomalies, “But what alternatives are there to the plume hypothesis?” This question may even be posed immediately after alternatives have just been described. The objective of this article is to lay out the alternative known as the “plate model,” so that future work may build on what has already been achieved rather than comprise reiterations of what has been done in the past. This article does not seek to describe in detail the plume model, nor to compare the plate and plume models, though a few passing remarks on these issues are necessary.

It is common also to hear the statement that no viable alternative to the plume model exists. What is “viable” is a matter of debate. At present no model, plume or otherwise, is without unresolved issues, and it would be premature to accept any without questioning its fundamental validity and subjecting it to rigorous tests (Foulger et al., 2005c; Foulger, 2006a).

The concept of a global “hotspot” phenomenon emerged shortly after plate tectonic theory had been established (Anderson and Natland, 2005; Glen, 2005). Plate tectonics provided an elegant explanation for much of Earth’s volcanism, and in so doing brought into focus the existence of many exceptions. Volcanic regions away from spreading plate boundaries or subduction zones, and high-volume on-ridge volcanism were considered to be unexplained by plate tectonics. The term “hotspot” was coined by Wilson (1963), who suggested that the time-progressive Hawaiian Island chain resulted from motion of the Pacific lithosphere over a hot region in the mantle beneath. The concept was extended by Morgan (1971), who suggested that a global constellation of approximately twenty such “hotspots” exists on Earth. He further postulated that they were fixed relative to one another and that this could be explained if they were fueled by source material delivered in hot plumes that rose by virtue of their thermal buoyancy from the deep mantle. The deep mantle was thought to be convecting only slowly and therefore able to provide a stable reference frame.

The model was further developed by laboratory and numerical convection modeling, which suggested that plumes rise from a thermal boundary layer at the core-mantle boundary and comprise bulbous heads followed by narrow tails (Campbell and Griffiths, 1990). Geochemical study of basalts from ocean islands thought to be “hotspots” revealed compositions distinct from mid-ocean ridge basalt (MORB), in particular, being more enriched in incompatible elements and containing high

$^3\text{He}/^4\text{He}$  ratios (Schilling, 1973; Kellogg and Wasserburg, 1990). Such basalts are called “ocean-island basalts” (OIB). Because it was assumed that “hotspots” are fueled by plumes from the deep mantle, their geochemical characteristics were thought to reveal lower mantle characteristics. The numbers of postulated “hotspots” and plumes rose to approximately fifty or more on many lists (Fig. 1).

The main features and predicted observables of the original classical plume model were recently laid out by Campbell (2006) in a clear and concise summary. These traits are a long period of precursory uplift followed by large igneous province (LIP) eruption, subsequent dwindling of the magmatic rate as the plume head exhausts itself and is replaced by plume tail volcanism, high temperatures, and a conduit extending from the surface to the core-mantle boundary. Fixity relative to other “hotspots” and OIB geochemistry are also expected. High  $^3\text{He}/^4\text{He}$  ratios were predicted following their initial observation in some basalts from Hawaii and Iceland, although this feature is not intrinsic to the original plume hypothesis.

Skepticism about the plume model that has arisen since the beginning of the twenty-first century has resulted largely from a growing awareness that most of these basic predictions are not fulfilled at most “hotspots.” Indeed, no single volcanic system on Earth unequivocally exhibits all the required characteristics listed above (Courtillot et al., 2003; Anderson, 2005b), a full set of which can only be garnered by combining the observations from more than one locality. Even for Hawaii, the type example of a plume, evidence is absent for a precursory LIP with uplift (Shapiro et al., 2006), the locus of volcanism has not remained fixed relative to the paleomagnetic pole (Tarduno and Cottrell, 1997), volcanism is increasing and not dwindling, and as yet a conduit extending from the surface to the core-mantle boundary has not been imaged seismically (Montelli et al., 2004). These difficulties have traditionally been dealt with in the contemporary (in contrast to the classical) plume context by ad hoc adaptations of the model to suit each locality. These adaptations include suggestions that predicted but unobserved features exist but have not yet been discovered, are fundamentally undetectable, or are due to the long-distance lateral transport of melt in the mantle from a distant plume (e.g., Ebinger and Sleep, 1998; Niu et al., 2002). More sophisticated models (e.g., of uplift and mantle convection) have also suggested that some of the original predictions were too simplistic, to the point of negating the very features that the original model was invented to explain (e.g., Steinberger et al., 2004).

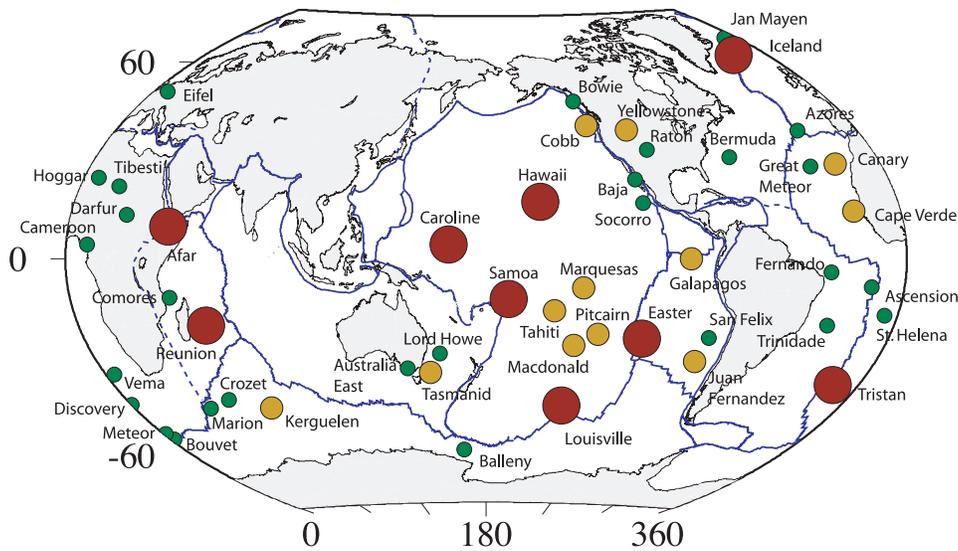


Figure 1. Map showing global “hotspots” compiled using the list of forty-nine of Courtillot et al. (2003). (See commentary on this list by Anderson, 2005b). Red, yellow, and green dots indicate “hotspots” proposed to be sourced at the core-mantle boundary, the base of the upper mantle, and within the lithosphere, respectively. Thin blue lines indicate plate boundaries. For additional maps and relevant resources, see Sandwell et al. (2005).

This approach is unacceptable to many scientists, who have in the past few years worked to develop alternative models. The most widely applicable of these alternatives attributes melting anomalies on Earth’s surface to shallow, plate tectonic–related processes (Anderson, 2000, 2001, 2005a; Foulger and Natland, 2003; Foulger, 2004; Foulger et al., 2005c). Fundamentally, this model is based on processes associated with Earth’s top thermal boundary layer—the surface—where heat is transferred from the solid Earth to the atmosphere. This alternative model is referred to here as the “plate model.” It is the inverse of the plume model, which attributes “hotspots” to processes associated with Earth’s bottom thermal boundary layer, the core-mantle boundary. Extraterrestrial models, involving meteorite impacts, form a separate category of mechanism and will not be dealt with in this article.

Much has already been achieved by a few researchers in a short time and has, most critically, highlighted problems that had gone largely unacknowledged in recent years. In this article, I summarize the basic fundamentals of the plate model, outline some key observations that have inspired its development, and briefly suggest what may be the most critical research goals for the future.

## THE PLATE MODEL

### *Statement of the Model*

The plate model attributes melting anomalies on the surface of Earth to shallow-based processes related to plate tectonics (Anderson, 2001). Simply put, it suggests that where the lithosphere is in extension, permissive volcanism will occur (Favela and Anderson, 1999; Natland and Winterer, 2005). The total magmatic volume is related to the fertility of the source material beneath, and the magmatic rate is related to factors that include the rate of extension, thickness of the lithosphere, fertility, temperature, and the availability of pre-existing melt.

Source fertility varies as a result of the dehomogenizing effects of processes related to plate tectonics, which recycle surface material and thereby refertilize the shallow mantle. At localities where extension occurs over refractory, infertile mantle, little melt will be produced. Where the source is fusible, fertile, and possibly partially molten to begin with, large volumes of melt will be emplaced.

Heat plays a role through being the ultimate driver of plate tectonics, and related temperature variations in the mantle and lithosphere will influence the volume of melt produced. Nevertheless, the plate model for the genesis of melting anomalies is essentially an athermal, top-driven one (Anderson, 2001). Heat is transported via processes including mantle convection, melt advection at mid-ocean ridges, slab subduction, gravitational instabilities, and delamination. Thus temperature and composition vary laterally as a result of plate tectonics. However, melting anomalies are primarily related to the structures and processes associated with plate tectonics, which are close to the surface thermal boundary layer, and not to point-source influxes of heat from the deep mantle that are essentially independent of plate tectonics.

### *Stress*

The plate model suggests that the locations of melting anomalies are controlled by stress (Anderson, 2002). Indicators of extensional stress in the lithosphere include spreading ridges in the oceans, rift zones on land, and magmatism itself. The plate model predicts that surface melting anomalies will be preferentially located at or near these features. This prediction is largely borne out by the observation that approximately one-third of all melting anomalies are on or close to mid-ocean ridges (Fig. 1). Of these, several are at localities where extension is exceptionally large as a result of ridge-ridge-ridge triple junctions (e.g., the Easter, Afar, Bouvet, and the Azores melting anomalies). The Easter and

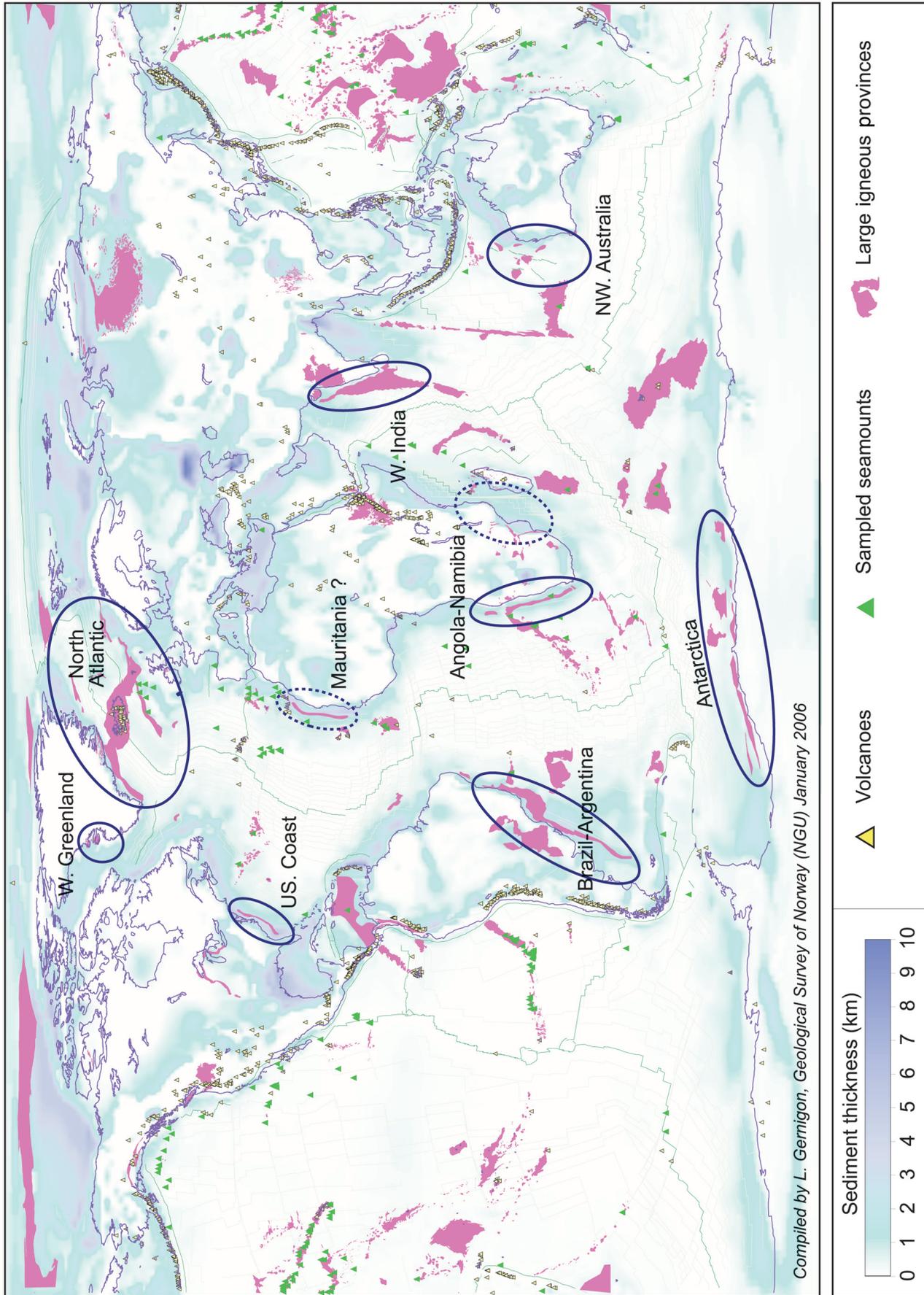


Figure 2. Map of global large igneous provinces. B&R—Basin and Range province; CRB—Columbia River basalts; EAR—East African rift; HLP—High Lava Plains; Y-SRP—Yellowstone–Snake River Plain. Base map by Laurent Gernigon, Geological Survey of Norway, January 2006. Data after Coffin and Eldholm (1994), Simkin and Siebert (1994), and Laske and Masters (1997). Coastline from USGS. Isochrons after Müller et al. (1997).

Iceland melting anomalies are also associated with oceanic microplates (Schilling et al., 1985a; Foulger, 2006b). Oceanic LIPs that formed at or near ridges include Shatsky Rise, which formed at a ridge-ridge-ridge triple junction (Sager et al., 1999; Sager, 2005) and the Iceland plateau, which is currently being created (Fig. 2). It is speculated that the Ontong Java plateau also formed near a ridge (Larson, 1997), and that it comprises only part of a much larger plateau that includes the Manihiki and Hikurangi plateaus (Taylor, 2006), but these suggestions remain to be confirmed.

Most melting anomalies in continental regions occur within or on the edges of extending regions or rift zones or at lithospheric discontinuities that predate the extension and magmatism. The extending Basin and Range province in the western United States is associated with widespread volcanism, including the persistent Yellowstone anomaly that lies on its northern edge. The East African rift is the location of much volcanism, including Afar at its northern end. Cameroon volcanism occurs on the southern flank of the parallel Benue Trough (Fitton and James, 1986). Many LIPs that now lie partially on continental lithosphere erupted where rifting led to continental breakup (e.g., the North Atlantic Igneous province, the Central Atlantic Magmatic province, and the Deccan Traps). Others formed as continents converged (e.g., Silver et al., 2006).

Some low-volume oceanic melting anomalies have been attributed to intraplate extension (e.g., due to thermal contraction; Sandwell et al., 1995; Lynch, 1999; Smith, 2003; Sandwell and Fialko, 2004). Global stress field maps show that volcanic regions occur in areas of tensile stress (Lithgow-Bertelloni and Richards, 1995; Lithgow-Bertelloni and Guynn, 2004). Calculations of the stress field from cooling for the whole Pacific plate predict large extensional stresses—in particular, near Samoa, Easter, and Louisville—and contraction normal to the orientation of the Hawaiian chain (Lithgow-Bertelloni and Guynn, 2004; Stuart et al., this volume). Globally, plate-circuit-closure calculations suggest that the total amount of thermal contraction occurring in the interior of oceanic plates is equivalent to a slow-spreading ridge (Gordon and Royer, 2005).

The question arises whether the presence of volcanism per se may be interpreted as evidence for extension (Favela and Anderson, 1999). Volcanic chains or lineations are expected to develop along extensional structures, such as fissures, faults, or cracks, and their orientation may thus be interpreted as indicating instantaneous orientation of extensional stress. Point sources of volcanism may be interpreted as localized extensional stress, and migration of the locus of volcanism with time may be interpreted as migration of the locus of extension. The building of volcanic edifices may also modulate stress locally and influence the location of subsequent eruptions (Hieronymus and Bercovici, 1999, 2000).

### Source Inhomogeneity

Plate tectonic processes that create a heterogeneous mantle include melt extraction at mid-ocean ridges and subduction of

oceanic lithosphere at trenches. Fertile and enriched material is continually advected from below into both the oceanic and continental mantle lithosphere, causing metasomatism and gravitational instability of over-thickened crust and lithosphere. Detachment recycles this material back into the mantle. Inhomogeneities are distributed throughout the shallow mantle by convection. As a result of these processes, the mantle is not uniformly comprised of peridotite or pyrolite but may vary radically in fertility from place to place (Meibom and Anderson, 2004; Anderson, 2005a; Fig. 3).

At mid-ocean ridges, plate separation proceeds hand in hand with the extraction of the most fusible 10–20% of the mantle source beneath, which is intruded and erupted to form new ocean crust. During the oceanic lithosphere’s journey across the ocean, its thickness continually increases as it cools, asthenospheric mantle is plated onto its bottom, it is metasomatized, and sediments accumulate on its top. The slab that is re-injected into the mantle at subduction zones is thus a complex lithological package that includes the various rock types that make up the crustal section and the mantle lithosphere. The basaltic and gabbroic portions become transformed to eclogite at depths of ~50–60 km. At subduction zones, the downgoing slab experiences geochemical alteration as a result of dehydration and selective extraction of components into the mantle wedge. At

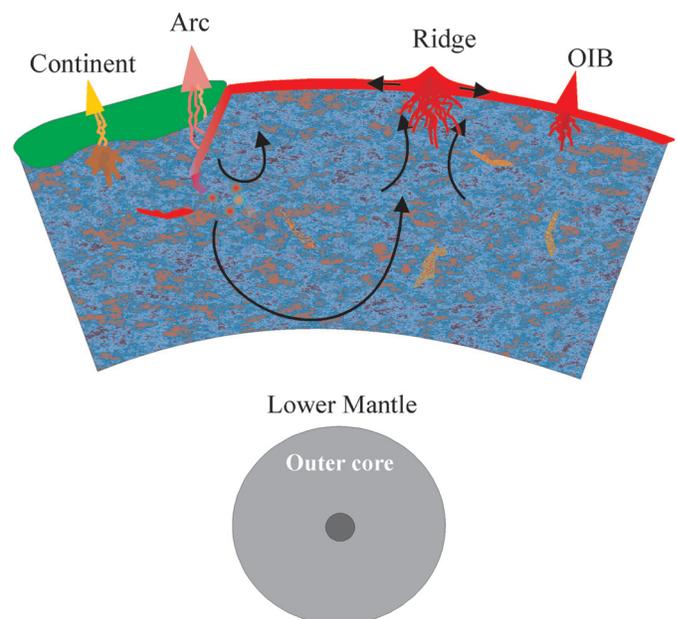


Figure 3. Schematic illustration (not to scale) of the upper mantle as a heterogeneous assemblage of depleted, infertile residues (bluish colors) and enriched, fertile, subducted oceanic crust, lithosphere, and sediments (reddish colors). The heterogeneities are statistical in nature and have wide ranges in shape, size, melting point, age, and origin (from Meibom and Anderson, 2004).

greater depths, different parts of the slab may become separated and recycled in the mantle separately.

Where the lithosphere becomes thickened (e.g., at island arcs and in collision belts), eclogitization of the lower crust and mantle lithosphere may occur, increasing its density and causing it to become negatively buoyant (Kay and Kay, 1993). A Rayleigh-Taylor (gravitational) instability may then develop, causing the mantle lithosphere and possibly also parts of the lower crust to detach and become recycled in the upper mantle (Tanton and Hager, 2000; Elkins-Tanton, 2005; Fig. 4). The continental mantle lithosphere thus recycled may be very old and may have been refertilized by metasomatism. Other mechanisms for detaching and recycling the lower continental crust into the mantle have also been suggested (Anderson, this volume).

To what depths are these materials recycled? Whereas lithospheric detachment events may be episodic, slab subduction is widespread and its locality well known. Seismology and mineral physics suggests that the depths to which slabs sink are variable. A high-velocity body that is apparently continuous from the transition zone to the core-mantle boundary has been detected seismically beneath North America and is commonly interpreted as a subducted slab (e.g., Grand, 1994). However, this scenario is the exception rather than the rule, and volume considerations cast doubt on whether a slab interpretation for this entire structure is plausible. Other down-going slabs clearly do not penetrate into the lower mantle (e.g., in the western Pacific; Ritsema et al., 1999; Gorbatov et al., 2000). It is likely that much material, both subducted and detached, is recycled in the upper mantle. Likely depths of density equilibration have been suggested by Anderson (2005a).

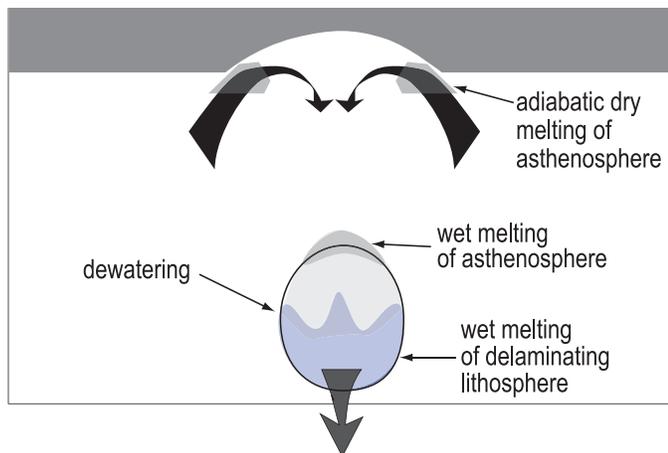


Figure 4. Schematic model for magma production during detachment of the mantle lithosphere as a result of gravitational instability. As the lithosphere sinks, asthenosphere is sucked into the resulting evacuated dome and may melt adiabatically. If the instability is hydrous, it may dehydrate as it sinks and heats, triggering melting of the mantle or of itself (from Elkins-Tanton, 2005).

## OBSERVATIONS

### Vertical Motion

The pattern of vertical motion associated with melt extraction is potentially a strong diagnostic of the source process. Differential vertical motion is expected to accompany transient volcanism, in particular LIP emplacement, and to scale with the volume of eruption. It can be studied using sedimentology on both terrestrial and marine sediments deposited prior to, during, and following eruption, and geomorphology. Of particular interest is comparing observations with the predictions of the classical, simple plume model that widespread regional domal uplift precedes LIP emplacement by 10–20 Ma (Campbell, 2006). This emplacement is expected to be followed by post-eruption subsidence as the LIP is carried by plate motion away from the hot plume (Griffiths and Campbell, 1991; Clift, 2005). Lithospheric detachment models, in contrast, predict that surface subsidence is associated with detachment of the dense, thickened lower lithosphere and precedes the LIP volcanism that occurs as mantle material from greater depths flows in to replace the lost lithosphere (Kay and Kay, 1993; Elkins-Tanton, 2005).

The picture that emerges from observations is variable, with precursory or syn-eruption uplift associated with some LIPs but not others. The Siberian Traps and the proposed composite Ontong Java–Manihiki–Hikurangi plateau represent the world’s largest continental and oceanic LIPs, respectively, and yet neither is associated with major precursory uplift (Neal et al., 1997; Czamanske, 1998; Ito and Clift, 1998; Ingle and Coffin, 2004; Ivanov, this volume). The Siberian Traps have a volume of some  $4 \times 10^6 \text{ km}^3$  (Masaitis, 1983) and were erupted at ca. 200 Ma onto old continental lithosphere with a thickness of 100–200 km. They overlie the largest coal resource in the world and have been the subject of intensive research, including the drilling of hundreds of boreholes. The geology and stratigraphy of the region are thus known in great detail. The results show that the region subsided rapidly and continually prior to and during LIP emplacement, and that this subsidence was substantial and widespread (Czamanske, 1998; Ivanov, this volume).

The proposed composite Ontong Java–Manihiki–Hikurangi plateau has a combined volume of  $\sim 10^8 \text{ km}^3$  (Taylor, 2006). A plume origin for such a vast LIP would predict pre-eruption uplift of several kilometers. However, paleodepth estimates from volatiles in glass samples and from the study of sediments immediately overlying the crust from several ocean drill holes give elevations ranging from approximately sea level down to  $\sim 2.5$  km below sea level at the time of eruption (Michael, 1999; Ingle and Coffin, 2004; Roberge et al., 2005). Subsidence subsequent to its formation has also been studied using oceanic sediments and was found to be slower than for normal oceanic lithosphere, consistent with an underlying mantle that is cooler than average, not warmer (Clift, 2005).

The pattern of vertical motion accompanying emplacement of the North Atlantic Igneous province is more complicated.

Volcanism occurred in two phases. The first occurred at 61–59 Ma and was distributed in a northwest-trending belt passing through northern Britain and east and west Greenland. The second phase occurred at 54 Ma, was associated with the opening of the North Atlantic, and was distributed in a NNE-trending belt along the newly formed passive margins. The first phase was preceded by limited uplift in the vicinity of some igneous centers. Kilometer-scale regional uplift did not occur until the second stage. It was contemporaneous with the second stage of volcanism and did not significantly predate it, as is expected for an impinging plume (MacLennan and Jones, 2006). It was closely associated with the new continental margins and comprised several episodes of uplift and magmatism. The main phase of uplift occurred in a short burst that may have lasted only a few tens of thousands of years. The pattern of subsequent subsidence permits only moderate temperature anomalies of no more than  $\sim 100$  °C down to  $\sim 100$  km depth and precludes very high temperature or very thick thermal sources (Clift, 2005).

Simpler, precursory uplift has been reported to be associated with the ca. 250-Ma Emeishan LIP in southwest China. There, variations in thickness of sedimentary rocks underlying the basalts suggest kilometer-scale domal uplift throughout a region  $\sim 900$  km wide (He et al., 2003; Xu and He, this volume). Even though other cases have been reported (Rainbird and Ernst, 2001), this very recently reported result has been cited as the best example of domal uplift associated with LIP emplacement (Campbell, 2006).

Brief mention may be made of other LIPs. There is no unequivocal evidence for regular, regional uplift prior to emplacement of the Deccan Traps, which erupted at ca. 67 Ma and had a volume of  $\sim 1.5 \times 10^6$  km<sup>3</sup> before erosion. Instead there is ev-

idence that such uplift did not occur (Sheth, 2006a, this volume). There is evidence of subsidence prior to emplacement of the Columbia River basalts and of uplift immediately afterward (Hales et al., 2005), though this sequence of events is contested (Hooper et al., this volume). Study of sedimentary cores from many oceanic plateaus has produced little evidence for post-emplacement subsidence (e.g., at Shatsky Rise, Kerguelen plateau, Walvis Ridge, Ninetyeast Ridge; Clift, 2005).

The picture that emerges in general is one of considerable complexity. At some well-studied LIPs, notably the world’s largest continental and oceanic ones, either subsidence or no uplift is observed prior to emplacement. Precursory, kilometer-scale uplift is not observed at most volcanic rifted margins. Uplift is associated with LIP emplacement at some localities, but it may not occur prior to eruption, and its pattern may be complex. Most oceanic plateaus subsided little, if at all, following volcanism, suggesting only minor temperature anomalies, if any, in the mantle when they formed.

In view of these variable observations, it is interesting to note the recent work of Burov and Guillou-Frottier (2005b), who modeled the uplift predicted to accompany the influx of hot source material using realistic continental lithosphere rheologies. They modeled lithosphere that included elastic-brittle-ductile properties and a stratified structure. Their results suggest that the simple classical expectation of regional domal uplift is oversimplified, and a more complex pattern might occur (Fig. 5). This work reopens the question regarding the vertical motions expected to accompany all source models. It suggests that a reassessment of predictions is required, and brings into question whether the occurrence of precursory domal uplift may be used to test the plume hypothesis.

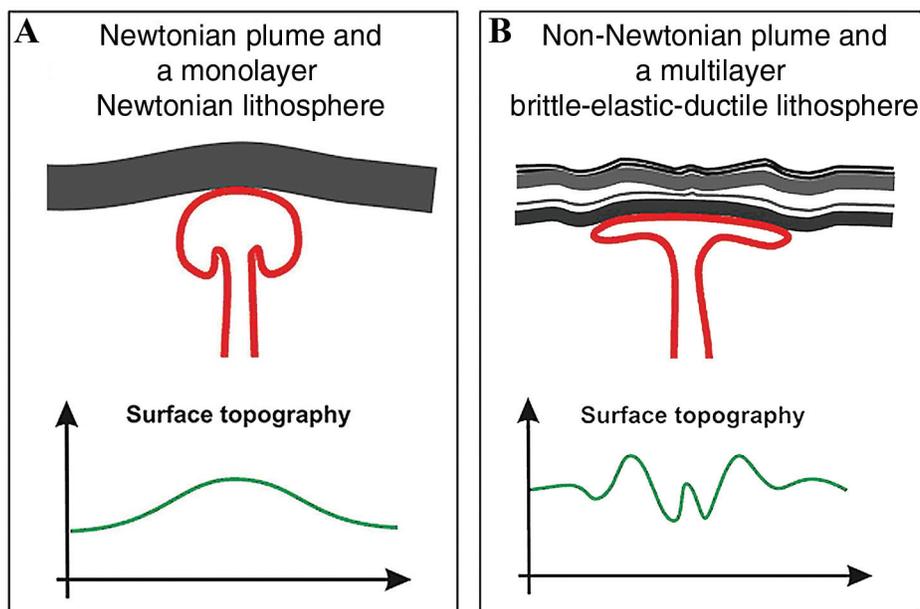


Figure 5. Sketches of plume-lithosphere interactions. (A) A “conventional” plume impinges on a single-layer viscous lithosphere, resulting in a single, long-wavelength topographic signature. (B) When realistic lithosphere rheology (brittle-elastic-ductile) and multilayer structure are considered, several wavelengths of surface topographic undulations are expected and both local uplifts and subsidences are predicted (from Burov and Guillou-Frottier, 2005a).

### Volume, Rate, and Chronology of Volcanism

The total volume of magma and rate and chronology of emplacement at melting anomalies all vary throughout their entire feasible ranges. The high-volume end of the spectrum is represented by LIPs, the largest of which is the  $\sim 10^8$  km<sup>3</sup> proposed composite Ontong Java–Manihiki–Hikurangi plateau (Taylor, 2006). Determining the lowest-volume LIP is, however, a matter of definition. LIPs were originally defined as having an areal extent of  $>0.1 \times 10^6$  km<sup>2</sup> (Coffin and Eldholm, 1994) but Sheth (2007a), as part of a proposed classification scheme for LIPs, recently suggested that  $0.05 \times 10^6$  km<sup>2</sup> is a more appropriate lower limit (see Sheth, 2006b, and linked Web pages for a substantial discussion; Bryan and Ernst, 2007). Such debate hints that the sizes of volcanic provinces may form a continuous spectrum and invites the question of whether there is a natural break in the size distribution that might form the basis for theories of genesis.

At the low end of the volume scale are such island chains as the Samoa, Tahiti, Marquesas, Pitcairn, and Louisville chains (Fig. 1). The eruption rates of these chains throughout their lifetimes have been much less than 0.01 km<sup>3</sup>/yr. In some cases, chains are discontinuous, and the eruption rate was almost zero for long periods (e.g., for the youngest ca. 30 Ma of the Louisville chain). In these cases, extrapolations or interpolations are speculative and model-dependent, and it is valid to question whether the volcanism results, in some sense, from a single “source.”

Magmatic rates also vary widely. Thick sequences of lavas with similar ages are observed at many LIPs, leading to the view that they are emplaced very rapidly at rates that may exceed 1 km<sup>3</sup>/yr and be sustained for one or more millions of years. However, age dating is insufficiently accurate to provide a precise figure, and where much of the LIP volume is inaccessible to sampling, such an assumption may not be supported by observational evidence. For example, it is commonly assumed that the Ontong Java plateau was emplaced rapidly, perhaps over just a few million years, giving a sustained magmatic rate of up to several km<sup>3</sup>/yr (Fitton et al., 2004). However, this rate is conjectural, as the thick lower crust cannot be sampled. Ito and Clift (1998) suggested that the slow postemplacement subsidence observed there might be explained if the lower crust continued to be thickened by intrusions from a plume tail over a period of several tens of million years, but this seems unlikely, if only because the plateau drifted by  $\sim 2000$  km between 120 and 90 Ma (Kroenke et al., 2004).

The very high eruption rates thought to occur at LIPs may be compared with the much lower rates at island chains. The magmatic rate along the Emperor and Hawaiian chains is shown in Figure 6. No pre-chain LIP is credibly associated with the Emperor chain. Its older end abuts the Aleutian Trench, but thick oceanic plateaus are probably not completely subductable (Cloos, 1993). There is no evidence for recycled subducted Emperor chain material in the lavas of Kamchatka (Shapiro et al., 2006).

The magmatic rate during emplacement of the Emperor and Hawaiian chains was modest for most of their history (Fig. 6).

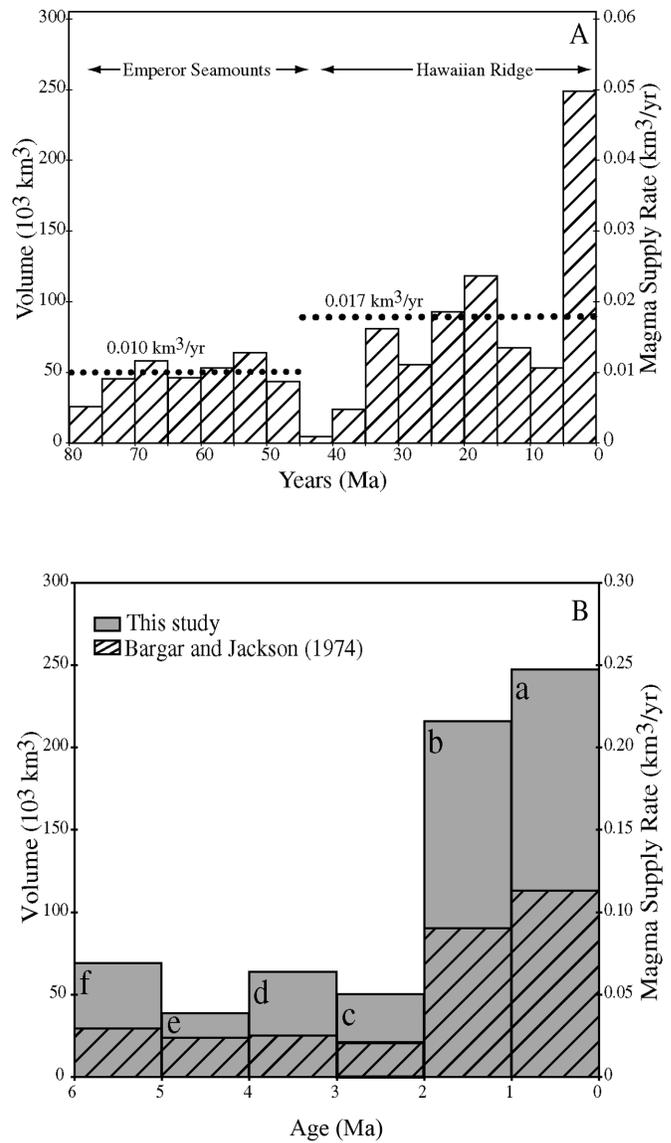


Figure 6. Volumes and magma supply rates for the Emperor and Hawaiian volcanic chains. (A) Histogram along the Emperor seamounts and Hawaiian ridge at 5-Ma intervals (from Bargar and Jackson, 1974). Dotted lines mark the long-term average magma supply rates. (B) Histogram of the Hawaiian Islands at 1-Ma intervals. Gray bars show volumes and magma supply rates calculated by Robinson and Eakins (2006), who take into account extra volume contained by subsidence of the Cretaceous sea floor. Hatched bars show the results from Bargar and Jackson (1974).

It was  $\sim 0.01$  km<sup>3</sup>/yr for much of the lifetime of the Emperor chain. This rate is approximately that of a single mid-ocean-ridge spreading segment. The volume rate declined to almost zero at the time of the bend (ca. 47 Ma; Sharp and Clague, 2006), and grew to somewhat higher rates, or up to  $\sim 0.17$  km<sup>3</sup>/yr for much of the Hawaiian chain. Very recently however, the magmatic production rate underwent a rapid increase, and is currently  $\sim 0.25$  km<sup>3</sup>/yr (Robinson and Eakins, 2006), which is of

the order of the magmatic rate inferred for some LIPs. The young, high-volume end of the Hawaiian chain that formed over the past ca. 5 Ma has an area of  $\sim 0.15 \times 10^6 \text{ km}^2$  and a volume of  $\sim 0.6 \times 10^6 \text{ km}^3$  (Joel Robinson, personal commun., 2006) which means it is a LIP, even according to the higher-end classification scheme of Coffin and Eldholm (1994). The Hawaiian archipelago thus comprises a LIP forming at the end of a time-progressive chain rather than at its beginning.

The Iceland region is another locality where magmatic rate has increased with time rather than dwindled. Magmatism in excess of local ocean-crust formation has occurred there for the  $\sim 54\text{-Ma}$  duration of the opening of the north Atlantic. Following continental breakup, the excess magmatic rate was less than  $\sim 0.1 \text{ km}^3/\text{yr}$  for the first  $\sim 10 \text{ Ma}$ , during which time the Iceland-Faeroe Ridge was built, with its modest north-south extent of  $\sim 100 \text{ km}$  (Fig. 7). Since ca. 44 Ma, however, the Icelandic Volcanic plateau has been built. It has a north-south extent of up to  $\sim 600 \text{ km}$  and a crustal thickness of up to  $\sim 20 \text{ km}$  greater than

the crust on the Reykjanes Ridge. It must thus have been built at an excess rate of  $\sim 0.25 \text{ km}^3/\text{yr}$ , approximately the same as at present-day Hawaii. The Iceland plateau presently covers an area of  $\sim 0.3 \times 10^6 \text{ km}^2$  and thus has an excess volume of  $\sim 5 \times 10^6 \text{ km}^3$ . It therefore also comprises a LIP that is forming late in the sequence of assumed associated magmatism rather than at the beginning.

The formation of a LIP followed by continuation of volcanism to form a time-progressive volcanic chain is a rare exception rather than the general rule (e.g., Anderson, 2005b; Beutel and Anderson, this volume). No trace of such associations exist, for example, for the Siberian Traps, the Ontong Java plateau, Afar, Hawaii, Samoa, or Louisville. Of the forty-nine “hotspots” cataloged by Courtillot et al. (2003) only thirteen associations of LIPs and volcanic chains are listed (Crozet, Easter, Fernando, Galapagos, Iceland, Kerguelen, Macdonald, Marion, Marquesas, Réunion, Tristan, Vema, and Yellowstone), and only two of these are cited as being unequivocal (Réunion/Deccan

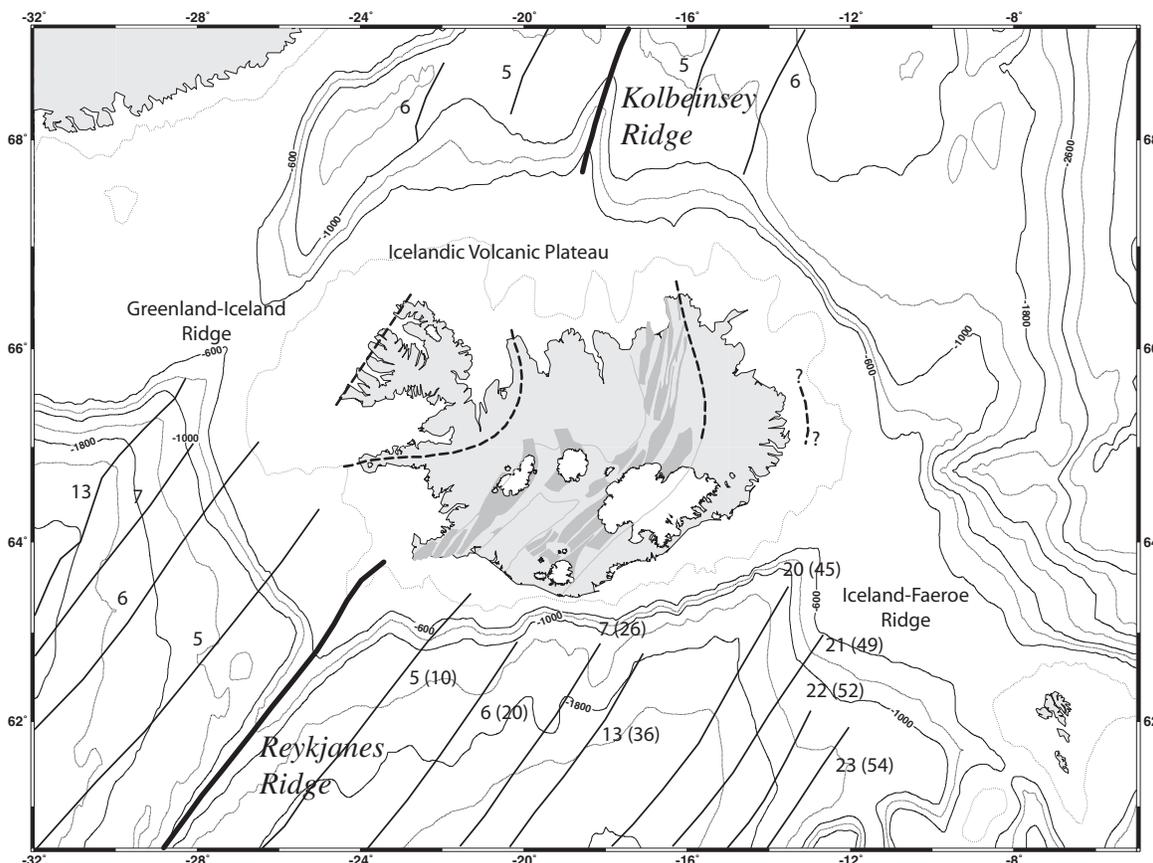


Figure 7. Map of the Iceland region, showing bathymetric contours, which are thought to indicate the extent of thick crust in the region. The island of Iceland makes up the innermost, subaerial part of the broader, mostly submarine, Icelandic volcanic plateau. Oceanic magnetic anomalies (Nunns, 1983) are labeled with anomaly number. Approximate ages in Ma are shown in parentheses after the anomaly number on the eastern flank of the Reykjanes Ridge. Thick black lines—axes of the Reykjanes and Kolbeinsey ridges; thin lines on land—outlines of the neovolcanic zones; gray—spreading segments; white—glaciers; dashed lines—extinct rift zones (two in west Iceland and two in east Iceland).

and Tristan/Parana). Both of these are contested by other authors, however (Burke, 1996; Fairhead and Wilson, 2005; Sheth, 2005).

### Structure of the Mantle from Seismology

Seismology is able to probe the mantle in a focused way that is not possible using any other method. Many different approaches are available that vary in target and resolution. Of particular interest is the question of what structures underlie currently active melting anomalies, whether they are thermal, and whether they are continuous between the surface and the deep mantle.

Teleseismic tomography uses regional seismometer networks and can resolve structure on a scale of 50–100 km down to depths roughly equal to the breadth of the network. This breadth is typically a few hundred kilometers. Such experiments have been conducted to study the Iceland, Yellowstone, and Eifel melting anomalies (Iyer et al., 1981; Ritter et al., 2000; Foulger et al., 2001; Christiansen et al., 2002). The results in all these cases are consistent with the presence of underlying low-velocity bodies that are confined to the upper mantle. No teleseismic tomography experiment has yet been conducted on a scale that enables structure through and beneath the mantle transition zone, 410–660 km deep, to be well imaged. Thus, the continuity of structures between the upper and lower mantle has not yet been studied using this method.

Global tomography provides a continuous image of the mantle but at the much lower resolution of 500–1000 km (e.g., Ritsema et al., 1999). The results show a sharp contrast in structural character between the upper and lower mantles. Variations in the spherical harmonic power of velocity throughout the mantle shows that the Earth is characterized by strong heterogeneity in the longer-wavelength components in the upper mantle but weak heterogeneity in the lower mantle (Gu et al., 2001). At individual melting anomalies (e.g., Iceland, Yellowstone, Eifel), global tomography generally shows that the underlying low-velocity structures are confined to the upper mantle. Recent reports of low-velocity anomalies beneath some “hotspots” traversing the entire mantle based on finite-frequency tomography (e.g., Montelli et al., 2004) have been seriously challenged and remain to be confirmed (van der Hilst and de Hoop, 2005).

To study in detail possible continuity of structures through the transition zone, receiver functions have been applied to determine topography on the 410-km and 660-km discontinuities that comprise its upper and lower boundaries, respectively. These discontinuities result from mineralogical phase changes, and their depths are predicted to change by measurable amounts in the presence of temperature or compositional anomalies (Bina and Helffrich, 1994; Presnall, 1995; Fig. 8). The results in general show broad global correlations, with a thin transition zone beneath oceans and a thicker one beneath continents, but within individual provinces, there is little correlation between the topography on the two discontinuities (Gu and Dziewonski,

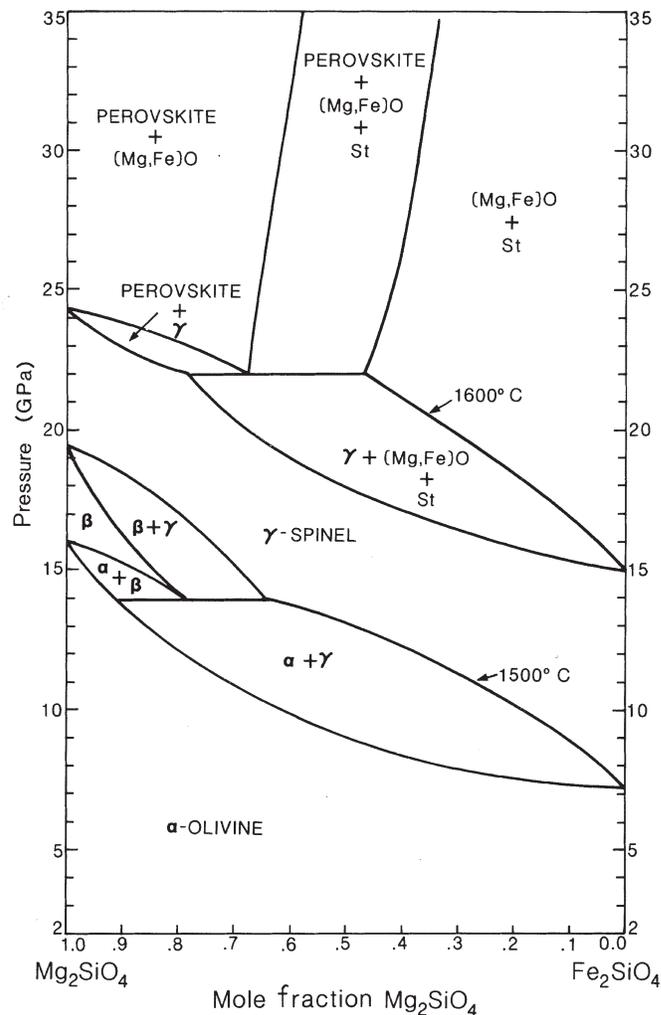


Figure 8. Mineralogical phase changes in the transition zone (from Presnall, 1995). This figure assumes a homogeneous mantle. For a discussion of the role of eclogite in the transition zone, see Anderson (this volume). Abbreviation: st—stishovite.

2001). This pattern is also observed at melting anomalies (e.g., Yellowstone, Iceland, Eifel). There, deflections on the 410-km discontinuity are observed but no corresponding deflections on the 660-km discontinuity (Dueker and Sheehan, 1997; Grunewald et al., 2001; Du et al., 2006).

Global tomography suffers from formidable problems of resolution and coverage, which become increasingly severe with depth, and there are many regions of the mantle devoid of seismic rays. Repeatability among different models is also a problem, with proposed lists of deeply sourced “hotspots” often exhibiting little overlap (Anderson, 2005b; Table 1). Nevertheless, the greatest barrier to using seismology to study melting anomalies is the inherent ambiguity in the physical interpretation of velocity anomalies. Seismic velocity varies because of variations in composition, mineralogical phase, anisotropy, the pres-

**TABLE 1. HOTSPOTS REPORTED TO BE UNDERLAIN BY SEISMIC ANOMALIES TRAVERSING THE UPPER MANTLE ONLY AND THE WHOLE MANTLE, AND HOTSPOTS DEFINED AS ARISING FROM D”**

Hotspot name	Traversing the upper mantle only*	Traversing the whole mantle†	Defined as arising from the core-mantle boundary‡
Afar	X		X
Hawaii	X		X
Bowie	X		
Samoa	X	X	
Macdonald	X		
Easter	X	X	X
Louisville	X		X
Iceland	X		X
Tahiti		X	
Azores		X	
Canaries		X	
Ascension		X	
Réunion			X
Tristan			X

Note: Only Easter is common to all three lists.

\*Traversing the upper mantle only—the deepest detected (Ritsema and Allen, 2003).

†From Montelli et al. (2004).

‡Arising from D”, according to Courtillot et al. (2003).

ence of partial melt and volatiles, and temperature (Anderson, 1989; Fig. 9). Interpreting anomalies is thus highly ambiguous, and nowhere is this difficulty better illustrated than in the case of the Pacific and south Atlantic “superplumes.” These are low-velocity bodies several thousand kilometers broad that extend from the core-mantle boundary to the shallow mantle. It is widely assumed that they are thermal in origin and fuel surface melting anomalies either directly or indirectly. However, Trampert et al. (2004) were able to separate out the thermal and compositional components using bulk sound wave and shear wave velocities obtained from normal modes. They found that the “superplumes” are largely compositional in origin and anomalously dense, not thermally buoyant. Very recent high-pressure experiments on the silicate phases that exist at core-mantle boundary depths suggest that the low-velocity anomalies in the D” layer are also dense and not thermal in origin (Mao et al., 2005).

In addition to having significant bearing on geodynamic models of the mantle, these results are an important cautionary tale concerning interpretation of seismic velocity anomalies. They also highlight the general scientific rule that it is dangerous to interpret observations under the assumption that one out of many possible interpretations is correct.

### Temperature and Heat

The extraction of large volumes of melt requires either high temperature, a fusible source, or both, along with an appropri-

ate state of lithospheric stress. Study of temperature variations in the presumed source region is thus important for determining the source process. This study is not an easy task, however. The interpretation of seismic velocity variations is ambiguous. Surface heatflow is insensitive to mantle temperature because thermal conduction in rocks is slow, and ground-water circulation may complicate matters. Thus, petrological methods may be important. However, in between melt formation and eruption on the surface, basalts experience a complex history, and as a result, the field of geothermometry is still controversial and developing rapidly.

Because the absolute potential temperature of the mantle is poorly known, temperature differences between volcanic provinces are generally studied. Specifically, melting anomalies are compared with mid-ocean ridges, which are assumed to represent average background mantle temperature. The most robust approach is to study olivine control lines. When a primary mantle-derived melt cools in a crustal magma chamber, its composition initially follows a liquid line of descent controlled by the precipitation of olivine only, during which time its MgO contents decrease. If samples of the instantaneous liquids formed during this cooling trajectory are preserved and erupted, they may be sampled. The original composition of the parental/primary liquid may then be back-calculated. Olivine is incrementally added (i.e., the reverse of crystal fractionation) to an evolved liquid composition (which itself lies on the olivine-only crystallization path) to obtain equilibrium with the most mag-

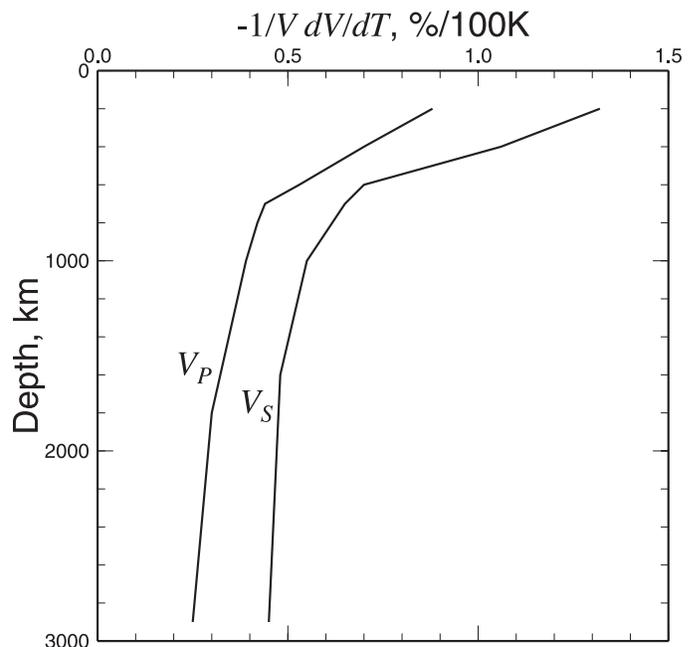


Figure 9. Sensitivity of seismic wavespeeds to temperature at different depths in the Earth’s mantle (after Karato, 1993). These functions depend on composition and mineralogy and are uncertain by perhaps 30% (reproduced from Julian, 2005).

nesian olivine phenocryst observed, using an olivine geothermometer (Green and Falloon, 2005). The composition of the parental liquid can then be compared with experimental mantle melt compositions to deduce a pressure and temperature of mantle equilibration and a mantle potential temperature. It is necessary to assume a number of parameters for this calculation, including source composition, degree of partial melting, pressure of initial melting, latent heat of fusion, and heat capacities.

The method has been applied to Hawaii, which is the only locality where high-MgO basaltic glasses have been found (Clague et al., 1991). Attempts have been made to apply it to other regions (e.g., Iceland), but the only high-MgO basalts found there are cumulate rocks that cannot safely be assumed to represent liquid compositions. Putirka (2005), using an olivine geothermometer described in the same paper, reported a temperature of  $\sim 245 \pm 52$  °C hotter for Hawaii compared with mid-ocean ridges for olivine crystallization temperatures of parental liquids. Based on this result, Putirka (2005) postulated that there is a difference in mantle potential temperature of  $\sim 250$  °C between the Hawaiian and MORB sources. Green et al. (2001), Green and Falloon (2005), and Falloon et al. (this volume), however, report very similar temperatures for the olivine crystallization temperatures of parental liquids for basalts from Hawaii, Iceland, and Réunion and a subset of MORBs. They followed a similar approach but used the olivine geothermometer of Ford et al. (1983). That geothermometer uses partition coefficients more appropriate for olivine crystallization at crustal pressures than does any other olivine geothermometer. The parental liquid compositions calculated by Green et al. (2001), Green and Falloon (2005), and Falloon et al. (this volume) show very similar pressures and temperatures of mantle equilibration when compared with the relevant experimental data. They thus conclude that it is unlikely that significant differences exist between the potential temperatures of the mantle sources for Hawaiian and the hottest MORB parental liquid compositions. The differences between the results of Putirka (2005) and Green and Falloon (2005) using the same approach can be explained entirely by the differences in performance of the olivine geothermometers used (Falloon et al., 2007a).

The question of heat is separate from that of temperature. The advection of large volumes of melt to the near-surface requires a mechanism for extracting large amounts of heat from the mantle. The heat required consists of the specific heat needed to raise the source material to its solidus, plus latent heat of melting. The amount of melt that can be produced by thermal upwellings was modeled numerically by Cordery et al. (1997) and Leitch et al. (1997). The results showed that a pyrolite upwelling is incapable of generating any melt at all for reasonable temperature anomalies and lithosphere ages greater than a few million years. Melting of entrained eclogite, which is more fusible, was therefore modeled. Figure 10 shows the volumes of melt produced assuming that all the melt came from the fusible eclogite component and the latent heat was obtained from conduction from the surrounding material, not supplied by decompression.

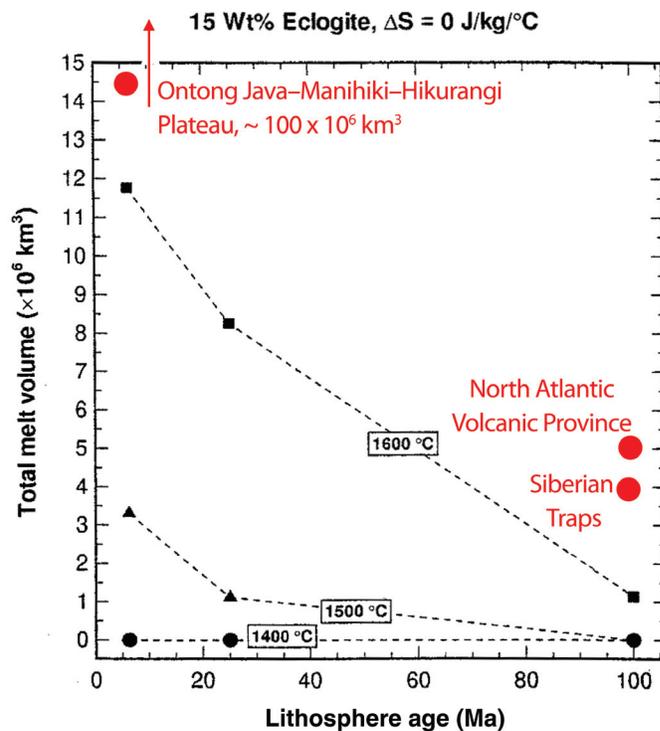


Figure 10. Volumes of melt calculated using finite-element modeling of plumes rising from the core-mantle boundary, plotted against lithosphere age, which is a proxy for lithosphere thickness. Numbers indicate temperature anomaly, assuming a normal mantle potential temperature of 1300 °C. It is assumed that all the melt comes from fusible eclogite entrained in the plume, and that the latent heat is obtained by conduction from the surrounding material. It is not supplied by decompression, and thus may be set to be zero in the calculations (adapted from Cordery et al., 1997). This model cannot account for the volumes of magma observed in the largest of the large igneous provinces for reasonable temperature anomalies.

Only small or moderate LIPs erupting through thin lithosphere could be simulated, but not LIPs such as the Siberian Traps, which are large and erupted through thick lithosphere. The lithosphere beneath Hawaii is ca. 100 Ma old, and thermal upwelling models can only simulate magmatic rates  $\ll 0.1$  km $^3$ /yr for reasonable temperatures.

In the case of near-ridge melting anomalies, the question arises whether the excess melting can be modeled by isentropic upwelling of fusible eclogite at otherwise normal parts of the ridge system. Such a mechanism has been suggested to account for the large volumes of melt produced at Iceland, the Ontong Java plateau, and other oceanic plateaus in the Atlantic and Indian oceans (Yaxley and Green, 1998; Anderson, 2005a; Foulger and Anderson, 2005; Korenaga, 2005). At present the energy required to melt the relevant minerals at the appropriate temperatures and pressures is not known sufficiently well to be able to answer this question. It has also been suggested that subducted ocean crust or fusible, metasomatized lithospheric mantle may warm by conduction of heat from its surroundings in the

upper mantle, causing it to rise and melt in a runaway fashion (Anderson, this volume). Melt accumulating from long-term warming of fusible material in the mantle by conduction may also pond and erupt on a much shorter time scale than that of accumulation (Silver et al., 2006). These ideas are currently speculative, however. At present no fully developed, robust numerical model currently exists that can explain the melt volumes and magmatic production rates thought to occur at large LIPs erupted through thick lithosphere.

### Geochemistry

Significant geochemical differences are observed among many OIBs and MORBs (Hofmann and White, 1982). In particular the former include samples that are relatively enriched in incompatible elements, such as U and Th, and in light rare earth elements relative to heavy ones. Radiogenic isotope ratios (e.g.,  $^{87}\text{Sr}/^{86}\text{Sr}$ ) in MORB are indicative of past melting events that caused relative depletion in the parent elements compared to the source of OIB. OIBs sometimes have high ratios of primordial noble gas isotopes relative to radiogenic ones.

These characteristics of OIB represent end-members, however, and a range of compositions may be observed that overlap those of MORB (e.g., at Iceland). Furthermore, all the characteristics listed above may not be found in every sample. For example, in basalts from Baffin Island, the ratio of  $^3\text{He}/^4\text{He}$  correlates with depletion in incompatible elements (Stuart et al., 2003). The observed geochemical signatures cannot be explained by simple binary mixing of two sources.

A remarkable aspect of OIB geochemistry is that similar compositions are seen in a wide variety of settings, including oceanic melting anomalies and low-volume magmas erupted in continental rifts, where a deep thermal origin is implausible (Natland and Winterer, 2005; Fitton, this volume). Eruption of similar compositions may furthermore continue for tens or even hundreds of millions of years at the same locality. For example, in the Scottish Midland Valley, OIB-like volcanism persisted for ~60 Ma, during which time the lithosphere is thought to have drifted ~15° relative to the deep mantle (Fitton, this volume). A fixed, focused source for the magmas observed is thus unlikely, and a source that is either ubiquitous (e.g., Anderson, 1995) or carried with the plate seems to be required.

The most likely source of OIB signatures is recycled near-surface materials (Hofmann and White, 1982). Throughout its history, the Earth has become chemically stratified by partial melting that preferentially extracts elements incompatible in mantle phases and transports them upward. The mantle has been extensively processed in this way throughout geological time. The source of OIB must be either some region of the mantle containing material relatively little processed by melt extraction during Earth history or recycled near-surface material. There is little geochemical support for the former scenario, which is at odds with the expectation that the mantle went through an early, largely molten stage (e.g., Anderson, 1989).

Candidate near-surface material that may be recycled into the mantle includes oceanic crust, along with overlying pelagic and terrigenous sediments and metasomatized mantle lithosphere. Ocean crust is reintroduced into the mantle at subduction zones. Metasomatized arc or continental lithosphere and lower crust may become detached during orogenic events and recycled back into the convecting upper mantle (e.g., McKenzie and O’Nions, 1995).

Chauvel and Hemond (2000), Breddam (2002), and Foulger et al. (2005b) have advocated a source for Icelandic tholeiitic basalts in discrete lithologies from subducted oceanic crust, in particular of Iapetus origin. Green et al. (1967) long ago pointed out the difficulty of producing from this source the silica-undersaturated basalts observed at many ocean islands. More recently, Niu and O’Hara (2003) objected to oceanic crust as a source, citing the inability of remelted oceanic crust to produce high-MgO parent melts, the isotopic depletion of ancient oceanic crust, and the depletion of water-soluble incompatible elements that it is expected to have undergone during subduction. Niu and O’Hara (2003) suggest that deeper portions of recycled oceanic lithosphere are a more likely source for OIB.

Pilet et al. (2002, 2004) recently proposed an origin in variably metasomatized mantle lithosphere by partial melting of metasomatic veins plus the enclosing lithospheric mantle. This model could generate isotopic and trace element ratio variations similar to those observed both in oceanic islands and continental rift zones. It would be applicable to both continental and oceanic lithosphere and might thus explain the ubiquity of OIB.

Helium isotope ratios make up a special category of geochemical observations because, whereas it is generally acknowledged that other geochemical characteristics are not diagnostic of source depth, high  $^3\text{He}/^4\text{He}$  is widely assumed to indicate a deep lower-mantle origin. In some samples from melting anomalies (e.g., Iceland, Hawaii), values of  $^3\text{He}/^4\text{He}$  as high as ~50 Ra (the atmospheric value of  $1.38 \times 10^{-6}$ ) are measured. This ratio may be compared with values of 6–10 times the atmospheric value, which is the maximum typically found in samples from mid-ocean ridges. The high values have traditionally been interpreted as resulting from high levels of the primordial isotope  $^3\text{He}$ , considered to be stored in a near-primordial lower-mantle region (e.g., Kellogg and Wasserburg, 1990). An alternative interpretation is that they result from low levels of  $^4\text{He}$ , a radiogenic decay product of U and Th (Fig. 11). In this model, He is stored in a low-U+Th matrix (e.g., dunite cumulates), and ancient, high values of  $^3\text{He}/^4\text{He}$  are preserved for long periods with little change (Anderson, 1998a,b; Natland, 2003; Meibom et al., 2005). Dunite cumulates occur in the lower parts of oceanic crust, and this model thus suggests that high  $^3\text{He}/^4\text{He}$  ratios may be explained by recycled subducted oceanic crustal lithologies in the same way as other OIB characteristics. The Ne isotope ratios found in some OIBs are low in  $^{21}\text{Ne}$  compared with those measured in MORBs.  $^{21}\text{Ne}$  is produced by the nucleogenic decay of  $^{24}\text{Mg}$  and  $^{18}\text{O}$  when irradiated by U and Th decay products. It is thus produced at the same rate as  $^4\text{He}$ ,

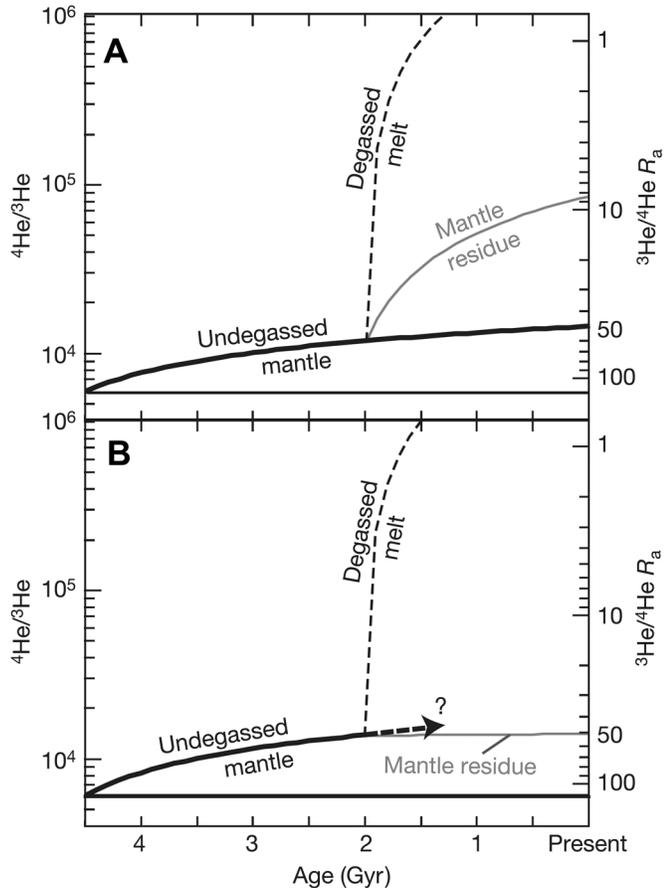


Figure 11. Scenarios for  $^3\text{He}/^4\text{He}$  isotope ratios for the lower-mantle (plume) and upper-mantle (plate) models (from Parman et al., 2005). Age is plotted versus He isotope ratio. Both (A) and (B) start with an initial  $^3\text{He}/^4\text{He}$  of 120 Ra (the atmospheric value of  $1.38 \times 10^6$ ) at 4.5 Ga. (A) the standard (plume) model attributes high  $^3\text{He}/^4\text{He}$  to high residual  $^3\text{He}$  from a little-degassed reservoir in the lower mantle. Undegassed mantle (thick black line), thought to comprise part or all of the lower mantle, evolves to a present-day value of  $\sim 50$  Ra and is assumed to be sampled by ocean island basalt (OIB). A melting event at 2 Ga increases the  $^{238}\text{U}/^3\text{He}$  of the melt relative to the undegassed mantle such that it evolves to a present-day value of 8 Ra (gray line). The standard model suggests that this is the mid-ocean ridge basalt (MORB) source. If the melt approaches the surface and degasses,  $^{238}\text{U}/^3\text{He}$  increases and crustal materials made from such melts will rapidly evolve very low  $^3\text{He}/^4\text{He}$  ratios (dashed line). (B) Alternative model in which He is more compatible than U + Th (line styles the same as in panel A). Here the highest  $^3\text{He}/^4\text{He}$  OIBs are melts of mantle residue. Melting at 2 Ga produces a mantle residue with a  $^{238}\text{U}/^3\text{He}$  one-tenth that of the undegassed mantle, which evolves to a present-day  $^3\text{He}/^4\text{He}$  of 50 (gray line). As in (A), the degassed melt rapidly evolves low  $^3\text{He}/^4\text{He}$ . Simply put, the plume model in (A) suggests that high  $^3\text{He}/^4\text{He}$  arises from a primordial source with high absolute concentrations of  $^3\text{He}$  whereas the model in (B) suggests that high  $^3\text{He}/^4\text{He}$  arises from a source depleted in U + Th by an earlier melting event and is thus low in  $^4\text{He}$ . Such a model was originally proposed by Anderson (1998a,b).

which suggests that the neon isotope ratios observed in OIB may be explained in the same way as the helium isotope ratios.

The implication for mantle dynamics of OIB geochemical signatures then revolves around the depths to which recycled near-surface materials are circulated, which involves considerations of density. Basalt transforms to the denser rock eclogite at depths of 50–60 km, which is expected to encourage sinking (O'Hara, 1975). The depth at which it reaches neutral buoyancy is disputed, however. Deep-source models cite tomographic evidence to propose that downgoing subducted slabs reach the core-mantle boundary and are recycled back to the surface from there in plumes (e.g., Kellogg and Wasserburg, 1990). The plate model proposes that neutral buoyancy for much recycled material is reached in the upper mantle, where it reheats by thermal conduction from the surrounding mantle (e.g., Anderson, 2006). The time scale of slab reheating is much less than the age of the plate on subduction, and because eclogite is more fusible than peridotite, its temperature may approach or even exceed its solidus in the upper mantle, providing potentially highly productive source material to supply melting anomalies.

## EXAMPLES

### *Atlantic Ocean*

The primary melting anomalies in the Atlantic Ocean include Iceland, the Azores, Bermuda, the Canary Islands, the Cape Verde Islands, the Ascension-Cameroon system, and Tristan (Figs. 1 and 12). All are very different in character. Plate models have been suggested for the Iceland and Tristan systems. The Bermuda melting anomaly is discussed in detail by Vogt and Jung (this volume). Plate models for other Atlantic melting anomalies have yet to be developed.

The melting anomaly associated with Iceland is by far the most voluminous in the Atlantic Ocean and the only one to produce large quantities of tholeiite. The magmatic rate has varied irregularly over time. High-volume volcanism along a  $\sim 2500$ -km zone of rifting accompanied the breakup of Laurasia and the opening of the north Atlantic Ocean at ca. 54 Ma. This volcanic activity subsequently dwindled as normal-thickness ocean crust began to form, but the magmatic rate continued at a high level along a  $\sim 100$ -km-long portion of the Mid-Atlantic Ridge, building the Iceland-Faeroe Ridge (Fig. 7). Starting at ca. 44 Ma, the part of the ridge that produced excess volcanism greatly lengthened to attain a present-day north-south extent of  $\sim 600$  km. If the whole of the  $\sim 30$ -km-thick seismic crust is melt (an assumption that is questioned; Björnsson et al., 2005; Foulger, 2005) the present-day excess melt production rate is  $\sim 0.25$  km<sup>3</sup>/yr. However, apart from precursory volcanism at 61–59 Ma, magmatism has always been centered on the Mid-Atlantic Ridge, and there is little evidence for a time-progressive volcanic track (Foulger, 2003, 2006b; Foulger and Anderson, 2005; Lundin and Doré, 2005). The area of the Icelandic platform is  $\sim 0.3 \times 10^6$  km<sup>2</sup>, so

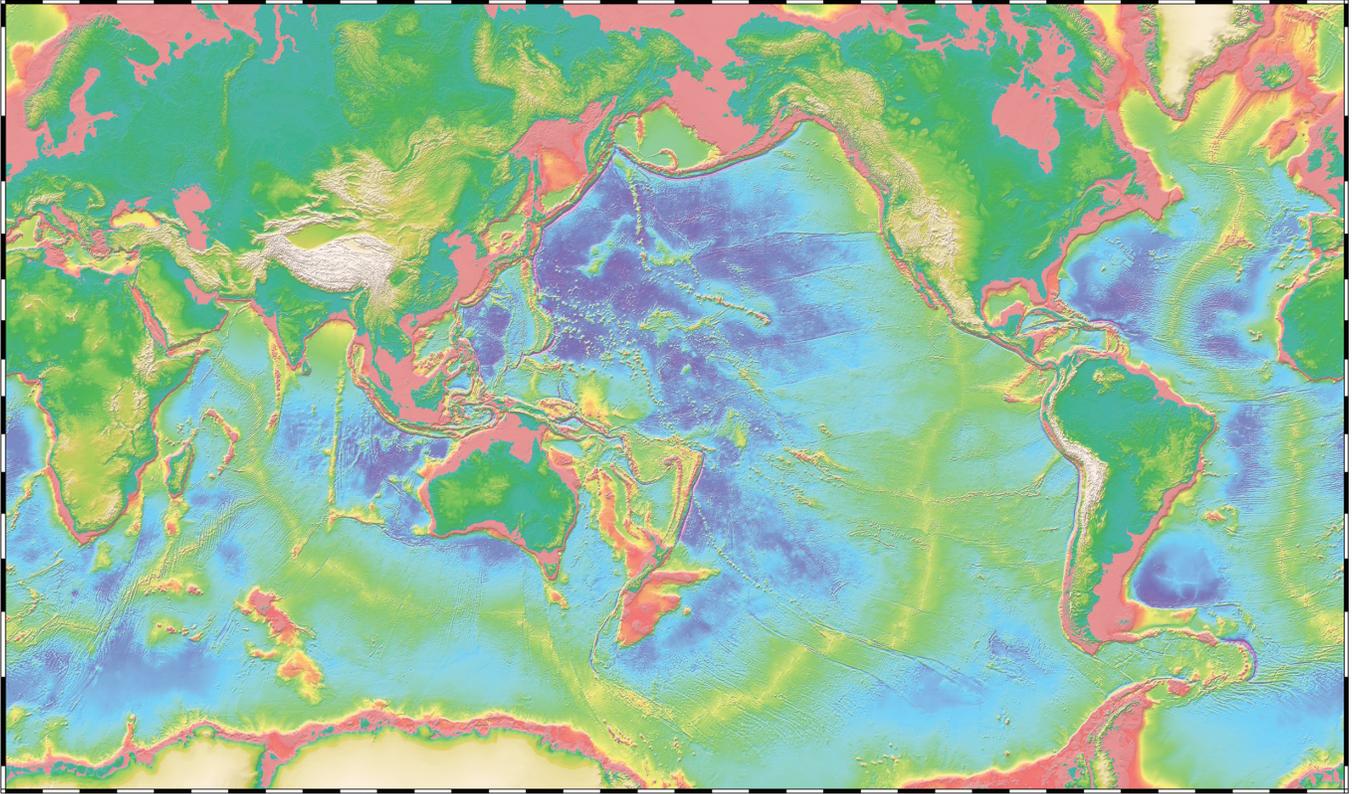


Figure 12. Topography of the Earth based on a global compilation of land data (GTOPO30, <http://edcdaac.usgs.gov/topo30/README.html>) and ocean data from a combination of sparse ship soundings and marine gravity anomalies derived from satellite altimetry (Smith and Sandwell, 1997).

it qualifies as a LIP (Coffin and Eldholm, 1994; Sheth, 2006b, 2007a).

Temperature estimates for the mantle beneath Iceland have been obtained using seismology, petrology, bathymetry, vertical motion, and heatflow (see Foulger et al., 2005a, for a review). All either require or are consistent with temperature anomalies no greater than 50–100 °C compared with the mean for mid-ocean ridges (Foulger et al., 2005a). Seismic experiments show that the whole north Atlantic is underlain by a low-velocity zone that extends from the surface to the base of the upper mantle (Ritsema et al., 1999). Its strength is consistent with either a temperature anomaly of ~100 °C throughout most of its volume or a fraction of a percent of partial melt, which might result from the presence of a small amount of carbonate (Presnall and Gudfinnsson, 2005). Detailed tomography and receiver function studies agree that the anomaly terminates in the transition zone (Ritsema et al., 1999; Foulger et al., 2001; Montelli et al., 2004; Du et al., 2006).

These results, coupled with observed tectonic correlations, have inspired development of a plate-based model for Iceland. The Greenland-Iceland-Faeroe Ridge formed colinearly with a northwest-trending zone of precursory magmatism that erupted

at 61–59 Ma. This trend is also colinear with the trend of the western frontal thrust of the Caledonian suture where it runs offshore in east Greenland and re-emerges on land in Britain. At this latitude, the Mid-Atlantic Ridge, newly formed at ca. 54 Ma, crossed the outer limit of the Caledonian suture. Here, the trend of the Mid-Atlantic Ridge also changes radically, from ~N35°E to N20°E. Tectonic style has been persistently locally complex in the Iceland region, involving migrating ridges, parallel-pair spreading, and both oceanic and continental microplate formation since the earliest opening of the ocean, when a transform fault existed at this latitude (Nunns, 1983; Foulger, 2006b).

The plate model relates excess magmatism in the Iceland region to processes related to extension (Foulger, 2002; Foulger and Anderson, 2005; Foulger et al., 2005a,b). The volcanism that accompanied the early opening of the north Atlantic was clearly associated with continental breakup (van Wijk et al., 2004). The Iceland melting anomaly comprises a portion of the spreading plate boundary where the mantle beneath is unusually fertile as a result of entrained eclogite from slabs trapped in the Caledonian suture when it closed at ca. 400 Ma. This entrained eclogite is now dispersed locally in the mantle below where the suture formerly lay. Although north Atlantic mantle in general

may be somewhat warmer than the global average, the excess magmatism is attributed essentially to isentropic upwelling of mantle with enhanced fertility and not to high temperature. This model naturally explains the co-location of the region with the Caledonian western frontal thrust, the persistence of the melting anomaly at the Mid-Atlantic Ridge, and the evidence for normal temperature or only small temperature anomalies. In this model, the mantle seismic anomaly represents a thickening of the ubiquitous global low-velocity zone, which is thought to be caused by small degrees of partial melt induced by volatiles, in particular CO<sub>2</sub> (Presnall and Gudfinnsson, 2005; Dasgupta and Hirschmann, 2006). It may also be related to a high eclogite content (Foulger and Anderson, 2005; Foulger et al., 2005b). Enhanced volatile content is observed in erupted lavas and is expected in a mantle made unusually fertile by recycled near-surface material. The unstable tectonics in the Iceland region is clearly a continuation of behavior that dates from the earliest opening of the ocean, when the newly formed ridge crossed a major structural divide (Skogseid et al., 2000).

The Azores melting anomaly lies at a ridge-ridge-ridge triple junction, where the Azores ultra-slow spreading ridge meets the Mid-Atlantic Ridge. The Azores plateau, which is underlain by crust ~10–14 km thick, has been emplaced since ca. 20 Ma by magmatism along all three ridge branches. The Azores branch is deduced from kinematic models, earthquake focal mechanisms, and bathymetry to be an oblique ridge, spreading at a rate of 3–4 mm/yr (Lourenco et al., 1998). As is also the case with Iceland, its geochemical footprint along the Mid-Atlantic Ridge is asymmetric, extending to 1000 km south of the Azores but only to 250 km to the north. Moreira et al. (1999) suggested that a recycled oceanic crustal component originating in subcontinental lithosphere delaminated during opening of the north Atlantic can explain the Pb isotope systematics observed there, where <sup>203</sup>Pb/<sup>204</sup>Pb is up to 11.3. The discovery of zircons dated at 330 Ma and 1.6 Ga in the Kane fracture zone ~1500 km south of the Azores lends support to this hypothesis (Pilot et al., 1998). A promising plate model for the Azores melting anomaly would then involve enhanced magmatism resulting from source fertility, peaking at the triple junction as a result of locally enhanced extension and focused flow (Georgen and Lin, 2002).

Magmatism in the Ascension-Cameroon region is complex. The Cameroon line has erupted along much of its length between ca. 130 Ma and the present, and no regular time progression is observed (Fitton and Dunlop, 1985; Fitton, 1987; Ernst and Buchan, 2002). It has been suggested that it is related to reactivation of the Central African Shear Zone (Moreau et al., 1987). Volcanism continues to the southwest as the St. Helena seamount chain, for which time-progressive ages are reported (O'Connor et al., 1999). Extrapolation of this age progression would suggest a present-day location for the locus of melt extraction ~500 km west of St. Helena. No current volcanism is observed there, however. Young volcanism in the region is, instead, scattered over many seamounts distributed throughout a

region some hundreds of kilometers broad, including Ascension Island, St. Helena, and the Circe seamount (O'Connor et al., 1999). It has been suggested that a nearby geochemical anomaly observed on the Mid-Atlantic Ridge results from lateral flow from this melting anomaly (Schilling et al., 1985b). However, gravity data suggest a crustal thickness anomaly of only ~4 km in the Ascension region, with gravity variations resulting from crustal thickness variations and not variations in mantle temperature (Bruguier et al., 2003). In view of this suite of observations, a likely plate model would involve diffuse permissive volcanism as a result of variable extensional stress in the African plate. The mantle beneath the Atlantic is variable in fertility (the “blob” model; e.g., Bruguier et al., 2003; Vogt and Jung, 2005). Variations in the African plate stress field may be linked to tectonic evolution of the northern collision boundary with Europe, and old sutures and fault zones may be preferentially activated (Fairhead and Wilson, 2005).

A similar plate model has been proposed for the Tristan melting anomaly (Fairhead and Wilson, 2005). Like the Ascension-Cameroon province, this region also comprises a southwest-orientated zone of seamount volcanism—the Walvis Ridge. There are few reliable radiometric dates (Baksi, 1999), but the youngest volcanism is at the southwestern end, some hundreds of kilometers east of the Mid-Atlantic Ridge. Like the Ascension-Cameroon province, it, too, is scattered over a region several hundred kilometers broad. In contrast with that province, however, the Walvis Ridge is mirrored by similar volcanism west of the Mid-Atlantic Ridge, the conjugate Rio Grande Rise on the South American plate. These two ridges exhibit entirely different morphologies, however.

Fairhead and Wilson (2005) combined satellite gravity and plate motion data from GPS (Global Positioning System) with extensive information constraining the tectonic history of Africa and South America. They concluded that changes in stratigraphy in African rift basins reflected variations in the state of stress of the African plate resulting from changes in distant plate motions (e.g., India-Eurasia and Africa-Europe collisions). They proposed a plate model whereby periodic stress release permitted episodic excess volcanism near the Mid-Atlantic Ridge and on flanking aseismic ridges. These include the Walvis Ridge and the Rio Grande Rise, and volcanism occurs along shear, wrench, and extensional deformation structures. Additional observations in support of this model are reported from seismic reflection studies of the Cabo Frio area, part of the Rio Grande Rise immediately offshore Brazil (Oreiro et al., 2006).

The Atlantic Ocean contains a number of other, smaller-scale melting anomalies, some of which comprise short archipelagos or zones of seamounts. These include the Canary Islands, the Cape Verde Islands, the Discovery seamount chain, and various seamounts in the Azores region. There is a paucity of reliable dates, but volcanism may be locally short- or long-lived. Such volcanism fits best a model of varying stress and source fusibility and fits poorly a model involving large thermal diapirs.

## Pacific Ocean

The most remarkable currently active melting anomaly in the Pacific Ocean is the Hawaiian archipelago at the southeast end of the Emperor and Hawaiian volcanic chains (Fig. 12). This system is unique on Earth. No other currently active volcanic system exhibits the same combination of longevity, extreme regularity of time progression, remoteness from plate boundaries and continental edges, and a high present-day production rate of tholeiitic basalt. The most widely suggested nonthermal model for the Emperor and Hawaiian volcanic chains is based on a propagating crack. This model was originally suggested by Dana (1849), who observed that the islands aged to the northwest and suggested that they formed over a major fissure zone caused by cooling and shrinkage of the outer layers of the Earth. This inspired insight forms the basis for some of the most recent efforts to explain linear volcanic chains in the Pacific (e.g., Sandwell and Fialko, 2004).

Models to explain Emperor and Hawaiian volcanism must account for a number of observations, including:

1. The Emperor chain started on or near a spreading ridge, and there is no evidence that it was preceded by LIP eruption (Shapiro et al., 2006; Norton, this volume);
2. Both chains are regularly time progressive (Sharp and Clague, 2006);
3. With respect to the geomagnetic pole, the locus of melt extraction propagated south by ~800 km during emplacement of the Emperor chain, but was stationary during emplacement of the Hawaiian chain (Tarduno and Cottrell, 1997);
4. The direction of propagation changed by ~60° at ca. 47 Ma, at the time of the “bend” (Sharp and Clague, 2006). No corresponding change in the direction of Pacific plate motion occurred at this time (Norton, 1995);
5. The melt extraction rate has increased by an order of magnitude during the past ~2 m.y. (Robinson and Eakins, 2006), resulting in LIP emplacement following the development of the time-progressive chains rather than preceding it (Fig. 6). This change in magmatic rate has apparently been accompanied by a very recent change in propagation direction from approximately southeast to approximately SSE;
6. The younger half of the Hawaiian chain is surrounded by a bathymetric swell (Fig. 12).

A great deal of research on the Hawaiian Islands has been directed at probing the underlying mantle structure. However, work is hampered by the small size of the “Big” Island and the large percentage of the Hawaiian archipelago that is under water. These factors limit the aperture of land seismic networks and geochemical sampling.

The temperature of the mantle beneath Hawaii has been investigated using sea-bottom heatflow measurements and petrology. Heatflow measurements across the Hawaiian swell do not

detect a heatflow anomaly of the kind expected for a thermal source (von Herzen et al., 1989), although recently it has been suggested that hydrothermal circulations mask expected anomalies (Harris and McNutt, 2005). Estimates from picrite glass samples of the mantle temperature anomaly vary from zero to ~250 °C (Green and Falloon, 2005; Putirka, 2005; Falloon et al., this volume).

The geochemistry of the lavas is variable, exhibiting both enriched and depleted signatures consistent with components of continental and marine sediments and oceanic crust, and having variable isotopic ratios. The geochemistry varies geographically where the Hawaiian and Emperor chains cross fracture zones (e.g., the Mendocino, Murray, and Molokai fracture zones; Basu and Faggart, 1996). Variations with time area also reported; for example, Mukhopadhyay et al. (2003) describe changes in  $^3\text{He}/^4\text{He}$  of up to 8 Ra during a single century in Kauai volcano. Observations suggest that different volcanoes are fed by different magma sources, suggesting a chemically heterogeneous source. This possibility is in keeping with recent work that suggests the present-day source contains up to 30% of recycled crust (e.g., Sobolev et al., 2005). The volume of magma erupted at Hawaii cannot currently be explained by any model, plume or plate, without appealing to a source much more fertile than lherzolite (Cordery et al., 1997; Leitch et al., 1997). The maximum  $^3\text{He}/^4\text{He}$  observed at Hawaii is 35 Ra in samples from Loihi (Graham, 2002). Calculations based on a lherzolite source suggest that the lavas were last in equilibrium at ~90–100 km depth, corresponding to the estimated base of the Cretaceous lithosphere on which Hawaii is emplaced.

Teleseismic tomography conducted using a network on the Big Island found little significant structure in the top ~100 km (Ellsworth, 1977). Larger-scale teleseismic tomography involving a ~600-km-long array of sensors on several islands found low-velocity anomalies beneath the islands of Maui and Molokai, 250 km northwest of the Big Island, but no low-wavespeed anomaly beneath the Big Island itself down to the maximum depth of good resolution there at ~150 km (Wolfe et al., 2002). Ray coverage throughout the Pacific Ocean is patchy, and whole-mantle tomography can only image very large-scale features (Julian, 2005). It reveals continuous low-velocity material between the surface in the Hawaiian region and the core-mantle boundary beneath a large swath of the Earth ranging from the New Hebrides and Samoa, throughout the south Pacific, and north along the East Pacific Rise, depending on the line of cross-section selected (Ritsema, 2005). This global-scale anomaly is associated with the “Pacific superplume,” which has recently been shown to be largely compositional in origin (Trampert et al., 2004). Finite-frequency tomography reveals low velocities beneath the Hawaiian region that die out in the mid-mantle (Montelli et al., 2004). Transition zone thickness beneath the Hawaiian region is ~229 km (Gu and Dziewonski, 2001), which is ~13 km thinner than the global average of 242 km but typical for the central Pacific as a whole.

Only a little work has been done to develop a numerical plate model for Hawaii involving a propagating crack (Shaw, 1973; Shaw and Jackson, 1973; Turcotte and Oxburgh, 1973; Turcotte, 1974; Jackson and Shaw, 1975; Jackson et al., 1975; Lithgow-Bertelloni and Guynn, 2004; Natland and Winterer, 2005). Stuart et al. (this volume) calculated the stress field that would result from cooling of the Pacific plate and found that the orientation of stress in the region of Hawaii was optimal for the southeastward propagation of a crack tip. It seems likely that the major changes in magmatic rate, from near zero at the time of the bend to the present-day maximum, must reflect variations in source composition rather than variations solely in temperature. A stress-based model could also naturally explain the time-progressive chains; variable propagation rates; and the sharpness of the bend, which argues against a cause in slowly varying thermal convection structures.

Favela and Anderson (1999) suggested that Hawaiian volcanism initiated as a result of a change in stress in the plate. Smith (this volume) suggests that the cause of this stress change was a nearly simultaneous change in the western part of the Pacific plate from being bounded by the Kula-Pacific and North New Guinea-Pacific ridges to being subducted beneath the Aleutian and Izu-Bonin-Mariana arcs. The volume of the bathymetric swell is proportional to the volume of magma emplaced on the surface and may be explained by buoyant residuum left over from melt extraction from the mantle (Phipps Morgan et al., 1995). Notwithstanding these speculations, a crack model for Hawaiian volcanism urgently needs to be tested, ideally by measuring extension normal to the chain. However, detecting extension in the deep ocean that might be on the order of 1 mm/yr and total no more than 1 km beneath a lava pile ~25 km thick is currently technologically challenging.

Aside from the Emperor-Hawaiian system, there is great diversity in the nature of melting anomalies in the Pacific. Low-volume seamounts of alkalic OIB are widely scattered over the ocean floor, and most do not plausibly form discrete chains. However, linear arrays of seamounts are common, but chronological progressions range from being regular (e.g., the Louisville chain; Watts et al., 1988) to highly variable (e.g., the Austral chain; McNutt et al., 1997). Many chains are of little significance volumetrically, are associated with essentially none of the characteristics expected of plumes, and may readily be explained by thermal contraction and cracking of the lithosphere (Sandwell et al., 1995; Sandwell and Fialko, 2004). Volumetrically larger melting anomalies do exist, for example, Samoa and the Galapagos, but neither of these is associated with time progression of volcanism. Dates from the Galapagos suggest a broad, heterogeneous melting anomaly. Volcanism in the Samoa region is located at the northern end of the Tonga Trench, where major extension is predicted by models of Pacific lithosphere cooling (Stuart et al., this volume).

### Continents

Magmatism on the continents resembles that of the oceans in that a broad spectrum of volume, period of emplacement, time

progressiveness, and relationship with tectonic events and features are exhibited. The volumes of magma observed range from very small to LIP-sized, with volumes up to  $\sim 4 \times 10^6 \text{ km}^3$  in the case of the Siberian Traps (Fig. 2). Much of the Siberian Traps is thought to have erupted within a few million years (Kamo et al., 2003; Ivanov, this volume). However, this intensity may be compared with volcanism in the Basin and Range province, western United States, where widespread, relatively low-volume eruptions have continued since ca. 20 Ma to the present day and are sourced beneath a region ~1000 km broad.

Basaltic continental magmas typically have OIB-like compositions and range from alkalic to tholeiitic types. Low-volume alkalic OIB-like basalts are widespread in many rift settings, where they are presumed to be associated with the rifting process (e.g., Hooper et al., this volume). High-volume basaltic LIPs generally also have OIB-like compositions, although the Siberian Traps is an exception (Fitton, this volume). It has recently been recognized that silicic provinces may be sufficiently large to be classified as LIPs, and their duration of emplacement may be up to 40 Ma (Bryan et al., 2002). LIPs may be dominantly basaltic or silicic, though at least one province with subequal proportions of each is known (Sensarma et al., 2004). Large-volume silicic melts must come from remelting the continental crust. These observations need to be accounted for in models proposed to explain the origin of continental volcanism and the time scales and mechanisms of heat supply.

The history of vertical motion associated with the largest LIPs also varies widely, and includes cases of subsidence (e.g., the Siberian Traps, the Columbia River basalts; Czamanske, 1998; Hales et al., 2005), lack of uplift (e.g., the Deccan Traps; Sheth, this volume), and uplift (e.g., the Emeishan basalts; Rainbird and Ernst, 2001; He et al., 2003; Xu and He, this volume). The chronological relationship between vertical motion and eruption is also variable. For example, uplift may precede volcanism (e.g., the Emeishan basalts), accompany it (e.g., the North Atlantic Igneous province; Maclennan and Jones, 2006) or postdate it (e.g., the Whitsunday province; Bryan et al., 2002).

Major time-progressive trails of volcanism are rare, the best example being the Yellowstone–Snake River Plain chain of silicic calderas that formed from ca. 17 Ma to the present (Christiansen et al., 2002). This chain is aligned with the plate direction but is paired with the High Lava Plains time-progressive volcanic track, which youngs in the opposite direction (Jordan, 2005).

Volcanism is often associated with significant tectonic structures or events. Many continental LIPs are coastal, erupted when continents broke up (e.g., the North Atlantic Igneous Province, the Central Atlantic Magmatic province, the Deccan Traps), and are clearly associated with the rifting process. Van Wijk et al. (2001) modeled decompression melting and found that the volumes observed at many volcanic margins can be accounted for simply by isentropic upwelling of asthenosphere in response to continental rifting. Volcanism commonly occurs in continental rifts that have not developed to the point of continental breakup (e.g., the east African rift) and may be particu-

larly intense at triple junctions (e.g., Afar). Well-documented pre-emplacement subsidence has led to the suggestion that some major continental LIPs erupt in response to the development of gravitational instability and detachment of the mantle lithosphere and possibly part of the lower crust (Tanton and Hager, 2000; Hales et al., 2005; Lustrino, 2005). Back-arc and slab deformation and fragmentation processes have been suggested to have triggered volcanism in such regions as Mexico, Turkey, the Basin and Range province, and east Asia (e.g., Ferrari, 2004; Keskin, this volume).

## DISCUSSION

The primary objective of this chapter is to present the rationale behind the plate model for the genesis of melting anomalies. Although some observations are consistent with the classical plume model originally defined by Morgan (1971) and recently reiterated by Campbell (2006), many features of volcanic regions are not naturally expected for a classical plume source, and the predictions of that model are often not fulfilled. These failures have, in recent years, inspired the quest for a theory that may explain the observations more wholly and have greater powers of prediction.

The term “plume” is used flexibly to describe a wide variety of phenomena, some of which are remote from the original model of Morgan (1971). Some, ironically, fall within the plate model. These include thermal uprisings in the mantle resulting from continental insulation (Burov and Guillou-Frotier, 2005b), conductive and radiogenic warming of eclogite (Davies and Bunge, 2006), and “lithospheric plumes” (Courtillot et al., 2003). Such broad and undefined usage of the term can clearly, in itself, be a barrier to progress.

To be viable, a model must be consistent with the holistic observations, including field geology, tectonics, igneous petrology, geochronology, geochemistry, and geophysics. This requirement presents a formidable challenge to researchers, given the extraordinary degree of specialization that now exists in all branches of modern Earth science. Integrated science is, however, imperative for progress in this highly cross-disciplinary field.

Although melting anomalies or “hotspots” have traditionally been considered a single phenomenological class, a general review reveals extreme variation in almost every respect. This variability includes the melt volume, pattern of vertical motion, chronology of eruption, petrology, underlying mantle structure, and suite of characteristics at a single locality. Under such circumstances, no single, simple generic model is possible for every “hotspot” that does not require much ad hoc special pleading. Melting anomalies are united essentially only by the commonality of unusual melt extraction. Even in this respect, there is a continuous spectrum in the volume produced—from extraordinarily large LIPs to trivial eruptions—and in composition. A natural bimodalism might be expected if fundamentally different genesis processes are at work, but no such bimodalism has been identified to date. A unifying model for anomalous vol-

canism is required that can account for the full spectrum of size, longevity, composition, and spatial and temporal distributions of volcanism.

Two LIPs are currently in the process of formation, Iceland (with a present-day area and volume of  $\sim 0.3 \times 10^6 \text{ km}^2$  and  $\sim 5 \times 10^6 \text{ km}^3$ , respectively) and Hawaii (with a present-day area and volume of  $\sim 0.15 \times 10^6 \text{ km}^2$  and  $\sim 0.6 \times 10^6 \text{ km}^3$ , respectively). Both are forming late in the sequence of associated volcanism. In the context of the plume model, these systems amount to plume-head volcanism occurring after plume-tail volcanism and would be interpreted as indicating a pulse in the temperature or flow of the plume tail. In the context of plate interpretations, an explanation would be expected in variations of extension rate or the fertility of the source material being tapped. Melting anomalies that exhibit LIP volcanism followed by low-volume, time-progressive volcanic chain formation, as predicted by the plume model, are at best very rare, and perhaps nonexistent.

There is no unequivocal evidence for systematically elevated temperature at LIPs or “hotspots” in general. Petrological approaches using olivine control lines are still contested. The traditional assumption that high-forsterite olivines and picrite cumulates require high temperature is now recognized to be incorrect (Green et al., 2001; Green and Falloon, 2005; Presnall and Gudfinnsson, 2005; Falloon et al., 2007a, this volume). The eruption of large volumes of tholeiitic magma at LIPs requires the removal of a large quantity of heat from the mantle, but direct evidence that this removal is engineered by temperatures that are hundreds of degrees above those beneath normal mid-ocean ridges remains elusive. However, the OIB geochemical signature of “hotspot” lavas is *prima facie* evidence for fusible material in the source, and the percentage in recent Hawaiian and Icelandic lavas may be large (e.g., Foulger et al., 2005b; Sobolev et al., 2005).

OIB geochemistry is usually but not always associated with melting anomalies. It is observed in eruptive sequences with a wide range of tectonic settings, distributions, volumes, and longevities. OIB-like eruptive sequences may appear to be stationary in some frame and to erupt independently of plate motion, giving rise to time-progressive sequences. At the other extreme, they may appear to travel with the plate. The generally agreed-upon association of OIB geochemistry with recycled, near-surface materials naturally suggests shallow-based models, though the precise sources remain unclear. The depth of origin of the noble gas signatures associated with OIB remains controversial, with recent work tending to support shallow sources (Stuart et al., 2003; Parman et al., 2005; Fig. 11).

Usually, the differences between the volcanism in the fast-spreading Pacific Ocean and the slow-spreading Atlantic Ocean are most strongly emphasized. However, there is considerable commonality. Each ocean contains a single major tholeiitic province (Hawaii and Iceland) and several melting anomalies associated with the ridge system. These include the Cobb, Galapagos, Easter, and Louisville anomalies in the Pacific and Iceland, the Azores, Ascension, Tristan, and the Bouvet triple

junction in the Atlantic. In both oceans, there are long, linear seamount chains, both time-progressive and nonprogressive, along with many shorter, low-volume chains and regions, where volcanism is scattered over broad areas.

The greatest challenge remains to explain quantitatively the largest melt volumes and volume rates that the Earth is capable of producing at the surface. Small volumes of alkalic lava, assumed to be formed by small-percentage melting, are a common observation in rifts and near extensional faulting and are assumed to be related to the lithosphere extension process. However, very large eruptive rates of tholeiitic basalt, which is thought to require a high degree of melting of a peridotite source or almost complete melting of an eclogite source, have yet to be fully explained. Quantitative modeling of decompression melting in plumes, even if containing fusible material, cannot reproduce, at reasonable temperatures, the volumes of the largest LIPs, where the lithosphere has significant thickness (Cordery et al., 1997). Decompression accompanying lithospheric rifting can explain the volcanic margins associated with continental breakup (van Wijk et al., 2001) and LIPs formed at mid-ocean ridges (Korenaga, 2005). The large temperature anomalies that have been suggested, but so far defy observation, are not required.

The problem for both thermal and athermal models then lies in explaining significant eruptive volumes at locations where the lithosphere is thick and large-scale extension is not observed. Where surface extension is minor, or where upwellings stall at the base of thick lithosphere, it is not clear how isentropic decompression can provide the energy required to melt large volumes of source material. The eruption of pre-existing, ponded melt has been suggested (e.g., Silver et al., 2006), but work still needs to be done to explain how extremely large volumes of melt can accumulate and be retained in the mantle for long periods. There are, however, few regions of this sort. Hawaii is the only presently active example, and has only become so in the past 2 Ma. Older examples include LIPs that erupted in the interiors of continents, far from rifted margins, such as the Siberian Traps (Ivanov, this volume).

The plate model for the source of melting anomalies is at an embryonic stage of development, and much work remains to be done before its full potential can be assessed. Urgent avenues of investigation include:

1. Mapping and modeling variations in stress in the lithosphere, both locally and on regional and plate-wide scales;
2. Numerical modeling of the melt volumes produced in different extensional settings;
3. Quantifying the volumes of melt produced by isentropic upwelling of nonperidotitic source material;
4. Modeling convection;
5. Solving the paradox of the origin of OIB;
6. Studying in detail the history of vertical motion associated with major LIPs and comparing with the predictions of plate models;

7. Understanding the mechanism that produces very large melt volumes and eruption rates;
8. Determining the continuity of seismic structure through the transition zone beneath active melting anomalies;
9. Understanding how to interpret seismic anomalies;
10. Resolving the present controversy regarding the source temperature of “hotspot” lavas;
11. Investigating the relationship between extensional faulting and related volcanism; and
12. Erecting and testing plate models for melting anomalies that have not yet been subject to this type of scrutiny.

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## DISCUSSION

### 26 December 2006, Rex H. Pilger

Foulger (this volume) writes that the Easter melting anomaly is located close to a ridge triple junction and is associated with a microplate. Pilger and Handschumacher (1981) showed that the

Easter “hotspot” could have produced both the Tuamotu and Nazca ridges as well as the Sala y Gomez trace, especially if the Hawaiian-Emperor bend were older than 43 Ma, significantly predating the triple junction and the Easter plate. The oblique trends of the Tuamotu and Nazca ridges to magnetic chrons

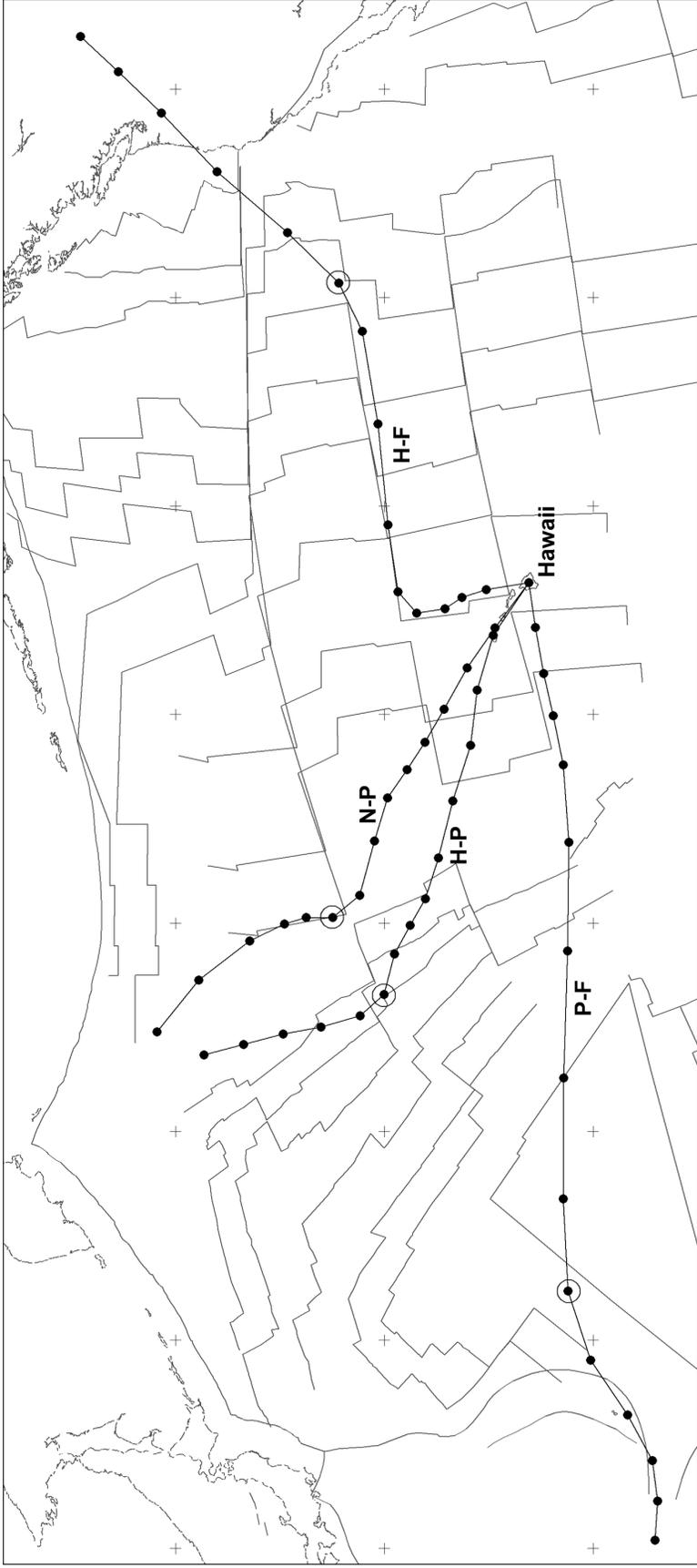


Figure D-1. Calculated loci at 5-m.y. intervals, from Hawaii: H-F—Hawaii to Farallon; H-P—Hawaiian “hotspot” relative to Pacific plate; N-P—North American to Pacific plate; P-F—Pacific to Farallon plate. Circled nodes are at 45 Ma. Gray crosses are latitude-longitude graticules at fifteen-degree intervals; the cross immediately southeast of Hawaii is at 15°S, 150°W. Gray line segments are magnetic isochrons from Müller et al. (1977).

implies that the first stress mechanism proposed by Beutel and Anderson (this volume) is not applicable, but the second, melting anomaly, mechanism (also part of the plate model of Foulger) could still apply.

Foulger also states, “The direction of propagation [of the Hawaiian-Emperor chain] changed by  $\sim 60^\circ$  at ca. 47 Ma, at the time of the ‘bend.’ . . . No corresponding change in the direction of Pacific plate motion occurred at this time.” There is indeed no evidence of a change in Pacific plate motion relative to adjacent plates at 43 Ma. However, there are ca. 47-Ma kinks in motion loci of all adjacent continental plates relative to Pacific plate (Norton, 2000, Fig. 10). North and South American–Pacific loci show comparable deflections to the Hawaiian-Emperor bend.

Loci of Pacific and Farallon (-Juan de Fuca) plate motion and North American and Hawaiian “hotspot” motion are shown in Figure D-1, interpolated at 5-m.y. intervals, using parameters from Pilger (2007). The circled nodes indicate the 45-Ma reconstructions and bends (finer interpolation would show bends from 45 to 48 Ma). A possible explanation for these correspondences is presented in Pilger (2007).

The oft-stated lack of correspondence of the bend with plate motions is apparently erroneous, but this does not invalidate the plate hypothesis for the origin of anomalous volcanic traces.

**8 February 2007, Gillian R. Foulger and William W. Sager**

In his comment of 26 December, Pilger makes two points. First, he writes that a fixed Easter “hotspot” could have produced the Tuamotu, Nazca, and Sala y Gomez ridges. In a region containing widespread volcanism it is to be expected that some can be found that conform with fixed hotspot models. However, the Easter and Sala y Gomez chains emanate from the southern edge of the Easter microplate, which is unlikely to be coincidence.

Pilger’s second comment is to take issue with a statement in Foulger (this volume) that there was no change in plate motion corresponding to the Hawaiian-Emperor bend. Models of plate motion based on the Hawaiian-Emperor and Louisville chains only, and assuming the fixed hotspot hypothesis predict a large change. However, other reconstructions do not, e.g., that of Raymond et al. (2000), which is based on hotspot propagation models for the Indian and Atlantic plates, reconstructing them to the Pacific plate using relative spreading models from magnetic lineations.

Pilger argues for a significant shift in plate motion based on kinks in his plate motion trends for the South and North American plates relative to the Pacific. However, the contemporaneous changes in his Figure D-1 for plates adjacent to the Pacific are small or nonexistent. Furthermore, the paths of the North and South American plates relative to the Pacific plate are highly uncertain, as a result of errors in the long chains of relative plate motions that must be used to make such reconstructions and uncertainty concerning the long-term rigidity of Antarctica as a

single plate. These uncertainties are large and are not shown in Figure D-1.

The fact that the bend in the Hawaiian-Emperor chain is accompanied by very small changes, if any, in the major fracture zones of the Pacific is a first-order observation that cannot be readily dismissed. The  $60^\circ$  change in trend in the Hawaiian-Emperor chain, now known to have occurred at ca. 50 Ma (Sharp and Clague, 2006), can correspond to only a small change in plate motion, if any. Atwater (1989) attributes a small change in plate motion at about this time to the breakup of the Farallon plate into north (Vancouver plate) and south parts. The small change in direction she postulates is accompanied by only very small changes in the fracture zones. If the plate changed direction, new plate rotation poles would be set up, and large changes in the transform faults would be expected. The orientations of fracture zones are, after all, *prima facie* testimony of the direction of plate motion, in contrast to other data that usually require careful processing to extract information on plate motion direction.

The time of the bend correlates with a radical change of behavior of the melting anomaly, when it ceased rapid southerly migration with respect to the geomagnetic pole (Tarduno et al., 2003; Sager, this volume). The evidence is consistent with most of the north-south motion implied by the Emperor seamounts being due to migration of the locus of melt extraction, and not by plate motion, whereas the reverse is the case for the Hawaiian chain. In the deep-rooted, fixed hotspot hypothesis, the occurrence of this change in behavior simultaneously with a major plate motion change would have to be coincidence. We find explanation of these observations as coincidence unlikely. The conclusion is inevitable: the bend in the Hawaiian-Emperor volcano chain is probably largely due to a change in migration behavior of the melt extraction anomaly, and not to a change in plate motion direction.

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