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 (“hot spots”) 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 “hot spots” 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 “hot spots” and “hot spot tracks” exhibit extreme variability. This 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 precursory-, contemporaneous- or post-emplacement uplift or subsidence. The total volume erupted ranges from trivial in the case of minor seamount chains to ~ 10^8 km^3 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 underway and are stimulating an unprecedented and healthy degree of critical debate regarding the results. Determining seismologically the physical conditions beneath melting anomalies are 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 at large igneous provinces requires knowledge of the volume and time-scale of emplacement, which are often unclear. Although ocean-island-basalt-type geochemistry is generally agreed to be derived from recycled near-surface materials, the specifics are not yet agreed. 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-progressive 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 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 at an ongoing stage of development, and poses a rich array of crucial but challenging questions that need to be addressed.

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 the current paper 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 paper 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 in places.

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, 2006a; Foulger et al., 2005c).

The concept of a global "hot spot" 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 doing so brought into focus the fact that much also exists that appears to be exceptional. Volcanic regions away from spreading plate boundaries or subduction zones, and large-volume on-ridge volcanism were considered to be not explained by plate tectonics. The term "hot spot" 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 ~ 20 such "hot spots" exists on Earth. He further postulated that they were fixed relative to one-another and that this could be explained if they were fuelled by hot 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 not convecting vigorously and therefore could 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 though to be "hot spots" revealed compositions distinct from mid-ocean ridge basalt (MORB), in particular being more enriched in incompatible elements and containing high $^{3}He/^{4}He$ ratios (Kellogg and Wasserburg, 1990; Schilling, 1973). Such basalts are called "ocean-island basalts" (OIB). Because it was assumed that "hot spots" are fueled by plumes from the deep mantle, their geochemical characteristics were thought to reveal lower mantle characteristics. The numbers of "hot spots" and plumes postulated rose to ~ 50 or above on many lists (Figure 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 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 “hot spots”, and OIB geochemistry, are also expected. High $^3He/^4He$ ratios were predicted following their initial observation in some basalts from Hawaii and Iceland, although this is not intrinsic to the original plume hypothesis.

The skepticism for the plume model that has arisen since the beginning of the 21st century has resulted largely from a growing awareness that most of these basic predictions are not fulfilled at most “hot spots”. Indeed, no single volcanic system on Earth unequivocally exhibits all the required characteristics listed above (Anderson, 2005b; Courtillot et al., 2003), 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 for a precursory LIP with uplift is absent (Shapiro et al., 2006), the locus of volcanism has not remained fixed relative to the palaeomagnetic pole (Tarduno and Cottrell, 1997), volcanism there 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* adaptions of the model to suit each locality. These include suggestions that predicted but unobserved features exist but have not yet been discovered, or are fundamentally undetectable, or due to the long-distance lateral transport of melt in the mantle from a distant plume (e.g., Ebinger and Sleep, 1998; e.g., 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).

This approach is unacceptable to many scientists, who have in the last few years worked to develop alternative models. The most widely applicable of these attributes melting anomalies on Earth’s surface to shallow, plate-tectonic-related processes (Anderson, 2000, 2001; Anderson, 2005a; Foulger, 2004; Foulger and Natland, 2003; Foulger et al., 2005c). Fundamentally, this 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 will be referred to herein as the “Plate model”. It is the inverse of the Plume model, which attributes “hot spots” to processes associated with Earth’s bottom thermal boundary layer, the core-mantle boundary. Extra-terrestrial models, involving meteorite impacts, form a separate category of mechanism and will not be dealt with in this chapter.

Much has already been achieved by a few in a short time frame and has, most critically, highlighted problems that had gone largely unacknowledged in recent years. In this chapter 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 dehomogenising 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, 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 is largely borne out by the observation that approximately 1/3 of all melting anomalies are on or close to mid-ocean ridges (Figure 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 Iceland melting anomalies are also associated with oceanic microplates (Foulger, 2006b; Schilling et al., 1985a). Oceanic LIPs that formed at or near ridges include Shatsky Rise, which formed at a ridge-ridge-ridge triple junction (Sager, 2005; Sager et al., 1999) and the Iceland Plateau which is currently being created (Figure 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 also the Manihiki and Hikurangi Plateau (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 & Range province in the western U.S.A. 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 small-volume oceanic melting anomalies have been attributed to intraplate extension, *e.g.*, due to thermal contraction (Lynch, 1999; Sandwell and Fialko, 2004; Sandwell et al., 1995; Smith, 2003). Global stress field maps show that volcanic regions occur in regions of tensile stress (Lithgow-Bertelloni and Guynn, 2004; Lithgow-Bertelloni and Richards, 1995). Calculations of the stress field for the whole Pacific plate from cooling 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 localised 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 (Figure 3; Anderson, 2005a; Figure 3; Meibom and Anderson, 2004).

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 its journey across the ocean, the thickness of the oceanic lithosphere continually increases as it cools, asthenospheric mantle is plated onto its bottom, metasomatised, 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 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, eclogitisation 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 (Figure 4; Elkins-Tanton, 2005; Figure 4; Tanton and Hager, 2000). The continental mantle lithosphere thus recycled may be very old and have been refertilised by metasomatism. Other mechansims 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 is detected seismically beneath north America and commonly interpreted as a subducted slab (e.g., Grand, 1994). However, this 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 (Gorbatov et al., 2000; Ritsema et al., 1999). 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).

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, 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 is expected to be followed by post-eruption subsidence as the LIP is carried by plate motion away from the hot plume (Clift, 2005; Griffiths and Campbell, 1991). 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 (Elkins-Tanton, 2005; Kay and Kay, 1993).

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 worlds largest continental and oceanic LIPs respectively, and yet neither is associated with major precursory uplift (Czamanske, 1998; Ingle and Coffin, 2004; Ito and Clift, 1998; Ivanov, this volume; Neal et al., 1997). The Siberian Traps have a volume of some 4 x 10^6 km^3 (Masaitis, 1983) and were erupted at ~ 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 ~ 10^8 km^3 (Taylor, 2006). A plume origin for such a vast LIP would predict pre-eruption uplift of several kilometers. However, palaeodepth estimates from volatiles in glass samples and from study of sediments immediately overlying the crust from several ocean drill holes give elevations ranging from approximately sea level down to ~ 2.5 km below sea level at the time of eruption (Ingle and Coffin, 2004; Michael, 1999; Roberge et al., 2005). Subsidence subsequent to its formation has also been studied using oceanic sediments and was slower than for normal oceanic lithosphere, consistent with underlying mantle 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 northwesterly 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 ka. The pattern of subsequent subsidence permits only moderate temperature anomalies of no more than ~ 100 ˚C down to ~ 100 km depth and precludes very high temperature or very thick thermal sources (Clift, 2005).

Simpler, precursory uplift has been reported associated the ~ 250 Ma Emeishan LIP in SW China. There, variations in thickness of sedimentary rocks underlying the basalts suggest kilometer-scale domal uplift throughout a region ~ 900 km wide (He et al., 2003; Xu et al., this volume). Despite the fact that 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 some other LIPs. There is no unequivocal evidence for regular, regional uplift prior to emplacement of the Deccan Traps, which erupted at ~ 67 Ma and had a volume of ~ 1.5 x 10^6 km^3 before erosion. Instead there is evidence that such uplift did not occur

(Sheth, this volume). There is evidence of subsidence prior to emplacement of the Columbia River Basalts and uplift immediately afterwards (Hales et al., 2005) though this is contested (Hooper et al., this volume). Study of sedimentary cores from many oceanic plateaus have produced little evidence for post-emplacement subsidence, *e.g.*, at Shatsky Rise, the Kerguelen Plateau, the Walvis Ridge and 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 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 uplift predicted to accompany the influx of hot source material using realistic continental lithosphere rheologies. They modelled lithosphere that included elastic–brittle–ductile properties and a stratified structure. Their results suggest that the simple classical expectation of regional domal uplift is over-simplified, and a more complex pattern might occur (Figure 5). This work re-opens 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.

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 ~ 10^8 km^3 proposed composite Ontong Java–Manihiki–Hikurangi Plateau (Taylor, 2006). Determining the lowest-volume LIP is, on the other hand, a matter of definition. LIPs were originally defined as having an areal extent of > 0.1 x 10^6 km^2 (Coffin and Eldholm, 1994) but Sheth (2007), as part of a proposed classification scheme for LIPs, recently suggested that 0.05 x 10^6 km^2 is a more appropriate lower limit (see Bryan and Ernst, 2006; Sheth, 2006b and linked webpages for a substantial discussion). Such debate hints that the sizes of volcanic provinces may form a continuous spectrum and invites the question whether there is a natural break in the size distribution that might form the basis for theories of genesis.

At the small end of the volume scale are island chains such as the Samoa, Tahiti, Marquesas, Pitcairn and Louisville chains (Figure 1). The eruption rates of these chains throughout their lifetimes have been much less than 0.01 km^3/a. In some cases, chains are discontinuous and the eruption rate was almost zero for long periods, *e.g.*, the youngest ~ 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 rate also varies 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^3/a 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 Ma, giving a sustained magmatic rate of up to several km^3/a (Fitton et al., 2004). However, this is conjectural as the thick lower crust cannot be sampled. Ito and Clift (1998) suggested that the slow post-emplacement subsidence observed there might be explained if the lower crust continued to be thickened by intrusions from a plume tail over a time period of several tens of Ma, but this seems unlikely, if only because the plateau drifted by ~ 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 Hawaii 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 (Figure 6). It was ~ 0.01 km^3/a for much of the lifetime of the Emperor chain. This is approximately the volume rate of a single mid-ocean-ridge spreading segment. The volume rate declined to almost zero at the time of the bend (~ 47 Ma; Sharp and Clague, 2006), and grew to somewhat higher rates, or up to ~ 0.17 km^3/a for much of the Hawaiian chain. Very recently however, the magmatic production rate underwent a rapid increase, and is currently ~ 0.25 km^3/a (Robinson and Eakins, 2006). This 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 last ~ 5 Ma has an area of ~ 0.15×10^6 km^2 and a volume of ~ 0.6×10^6 km^3 (Joel Robinson, personal communication, 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. There, magmatism in excess of local ocean-crust formation has occurred for the ~ 54 Ma duration of the opening of the north Atlantic. Following continental breakup, the excess magmatic rate was less than ~ 0.1 km^3/a for the first ~ 10 Ma, during which time the Iceland-Faeroe ridge was built, with its modest north-south extent of ~ 100 km (Figure 7). Since ~ 44 Ma, however, the Icelandic Volcanic Plateau has been built. It has a north-south extent of up to ~ 600 km and a crustal thickness of up to ~ 20 km greater than the crust on the Reykjanes ridge. It must thus have been built at an excess rate of ~ 0.25 km^3/a, approximately the same as at present-day Hawaii. The Iceland plateau presently covers an area of ~ 0.3×10^6 km^2 and thus has an excess volume of ~ 5×10^6 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 a 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 49 "hot spots" cataloged by Courtillot *et al.* (2003) only 13 associations of LIPs and volcanic chains are listed (Vema, Reunion, Marion, Crozet, Kerguelen, Macdonald, Marquesas, Easter, Yellowstone, Galpagos, Iceland, Fernando and Tristan), and only two of these are cited as being unequivocal (Reunion/Deccan and Tristan/Parana). Both of these two 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 is typically a few hundred kilometers. Such experiments have been conducted to study the Iceland, Yellowstone and Eifel melting anomalies (Christiansen et al., 2002; Foulger et al., 2001; Iyer et al., 1981; Ritter et al., 2000). 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, at 410-660-km depth, 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 mantle. Variations in the spherical harmonic power of velocity throughout the mantle shows that the Earth is characterised by strong heterogeneity in the longer-wavelength components in the upper mantle but low heterogeneity in the lower mantle (Gu et al., 2001). At individual melting anomalies, *e.g.*, Iceland, Yellowstone and 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 hot spots 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)

In order 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. 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 (Figure 8; Bina and Helffrich, 1994; Presnall, 1995). 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, 2001). This pattern is also observed at melting anomalies, *e.g.*, Yellowstone, Iceland and Eifel. There, deflections on the 410-km discontinuity are observed but no corresponding deflections on the 660-km discontinuity (Du et al., 2006; Dueker and Sheehan, 1997; Grunewald et al., 2001).

Global tomography suffers from formidable problems of resolution and coverage, which become increasingly severe with depth, and with many regions of the mantle devoid of seismic rays. Repeatability between different models is also a problem, with proposed lists of deeply sourced “hot spots” often exhibiting little overlap (Table 1; Anderson, 2005b). 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 presence of partial melt and volatiles and temperature (Figure 9; Anderson, 1989). Interpreting anomalies is thus highly ambiguous and nowhere is this 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 suggests 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 comprise 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 appropriate state of lithospheric stress. Study of temperature variations in the presumed source region is thus important for determining the source process. This is not an easy task, however. The interpretation of seismic velocity variations is ambiguous. Surface heat flow 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 crystallisation path) to obtain equilibrium with the most magnesian 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, 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 be assumed to represent liquid compositions. Putirka (2005), using an olivine geothermometer described in the same paper, reported a temperature of ~245 ± 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 ~ 250˚C between the Hawaiian and MORB sources. Green *et al.* (2001), Green and Falloon (2005) and Falloon *et al.* (this volume), on the other hand, 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 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., 2006).

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 comprises 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 modelled 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 Ma. Melting of entrained eclogite, which is more fusible, was therefore modelled. 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, and not supplied by decompression. Only small to 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 ~ 100 Ma old, and thermal upwelling models can only simulate magmatic rates << 0.1 km^3/a for reasonable temperatures.

In the case of near-ridge melting anomalies, the question arises whether the excess melting can be modelled 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 (Anderson, 2005a; Foulger and Anderson, 2005; Korenaga, 2005; Yaxley and Green, 1998). 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, metasomatised 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 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 between many basalts from ocean islands (OIB) and mid-ocean ridges (MORB) (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.*, $^{87}Sr/^{86}Sr$) in MORB are indicative of past melting events that caused relative depletion in the parent elements compared to the source of OIB. OIB 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 $^{3}He/^{4}He$ 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 small-volume magmas erupted in continental rifts where a deep thermal origin is implausible (Fitton, this volume; Natland and Winterer, 2005). Eruption of similar compositions may furthermore continue for tens or even hundreds of Ma at the same locality, *e.g.*, in the Scottish Midland Valley. There, OIB-like volcanism persisted for about 60 Ma during which time the lithosphere is expected to have drifted about 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 which preferentially extracts elements incompatible in mantle phases and transports them upwards. 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 materials. 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 materials that may be recycled into the mantle include oceanic crust along with overlying pelagic and terrigenous sediments and metasomatised mantle lithosphere. Ocean crust is re-introduced into the mantle at subduction zones. Metasomatised 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 in water-soluble incompatible elements 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.* (2004; 2002) recently proposed an origin in variably metasomatised mantle lithosphere, through 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 comprise a special category of geochemical observations because, whereas it is generally acknowledged that other geochemical characteristics are not diagnostic of source depth, high $^3He/^4He$ is widely assumed to indicate a deep lower-mantle origin. In some samples from melting anomalies *e.g.*, Iceland and Hawaii, values of $^3He/^4He$ as high as ~ 50 times Ra (the atmospheric value of 1.38 x 10^{-6}) are measured. This 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 ^3He, 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 ^4He, a radiogenic decay product of U and Th (Figure 11). In this model, helium is stored in a low-(U+Th) matrix, *e.g.*, dunite cumulates, and ancient, high-$^3He/^4He$ values are preserved for long periods with little change (Anderson, 1998a, 1998b; Meibom et al., 2005; Natland, 2003). Dunite cumulates occur in the lower parts of oceanic crust, and this model thus suggests that high-$^3He/^4He$ ratios may be explained by recycled subducted oceanic crustal lithologies in the same way as other OIB characteristics. The neon isotope ratios found in some OIB are low in ^{21}Ne compared with those measured in MORBs. ^{21}Ne is produced by the nucleogenic decay of ^{24}Mg and ^{18}O when irradiated by U and Th decay products. It is thus produced at the same rate as ^4He, which suggests that the neon isotope ratios observed in OIB may be explained in the same way as the helium isotope ratios.

The question of the mantle dynamical implications 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 timescale of slab reheating is much less than the age of the plate on subduction, and since 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

The 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 (Figures 1 & 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. Large-volume volcanism along a ~ 2,500-km zone of rifting accompanied the breakup of Laurasia and the opening of the north Atlantic ocean at ~ 54 Ma. This subsequently dwindled as normal-thickness ocean crust began to form, but the magmatic rate continued at a high level along a ~ 100-km-long portion of the mid-Atlantic ridge, building the Iceland-Faeroe ridge (Figure 7). Starting at ~ 44 Ma, the part of the ridge that produced excess volcanism greatly lengthened to attain a present-day north-south extent of ~ 600 km. If the whole of the ~ 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 ~ 0.25 km^3/a. 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 ~ 0.3 x 10^6 km^2, so it qualifies as a LIP (Coffin and Eldholm, 1994; Sheth, 2007).

Temperature estimates for the mantle beneath Iceland have been obtained using seismology, petrology, bathymetry, vertical motion and heat flow (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 (Du et al., 2006; Foulger et al., 2001; Montelli et al., 2004; Ritsema et al., 1999).

These results, coupled with observed tectonic correlations have inspired development of a Plate-based model for Iceland. The Greenland-Iceland-Faeroe ridge formed co-linear with a northwesterly 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 ~ 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 (Foulger, 2006b; Nunns, 1983).

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; Foulger et al., 2005b). 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 ~ 400 Ma and 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 caused by volatiles, in particular CO_2 (Dasgupta and Hirschmann, 2006; Presnall and Gudfinnsson, 2005). 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 behaviour 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 ~ 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/a (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 ^{203}Pb/^{204}Pb is up to 11.3. The discovery of zircons dated at 330 Ma and 1.6 Ga in the Kane fracture zone ~ 1,500 km south of the Azores lends support to this (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 there (Georgen and Lin, 2002).

Magmatism in the Ascension-Cameroon region is complex. The Cameroon line has erupted along much of its length between ~ 130 Ma and the present, and no regular time progression is observed (Ernst and Buchan, 2002; Fitton, 1987; Fitton and Dunlop, 1985). 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 with extensive information constraining the tectonic history of Africa and South America. They concluded that changes in stratigraphy in African rift basins reflected changes 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. Additinal observations in support of this are reported from seismic reflection studies of the Cabo Frio area, part of the Rio Grande Rise immediately offshore Brazil (Oreiro et al., 2005).

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.

The Pacific ocean

The most remarkable currently-active melting anomaly in the Pacific ocean is the Hawaiian archipelago at the SE end of the Emperor and Hawaiian volcanic chains (Figure 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 high present-day production rate of tholeiitic basalt. The most widely suggested non-thermal 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 (Norton, this volume; Shapiro et al., 2006);
1. Both chains are regularly time progressive (Sharp and Clague, 2006);
1. 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 ~ 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 last ~ 2 Ma (Robinson and Eakins, 2006), resulting in LIP emplacement *following* the development of the time-progressive chains rather than preceding it (Figure 6). This change in magmatic rate has apparently been accompanied by a very recent change in propagation direction from approximately southeast to approximately south-southeast;
6. The youngest half of the Hawaiian chain is surrounded by a bathymetric swell (Figure 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. This limits the aperture of land seismic networks and geochemical sampling.

The temperature of the mantle beneath Hawaii has been investigated using sea-bottom heat flow measurements and petrology. Heat flow measurements across the Hawaiian swell do not detect a heat flow anomaly of the kind expected for a thermal source (von Herzen et al., 1989) although recently it has been suggested that hydrothermal circulations masks expected anomalies (Harris and McNutt, 2005). Estimates from picrite glass samples of the mantle temperature anomaly vary from zero to ~ 250˚C (Falloon et al., this volume; Green and Falloon, 2005; Putirka, 2005).

The geochemistry of the lavas is variable with both enriched and depleted signatures consistent with components of continental and marine sediments and oceanic crust, and variable isotopic ratios. The geochemistry varies geographically where the Hawaiian and Emperor chains cross fracture zones, *e.g.*, the Mendocino, Murray and Molakai fracture zones (Basu and Faggart, 1996). Variations with time area also reported, for example, Mukhopadhyay *et al.* (2003) describe changes in $^{3}He/^{4}He$ 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 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 $^{3}He/^{4}He$ 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 only 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-wave-speed 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 Hawaii region and the core-mantle boundary beneath a large swathe 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 peter out in the mid-mantle (Montelli et al., 2004). Transition zone thickness beneath the Hawaii 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.

A little work only has been done to develop numerically a Plate model for Hawaii involving a propagating crack (Jackson and Shaw, 1975; Jackson et al., 1975; Lithgow-Bertelloni and Guynn, 2004; Natland and Winterer, 2005; Shaw, 1973; Shaw and Jackson, 1973; Turcotte, 1974; Turcotte and Oxburgh, 1973). 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 there. 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 near simultaneous change from the western part of the Pacific plate being bounded by the Kula-Pacific and North New Guinea-Pacific ridges, to subducting 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 naturally 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 needs urgently to be tested, ideally by measuring extension normal to the chain. However, detecting in the deep ocean extension that might be of the order of 1 mm/a 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. Small-volume seamounts of alkalic ocean-island basalt are widely scattered over the ocean floor and most do not plausibly form discrete chains. Linear arrays of seamounts are common, on the other hand, 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 and Fialko, 2004) (Sandwell et al., 1995). Volumetrically larger melting anomalies do exist, *e.g.*, 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).

The 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 is exhibited. The volumes of magma observed range from very small to LIP-sized with volumes up to ~ 4 x 10^6 km^3 in the case of the Siberian traps (Figure 2). Much of the Siberian traps is thought to have erupted within a few Ma (Ivanov, this volume; Kamo et al., 2003). However, this may be compared with volcanism in the Basin and Range province, western USA, where widespread, relatively small-volume eruptions have continued since ~ 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. Small-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). Large-volume basaltic LIPs generally have OIB-like compositions also, although the Siberian traps is an exception (Fitton, this volume). It has recently been recognised 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 timescales 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; He et al., 2003; Rainbird and Ernst, 2001; Xu et al., 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 post-date 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 which formed from ~ 17 Ma to the present (Christiansen et al., 2002). This chain is aligned to 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 and 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 particularly 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 development of gravitational instability and detachment of the mantle lithosphere and possibly part of the lower crust (Hales et al., 2005; Lustrino, 2005; Tanton and Hager, 2000). Back-arc and slab-deformation and fragmentation processes have been suggested to have triggered volcanism in regions such as Mexico, Turkey the Basin and Range province and east Asia (e.g., Ferrari, 2003; 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. It is fundamentally this that has, 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-Frottier, 2005b) or 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.

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

Although melting anomalies or "hot spots" have traditionally been considered a single phenomenological class, a general review reveals extreme variation in almost every respect. This 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 "hot spot" 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 volcanism is required that can account for the full spectrum of size, longevity, composition and spatial and temporal distribution of volcanism.

Two LIPs are currently in the process of formation, Iceland (with a present-day area and volume of ~ 0.3×10^6 km^2 and ~ 5×10^6 km^3 respectively) and Hawaii (with a present-day area and volume of ~ 0.15×10^6 km^2 and ~ 0.6×10^6 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 extension rate or in the fertility of the source material being tapped. Melting anomalies that exhibit LIP volcanism followed by small-volume, time-progressive volcanic chain formation as predicted by the Plume model are at best very rare, and perhaps non-existent.

There is no unequivocal evidence for systematically elevated temperature at LIPs or “hot spots” in general. Petrological approaches using olivine control lines are still contested. The traditional assumption that high-Fo olivines and picrite cumulates *require* high temperature is now recognised to be incorrect (Falloon et al., this volume; Falloon et al., 2006; Green and Falloon, 2005; Green et al., 2001; Presnall and Gudfinnsson, 2005). 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 is engineered by temperatures elevated by hundreds of degrees above those beneath normal mid-ocean ridges remains elusive. On the other hand, the OIB geochemical signature of “hot spot” lavas is prima facie evidence for fusible material in the source and the percentage in recent Hawaiian and Icelandic lavas may be very large (e.g., Foulger et al., 2005b; e.g., 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 independent of plate motion, giving rise to time-progression. At the other extreme they may appear to travel with the plate. The generally agreed 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 (Figure 11; Parman et al., 2005; Stuart et al., 2003).

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 non-progressive, along with many shorter, small-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 modelling of decompression melting in plumes, even if containing fusible material, cannot reproduce the volumes of the largest LIPs at reasonable temperatures 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 defied 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 last 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 modelling variations in stress in the lithosphere, both locally and on regional- and plate-wide scales;
2. Numerical modelling of the melt volumes produced in different extensional settings;
3. Quantification of the volumes of melt produced by isentropic upwelling of non-peridotic source material;
4. Convection modeling;
5. Solving the paradox of the origin of OIB;
6. Detailed study of the history of vertical motion associated with major LIPs and comparisons 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 "hot spot" lavas;
11. Investigating the relationship between extensional faulting and related volcanism;
12. Erecting and testing Plate models for melting anomalies that have not yet been subject to this type of scrutiny.

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Table 1: Hotspots reported to be underlain by seismic anomalies traversing the upper mantle (traversing the upper mantle only – the deepest they detected; Ritsema and Allen, 2003), traversing the whole-mantle (Montelli et al., 2004) and those defined as arising from D" by Courtillot *et al.* (2003). Only Easter is common to all three lists.

	Traversing the upper mantle (Ritsema and Allen, 2003)	Traversing the whole-mantle (Montelli et al., 2004)	Defined as arising from the core-mantle boundary (Courtillot et al., 2003)
Afar	√		√
Hawaii	√		√
Bowie	√		
Samoa	√	√	
Macdonald	√		
Easter	√	√	√
Louisville	√		√
Iceland	√		√
Tahiti		√	
Azores		√	
Canaries		√	
Ascension		√	
Reunion			√
Tristan			√

Figure captions

Figure 1: Map showing global "hot spots" compiled using the list of 49 of Courtillot *et al.* (2003). (See commentary on this list by Anderson (2005b)). Red, yellow and green dots indicate "hot spots" 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)

Figure 2: Map of global LIPs. EAR: East African Rift, B&R: Basin & Range province, CRB: Columbia River Basalts, Y-SRP: Yellowstone-Snake River Plain, HLP: High Lava Plains. Basemap by Laurent Gernigon.

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).

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 dewater as it sinks and heats, triggering melting of the mantle or of itself (from Elkins-Tanton, 2005).

Figure 5: Sketches of plume-lithosphere interactions: (a) a "conventional" plume impinges on a single-layer viscous lithosphere resulting in a single, large-wavelength topographic signature, (b) when realistic lithosphere rheology (brittle – elastic – ductile) and multi-layer 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).

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).

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 comprises the innermost, subaerial part of the broader, mostly submarine, Icelandic Volcanic Plateau. Oceanic magnetic anomalies (Nunns, 1983) are labelled 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, grey:

spreading segments, white: glaciers. Dashed lines: extinct rift zones (two in west Iceland and two in east Iceland).

Figure 8: Minerological 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).

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 to perhaps 30% (reproduced from Julian, 2005).

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 from 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 LIPs for reasonable temperature anomalies.

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

Figure 12: Topography of the Earth based on a global compilation of land data (GTOPO30, http://edcdaac.usgs.gov/gtopo30/README.html) and ocean data from a combination of sparse ship soundings and marine gravity anomalies derived from satellite altimetry (Smith and Sandwell, 1997).

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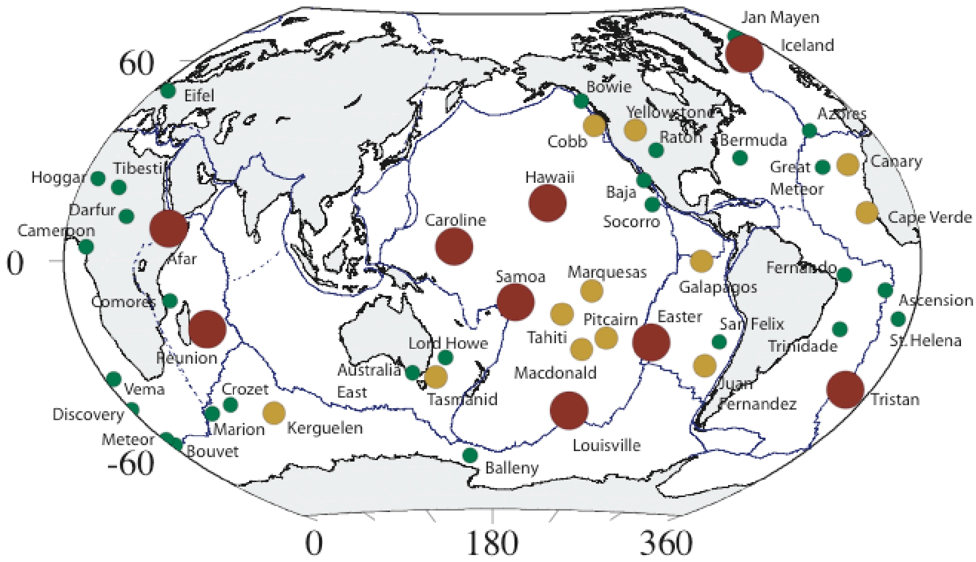

Jan Mayen
Iceland
60
Eifel
Bowie
Yellowstone
Azores
Cobb
Raton
Bermuda
Hoggar
Tibesti
Great
Meteor
Canary
Hawaii
Baja
Darfur
Socorro
Cape Verde
Cameroon
Caroline
Afar
0
Galapagos
Fernando
Samoa
Marquesas
Ascension
Comores
Tahiti
Pitcairn
Easter
San Felix
St. Helena
Reunion
Lord Howe
Trinidade
Macdonald
Australia
East
Juan
Fernandez
Tristan
Vema
Crozet
Tasmanid
Discovery
Marion
Kerguelen
Meteor
Louisville
Bouvet
-60
Balleny
0
180
360

Figure 1

World sedimentary basins, Large Igneous Provinces and distribution of volcanic rifted margins

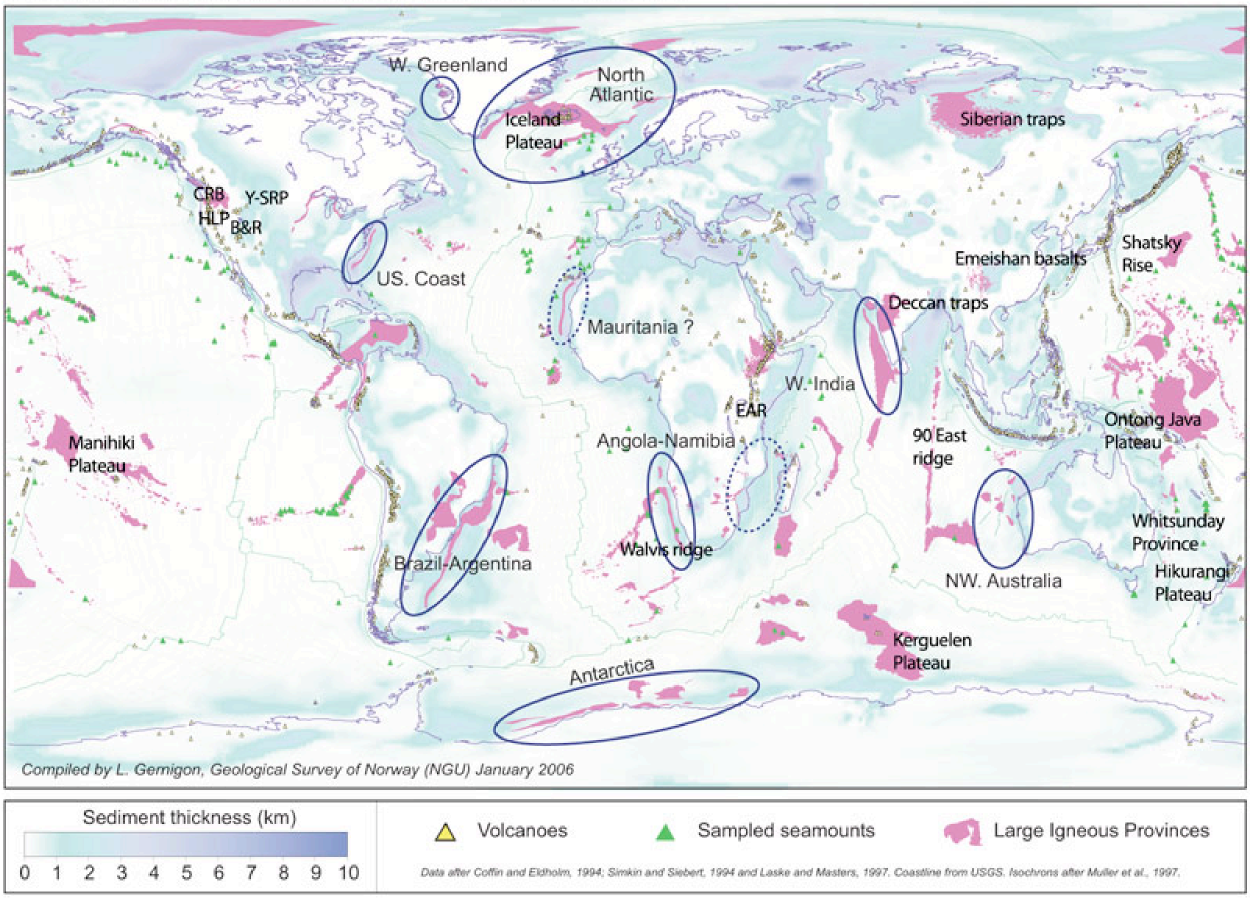


Figure 2

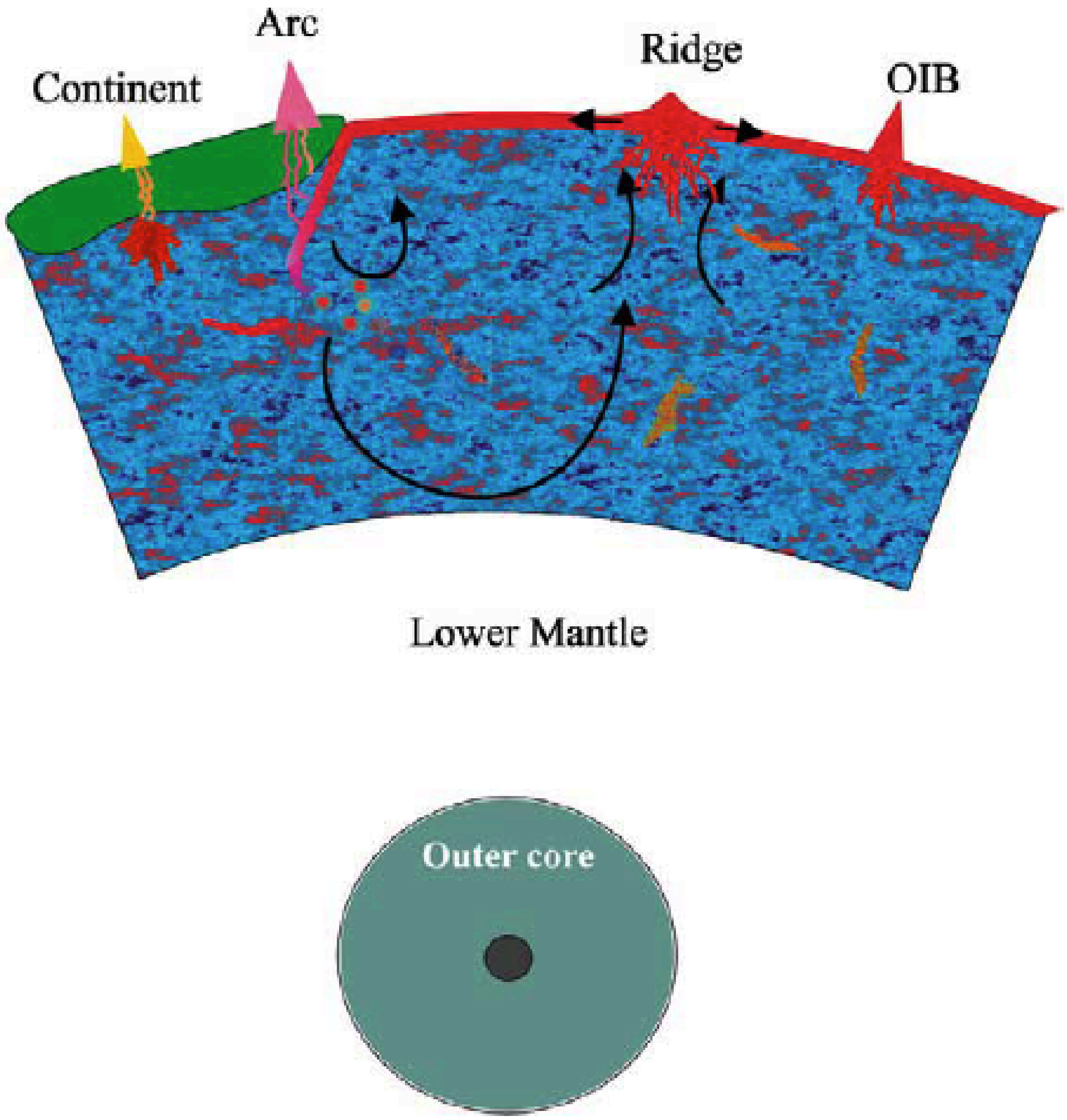


Figure 3

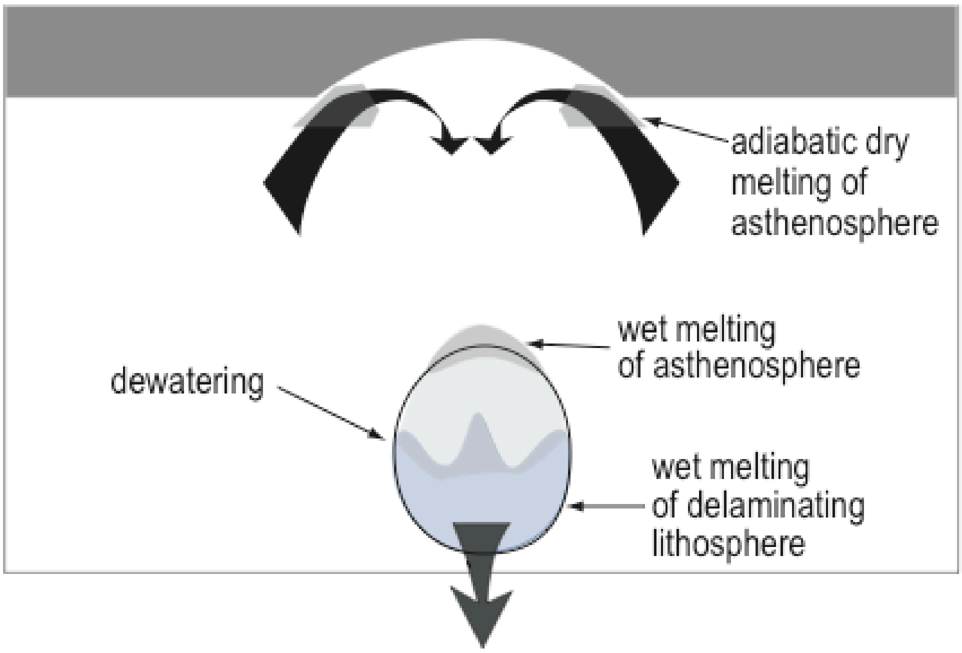

adiabatic dry
melting of
asthenosphere
wet melting
of asthenosphere
dewatering
wet melting
of delaminating
lithosphere

Figure 4

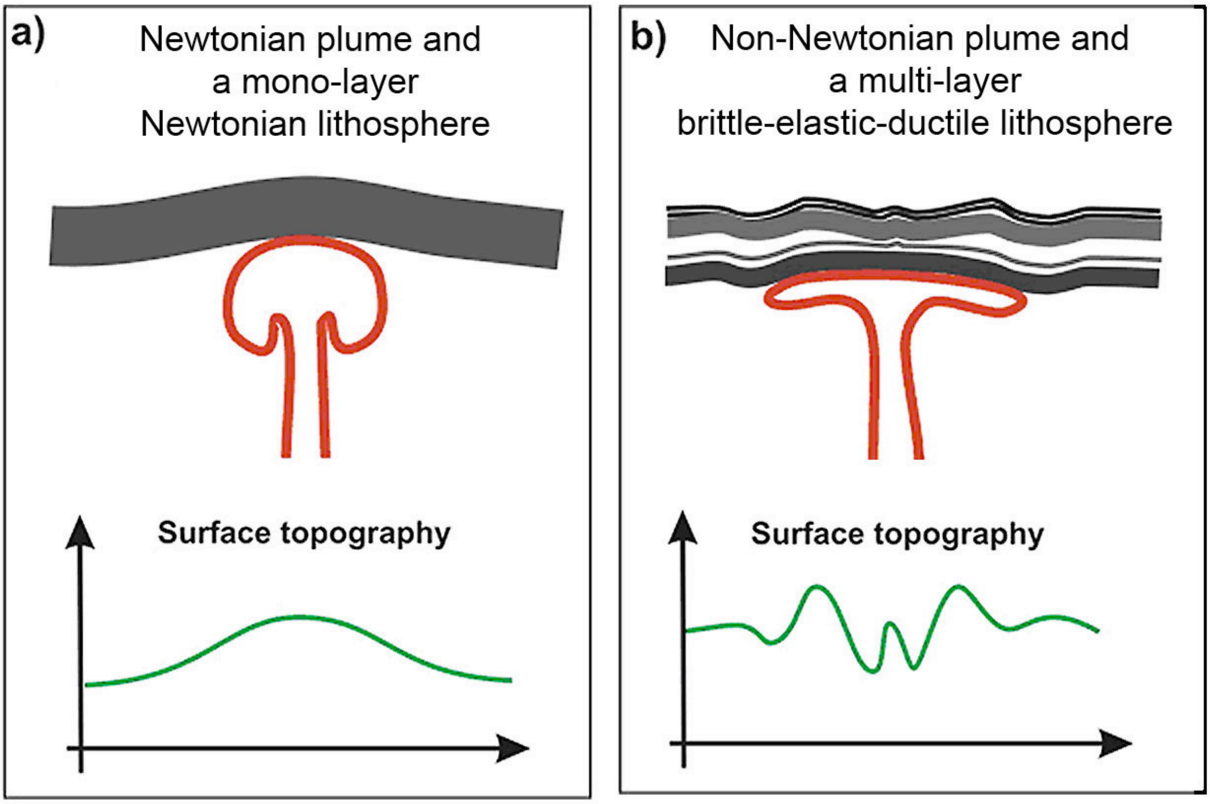

a)
Newtonian plume and
a mono-layer
Newtonian lithosphere
Surface topography
b)
Non-Newtonian plume and
a multi-layer
brittle-elastic-ductile lithosphere
Surface topography

Figure 5

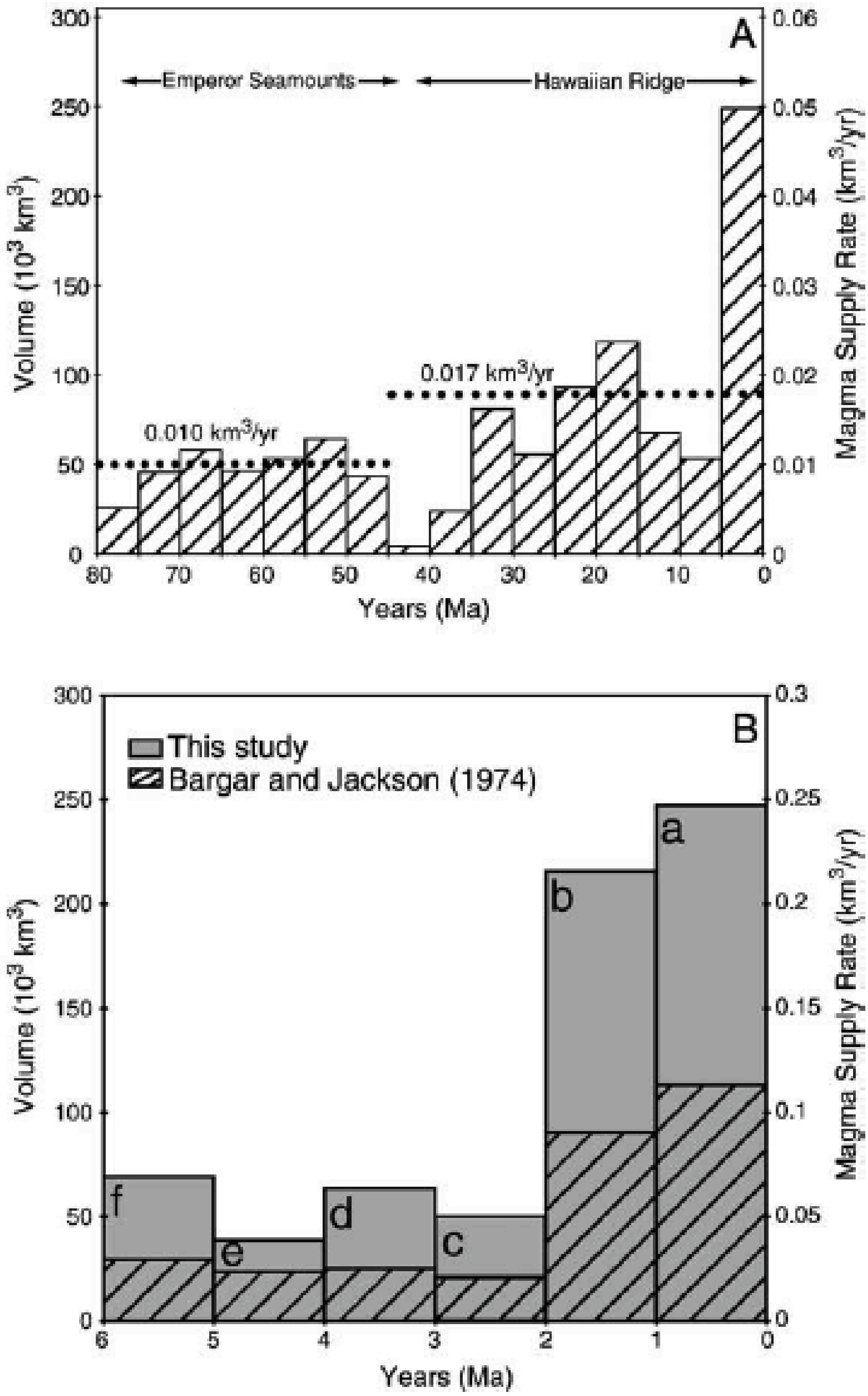

A
Emperor Seamounts
Hawaiian Ridge
0.017 km³/yr
0.010 km³/yr
Volume (10³ km³)
Magma Supply Rate (km³/yr)
Years (Ma)
B
This study
Bargar and Jackson (1974)
a
b
c
d
e
f
Volume (10³ km³)
Magma Supply Rate (km³/yr)
Years (Ma)

Figure 6

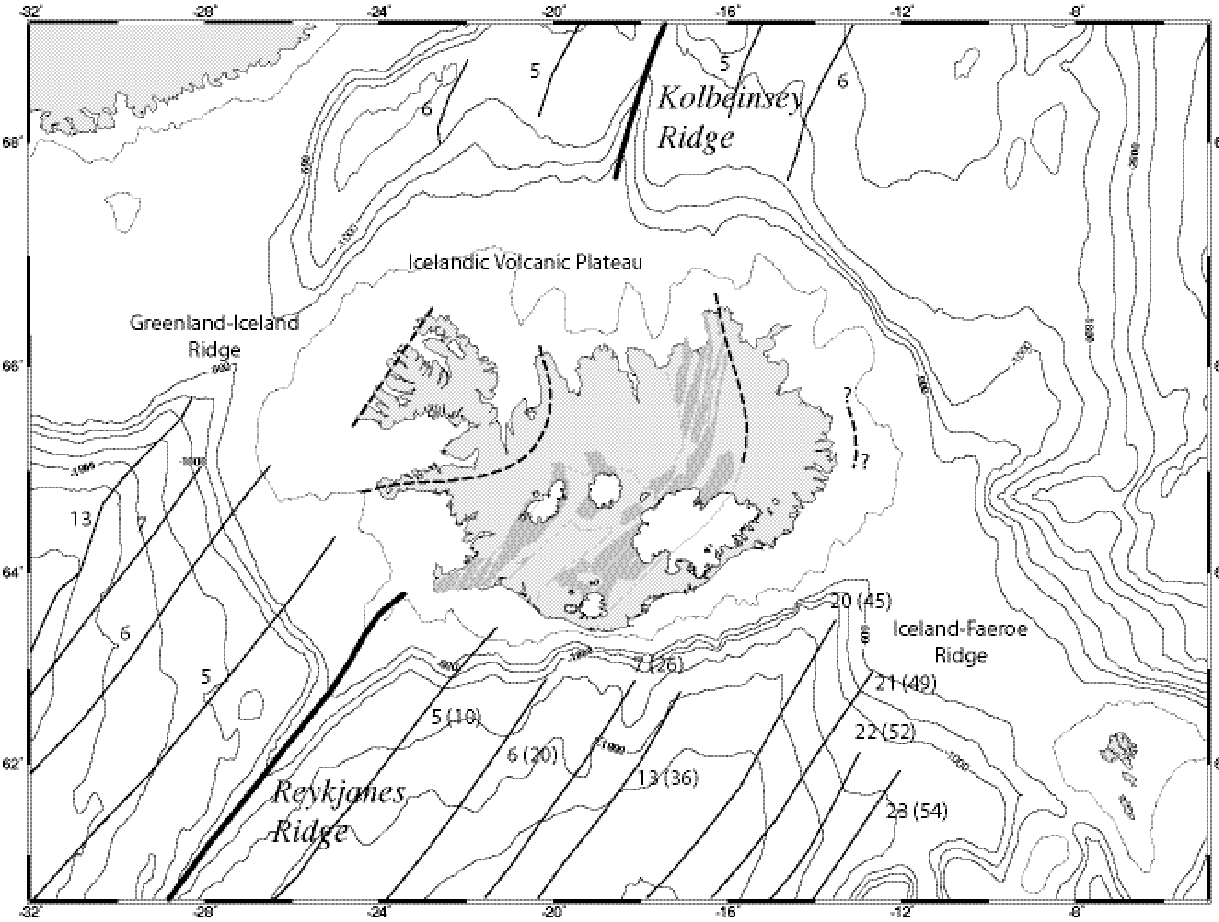

-32°
-28°
-24°
-20°
-16°
-12°
-8°
68°
66°
64°
62°
Kolbeinsey Ridge
Icelandic Volcanic Plateau
Greenland-Iceland Ridge
Iceland-Faeroe Ridge
Reykjanes Ridge
5
6
13
20 (45)
7 (26)
21 (49)
5 (10)
6 (20)
13 (36)
22 (52)
23 (54)
?

Figure 7

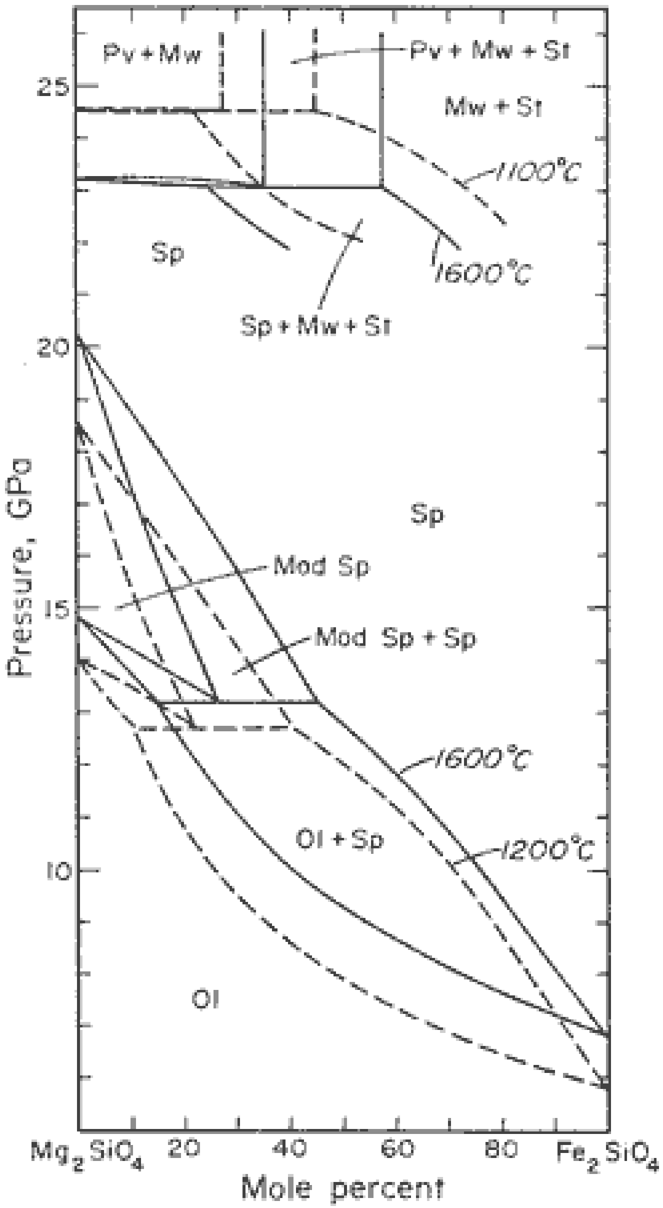

Pv + Mw
Pv + Mw + St
Mw + St
1100°C
Sp
1600°C
Sp + Mw + St
Sp
Mod Sp
Mod Sp + Sp
1600°C
Ol + Sp
1200°C
Ol
25
20
15
10
Pressure, GPa
Mg_2SiO_4
20
40
60
80
Fe_2SiO_4
Mole percent

Figure 8

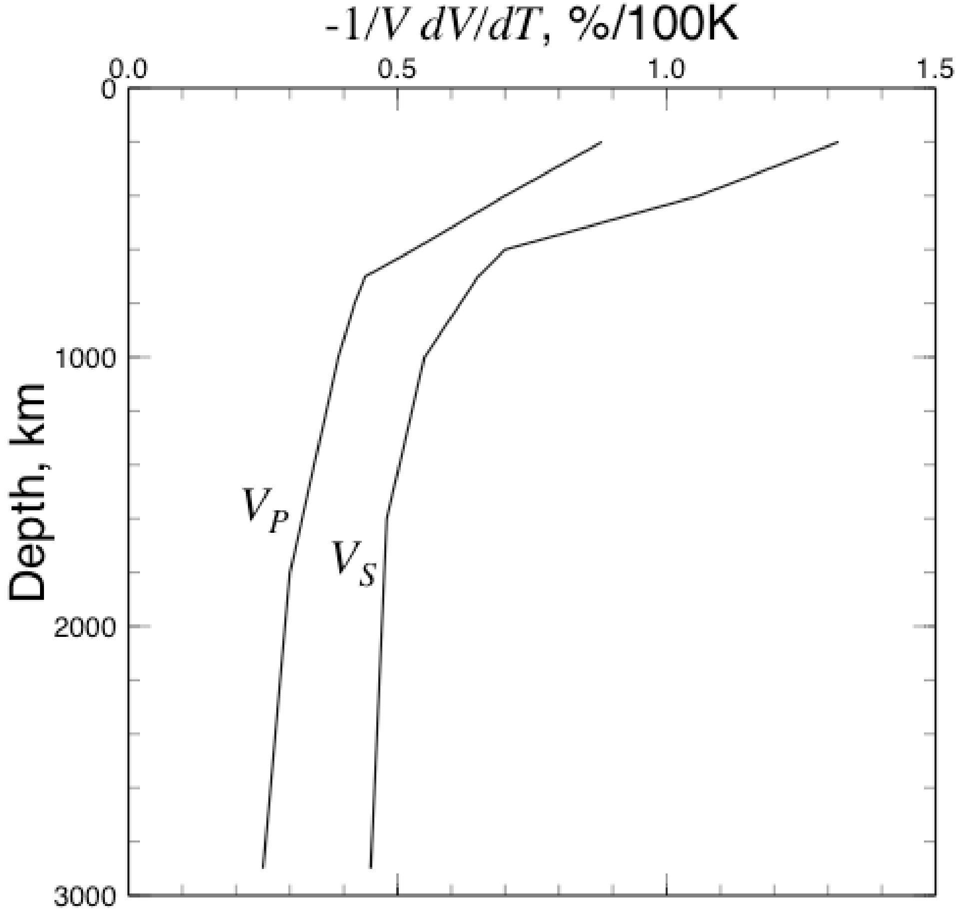


Figure 9

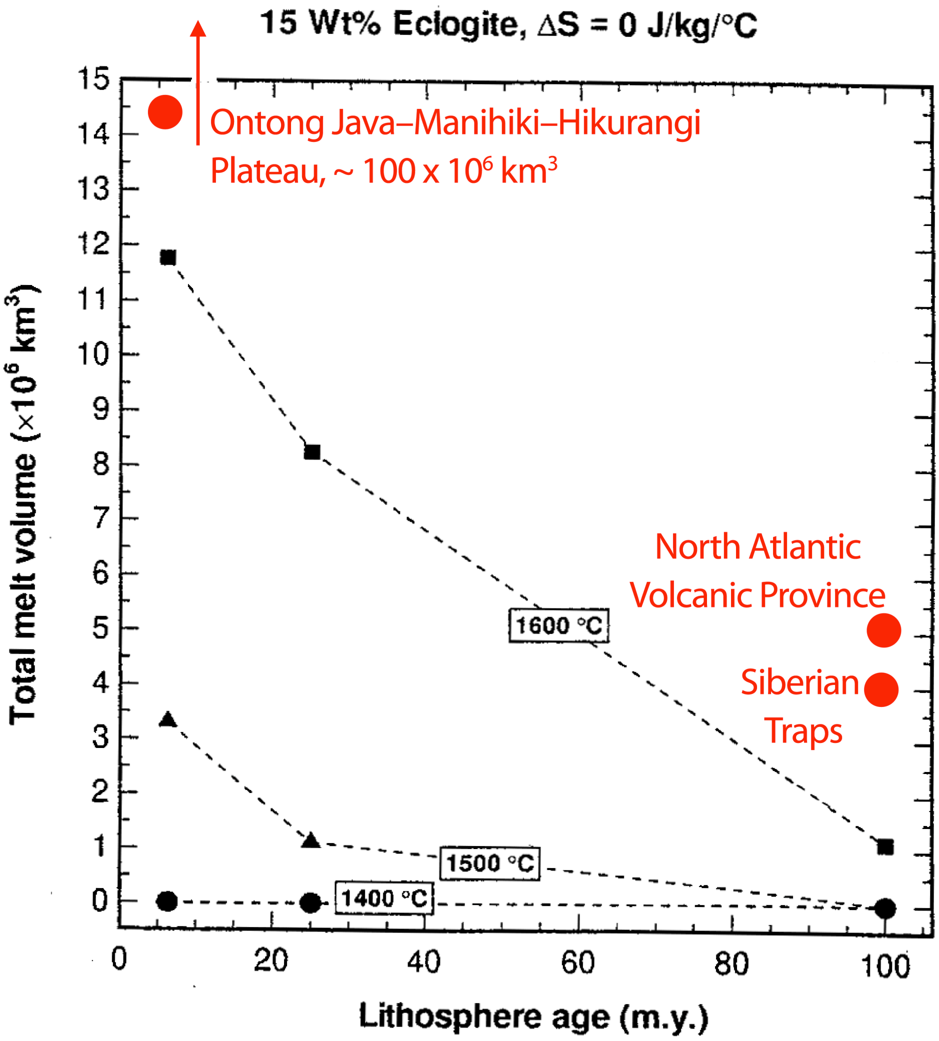
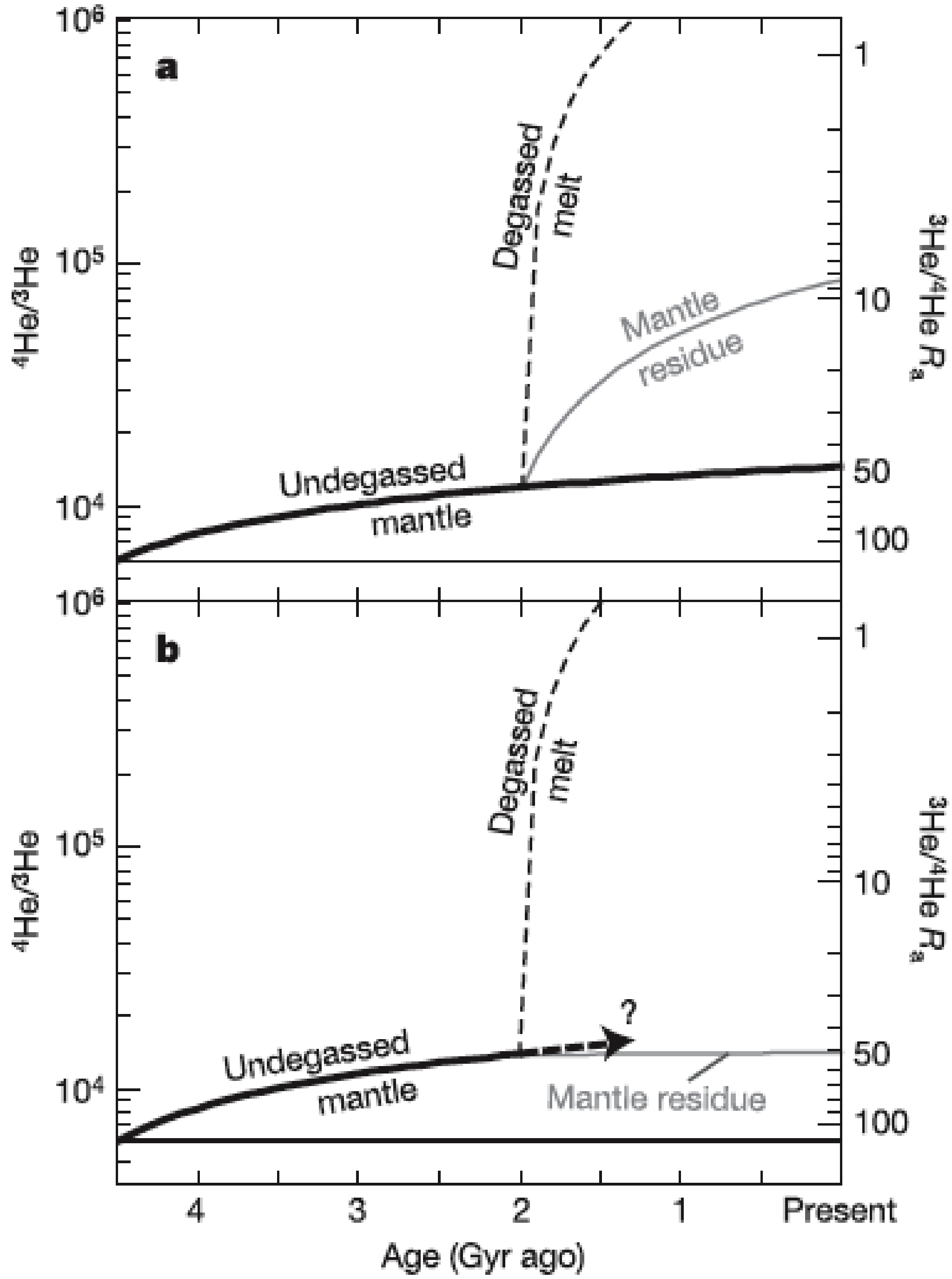

15 Wt% Eclogite, ΔS = 0 J/kg/°C
Ontong Java–Manihiki–Hikurangi Plateau, ~ 100 x 10^6 km^3
North Atlantic Volcanic Province
Siberian Traps
1600 °C
1500 °C
1400 °C
Total melt volume (×10^6 km^3)
Lithosphere age (m.y.)
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
0
20
40
60
80
100

Figure 10



a
b
10^6
10^5
10^4
$^4He/^3He$
$^3He/^4He$ R_a
1
10
50
100
Degassed melt
Mantle residue
Undegassed mantle
?
4
3
2
1
Present
Age (Gyr ago)

Figure 11

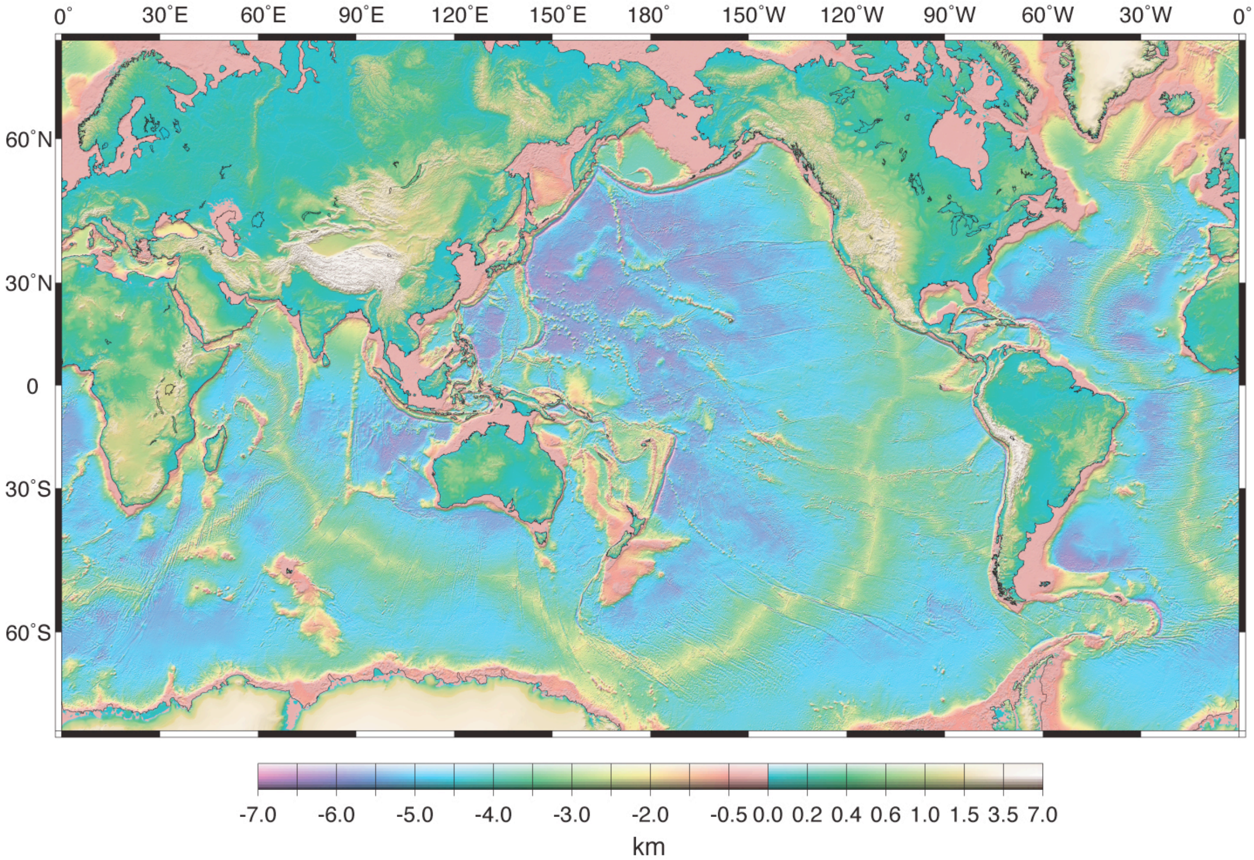

0˚
30˚E
60˚E
90˚E
120˚E
150˚E
180˚
150˚W
120˚W
90˚W
60˚W
30˚W
0˚
60˚N
30˚N
0
30˚S
60˚S
-7.0 -6.0 -5.0 -4.0 -3.0 -2.0 -0.5 0.0 0.2 0.4 0.6 1.0 1.5 3.5 7.0
km

Figure 12