ARE 'HOT SPOTS' HOT SPOTS?

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Abstract

The term 'hot spot' emerged in the 1960's from speculations that Hawaii might have its origins in an unusually hot source region in the mantle. It subsequently become widely used to refer to volcanic regions considered to be anomalous in the then-new plate tectonic paradigm. It carried with it the implication that volcanism a) is emplaced by a single, spatially restricted, mongenetic melt-delivery system, assumed to be a mantle plume, and b) that the source is unusually hot. This model has tended to be assumed *a priori* to be correct. Nevertheless, there are many geological ways of testing it, and a great deal of work has recently been done to do so. Two fundamental problems challenge this work. First is the difficulty of deciding a 'normal' mantle temperature against which to compare estimates. This is usually taken to be the source temperature of mid-ocean ridge basalts (MORB). However, Earth's surface conduction layer is ~ 200 km thick, and such a norm is not appropriate if the lavas under investigation formed deeper than the 40-50 km source depth of MORB. Second, methods for estimating temperature suffer from ambiguity of interpretation with composition and partial melt, controversy regarding how they should be applied, lack of repeatability between studies using the same data, and insufficient precision to detect the 200-300˚C temperature variations postulated. Available methods include multiple seismological and petrological approaches, modelling bathymetry and topography, and measuring heat flow. Investigations have been carried out in many areas postulated to represent either (hot) plume heads or (hotter) tails. These include sections of the mid-ocean spreading ridge postulated to include ridge-centred plumes, the North Atlantic Igneous Province, Iceland, Hawaii, oceanic plateaus, and high-standing continental areas such as the Hoggar swell. Most volcanic regions that may reasonably be considered anomalous in the simple plate-tectonic paradigm have been built by volcanism distributed throughout hundreds, even thousand of kilometres, and as yet no unequivocal evidence has been produced that any of them have high temperature anomalies compared with average mantle temperature for the same (usually unknown) depth elsewhere. Critical investigation of the genesis processes of 'anomalous' volcanic regions would be encouraged if use of the term 'hot spot' were discontinued in favor of one that does not assume a postulated origin, but is a description of unequivocal, observed characteristics.

1. Introduction

The origin of the term 'hot spot', also written 'hotspot', is obscure. It emerged in the 1970s to signify an active volcanic region that appeared to not fit the then-new plate tectonic hypothesis. It was originally understood to signify an unusually hot region in the mantle that gave rise to surface volcanism unconnected with a plate boundary [Wilson, 1963]. Development of the concept was inspired by Hawaii – a unique phenomenon, given its intraplate setting, huge present-day volcanic production rate, and exceptionally long, narrow, time-progressive volcanic chain.

The term 'hot spot' subsequently became inexorably linked with the plume hypothesis, which was developed to explain how 'hot spots' could be maintained over long periods of time [Morgan, 1971]. Mantle plumes were originally envisaged as diapirs, rising by virtue of their thermal buoyancy, from the core-mantle boundary. In this way, they tapped heat from an essentially inexhaustible reservoir, and could maintain high temperatures in 'hot spot' source regions for as long as necessary.

Plumes are considered to be localised heat- and melt-delivery structures, but high temperature is their most fundamental characteristic. The question whether 'hot spots' are really hot is thus of critical importance, because it amounts to a test of the mantle plume hypothesis. Since the turn of the 21st century, this has been seriously questioned [e.g., Foulger, 2002; Foulger, 2007; 2010]. This questioning arose from widespread realisation that geological and geophysical observations do not fit the predictions of the hypothesis, and it led to development of the alternative 'plate' hypothesis. This attributes volcanism to permissive leakage of pre-existing melt from the asthenosphere, in response to lithospheric extension consequential to plate tectonic processes. In the 'plate' hypothesis, variations in magma productivity are attributed to variations in source composition. Unusually high magma source temperatures are not required.

In order for a diapir to rise from the core-mantle boundary, through the mantle and to Earth's surface through thermal buoyancy, a temperature anomaly of at least 200 - 300˚C is required [Courtney & White, 1986; Sleep, 1990; Sleep, 2004]. The hottest part of a plume is predicted to be the centre of the head, and the tail immediately beneath. Most of the plume head is expected to be cooler, as it is predicted to entrain large amounts of normal-temperature ambient mantle as it rises and overturns. The temperature in a plume head is thus expected to reduce from a maximum at its centre over the tail, to approach ambient mantle temperature at its periphery [Campbell, 2006; Davies, 1999]. As a consequence, the highest temperatures are expected to be found near the centres of flood basalts and at currently active 'hot spots'.

The plate hypothesis predicts that the sources of melt erupted at the surface lie mostly within Earth's surface conductive layer, which comprises the shallowest ~ 100-200 km [Anderson, 2007; 2010; Foulger, 2010]. Within this layer, the potential temperature (T_P – Section 3.1) generally increases with depth. Melt drawn from deep within it will thus have a higher T_P than melt that forms at shallower levels (Figure 1). Some variation is thus expected in the source temperatures of melts erupted at 'hot spots', but this is expected to be largely related to depth of extraction. Lateral variation in mantle temperature is also expected. However, the predictions of the plate hypothesis contrast with those of the plume hypothesis in that the mantle is not considered to be essentially isothermal everywhere except for discrete, isolated, spot-like high-T_P anomalies of several hundred degrees Celsius. Unusually productive parts of the mid-ocean ridge system are expected to be sourced from similar depths and temperatures to those parts with average productivity. Exceptionally large melt production rates at some localities are attributed to a more fusible source composition, and corresponding differences in the geochemistry of the lavas is expected. Temperatures may be elevated by a few tens of degrees Celsius in some regions, where lithospheric structure encourages lateral flow from deeper within the conductive layer, *e.g.*, at plate boundary junctions such as ridge-transform intersections and triple junctions. Flood basalts and oceanic plateaus are not predicted to have a radial thermal structure.

The term ‘hot spot’ implies localised, high temperatures. Along with the plume hypothesis, it was exported from Hawaii and became widely applied to unusual volcanic regions all over the world. Unfortunately, the expression has traditionally been used with little questioning, for many localities where there is no evidence that it describes the phenomenon in question well, and even where there is *prima facie* evidence that it does not. This is now changing. Efforts are being made to test whether ‘hot spots’ are indeed hot, and to address the difficult task of developing reliable methods to do so. Although there is a surprisingly large number of ways to estimating temperature, and variations in temperature, in the mantle, the endeavour is plagued by fundamental technical and philosophical problems. As a consequence, testing whether ‘hot spots’ are hot is a surprisingly difficult task.

Casual and widespread use of the term ‘hot spot’ has been defended on the grounds that it may simply signify active surface volcanism. This cannot be justified because volcanoes at mid-ocean ridges and subduction zones would then also qualify as ‘hot spots’. What is potentially more problematic is that its use tends to encourage the presumption, without testing, that unusual volcanic regions arise from exceptionally hot mantle source rocks. For this reason, *a priori* use of the term is undesirable.

In addition, the term carries with it the concept that volcanism in particular regions may be attributed to some sort of specific, local entity (a ‘spot’) with an independent life of its own. This bolsters models of compact, localized melt delivery systems which are also, in general, unjustified. Such a view is particularly lacking in support in regions where simultaneous, ongoing volcanism is widespread over regions hundreds of kilometres broad, *e.g.*, Iceland, Yellowstone, the East African Rift, and Europe.

The term ‘hot spot’ is a neat and elegant term. However, it is flawed in that it is used to refer to volcanic regions in terms of an implied genesis process (mantle plumes), instead of reflecting uncontestable, observable characteristics. It has tended to ingrain assumptions and de-emphasise the necessity of testing assumptions scientifically. This has slowed progress in understanding unusual volcanic regions, and in explaining the many features which remain enigmas today.

The primary focus of this review is to summarise the methods for probing the source temperature of volcanic rocks, and the mantle, and to discuss critically the evidence for intrinsically high temperatures beneath regions commonly designated as ‘hot spots’. To begin with, however, some comments will be made on whether these regions may reasonably be referred to as ‘spots’.

2. Are ‘hot spots’ spots?

The concept that the melt delivery systems of intraplate- and unusually voluminous on-ridge volcanism are compact, spot-like entities is at odds with observations. It arose originally from speculations on the origin of Hawaii, where current, high-volume volcanism is centered on the Big Island. The concept is intrinsically linked to the plume hypothesis, which postulates that melt is supplied from the deep mantle through relatively narrow conduits [Morgan, 1971].

The widespread use of the term ‘hot spot’, along with assumption of the plume model, has resulted in views such as:

- attribution of the volcanism to an independent, self-sustaining entity that is compact, coherent, and exists in its own right;

- the existence of a relatively compact, primary point of melt delivery, possibly identifiable with a single volcano no more than a few kilometres or tens of kilometres across, from which any more wide-ranging volcanism is derived by lateral flow;

- 'hot spot tracks', 'hot spot migrations', current and past 'hot spot locations', and 'hot spot centres'.

Examination of the facts at almost any volcanic region shows that this does not fit the observations well. At Hawaii, recent volcanic activity has been distributed along 350 km of the chain, extending from the Big Island back as far as the island of Koolau to the northwest [Clague & Dalrymple, 1987]. The Pacific plate is moving at ~ 10.4 cm/a. Thus, if Koolau volcanoes are sourced from a fixed 'hot spot' now centred beneath Loihi, a submarine volcano just a few kilometres wide, lateral flow of hundreds of kilometres, maintaining volcanoes for up to ~ 3.5 Ma after they have migrated away, is required.

At Iceland, despite the widespread assumption that the "hot spot" is centred on the 8-km-wide Grimsvötn volcano, post-glacial volcanism (0 - 10,000 a) is spread over ~ 400 km normal to the spreading direction [Johannesson & Saemundsson, 1998]. Elsewhere, many linear chains of volcanoes are simultaneously active along much of the chain length, *e.g.*, the Cameroon line [Fitton, 1987] and the Austral-Cook islands [Davis *et al.*, 2002]. At the Galapagos, the entire ~ 1,000-km-long aseismic Cocos Ridge, and 600 km of the Carnegie Ridge, of which the Galapagos islands form a part, have been volcanically active in the last ~ 3 Ma [O'Connor *et al.*, 2007].

In continental areas, volcanism is commonly scattered over regions thousands of kilometres broad, *e.g.*, in Europe [Abratis *et al.*, 2007; Lustrino & Wilson, 2007], Africa [Bailey, 1992], western North America and Asia. Volcanic systems that have been built by magmatism that was only ever active at a single site, and migrated in a uniform direction, are rare. Having said this, many are too poorly dated for the results to be definitive, *e.g.*, the Tristan-Walvis Ridge system [Baksi, 1999; 2005; O'Connor *et al.*, 1999] and the Lakshadweep-Chagos-Reunion system [Baksi, 1999; 2005]. The best dated examples from long, linear, oceanic volcano chains are from the Ninetyeast Ridge [Pringle *et al.*, 2008], the Louisville seamount chain [Clouard & Bonneville, 2005] and the Emperor and Hawaiian chains [Sharp & Clague, 2006].

The picture that emerges from most currently active volcanic regions styled as 'hot spots' is one of volcanism distributed over broad areas, and not produced from the local 'spots' generally assumed. In those cases, expectations of coherent, simple, narrow, time-progressive volcano chains, and singular sources, are assumption-driven and not based on the observations.

3. Are 'hot spots' hot?

3.1 Temperature inside Earth

The term 'hot spot' embodies the concept that volcanic islands such as Hawaii form over regions in the mantle that are hotter than surrounding mantle at the same depth. Testing for high temperature is not straightforward, however. Not only must the temperatures of the mantle sources of lavas be measured, but also a reference value for Earth against which to compare them is needed. These are both surprisingly difficult to determine. The problem is further exacerbated by the fact that the absence of evidence for high temperature is not evidence for its absence, and the widespread use of the term 'hot spot' then tends to make high-temperature the *a priori* assumption that is retained in the absence of evidence for it.

Temperature is not laterally invariant in Earth. Heat is transported around inside Earth by advection, conduction and radiation. Conduction and radiative heat transfer proceed only slowly [Hofmeister & Criss, 2005; Toksöz & Kehrer, 1972]. By far the most efficient method is by the advection of fluids, including melt percolation, intrusion, eruption, hydrothermal circulation and degassing. Solid-state mantle convection also redistributes heat by advection. Convection involves both warm risers and cold sinkers, including subducting slabs and material delaminated from the underside of the lithosphere. As a result of all these processes, combined mantle temperature varies both vertically and laterally.

There are two major thermal boundary layers in Earth, *i.e.*, layers across which the temperature rises rapidly. These are at the surface, and the core-mantle boundary. A rise in temperature of approximately 1,500˚C occurs between Earth's surface and the mantle across a depth interval of ~ 100-200 km (Figure 1). This region is more accurately referred to as the 'conductive layer'. The temperature increase occurs because heat is lost to space, cooling the surface. In the continents, the rate of heat loss is ~ 50-100 mW/m^2, with a median of ~ 60 mW/m^2 [Anderson, 2007]. In the oceans, it varies in the range ~ 25-300 mW/m^2, with a median value about the same as for continents [Anderson, 2007]. Heat flow decreases with the age of the ocean floor up to ~ 50 Ma, but is approximately constant for older crust.

A commonly assumed model treats the oceanic lithosphere as a cooling plate underlain by asthenosphere maintained at a constant temperature by convection and radioactivity [Parsons & Sclater, 1977; Stein & Stein, 1992]. Such a model predicts that heat flow dies off as the square root of lithospheric age. It fits the bathymetric observations but not heatflow (Figure 2).

The largest generator of heat in Earth is radioactivity. The main contributors are the isotopes ^{40}K, ^{238}U, ^{235}U and ^{232}Th, much of which resides in the continental crust. Another source of heat in the lithosphere is strain from tectonic processes [Anderson, 2010]. A recent study of the thermal diffusivity and specific heat capacity of crustal materials suggests positive feedback between strain heating in shear zones and thermal insulation. This removes the requirement for unusually high radiogenic heat production to achieve crustal melting temperatures, and suggests that partial melting may occur in both the crust and the mantle in many tectonic settings [Whittington *et al.*, 2009]. There is little radioactive material in oceanic crust or in the mantle.

Approximately 18% of the heat lost at the surface comes from core via the core-mantle boundary, Earth's lower thermal boundary layer [Davies, 1988; Hofmeister, 2007; Sleep, 1990]. There, T_P increases by ~ 1,000˚C across a depth interval of ~ 200 km, from ~ 2,000˚C in the lowermost mantle to ~3,000˚C in the outer core. The surface area of the core is only ~40% that of Earth's surface. Thus, if the total heat flux from the core is ~ 18% that from the surface, the heat flux per unit area from the core is ~ 72% of that from the surface.

Knowledge of the source depth of lavas, and of the variation in temperature with depth in Earth is necessary in order to correctly interpret temperature estimates. Temperature increases with depth at ~ 20˚C/km in tectonically stable continental interiors, ~ 50˚C/km in rift valleys and ~ 100˚C/km on volcanically active oceanic islands, *e.g.*, Iceland. Locally, at volcanoes and geothermal areas, it may rise even faster. At great depth, progressive compression of minerals with increasing pressure – adiabatic compression – causes absolute temperature to increase with depth, regardless of other effects. The adiabatic temperature gradient (the 'mantle adiabat') in the shallow mantle is ~ 0.4˚C/km, and the total increase in absolute temperature throughout the mantle is ~ 500-900˚C (Figure 3). The adiabatic gradient changes abruptly where mineralogical phase changes occur in the transition-zone, and it will change if the amount of radioactive elements varies. Interestingly, because of the negative Clapeyron slope of the mineraological phase change in olivine at the 650-km discontinuity, temperature drops across that boundary (Figure 3) [Katsura *et al.*, 2004].

The adiabatic temperature increase in Earth is of little interest because if material rises and its confining pressure decreases, its absolute temperature simply reduces adiabatically. Adiabatic temperature variations are thus typically ignored, and the temperatures in Earth's mantle are expressed as potential temperature (T_P). This is the temperature that material would have at Earth's surface if it rose adiabatically (*i.e.*, without loss or gain of heat), without change of state or phase, from its original depth [McKenzie & Bickle, 1988]. If temperature in the mantle were purely adiabatic and laterally invariant, T_P would be the same everywhere. However, this cannot be so because dynamic geological processes such as subduction occur, and the mantle is radioactively inhomogeneous. Furthermore, the mantle must have variable T_P in order to convect.

T_P at the base of the surface conductive layer is not accurately known, neither its average value, nor its value in key local areas. It is commonly assumed to be the same as the temperature of the source of mid-ocean-ridge basalts (MORB). However, there is no observational evidence to support this, and it is self evident that this cannot be true everywhere. For example, in the neighborhood of subducting slabs, lateral variations in T_P of hundreds of degrees Celsius occur. Melt extraction at volcanoes, thermal disruption of the shallow mantle by continental breakup, lithosphere delamination, and mantle convection also cause T_P to vary.

It is self-evident that melt exists in the interior of Earth, in both the crust and mantle, since it erupts periodically onto the surface. The depth of formation of this melt must be known if calculated source temperatures are to be interpreted with confidence.

The mere presence of melt, in particular in large volumes, is often taken as evidence for unusually high temperature. This is not safe, as different lithologies melt at different temperatures. The mantle is conventionally assumed to be homogeneous and made of peridotite. This cannot be correct, however, because numerous processes related to plate tectonics remove melt from it (*e.g.*, at spreading ridges), reinject fusible near-surface materials (*e.g.*, at subduction zones) and differentiate and transport material within it. Other petrologies that must exist in the mantle include eclogite, which forms when basaltic oceanic crust subducts to depths greater than ~ 60 km. The solidus and liquidus of eclogite are typically ~ 200˚C lower than those of peridotite, and under some conditions eclogite may be completely molten at a temperature lower than the solidus of peridotite (Figure 4) [Cordery *et al.*, 1997].

Volatiles, including water and carbon dioxide, lower the solidus of mantle rocks by up to several hundred degrees Celsius (Figure 5) [Green *et al.*, 2010; Hall, 1996]. The effect of water is clearly illustrated at subduction zones, where arc volcanism arises from the fluxing of mantle rocks by hydrous fluid rising from the down-going slab. The effect of CO_2 on mantle rocks has been studied in the laboratory using chemical assemblages that reproduce the major-element compositions of fertile peridotites (*e.g.*, the CaO-MgO-Al_2O_3-SiO_2-Na_2O-FeO-CO_2 (CMASNF-CO_2) system) (Figure 6). Experiments have been conducted at pressures up to ~ 8 GPa, corresponding to a depth of ~ 250 km. At depths greater than ~ 60 km, the solidus of peridotite drops from ~1,300˚C to ~ 1,000˚C in the presence of CO_2. It thus probably falls below the ambient mantle temperature at these depths. The partial melt thereby formed may be responsible for the seismic low-velocity zone, which is typically ~ 100 km thick and widespread beneath the oceans. In it, V_S is reduced by ~ 5-10% and anisotropy is at a maximum [Presnall & Gudfinnsson, 2008].

The Earth is continually out-gassing CO_2, and this must induce continual replacement by upward flow from depth. It is transported through the mantle by the upward percolation of CO_2-rich melts in the form of carbonate-rich (hydrous-silicate) melts. These melts rise by virtue of their low density and viscosity, and form carbonated, partially molten, CO_2-rich regions of the mantle that are expected to appear in seismic tomography images as low-velocity regions. They must have a major effect on the transport of trace elements, on the rheology and seismic properties of minerals, and on mantle structure and convection [Dasgupta & Dixon, 2009]. Recently, the H_2O-storage capacity of the mineral pargasite has been found to be strongly depth-dependent and it may control the depth of the lithosphere-asthenosphere boundary by inhibiting volatile-related melting at pressures less than ~ 3 GPa (depths shallower than ~ 100 km) [Green & Falloon, 2010].

Carbonatite lavas that erupt at the surface are unusual in containing almost no SiO_2 and comprising over 50% carbonate minerals. The only currently active volcano that produces such lavas is Oldoinyo Lengai, Tanzania. The volcanic gases erupted simultaneously are indistinguishable from those emitted at mid-ocean ridges, despite the continental setting of Oldoinyo Lengai. This suggests that a globally homogeneous reservoir of volatiles exists in the upper mantle, that supplies volatiles to volcanism everywhere [Fischer *et al.*, 2009].

Massive and continuous outgassing of CO_2 is an almost ubiquitous feature of ‘hot spots’, even during periods of inactivity. There is also abundant geochemical evidence for water and carbonatite contributions to the melt sources, *e.g.*, at Hawaii [Huang *et al.*, 2009; Sisson *et al.*, 2009], and the western Mediterranean [Frezzotti *et al.*, 2009]. Concentrations of CO_2 of up to 6% have been determined for some Hawaiian magmas [Barsanti *et al.*, 2009]. It has been suggested that ‘hot spots’ might, instead, be ‘wet spots’, and owe their excess magmatism solely to high volatile contents [Bonatti, 1990].

3.2 Methods of measuring the temperature of mantle source rocks

There are a surprisingly large number of ways of estimating temperature inside Earth. However, achieving the accuracy required to determine if a significant temperature anomaly exists at a locality, is problematic. Absolute temperature is difficult to estimate because it is dependent on many poorly known factors. These include mantle composition, volatile content, and the physical properties of common rock-forming minerals at conditions corresponding to mantle depths in Earth. Absolute temperature is not needed to test the hypothesis that ‘hot spots’ arise from anomalously hot sources, however. This can be done simply by measuring the variation in source temperature between ‘hot spot’ lavas and others presumed to come from ‘normal temperature’ mantle at the same depth. Nevertheless, this is still a vexed problem.

First, at what depth did the melt form? The surface conductive layer – the layer throughout which T_P progressively increases with depth – may be up to ~ 200 km thick. T_P increases strongly with depth throughout this layer, and if temperature estimates for different regions correspond to different depths, large differences are expected that do not reflect lateral variations in temperature. The ‘normal’ temperature of the mantle, against which all else is compared, is commonly taken to be that of the source region of MORBs. These are generally assumed to be extracted from the top few tens of kilometres of the mantle. However, the depth of extraction of melt at ‘hot spots’ is almost always poorly known, and may be deeper over the thicker lithosphere away from mid-ocean ridges.

3.2.1 Seismology

3.2.1.1 The depth to the base of the seismogenic layer

The maximum depth of earthquakes is temperature dependent because where the rocks are hot enough to flow, earthquakes do not occur. By mapping the depth at which earthquakes cease to occur, knowing the petrology, volatile content, and strain rate of a region, the temperature can be estimated [*e.g.*, Foulger, 1995; Hill, 1989].

3.2.1.2 The thickness of igneous crust

The thickness of igneous crust (*i.e.*, oceanic crust, or basalt flow sequences in continental areas) is potentially a proxy for the amount of melt formed, and can be measured using seismology. A larger melt volume, and thicker igneous crust, is expected for higher-temperature melting. However, the same result could arise at normal temperatures in the case of a more fusible source composition. The two can be distinguished by combined analysis of seismic velocity (V_P) and crustal thickness. Data of this kind can be acquired from high-fidelity explosion seismology. Igneous crust formed at higher temperatures will not only be thicker, but also richer in olivine and MgO and as a result have higher V_P. Crustal thickness and V_P will thus be correlated for a hot source (Figure 7) [Korenaga & Kelemen, 2000; Korenaga *et al.*, 2002]. For a source at normal temperature, enhanced melt volume is expected to be related to source fusibility, not temperature, and thus a negative correlation, or no correlation at all, between crustal thickness and V_P is predicted.

3.2.1.3 Seismic velocity

The most common seismic parameter used to infer temperature is wave speed, usually referred to as velocity. Low velocities in either V_P or V_S, in particular in seismic tomography images, are commonly assumed to indicate high temperature. In the case of teleseismic tomography experiments, the reference V_P and V_S values that are assumed to represent mantle with 'normal' temperature, are typically the average V_P or V_S within the study volume. For an active plume, a hot tail extending from the surface down to the core-mantle boundary is predicted, and this is expected to be characterised by low V_P and V_S. For a plume stem ~ 300˚C hotter than ambient mantle, reductions in V_P and V_S of ~ 2-3% and 3-4.5% respectively are expected in the upper mantle, reducing to ~ 0.5-0.75% and 1-1.5% respectively in the deep mantle (Figure 8) [Julian, 2005; Karato, 1993]. For melting anomalies sourced by normal-temperature mantle, low-velocity seismic anomalies beneath currently active melting anomalies are predicted to be essentially confined to the upper mantle and to arise from enhanced melt content. Quasi-cylindrical, quasi-vertical, low-velocity structures extending from the surface down to the core-mantle boundary are not expected.

Unfortunately, although high temperatures usually (but not always) reduce seismic velocities, low seismic velocities cannot be assumed to indicate high temperature because velocity is dependant on other effects. These include petrology, composition, mineralogical phase, and physical state (liquid or solid). Of all influences, composition has the strongest effect and temperature the weakest [Table 1; [e.g., Foulger, 2007; 2010]. Laboratory elasticity data show, for example, that a decrease of 1% in the Mg/(Mg+Fe) ratio in olivine reduces the velocity by an amount equivalent to a temperature increase of 70 K [Chen *et al.*, 1996].

It is not usually possible to separate out these several effects with measurements of only V_P or V_S, or both. In addition, where V_P and V_S have been measured in different experiments, the errors may be too large to meaningfully interpret them jointly to investigate temperature. For example, if V_P is mapped using explosion seismology, and V_S using surface waves or receiver functions from earthquakes, Earth will have been sampled on different spatial scales. An additional problem that hinders the application of seismic velocity to infer temperature in the mantle is that the progressive increase in pressure with depth in the mantle reduces the sensitivity of velocity to temperature by a factor of ~ 3 from the top to the bottom (Figure 8).

The ratio V_P/V_S is sensitive to the presence of partial melt, which lowers V_S more than V_P. High V_P/V_S ratios are thus expected for volumes containing partial melt. Correlations between V_P/V_S and Mg# in peridotites are also commonly assumed, but detailed testing of this shows that it is valid only for a restricted range of mineral assemblages and temperatures. For most of the upper mantle the variations in V_P/V_S predicted for compositional variations are too small to be resolved using current seismological methods. In general, compositional anomalies in the mantle cannot be separated from thermal anomalies on the basis of seismological studies only [Afonso *et al.*, 2010; Foulger, 2010].

The case of the 'superplumes' is a rare exception that also illustrates the ambiguity problem. The 'superplumes' are vast, low-seismic-velocity bodies in the lower mantle that underlie the eastern Pacific Ocean, and the South Atlantic. They were originally assumed to be high-temperature bodies (and were thus named 'superplumes'), and proposed to be the source of heat responsible for volcanism in the east Pacific and in Africa. However, seismic evidence from normal modes indicates that these bodies actually have positive anomalies in the elastic bulk modulus. Because temperature affects bulk modulus and V_S similarly, the combination of positive bulk modulus anomalies and negative-V_S anomalies is not consistent with high temperature alone. The observations require that they are dense, not buoyant, and caused primarily by chemical heterogeneity [Brodholt *et al.*, 2007; Trampert *et al.*, 2004]. It is unfortunate that the misnomer 'superplume' has now become entrenched in geophysical literature, and continues to reinforce an incorrect assumption that they are heat sources [Garnero *et al.*, 2007].

3.2.1.4 Seismic attenuation

Seismic attenuation increases with temperature [Anderson, 2007] but again, temperature is not the only factor that influences it. Partial melt and composition are also important. Calculating attenuation from seismic data can reduce interpretive ambiguity inherent in seismic velocity data alone. For example, at the Ontong Java plateau, a thermal interpretation for a low-seismic velocity body in the underlying mantle could be ruled out because the body was also found to have low attenuation. This combination of seismological characteristics indicates a high-viscosity body that owes its low seismic velocities to anomalous composition, not high temperature [Klosko *et al.*, 2001].

3.2.1.5 Thickness of the mantle transition zone

The mantle transition zone is the region between the major seismic velocity discontinuities at ~ 410- and ~ 650-km depths. Their exact depths have been used widely in attempts to map temperature variations at these levels. Elevated temperature is predicted to deepen the 410-km discontinuity from its global, average value by ~ 8 km/100˚C. For a plume with a 200-300˚C temperature anomaly, a downward deflection of ~ 15-25 km is expected. The effect of temperature on the 650-km discontinuity is more complex. At that depth, the effect of elevated temperature on olivine is expected to deepen the discontinuity, but the effect on garnet minerals is the opposite. These effects may partially or wholly cancel out. Topography on the 650-km discontinuity is thus currently too poorly understood to be used for temperature estimations.

Complicating the problem is the fact that, in addition to temperature, lateral variations in composition, volatiles, and phase may also cause deflections on the transition-zone-bounding discontinuities (Figure 9).

Using variation in the depth of the 410-km discontinuity to infer temperature is further hindered by the difficulty in making sufficiently accurate measurements. Many tens of teleseismic records must be stacked to reduce formal errors to as little as ~ 5 km, and repeatability of the results between independent experiments is often poor. An error of 5 km is large compared with the signals being sought, which might amount to no more than a 10-15 km deflection.

More problematic is the ambiguity in interpretation. In addition to high temperature, high Mg (*i.e.*, a refractory composition), and low water contents also have the effect of deepening the 410-km discontinuity (Figure 9) (Figure 10). Conversely, high Fe (*i.e.*, a fertile composition), and hydrous conditions, in addition to low temperatures, can elevate the 410-km discontinuity. Most likely, topography on the discontinuity results from a combination of all three effects.

3.2.2 Petrological and geochemical methods

There are numerous petrological approaches for estimating the temperature of the mantle source of igneous rocks. These include methods for estimating both absolute temperature and regional temperature variations. In the case of the latter, a 'normal' or 'average' source temperature is needed against which other estimates may be compared. This is usually taken to be the source temperature of MORB. This reasoning rests on the perception that the global spreading ridge system is a mantle-sampling zone that is both widespread and relatively uniform in process. Temperature estimates from 'hot spots' are commonly compared with 'the MORB average'. Nevertheless, a problem with this is immediately obvious. It cannot necessarily be assumed that all 'hot spot' lavas arise from the same depths in the mantle as MORB.

3.2.2.1 The Global Systematics

The approach to temperature estimation known as 'the Global Systematics' was based on the belief that Na_8 and Fe_8 (the Na_2O and FeO contents of basalts, inferred for a reference value of 8% MgO) in basalts from mid-ocean ridges correlate with bathymetric depth of eruption. This was explained using a model whereby higher source temperatures in an adiabatically upwelling mantle result in melting beginning deeper and at higher pressures, leading to a higher extent of melting, higher Fe_8, lower Na_8, a larger melt volume, thicker crust and shallower bathymetry [Klein & Langmuir, 1987; Langmuir *et al.*, 1992] (Figure 11).

The Na_8 contents in MORB were calibrated against temperature using localities where crustal thickness had been measured seismically, assuming that the crust formed from adiabatic upwelling decompression melting at ridges. It was concluded that melting occurred in the depth range ~ 40-130 km (corresponding to pressures of ~ 1.3-4.3 GPa) and the potential temperature beneath ridges varied in the range ~ 1260-1590°C. The model assumes that the composition of the mantle source of lavas is uniform, including both 'normal' ridges and 'hot spots'. This is despite the fact that a deep-mantle plume origin for lavas at 'hot spots' is largely based on their distinct chemistry, and that this chemistry is attributed to a component of near-surface, fusible material [Hofmann & White, 1982].

The 'Global Systematics' have now been shown by observation to be invalid. The larger set of data on MORB compositions that has accumulated since the theory was originally proposed shows that the postulated correlations between Na_2O, FeO and axial depth along ridges do not, in fact, occur (Section 3.3) [Niu & O'Hara, 2008; Presnall & Gudfinnsson, 2008]. Nevertheless, the 'Global Systematics' still continue to be assumed to be correct and applied [*e.g.*, Singh *et al.*, 2010].

3.2.2.2 Mineralogical phase relationships

The mineralogical phase boundaries in mantle minerals have been explored in the laboratory over a wide range of pressures and temperatures. The clearest results have been obtained using chemical assemblages that replicate the compositions expected for mantle rocks. The results obtained for the six-component CaO-MgO-Al_2O_3-SiO_2-Na_2O-FeO (CMASNF) system in the pressure range ~ 0.9-1.5 GPa (~ 25-45 km depth) simulate the behaviour expected at the natural plagioclase/spinel lherzolite transition where most MORBs are thought to be generated by partial melting. Comparing the compositions of the melts produced in these experiments with those of natural samples has revealed the systematics of MORB generation under variable temperature, pressure and source-composition conditions [Presnall *et al.*, 2002]. Using these results, the relative importances of temperature and compositional heterogeneity in producing the compositional variations seen in MORBs have been determined. The results suggest that the chemical variations observed owe much more to source compositional variations than to variations in temperature. Since many postulated 'hot spots' lie along ridges, this casts doubt on whether the mantle at those locations is any hotter than elsewhere beneath the spreading ridge system.

3.2.2.3 Olivine control lines

When melts that are high in Mg cool, the first crystalline phase to form is olivine. The first olivine to crystallize is the highest in Mg (*i.e.,* the most forsteritic) and the Mg content of subsequent crystals progressively decreases as the temperature of the melt reduces. As this proceeds, the Mg content of the remaining liquid progressively reduces. This evolution in composition is known as an olivine-controlled liquid line of descent. In order to detect confidently olivine-controlled crystallization, melts sufficiently high in Mg to classify as picrites (*i.e.*, with MgO > 12%) are required.

If samples of the instantaneous melts formed during olivine-controlled crystallization are analyzed, they can be used to estimate the temperature of the original parent liquid. This is done by mathematically modelling the olivine crystal fractionation process (Figure 12). In order to do this, an estimate of the composition of the first olivine phenocryst is needed. The most magnesian olivine sampled, or the most magnesian olivine commonly observed in the study area, is typically used. In order to deduce the temperature at which the calculated melt composition began to crystallize, an 'olivine geothermometer' is needed. These are derived empirically from laboratory measurements of the temperatures at which rock samples are observed to melt [Falloon *et al.*, 2007b].

The method is dependent on many assumptions and a large range of temperatures may be obtained for a particular locality, using different samples, assumptions, and/or olivine geothermometers. Variations of up to several hundred degrees Celsius may be obtained. If this is the case, the method clearly has little ability to test whether 'hot spot' lavas arise from an unusually hot source. The uncertainty in the results is of the order of the total temperature anomaly expected.

The reasons for these large variations include:

- It must be assumed that the samples used to construct the olivine-controlled liquid line of descent were once liquids. It is only safe to assume this if the samples are glass, because crystalline rocks may comprise composite assemblages of crystals accumulated from several different sources [Natland, 2010; Presnall *et al.*, 2009]. Unfortunately, picritic glass is virtually never observed. Crystalline samples have thus been used in several recent studies, simply because glass is not available.

  The danger in using cumulate rocks has been confirmed. Detailed study of picrite cumulates, *e.g.*, from Hawaii [Clague *et al.*, 1995], shows that the crystals in picrite cumulates have been scavenged from several different magmas or crystal mushes. Furthermore, the mixing of a partially differentiated mush with a more primitive melt will increase the FeO content for a given MgO content. This will result in erroneously high apparent temperatures from olivine-control-line analysis. Simply put, olivine-control-line analysis results can only be trusted if it is based on picritic glass samples, and if cumulate rocks are used, temperatures are expected to be incorrectly high;

- If the cumulate starting samples contain plagioclase and pyroxene crystals, the mathematical olivine crystal addition method is invalid. The presence of plagioclase and pyroxene crystals is evidence anyway that the sample does not form part of an olivine-crystallization-only array. Plagioclase and pyroxene crystallization can be corrected for, but this introduces further uncertainties into the analysis [Falloon *et al.*, 2007a];

- The forsterite composition that should be used for the target phenocryst cannot be known with certainty. The higher the forsterite content chosen, the higher the temperature obtained. There are various ways of making the choice, *e.g.*, the most-forsteritic crystal, or the most common value, but no way of knowing which, if any, is correct;

- Different olivine geothermometers have been used in different analyses [e.g., Beattie, 1993; Falloon *et al.*, 2007b; Ford *et al.*, 1983; Helz & Thornber, 1987; Herzberg & O'Hara, 2002; Putirka, 2005b]. These may use different partition coefficients, which describe how chemical species partition between the solid and liquid phases. They may also model the olivine liquidus temperature and crystal composition as a function of different subsets of the controlling variables, which include melt composition, pressure, and temperature. The use of different geothermometers is thus a significant source of variation in the resulting source temperature calculated. Tests performed by comparing their predicted temperatures with those measured by melting in the laboratory show that differences of > 200˚C can be obtained, starting from the same petrological observations (Figure 13) [Falloon *et al.*, 2007b];

- The conclusion regarding whether studied samples indicate a source temperature anomaly depends on the choice of 'normal' temperature with which to compare the results. This temperature is usually what is considered to be typical for MORB source. This is, nevertheless, problematic to decide and the estimates for different parts of the global ridge system are often large. Some investigators have taken the maximum temperature calculated for MORB, whereas others have rejected the higher temperatures, attributing them to near-ridge 'hot spots'. The latter approach assumes a bimodal temperature distribution in the mantle, thereby locking into the experiment, *a priori*, an inevitable outcome. Thus, even if there is agreement about the source temperature of lavas from a 'hot spot', if different mid-ocean-ridge reference temperatures are assumed, different conclusions may be reached.

3.2.2.4 Rare-earth-element modelling

The atomic radii of the rare-earth elements – the Lanthanide-Lutetium (La - Lu) series – decrease systematically as atomic number increases. The smaller atoms are more easily retained in crystal lattices, and as a result, the elements with larger atomic radii systematically partition more readily into the liquid during partial melting (Figure 14).

The pattern of rare-earth elements observed in a liquid varies according to which minerals melted and to what degree. The stability fields of plagioclase, spinel and garnet are depth dependent, and the rare-earth element content can thus be used to estimate melting degree and which minerals melted. From this, the depth of onset of melting may be deduced. The temperature of the source can be deduced if it is assumed that the melt arises from isentropic upwelling, [McKenzie & O'nions, 1991].

The method involves several assumptions, including:

- The chemistry and mineralogy of the source. Usually a peridotite mineralogy and a composition similar to primitive mantle are assumed;
- The sources of melts from different regions are assumed to have the same compositions;
- All the samples are assumed to represent single melts that formed by fractional melting, not batch melts, nor aggregates from more than one melt.

If any of these assumptions are wrong, the differences between the temperatures calculated will also be wrong. Since the question at issue is commonly whether variations in the amount of melting results from variations in temperature or composition, rare-earth element modelling is intrinsically unsuited to resolving the debate.

3.2.2.5 Komatiites

Komatiites are generally held to have been produced at unusually high temperatures, perhaps exceeding 1,600˚C, compared with ~1,300˚C for MORBs. Komatiites are basaltic rocks that contain 18-30 wt.% MgO, and are thought to have come from parental melts that had similarly high MgO contents. This is higher than the 12-18% MgO which defines picrites, and much higher than the 8-10% that is found in most MORBs [Le Bas, 2000]. Komatiites are widespread in Archean rocks. They are rare in younger rocks, however, and the only known examples from the Phanerozoic occur on Gorgona Island, in the Pacific ocean [Echeverría, 1980; Kerr, 2005]. They have been variously attributed to laterally migrating plumes postulated to underlie currently either the Galapagos or the Caribbean. Recent, careful age-dating work shows, however, that magmatism at Gorgona spanned the ~ 38 Ma period ~ 99-61 Ma [Serrano *et al.*, 2011b]. Also, the whole Caribbean igneous province is characterised by a long period of diffuse magmatism without an obvious spatial pattern of migration [Serrano *et al.*, 2011a].

Another environment in which komatiites can be produced is at low pressures in subduction zones under hydrous melting conditions [Parman *et al.*, 2001]. Evidence for this is the high SiO_2 contents of some komatiites, which suggests melting at low pressures, and also the presence of hydrous minerals such as amphibole and high concentrations of water in melt inclusions. Such komatiites overlap compositionally with boninites, which are produced in island arcs above subduction zones. Boninites have similarly high SiO_2 at high MgO contents, which is attributed to their arising from high-degree melts that formed at shallow depths. Large degrees of melting are possible at shallow depth beneath volcanic arcs because dehydration of the subducting slab fluxes the mantle wedge above with abundant water [Crawford *et al.*, 1989]. These two possible interpretations of komatiites – whether they indicate very high-temperature magmas or whether they are simply the product of low-pressure, water-fluxed melting beneath subduction zones, is currently a subject of debate [Parman & Grove, 2005; Parman *et al.*, 2001; Puchtel *et al.*, 2009].

3.2.3 Ocean floor bathymetry

The depth of the ocean floor increases with age. This is attributed to the cooling of the lithosphere with time, following creation of the crustal part at the mid-ocean ridge. This simple model implies that anomalously shallow sea floor can be explained by an anomalously hot lithosphere. Using this model, ocean-floor depth has been widely interpreted in terms of temperature variations.

Two approaches to this kind of analysis have been used:

1. The variations in present-day bathymetric depth to the sea floor have been interpreted directly as indicating regional variations in lithosphere temperature. For example, conductive cooling models have been used to account for anomalies such as the Hawaiian swell [Ito & Clift, 1998; Sleep, 1987].

2. Temporal variations in depth have been studied. The histories of sea-floor subsidence at individual localities have been compared with theoretical "normal" subsidence curves to deduce the original temperature of formation [Clift, 1997]. If the lithosphere is postulated to have formed over asthenosphere that was hotter than normal, subsequent anomalously rapid subsidence is predicted as the region is transported over normal-temperature asthenosphere by plate motion.

   Subsidence curves for regions postulated to have been emplaced anomalously hot have been estimated from the sedimentary sections of marine drilling cores (Figure 15). Application of the method to oceanic regions is straightforward compared with continental regions because the history of vertical motion for the latter is much more complicated to assess because the effects of tectonism and erosion are difficult to estimate. In contrast, subsidence of the ocean floor is relatively simple. This method has been used to study oceanic plateaus, which are postulated to represent the eruptions accompanying the arrival of hot plume heads. This model predicts that they subside anomalously rapidly following their formation. If, on the other hand, they arise from the melting of unusually fusible source material at normal asthenopshere temperatures, they are expected to have subsided at a normal rate as they drifted away from their original locations.

A significant problem with using bathymetry to estimate temperature is the difficulty of making accurate estimates of the variation in ocean depth with time. Isostatic adjustments in response to progressive sediment loading must be corrected for. Water depth at the time of deposition is estimated from the type of sediment and fossils in the cores. Benthic foraminifera are particularly sensitive to depositional environment. Water depth estimation is most reliable for shallow water (< 200 m), but errors may be as large as ~ 1,000 m for greater depths. Variations in sea-level introduce an additional uncertainty of ~ 100 m that cannot be corrected for because the timing and magnitude of these variations are currently insufficiently well known.

The method makes the simplistic assumption that all anomalous bathymetric depth estimates can be attributed to temperature. This is known to be not the case, however. Sources of compositional buoyancy include the light residuum that remains following melt extraction, and this may even be able to account for the entire Hawaiian swell [Section 2.3, 2.5.3; Leahy *et al.*, 2010; Phipps-Morgan *et al.*, 1995]. Shallow bathymetry at currently active melting anomalies can thus potentially be explained either by anomalously hot asthenosphere, or by the residuum left following melt extraction.

3.2.4 Heat flow

Surface heat flow can indicate regional variations in crustal and lithospheric temperature. It is measured both on continents and in the oceans. Oceanic crust is more uniform and simpler to understand than continental crust. Most of the world's postulated 'hot spots' lie in oceanic regions, and thus heat-flow measurements from oceans have been analyzed most extensively in that context.

Currently active plume stems are expected to reheat the lower lithosphere to asthenospheric temperatures over a circular region several hundred kilometres in diameter [von Herzen *et al.*, 1982]. Where a plume underlies a thick, rapidly moving plate, no measurable local surface heat-flow anomaly is expected (except that resulting directly from the advection of melt to the surface) because a substantial time is required for the heat to conduct through the lithosphere to the surface. The maximum heat-flow anomaly is predicted to occur some distance along the volcano chain where sufficient time has elapsed for the heat to conduct to the surface (Figure 16). In the case of 'hot spots' produced by extraction of mantle melts at normal temperatures, heat flow is expected to be normal for lithosphere of that age.

In order to obtain consistent, reliable measurements, sites must be chosen carefully. The best sites are relatively flat, minimise local effects due to hydrothermal circulation, and have a thick (>100 m) cover of uniform sediment. Final values for heat flow are calculated by combining several tens of measurements in an area, thereby reducing the error by a factor of approximately the square root of the number of measurements. The difficulty of measuring thermal conductivity, which may well be anisotropic, adds to the error budget.

Measuring surface heat flow is the most direct way of testing for high mantle temperature. Nevertheless, there are significant challenges to deducing mantle temperature from the results. As with other methods, individual measurements must be compared with some global average for lithosphere of the same age. For this, a reliable global model is needed. Many older interpretations of oceanic heat flow were flawed because older global models underestimated average heat flow [Stein & Stein, 1992]. Also, it must be assumed that the ocean bottom temperature has remained constant for 10^4 - 10^5 years, which may not be correct, in particular in volcanically active areas.

Local geological conditions may complicate heat flow. Recent magmatism will increase it, and hydrothermal circulation may reduce it by transporting heat away [Harris *et al.*, 2000]. Site-specific assessments are necessary to assess whether volcanism or hydrothermal circulation affects individual sites [Stein & Von Herzen, 2007].

Most serious is the problem that heat-flow measurements are relatively insensitive to mantle temperature, particularly beneath thick lithosphere. In addition, there may be a long delay between a change in asthenosphere temperature and a measurable effect arriving at the surface. For asthenosphere several hundred degrees Celsius hotter than normal, and lithosphere several tens of kilometres thick, it takes ~ 100 Ma for a surface heat-flow anomaly of a few mW/m^2 to develop. Such an anomaly is at the lower limit of what is measurable, even for the highest-quality heat-flow measurements. Numerical modeling also suggests that small-scale convection could take place without noticeably disturbing surface heat flux [Korenaga, 2009].

3.3 Observations

3.3.1 Mid-ocean ridges

The early pioneering work of McKenzie and Bickle [1988] on mantle T_P beneath mid-ocean ridges assumed that melt is formed by isentropic upwelling of a standard peridotite lithology. This study assumed that crustal thickness was a direct measurement of melt thickness, and a direct proxy for average mantle T_P. With these assumptions, a T_P of 1,280˚C was deduced for the mantle beneath mid-ocean ridges where the crust is 7 km thick. A magnesium content of ~11% MgO was calculated for the parent melt. Using a similar approach, the temperature required to produce a ~ 27-km thickness of melt from a hot plume was estimated. A T_P of ~ 1480˚C was found to be required, which generated magma with up to ~17% MgO. This suggested a temperature difference of ~ 200˚C between mantle plumes and mid-ocean ridges.

This work was influential in shaping subsequent thinking. However, it was founded on three critical, simplifying assumptions that are known not to be universally correct. These are that:

- mantle composition is uniform;
- crustal thickness is a proxy for T_P; and
- the T_P of the source of MORBs is the same as that of the mantle as a whole with the exception of plume-fed 'hot spots'.

This approach is somewhat analogous to that of 'Global Systematics' geochemistry, which also assumes that mantle composition is uniform. This method suggested that there are variations in mantle T_P of up to ~ 330˚C along the mid-ocean ridge system. Some regions do not fit the 'Global Systematics' model, however. For example, the geochemistry of the Azores and Galapagos regions, widely assumed to result from high-temperature 'hot spot' melting, do not conform to the predictions. In contrast, the geochemistry of the mid-Atlantic ridge in the Iceland region does. It was suggested that the "Global Systematics" should be applied only to mid-ocean ridges away from melting anomalies [Langmuir *et al.*, 1992]. This, ironically, is tantamount to suggesting that this particular geothermometer does not work where temperatures vary.

This illogical state of affairs recently led to a re-examination of the Global Systematics using the much larger geochemical databases that have accumulated subsequent to the proposal of the original model [Niu & O'Hara, 2008; Presnall & Gudfinnsson, 2008]. The Na_8 and Fe_8 contents of basalts from the mid-Atlantic ridge between ~ 55˚S and ~ 73˚N are anti-correlated, as the Global Systematics predict, for the entire data set. However, they are positively correlated in individual regional data subsets (Figure 17). The overall, global anti-correlation of the data results from the progressive offset of the array of individual, positively correlated subsets (Figure 18). The method as a whole is additionally flawed in that the predicted correlations of Na_8 and Fe_8 with axial depth hold only for the Iceland region and not for other parts of the mid-Atlantic ridge (Figure 19). Similar observations are reported for other parts of the global mid-ocean ridge system, *e.g.*, the East Pacific Rise (J. Natland, personal communication).

The observations are explained better by simple melting of peridotite at relatively uniform temperature and pressure, in the presence of minor compositional variations between sub-sections of the ridge [Presnall & Gudfinnsson, 2008]. Laboratory melting experiments using CMASNF samples (Section 3.2) reveal a positive correlation between Na_8 and Fe_8 for melts produced in the depth range 25-45 km (a pressure range of ~ 0.9-1.5 GPa). Small variations in the FeO/MgO content will produce local subsets that are arrayed in Na_2O-FeO space such that the trend of the entire composite data set (the 'global trend') shows an overall anti-correlation between Na_8 and Fe_8 (Figure 18). The data are consistent with fertile peridotite melting under largely uniform pressure and temperature conditions, but with minor variations in composition from region to region. This model can explain the geochemical observations consistently for the entire spreading ridge system. It also implies that there is little or no increase in source T_P in the vicinity of high-productivity ridge sections such as Iceland, the Azores, Tristan da Cuhna, and the Galapagos Islands.

Olivine control line analysis has also been applied to mid-ocean ridges. These studies have all been done using olivine cumulate rocks, because not one, single sample of glass with picritic composition has ever been retrieved from the mid-ocean ridge system, despite tens of thousands of samples having been collected in dredge hauls and drilling cores. Cumulate rocks cannot confidently be claimed to provide evidence for olivine-only crystallization, and the temperatures deduced are thus unreliable (Section 3.2.2.3). Nevertheless, the results obtained have been used widely as a reference temperature for 'normal' mantle, against which temperature estimates at 'hot spots' determined using the same method may be compared. T_P obtained for ridges using olivine control line analysis vary from 1,243˚C to 1,475˚C (Table 2).

These temperature estimates from olivine control line analysis vary so widely that they do not provide a useful guide to the T_P of the source of MORBs. Even if this problem can be resolved, it is not obvious that the T_P of the source of MORBs is a useful temperature against which to compare estimates from intraplate 'hot spots', as they may arise from different depths.

Rare-earth-element modelling of MORBs has yielded a generalized T_P of 1,300˚C for the source of 'normal MORB' [N-MORB; Section 7.2.7; McKenzie & O'nions, 1991]. This method also assumes a compositionally uniform peridotite source. In view of the evidence from mineral phase relations for influential source composition variations [Presnall *et al.*, 2002], this approach is unreliable.

3.3.2 The North Atlantic Igneous Province

The North Atlantic Igneous Province, including Iceland, is arguably the most extensively studied 'hot spot' as regards source temperature (Figure 20). Almost every method available to geology and geophysics has been applied. Nevertheless, unequivocal evidence for high temperatures is lacking.

Iceland itself is the largest subaerial exposure of spreading plate boundary on Earth, and is widely assumed to be the manifestation of excess melt production from a hot plume. The perception that there is anomalously large melt production at Iceland is key to the acceptance of this hypothesis there. Nevertheless, despite extensive seismological investigations, it is still not fully understood precisely how much excess melt is produced.

Explosion seismology was first used in the 1960s to explore the structure of the crust beneath Iceland and its offshore continuations, the Greenland-Iceland- and Iceland-Faeroes ridges. The base of the crust is defined seismologically by the Mohorovocic discontinuity (the "Moho"), where V_P rises abruptly from ~ 7.3 km/s to ~ 7.8-8.2 km/s [Mohorovicic, 1909]. In oceanic settings, it is simplistically assumed that the crust is formed by solidification of melt from the mantle, and that the mantle beneath is devoid of melt. This is known to not be strictly true [*e.g.*, Cannat, 1996], but it is nevertheless commonly assumed to be a reasonable approximation. A great deal of explosion seismology has thus been conducted in Iceland with the objective of determining the depth to the Moho, and thus the amount of excess melting there [see Foulger *et al.*, 2003 for a review].

In order to detect unequivocally a regional seismic discontinuity such as the Moho, refracted head waves must be observed. The deeper the discontinuity, the longer the explosion profile must be for this to be achieved. For a discontinuity tens of kilometres deep, profiles hundreds of kilometres long are needed, necessitating large explosions.

In the 1960s and early 1970s, it was assumed that the Icelandic crust had a thickness of the same order as North Atlantic oceanic crust, *i.e.*, 5-10 km [Foulger & Anderson, 2005]. Early refraction profiles in Iceland were relatively short, and did not yield refracted head waves from a clear Moho, so it was concluded that the base of the crust lay at ~ 10 km depth where the vertical velocity gradient abruptly reduced [Flovenz, 1980; Palmason, 1980]. Below ~ 10 km, the rocks had a fairly uniform V_P of ~ 7.0 km/s. This was interpreted as mantle with unusually low velocities as a result of retained partial melt. From this, it was concluded that the mantle beneath Iceland was anomalously hot [Björnsson *et al.*, 2005]. This model was known as the 'thin/hot' crustal model. It was considered to be consistent with the plume hypothesis, since it envisaged an anomalously hot mantle.

In the 1980s and 1990s, more ambitious seismic refraction experiments were performed. The submarine Greenland-Iceland and Iceland-Faeroe ridges were studied with profiles hundreds of kilometres long. Long profiles were also shot on Iceland itself [Bott & Gunnarsson, 1980; Darbyshire *et al.*, 1998; Holbrook *et al.*, 2001]. The crust beneath these ridges was concluded to be ~ 30 km thick. This finding was strongly supported for the Iceland-Faeroe ridge, which yielded clear, Moho refracted head waves. On the Greenland-Iceland ridge, and in Iceland itself, it was less strongly supported because reflections only were observed, not refracted waves. Reflections may occur from relatively thin layers, and do not necessarily indicate major seismic discontinuities. The anomalous nature of the base of the layer in Iceland with crust-like velocities is also revealed by analysis of receiver functions using teleseismic earthquakes [Du & Foulger, 1999; Du *et al.*, 2002; Du & Foulger, 2001; Foulger *et al.*, 2003], which shows that a sharp Moho is absent.

In Iceland, a reflective horizon had originally been reported in 1960 at ~ 20-40 km depth and interpreted as the Moho [Bath, 1960]. The report in the 1990s was thus a re-discovery [Bjarnason *et al.*, 1993; Björnsson *et al.*, 2005]. Nevertheless, it resulted in the material previously interpreted as hot, partially molten mantle being reinterpreted as sub-solidus, gabbroic, lower crust. A sub-solidus state for the material in the depth range ~ 10-30 km is supported by the observations that seismic attenuation is low, and V_P/V_S ratios are normal (Section 3.2.1.3). These results indicate that temperature does not exceed ~ 900˚C down to ~ 30-40 km depth. This interpretation implied, in turn, that the lower crust as defined by seismology is cooler than the same depth interval beneath the East Pacific Rise [Menke & Levin, 1994; Menke *et al.*, 1995]. The 'thin/hot' model was thus superceded by the 'thick/cold' crustal model.

Far from radically changing views of the fundamental origin of Iceland, this complete reversal of interpretation of the nature of Icelandic-type crust made no difference to mainstream assumptions. The 'thin/hot' model was considered to be consistent with the plume hypothesis on the grounds that it involved a hot mantle. The 'thick/cold' model was considered to support it since a thick crust was assumed to represent an anomalously large volume of melt.

The question concerning how much of the layer with crust-like seismic velocities is truly melt arose immediately following the new interpretation. Despite the fact that present-day Iceland happens to protrude above sea level, in a regional context it is a relatively minor topographic anomaly. It is the summit of a vast bathymetric swell ~ 3,000-km in north-south extent (Figure 21) [Vogt & Jung, 2005]. Iceland itself rises only ~ 1 km higher than a smooth continuation of the swell. Part of this small topographic anomaly may be a consequence of the lower density of the subaerially erupted volcanics in Iceland and on the Iceland-Faeroe ridge. Part must also arise from excess crustal thickness from a submerged microplate, analogous to the Easter microplate, which is required by palinspastic reconstructions [Foulger, 2006; Foulger & Anderson, 2005].

Because of the low topography of Iceland, the thick layer with crust-like seismic velocities cannot be entirely made up of rocks with basalt- or gabbro-like densities. If it were, Iceland would rise to over 4 km above sea level instead of the ~ 1 km observed [Gudmundsson, 2003; Menke, 1999]. The small topographic anomaly could be supported isostatically by just a few kilometres thickness of excess material with crust-like densities. The intriguing possibility that only part of the material with crust-like seismic wavespeeds corresponds to melt, is supported by petrological considerations. Mineralogical phase diagrams require that Icelandic lavas are extracted from the depth range ~ 40-50 km (~ 1.2-1.5 GPa) and at temperatures of 1240-1260˚C [Presnall & Gudmundsson, 2011]. This is partly within the seismically defined crust itself, and an abrupt jump from temperatures of ~ 900˚C to ~ 1250˚C seems implausible. This serious problem is usually ignored.

In summary, the evidence from crustal thickness as measured by seismology, inferences concerning temperature from attenuation and V_P/V_S measurements, and topographic and isostatic considerations, are inconsistent with the full 40-km-thick layer with crust-like seismic velocities being melt. The question then remains how much, excess melt there is at Iceland, if any, and how strong the crustal-thickness evidence is for high source temperatures.

Thick lava sequences are confirmed along the volcanic margins of the north Atlantic. Methods of generating these were investigated using finite-element modelling of super-continent breakup. The results show that a T_P anomalously high compared with that of the source of MORBs is not required to produce the observed melt thicknesses. These formed when the Eurasian continent broke up at ~ 54 Ma, splitting continental lithosphere up to 200 km thick, and allowing material to rise from those great depths [van Wijk *et al.*, 2004; van Wijk *et al.*, 2001].

In addition to indications of mantle temperature based on estimates of crustal thickness, a wide variety of other techniques have been used. Korenaga et al. [2002] studied the Greenland-Iceland ridge, and found that crustal thickness and V_P are negatively correlated. This is consistent with the excess melt there arising from a source with enhanced fusibility and not high temperature (Figure 7).

The temperature of the mantle has been estimated from seismic velocities, measured using teleseismic tomography and receiver functions. Low-V_P and V_S anomalies up to a maximum of ~ 2% and ~ 4% respectively occur throughout a region several hundred kilometres wide, extending from near the base of the crust to deep withing the mantle transition zone [*e.g.*, Foulger *et al.*, 2001; Hung *et al.*, 2004]. Receiver functions require much stronger mantle anomalies, in particular in the global low-velocity zone in the depth range ~ 80-120 km. There, anomalies of up to 10% in V_S are found [Vinnik *et al.*, 2005]. If due to temperature alone, a seismic anomaly as strong as this would require temperatures as much as ~ 600˚C higher than the background mantle T_P. This interpretation cannot be correct because it is at odds with all other temperature estimates. It can only be explained by partial melt, and if partial melt is required, it may be responsible for all of the low-velocity anomaly. The depressed velocities could alternatively be explained by a decrease of 5-10% in the Mg/(Mg+Fe) content of olivine in the mantle (Figure 9) (Figure 10). The seismic observations thus permit partial interpretation as elevated temperature, but they do not require it.

The seismic anomaly detected using tomography weakens with depth more rapidly than can be explained simply as a reduction in temperature sensitivity with depth (Figure 8). Critically, it does not continue down below the transition zone [Foulger & Pearson, 2001]. Thus, if the seismic anomaly in the mantle beneath Iceland is partly or wholly due to temperature, then it must reduce with depth and approach zero at the base of the upper mantle. The 410-km discontinuity is at normal depth beneath most of Iceland, providing no evidence for a temperature anomaly. It is ~ 20 km deeper than average beneath a small region in central-southern Iceland [Du *et al.*, 2006; Shen *et al.*, 2002]. This deepening can be explained by ~ 200˚C higher T_P, a few mole % higher magnesian olivine, relatively dry conditions, or a combination of some or all these effects.

Taken as a whole, the seismic data from Iceland require normal or even unusually low temperatures in the crust. Data from the mantle require at least some partial melt and permit, but do not require, elevated temperatures. If the observations are interpreted in terms of high temperatures, then these are confined to the upper mantle and do not extend down into the lower mantle. The entire suite of seismic observations could be interpreted as a compositional anomaly and partial melt in a fusible mantle at normal temperatures.

Petrological methods have also been used to estimate the T_P of the basalt source at Iceland and around the North Atlantic. Because Iceland has the thickest crust anywhere along the global mid-ocean ridge system, it was a key region used to construct the Global Systematics hypothesis [Klein & Langmuir, 1987; Langmuir *et al.*, 1992]. This hypothesis was developed when the 'thin/hot' crustal model for Iceland was in vogue, and a crustal thickness of 14 km was assumed for Iceland.

Contrary to the predictions of the Global Systematics, Na_8 and Fe_8 are essentially uncorrelated in Iceland. Basalts are high in Na_8, and highest above the thickest crust in central Iceland [Foulger *et al.*, 2005b]. This is the opposite of what is predicted. Primitive Icelandic lavas have Na_8 similar to basalt glasses from the East Pacific Rise where the crust is only 7 km thick. As for most parts of the mid-ocean-ridge system, the Na_8 and Fe_8 systematics are more readily explained by mineralogical phase relationships which indicate melt extraction under broadly uniform pressure and temperature conditions of 1.2 – 1.5 GPa (~ 40-50 km depth) and ~ 1240-1260˚C [Presnall & Gudmundsson, 2011]. The compositional variations observed are consistent with moderate compositional variations in the source. A source enriched in fusible components produces larger volumes of melt for the same mantle potential temperature. The ridge spanning Iceland taps a source strongly enriched in FeO and unusually low in MgO/FeO, consistent with enrichment with recycled crust [Presnall & Gudfinnsson, 2008]. This is also consistent with an interpretation of part or all of the low-velocity seismic anomaly detected by tomography, as an increase of 5-10% in the Fe/(Mg+Fe) content of olivine.

Estimates of the temperature of the source of Icelandic basalts obtained using olivine-control-line modelling vary widely. Values of T_P range from as low as 1,361˚C (compared with 1,243-1,351˚C obtained for mid-ocean ridges using the same method) to as high as 1,637˚C (compared with 1,453-1,475˚C obtained for mid-ocean ridges using the same method) [Falloon *et al.*, 2007a; Putirka, 2005a]. These results imply T_P anomalies for the mantle beneath Iceland of 10˚C and 162˚C respectively.

Olivine-control-line modelling has also been applied to rocks from West Greenland. The largest accumulation of picrite cumulate rocks known to have erupted in the Phanerozoic, some ~ 22,000 km^3, forms part of the volcanic margin there. They are widely assumed to require a hot source for their formation, and have been attributed to the head of a plume now assumed to underlie Iceland. Melt temperatures of 1,515 - 1,560˚C have been calculated for parental picritic liquids in equilibrium with olivines with compositions up to $Fo_{92.5}$ [Larsen & Pedersen, 2000]. The same method has been used to suggest that the source temperature of the postulated plume subsequently cooled to ~ 1,460˚C [Herzberg & Gazel, 2009]. It has been argued, alternatively, that olivine control line modelling cannot be applied to cumulate rocks (Section 3.2.2.3) and an alternative estimate of T_P, derived from a spinel-based geothermometer, of 1,340˚C has been proposed [Natland, 2008; Poustovetov & Roeder, 2000]. These disagreements notwithstanding, the source T_P of plume heads is predicted to be cooler than that of plume tails, the opposite of what is reported for evidence cited in support of an Icelandic plume.

No picrite glass samples have ever been found in Iceland. The most magnesian glass so far reported from there contains just 10.6% MgO (at least 12% is required for a rock to be classified as a picrite) [Breddam, 2002], and plagioclase and pyroxene crystals generally co-exist with olivine. Their crystallization should also be modelled, but this is usually ignored. No glasses are high enough in MgO to unequivocally indicate olivine-only crystallization. Small melt inclusions with MgO up to 16% are found in olivine crystals from Iceland, but it cannot be demonstrated that they have not gained MgO through re-equilibrating with surrounding olivine crystals. As a result, all olivine-control-line estimates of T_P for Iceland rest on the questionable assumption that the cumulate picrites used represent uncontaminated parental melts that were derived directly from the mantle, and did not lose or gain any material from conduits or magma chambers on their way to the surface.

Estimates of the T_P of the source of Icelandic basalts have been made using rare-earth element inversions [White *et al.*, 1995]. Assuming a peridotite mineralogy, values of ~ 200˚C higher than mid-ocean ridges have been obtained. These estimates are unsafe, and non-unique. Major-element variations both in Iceland and along the mid-Atlantic ridge provide strong evidence for variations in source composition, invalidating the assumption of uniform peridotite mineralogy. The observed rare-earth element patterns can be explained with no temperature anomaly, simply by reasonable alternative source compositions, *e.g.*, a mixture of abyssal gabbro and ocean-island basalt, which also match other geochemical characteristics of these basalts [Foulger *et al.*, 2005b].

Bathymetry, and the history of sea-floor subsidence in the North Atlantic has been used to estimate asthenosphere temperature [Clift, 2005; Clift *et al.*, 1998; Ribe *et al.*, 1995]. The sea floor is anomalously shallow in the North Atlantic, and it has been suggested that this results from support by thermally buoyant plume-head material. A mantle T_P anomaly of ~ 70˚C has been calculated, from these assumptions [Ribe *et al.*, 1995]. Similar T_P anomalies, of somewhat less than 100˚C, have been calculated by modelling the history of subsidence on the volcanic margins of east Greenland, Norway and Britain. The oldest oceanic lithosphere subsided more rapidly than the average for normal sea floor, consistent with asthenosphere that was warmer than normal when the volcanic margins formed, and cooled rapidly thereafter. An exception to this general rule is, ironically, the Iceland-Faeroe Ridge, which subsided at the average rate for normal ocean crust.

No significant heat flow anomaly has been found on the ocean floor north and south of Iceland [DeLaughter *et al.*, 2005; Stein & Stein, 2003]. Heat flow is lower west of the Mid-Atlantic Ridge than east of it, the opposite of what is predicted by the conventional model of an eastward-migrating plume [Lawver & Muller, 1994]. Heat flow is intrinsically insensitive to mantle temperature, however, and cannot rule out an anomaly of 100-200˚C.

Taken holistically, the seismic, petrological, bathymetric and heat-flow observations from the entire North Atlantic Igneous Province are all consistent with normal T_P or elevations of only a few tens of degrees Celsius. Some data require this result. Reports of high temperatures, which are widely cited, are unsafe and/or not required by the data. Unequivocal evidence from any method for temperature anomalies of ~ 200˚C is lacking. The absence of expected observations such as picrite glass, is an enigma to high-temperature genesis models.

3.3.3 Hawaii

The structure of the mantle beneath the Hawaiian region (Figure 22) has been studied in seismic experiments of a similar nature to those conducted in Iceland. The results are more limited, however, because the smaller extent of the landmass restricts the aperture of seismic arrays. Recently, there have been efforts to image the mantle in detail using seismic tomography.

A relatively high-resolution global V_P model was derived using diverse seismic-wave phases and available global stations [Li *et al.*, 2008]. It detected a strong low-velocity anomaly in the upper mantle beneath the Big Island (Figure 22). A weaker extension of this anomaly continues down into the lower mantle to a depth of ~ 1,500 km, dipping at approximately 45˚ to the northwest.

A later seismic tomography study involved a challenging 2-year deployment of 36 ocean-bottom seismographs over a 1,500-km-broad region of sea floor around the Big Island, supplemented by ~ 10 stations on land [Wolfe *et al.*, 2009; Wolfe *et al.*, 2011]. That study confirmed the presence of a strong, low-velocity body in the upper mantle. It was found to extend down to ~ 500 km depth, to involve anomalies of up to ~ 3% in V_S, and to pinch out rapidly at ~ 550 km depth. Beneath this, it was found to continue down in much weaker anomalies (up to ~1%) to ~ 1,500 km (Figure 22) (Figure 23). A smaller downward continuation of these anomalies dips to the northwest at an angle of ~ 40˚, and is interpreted as an artifact. The other dips to the southeast at an angle of ~75˚ and is interpreted by Wolfe et al. [2009] as a hot mantle plume with a temperature anomaly of up to 300˚C. The V_P model calculated using the same data set shows significant discrepancies with the V_S model, in particular in the upper mantle, but also images the deeper structure interpreted as a plume [Wolfe *et al.*, 2011]. The discrepancies in the upper mantle are surprising because similar results are expected if the same data set is used, and that is also the best-constrained region of the study volume.

The strong anomaly in the upper mantle is well resolved by both experiments, and resembles similar anomalies beneath Iceland, Yellowstone, and the Ontong Java plateau. Like those anomalies, it could result from partial melt, composition, temperature or a combination of some or all of these effects (Section 3.2.1.3). Remarkably, the weak anomalies imaged to extend down into the lower mantle in the studies of Wolfe et al. [2009] and Li et al. [2008] dip in opposite directions – there is no repeatability between the independent studies.

The reason for this is most likely that neither study can resolve structure in this part of the mantle because of the experimental limitations of studying a region remote from most of the earthquake sources and with limited station coverage. In order to image a region well using teleseismic tomography, a rich set of crossing ray paths is needed. These are not available in the lower mantle in the experiment of Wolfe et al. [2009]. The anomaly dipping to the southeast is constrained only by *ScS* arrivals, which were reflected off the surface of the core and approach Hawaii steeply upwards in a bundle of quasi-parallel rays. The low velocities imaged using this ray bundle could thus equally well originate anywhere along the ray paths between the base of the study volume and the earthquake sources in the Chile trench. In addition to this problem, the teleseismic tomography method cannot fully account for structure outside of the study volume, which has now been shown to significantly corrupt inversion images [Masson & Trampert, 1997].

The plume-like, low-velocity body imaged by Wolfe et al. [2009] in the lower mantle is almost certainly an artifact [Cao *et al.*, in press, J.R. Evans, personal communication][1]. The mirror-image anomaly imaged by Li et al. [2008] is also not resolved, and also almost certainly an artifact [R. van der Hilst, personal communication].

Hawaii is unique because it is the only place on Earth where volcanic glass with picritic composition has been found. It is extremely rare, however, and known only from a few terrestrial samples and sand grains < 300 μm in size. The latter were dredged from Puna Ridge, the submarine extension of the East Rift Zone of Kilauea volcano, south east of the Big Island. They were discovered in 1988 in a core retrieved from 5,500 m depth [Clague *et al.*, 1991]. With MgO of up to 15.0 wt%, they define, together with the less-magnesian terrestrial samples, an undisputed olivine control line (Figure 24). These data are free of the uncertainty associated with the crystalline samples used everywhere else, regarding whether or not they represent a single melt.

The temperature at which the melt formed, along with the depth of formation of the alkalic and tholeiitic basalts from Hawaii, has been studied using olivine-control-line modelling and laboratory phase-equilibrium data (Section 3.2). This approach suggests that Hawaiian tholeiitic basalts arise from melt that formed at depths of ~ 130-150 km (~4-5 GPa) at a T_P of ~ 1,450-1,500˚C. This is much deeper and hotter than calculated for the source of magmas at the mid-ocean ridge system. These are thought to arise from 40-50 km depth (1.2-1.5 GPa), at temperatures of 1,240-1,260˚C [Clague *et al.*, 1995; Presnall & Gudmundsson, 2011].

If the conductive layer is 100-200 km thick, these variations in source temperature may simply reflect the different depths of extraction [Presnall & Gudmundsson, 2011; Presnall *et al.*, 2009]. A source for Hawaiian lavas in the seismic low-velocity zone could explain the eruptive sequence of initial low-volume alkalic magmas, followed by voluminous tholeiitic magmas showing low-pressure olivine-controlled crystallization, and terminated by low-volume alkalic magmatism, as it matches the layering of melts in the low-velocity zone [Presnall & Gudmundsson, 2011].

Numerous other investigations of the same samples have been conducted, and the temperatures of formation deduced vary greatly (Table 2). Estimates range from as low as 1,286˚C [Falloon *et al.*, 2007a] to as high as 1,688˚C [Putirka, 2005a]. The temperature differences deduced between the source of Hawaiian basalts and that of mid-ocean ridge basalts ranges from essentially zero [Green & Falloon, 2005; Green *et al.*, 2001] to 270˚C [Herzberg *et al.*, 2007]. This variability is due almost entirely to the use of different olivine geothermometers (Figure 13) [Falloon *et al.*, 2007b]. Thus, although studies of the source T_P of Hawaiian basalts are not limited by the lack of

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[1] http://www.mantleplumes.org/TomographyDiscussion.html

picrite glass, as they are at other localities, petrological modelling techniques are still so variable that the question of whether there is petrological evidence for anomalously high temperatures in the mantle beneath Hawaii is still controversial. A T_P of ~ 1,500˚C has been derived for the mantle beneath Hawaii using rare-earth-element modelling [McKenzie & O'nions, 1991]. This is unsafe because the source of Hawaiian basalts cannot be the standard peridotite assumed (Section 3.2.2.4) [*e.g.*, Kogiso *et al.*, 2003; Sobolev *et al.*, 2005].

The temperature of the mantle in the Hawaiian region has also been investigated using heat-flow measurements. The Hawaiian island chain is the subaerial summit of a 200-km-wide ridge, which overlies a ~1,200-km-wide bathymetric swell. There are two main candidate explanations for this swell. The first is that it is supported by thermal buoyancy which has replaced the lower part of the lithosphere [Crough, 1983]. The second explanation is that it is simply a consequence of the volcanism itself, *i.e.*, that it is supported by buoyant mantle residuum left over after melt extraction.

The thermal hypothesis has been tested using heat-flow measurements on the sea floor. The Pacific plate near Hawaii is moving at ~ 10.4 cm/a and the ~ 90-100 Ma lithosphere there is ~100 km thick. The heat flow anomaly from a plume beneath the Big Island is predicted to increase from zero at its centre, where there has been insufficient time for the heat to conduct to the surface, to a maximum of ~25 mW/m^2 after ~15 – 20 Ma, at a distance of ~ 1,800 km along the volcanic chain [von Herzen *et al.*, 1989]. The anomaly is predicted to peak on the axis of the chain and to decrease perpendicular to it, approaching zero at ~600 km distance (Figure 16).

A transect perpendicular to the chain ~ 1,800 km northwest of the Big Island does not reveal such a pattern (Figure 25). Furthermore, along the chain, the many heat-flow measurements that have been made show a large scatter which, when averaged, indicate similar heat flow to lithosphere of comparable age elsewhere. Heat flow is thus essentially constant from 800 km southeast of the Big Island to 800 km to the northwest (Figure 26) [DeLaughter *et al.*, 2005]. It has been suggested that shallow water circulation could mask the predicted heat flow signal [Harris & McNutt, 2007] but this was found to be only a local effect in old oceanic crust, and to not mask heatflow patterns on a regional scale [Stein & Von Herzen, 2007].

The hypothesis that the swell is due entirely to buoyant residuum rests on the fact that the removal of partial melt from common mantle materials, *e.g.,* garnet peridotite, leaves residuum that is as much as ~ 1% less dense than the original source rock. This is a consequence of the preferential removal of iron (Fe) from olivine and the heavy minerals garnet and orthopyroxene [Schutt & Lesher, 2006]. The density reduction from chemical depletion of this kind is comparable to that produced by the temperature excesses commonly proposed for plumes. For example, the density reduction that results from the removal of 20% partial melt from spinel lherzolite is similar to that produced by raising the temperature by ~ 200˚C. It is interesting to note that when mantle peridotite and similar petrologies partially melt, both the liquid and the solid residuum parts are less dense than the original protolith.

The volume of the swell along the Hawaiian chain is almost everywhere approximately proportional to the volume of the magma produced [Phipps-Morgan *et al.*, 1995]. If it is supported entirely by the residuum, the ratio of swell volume to magma volume is expected to be ~ 0.25 [Foulger, 2010]. As yet, a detailed study of whether the entire swell can be explained by this mechanism, taking into account the unusual mantle petrology indicated by the geochemistry of Hawaiian lavas, has not yet been done.

3.3.4 Oceanic plateaus

The temperature of the mantle beneath oceanic plateaus is of interest because they are commonly attributed to plume-head volcanism, with temperature anomalies of 200-300˚C. Mantle temperature beneath plateaus, ridges and seamounts, has been studied using their subsidence histories (Figure 27) [Clift, 2005]. The results can be divided into four categories (Figure 28):

a. Some sites subsided more quickly than average oceanic crust, consistent with formation over mantle hotter than normal, and subsequent transportation, by plate motion, over cooler mantle. Several sites in the North Atlantic were found to have such subsidence patterns;

b. Some sites subsided at the average rate for oceanic crust, consistent with formation over normal-temperature mantle. Examples include the Iceland-Faeroe Ridge, the Ninetyeast Ridge, Shatsky Rise, the Kerguelen Plateau, and Walvis Ridge;

c. Some sites subsided more slowly than average, consistent with formation over mantle cooler than normal. Examples include the Magellan Rise, the Manihiki Plateau, and several sites along the Ninetyeast Ridge, the Kerguelen Plateau, and the Rio Grande Rise;

d. Where the volcanism of interest occurred away from a spreading ridge, when the oceanic crust was already considerably aged, interpretation of the observations is less straightforward. The subsidence observed could either be compared with that expected for crust of the age of the local sea floor, or it could be compared with zero-age crust. The latter case would be appropriate if the crust were completely thermally reset by the magmatism. Late-stage volcanism occurred on several of the sites studied, including the Ontong Java Plateau, the mid-Pacific Mountains, the Nintoku Seamount, Hess Rise and MIT Guyot.

The only sites where an unusually rapid pattern of subsidence suggests crustal formation over anomalously warm mantle, are at the volcanic margins around the North Atlantic. The observations are, however, consistent with small temperature anomalies throughout relatively thin layers only. Modelling has suggested temperature anomalies no larger than ~100˚C and layer thicknesses no more than ~100 km.

Every other oceanic volcanic area studied has a subsidence history that suggests formation over mantle that either had an average temperature, or was anomalously cool. In some cases, the errors in water depth estimates are sufficiently large that a mantle warmer than average cannot be ruled out. However, only the sites in the North Atlantic require elevated temperature. In some cases, the maximum thickness and temperature anomaly of permitted thermal sources has been estimated. This has allowed the subsidence data to rule out thermal anomalies of the size and magnitude required to produce some magmatic features. The Ontong Java Plateau, the largest volcanic province on Earth, is one of these [Ito & Clift, 1998].

Present-day mantle temperature beneath the Ontong Java plateau has been estimated using seismology. Because it is almost all oceanic, seismic stations have not yet been installed on its surface. It has thus been studied using surface waves from earthquakes in the subduction zones to the south, recorded on seismic stations on the Caroline Islands to the north [Richardson *et al.*, 2000].

The plateau is underlain by a body with V_S up to 5% lower than the global average that extends from near the surface down to ~ 300 km depth. This anomaly is similar to that found beneath Iceland (Section 3.3.2). If interpreted as being caused by anomalously high temperature only, a maximum temperature ~ 300˚C higher than average is predicted. However, the observations could also be explained by low-degree (< 1%) partial melt or by a compositional anomaly such as a reduction by ~ 5% in the Mg/(Mg+Fe) content of olivine (Table 1). In order to reduce the ambiguity between partial melt, composition and temperature, attenuation was calculated. If the anomalously low seismic velocities result from partial melt of high temperature or melt, high attenuation is expected.

Seismic *ScS* phases, shear waves that travel downwards from the earthquake source, are reflected off the surface of the core, and return again to the surface, reveal that attenuation within the anomaly beneath the Ontong Java plateau is, on the contrary, exceptionally low compared with the Pacific mantle in general [Gomer & Okal, 2003]. This rules out an explanation for the low velocities in temperature alone, and requires a chemical interpretation, and one that implies a high-viscosity body [Klosko *et al.*, 2001].

Separate from these considerations, in common with the 'superplumes', this body is an example of a low-velocity seismic anomaly that must be chemical, and not thermal in nature. Furthermore, it is comparable in terms of velocity to the low-V_S anomaly reported to underlie Iceland. It is important in that it illustrates that seismic velocity cannot be assumed to be a thermometer, and in particular low seismic velocities cannot be assumed to reflect high-temperature, buoyant material (Section 3.2.1.3).

Other methods used to investigate the temperature of the mantle source from which the Ontong Java plateau was derived have yielded much higher temperatures, at odds with the low temperatures apparently required by the subsidence history. Olivine control line analysis has been applied to cumulates that contain up to 10.9% MgO. This analysis has yielded estimates of source T_P of 1,500˚C, suggesting a temperature 100-220˚C hotter than mid-ocean ridges [Herzberg *et al.*, 2007]. This analysis is, however, flawed for the same reasons as cumulate-based analyses of this kind elsewhere (Section 3.2.2.3). The high levels of incompatible element concentrations have also been modelled. If the source rock was a near-primitive peridotite, a degree of partial melt of ~30% is required [Fitton & Godard, 2004]. Such a high degree of partial melt would require a source T_P of at least 1,500˚C. However, this line of reasoning is immediately invalidated if the source was not a near-primitive peridotite. A more fertile source could melt to produce the observed incompatible element abundances, and other geochemical characteristics, with no temperature anomaly at all, and such a process could produce the volume of melt observed [Korenaga, 2005].

3.3.5 Swells

A hot, buoyant underlying mantle is commonly suggested to explain areas that stand high, both oceanic and terrestrial. This explanation has been particularly favoured where there is no obvious link to tectonism, such as a nearby active plate boundary. Localities on land where it has been suggested include the Hoggar swell, in north Africa, and the Colorado Plateau. Unfortunately, it takes several tens of millions of years for heat to conduct through thick continental lithosphere, to cause measurable heat flow anomalies, so the suggestion can rarely be tested using that method. As a result, while a thermal model can often fit the observations that do exist, there is typically essentially no corroborating evidence and thus it is only one of several permitted candidates.

The Hoggar volcanic province is an example. It has been active from ~ 35 Ma almost up to the present day [Liégeois *et al.*, 2005]. The swell is ~ 1 km high and has a diameter of ~ 1,000 km. On it lie ~ 14,000 km^3 of mafic lavas. It has been proposed that the uplift and volcanism is due to an underlying hot mantle plume [*e.g.*, Sleep, 1990]. Heat flow over the swell is normal for the age of the lithosphere (~50 mW/m^2), and even lowest at the highest basement elevations. Enigmatically, heat flow is higher north of Hoggar, outside the swell, in the region of the Saharan basins.

Nevertheless, the heat flow data alone cannot rule out a hot source because 35 Ma is insufficient for heat to have conducted through the lithosphere and caused a measurable surface heat flow anomaly. In this time, a layer with a temperature anomaly of ~ 250°C, 100 km thick, emplaced beneath lithosphere 120 km thick, would only give a heat flow anomaly of 0.2 mW/m^2. It would take 100 Ma for surface heat flow to increase by 2 mW/m^2. The standard deviation of heat-flow measurements on land is typically ~ 10 mW/m^2, so such a small anomaly would be unresolvable.

The Colorado Plateau, in western North America, comprises a deeply eroded land surface lying at a somewhat lower elevation than the surrounding mountains. Its most famous deep erosion feature is the Grand Canyon. Volcanic rocks are confined to its margins, and there is no associated flood basalt eruption.

The exact timing and cause of Rocky Mountain and Colorado Plateau uplift is the subject of a longstanding debate [*e.g.*, Pederson *et al.*, 2002; Sahagian *et al.*, 2002]. The current consensus is that since the Late Cretaceous there has been ~ 2 km of largely amagmatic Cenozoic true rock uplift [England & Molnar, 1990]. Several explanations have been put forward regarding what has caused this, including thermal explanations. The uplift has been linked to the Laramide orogeny, which occurred at ~ 80-40 Ma [McMillan *et al.*, 2006] and was associated with the reorganisation of the plate boundary on the west coast of North America. Other suggestions include flat subduction of the eastward-moving Farallon slab, and delamination of the continental lithosphere. Warming of the thick Colorado Plateau lithosphere over 35-40 Ma following the removal of the Farallon slab can explain the time-scale of migration of Cenozoic magmatism from the margins towards the plateau interior, the lower seismic velocities, and the gravity field [Roy *et al.*, 2009]. Edge-driven, small-scale mantle convection can explain the large gradients in seismic velocity across the margins of the plateau, the low lithospheric seismic velocities, continued, low-level volcanism, and the uplift [Van Wijk *et al.*, 2010]. These proposed explanations do not invoke extremely high temperatures in the mantle, but only involve shallow convection. Interestingly, they predict more uplift than is observed, a problem that they share with thermal models for Iceland and the Ontong Java plateau.

Cape Verde, Bermuda, Crozet and Réunion are examples of oceanic swells where elevated heat-flow measurements have been reported and cited in support of high-source-temperature models. Ironically, elevated heat flow is absent at Hawaii, but this has been explained as hydrothermal circulation removing the expected signal [Harris *et al.*, 2000]. Because of this inconsistent interpretive approach to heat-flow measurements from swells, the data were recently re-appraised using a correct model of average global heat flow, and assessing local effects [Stein & Von Herzen, 2007]. Very few reliable measurements indicate heat flow sufficiently elevated to support a high mantle temperature explanation (Figure 29). The majority are small and within one standard deviation of the measurements which is typically ~ 5 mW/m^2. Multiple high-heat-flow measurements at individual swells, that might suggest regional elevated temperature, are lacking. There is thus no compelling evidence for broad regions of elevated heat flow over swells.

A large region of the south Pacific sea floor, including the Darwin Rise and the South Pacific 'superswell' has been explained as a global-scale uplift caused by elevated underlying mantle temperatures (Figure 30) [Larson, 1991; McNutt & Judge, 1990]. A problem with this model is that the Darwin Rise is not, in general, anomalously shallow, but merely a region of abundant seamounts emplaced on sea floor of normal depth. The 'superswell' is, in contrast, up to ~ 500 m shallower than Pacific sea floor of comparable age (Figure 31, top). This shallowing could be explained by removal of the lower lithosphere by thermal erosion, and its replacement with warmer asthenosphere. Heat flow elevated by ~ 15-25 mW/m^2 above the global average for sea floor of comparable age is predicted by this model. Heat flow in the 'superswell' region is, however, normal (Figure 31, bottom) [DeLaughter *et al.*, 2005; Stein & Abbott, 1991; Stein & Stein, 1993]. The shallow observed bathymetry may therefore merely be a result of widespread volcanic products from the numerous seamounts and volcanic islands that carpet the Pacific Ocean seafloor in this region [Levitt & Sandwell, 1996].

4. Discussion

Melting anomalies are almost never spots, and there is little equivocal evidence that they are hot. The term 'hot spot' encourages the assumptions that they are, it tends to be defended rather than tested. It is thus a barrier to progress.

The term 'hot spot' was originally used to describe the Hawaiian system. The plume hypothesis was inspired by that system, which arguably fits it best. Hawaii is far from any plate boundary, occurs in oceanic lithosphere which is structurally simple compared with most continental lithosphere, and it has a relatively regular time-progressive volcanic chain. The present volcanic production rate at Hawaii is extraordinarily large [Robinson & Eakins, 2006]. No other melting anomaly fits the plume hypothesis so well, and it is questionable whether the hypothesis would have developed at all if Hawaii did not exist.

Nevertheless, close scrutiny of the Hawaiian system has tended to undermine support for the model rather than strengthen it [Foulger, 2010]. It was known when the plume hypothesis was originally suggested that the locus of melt extraction was not 'fixed', but had drifted substantially [Karpoff, 1980; Kono, 1980]. This drift has now been measured and found to be of the order of 2,000 km [Sager, 2007; Tarduno & Cottrell, 1997]. The present-day, high volcanic rate has persisted only for the last ~ 2 Ma. The volcanically active region is ~ 350 km long, and is therefore not well-described as a 'spot'.

Young volcanism at other melting anomalies considered to be the strongest 'hot spot' candidates, is distributed over even wider regions, *e.g.*, 400 km (erupted over the last 10,000 a) at Iceland [Johannesson & Saemundsson, 1998] and 1,000 km (erupted over the last 3 Ma) in the Galapagos region [O'Connor *et al.*, 2007]. At Yellowstone, volcanics younger than 4 Ma extend for 450 km along the Eastern Snake River Plain, and for 1,000 km in total, from Yellowstone, across the entire northern boundary of the Basin and Range province as far as the Cascades in Oregon [Christiansen *et al.*, 2002].

Volcanic regions that are hundreds, or even thousands of kilometres broad, are not well-described as 'spots', and this brings into question the concept of a narrow, localised melt-delivery system. As progressive research shows volcanic regions to be wider, so do postulated plume stems widen until eventually multiple 'plumelets' or 'baby plumes', rising from broad, regional, hot bodies are suggested. Systematic spatial patterns of eruption, *e.g.*, along linear features such as rift valleys, are attributed to lateral flow along weak zones, or 'upside-down drainage' [*e.g.*, Sleep, 1997]. At this point, the predictions of the modified plume hypothesis can become indistinguishable from those of the plate hypothesis [Foulger, 2010], except for the question of source temperature.

Investigating whether a melt arises from an unusually hot source is surprisingly difficult. First, the issue is confused by the difficulty of deciding on a standard against which to compare estimates. The average T_P of MORB is the most common norm used. However, estimates of the average T_P of MORB source vary widely. They range from 1,260˚C to 1,400˚C, and estimates of the variations along ridges range from ± 20˚C to ± 200˚C [Anderson, 2000; Anderson, 2007; Kaula, 1983; McKenzie & Bickle, 1988]. The problem is further exacerbated by the practise of some workers of eliminating from calculated averages, estimates from ridges considered to be 'plume influenced'. Such an approach clearly cannot be used in experiments designed to test whether 'hot spots' are hot.

The second major problem is that of depth. The temperature profile in the mantle is not well known. The simple, traditional view considers the asthenospheric melt source to be thermally homogenised by convection and to have a uniform T_P that is identical to that of MORBs, except at 'hot spots' with temperature excesses of 200-300˚C. This model cannot be correct. Mantle T_P must vary, both laterally and with depth. Seismology provides abundant evidence that the asthenospheric mantle is structurally and compositionally inhomogeneous, in particular at shallow depths. Factors such as the variable distribution of radioactive elements, the insulating effect of continents, and disruption of the asthenosphere by continental breakup and subduction cause T_P to vary laterally by tens of degrees Celsius [Anderson, 1982; Coltice *et al.*, 2007]. Mantle convection would not occur if T_P did not vary. It must also rise with depth throughout the 100-200-km-thick surface conductive layer [Anderson, 2007]. MORB is thought to be extracted from the topmost few tens of kilometres of the asthenosphere. Comparisons with the T_P of MORB source, even if it could be determined reliably, are thus only valid for melts extracted from similar depths. Melts that arise from the asthenosphere beneath thick lithosphere, *e.g.*, in older ocean basins or continental regions, are drawn from deeper in the surface conductive layer where T_P is higher, and thus their source temperatures cannot validly be compared with those of MORB.

These fundamental philosophical problems are exacerbated by the difficulty in estimating T_P using the methods currently available. Seismology alone cannot yield temperature, as composition and partial melt have strong, often stronger, effects on almost all seismological parameters. Only in rare cases can seismology alone constrain temperature. The seismological demonstrations that the 'superplumes', and the body beneath the Ontong Java plateau, are not hot are salutary cases. Combining seismology with other disciplines, *e.g.*, petrology, is a powerful approach that is able to distinguish between thermal and chemical origins for thick sequences of melt in flood basalts and at oceanic plateaus [Korenaga *et al.*, 2002]. Seismic tomography cross sections of the mantle cannot be viewed as maps of the temperature. Red is not equal to hot and blue is not equal to cold. Topography on the transition-zone-bounding discontinuities is an unreliable geothermometer, as it is affected by composition and the presence of water, in addition to temperature.

Similar problems limit petrological methods. The T_P anomalies predicted by even the hottest plume models are small compared with the uncertainties and the repeatabilities between results from different methods. For example, estimates of T_P calculated by different investigators for MORBs using olivine-control-line analysis vary by 195˚C. Estimates for T_P for the source of Hawaiian basalts vary by ~ 400˚C (Table 2). Some studies conclude that there is very little difference in the source temperatures of MORB and lavas from 'hot spots', in particular when volatile content and possible variations in melt fraction are taken into account [*e.g.*, Green & Falloon, 2010]. There is still considerable disagreement over the correct way to apply olivine-control-line analysis, and even whether it is valid to use picrite cumulates or whether glass only is reliable. Other methods, such as the Global Systematics, and rare-earth-element modelling have either been shown to be invalid, or do not apply for lavas with different source compositions.

Measuring temperature and temperature variations in Earth is difficult. Nevertheless, despite extensive study, there is little equivocal evidence for unusually high T_P at ‘hot spots’. This is despite diverse methods being applied to localities that range from long-extinct flood basalts to the currently active volcanic regions considered to be the strongest candidates for the products of hot mantle plumes.

The only localities where reliable evidence for unusually high T_P has been reported are at the North Atlantic volcanic margins and Hawaii. Anomalously rapid subsidence of the oldest ocean floor at the margins of the North Atlantic suggest that the mantle at the time of breakup was ~ 100˚C warmer than average. No unequivocal evidence for high T_P has been found in Iceland itself. Hawaii, is unique in being the only place on Earth where picrite glass has been found. At all other ‘hot spot’ localities, olivine-control-line modelling of T_P relies on cumulate samples of disputed reliability. At both the North Atlantic margins and Hawaii, the melt may be sourced from deep within the conductive layer. The North Atlantic volcanic margins formed when continental lithosphere possibly as thick as ~ 200 km ruptured, potentially sourcing melt from such great depths. The lithosphere at Hawaii is ~100 km thick, and the tholeiite erupted there may come from beneath it. Although the melt source appears to be unusually hot at both these localities, this may thus simply be a consequence of their unusually deep origins within the conductive layer.

The failure to unequivocally detect high temperatures at ‘hot spots’ has done little to shake confidence in the assumption that excess volcanism is fueled by unusually hot mantle source regions. It has instead been attributed to the difficulty in making observations, or to the expected anomalies being too small [*e.g.*, Ribe *et al.*, 1995]. For example, the absence of picrite glass, *e.g.*, at Iceland, has been attributed to its high density precluding it from rising to the surface [Larsen & Pedersen, 2000]. This does not explain, however, why the picrite cumulates that are often assumed to represent original melts, are common. The absence of plateau subsidence predicted by cooling models has been attributed to post-eruption magmatic underplating causing continuous crustal growth and bathymetric support [Clift, 2005] or to support from buoyant, depleted mantle residuum remaining after melt extraction, *e.g.*, at the Ontong Java plateau [Robinson, 1988] and the Hawaiian swell [Phipps-Morgan *et al.*, 1995]. On the other hand, the lack of such support in the North Atlantic has been explained by the suggestion that vigorous convection removed the residuum there. In a similar interpretive style, measurement of high heat-flow at oceanic swells are attributed to high mantle temperatures, but the absence of high heat-flow has been explained as hydrothermal circulation removing the expected signal.

There is no unequivocal evidence, from seismic, petrological, bathymetric or heat flow data from the vast majority of ‘hot spots’ that they are, indeed, hot. In some cases there are indications that they are not. The term ‘hot spot’ is thus a misnomer. It is, however, not only that. It pre-supposes a mechanism for the creation of anomalous volcanic regions, and in doing so discourages hypothesis testing. The label ‘hot spot’ should be abandoned in favor of one that names the phenomenon under scrutiny not by an unverified, presumed genesis process, but in terms of its unequivocal, observed characteristics.

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Table 2 Estimates of mantle source T_P.

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Phase	Partial melt (per 1% increase in melt content)	Composition (per 10% reduction in Mg/(Mg+Fe) in olivine)	Temperature (per 100 K increase)
V_P	2-3%	7%	1%
V_S	3-10%	12%	1.5%

Table 2 Estimates of mantle source T_P.

	Global	'normal' mid-ocean ridges	Hawaii	Iceland	West Greenland	Gorgona	Réunion
Olivine control lines [Herzberg *et al.*, 2007]		1,280-1,400	1,550	1,460		1,460	
Olivine control lines [Green *et al.*, 2001]		1,430	1,430				
Olivine control lines [Falloon *et al.*, 2007a]		1,243-1,351	1,286-1,372	1,361			1,323
Olivine control lines [Putirka, 2005a]		1,453-1,475	1,688	1,637			
Olivine control lines [Clague *et al.*, 1991]			1,358				
Olivine control lines [Larsen & Pedersen, 2000]					1,515 - 1,560		
Global systematics [Klein & Langmuir, 1987; Langmuir *et al.*, 1992]		1,300-1,570					
Geophysics [Anderson, 2000]	1,400-1,600						
Plate velocities [Kaula, 1983]	± 180 K						
Mineralogical phase diagrams [Presnall & Gudfinnsson, 2008]		1,250-1,280		1,250-1,280			

Figure captions

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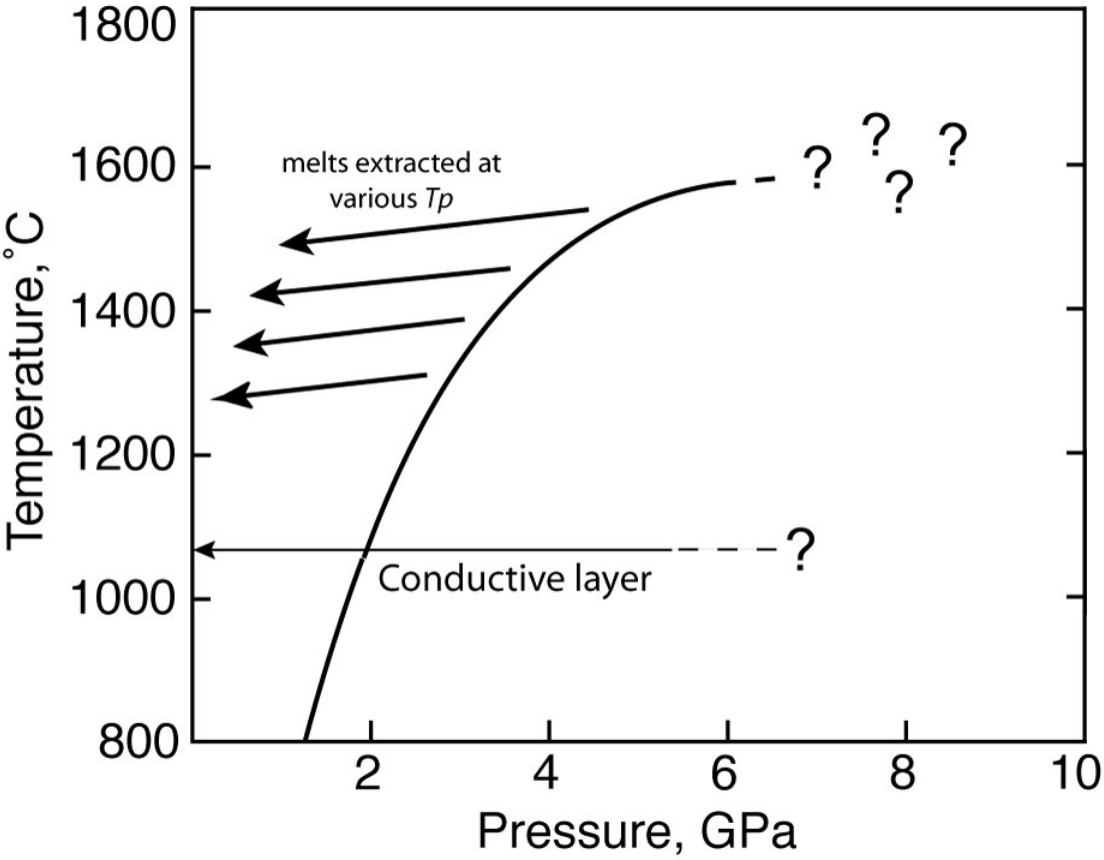
Figure 27 Deep Sea Drilling Project and Ocean Drilling Program drill sites where the temperature of the mantle has been estimated using subsidence histories from sedimentary cores. Faster subsidence than mid-ocean ridges (MORs) indicates higher initial mantle temperatures, and slower subsidence than MORs indicates lower initial mantle temperatures [from Clift, 2005].

Figure 28 Reconstructed depth-to-basement histories for a selection of sites that subsided a) faster than average, b) at an average rate, and c) more slowly than average. Vertical bars show uncertainties in water depth estimates from sediments or microfossils. Black bars indicate high confidence, and gray bars low confidence. Solid curves shows the predicted subsidence history for normal oceanic crust. The end points of these curves are generally known accurately. d) shows sites where volcanism occurred away from a mid-ocean ridge. The grey curves show the predicted curve for crust of the same age as the volcanic rocks (equivalent to total thermal resetting of the lithosphere), and the black curves show predicted subsidence for crust of the age of the host crust [from Clift, 2005].

Figure 29 Pogo heatflow data vs. crustal age for the Hawaii, Réunion, Crozet, Cape Verde and Bermuda swells (triangles) and non-swells (squares). There is no significant evidence for elevated heat flow over swells [from Stein & Von Herzen, 2007].

Figure 30 Map of the south Pacific showing the Darwin Rise and the Superswell [from Stein & Stein, 1993].

Figure 31 Top: Bathymetric depths from the Superswell and other regions of the same age in the Pacific for comparison. Bottom: heat flow observed, and predicted by thinning models [Stein & Stein, 1993].



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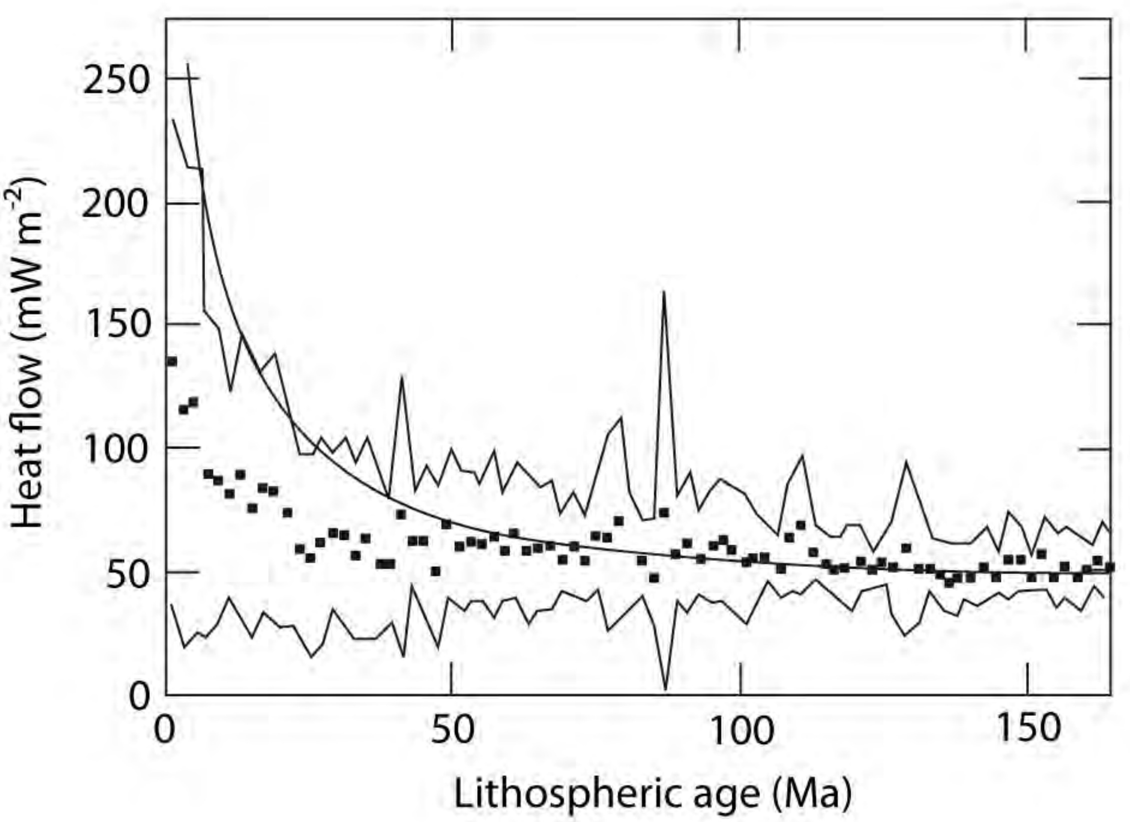

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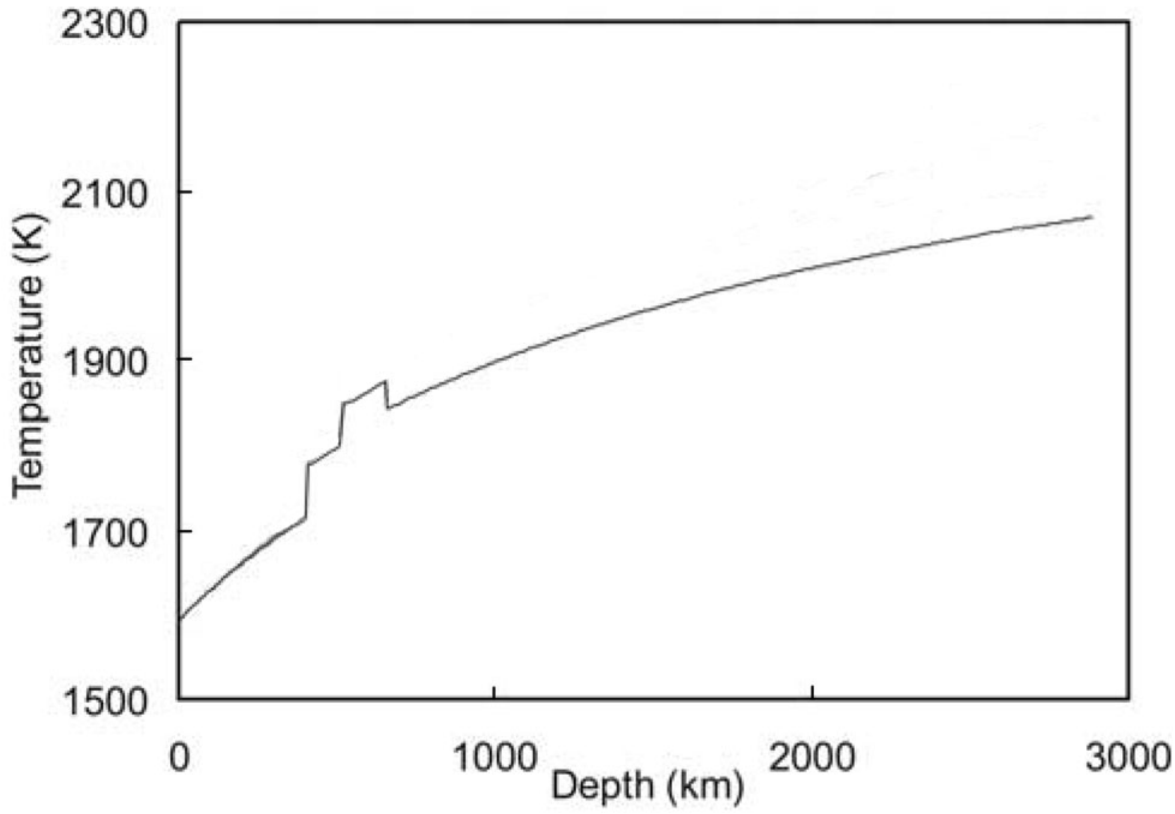

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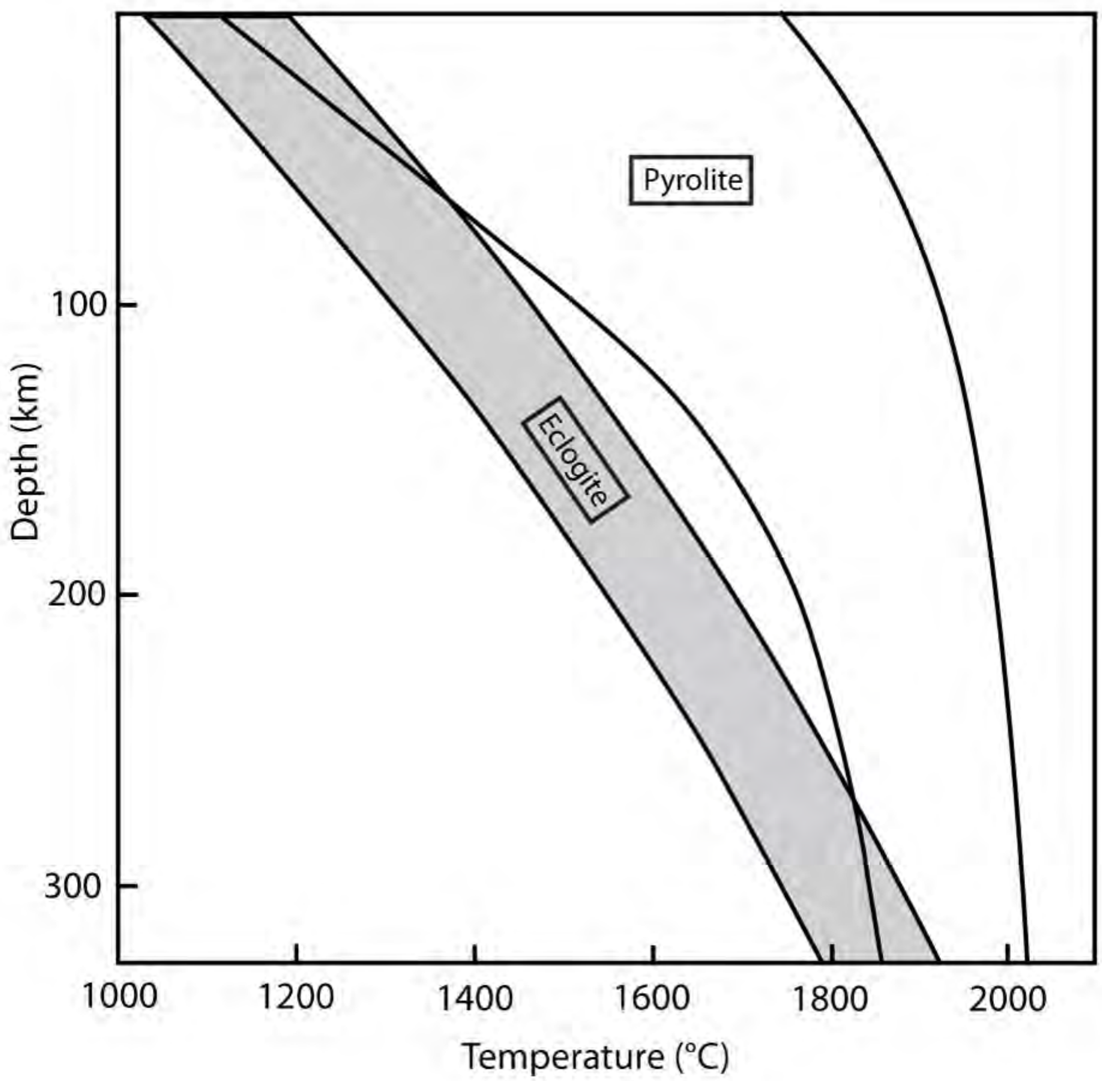

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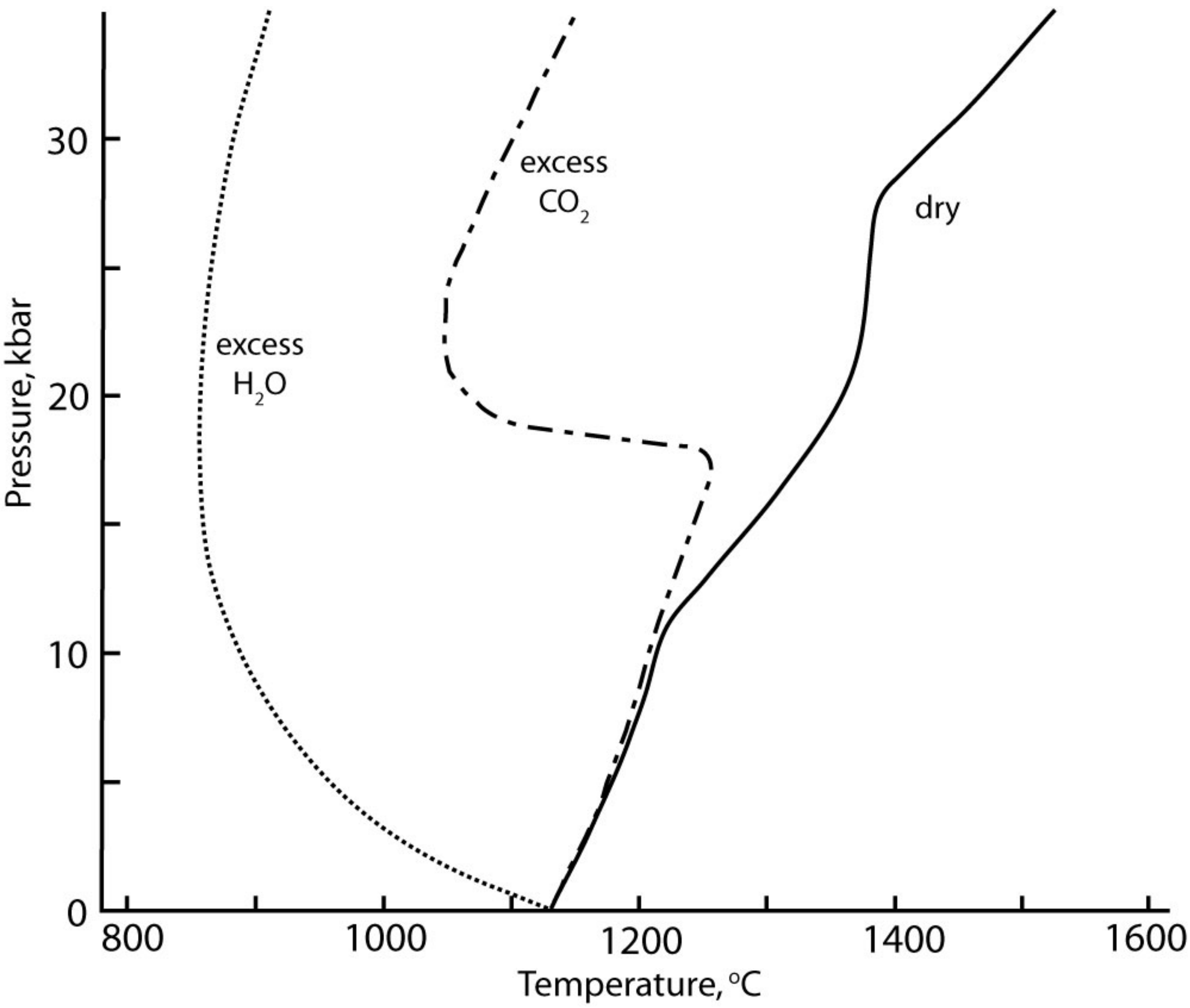

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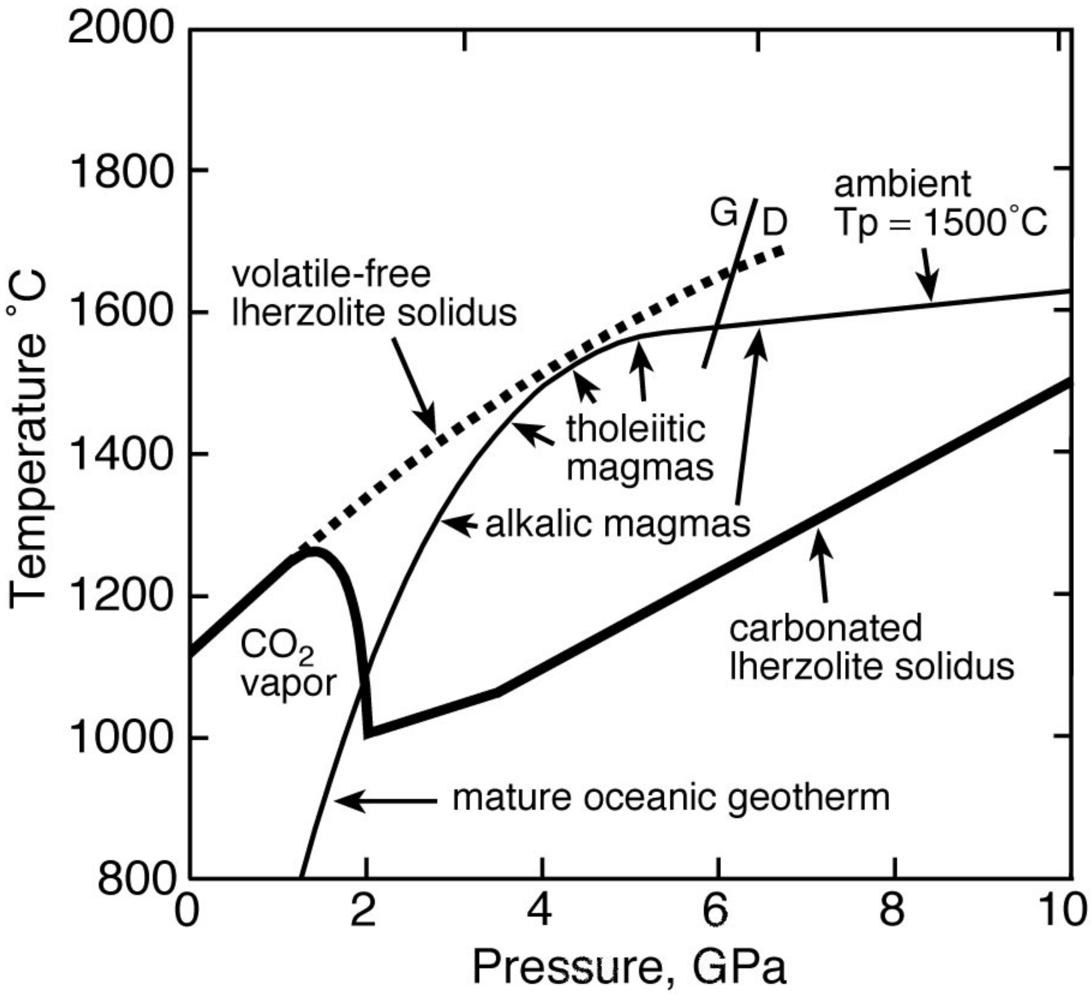

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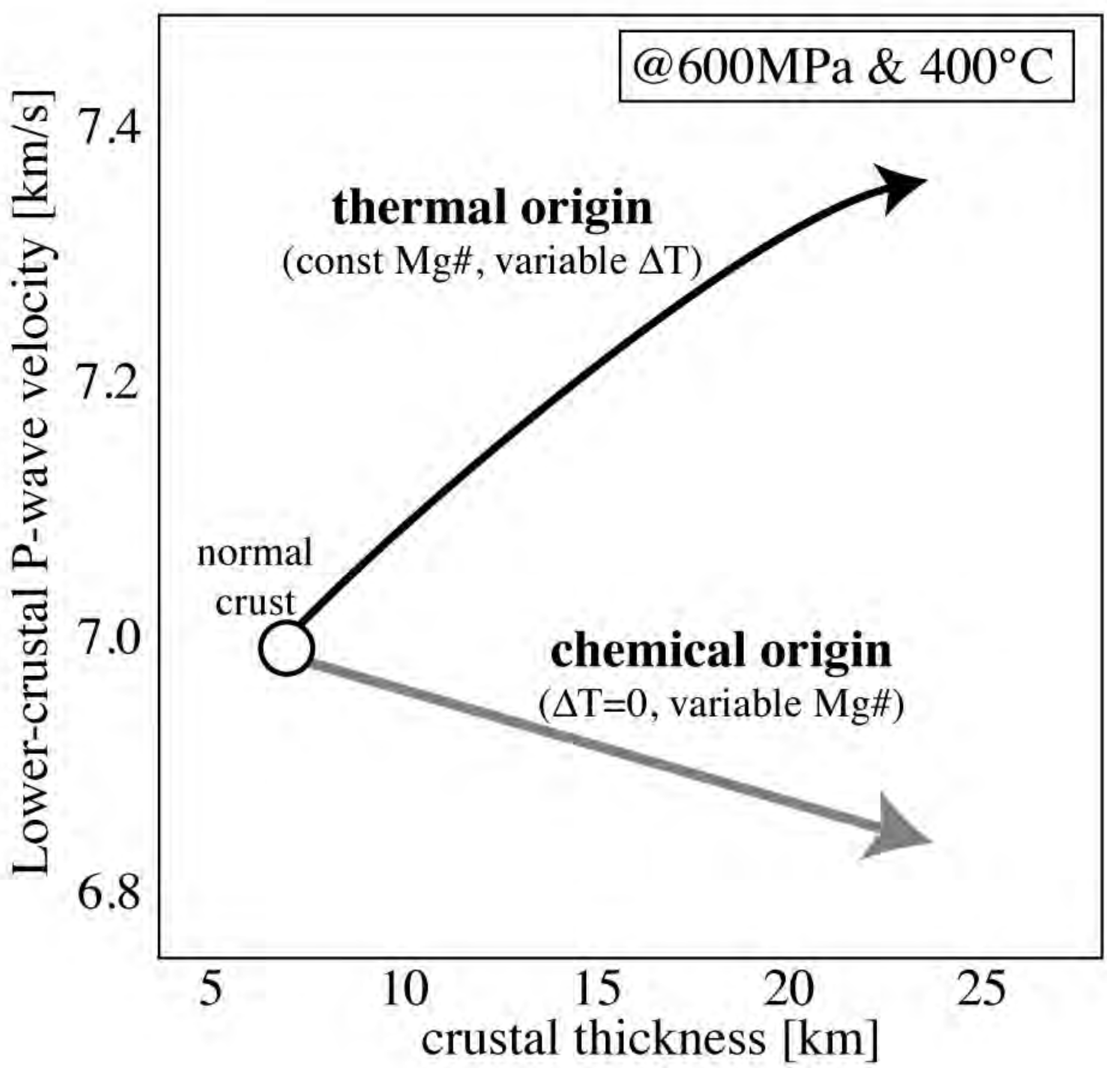

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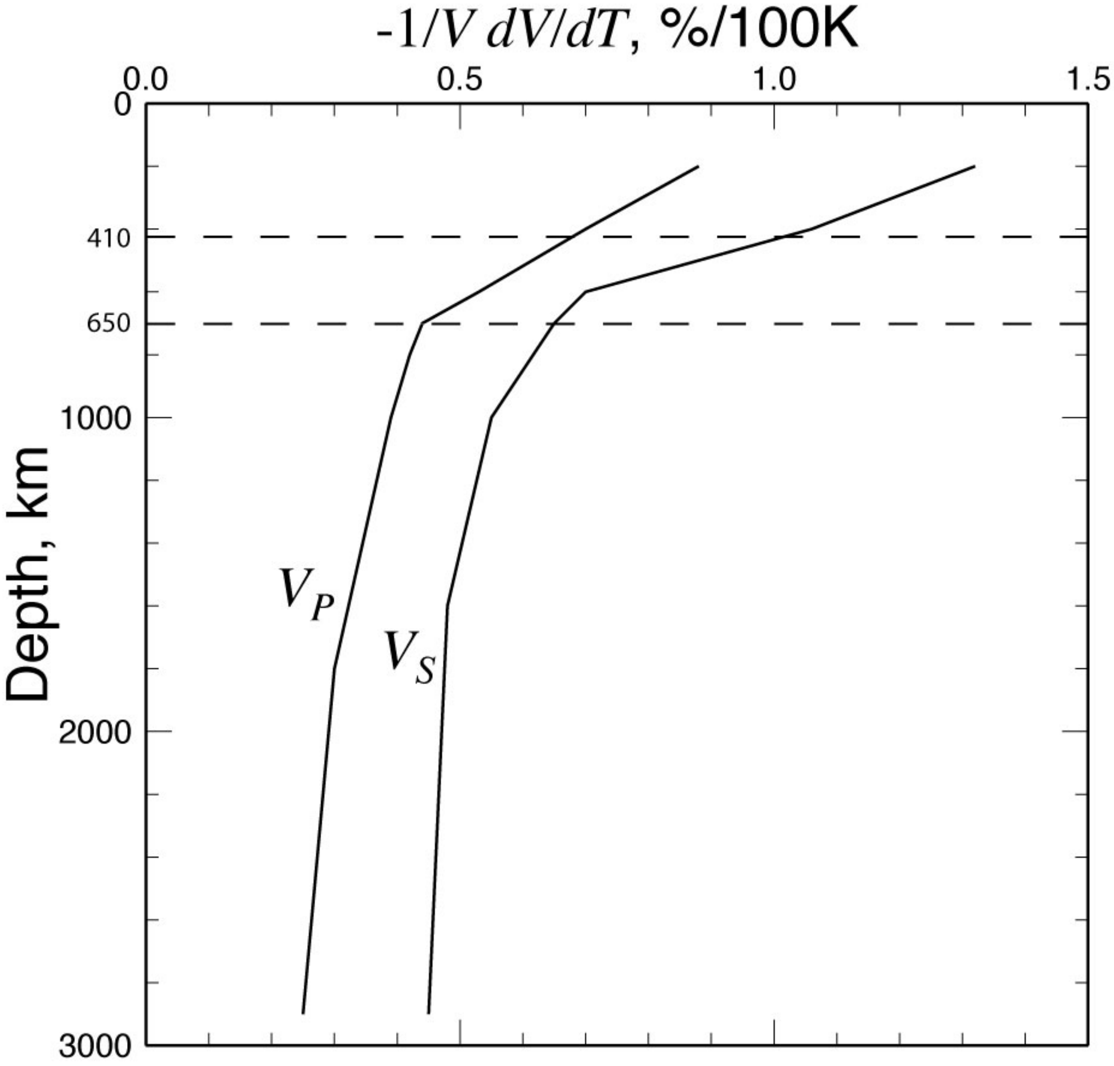

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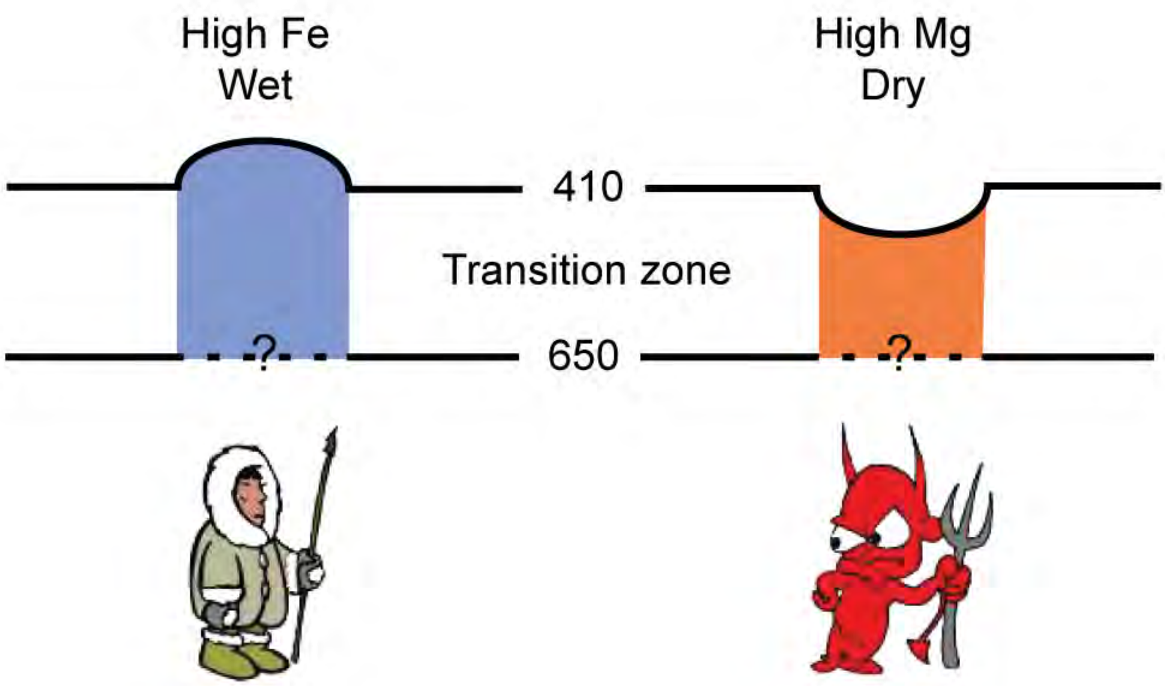

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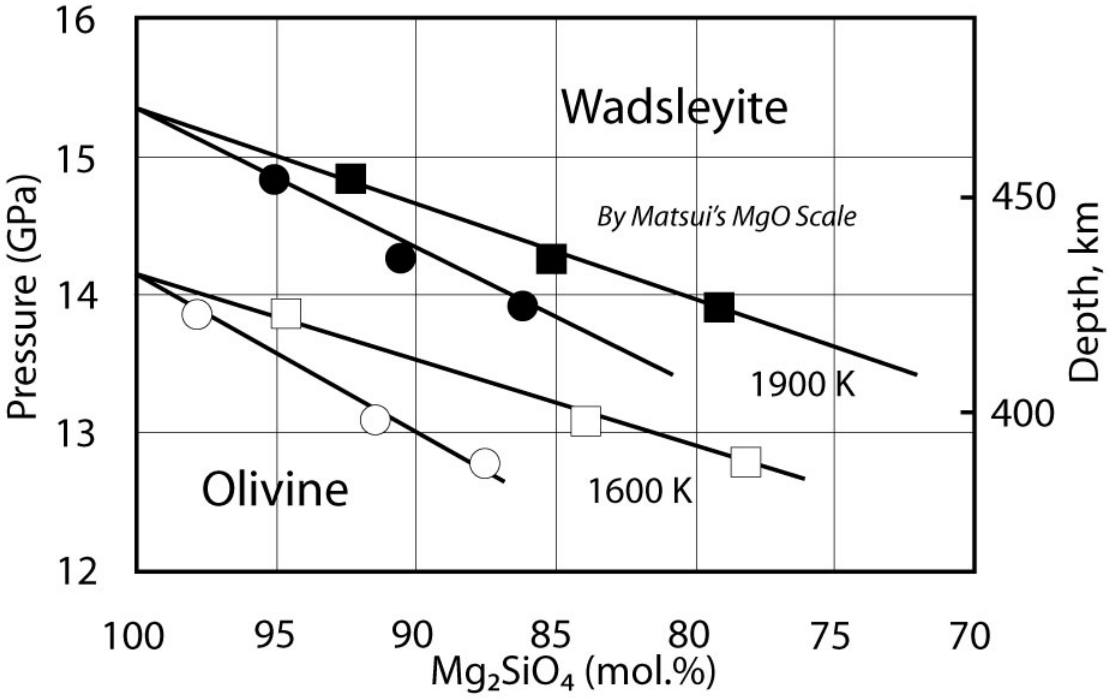

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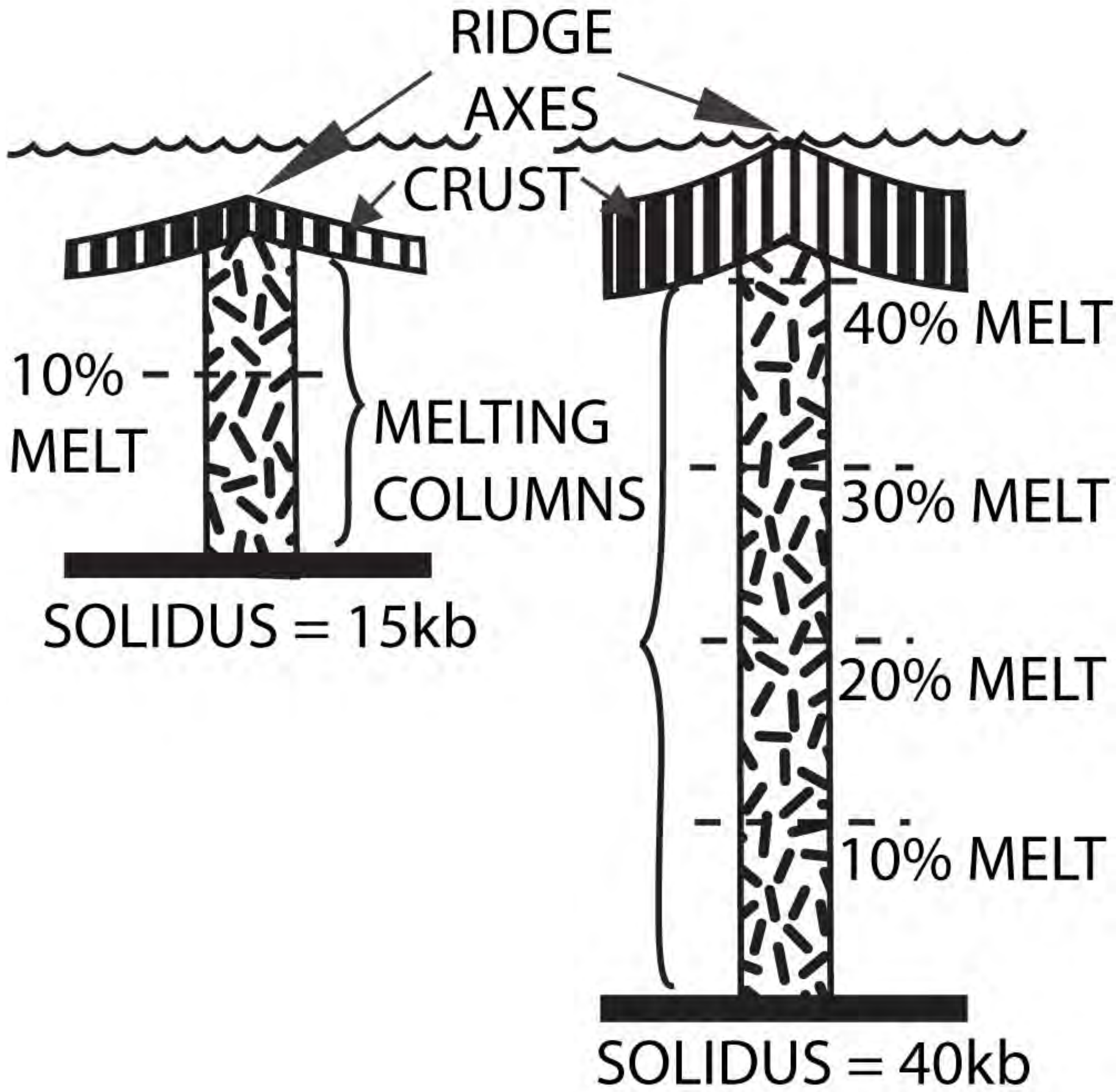

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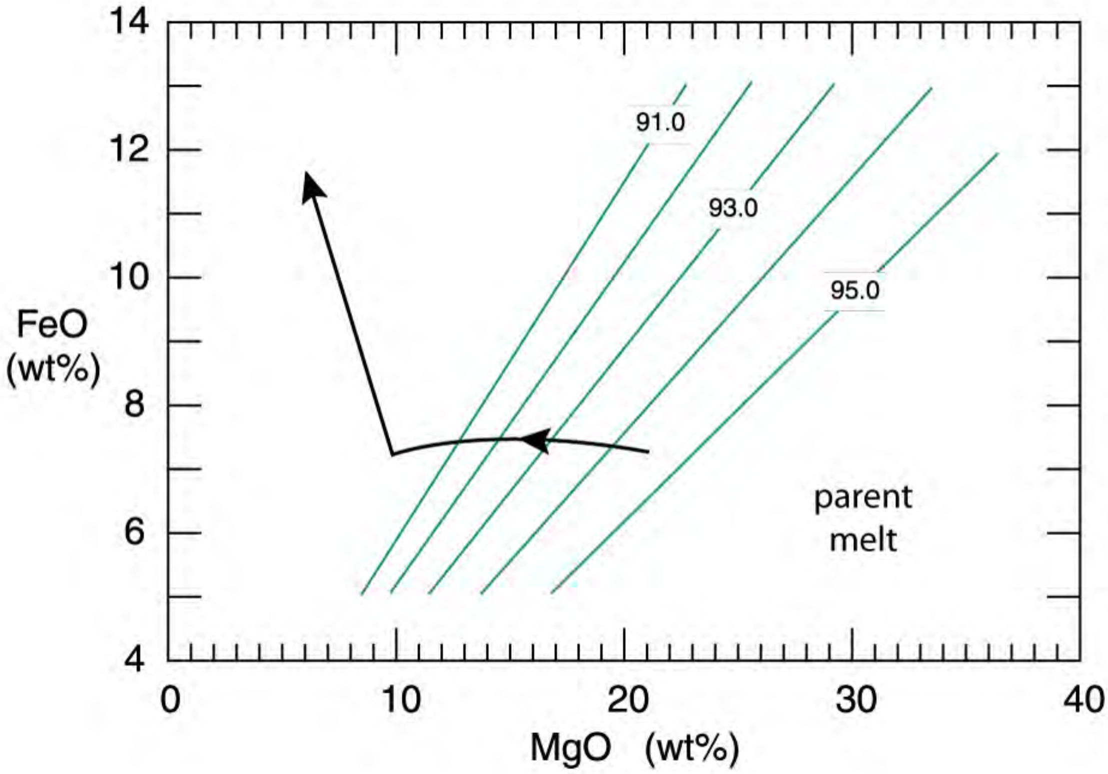

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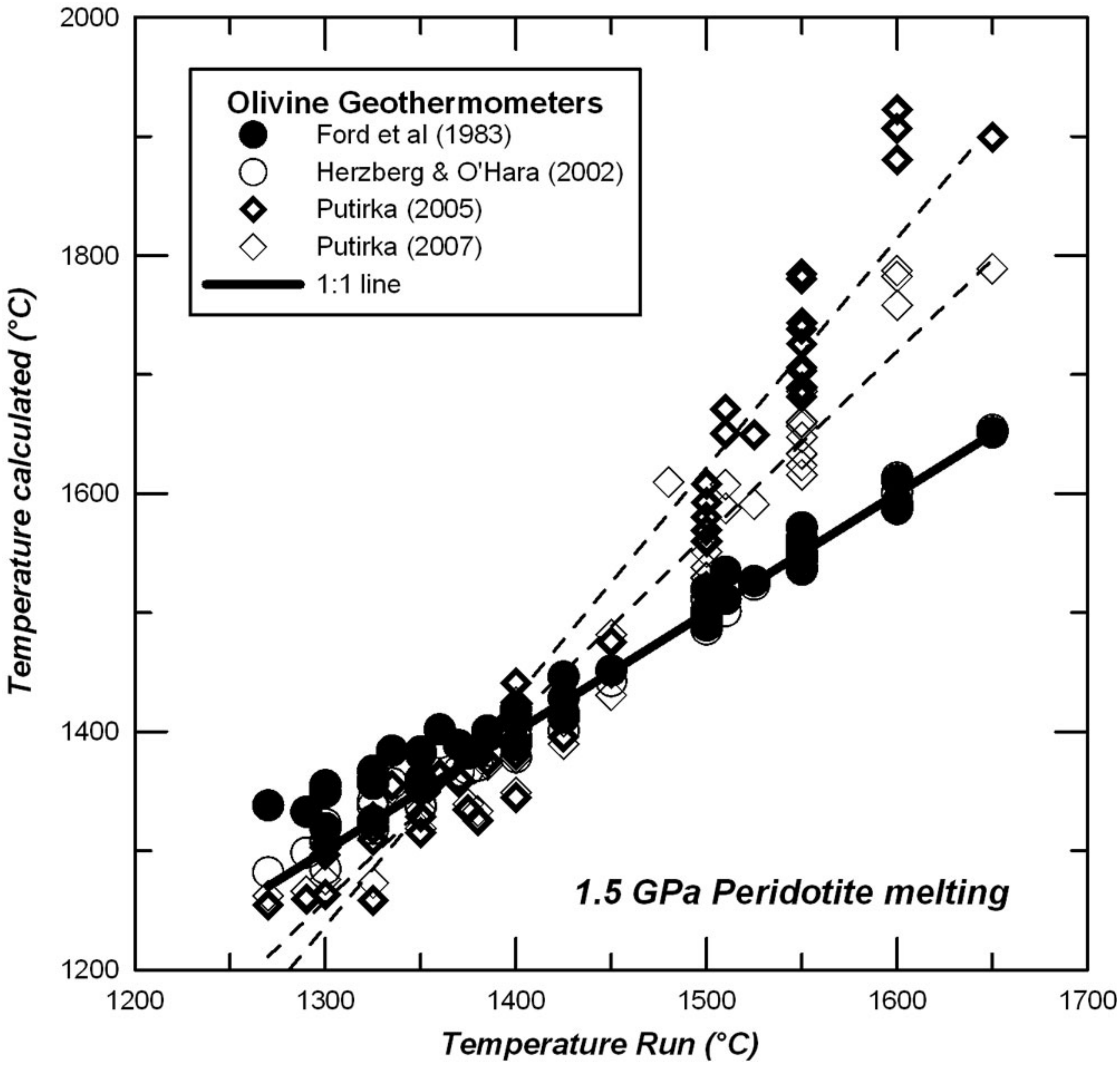

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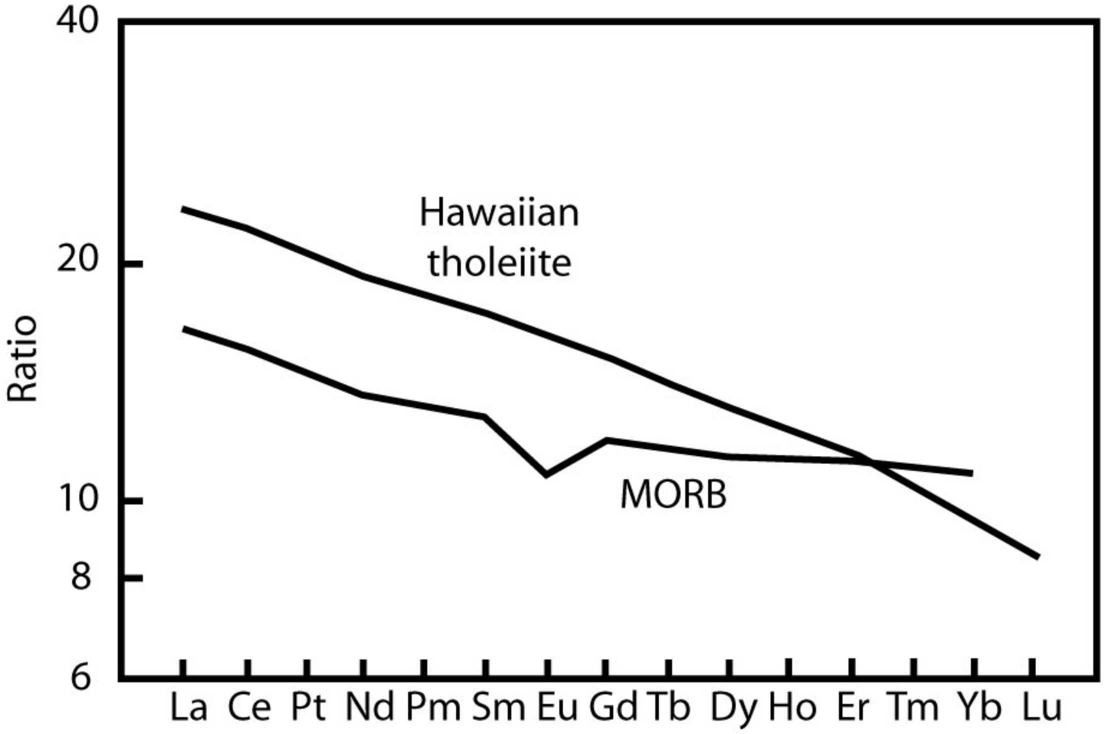

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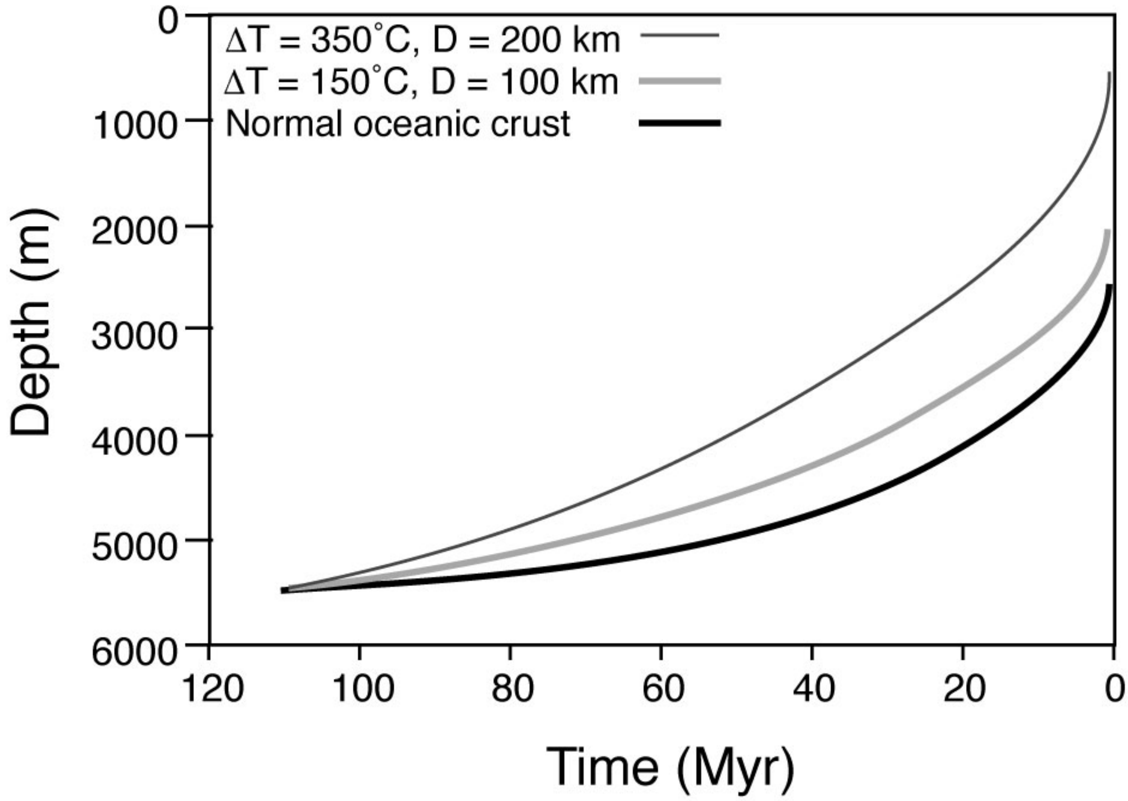

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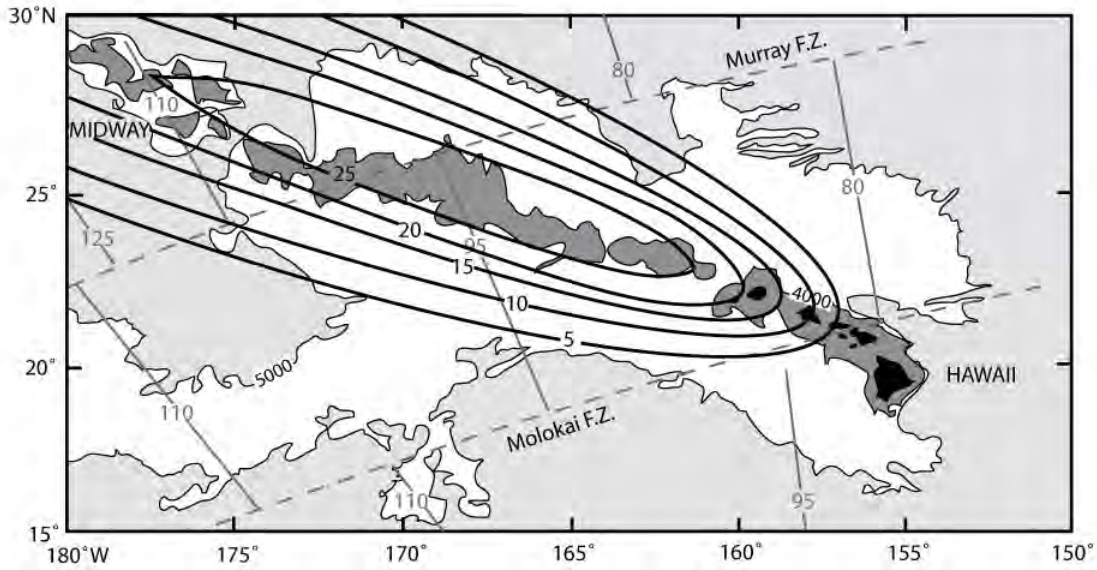

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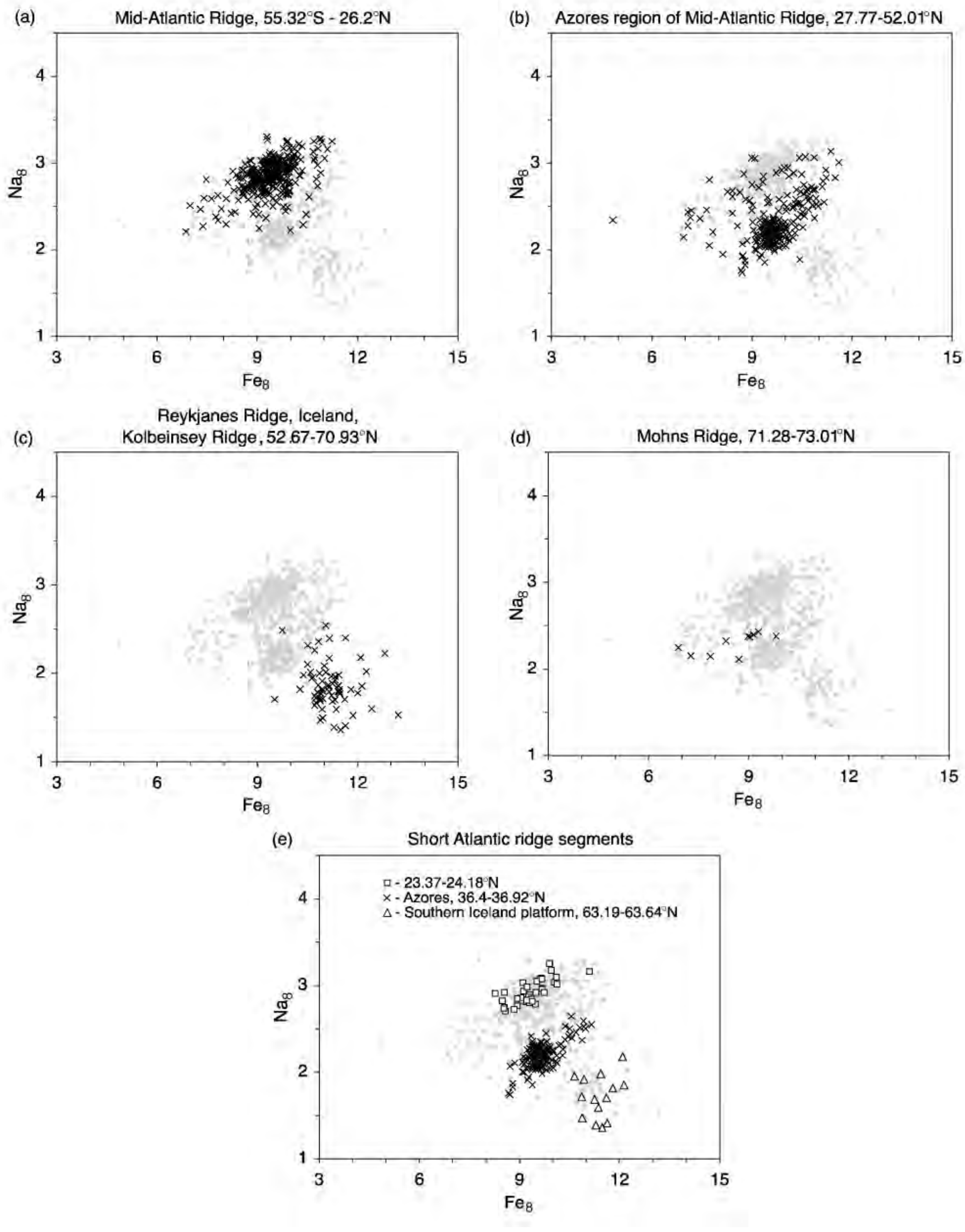

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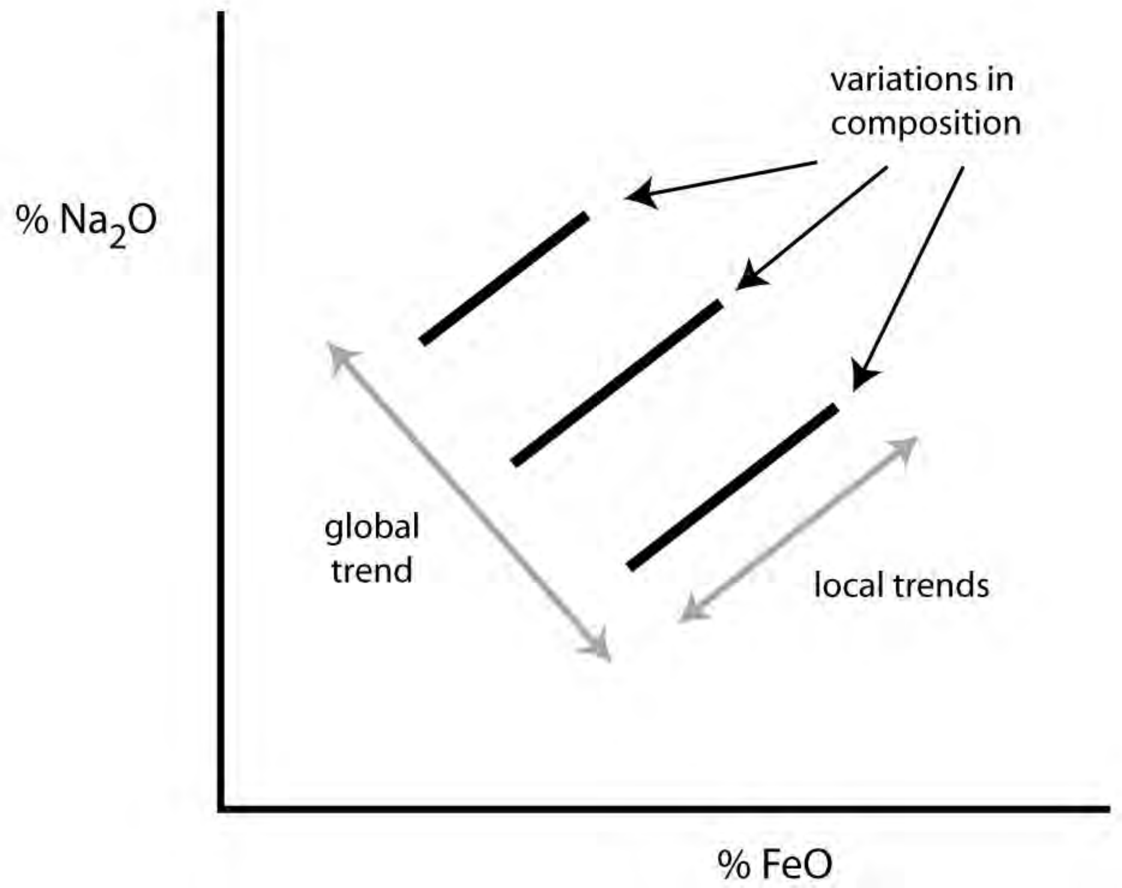

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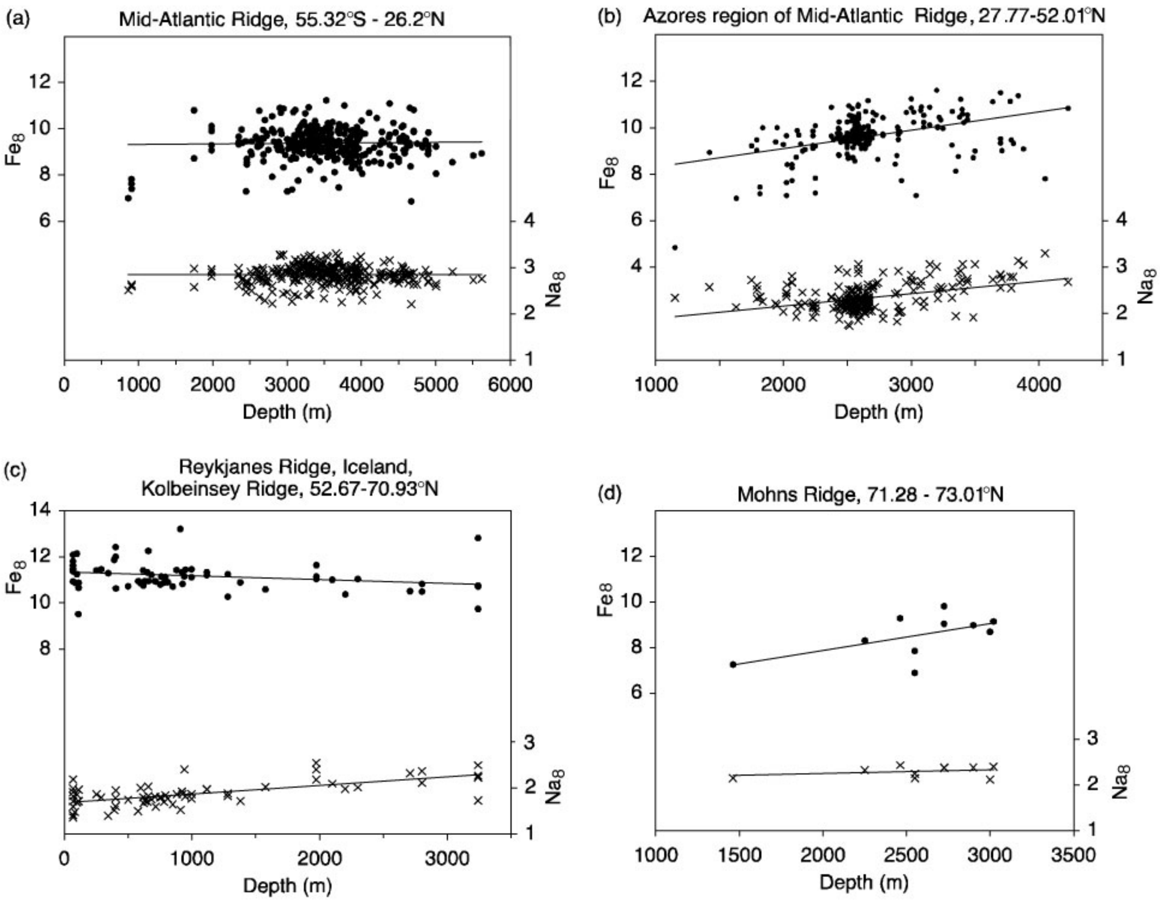

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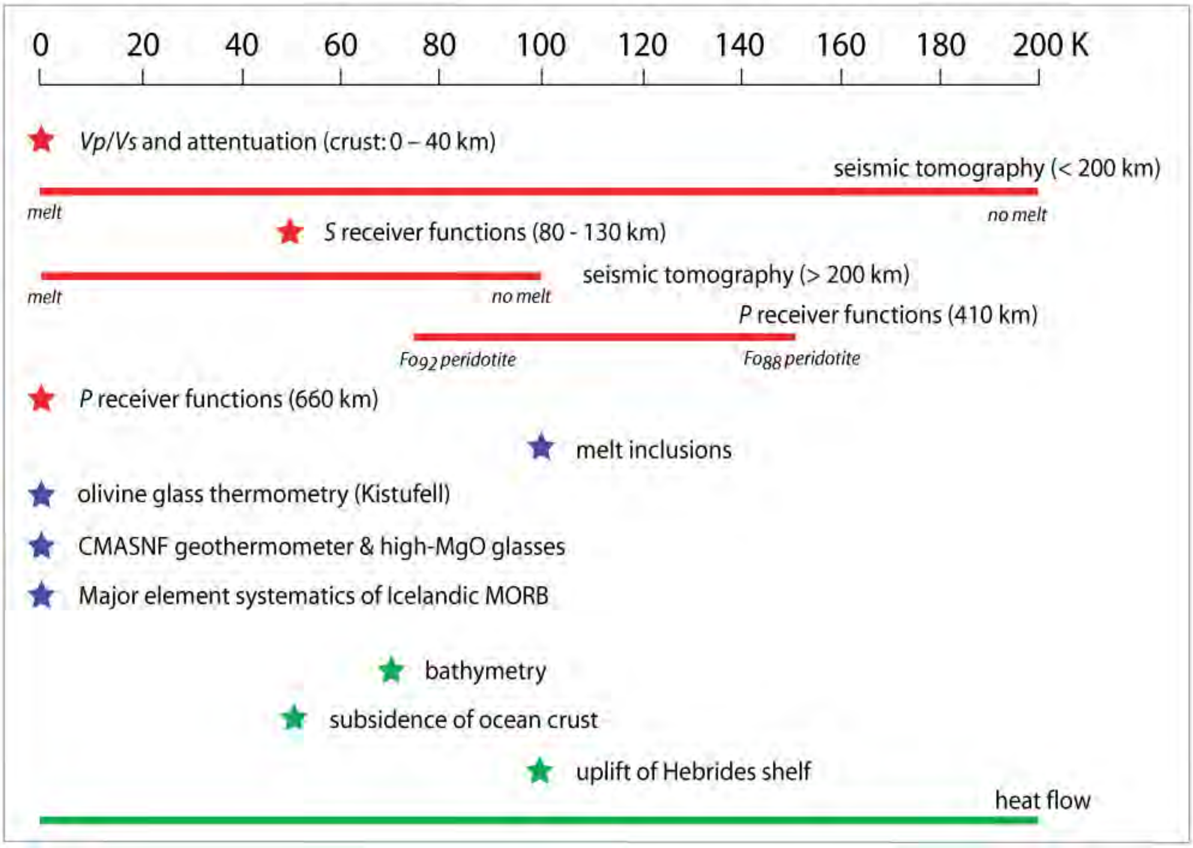

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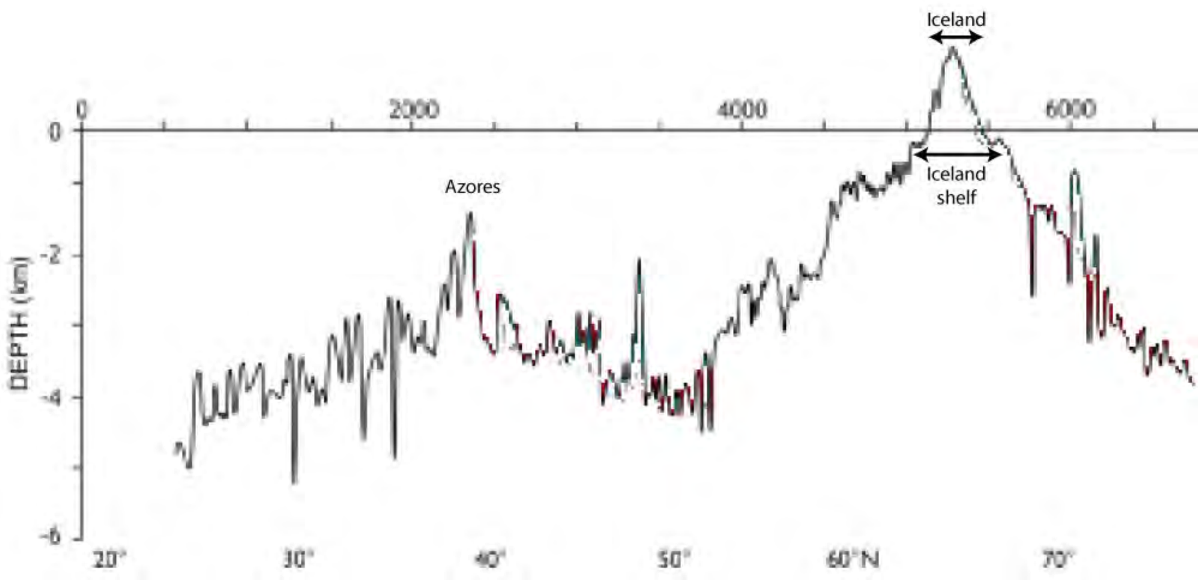

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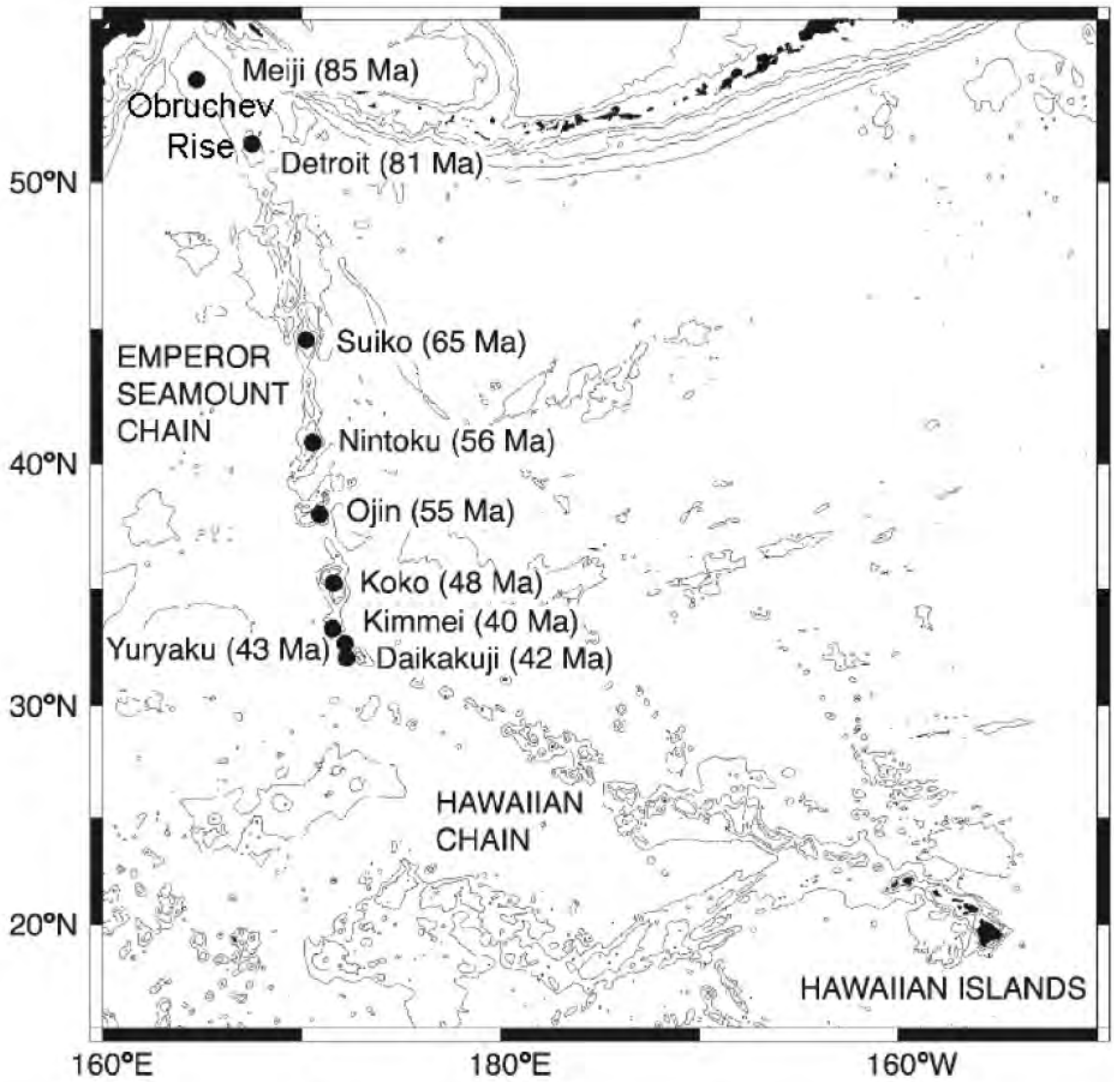

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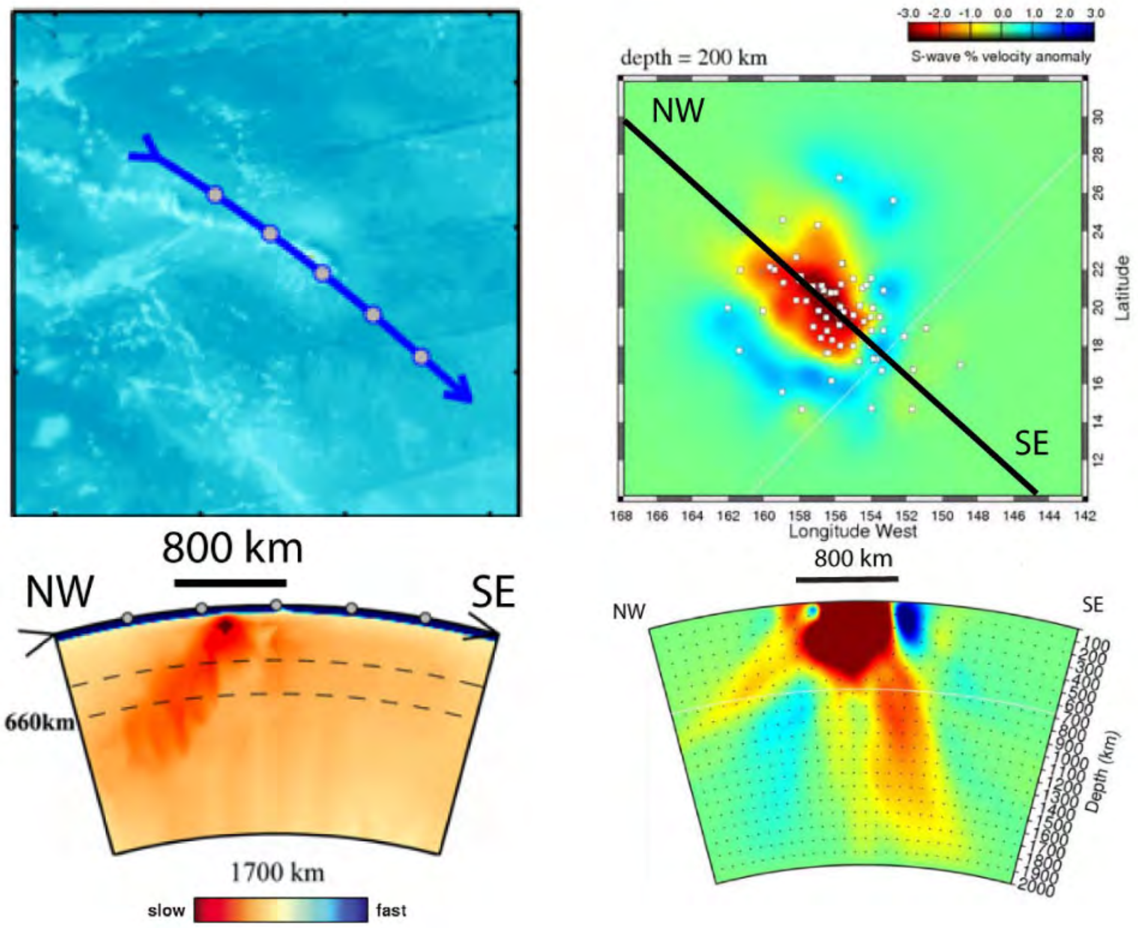

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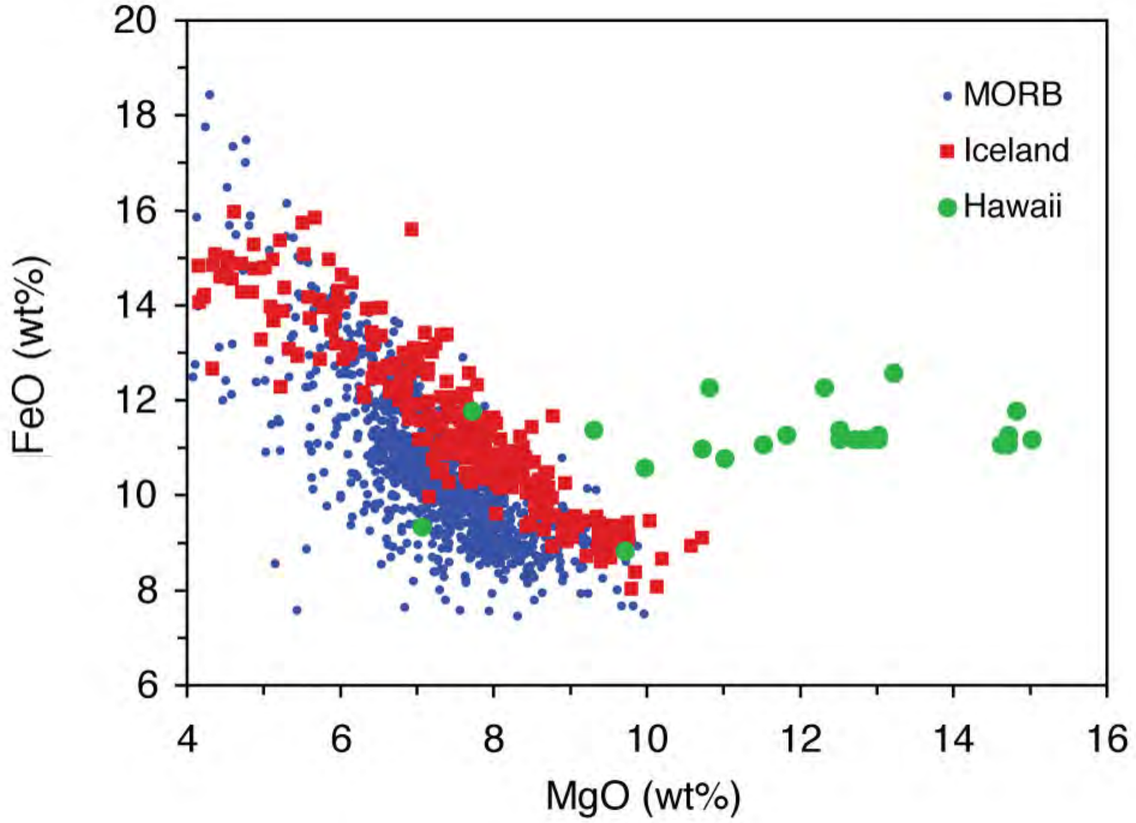

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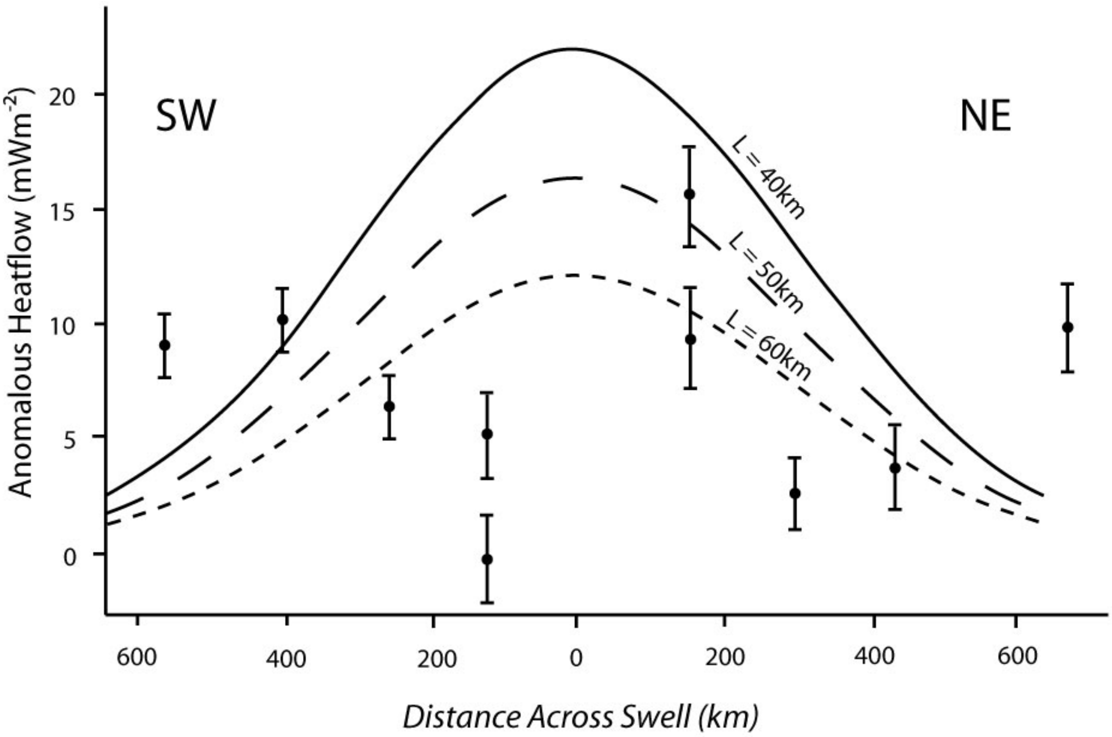

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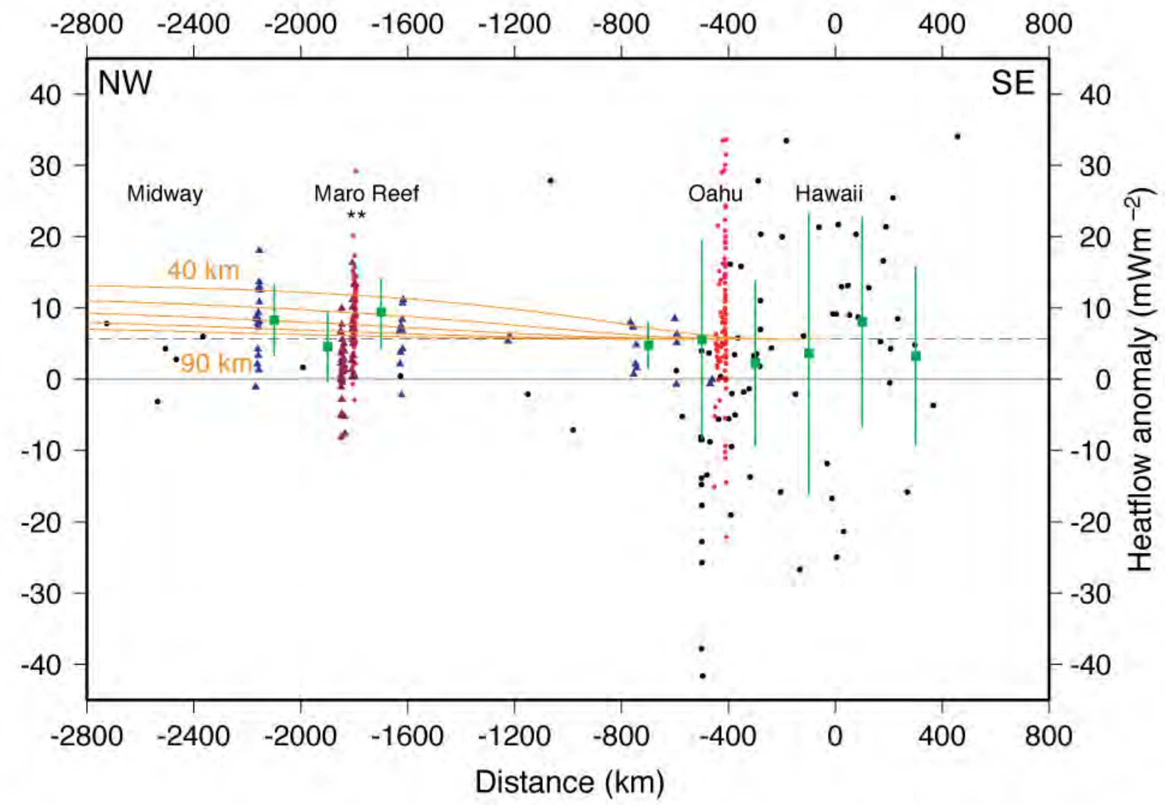

Figure 26 Heat flow data along the Hawaiian volcano chain. The heat flow anomalies are relative to the GDH2 average oceanic model (solid gray line) and to the average heat flow for lithosphere incoming from southeast of Hawaii (dashed gray line). Predicted heatflow (orange curves) for different depths of lithosphere reheating [from DeLaughter *et al.*, 2005].

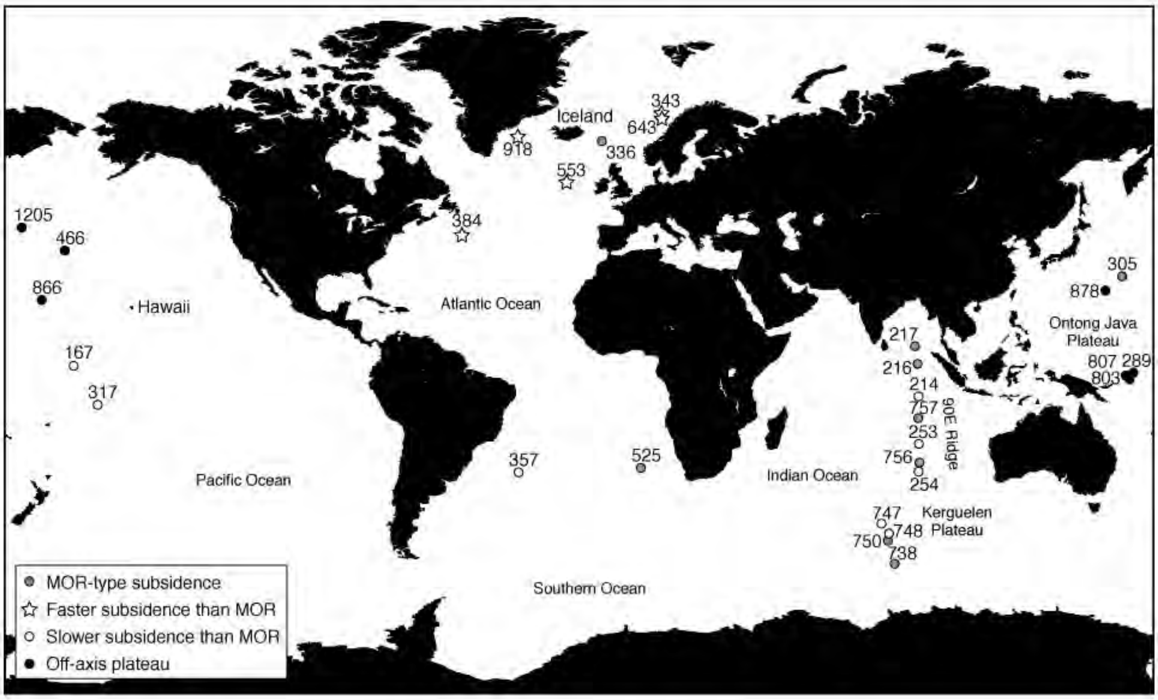

Figure 27 Deep Sea Drilling Project and Ocean Drilling Program drill sites where the temperature of the mantle has been estimated using subsidence histories from sedimentary cores. Faster subsidence than mid-ocean ridges (MORs) indicates higher initial mantle temperatures, and slower subsidence than MORs indicates lower initial mantle temperatures [from Clift, 2005].

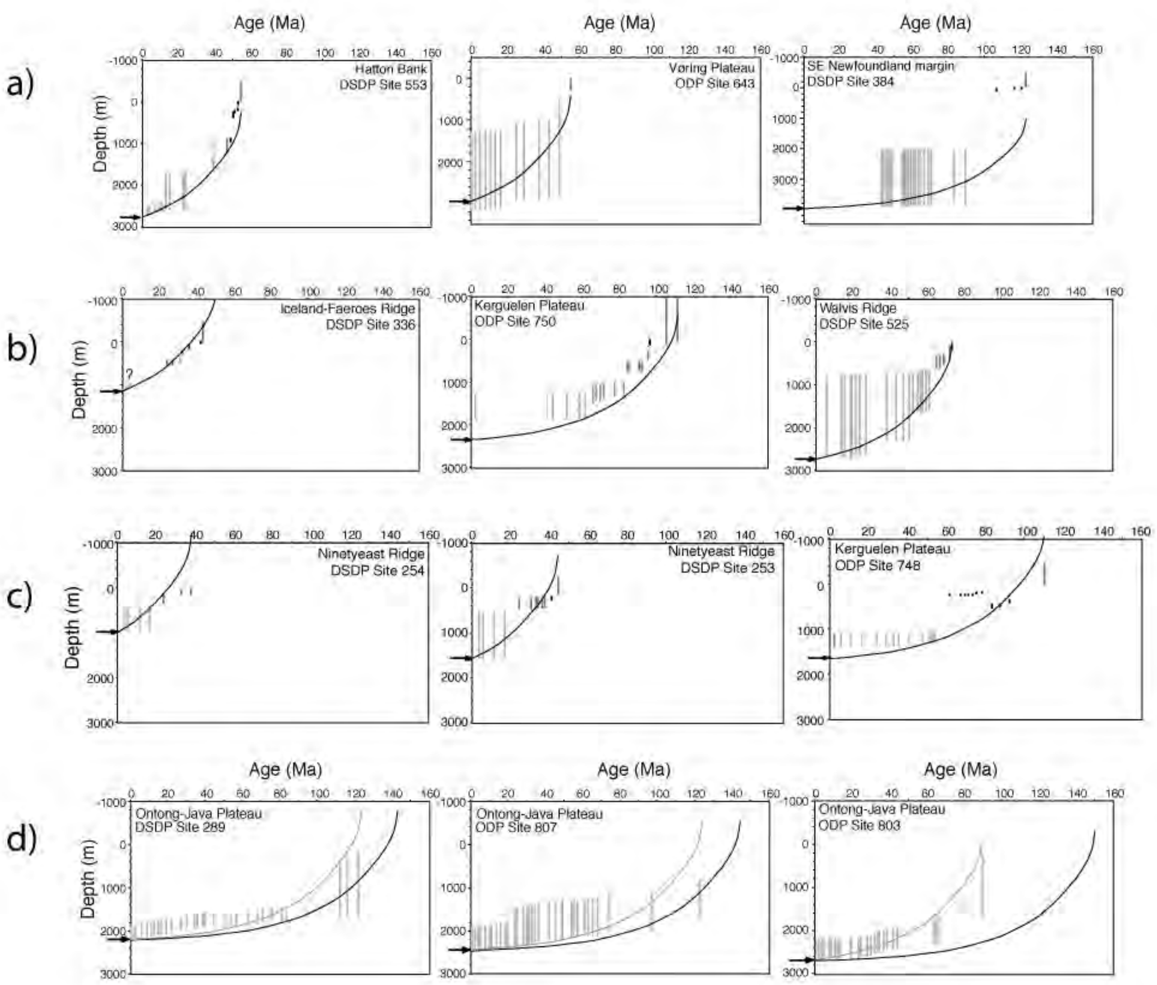

Figure 28 Reconstructed depth-to-basement histories for a selection of sites that subsided a) faster than average, b) at an average rate, and c) more slowly than average. Vertical bars show uncertainties in water depth estimates from sediments or microfossils. Black bars indicate high confidence, and gray bars low confidence. Solid curves shows the predicted subsidence history for normal oceanic crust. The end points of these curves are generally known accurately. d) shows sites where volcanism occurred away from a mid-ocean ridge. The grey curves show the predicted curve for crust of the same age as the volcanic rocks (equivalent to total thermal resetting of the lithosphere), and the black curves show predicted subsidence for crust of the age of the host crust [from Clift, 2005].

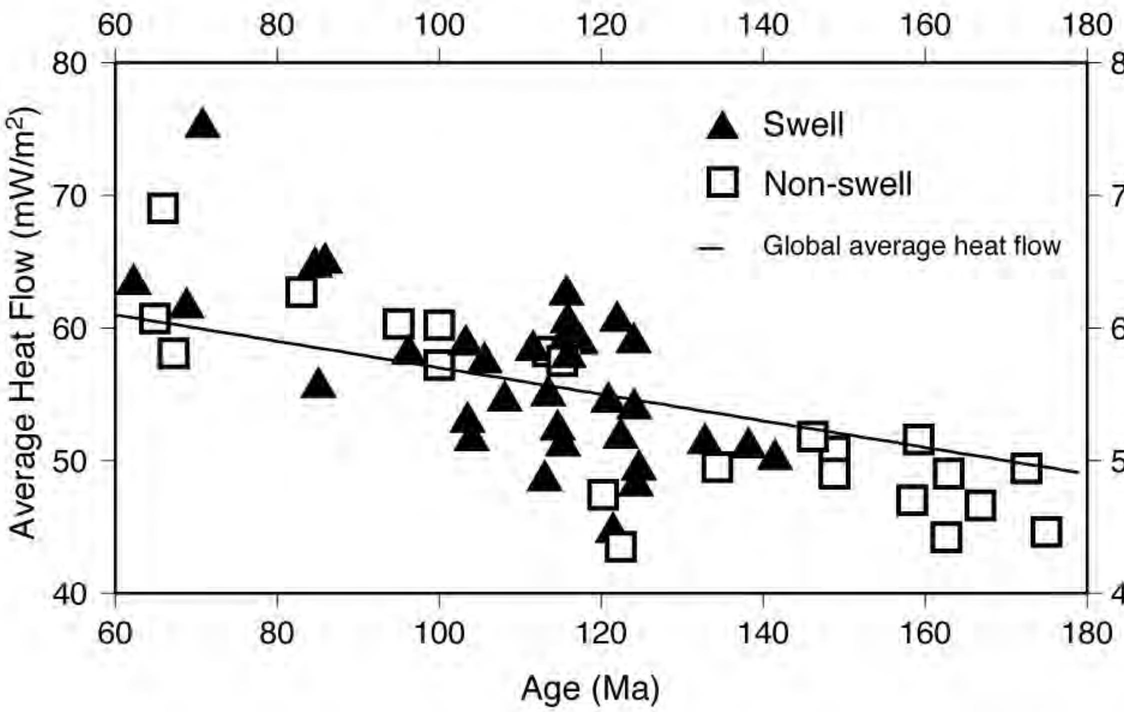

Figure 29 Pogo heatflow data vs. crustal age for the Hawaii, Réunion, Crozet, Cape Verde and Bermuda swells (triangles) and non-swells (squares). There is no significant evidence for elevated heat flow over swells [from Stein & Von Herzen, 2007].

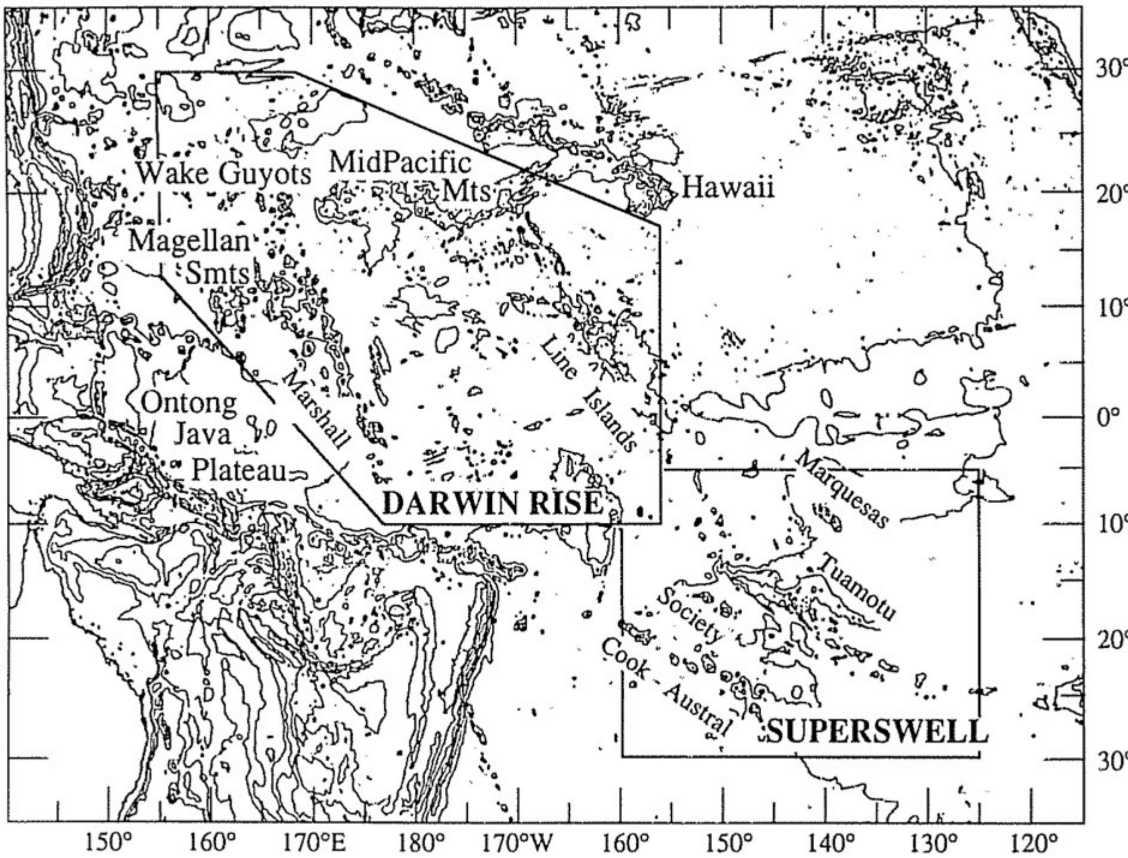

Figure 30 Map of the south Pacific showing the Darwin Rise and the Superswell [from Stein & Stein, 1993].

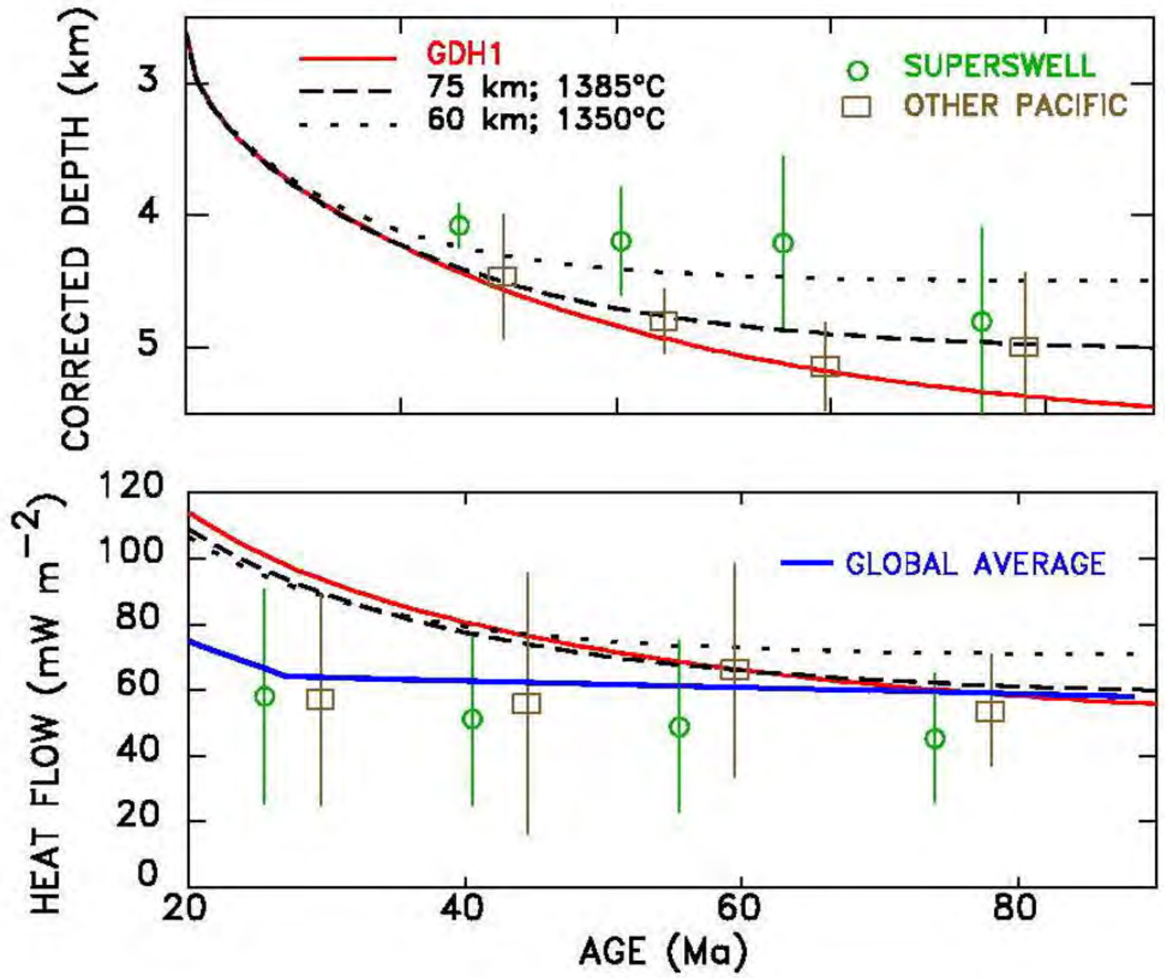

Figure 31 Top: Bathymetric depths from the Superswell and other regions of the same age in the Pacific for comparison. Bottom: heat flow observed, and predicted by thinning models [Stein & Stein, 1993].

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