

# *Origins of the plume hypothesis and some of its implications*

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## **ABSTRACT**

**Hotspots are regions of voluminous volcanism including Hawaii and Iceland. However, direct tests by mid-mantle tomography and petrology of lavas do not conclusively resolve that mantle plumes and mantle high temperature cause these features. Indirect and model-dependent methods thus provide constraints on the properties of plumes if they exist. Classical estimates of the excess (above MORB) potential temperature of the upwelling mantle (before melting commences) are 200-300 K. The estimate for Iceland utilizes the thickness of the oceanic crust. Estimates for Hawaii depend on the melting depth beneath old lithosphere. Flux estimates for Iceland depend on the kinematics of the ridge and those Hawaii or the kinematics and dynamics of the Hawaiian swell. These techniques let one compute the global volume and heat flux of plumes. The flux is significant, about 1/3 of the mantle has cycled through plumes extrapolating the current rate over geological time. On the other hand, obvious vertical tectonics may not be evident for many hotspots. For example, the relatively weak Icelandic hot spot (1/6 of the volume flux of Hawaii) would produce only modest volcanism and swell uplift if it impinged on the fast-moving Pacific plate or the fast-spreading East Pacific rise.**

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## **INTRODUCTION**

The Hawaii-Emperor seamount chain is a prominent feature volcanic in the middle of the Pacific plate. Iceland is a region of voluminous volcanism on the mid-Atlantic ridge. I spent considerable time early in my career attempting to find alternative causes to plumes for these features. By 1987, I had found my alternative hypotheses contrived and wanting. I acknowledge that there still are multiple viable alternatives to thermal plumes (e.g., Foulger et al., 2005a, 2005b) and that data sets that relate to one hypothesis may not bear on another. I do not attribute all off-axis volcanism to plumes. Lithospheric cracks may well tap low-volume source regions at normal mantle temperatures. Cracks are also necessary for plume magma to ascend in midplate regions.

Plumes meet the minimum requirement for a scientific hypothesis in that they putatively explain a class of observations on hotspots. They lead to testable predictions. Sleep (2006) reviewed deep aspects of the plume hypothesis starting with convection at the base of the lithosphere. I concentrate on geologically observable surface features in this paper. I first discuss a well-posed form of the plume hypothesis to obtain testable implications. Plumes, as I envision, are mainly thermal features from convection heated from below. Hot material rises buoyantly through the mantle through cylindrical low-viscosity conduits.

The deep part of the issue would be settled if one could resolve the structure at mid-mantle depths. Seismic data do provide evidence that such conduits exist in the expected places (Montelli et al., 2004). However, the features are near the limit of resolution (e.g., van der Hilst and de Hoop, 2005). These data are far from resolving whether the plumes are thermal or chemical features.

Petrological studies of lavas bear on the shallow part of the hypothesis. They have the potential to determine if the mantle beneath hotspots is in fact hotter than the mantle that which supplies most midoceanic ridge axes and whether the source regions are chemically different. Green and Falloon (2005), for example, state that there is no systematic difference in source temperature between MORB and hotspot lavas. Putirka (2005) contends that hotspot sources are in fact hotter than MORB source. I make no attempt in this paper to derive source temperatures from magma compositions.

In the absence of firm seismological and petrological evidence, earth scientists use data sets that indirectly constrain the properties of plumes, if they exist. In this paper, I apply heat and mass balance constraints to quantify the implications of the plume hypothesis. I center my arguments on the prominent hotspots, Iceland and Hawaii. I use a reliable data set, topography, as much as possible. Conversely, I point out lines of evidence that currently do not have the resolution to provide strong constraints.

## **THERMAL PLUME HYPOTHESIS**

I discuss plumes as thermal features to put the hypothesis in context. I start with melting temperature and melt volume as the hypothesis relates voluminous volcanism. I then discuss the flow of plume material along the base of the lithosphere with regard to Iceland and Hawaii. I discuss starting plume heads in the last part of this section.

### ***Adiabatic melting and mantle source temperature***

Plumes are supposedly hotter than the “ordinary” mantle that ascends at mid-oceanic ridges. The temperature difference is considered to be greater than the ambient temperature variations in the rest of the asthenosphere. That is, the hypothesis implies one can to the first order partition temperature variation into “normal” mantle near the MORB adiabat, a plume adiabat, and sinking slabs.

I present a linearized version of pressure-release melting to obtain semi-schematic diagrams and to consistently define terms. The fraction of melting  $f$  depends on the depth  $Z$  and temperature  $T$  (Figure 1). Such petrological grids are widely used and can be constructed for any putative mantle composition (Klein and Langmuir, 1987; McKenzie and Bickle, 1988; White et al., 1992; Plank and Langmuir, 1992; Klein, 2003).

Using potential temperature referenced to the surface on the grid results in some simplification,

$$T_p \equiv T - \gamma Z, \quad (1)$$

where  $\gamma$  is the solid adiabatic gradient. I use a linear expression for the fraction of melting,

$$f = \frac{T_p - T_0 - \frac{\partial T_m}{\partial Z} Z}{T_1}, \quad (2)$$

where  $T_0$  is the solidus at the surface,  $T_1$  the temperature difference between the solidus and the liquidus, and  $\partial T_m / \partial Z$  is the gradient of melting temperature with depth.

Ascending mantle material is initially solid with a potential temperature  $T_s$ . It starts to melt at a depth

$$Z_s \equiv (T_s - T_0) \left( \frac{\partial T_m}{\partial Z} \right)^{-1}. \quad (3)$$

A balance between specific heat and latent heat determines the temperature within an adiabatic column above this depth

$$\rho C (T_s - T_p) = f \rho L, \quad (4)$$

where  $\rho$  is density,  $C$  is specific heat per mass, and  $L$  is latent heat per mass. One solves (2) and (4) for the temperature and the melt fraction

$$f = \frac{T_s - T_0 - \frac{\partial T_m}{\partial Z} Z}{T_1 + \frac{L}{C}} \quad (5)$$

(Figure 2). The ratio of latent heat to specific heat  $L/C$  has dimensions of temperature.

Finally, the total thickness of segregated melt from the column is

$$Z_c = \int_{Z_{\text{top}}}^{Z_s} \frac{f}{1-f} df, \quad (6)$$

where  $Z_{\text{top}}$  is the depth to the top of the column and the factor  $1/(1-f)$  accounts (to the first order) for the geometrical effect that the melt that segregates and ascends is replaced by more ascending material (Figure 2). At sufficiently fast-spreading ridge axes, the base of the oceanic crust defines the top of the column. That is,  $Z_c = Z_{\text{top}}$ . Elsewhere, the thermal (or perhaps chemical) base of the lithosphere defines the top.

Figures 1 and 2 use parameters calibrated to give 6-km thick crust at “normal” ridge axes where the source potential temperature is 1300°C. The surface solidus is 1132°C. The melting point gradient is 3 K km<sup>-1</sup>. The temperature  $L/C$  and the temperature  $T_1$  are both 365.5°C. The solid adiabatic gradient is 0.3 K km<sup>-1</sup>. Melting begins at 56 km depth at normal ridge axes.

I obtain example classical estimates for the source temperature differences for Iceland and Hawaii from the petrological grid for Iceland and Hawaii. They do not involve detailed petrology. They effectively give the average temperature anomaly of the region in the core of the plume that actually melts. (Plank et al. (1995) discuss the semantics and systematics of the average fraction of melting in a column.) I use a rounded estimate of 26 km for Iceland. (This is 20 km thicker than normal crust. It is necessary to average over the edifice because plume material melts within the rising plume and then spreads laterally.) This implies that the Icelandic source region is 200 K hotter than normal mantle. Voluminous melting beneath Hawaii begins at a rounded depth of 116 km (60 km greater than normal mantle). This implies an excess temperature of 180 K. Analysis using more sophisticated petrological grids yield somewhat higher excess temperatures. For example, McKenzie and Watson (1991) obtain ~280 K from numerical models of melting beneath Hawaii. Sobolev et al. (2005) consider the detailed sequence of melting beneath Hawaii. Recycled oceanic crust melts first and reacts with surrounding peridotite to form pyroxenite. The pyroxenite then melts to form voluminous Ni-rich magmas. They obtain an excess temperature of 250-300 K.

I reiterate the assumptions for the petrological grid to introduce alternatives to thermal plumes. (1) The ascending material is geometrically that needed to make the oceanic lithosphere. The velocity is upward and no material recirculates. (2) Much of the melt segregates and ascends. (3) The material is homogeneous and the composition is the same beneath hotspots and normal ridges. The validity of these assumptions and alternatives are potentially detectable from petrological studies.

Edge-driven convection could conceivably recirculate the material beneath the ridge axis leading to much more melting than implied by the model (e.g., King and Anderson, 1998). For example, 26-km-thick ocean crust beneath Iceland could form a hot adiabatic column as assumed in model. It could conceivably form by recirculating a column of normal-temperature mantle (that produces 6 km of crust)  $26/6 = 4.3$  times. The composition of the melt in the two cases should differ (e.g., Plank et al., 1995).

The second assumption is valid for mass-balance calculations as long as most of the melt segregates and ascends. One cannot appeal to inefficient melt segregation to explain the difference between Iceland and normal ridges. Using the example thickness of 6 and 26 km, 20 km of melt would have to be retained in the mantle beneath normal ridges. If it were distributed over the upper 100 km of the lithosphere, there would be 20% partial melt. This feature would be obvious from seismology.

The third assumption is obviously incorrect in detail, radiogenic isotopes indicate that the mantle is heterogeneous and has been so for billions of years. Still one can construct a petrological grid for each heterogeneous domain. The heat balance in (4) still applies in an integral sense over the domains. Small heterogeneities melt over different depth intervals as they ascend. Latent heat cools the first domains that melt and heat flows in from the surrounding unmelted regions (Sleep, 1984). This enhances the fraction of melting where it is already occurring and suppresses melting elsewhere.

Major element (including water and maybe CO<sub>2</sub>) variations from normal source composition are needed to produce voluminous volcanism at normal mantle temperatures. I discuss three useful end members. First as already noted, a small easily melted domain does not affect the heat balance. The surrounding material maintains its temperature at the solid adiabat. This effect explains low-volume hydrous magmas but not voluminous volcanism.

Second, I consider that the ascending mantle is a eutectic composed of the minimum melting material on the grid. The parameter  $T_1$  is then 0 rather than about  $L/C$ , which is  $\sim 1$  and  $T_0$  is unchanged. This modification has the effect of doubling the melt production in (5). That is, 26-km-thick crust cannot be obtained from normal temperature material with the properties of the near-solidus melts at normal ridges.

Third, the solidus of the material may be less than the ridge solidus. In terms of the grid, one may reduce the surface solidus  $T_0$  by  $\sim 200$  K rather than increasing the source potential temperature  $T_s$  by that amount and leave  $T_1$ .

Finally to appraise the thermal plume hypothesis, one would like to determine the source potential temperature  $T_s$  of the material before significant melting begins. However, complicated magma processes at depth make this estimate far from straightforward. MORB is the partly pooled series of melts from the adiabatic column. Melt ascending on the liquid adiabat is superheated and may assimilate mantle wall rock. The melts pool within the axial magma chamber and fractionally crystallize. Midplate source regions may cool by conduction into the overlying lithosphere and melts may have complex histories within deep and high-level magma chambers.

Careful sampling does show some enriched MORB samples away from hotspots on the slow spreading mid-Atlantic ridge as expected if local heterogeneities are ubiquitous (Donnelly et al., 2004). Many low-volume magmas do not reach the surface. They freeze within the mantle producing new heterogeneities (Donnelly et al., 2004). These domains are sources of enriched MORB when they remelt beneath ridge axes. Radiogenic isotopes indicate they have  $\sim 300$  m.y. survival times. Off-axis, these domains are likely sources for low-volume magmas that are not associated with plumes.

### *Iceland*

I apply the petrological grid formalism to obtain the flux of the on-ridge hotspot Iceland and then discuss some geometrical complications. The plume supplies material that melts and eventually forms lithosphere along a length  $L$  of the axis (Fig. 3). That is, the volume flux is

$$V = U_R L D, \quad (7)$$

where  $U_R$  is the full spreading rate of the ridge,  $L$  is the length of ridge axis supplied by the plume, and  $D$  the thickness of lithosphere formed from plume material. The thickness  $D$  is  $\sim 100$  km the depth where voluminous melting begins. Petrology and the thermal thickness of the lithosphere away from the axis provide further constraints (Schilling, 1991).

About a 1000-km length of ridge is involved (Fig. 3). The aseismic Iceland-Greenland and Iceland-Faeroe ridges extend to the adjacent continental margins. The full spreading rate is  $\sim 20 \text{ mm yr}^{-1}$ . This gives a volume flux of  $63 \text{ m}^3 \text{ s}^{-1}$ .

The geometry of mantle flow beneath Iceland presents complications in determining the source temperature from rocks. Plume material ascends beneath the hotspot and flows laterally long the ridge axis (e.g., Albers and Christensen, 2001). The entire hotspot is to the first order the product of an adiabatic column, but the local melts pooled in a crustal magma chamber are not. For example, the mantle material at the distal ends of the hotspot (the Reykjanes ridge on the south) has already partially melted in the plume conduit.

### *Hawaii*

The concept of buoyancy flux is relevant to off-ridge hotspots. In terms of the plume, the buoyancy flux is

$$B = \Delta\rho V, \quad (8)$$

where  $\Delta\rho$  is the density contrast of the plume material with normal mantle and  $V$  is the volume flux through the plume.

One obtains this quantity indirectly from “swells” along hotspot tracks. Hawaii is the best example. The swell is over 1000 km wide and over a kilometer high along the axis of the chain (Fig. 4). The measured buoyancy flux from the swell is

$$B = (\rho_m - \rho_w)U_pWH, \quad (9)$$

where  $\rho_m$  is the density of the mantle,  $\rho_w$  is the density of ocean water and  $U_p$  is the velocity of the plate relative to the hotspot. I express the cross sectional area of the swell compactly as the product of the width  $W$  with the height  $H$ . Estimation of these quantities requires correction of relief associated with the volcanic edifice and its flexural moats. In addition, the plume material is fluid, not a rigid layer attached to the plate. The estimate in (9) thus does not precisely give the desired plume quantity in (8) (Ribe and Christensen, 1994).

The Hawaii-Emperor chain is the clearest hotspot track (Fig. 4). Any hypothesis to explain this feature must account for the Hawaii swell as well as volcanism along the axis of the chain. The swell is a major submarine uplift. Its full width is 1500 km and its maximum height is 1.4 km. Its cross sectional area is  $1400 \text{ km}^2$  (Sleep, 1992). The plate velocity is  $83 \text{ mm yr}^{-1}$ , giving a swell buoyancy flux of  $8.9 \text{ Mg s}^{-1}$ .

To compare volume flux and buoyancy flux, I represent the density change in terms of the temperature change

$$\Delta\rho = \rho\alpha\Delta T, \quad (10)$$

where  $\rho$  is the density of average mantle ( $3400 \text{ kg m}^{-3}$ ),  $\alpha$  is the volume thermal expansion coefficient ( $3\times 10^{-5} \text{ K}^{-1}$ ), and  $\Delta T$  is the temperature contrast (here  $\sim 250 \text{ K}$ )

between the plume material and the normal mantle. This gives the volume flux for Hawaii of  $350 \text{ m}^3 \text{ s}^{-1}$ . It is also useful to have the excess heat per volume of material is

$$E = \rho C \Delta T, \quad (11)$$

where  $C$  is specific heat per mass. The ratio of density change to excess heat,

$$\frac{\Delta \rho}{E} = \frac{\alpha}{C}, \quad (12)$$

involves only measurable physical parameters. The uncertainty in these parameters is much less than the uncertainty in volume or buoyancy flux estimates.

The “nose” of the swell extends upstream from the hotspot by about 400 km (Fig. 4). I initially investigated its kinematics to disprove the involvement of a plume. However, a plume readily explains the existence and shape of the swell. I discuss both a kinematic model and a more sophisticated dynamic model.

The kinematic model illustrates the geometry (Sleep, 1990a, 1992). Close to the hotspot, plume material flows radially into the asthenosphere from the hotspot. Plate motion sweeps the plume material and the asthenosphere downstream. The depth-average velocity of the ponded plume material is

$$\bar{U} = \frac{V}{2\pi Ar} \hat{r} + \frac{U_p}{2} \hat{x}, \quad (13)$$

where  $A$  is the thickness of the layer of spreading plume material,  $r$  is horizontal distance from the center of the plume,  $\hat{r}$  is the radial unit vector, and  $\hat{x}$  is the unit vector in the downstream direction. The factor of 2 in the second right-hand term arises because the velocity from drag of the plate goes from plate velocity at the top of the plume material to near the hotspot frame at its base. A stagnation point where the radial velocity from the plume balances the velocity from the drag of the plate occurs upstream where

$$s = \frac{V}{\pi A U_p}. \quad (14)$$

The parameters in the equation are grossly known. The volume flux is  $350 \text{ m}^3 \text{ s}^{-1}$ , the thickness of flowing material is 100 km, and the plate velocity is  $83 \text{ mm yr}^{-1}$ . This yields a distance from the nose to the hotspot of 425 km, which is acceptable. The streamline through the stagnation point separates material supplied by the plume and normal asthenosphere. It approximately parabolic shape fits the swell particularly to the north (Fig. 4).

Ribe and Christensen (1994) considered buoyant fluid ponded at the base of the moving plate. This dynamic modeling better represents the physics of the swell. I retain the thickness  $A$  of the plume material and the distance to the nose of the swell  $s$  as dimensional parameters. I let the base of the lithosphere be flat (see Sleep (1996) for a discussion of the effect of relief at the base of the lithosphere). The flux (volume per horizontal length) of material away from the plume is then dimensionally

$$F \approx \frac{\partial A}{\partial x} \frac{\Delta \rho g A^3}{\eta_p} \approx \frac{\Delta \rho g A^4}{s \eta_p}. \quad (15)$$

where  $g$  is the acceleration of gravity,  $\eta_p$  is the viscosity of the plume material, and the slope of the bottom of the plume material  $\partial A/\partial x$  in the upstream direction drives flow. The volume flux is the product of  $F$  and the circumference of a circle of radius  $s$  around the plume. This is dimensionally,

$$V = 2\pi s = \frac{\Delta \rho g A^4}{\eta_p}. \quad (16)$$

Solving, the thickness of plume material

$$A = \left[ \frac{V \eta_p}{\Delta \rho g} \right]^{1/4}, \quad (17)$$

depends weakly of the parameters and is independent of plate velocity. The distance  $s$  to the nose of the swell is obtained by noting that the upstream velocity is dimensionally the plate velocity  $U_p$ ,

$$U_p = \frac{F}{A} = \frac{\Delta \rho g A^3}{s \eta_p}. \quad (18)$$

Solving (17) and (18) yields the nose distance

$$s = \frac{1}{U_p} \left[ \frac{\Delta \rho g V^3}{\eta_p} \right]^{1/4}. \quad (19)$$

The kinematic and dynamic models differ in that the thickness of plume material in (17) is not constant in the latter. The distance to the nose of the swell is inversely proportional to plate velocity in both models. Note that the models do not explicitly assume that thermal buoyancy drives the flow. Part of the uplift of the swell may owe to the buoyant residuum from the partial melting that produced the volcanic chain (Phipps Morgan et al., 1995).

Finally, I treat Hawaii at 90 Ma as an on-ridge hotspot. Figure 8 of the reconstruction of Norton (this volume) reconstruction, the hotspot occupies about 300 km of ridge axis. The full spreading rate is unknown as this time is during the long Cretaceous interval of normal magnetic polarity. I estimate 180 mm yr<sup>-1</sup>. I let the thickness of affected lithosphere be 100 km. This yields a volume flux of 170 m<sup>3</sup> s<sup>-1</sup>, which is somewhat less than my estimate for the current flux.



### ***Volume of starting plume heads***

Starting plumes heads impinge on the base of the lithosphere. One would like to include their long-term average in estimates of global plume flux. In the standard plume hypothesis, starting heads produce radial dike swarms and flood basalts. The total volume of material in the plume head once ponded is

$$Q = \pi R_H^2 D_H, \quad (20)$$

where  $R_H$  is the radius of the region underplated by the plume head and  $D_H$  is the thickness of the underplated material. Estimation of either quantity is not straightforward. Plume material spreads laterally beneath regions of thin lithosphere. Dike swarms are not truly radial (McHone et al., 2005). This is expected as the ambient stress in the plate adds tensorially to the stress associated with the plume. This effect is likely to deflect the distal parts of the swarm where plume related stresses are weak. Hill et al. (1992) obtain an average heat flow of 1.2 TW from starting heads as an average for the last 100-200 m.y.

## **GLOBAL FLUX ESTIMATE**

Continuing with kinematic implications of the plume hypothesis, I discuss the global flux of hotspots. Davies (1988) and Sleep (1990) attempted to quantify the vigor of individual hotspots and the global amount of heat and mass transfer from plumes. Their compilations are still cited, not because of high accuracy, but because of the dearth of significant subsequent improvement.

Both Davies (1988) and Sleep (1990) summed the flux of their individual hotspots. The more vigorous measurable examples, like Hawaii, contributed a significant fraction of their fluxes. They did not attempt to estimate the flux from starting plume heads. I use a more recent global estimate of Anderson (2002) who includes that starting plume heads from Hill et al. (1992). Plume tails currently supply a heat flux of 2.3 TW and starting plume heads average 1.2 TW giving total of 3.5 TW. From my experience in estimating the fluxes of individual hotspots, I give a qualitative estimate of the uncertainty as  $\pm 30\%$ , provided the gross concept is correct. This uncertainty is small enough that it does not affect the gist of the issues that I raise below.

The global volume flux is convenient of topics involving the lithosphere. The volume heat capacity  $\rho C$  is  $\sim 4 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ . This yields a volume flux of  $2900 \text{ m}^3 \text{ s}^{-1}$  from (12). The mass flux is convenient for the whole mantle as density increases with depth. For a shallow mantle density of  $3400 \text{ kg m}^{-3}$ , the mass flux is  $10^7 \text{ kg s}^{-1}$ .

The flux is significant in terms of available reservoirs; it recycles the mass of the mantle,  $4 \times 10^{24} \text{ kg}$  in 12.7 billion years. Equivalently, the flux cycles 1/13 of the mass of the mantle in a billion years or 35% of it since the Earth formed at 4.5 Ga, extrapolating the present rate.

The estimated heat flux, 3.5 TW, is a significant fraction of the global mantle heat flux of 36 TW. The actual heat budget of plumes is greater (Anderson, 2002; Labrosse, 2002, 2003; Bunge, 2005; Mittelstaed and Tackley, 2006). The bottom hot thermal boundary layer of the mantle warms cool subducted material to the MORB adiabat before

it imparts the excess temperature of plumes. Both heat fluxes are comparable so the total heat flux to plume from the base of the mantle is around 7 TW.

This flux allows modeling the thermal history of the core (Anderson, 2002, Labrosse, 2002, 2003; Nimmo et al., 2004). The specific heat per unit mass of the core is about half that of the mantle  $0.625 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$  and the mass of the core is  $2 \times 10^{24} \text{ kg}$ . A heat flux of 7 TW cools the core at  $176 \text{ K B.Y.}^{-1}$  in the absence of contributions from radioactive decay and latent heat. This rate requires that the core cools faster than the mantle. Radioactive decay of potassium in the core is a possible heat source. A chemically dense layer at the base of the mantle may contain radioactive elements that augment the heat from the core (Davaille, 1999; Kellogg et al., 1999). There is no hard evidence demonstrating a hidden chemical reservoir, but U and Th antineutrino detectors will soon have the capability of detecting a heat source from a chemically dense “dregs” layer at the base of the mantle (Araki et al., 2005; Fiorentini et al., 2005; Enomoto et al., 2006). This is especially true if the layer is locally thicker (passive cusps entrained by the plume or buoyant hot “lava-lamp” upwellings) beneath hotspots as seafloor detectors could resolve lateral variations in antineutrino flux within the Pacific basin. Precise seismic tomography could conceivably independently resolve these features and the deepest conduit regions of the plumes. Potassium antineutrino detectors are not yet practical.

It is also illustrative to compare the rate at which plumes recirculate mantle with the rate at which plate processes segregate it. Modern oceanic crust is  $\sim 6 \text{ km}$  thick; the underlying depleted residuum is  $\sim 50 \text{ km}$  thick (Klein and Langmuir, 1987, Plank and Langmuir, 1992; Klein, 2003). White et al. (1992) give a precise estimate of  $3.3 \text{ km}^2 \text{ yr}^{-1}$  for the global rate of seafloor production. I use the rounded value  $\sim 3 \text{ km}^2 \text{ yr}^{-1}$  as a recent long-term estimate. The rates of crustal production and residuum production are therefore  $570$  and  $5300 \text{ m}^3 \text{ s}^{-1}$ . This total volume is only twice the estimated volume flux of plume material. At the current rate and depth of melting, a mass equivalent to 70% of the mantle has passed through the melting zone of the lithosphere at ridges since 4.5 Ga. The actual fraction is higher as the melting depth was higher in the past when the mantle was hotter. Rates of plate tectonics on the early Earth conceivably were faster (slower or nonexistent) than at present.

The lack of an obvious water source is relevant to “wet spot” hypotheses (e.g., Bonatti, 1990). Estimates for the amount of water  $1700 \times 10^{10} \text{ mol yr}^{-1}$  carried down with subducted crust (Staudigel, 2003, cf., Rüpke et al., 2004 who give 1.8 times this) are similar to the flux through arc volcanoes (Oppenheimer, 2003; Wallace, 2005). Even the sign of the net long-term flux is unknown. I agree that moderately water-rich regions (700 ppm  $\text{H}_2\text{O}$ , compared with  $\sim 150$  ppm in depleted mantle (Saal et al., 2002)) may exist and contribute to medium-volume hotspots like the Azores (Asimow, 2004).

## **ISSUES WITH THE PLUME HYPOTHESIS**

I cannot in a reasonably short paper attempt to refute or even summarize all the objections to and alternatives to the plume hypothesis. As I noted in the introduction, the direct lines of evidence do not have sufficient resolution, including mid-mantle tomography and lava petrology to obtain the potential temperature of the mantle source. I concentrated on reliable data sets including the thickness of crust beneath Iceland, the depth of extensive melting beneath Hawaii, and the existence of the Hawaiian swell. I

continue with other lines of evidence that bear on the plume hypothesis, but are currently inadequate to provide definitive tests.

### ***Plumes as dynamic features***

Some confusion arises from cartoon models of plumes where the hotspot is a point source of heat rather than a source of buoyant material. The actual dynamics of plumes are more complicated, but lead to observable features. (Here semantics clouds thinking. “Spot” in English can mean “point” or “region” like a spot on a dog or a sunspot. Note that the French term is “*le point chaud*” not “*la tache chaude*.”)

First, the strong dependence of viscosity on temperature explains both the excess volcanism and its spatial localization. The bottom thermal boundary layer in the mantle acts as a planar conduit. Plume material flows radially inward toward the tail conduit. The tail conduit is a low-viscosity channel. Plume material impinges on the base of the lithosphere it spreads laterally away from the plume conduit as a buoyant fluid. I considered this effect with regard to the Hawaiian swell and briefly with the Reykjanes ridge.

Plumes, like hurricanes in meteorology and sunspots, focus attention on part of the larger global flow pattern. Like rigid plate tectonics, earlier fixed hotspot theory was an approximation. It is inevitable that the source of plumes at depth and their mid-mantle conduits advect with the rest of the flow. This attribute is predictable from fluid dynamics (Steinberger, 2000; Steinberger et al., 2004). The main difficulties are obtaining accurate relative plate velocities and track velocities (Gripp and Gordon, 2002). Steinberger et al. (2004) jointly analyzed hotspot tracks, relative velocities, and plume advection and found the results consistent with the existence of plumes.

### ***Delamination alternative***

I discuss a well-posed non-plume hypothesis for Hawaii that has simple implications. A crack propagates through the plate producing volcanism at its tip and causing the lower lithospheric mantle to delaminate on its flanks over a region perpendicular to the track forming the swell. Delamination to produce the swell and the volcanism has testable implications that distinguish it from plumes. I summarize them. Some data sets are consistent with either delamination or a plume. Others need to be better resolved to be definitive.

Plate cracking is evident along the Hawaiian chain. Solomon and Sleep (1974) pointed out the strike of the chain aligns perpendicular to the expected axis of intraplate tension (Stuart et al., this volume). En echelon lineaments of volcanic edifices align along the trends of the numerous chains including Hawaii (e.g., Jackson and Shaw, 1975). These features are compatible with a propagating crack, but also with plumes. Plume-derived melts need cracks to ascend to produce surface hotspots. These cracks preferentially align with the local intraplate stress.

I use the 1400 m uplift at the axis of the Hawaiian swell as an example. The amount of thinning of the lithosphere by delamination is essentially that of rejuvenation models of hotspots (Crough, 1983). For an example, I choose integers that are perfect squares; changing 81-m.y. lithosphere into 25-m.y. lithosphere produces an uplift of 1400

m. (Depth in m varies as 350 times the square root of age in millions of years in my examples. The lithosphere thickness in km is ~12 times the square root of age in millions of years.) The base of the lithosphere of 95-million-year-old lithosphere would thin from ~117 to ~69 km if the uplift is 1400 m.

The delaminated lithosphere needs to sink. Ambient mantle upwells to replace the lithosphere that delaminates. Its source temperature cannot be greater than that of the underlying asthenosphere. The material immediately underlying the lithosphere is just ambient asthenosphere. With good tomographic resolution, shallow lateral variation of seismic velocity would be confined about the depth to the pre-delamination base of the lithosphere.

The temperature anomaly of the delaminated lithosphere is equivalent to that of subducted of 16-m.y. lithosphere  $\sqrt{16} = \sqrt{81} - \sqrt{25}$ . There is in fact a broad positive geoid anomaly around Hawaii. This feature might indicate dense delaminated material as slabs produce positive long-wavelength geoid anomalies. However, the low viscosity and buoyancy of the plume tail conduit produces a similar geoid anomaly (Richards et al., 1988).

Tomography at mid-mantle depths would resolve the issue. One needs only the sign of the anomaly. A plume produces a narrow negative seismic velocity anomaly while delaminated material produces a broad positive. Montelli et al. (2004) resolve a negative anomaly beneath Hawaii, as do Lei and Zhao (2006).

### ***Uplift and subsidence from hotspots***

Returning to Hawaii, the excess elevation of the swell is evident even before one accounts for the dependence of seafloor depth on age (Fig. 4). The correction is straightforward and modest in the nose region of the swell (Phipps Morgan et al., 1995, plate 1.). One estimates the uplift rate of a point on the crust on the nose of the swell by assuming that the nose propagates across the seafloor with the hotspot velocity. This predicted rate of uplift in the last few million years could be appraised by careful studies of benthic fossils obtained from cores. Such a study could eliminate the possibility that the swell is an older feature unrelated to the current hotspot.

The behavior of the seafloor after the passage of a hotspot provides information on the underlying process. hotspots]. The long-term subsidence of oceanic islands has been known since the time of Darwin. However, quantifying this process is not simple. We may have a good constraint on seafloor age and edifice age. We may have paleo-depth indications back in geological time. I show that it is difficult to resolve the effects of plumes from alternatives with seafloor subsidence data. The melting sequence beneath Hawaii discussed by Sobolev et al. (2005) indicates that the buoyancy is mainly thermal.

First, ponded plume material produces modest uplifts. For example, a ponded layer 100-km thick with a temperature contrast of 250 K produces 1060 m of uplift. The material spreads laterally near the hotspot and transfers its heat into the overlying lithosphere. I use the 1400-m uplift at the axis of the Hawaiian swell in examples. The buoyancy causing uplift at the active hotspot is partially from ponded plume material and partial from lithospheric thinning. The excess heat of the plume thins the lithosphere downstream and the heat anomaly thereafter moves with the lithosphere.

Studies of ancient edifices need to account for the general subsidence of seafloor with age. That is, one must resolve the additional cooling from the loss of heat added by plumes. Mathematically, the O'Connell relationship governs the net increase rate of seafloor depth with time:

$$\frac{DS}{Dt} = \left( \frac{\rho_m}{\rho_m - \rho_w} \right) \left( \frac{\alpha}{\rho C} \right) (q - q_b) \quad (21)$$

where the substantive derivative indicates subsidence of a place on the seafloor as would be sampled by drilling, the first term represents the effect of isostasy for subsidence occurring beneath water, and the third term is the difference between the heat flow out of the top of plate and heat flow into the bottom. A rejuvenation model treats cooling as a half space, so there is no heat flow into the bottom except during the rejuvenation event. The seafloor depth is then

$$S = S_{\text{ridge}} + \left( \frac{2\alpha\Delta T_L\rho_m}{\rho_m - \rho_w} \right) \left( \frac{\kappa t}{\pi} \right)^{1/2}, \quad (22)$$

where  $S_{\text{ridge}}$  is the depth at the ridge axis,  $\Delta T_L$  is the temperature contrast between the surface and the base of the lithosphere,  $t$  is apparent plate age, and  $\kappa$  is thermal diffusivity.

A plume model is more complicated. Uplift occurs when the plume material ponds beneath the lithosphere. Subsidence occurs later if the plume material spreads laterally and thins. Finally subsidence occurs from the heat flow escaping from the top of the plate. The plume material is buoyant relative to the underlying mantle. This suppresses small-scale “stagnant-lid” convection into the base of the plate. The long-term subsidence thus differs from ordinary lithosphere where stagnant-lid convection inputs heat to the bottom.

I use two end-member models to illustrate the difficulties in resolving the effect of plumes. At one end, the lithosphere is not thermally affected by the hotspot and subsides with the square root of its crustal age.. At the other end, it behaves as rejuvenated lithosphere. The predicted difference in subsidence for a mid-plate plume is small. I use these models, as a best case. If one cannot resolve the predicted difference for Hawaii, one is unlikely to resolve the difference for a weaker midplate hotspot.

For example, consider lithosphere now near the Hawaii-Emperor bend that was rejuvenated 50 million years ago from a crustal age of 81 m.y. to a rejuvenated age 25 m.y.. (I use perfect squares for ages and retain extra digits on depths to aid in following the calculations.) Over the last 50 m.y., the rejuvenated area subsided 1281 m and the reference region subsided 856 m with the square-root subsidence model. The difference, 425 m, would be readily apparent if we knew the age of a surface eroded to sealevel on a seamount exactly at the time of the passage of the hotspot. Resolving this extra subsidence with real data is problematic. In an ideal case, we would have the current depths at each site of sealevel erosion surfaces formed at the same time, so eustatic sea-level variation does not cause uncertainty. An erosion surface is likely for a seamount chain, but less likely for a reference site away from the chain.

More realistically, erosion beveled the edifice downstream of the hotspot. The time taken to erode accrues significant source error unless there are tight age constraints. For example, if the beveled surface formed 10 m.y. downstream on rejuvenated lithosphere, it would have already subsided 320 m. This crust would have subsided  $1281 - 320 = 961$  m in the subsequent 40 m.y. If one incorrectly interpreted that the surface formed at the hotspot at 50 Ma, one would conclude that the crust subsided about the amount predicted for ordinary lithosphere, 856 m.

On-ridge hotspots are also difficult to analyze precisely. I again use the half-space model for illustration. It may apply near the axis where lithosphere is underlain by plume material. It does not apply far from the axis where normal lithosphere lies beneath the plate. The analytical formula for subsidence (22) implies that the depth change from the axis to aged seafloor depends linearly on the half space temperature. Making the plume source mantle as hot as reasonable from the petrological grid,  $1600^{\circ}\text{C}$  versus  $1300^{\circ}\text{C}$  for normal mantle implies a subsidence of 430 m at 1 m.y. versus 350 m. For example, the two regions subside 3010 and 2450 m, respectively in 49 m.y. The difference of 560 m is readily resolved if we find a subsided sealevel surface formed right at the axis. The time to erode to sealevel, however, is even an even more serious problem than with mid-plate hotspots. For example, a surface beveled 4 m.y. from the ridge axis would have already subsided 860 and 700 m, respectively, more than the expected difference between plume and ordinary lithosphere.

Clift (2005) analyzed data obtained from paleontological studies of cores obtained by deep sea drilling. He correctly concluded that he did not see large anomalous elevation changes in deep marine data from ancient hotspot edifices. The resolution of his data is inadequate to resolve small differences predicted between rejuvenated and unaffected lithosphere. This compiled data partitions depths to 0-150 m, 150-500 m, 500-2000 m, 2000-4000 m, and deeper. Only near-sea-level erosion surfaces and the current depth provide reliable information.

The size of the deeper bins is larger than the expected amounts of subsidence. In addition, the frustrating compilation method produces spurious precision (exactly 2000 m) when the paleo-depth in a borehole changes, say from the 500-2000-m bin to the 2000-4000-m bin. It also presumes that the depths of water masses inhabited by index benthic organisms did not vary with time. A careful reexamination of available core is certainly warranted.

### ***Hotspot track continuity***

Physically the plume source is a thermal boundary layer beneath a much thicker region, the rest of the mantle. In this case, the flux into the tail conduit depends on the properties of the deep part of the mantle, including flow stirred by plates and slabs and thermal anomalies from sunken slabs. It does not depend otherwise on the condition of the shallow mantle. Starting plume heads and plume tails thus impinge on a variety of surface tectonic environments (Fig. 5).

One would thus expect plumes to track across a variety of environments. Equivalently, geologists call both Hawaii and Iceland “hotspots.” This nomenclature implies both long-lived features owe to a common mechanism. Sleep (1990a, 1990b,

1992, 2002a) argued that some hotspots have evolved from off-ridge to on-ridge while others have moved from the axis into plate interiors.

The trend through the Montereian Hills, the New England Seamounts, the Corner Seamounts, and Great Meteor Seamount is the best example (Fig. 5). If this interpretation is correct, it is strong evidence for a deep cause. Tristan-Gough (ridge leaving), Réunion (ridge crossing), Louisville (ridge approaching), Ninetyeast (ridge leaving), and Foundation (ridge approaching) are additional examples.

Hawaii should be added to the list as a ridge-leaving hotspot. Norton (this vol.) shows it as an on-ridge hotspot at ~90 Ma when the present northern most part of the track formed. The other “half” of the track (formed before the hotspot crossed the ridge axis) was thus on another plate and is not now adjacent to the northern Emperor chain.

McHone (1996) doubts that a plume produced the New England Seamount track. He gives well-posed points that apply in general to hotspot tracks. They are based on valid data. I summarize them, give related objections by other authors, and briefly reply.

(1) There is no age progression evident on the land part of the track in New England. Further Bear Seamount on the continental rise is older 120 Ma (Swift et al., 1986) about 20 m.y. older than the age expected from a simple track model. These observations are correct, but expected from a dynamic model of a plume that is a source of hot buoyant material. From (19), the time for a place on the crust to track from the nose of the swell to the plume is dimensionally

$$t_{\text{nose}} \equiv \frac{s}{U_p} = \frac{1}{U_p^2} \left[ \frac{\Delta \rho g V^3}{\eta_p} \right]^{1/4}, \quad (23)$$

which becomes large when the plate velocity  $U_p$  is small. This time is about 5 m.y. for Hawaii where the track velocity is  $8.3 \text{ mm yr}^{-1}$ . This scale time indicates how long hot plume material remains beneath a patch of lithosphere.

I obtain the nose time, from parameters estimated by Sleep (1990b). The buoyancy flux for the western seamounts is about  $\frac{1}{2}$  that of Hawaii,  $\sim 2 \text{ Mg s}^{-1}$ . The track velocity about  $\frac{1}{2}$  that of Hawaii,  $4.7 \text{ mm yr}^{-1}$ . The nose time for the western seamounts is a factor of  $\sim 2^{1.25} = 2.8$ , or 12 my.

Sublithospheric relief may have caused plume material to spread upstream (east) when the plume was beneath New England. The base of the lithosphere shallows to the east from the craton north of Montréal, across old Appalachian continental boundary, and to the younger passive margin. This slope acted as an upside-down drainage pattern to the east. Bear seamount formed ahead of the plume from buoyant material that cascaded up the lithospheric escarpment at the passive margin.

A general age progression does exist when the track is viewed a whole. Baksi (2005) pointed out that the radiometric ages for the track by Duncan (1984) are not up to modern standards. Stratigraphic (Swift et al., 1986) and paleontological ages (Uchupi et al., 1970), however, support the progression. Collection of more samples particularly by drilling and their dating with modern radiometric methods and paleontology is warranted.

(2) The track is not straight. True. Given its duration one would not expect the velocity of the Atlantic plate to stay constant.

(3) The land volcanism follows local features. This is true when the track is view on a local scale. This is also true of the marine track (Beutel, 2005). This is expected in that

dikes are perpendicular to the most tensile stress and tend to follow local weaknesses. There is a tendency for flexural stresses to cause regions of horizontal tension and horizontal compression at the passive margin and hence domains of enhanced and sparse volcanism.

(4) The Early Jurassic magmas in New England are similar to the lavas associated with the track. There is no obvious chemical signature of the plume. There are other volcanic provinces including those Newfoundland and the southeastern United States. True. I would associate the Early Jurassic rocks with the Bahama-Cape Verde starting plume head centered in southern Georgia and the break-up of the Atlantic. Newfoundland may be part of the Azores track and the southeastern United States part of the Bermuda track. I reiterate that I consider plumes to be mainly thermal features.

(5) The uplift associated with the track is small. True and I have already noted that plume material produces only modest uplifts.

(6) There is Eocene to Miocene coralline algae on Mytilus seamount (Swift et al., 1986). This indicates a near sealevel surface at that time. It also indicates an episode of renewed volcanism in Tertiary, well after the track passed. The algal fossils are mixed with Cretaceous fossils (Uchupi et al., 1970, who quote an internal WHOI report by J. M. Zeigler). An organic reef grew on this seamount and the dredged fossils are likely to be in place. They indicate the death age of the surface reef not an age of volcanic activity. Similar fossils from Bear seamount are less likely to be in place. In general, feeble volcanism at a later date does not preclude earlier presence of a plume.

### ***Why it is hard to see evidence of other Pacific plumes***

In the plume hypothesis, Hawaii is a vigorous plume in the middle of a fast-moving plate. It leaves a clear track. Iceland is a much less vigorous plume, by a factor of ~6. It impinges at a slow-spreading ridge axis. This allows the maximum amount of pressure-release melting. The thickened old lithosphere on each side of the ridge confines the buoyant plume material channeling it along the axis. These favorable circumstances do not exist for many hotspots and the ages of many seafloor edifices are unknown.

I place the Iceland plume with is 1/6 of Hawaii (volume and buoyancy) flux beneath the Pacific plate as a thought experiment. Its buoyancy flux produces a swell cross section of  $1400/6 = 230 \text{ km}^2$ . The swell nose from (19) is a factor of  $(1/6)^{3/4} = 0.26$  closer to the plume or ~100 km. The swell thus is confined to the area of the volcanic edifices and their flexural moat, which obscures its existence. The thickness of the plume material beneath the hotspot *A* is a factor of  $(1/6)^{1/4} = 0.63$  of that beneath Hawaii.

When I place the Iceland plume beneath the East Pacific rise where it is spreading ~8 times faster than at Iceland, it can supply only  $1000/8 = 125 \text{ km}$  of ridge length with a strong flux. More likely, the buoyant plume material spreads along the axis producing a broader region of mildly enhanced volcanism. Kinematically, a plume cannot stay near a fast ridge axis like one does near Iceland. The “absolute” velocities of ridge axes are much greater in the Pacific than the Atlantic.

## **CONCLUSIONS**



The plume hypothesis is a straightforward explanation of the salient features of hotspots. Hot buoyant material ascends from deep in the mantle. It ponds at the base of the lithosphere and spreads laterally. Interaction with the drag of the plate provides a ready explanation for the Hawaiian swell. It also explains numerous flow-line features between a near-ridge hotspot and the ridge axis (Morgan, 1978; Sleep, 2002a, 2002b; Kopp et al., 2003). Buoyant material at the base of the lithosphere flows in that direction. Numerical modeling of this flow is not easy because the strong dependence of buoyancy and viscosity on temperature.

The excess temperature of plume material explains voluminous magma production. Starting plume heads explain the sudden outbreak of magmatism of broad previously quiescent regions. Plume tails heated at a basal boundary layer explain hotspot tracks. The deep parts of this process are at the limit of tomographic resolution. I do not have a good constraint on the source depth of plumes. With regard to the mid-mantle, plumes have been putatively resolved where expected beneath many hotspots (Montelli et al., 2004) but this claim is controversial (e.g., van der Hilst and de Hoop, 2005). Numerical codes cannot yet handle the viscosity contrasts and resolution necessary for plumes to self-organize in three-dimensional flow with plates.

It is thus helpful to go over the shallow implications of plume theory. We know the depth of the seafloor and its age over most of the ocean basins. I contend that one should concentrate on the most reliable data sets first. These are the current bathymetry and gravity. The latter is useful for inferring bathymetry where data are sparse and for locating the swells of weak hotspots. Seismic data need improved resolution at lithospheric and asthenospheric depths; that is hot plume material in the asthenosphere versus thinned lithosphere. Limited S-wave splitting data are compatible with the flow pattern expected for the Hawaiian swell (Walker et al., 2005), but are not available globally. Paleobathymetric data provide only gross constraints. The scatter in heat flow data precludes their use to find temperature anomalies beneath old seafloor.

The gross bathymetry of hotspots provides estimates of the volume of material involved with plumes and with any alternative mechanism. The global volume flux  $2900 \text{ m}^3 \text{ s}^{-1}$  and the heat flux 3.5 TW are large in many respects. They are a major heat sink on the deep mantle. If we extrapolate over time, a significant fraction about 1/3 of the mantle has gone through plumes in the Earth's history. Mass requirements alone preclude non-renewable small volume sources for chemical melting anomalies. Conversely, most of the mantle has melted at least once at a ridge axis and been subducted. Sobolev et al. (2005) contend that this recycled crust along with elevated temperature control the chemistry of Hawaiian magmas.

Conversely, some effects of plumes are small and local as they constitute only 10% of the global heat flux. This is particularly true on the fast-moving Pacific plate. The Iceland plume would have only subtle features if it impinged there or at the East Pacific Rise. The vertical tectonics associated with plumes is also modest. It is difficult to resolve subsidence along hotspot tracks from the ordinary subsidence of seafloor. It is difficult to resolve the timing and amount of ancient uplift of the deep seafloor. The poor resolution of much of relevant data is cause for improving analytic techniques and collecting more data, particularly seismic, paleo-bathymetry and the ages of seafloor edifices. In particular, we still do not know the age of many volcanic edifices. Unreliable dates make it difficult to appraise the age progression of the New England track.

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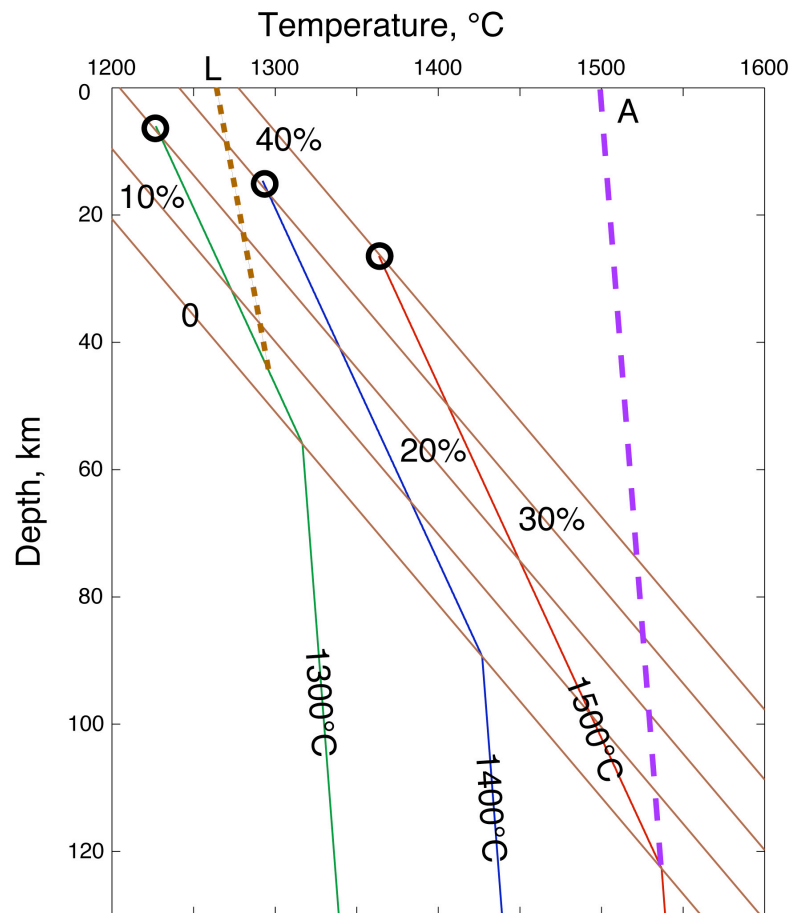


Figure 1. Simple petrological grid shows isopaths of the fraction of partial melting and the paths of abiabatic ascent in temperature-depth space. The paths at a ridge axis terminate at the base of the crust (circles). The potential temperature of the source before melting is the solid adiabat extrapolated to the surface (dashed line A). Magma ascending along the liquid adiabat (path L) becomes superheated.

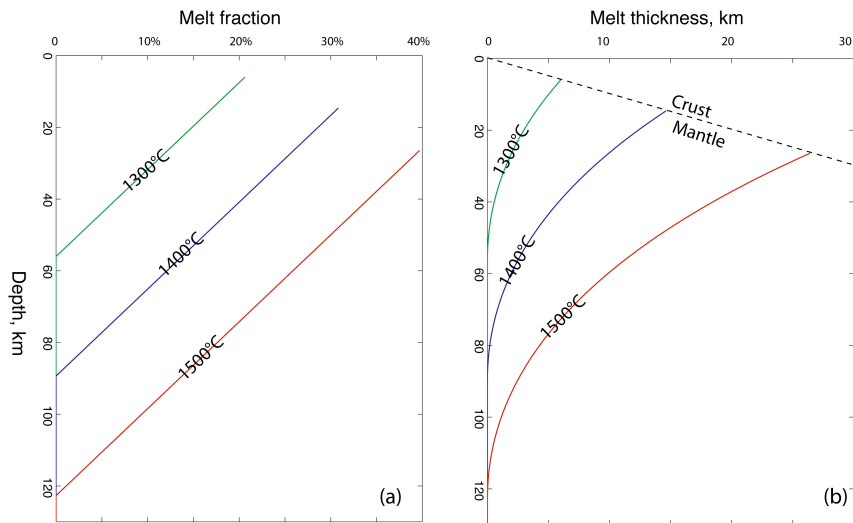


Figure 2. (a) The fraction of melt as a function of depth for potential temperatures of 1300, 1400, and 1500°C. The lines end at the base of the oceanic crust. (b) The total thickness of melt produced during adiabatic ascent. The curves end where the thickness equals the depth to the base of the oceanic crust.

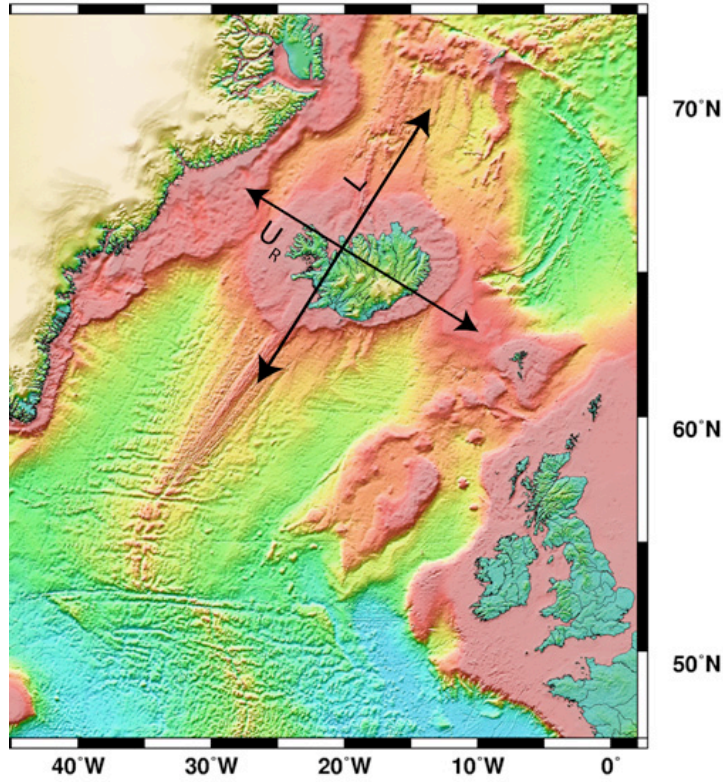


Figure 3. Topographic map of the Iceland area after Smith and Sandwell (1997). The hotspot strongly affects the region between the Faeroe Islands and Greenland for over 1000 km of strike distance. The starting plume head affected the Irish Sea and the West side of Greenland (not shown). Stretched continental margins occupy a much smaller surface area.

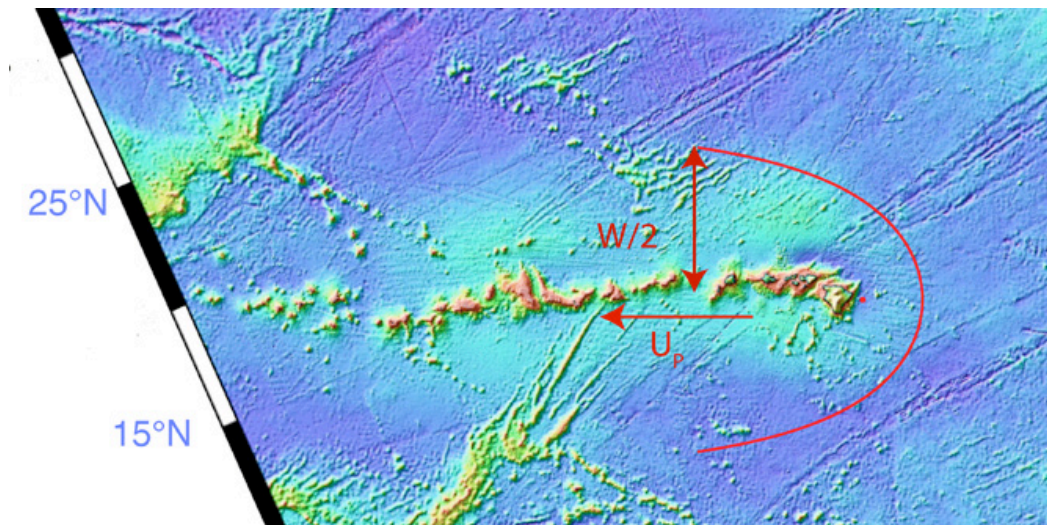




Figure 4. The light blue region of the Hawaiian swell is evident on the topographic map after Smith and Sandwell (1997). The red “parabola” from the kinematic model outlines the nose. The swell becomes more symmetrical north-south when one plots the difference between actual seafloor depth and the expected depth of the age of the seafloor (Phipps Morgan et al., 1995). Note that there is no increase in the rate of volcanism associated with the youngest edifice. The “big island” of Hawaii appears larger because the crust is loaded on only its west side (Watts and Ten Brink, 1989).

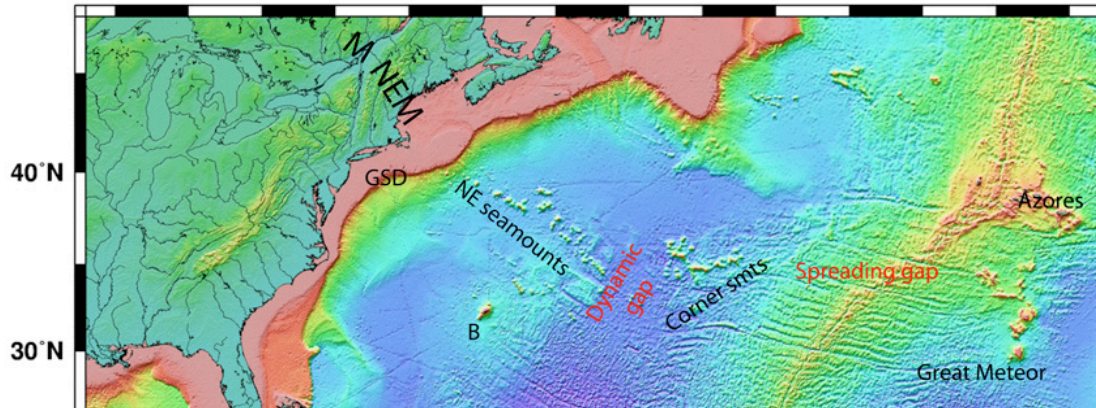


Figure 5. Topographic map after Smith and Sandwell (1997) shows Monteregian hotspot track. The older end of the track becomes evident in the Monteregian Hills (M) at the edge of the craton. It produced igneous events in New England (NEM). It crossed the passive margin. The Great Stone Dome (GSD) is an intrusion into passive margin sediments on the flanks of the track. Flow of plume material over the lithospheric escarpment beneath the passive margin lead to pressure-release melting. The New England (NE) seamounts formed in the middle of the plate. The track then jumped to the on-ridge Corner seamounts. It crossed the ridge axis and is now forming the mid-plate Great Meteor seamount. The gap between the New England and Corner seamounts is dynamic in origin (Sleep, 2002a). The plume became close enough to the ridge that plume material flowed along the base of the lithosphere toward the axis. The gap between Corner and Great Meteor seamounts is kinematic in origin from seafloor spreading that has occurred after the track crossed the axis. This interpretation illustrates that hotspot tracks are somewhat oblivious of shallow tectonic features and that on-ridge and off-ridge hotspots share a common cause.