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HOTSPOT VOLCANISM AND MANTLE PLUMES

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INTRODUCTION

Voluminous volcanism occurs both along mid-oceanic ridge axes and along island arcs. Plate tectonics provides a framework for discussing ridge volcanism in terms of upwelling mantle material which forms a diverging lithosphere of oceanic plates. Arc volcanism is more complicated but clearly related to the subduction of oceanic lithosphere beneath the arc. However, midplate volcanism that forms seamount chains, such as the Hawaiian Islands, and segments of the mid-oceanic ridge, such as Iceland, with excessive volcanism are not as obviously related to plate tectonics. Both types of features were called "hotspots" by Morgan (1971, 1972) and attributed to the upwelling of plumes of hot material from the deep mantle beneath the active volcanism.

The explanation for midplate seamount chains is relatively simple. The moving plate passes over an upwelling of hot material from great depths in the mantle (Figure 1). The lithosphere of the plate is heated and volcanism occurs above the upwelling. A sequence of volcanoes are each active when they are over the plume and then die as the plate moves away. Thus, the age of volcanism becomes progressively older as one moves down the chain away from the active volcano. This hypothesis was intended to explain basic features of midplate seamount chains: an active volcano at one end and the formation of atolls and submerged seamounts down the chain by erosion of the volcanic edifice, subsidence of the seafloor, and deposition of carbonate reefs. In addition, the parallelism of several Pacific

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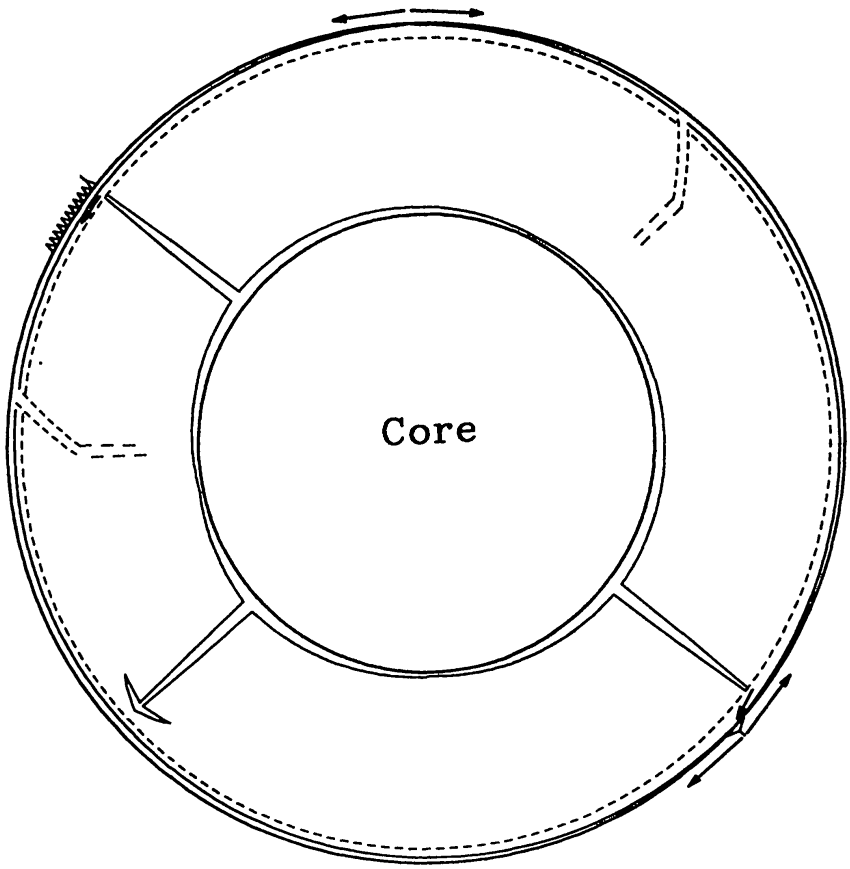

Figure 1 Various aspects of mantle plumes and convection within the Earth are shown schematically more or less to scale. From the bottom up: a thermal boundary layer is produced by heating from the core; plumes arise where the boundary layer is thick (4:30, 7:30, and 10:30); the boundary layer is thinner where slabs impinge (2:00 and 9:00); the cylindrical conduit of the plume tapers upward because viscosity decreases upward; new plumes start with a large head (7:30); hotspot chains are generated as a plate moves over a plume and hot material is entrained from the asthenosphere by the plate (10:30); normal ridges tap the adiabatic interior (12:00); plume material is entrained into hotspot ridges (4:30).

seamount chains in the plumes was presumed to arise from deep regions that are relatively fixed compared with the moving plates.

The history of ideas about the origin of seamount chains with particular emphasis on the Hawaiian Islands is given by Clague & Dalrymple (1987, 1989). A short discussion of this history is relevant to understanding the current controversies about hotspots and mantle plumes. The age progression of the Hawaii chain was known qualitatively in the last century and to some extent before that by native Hawaiians. The idea of a moving heat source also dates from the last century. Explanations of Hawaii and other Pacific seamount chains in terms of mantle plumes were presented even before plate tectonics was well understood and generally accepted

(Wilson 1963, 1965). However, mantle plumes did not receive the general acceptance of the rest of plate tectonics and alternative explanations, again adapting century-old ideas to plate theory, were proposed in terms of propagating cracks in the lithosphere.

Several reasons can be offered for the relatively slow progress toward agreement and understanding on the existence and nature of mantle plumes. First, the physics and kinematics of plumes are complicated. Plumes are supposed to arise from the deep mantle, probably near the core-mantle boundary. That is, the whole Earth, including the core and the deep mantle, is significantly involved. Second, hotspots occur globally and relevant observations involve geochemistry of the lavas, paleontological and radioactive dating methods, sedimentological studies to monitor uplift and subsidence, seismic studies of all depths in the Earth, and geoid and heat flow anomalies. That is, much of geology is involved on a global basis. Thirdly, on a more cultural basis, both Wilson (1963, 1965) and Morgan (1971, 1972, 1981) were vague on how and where plumes would originate at depth from mantle convection. For example, this allowed me (Sleep 1984) to contend that mechanisms involving lithospheric stress and cracks were inherently more testable than those involving plumes. That is, the present author entered the topic of mantle plumes as a skeptic with the working hypothesis that plumes did not exist.

I focus this review on lines of evidence that are useful for showing that mantle plumes really occur and that are thus also useful for demonstrating the physics of plumes and their effects at shallow depths. The concentration is on topics leading to the current level of understanding (Figure 1). That is, plumes are narrow cylindrical conduits of hot low-viscosity material that ascend from near the core-mantle boundary; material enters plumes by horizontal conduit flow along the lowest part of the thermal boundary at the base of the convecting mantle; and hotspots are the result of plume material ponding at the base of the lithosphere or being entrained into mid-oceanic ridge axes. I work from the bottom up starting with the convection in the deep mantle, going to the interaction of hotspots and ridges, and finally returning to midplate hotspot swells, a topic reviewed by Crough (1983). The cleanest evidence comes from isolated hotspots like Hawaii and Iceland. However, many hotspots occur in families including the Pacific superswell hotspots of Caroline, Samoa, Pitcairn, Marqueses, Macdonald, Tahiti, Easter, Juan Fernandez, and San Felix (Bonatti et al 1977, McNutt & Fischer 1987). Such lines of hotspots may form above two-dimensional tabular upwellings at depth (Houseman 1990, Sleep 1990a), but the process is too poorly understood to review usefully. I also concentrate on the physical aspects at the expense of the chemical ones. I do this because the physics is better understood at present (particularly by me) than the chemistry although chemical methods are likely to eventually provide the best constraints.

SOURCE OF MANTLE PLUMES

Several observable features of hotspots are the result of differences between plumes, which are convection heated from below, and the rest of the convection associated with plate tectonics which is heated mainly from within. The differences can be qualitatively understood by considering how heat gets out of and into the mantle. Heat can escape from the interior as surface heat flow only when hot material approaches the surface by convection because the mantle is much thicker than the distance over which heat can be carried by conduction even over the age of the Earth. The heat supplied by radioactive decay and by cooling of the mantle over time is distributed throughout the mantle. Thus the heat can be removed only by cycling all the convecting material near enough to the surface so that conduction can act—that is, through the oceanic lithosphere. Therefore, there are no fixed cores of convection cells—which is expected because ridges migrate with respect to trenches. In addition, the widespread heat sources do not create locally hot areas. In contrast, convection heated from below involves a hot basal boundary layer which is hotter (in terms of "potential" temperature adjusted for adiabatic effects and referenced to a common depth) than the rest of the fluid. Limited volumes of the basal boundary layer, which are heated at any one time and place ascend as isolated upwellings. The upwellings rise until they pond beneath the more rigid lithosphere or are entrained into mid-oceanic ridges.

Detection of local hot upwellings is thus fundamental to establishing the existence of plumes and to identifying the depth of the basal boundary layer. I begin with the latter topic and then discuss evidence that new plumes originate from instabilities on the basal boundary layer. The topic of whether the mantle source region of hotspot volcanism is really hotter than normal mantle is conveniently discussed with regard to the interaction of hotspots with mid-oceanic ridges.

Depth of the Source Region of Plumes

As plumes presumably arise from a basal boundary layer, they are a convenient probe to determine the depth to that layer and hence the depth of convection in the mantle. It is known that the convection in the Earth associated with plate tectonics extends down to at least the depth, 680 km, of the deepest earthquakes associated with downgoing slabs (Frohlich 1989). Thus the basal boundary layer could exist either near that transition between the upper and the lower mantle or just above the core. Both cases imply a compositional boundary in which the lower material has a more dense composition than the upper one so that material does not cross the boundary. Therefore, heat is transferred across the boundary by conduction.

The core-mantle boundary is clearly a compositional boundary between molten iron alloy and solid silicates. The 680 km transition zone is more problematic because most of the difference in density and seismic velocity is associated with a phase change. Laboratory and seismic data are not sufficiently accurate to determine whether an additional compositional density difference exists (cf Jeanloz & Morris 1986, Bina & Silver 1990, Bukowinski & Wolf 1990). Thus a controversy has persisted since the early days of plate tectonics as to whether convection is mantle wide or whether convection systems in the upper and lower mantle are separated by a compositional density difference (see Silver et al 1988).

Davies (1988a) used the properties of convection heated from below to show that plumes are more easily explained if they ascend from near the core-mantle boundary and if the convection associated with plate tectonics is also mantle wide. The relative sizes of the core and mantle are then such that the system behaves as convection heated from within with a modest perturbation from heat from below. Specifically radioactive elements (U, Th, and K) and hence heat generation are probably strongly concentrated in the mantle rather than the core. About 4/5 of the Earth's heat capacity is also in the mantle while 1/5 is in the core (Stacey 1980). As radioactivity and specific heat from cooling are comparable sources at present (Christensen 1985), about 10% of the total heat budget would come from the core if both the mantle and core cooled at equal rates. In contrast the upper mantle is only 18% of the mass of the Earth and 27% of the mass of the mantle. Thus separate upper mantle convection would be essentially convection heated from below.

Continuing with Davies' (1988a) line of reasoning, plate tectonics behaves mainly like convection heated from within. That is, mid-oceanic ridges are passive cracks that tap a relatively adiabatic reservoir from below. The depth-age relationship of oceanic crust is explained by conductive cooling of the oceanic lithosphere as it moves away from ridge axes. In contrast, the temperature contrast across the basal boundary layer associated with convection heated from below is comparable to the temperature contrast across the upper boundary layer—in this case the lithosphere. Ridges would be expected to migrate off upwellings from below. Thus large source temperature anomalies would be expected along the ridge axis depending on how effectively upwellings are locally tapped. Topographic and geoid anomalies comparable to those of ridges would be expected beneath upwellings not associated with ridges.

To be sure, hotspots do exhibit these characteristics of convection heated from below, but the effects are local and minor on a global basis. Using methods reviewed below, Davies (1988a) and Sleep (1990a) determined the heat flux of individual hotspots and summed the total globally. About 6% of the heat flux from the mantle is associated with hotspots. As emphasized by those authors, this is an appropriate amount to be supplied by the core but much too little to be supplied by separate convection beneath the upper mantle. In addition, hotspots tend not to overlie regions where the geometry of mantle-wide convection implies that slabs have collected near the base of the mantle over the last 100 m.y. The cool slabs are expected to make the basal boundary layer locally thin and thus locally suppress plumes (Chase 1979, Chase & Spowl 1983, Richards et al 1988).

Temperature Contrast across the Basal Boundary Layer

The structure at the base of the mantle like the structure of the upper mantle-lower mantle transition is not adequately resolved by seismic studies (see Young & Lay 1987). It is possible that an intrinsically dense layer of dregs exists beneath the basal thermal boundary layer associated with convection in the overlying mantle. This implies that part of the temperature difference between the core and the interior of the mantle occurs across this chemical layer while part occurs across the basal thermal boundary layer. Plumes are a direct probe of the thermal boundary layer and thus yield information whether a chemical layer exists (Jeanloz & Richter 1979, Jeanloz & Morris 1986).

I first consider flow to the plume in the absence of a chemical boundary layer. The highest temperatures occur at the base of the thermal boundary layer in contact with the core. As viscosity is strongly temperature dependent, the lower part of the thermal boundary layer acts as a conduit letting material flow to the plume (see Loper 1984, Stacey & Loper 1984, Loper & Eltayeb 1986, Sleep et al 1988, Davies 1990). The highest flow velocities are at the base of the layer because the core acts as a free-slip boundary. The free-slip nature of the boundary also allows this rapidly flowing material to leave the boundary layer to enter plumes. Thus the initial maximum excess temperature of the upwelling material is the temperature contrast across the thermal boundary layer. The excess temperatures of plumes are thus expected to be globally similar except for the loss of heat to the surrounding mantle on the way up. In contrast, a chemical boundary layer might cause the temperature difference to be much less. Secondarily it might act as a no-slip boundary and retard the flow of the hottest convecting material.

The physics of heat loss from a plume to its surroundings are tolerably understood (Griffiths & Campbell 1991). Heat loss from a vertical plume by conduction to its surrounding is relatively inefficient. However, plumes are tilted by cross currents in the mantle. This causes the hot central region of the plume to underlie cooler surrounding material. Thermal plumes are unstable to convection in the vertical plane perpendicular to the tilt direction. The hot center rises and cooler material from the conductive halo around the plume is entrained. The net effect is to mix the plume with its surroundings and reduce the temperature contrast. Modeling indicates that this effect is important for the weakest plumes, but minor for strong plumes, like Réunion and Hawaii.

As reviewed below, the average excess temperature of hotspot material for strong hotspots relative to normal mid-oceanic ridges is around 250°C implying that the maximum excess temperature is around 300°C. A temperature estimate at the base of the convecting layer is obtained by adjusting the upper mantle plume temperature $\approx$ 1600°C to 2600°C for adiabatic effects down to 2900 km. This temperature is much less than laboratory and thermodynamic estimates of the temperature at the top of the core which are often 1000°C or more higher (cf Jeanloz 1990, Boness & Brown 1990, Boehler et al 1990, Williams et al 1991). Thus an additional boundary—either dregs at the base of the mantle or between separate upper and lower mantle convection systems—is believed to take up some of this difference (Jeanloz & Richter 1979, Jeanloz & Morris 1986). Alternatively, alloying reduces the melting temperature in the core enough so that the temperature contrast above the core is only a few 100°C.

For a stable layer of dregs to persist, it needs to be sufficiently dense so that it is not entrained by the overlying fluid (Sleep 1988, Olson & Kincaid 1991). A more complicated constraint is that the layer needs to be internally stratified or sufficiently viscous so that it does not convect rapidly. Theoretical estimates based on the theory of well developed convection indicate that the temperature dependence of viscosity will make the dense layer sufficiently fluid to convect efficiently enough so that most of the temperature contrast ends up in the overlying thermal boundary layer (Sleep 1988). However in an experiment by Olson & Kincaid (1991), the dense layer convected slowly because it was too thin for well developed convection and only 22% of the total temperature contrast occurred across the overlying boundary layer. It should be noted that their dense layer was intrinsically more viscous by a factor of 3.5 than the upper layer while the additional iron to increase density in the mantle should decrease viscosity. The questions involving convection in the dense layer and entrainment could be resolved by modifying Olson & Kincaid's (1991) experiments so that the intrinsic viscosity difference between the upper and the lower fluid varied independently of density.

Initiation of Plumes: Heads and Tails

It is intuitively expected that new plumes should arise from time to time from instabilities of the basal boundary layer. This topic has received considerable laboratory and theoretical interest and more recently geological interest because of observable effects: massive flood basalts and radial dike swarms. For purposes of this review, the topic is relevant because it demonstrates that plumes consist of lower viscosity material upwelling through the much higher viscosity mantle and because flood basalts and dike swarms are the best evidence of the existence of plumes in the past.

The basic physics describing the upwelling of material which is two or three orders of magnitude less viscous than its surroundings is well understood and similar for either hot plumes or compositionally less dense plumes (Whitehead & Luther 1975, Olson & Singer 1985, Griffiths & Campbell 1990, Loper 1991). The first batch of fluid to rise through the mantle ascends as a mushroom-shaped head which is fed from below by a narrow (less than 100 km diameter) cylindrical conduit (Figure 1). The head ascends by displacing the surrounding mantle at the Stokes' law rate for a sphere of the less dense material moving through the higher viscosity of the mantle. Thus the head spends most of its ascent time and acquires most of its plume material in the deep high-viscosity part of the lower mantle (D. E. Loper, personal communication 1991). Material within the conduit moves as pipe flow at a rate controlled by the lower viscosity of the plume. The head of the plume ponds when it approaches the upper surface and the plume is thereafter a throughgoing (tail) conduit from the basal boundary layer to the uppermost mantle. Normal mantle is entrained into the plume head. Thermal plumes differ from chemical ones mainly because heat diffuses in the head of the plume warming the entrained material and cooling the plume material (Campbell & Griffiths 1990). This causes the head to widen as it ascends.

The geological implications of this process are evident in many plateau basalt provinces (Courtillot et al 1986, Richards et al 1989, Campbell & Griffiths 1990, Duncan & Richards 1991, Richards et al 1991, Hill 1991). The approach of the plume head toward the surface implies the sudden arrival of hot material over a broad area. Thus initial uplift is followed by short-lived voluminous eruption of flood basalts which form a plateau, like the Deccan traps, on land, or a submarine plateau, like Ontong-Java, in the ocean basins. This volcanism occurs over a roughly circular 2000-km diameter region. The 800-km distance of excess volcanism on the Labrador Sea spreading center in the Davis Strait owing to the hotspot location of Vink (1984) is particularly useful for giving a minimum diameter of the Iceland plume head because entrainment of hot mantle rather than possible propagation of dikes away from the source region is involved. This 1600-km diameter is appropriate if the plume head ascends from the base of the mantle but inappropriate for plumes originating at 680 km (Campbell & Griffiths 1990, Hill 1991). Subsequent volcanism is less voluminous and is restricted to the long-lived hotspot track of the plume tail.

An alternative hypothesis by White & McKenzie (1989) is that flood basalts are the result of continental breakup along a hotspot track. Plume material ponds beneath the continental lithosphere. Voluminous volcanism occurs because much plume material ascends to shallow depths along the new ridge axis and beneath highly stretched parts of the continent. As discussed below, entrainment of hot material into ridge axes does produce voluminous volcanism and submarine plateaus, like Iceland. The key difference between the two hypothesis is that lithospheric rifting is not a required side effect of a plume head but rather a fundamental part of the breakup hypothesis.

The case for a plume head is probably best demonstrated for the Deccan traps as reviewed by Duncan & Richards (1991). First, volcanism preceded rifting of the west coast of India (Hooper 1990). Second, the Indian continent was moving rapidly northward at the time making it difficult for the plume to supply a large accumulation of material to the base of the plate. Duncan & Richards (1991) also note that extension is mild to nearly absent in several possible plume head events, particularly the Siberian traps which erupted in a plate interior. Hill (1991) notes that plume head events typically preceed rifting by a few million years when the plate rearrangement is basically a ridge jump and tens of million years for a major rearrangement.

A key feature of plume heads is that they produce a brief widespread episode of volcanism and dike swarms which can be recognized in the geological record. This provides evidence that plumes existed before well developed recent (younger than 200 Ma) tracks on oceanic crust. In particular, the presence of a plume head indicates that the plume was significantly hotter and less viscous than the surrounding mantle. The oldest currently demonstrated plume head event is the MacKenzie dike swarm, which radiated over 1000 km from the associated Muskox intrusion and Coppermine basalts in northern Canada (LeCheminant & Heaman 1989). These events were synchronous at 1267 Ma to within the resolution of the data—a few million years. In this case, continental rifting occurred north of Coppermine, and the hotspot track probably ended up in the new ocean basin where it was not preserved.

Also, plume head events and large submarine plateaus are empirically negatively correlated with the frequency of magnetic reversals. The suggestion that these events involve more rapid heat loss from the core (Larson

1991), if correct, implies that a thick nonconductive chemical layer is absent at the base of the mantle. That is, if such a layer existed, the core would sense changes in the temperature of the overlying mantle only by thermal diffusion through the layer. A lag of approximately the layer thickness squared divided by the thermal diffusivity thus would occur between the start of plume ascent and the change in reversal rate.

Boundary Layer Viscosity Contrast

The presence of plumes, particularly the presence of plume heads, indicates that the hot material in the basal boundary layer is two or three orders of magnitude lower than the overlying normal mantle. This situation does not occur in laboratory experiments, rather an equilibrium temperature and viscosity gradient is established with about one order of magnitude contrast (Nataf 1991). Two reasons why a large viscosity contrast and hence plumes may exist in the Earth are qualitatively evident but not well understood (Nataf 1991).

First, the viscosity increases a few orders of magnitude from the low viscosity asthenosphere to the normal mantle just above the basal boundary layer. (Depth dependent viscosity is impractical in the laboratory because large pressures are needed to change viscosity.) This tends to retard flow in the lower mantle and provide a high-viscosity lid over the basal boundary layer. Otherwise, plate motions including sinking slabs would stir the basal mantle to plate velocities. This would preclude relatively stationary hotspots (Duncan & Richards 1991, Richards 1991). In addition, slab material tends to pond in the lower mantle and increase the temperature and viscosity contrast of the lid.

The relationship between stirring and viscosity is easily quantified dimensionally. Stirring from flow in low-viscosity upper layers scales to the inverse of viscosity because the shear stress,

$$\tau = \eta \frac{v}{X} \tag{1}$$

(where η is viscosity, v is velocity, and X is the length scale over which velocity varies), remains approximately constant with depth. Slabs are more effective at stirring because they pond (or thicken) as they enter higher viscosity layers. The velocity times the thickness of the slab (vS) remains constant ($v_0 S_0$) with depth. The shear stress scales to the thickness and the density contrast of the slab as

$$\tau = S\Delta\rho g \approx \frac{v_0 S_0}{v} \Delta\rho g. \tag{2}$$

Combining (1) and (2) gives a velocity that varies inversely with the square root of viscosity. Thus, if hotspot relative velocities are several millimeters per year, a viscosity increase by a factor of about 50 is required between the upper mantle beneath the asthenosphere and the deep lower mantle.

Second, nonsteady convection may increase the viscosity contrast near the base of the mantle (Sleep et al 1988, Nataf 1991). Both the core and mantle of the Earth are cooling. The higher viscosity of the lowermost mantle may cause convection associated with the basal boundary layer to be relatively inefficient, such that plate tectonics would cool the mantle faster than plumes could cool the core. Thus, a nonequilibrium temperature contrast may build up above the core-mantle boundary during the evolution of the Earth.

INTERACTION OF PLUMES AND RIDGES

A key element in the early reasoning in favor of mantle plumes is that hotspot tracks cross ridge axes (Morgan 1971, 1972, 1981; Duncan 1984). This implies that an off-axis hotspot evolves into an on-axis hotspot, and then back into an off-axis one. Therefore, off-axis hotspots, like Hawaii, and on-axis ones, like Iceland, are different aspects of the same phenomenon: plumes. Conversely, a crack in a plate—an alternative mechanism for off-axis hotspots—would not be expected to cross the free edge of a ridge axis. In addition, on-ridge hotspots are relevant here for demonstrating that hotspots really have exceptionally hot material beneath them and for understanding the dynamics of ridge axes.

Excess Source Temperature of Plumes

The ridge axis at on-axis hotspots is subaerial or rather shallow compared with the ~2.5 km depth for normal ridges. Paired submarine plateaus are formed when the thickened crust spreads off axis in both directions. Examples include the Iceland-Faeroe and Iceland-Greenland plateaus, the Rio Grande and Walvis rises from the Tristan hotspot, and the Cocos and Nazca rises from the Galapagos hotspot. That is, part of the uplift is permanent and associated with crust that is thicker than the normal 5–7 km thick oceanic crust. For example, the crustal thickness beneath Iceland is poorly resolved in part because it is not clear whether a 7.2 km s^{-1} layer beginning at 10–15 km depth is crust or anomalous mantle. The crustal thickness may reach 30–35 km beneath the Iceland Faeroe plateau (see Flóvenz & Gunnarsson 1991).

The origin of the excess melting that forms the thickened crust is explained in terms of adiabatic upwelling of hot plume material to the ridge axis (Schilling 1983; Bickle 1986; Klein & Langmuir 1987, 1989;

McKenzie & Bickle 1988). For simplicity, I illustrate the thermodynamics using a eutectic system with a single melting curve before discussing the real Earth. The adiabat of upwelling material intersects the melting curve at some depth because the melting point increases much more rapidly with depth (3°C km^{-1}) than the solid adiabatic gradient (0.3°C km^{-1}). The temperature and pressure of partial molten material lie on the melting curve. The fraction of melt is obtained from energy conservation by assuming for purposes of calculation that the material remains a superheated solid until it partially melts at some depth. The specific heat times the temperature difference between the extrapolated solid adiabat and the melting curve, $C_p\Delta T$, equals the latent heat for the fraction of melt, FL. The fraction of melt is thus

$$F = \frac{C_p \Delta T}{L} \tag{3}$$

and the fraction of melt increases with decreasing depth by

$$\frac{\partial F}{\partial z} = \frac{C_p}{L} \frac{\partial \Delta T}{\partial z}. \tag{4}$$

Beneath ridges, the fraction of melt increases by 1% km^{-1} assuming $L = 0.3$ MJ kg^{-1} and $C_p = 1.15$ kJ kg^{-1} $°C^{-1}$.

The total thickness of melt generated is approximated by integrating the fraction of melt generated from the depth z_m, where melting begins, to the surface:

$$C = \int_0^{z_m} F\,dz. \tag{5}$$

(The thickness of oceanic crust is more precisely estimated by integrating from the base of the crust, including the kinematic effects of compaction which segregates melt, and assuming that a small fraction of the melt does not segregate.) Melting beginning at 35 km depth generates 6 km of crust in the simplified model presented above. It is also evident that the crustal thickness increases with increasing source temperature of the upwelling material because z_m increases by $(\partial \Delta T/\partial z)^{-1}$. The actual mantle is not eutectic but rather melts over a temperature range. The thermal implications are easily (but messily) included in the model by making ΔT decrease with increasing fraction of melting F. Thus a greater depth of initial melting is needed to generate 6 km of oceanic crust than in the eutectic model.

More importantly the chemical effects of noneutectic melting are observable and give information on the temperature and depth of the source region for basaltic magmas (Klein & Langmuir 1987, 1989; McKenzie &

Bickle 1988). The mantle can be considered to be a mixture of an aluminous phase, clinopyroxene Ca(Mg, Fe) $(SiO_3)_2$, orthopyroxene (Mg, Fe)SiO_3, and olivine $(Mg, Fe)_2SiO_4$. At the pressures relevant to ridge axes, the phases melt preferentially in that order and the FeO components melt preferentially with respect to the MgO components. Melts thus become more magnesian with increasing fraction of melting. High MgO glasses are thus clear evidence of high source temperatures, as in Hawaii (Clague & Weber 1991). Unfortunately, high MgO lavas are quite rare and hard to observe directly because olivine also crystalizes first in shallow magma chambers lowering MgO.

The Na_2O component is more useful because it preferentially enters the melt phase relative to the solid phases (Klein & Langmuir 1987, 1989). Thus, the Na_2O abundance decreases with increasing fraction of melting as more Na_2O-depleted solid is added to the melt. This component does not enter olivine and thus is not strongly affected by early fractional crystallization. In practice, Na_2O abundances are adjusted for fractional crystallization to the abundance where MgO is 8%. The more complicated behavior of CaO, FeO, SiO_2, and Al_2O_3 allows the depth of melting to be inferred independently of the fraction of melting determined from Na_2O. Melts found at lower lithospheric depths with gamet remaining in the source region are enriched in light rare earth elements.

Klein & Langmuir (1987, 1989) found that chemical differences among mid-oceanic ridge magmas are explained by adiabatic melting of ascending material. That is, hottest source regions start melting at a greater depth and melt more to produce hotspot ridges. Solid adiabat temperatures vary 300°C on a global basis with plumes about 250°C hotter than typical ridges. In basic agreement with this estimate, Watson & McKenzie (1991) used a convection model of melting beneath Hawaii and chemical composition to find a maximum excess temperature relative to typical ridges of 278°C. The expected correlations exist between Na_2O and axial depth and crustal thickness. The correlation between original axial depth and Na_2O also holds for older oceanic crust (Keen et al 1990). Anomalies to the general trend do exist. Obviously material ascending near cool lithosphere—such as to the tip of a ridge segment at a transform fault—does not remain adiabatic. Chemical and isotopic differences between plume material and normal mantle probably also exist. These differences are useful for the entrainment of tracing plume material independent of temperature anomalies (Rideout & Schilling 1985; Schilling 1985, 1991; Schilling et al 1985). Possible origins of heterogeneities are easily envisioned in terms of plate tectonics—for example, subduction of sediments and altered oceanic crust—but quantification is still lacking.

It is likely that the temperature of normal mantle has decreased with time. Thus the process described above for hotspot ridges may be applicable to normal ridges in the Archean (Sleep & Windley 1982, Bickle 1986). The role of hotspots in generating modern Mg-rich lavas (komatiites) and the possible applicability to Archean komatiites has been pointed out (Campbell et al 1989, Storey et al 1991). However, it is possible that the mantle has cooled faster than the core making the temperature contrast above the core-mantle boundary in the Archean too small to produce plumes (Sleep et al 1988). It is difficult to tell whether an isolated exposure of Archean volcanic rocks was produced by a localized plume or from globally hot mantle. Identification of plume heads by widespread synchronous volcanism appears to be the best way to recognize plumes in the Archean.

Entrainment of Plume Material into Ridge Axes and Hotspot Jumps

The subaerial segments of ridge axis crossing Iceland are several hundred kilometers long. In addition several hundred more kilometers of ridge axis north and south of Iceland are moderately affected by plume material. Thus a long segment of the ridge is supplied by a much narrower plume. Two processes are involved. First the plume material is entrained to a place on the ridge axis. Then the plume material flows along the strike of the ridge axis mixing with normal mantle away from the hotspot.

The physics of entrainment is qualitatively explained by noting that the upwelling at normal ridge axes is passive in the sense that the adiabatic interior supplies the necessary material to form the new oceanic lithosphere (Rideout & Schilling 1985; Schilling 1985, 1991; Schilling et al 1985). The plume supplies material to a segment of the ridge making the normal passive upwelling unnecessary. The normal flow in the asthenosphere away from the ridge axis is thus suppressed and plume material is deflected to the ridge axis (Figure 1).

A kinematic consequence of entrainment is that the observed (track) position of an on-ridge hotspot stays on the ridge axis while a fixed plume position would be expected to cross the axis. In the case of Iceland, absolute velocities imply that movement to the east of a few hundred kilometers from under Greenland to the eastern part of Iceland (Vink 1984). The ridge jump at 36 Ma was west toward the expected hotspot position while the current ridge jump is east toward the expected hotspot.

In addition to ridge jumps which can be qualitatively explained by the plume material creating a zone of weakness in the plate, entrainment implies that a hotspot track is captured by the ridge axis for a period of time. This tends to create a jump in the track of an off-ridge hotspot approaching the ridge (Figure 2). When the hotspot is far away from the ridge it moves with the expected track velocity and does not significantly

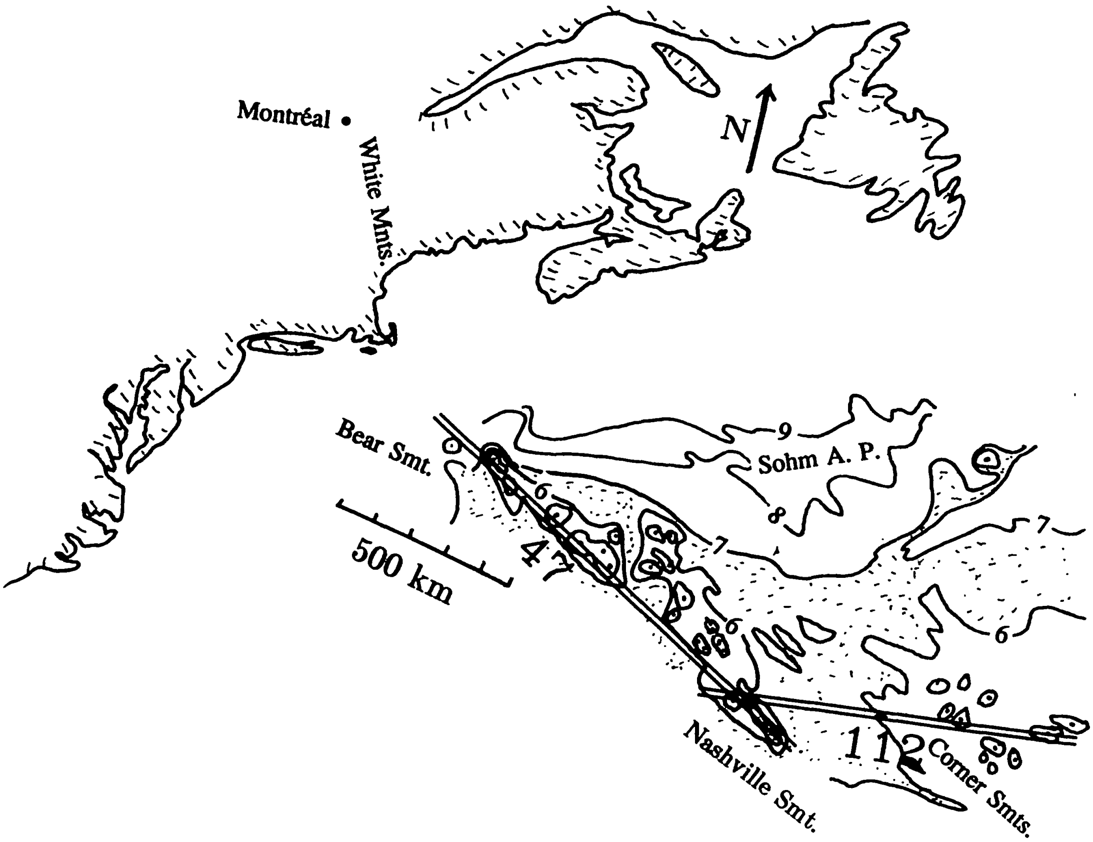

Figure 2 The track of the New England seamount plume (*double lines*) is shown on a depth to basement map modified from Tucholke (1986, Plate 5) by Sleep (1990b). The 5 km contour indicates the base of seamounts. The top of seamounts is indicated by dots. Between Bear and Nashville seamounts the track moved at 47 mm yr^{-1}; between Nashville seamount and the Corner Seamounts the track jumped toward the ridge axis at 112 mm yr^{-1} (Tucholke & Smoot 1990).

supply material to the ridge axis. At some time, the hotspot comes close enough to the ridge axis for entrainment to occur. The observed position of the hotspot then jumps to the ridge axis and remains on the ridge axis until the plume is too far on the other side. A jump off the ridge axis then occurs. The Icelandic hotspot has remained trapped on the ridge axis and has not jumped throughout its history.

The New England-Great Meteor hotspot jumped as it approached the ridge axis from the west (Figure 2). Between 85 and 76 Ma, the hotspot moved at a rate of 112 mm yr^{-1} (compared with the previous rate of 47 mm yr^{-1}) from the off-ridge Nashville seamount to the on-ridge Comer seamounts (Tucholke & Smoot 1990). Only small seamounts occur in the gap (Epp & Smoot 1989). Tristan has jumped in a more complicated way between segments of the ridge axis offset by transform faults [see Figure

6 of O'Conner & Duncan (1990)] and finally off the ridge to the east producing a gap west of the island of Tristan. The New England gap was mistaken for real weakening of the hotspots by Sleep (1990b). The Tristan gap is more complicated because it is not clear whether Gough is a separate hotspot and because the kinematics of a jump off the axis may be more complicated than a jump on. Schilling (1991) gives a low flux for Gough on the basis of a kinematic model of entrainment.

Entrainment, Small-Scale Convection, or Cracks

Plumes as envisioned here are sources of hot material that can flow rather than point sources of heat beneath the lithosphere. This model both provides an explanation for some weak hotspots while making it more difficult to recognize processes unrelated to plumes. Specifically, the discussion of hotspot swells below indicates that material hotter than normal adiabatic mantle persists beneath the lithosphere for a considerable time after it passed over the hotspot. Later this moderately hot material is tapped by cracks through the plate or entrained into the ridge axis. The island of Cocos northeast of the Galapagos hotspot is an example of the former process (Castillo et al 1988), while Rodriguez results from Réunion material entrained to the ridge axis (Morgan 1978).

These mechanisms are contrasted with tapping of normal adiabatic mantle by rifts and cracks in the lithosphere. As noted in the introduction, this mechanism has been applied in the past to even strong hotspots like Hawaii. Petrological indications of source temperature (Klein & Langmuir 1987, 1989; McKenzie & Bickle 1988; Clague & Weber 1991; Watson & McKenzie 1991) are the most definitive way to recognize hot material from plumes. Conversely, a source of stress is often evident, such as the Alpine collision near recent volcanism in the Massif Central in France and the Eifel hotspot near the Rheingraben. Many features including continental breakups may owe their existence to both mechanisms as hotter plume material creates regions of weakness which start cracks for passive upwelling.

A third class of mechanism involves small-scale convection in the asthenosphere driven by cooling at the base of the plate. Stress in the lithosphere caused by drag from the diverging flow at upwellings of this convection as well as the higher temperature of the upwellings themselves was proposed in the past for Hawaii (Solomon & Sleep 1974) and more recently for weak hotspots (Vogt 1991). It first should be noted that this mechanism is hard to distinguish petrologically from ordinary cracking because the upwelling temperature is that of normal adiabatic mantle. Secondly, analysis of topographic and geoidal anomalies in the ocean basins indicates that this mode of convection transports little heat (Davies

1988b). The only region where there is positive evidence of its occurrence is west of the East Pacific Rise where the asthenosphere probably consists of hot entrained plume material rather than normal adiabatic mantle.

More understanding of these issues, particularly with regard to weak hotspots on the continents, are relevant to petroleum exploration. Hotspots, small-scale convection, and cracks all may heat the sedimentary column and the lithosphere causing uplift unconformities and subsequent thermal basin subsidence (cf Crough 1981, Vogt 1991).

HOTSPOT SWELLS

Widespread vertical tectonics are associated with the passage of hotspot track across plates (Crough 1983). The resulting uplift as the hotspot approaches and subsidence after it passes are more easily observed in the ocean where the depth-age relationship of normal oceanic crust and sea level surfaces on atolls are useful references. In particular, a several hundred kilometer wide swell of elevated seafloor flanks the hotspot track; atolls along the axis of the track mark the general subsidence of the swell. Hypotheses regarding the origin of swells were reviewed by Crough (1983) who concluded that thermal expansion was the primary cause of the uplift. I start with this basic mechanism and then discuss extensions and difficulties of this hypothesis that have arisen since 1983.

Basically material from the plume is supposed to heat the lithosphere near the hotspot. The uplift from isostatically compensated thermal expansion is conveniently expressed in terms of an apparent thermal rejuvenated age of the oceanic lithosphere using the depth-age relationship of normal oceanic crust:

$$D = D_0 + 350\,\mathrm{m}\sqrt{\mathrm{age(c), m.y.}}, \tag{6}$$

where D_0 is the depth at the ridge axis. The rejuvenated depth of the swell defines the rejuvenation age,

$$D_r \equiv D_0 + 350\,\mathrm{m}\sqrt{\mathrm{age(r), m.y.}}, \tag{7}$$

and the swell uplift is $E = D - D_r$. To first order, swells and atolls on them subside at the rate predicted for oceanic crust of age(r) at the time of hotspot passage.

Heat Flux from Swell Topography

The hypothesis that the uplift of swells is caused by thermal expansion allows the heat flux to hotspots to be estimated (Davies 1988a, Richards et al 1988, Sleep 1987, 1990a). Assuming isostasy implies that the mass deficiency from hot material at depth is equal to the mass excess of the topography of the swell. The rate at which mass deficiency at depth is produced by the swell is then equal to the track velocity v_p times the mass excess of topography per length of track:

$$B = (\rho_m - \rho_w)\overline{WE}\,v_p. \tag{8}$$

Here ρ_m is the density of the mantle, ρ_w is the density of water, and $\overline{WE}$ is the cross-sectional area of the swell. In practice, the uplift of the swell is obscured by volcanic edifices along the swell axis and flexural downwarps flanking them (Figure 3). (Note: the terminology for this equation is not standardized, the quantity B is called buoyancy flux in papers where Sleep is an author, Davies does not multiply by density and expresses plume magnitude in $m^3\,s^{-1}$; Campbell & Griffiths (1990) multiply by g to express buoyancy flux in $N\,s^{-1}$.) The buoyancy flux is converted to heat flux by multiplying by the specific heat per mass and dividing by the thermal expansion coefficient:

$$H = BC_p/\alpha. \tag{9}$$

Davies (1988a) and Sleep (1990a) have used Equation (8) to determine buoyancy fluxes on a global basis and obtained the conclusion cited above that hotspots are about 6% of the mantle heat flux. The most vigorous hotspot is Hawaii with a buoyancy of around 8 Mg s^{-1}. Réunion has a flux of about 2 Mg s^{-1}. Several weak hotspots including Juan de Fuca have fluxes around 0.5 Mg s^{-1}. Equation (8) can also be applied to swells along old hotspot tracks if the initial uplift can be inferred from the subsidence of atolls through Equation (5) or from the current uplift of the swell relative to its flanks (Sleep 1990b).

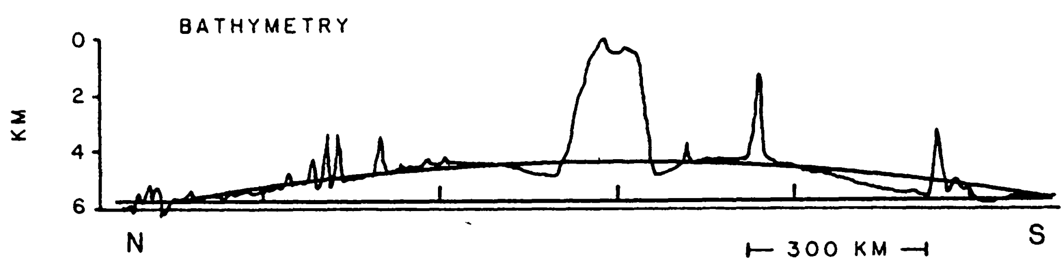

Figure 3 The bathymetry of the Hawaiian swell near Oahu is shown with a parabola to indicate the excess elevation above a constant depth line. The parabola has a halfwidth of 750 km, a height of 1.4 km, and a cross sectional area of 1400 square kilometers. Modified after Watts (1976) and Crough (1983).

Kinematics of Swell Formation

The nose of the Hawaiian swell extending upstream of the active hotspot is crudely parabolic (Figure 4). This shape is explained by the kinematics of flow from a relatively narrow plume through the asthenosphere to the flanks of the swell. A very simple geometric model of this process was presented by Sleep (1987, 1990a) and Richards et al (1988). Flow is represented as radial motion away from the plume,

$$v_{\text{plume}} = \frac{v_f r_f}{r}, \tag{10}$$

where r is the distance from the plume and the velocity is v_f at the reference distance r_f. The volume flux from the hotspot is

$$Q = 2\pi v_f r_f A, \tag{11}$$

where A is the asthenospheric channel thickness. The buoyancy flux is the volume flux times the density difference. The radial flow interacts with flow in the asthenosphere driven by the drag of the overlying plate at

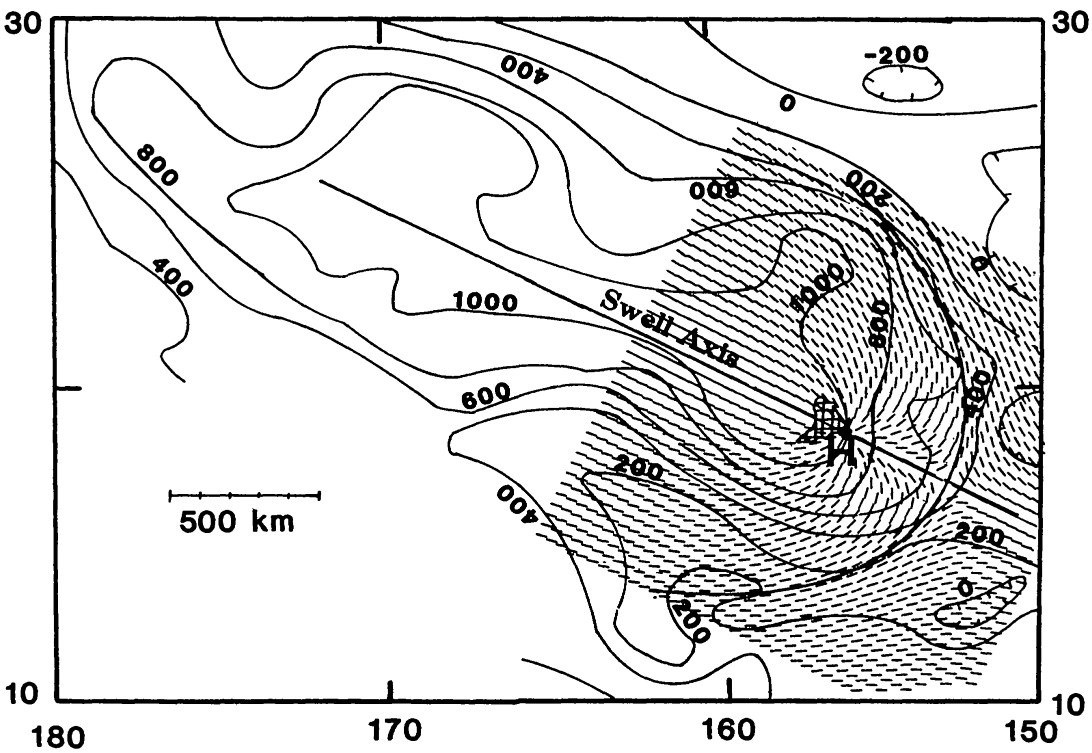

Figure 4 The excess elevation of the Hawaiian swell with the volcanic edifices removed [modified after Schroeder (1984)] is shown with flow directions in the asthenosphere near the nose of the swell [modified after Sleep (1990a)]. The thick parabola-like curve on the nose of the swell separates plume material and normal asthenosphere.

approximately half the plate velocity. Directly upstream of the hotspot, the radial flow from the plume is in the opposite direction of the drag-driven flow and a stagnation point occurs at r_s where

$$\frac{v_f r_f}{r_s} = \frac{v_L}{2}. \tag{12}$$

More generally the streamline through the stagnation point separates asthenosphere from the plume from normal asthenosphere (Figure 4). Watson & McKenzie (1991) obtain a similarly shaped streamline assuming that velocity varies away from the plume as $r^{-0.779}$ rather than r^{-1}. Thus, the shape of the nose of the swell is not a good indication of the detailed dependence of velocity on distance from the plume.

It is evident from this formalism and also Equations (8) and (9) that relatively weak plumes on fast plates should produce narrow swells. This creates one difficulty in that a narrow thermal swell is much more obscured by volcanic edifices and flexure than a wide one. A second difficulty is that very weak hotspots do not produce any tracks at all on fast plates. For example, this would preclude finding the weaker Atlantic hotspots on the Pacific plate (McNutt 1990).

Difficulties with the Thermal Uplift Theory

Crough (1978) and Detrick & Crough (1978) proposed the thermal rejuvenation and subsidence theory because individual atolls and also the axis of the Hawaiian swell appeared to subside at the rate predicted by the theory. An additional prediction of the theory is that the deeper part of the lithosphere is hotter than normal beneath the swell and that heat flow is elevated downstream of the hotspot. I illustrate the general relationship between heat flow and subsidence by giving the age-heat flow relationship,

$$q = q_0(\text{age})^{-1/2}, \tag{13}$$

where the constant q_0 implies 500 mW m^{-2} heat flow on 1 m.y. crust. The depth-age (6) and heat flow-age (13) relationships combine to yield a more general relationship between heat flow and subsidence:

$$q = -\text{constant}\frac{\partial E}{\partial t}. \tag{14}$$

Early heat flow measurements along the axis of the Hawaiian swell show the expected result (Von Herzen et al 1982). However, a later profile across the axis of the swell where the heat flow anomaly is expected to be largest showed much scatter but not the expected excess heat flow (Von Herzen et al 1989). It is hard to envision a time-dependent subsidence mechanism that moves with the lithospheric plate and does not involve thermal contraction by cooling to the surface. Equation (14) indicates that any such mechanism is inappropriate if the latter heat flow data are representative.

Another difficulty of the thermal uplift hypothesis comes from surface wave velocities along the Hawaiian swell (Woods et al 1991). The data show the expected velocity structure of unrejuvenated 80 m.y. lithosphere, rather than the expected rejuvenation age of 25 m.y. However, converted S to P waves indicate that hot material exists at 75 km depth beneath Oahu (Bock 1991).

Thermal subsidence is most easily observed on the rapidly moving Pacific plate where hotspots move quickly away from any dynamic uplift and subsidence mechanisms related to flow from the plume. Thus I return to the locality of Enewetak atoll which Detrick & Crough (1978) used to show that the cause of uplift (and subsidence) moves basically with the lithospheric plate as expected for thermal rejuvenation. As noted by Sleep (1990a), the asthenosphere moves more slowly relative to the hotspot than to the plate. The geometry of hotspot tracks in the Pacific allows this effect to be observed. Track directions before the Hawaii-Emperor bend at 43 Ma are from the NNE at a high angle to the current track direction from the ENE. The lithosphere beneath track segments older than 43 Ma (as well as those somewhat younger) has had time to move away from the plume asthenosphere originally below it. Thus a primarily asthenosphere uplift and subsidence mechanism would imply rapid subsidence after 43 Ma when the plate cleared hot asthenosphere and little subsidence thereafter. Enewetak erupted at 56 Ma (Lincoln & Schlanger 1991), close enough to the bend age to provide a useful test: The later history of Enewetak is compatible with thermal subsidence of rejuvenated lithosphere and if anything more rapid in the last 15 Ma than predicted by the theory (Quinn 1991, Lincoln & Schlanger 1991).

The seismic data from the Hawaiian swell are probably compatible with the subsidence data from Enewetak. The surface wave data are sensitive to intermediate lithospheric depths and thus imply that rejuvenation is not simply using the heat of plume material to change the geotherm to that of young oceanic crust over normal asthenosphere. However, hotter material moving with the lithosphere at greater depths would produce uplift and converted P-waves, but not high temperatures at the depths sensed by the surface wave experiments. More seismic data sensitive to thermal anomalies at greater depths and subsidence data from the same area would help resolve these issues.

CONCLUSIONS

The objective of this review is to discuss the current level of understanding of the causes of hotspots with concentration on physical processes in the interior of the Earth. At one end, certain conclusions can be made with reasonable confidence, most importantly that mantle plumes exist beneath vigorous hotspots with well developed tracks, like Iceland and Hawaii. The properties envisioned for plumes in Figure 1 have an intermediate level of confidence as the dynamics are still only qualitatively understood. Currently unresolved issues include the origin of hotspot swells and the fate of the hot material supplied by plumes. Incomplete understanding of the surficial aspects of hotspots detracts from understanding the deep interior. For example, the number of plumes and hence the structure of flow within the basal boundary layer are poorly understood because it is not evident which weak hotspots are supplied by their own plumes.

I expect the rapid progress toward consensus and understanding to continue in the immediate future. A wide variety of geological and geophysical methods have proved useful. Integration of massive amounts of petrological data into the physical framework, untangling the complicated history of closely spaced Pacific and Atlantic hotspots, and improved subsidence and uplift histories of hotspot swells are important tasks for surface observers. Direct mapping of the basal boundary layer by seismologists seems possible. The physical aspects of the hypothesis are now well enough defined that theoretical fluid dynamicists can include the salient features in their models. Conversely, understanding of hotspots will lead to progress on other topics of Earth science, particularly with regard to the effects of hotspot tracks on field-scale geology and tectonics and the dynamics of the interaction of plumes with flow associated with plate tectonics.

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