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# MANTLE PLUMES AND HOT SPOTS

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## Introduction

Hotspots are defined as anomalous volcanism that cannot be attributed to plate tectonics, unlike that associated with island arcs and spreading ridges. Mantle plumes, which are upwelling instabilities from deep in Earth's mantle, are thought to be responsible for hotspots that are relatively stationary, resulting in chains of islands and seamounts on moving oceanic plates. The volcanic rocks associated with hotspots have signatures in trace elements and isotopes distinct from those observed at mid-oceanic ridges and island arcs. Seismic imaging has revealed low-velocity anomalies associated with some deep-rooted hot mantle plumes, but images of their full-depth extent are of limited resolution, thus evidence for plumes and hotspots is primarily circumstantial. Commonly, it is not even clear which areas of intra-plate volcanism are underlain by a mantle plume and should be counted as a hotspot.

## Surface Expression of Hotspots

The primary surface expression of mantle plumes consists of hotspot tracks. These are particularly evident in the oceans as narrow ($\approx 100$ km) chains of islands and seamounts, such as the Hawaiian–Emperor chain, or as continuous aseismic ridges, such as the Walvis Ridge, up to several kilometres high. These tracks are thought to form as lithospheric plates move over plumes. The active hotspot is at one end of the chain; radiometric dating has determined that the ages of the volcanics along the chain tend to increase with distance from the active hotspot. Interpretation of age data is complicated, because volcanics do not necessarily erupt directly above a plume. Late-stage volcanism may occur several million years (My) after passage over a plume. Many hotspot tracks begin with a flood basalt or large igneous province. Volcanic volumes and age data indicate that these form during short time-spans with much higher eruption rates than are found at present-day hotspots. Examples of continental flood basalts (CFBs) include the Deccan Traps (associated with the Reunion hotspot) and the Parana basalts (associated with the Tristan hotspot). The Deccan Traps have erupted a volume of $\approx 1.5 \times 10^6$ km$^3$ within less than 1 My, whereas the present-day eruption volume at the Reunion hotspot is $\approx 0.02$ km$^3$ year$^{-1}$. For other tracks, older parts have been subducted, and yet others, particularly shorter ones, begin with no apparent flood basalt. The length of tracks shows that hotspots may remain active for more than 100 My. For example, the Tristan hotspot track indicates continuous eruption for 120 My. Numerous shorter tracks exist as well, particularly in the south central Pacific, commonly without clear age progression. This may indicate either that the region is underlain by a broad upwelling or that widespread flow from a plume is occurring beneath the lithosphere, with locations of volcanism controlled by lithospheric stresses. Geometry and radiometric age data of hotspot tracks indicate that the relative motion of hotspots is typically slow compared to plate motions. However, for the Hawaiian hotspot between 80 and 47 million years ago (Ma), inclination of the magnetization of volcanics indicates formation at a palaeolatitude further north than Hawaii, with hotspot motion southward of several centimetres per year. The Hawaiian–Emperor bend may therefore represent more than a change in Pacific plate motion. In most other cases in which palaeolatitude data are available, inferred hotspot motion is slow or below detection limit. Associated with many tracks is a hotspot swell ($\approx 1000$ km wide, with up to 3 km anomalous elevation). Swells are associated with a geoid anomaly. Swell height slowly decreases along the track away from the active hotspot, and the swell also extends a few 100 km 'upstream' from the hotspot. The geoid-to-topography ratio remains approximately constant along swells, and this value indicates isostatic compensation at depths $\approx 100$ km. From the

product of swell cross-sectional area and plate speed relative to the hotspot, plume anomalous mass flux (volume flux times density anomaly of plume relative to ambient mantle - this is the quantity that is directly estimated from observations), volume flux (volume flowing through plume conduit per time unit), and heat flux, (volume flux density times heat capacity times temperature anomaly of plume relative to ambient mantle), can be estimated. For Hawaii, the hotspot track and swell are both evident ([Figure 1]). The anomalous mass flux determined for Hawaii is the largest of any plume. Its estimated heat flux corresponds to ≈16% of global hotspot heat transport; global hotspot heat transport in turn is estimated to be ≈5% of total global heat flux.

Direct measurements of heat flow above hotspot swells yield only small anomalies ($\approx 5$–$10\,\mathrm{m\,W\,m^{-2}}$) or no anomalies at all. This may indicate that the swell is caused by buoyant uplift due to hot plume material spreading beneath the lithosphere, rather than by thermal erosion and heating of the lithosphere. The amount of volcanics produced at hotspots is estimated to be $<1\,\mathrm{km^3\,year^{-1}}$, ≲5% of volcanics produced at mid-ocean ridges. Magma production at mid-plate hotspots thus appears less efficient than at mid-ocean ridges. However, during formation of large igneous provinces, magma production rate and heat flow were higher than at present.

Hotspots cluster in two antipodal regions around the Pacific and Africa. They are mostly absent from regions where subduction has occurred in the past 100 My, and tend be in highs of the 'residual' geoid (actual geoid minus contribution of subducted slabs). Probably the hotspot plumes are not the primary cause of these geoid highs. Rather, both may be due to the same cause: a less dense and presumably hotter lower mantle in these regions. Plate boundaries, in particular mid ocean ridges, may move across hotspots; hotspots may hence, successively or simultaneously, leave tracks on different plates. Examples are the Reunion hotspot, with tracks on the Indian and African plates, and the Tristan hotspot, with tracks on the South American and African plates. If a plume is located close to a spreading ridge, eruption of plume material may occur not only directly above the plume, but also at the section of the ridge
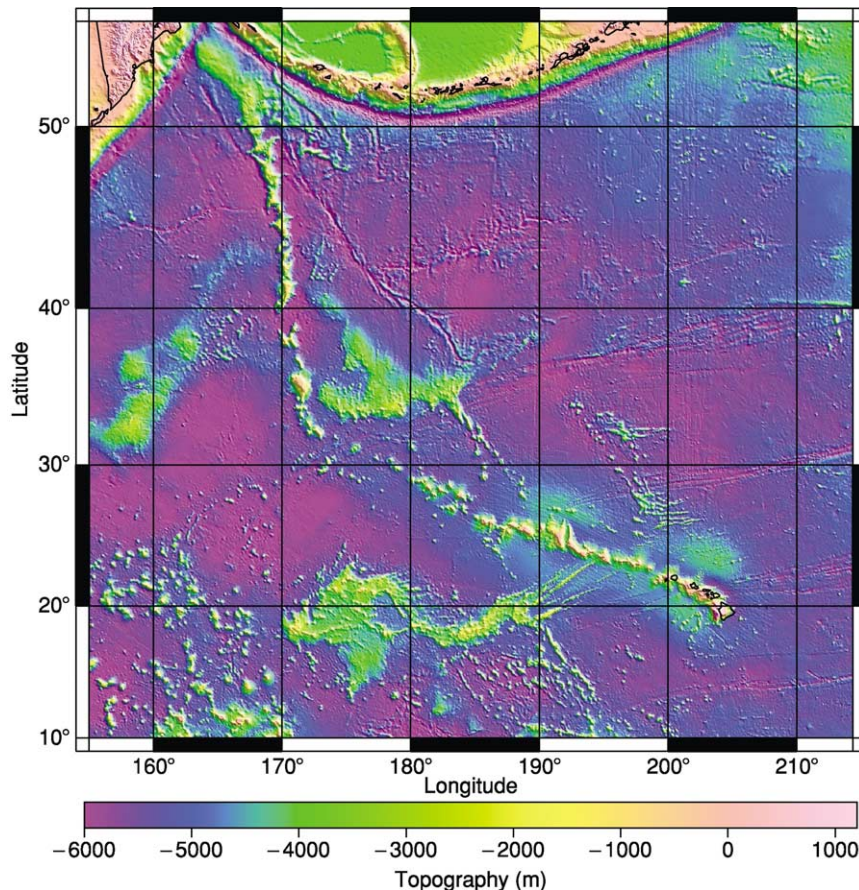


**Figure 1** Topographic map of the North Pacific, showing both the swell and the track associated with the Hawaiian plume.

close to the hotspot. Elongated aseismic ridges may be formed by volcanics above channels along which plume material flowed to the spreading ridge (for example, in the Musicians Seamounts near 200° E, 25° N) ([Figure 1](#)). Hotspots and their tracks are less obvious on continents. The Yellowstone National Park area in the United States is frequently regarded as a hotspot plume, with the Snake River Plain being its hotspot track.

## Seismic Images of the Mantle Plumes

Column-like anomalies of low seismic velocity associated with high temperatures are expected under hotspots, if hotspots are the surface expression of mantle plumes. Seismic imaging of hotspots has advanced in the past decade, and seismic images beneath some hotspots have been obtained. Commonly, these are imaged only at specific depths. For some hotspots, no low-velocity anomalies have been found.

### Upper Mantle

Global mapping of the upper mantle by long-period (50–300 s) surface waves has revealed low seismic velocities associated with hotspots and spreading ridges. There is a distinct difference between hotspots and ridges, concerning the depth extent of these low-velocity regions : low seismic velocities beneath hotspots extend to a depth of 200 km, whereas low seismic velocities beneath ridges are confined to the upper 100 km. This suggests that hotspots are caused by active upwellings (mantle plumes) with deeper sources, compared to ridges, which may be caused by passive upwelling. The 200-km depths of the slow velocities under hotspots do not necessarily correspond to the actual source depths of mantle plumes, but rather to the depths to which surface waves can resolve. Seismic array observations have been carried out in hotspot regions to resolve fine structures such as plume conduits. A recent example of an S-velocity model beneath the Icelandic hotspot was obtained from body and surface wave data recorded by a temporary seismic array; a low-velocity plume can be seen beneath the hotspot ([Figure 2](#)). A 200-km-thick low-velocity zone extends laterally beneath Iceland; a vertical column of low velocities under central Iceland extends to a depth of at least 400 km. Similar array observations carried out at other hotspots (e.g., Hawaii, Yellowstone, and Massif Central) detected low-velocity anomalies extending to sublithospheric depths in the upper mantle.
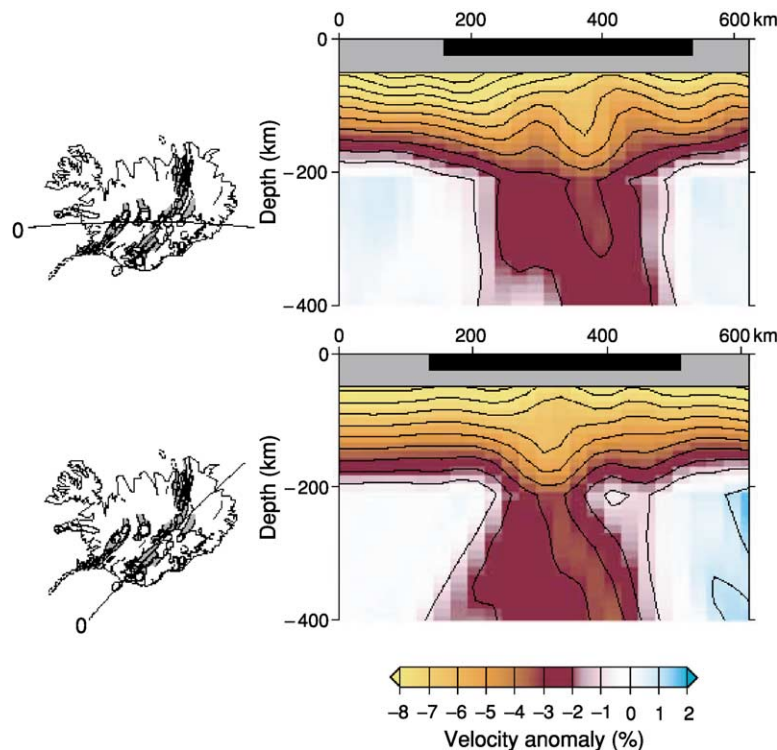


**Figure 2** The S-velocity profile beneath the Icelandic hotspot. Reproduced with permission from Allen RM, Nolet G, Morgan WJ, *et al.* (2002) Imaging the mantle beneath Iceland using integrated seismological techniques. *Journal of Geophysical Research—Solid Earth* 107(B12): 2325, doi:10.1019/2001 JB000595.

## The Transition Zone

The mantle transition zone is bounded by seismic discontinuities at depths around 410 and 660 km (referred to hereafter simply as '410' and '660'); the discontinuities at these two depths are interpreted to be due to olivine–β-olivine and postspinel phase changes, respectively. Topography of the 410 and 660 provides important data for understanding thermal structure and the ascending process of hot plumes; the 410 and 660 are expected to be depressed and elevated, respectively, in and near a hot plume, because the olivine–β-olivine and the postspinel phase changes have positive and negative Clapeyron slopes, respectively. The thickness of the transition zone (defined as the interval between the 410 and 660) is also used as a measure for thermal anomalies, because estimates are less affected by upper mantle velocity structure and therefore can be more reliably determined. Discontinuity depths are determined by detecting waves reflected or converted at the discontinuities, and measuring their timings. Using P- to S-converted waves recorded by regional seismograph arrays, the transition zone beneath Iceland and Hawaii has been estimated to be thinner than in the surrounding mantle by ≈20 and ≈40 km, respectively, for areas 400 km in diameter. Amounts of thinning can be converted to temperature anomalies, with the Clapeyron slope obtained experimentally, suggesting temperatures of ≈150 K and ≈300 K beneath Iceland and Hawaii, respectively. The transition zone beneath the Society hotspot in the South Pacific Superswell was estimated, using S waves reflected beneath

the discontinuities, to be 20–30 km thinner than it is in the surrounding area (an area 500 km in diameter), suggesting temperatures that are 150–250 K hotter (Figure 3). These results suggest that mantle plumes beneath these hotspots originate at least from transition zone depths, and possibly from the lower mantle.

## Lower Mantle and D″ Layer

To explore the deeper structure beneath hotspots requires analysis of body waves, which penetrate much deeper in the mantle. Global tomographic studies using body waves in the lower mantle have identified possible 'plume conduit signatures' beneath some hotspots, but not all. At present, different studies using different data and methods give somewhat inconsistent results for the same hotspot, making it difficult to derive definitive conclusions about the geometry of plume conduits and the source depths of hotspots. Different depth extents may indicate various source depths for hotspots. Two robust features – broad, low-velocity anomalies – have been identified beneath the South Pacific and Africa (Figure 4); often referred to as 'superplumes', these features have diameters of 1000–2000 km. The latter is tilted from south-west to the north-east, suggesting that the plume conduit may be advected by large-scale mantle flow. These superplumes may be related to the hotspots in the South Pacific Superswell and Africa, although robust seismic images of the plume conduits to each of the hotspots have not been obtained.
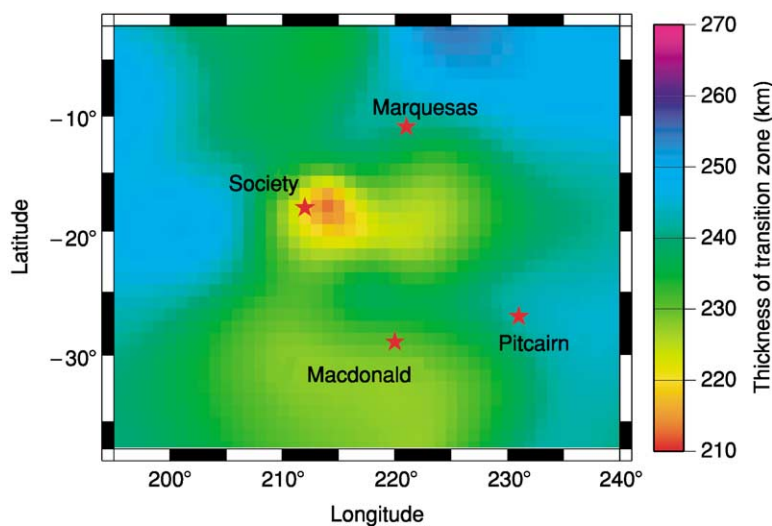


**Figure 3** Thickness of the transition zone beneath the South Pacific Superswell. Deviation from the global average is shown (colour bar, kilometres). Stars represent hotspots. Reproduced with permission from Niu F, Solomon SC, Silver PG, Suetsugu D, and Inoue H (2002) Mantle transition-zone structure beneath the South Pacific Superswell and evidence for a mantle plume underlying the Society hotspot. *Earth and Planetary Science Letters* 198: 371–380.
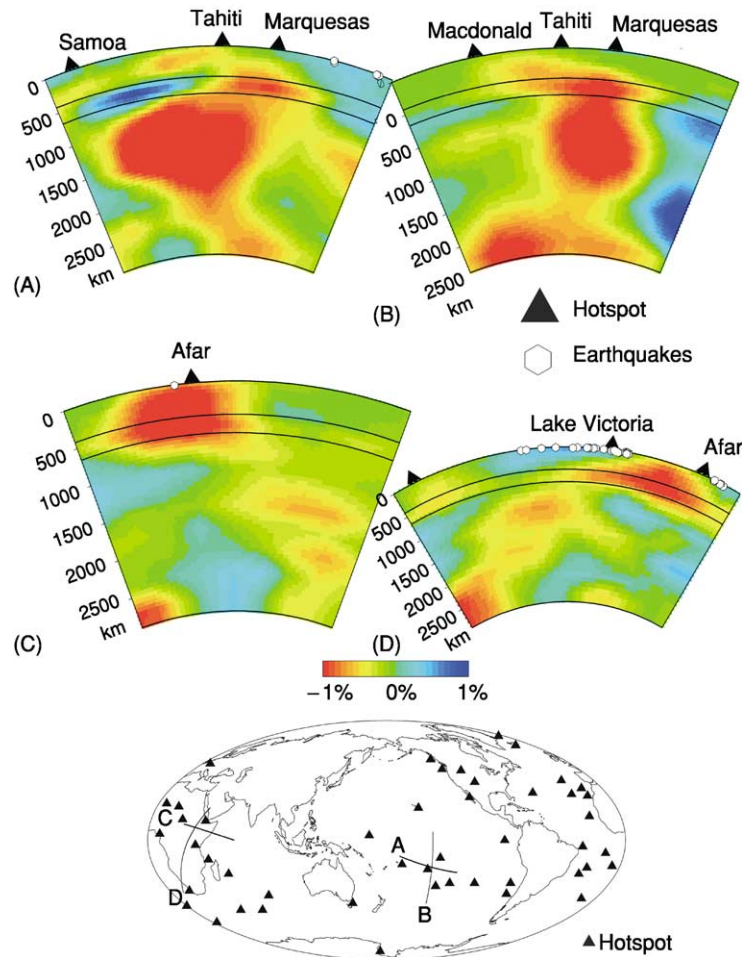
**Figure 4** The P-velocity profile of the whole mantle beneath the South Pacific Superswell (top) and Africa (middle). (Bottom) Locations of profiled areas. Reproduced with permission from Zhao D (2001) Seismic structure and origin of hotspots and mantle plumes. *Earth and Planetary Science Letters* 192: 251–265.

Over most of the lower mantle depth range, lateral variation of seismic velocities is less than it is in the upper mantle, but pronounced lateral variation does occur at depths below 2500 km. Broad, low-velocity anomalies located beneath the South Pacific Superswell and Africa seem to be linked to the super-plumes; high-velocity anomalies beneath circum-Pacific regions may be associated with subducted slabs. At the bottom of the D″ layer (the lowermost part of the mantle), above the core and mantle boundary (CMB), 5- to 50-km-thick zones of very low velocities have been found; these ultra-low-velocity zones (ULVZs) have low P and S velocity anomalies (10% or greater) and may represent a layer of partial melting and/or chemical heterogeneities. There is a spatial correlation between the ULVZs and hotspots, suggesting that the ULVZs may be a possible source of hotspots. Many plume conduits may be too narrow to be detected by present seismic data and techniques;

dense seismic arrays on the land and seafloor and new analysis techniques may be required to resolve mantle plumes at full depth.

## Petrological and Geochemical Signatures of Hotspot Rocks

Volcanic rocks associated with plume-related (hotspot) volcanism are quite variable in rock type and chemical composition and show significant correlation with the tectonic environment in which they erupt. Within the oceanic environment, hotspot volcanism produces mainly basaltic lavas, often called ocean island basalts (OIBs); these have chemical characteristics distinct from those of mid-ocean ridge basalts (MORBs), of which most of the ocean floor is composed. Hotspot volcanism in the oceanic environment sometimes produces voluminous oceanic plateaus or rises (e.g., the Ontong Java Plateau). In

continental environments, mantle plume activity is commonly manifested by eruption of huge floods of basaltic lavas, commonly called continental flood basalts (CFBs). Less voluminous magmatic activities also occur in continents with extensional tectonics such as intracontinental rifts and grabens. Volcanics in continental rift systems are thought to be products of mantle plume activity, although some continental rift volcanism is not clearly related to mantle plumes.

OIBs are commonly olivine-bearing lavas with sub-alkalic (tholeiitic) and alkalic compositions. Some large ocean islands, such as Hawaii and Iceland, are dominantly composed of tholeiitic basalts and their derivatives, with small amounts of alkalic rocks at later stages of eruptive sequences. A vast majority of ocean islands are composed entirely of alkalic basalts and their derivatives. The major element composition of tholeiitic OIBs is generally similar to that of MORBs, although lavas with higher MgO content are more abundant than at mid-oceanic ridges. Alkali olivine basalt is the most abundant type in alkalic OIB suites, but more alkalic lavas, such as basanite and nephelinite, also commonly erupt in many ocean islands. Compared with MORBs, both tholeiitic and alkalic OIBs have higher concentrations of incompatible trace elements, and their chondrite-normalized rare earth element (REE) patterns show variable but strong enrichment in light REEs relative to heavy REEs. Alkalic OIBs are typically more enriched in incompatible elements, including light REEs, compared to tholeiitic OIBs. Sr–Nd–Pb–Os isotopic ratios of OIBs show quite large variations relative to MORBs, which extend from nearly MORB-like values towards one component with highly radiogenic Pb isotopes (designated by the term 'HIMU', referring to the high U/Pb source ratios) and other components with nonradiogenic Pb isotopes (designated as 'enriched mantle 1', or EM1) and with high $^{87}Sr/^{86}Sr$ ratios (EM2). Variation in helium isotopic ratios of OIBs extends considerably towards the primordial value with high $^{3}He/^{4}He$ ratios, although some OIB suites have slightly lower $^{3}He/^{4}He$ ratios than MORBs have. Volcanic rocks of oceanic plateaus and rises are commonly aphyric lavas with tholeiitic composition. Their major element composition is relatively homogeneous compared to tholeiitic OIBs and more akin to MORBs. Oceanic plateau basalts have trace element characteristics similar to those of MORBs, but they are slightly more enriched in incompatible elements. Chondrite-normalized REE patterns are typically flat or slightly enriched in light REEs. Radiogenic isotope ratios of oceanic plateau basalts are quite variable, but their entire range is smaller than that of OIBs.

Petrological features of CFBs are similar to those of oceanic plateau basalts. Most CFB suites are composed of voluminous aphyric lavas with tholeiitic composition, and their major element compositions cover similar ranges as those covered by MORBs. However, compared to MORBs, CFBs generally have higher FeO/MgO ratios, and $SiO_2$-rich (andesitic) lavas are more dominant in CFBs. Most CFB suites are associated with minor amounts of alkalic rocks. In contrast to major elements, trace element features of CFBs are more like those in OIBs than in oceanic plateau basalts and MORBs. CFB suites are highly enriched in incompatible elements, sometimes even more than OIBs are, and have REE patterns with strong enrichment in light REEs. Sr–Nd–Pb–Os isotopic ratios of CFBs are much more variable than those of OIBs are, extending beyond the highest $^{87}Sr/^{86}Sr$ and the least radiogenic Pb isotopes of OIBs. Volcanic rocks associated with continental rift systems show considerable diversity in rock type, from transitional olivine basalt-rhyolite to strongly alkalic nephelinite–phonolite series, melilitites, and even carbonatite lavas (Ol Doinyo Lengai, Tanzania). A noticeable feature of continental rift volcanics is that acidic rocks, such as rhyolite, trachyte, and phonolite, are more abundant than in other environments. Enrichment of incompatible elements in continental rift volcanics is as great as or sometimes much greater than in alkalic OIBs. Variations of radiogenic isotope ratios are as wide as those in CFBs.

Geochemical diversity in plume-related volcanic rocks is generally attributed to chemical heterogeneity of mantle plume material. In particular, isotopic variations of OIBs, which are least affected by crustal material, clearly indicate that mantle plumes contain several geochemical components. The origin of such geochemical components in the mantle is still in dispute, but the most commonly accepted model is that subducted crustal materials (oceanic basaltic crust plus silicic sediments) contribute to formation of distinct geochemical reservoirs in the mantle, such as HIMU, EM1, and EM2. Chemical interaction between core and mantle has also been proposed to explain isotopic diversity of OIBs, especially for Pb and Os. High $^{3}He/^{4}He$ ratios in some OIB suites require the existence of some primordial volatile component in mantle plumes. Extremely radiogenic Sr isotopic ratios observed in some CFBs and continental rift volcanics have been explained by contamination or assimilation of continental crust. Major element diversity of plume-related volcanic rocks has been interpreted in terms of potential temperature of mantle plumes. A conventional view is that tholeiitic basalts in relatively large ocean islands and CFB suites are produced at higher degrees of melting in hot

mantle plumes, and alkalic rocks are produced at lower degrees of melting in colder plumes. On the other hand, the major element diversity of plume rocks can be explained by heterogeneity in major element composition of mantle plumes, because subducted crustal material, if contained in mantle plumes, may produce both tholeiitic and alkalic magmas under mantle conditions. In this case, no significant excess temperature is required for the plume, because the melting temperature of crustal materials is much lower than that of mantle peridotite.

## Dynamics of Mantle Plumes

In Earth's mantle, the Rayleigh number, a measure of the vigour of convection, is estimated to be $5 \times 10^6$ to $5 \times 10^7$. For this range, time-dependent convection, with instabilities originating at thermal boundary layers, is expected. Plume formation may, however, be suppressed by large-scale flow related to plate motions, which, if sufficiently fast, advects growing instabilities towards large-scale upwellings before individual plumes can form. For somewhat slower large-scale flow, individual plumes may exist, but are advected towards, and hence cluster in, regions of large-scale upwellings. A thermal boundary layer exists above the core–mantle boundary, which is therefore a likely plume source depth. Another thermal boundary and possible source depth is at the 660. Plumes may also originate at chemical boundaries within the mantle. No conclusive evidence exists for those, but anticorrelation of bulk sound and shear wave anomalies in the lowermost mantle indicates that the D″ layer at the base of the mantle may be chemically distinct from the overlying mantle. Topography of the chemical boundary could be up to a few 100 km, and plumes may rise from the high points (cusps) on the boundary. Plumes may entrain material from the underlying layer; the amount of entrained material depends on the density difference between the two layers, due to chemical and temperature differences. Both laboratory and numerical experiments show that a plume that is less viscous than the surrounding mantle tends to have a large, roughly spherical or mushroom-shaped head, followed by a narrow tail (conduit) connecting the head and the source region (Figure 5). Scaled to Earth dimensions, a head diameter of several 100 to 1000 km is expected, if it rises from the lowermost mantle. Such a size is also required to explain the volume of many large igneous provinces, thus adding support to the notion that these originate from upwellings from the lowermost mantle, whereas hotspots not associated with flood basalts may be caused by plumes from shallower depth. Conduits should be only ≈100 km wide. Thinner conduits may occur for strongly temperature-dependent viscosity. If mantle viscosity increases strongly with depth, conduit diameter could be up to a few hundred kilometres in the lower mantle. Ascent times of plume heads through the whole
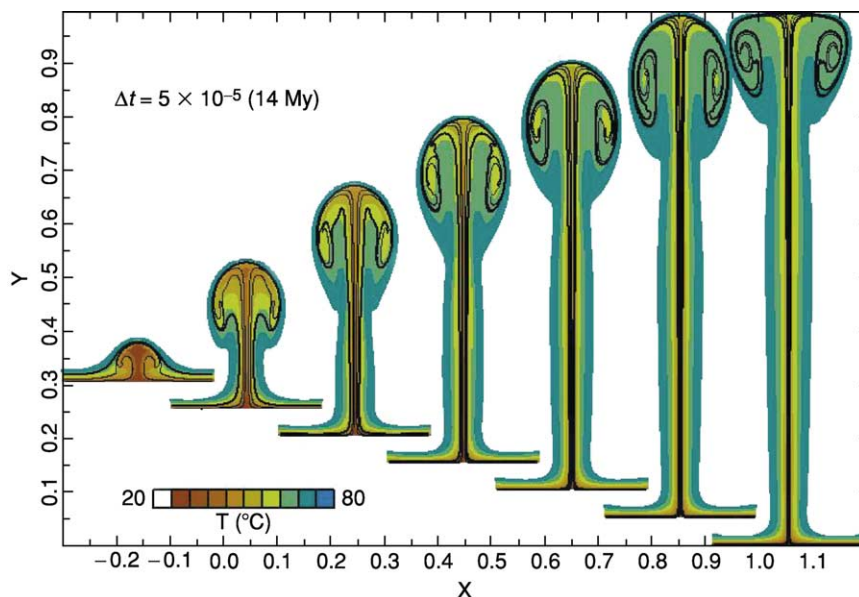


**Figure 5** Numerical model of an axisymmetric mantle plume with strongly temperature-dependent viscosity. Colours indicate temperature (red = hottest, blue = ambient mantle); black lines indicate marker chains. Plots are at 14-My time intervals. Reproduced with permission from van Keken P (1997) Evolution of starting mantle plumes: a comparison between numerical and laboratory models. *Earth and Planetary Science Letters* 148: 1–11.

mantle are estimated to be $\approx 10$ My to several tens of millions of years. For nonlinear mantle rheology, decrease of effective viscosity due to larger stresses around plume heads may cause rise times shorter than for Newtonian viscosity. As a plume head rises, the surrounding mantle heats up and becomes buoyant and less viscous. Mantle material is entrained into the rising plume, which therefore contains a mixture of materials from the source region and ambient mantle. Formation of a flood basalt is thought to occur when a plume head reaches the base of the lithosphere. Subsequently, the conduit may remain in existence as long as hot material is flowing in at its base – for 100 My or longer. The conduit consists of a narrow core, where most of the material transport occurs, and a thermal halo. Due to thermal diffusion, heat is lost from plume conduits as they traverse the mantle. Though strong plumes, such as the Hawaii plume, are not significantly affected, heat loss significantly reduces the temperature anomaly expected for weaker plumes. For weak plumes from the core–mantle boundary, with anomalous mass flux of $\lesssim 500$–$1000\,\mathrm{m}^3\,\mathrm{s}^{-1}$, the sublithospheric temperature anomaly is low and no melting is expected. Weak hotspots may therefore have shallower origins. On the other hand, the temperature anomaly of strong plumes, inferred from observations, is much less than expected from the temperature drop across a thermal boundary layer between core and mantle. This may indicate a chemically distinct layer at the base of the mantle, with plumes rising from its top. Mantle plumes may coexist with superplumes, and conduits are expected to be tilted and distorted in large-scale mantle flow. The rising of a tilted conduit may cause further entrainment of ambient mantle material. If the tilt exceeds $\approx 60°$, the conduit may break into separate diapirs, which may lead to extinction of the plume. As a consequence of conduit distortion, overlying hotspots are expected to move. Hence, mantle plumes probably do not provide a fixed reference frame. However, if plumes arise from a high-viscosity lower mantle, hotspots should move much more slowly than lithospheric plates move. Conduits are likely to be time variable, with disturbances traveling along them; these may be wave-like or may take the shape of secondary plume heads. Waves are associated with increased conduit flux, which may explain flux variations in mantle plumes. Ascending plumes interact with mantle phase transitions. The 660 somewhat inhibits flow across but is unlikely to block penetration of plumes. Experiments involving high pressure suggest phase relations of a pyrolite mantle such that, at the high temperatures of mantle plumes, this phase boundary does not hinder flow across. Beneath the lithosphere, buoyant plume material

flows out of the conduit, spreads horizontally in a low-viscosity asthenosphere, and is dragged along with moving plates. Plume material buoyantly lifts up the lithosphere and causes a hotspot swell. Partial melt extraction at the hotspot may leave behind a buoyant residue that also contributes to swell formation. Plume material does not necessarily erupt directly above the conduit. It may also flow upward along the sloping base of the lithosphere, and enhanced melting may occur at steep gradients. A sloping base exists near spreading ridges. If ridge and hotspot are less than a few hundred kilometres apart, eruption of volcanics may occur at the ridge rather than, or in addition to, directly above the plume. Also, in other cases, such as in Africa, the spatial distribution of plume-related melting and magmatism may be controlled by the lithosphere rather than by the plume position. Formation of vertical fractures and ascent of magma through the lithosphere preferably occur for tensile lithospheric stresses. Loading of the lithosphere by hotspot islands causes stresses that may influence formation of fractures and therefore determine the spacing of hotspot islands along tracks. Feeding of plume material to a nearby ridge may put the lithosphere above the plume under compression and shut off eruption directly above the plume. If a hotspot (e.g., Iceland) is located close to the ridge, the viscosity contrast between plume and ambient mantle may become a factor 1000 or more beneath thin lithosphere. Such large viscosity variations facilitate ridge-parallel flow of plume material and help to explain geochemical anomalies south of Iceland. Propagation of pulses in plume flux explains the V-shaped topography and gravity anomalies at the Reykjanes Ridge. Probably not all intraplate volcanism is caused by plumes as described. In many cases, the origin of intraplate volcanism may be shallow, due to cracks in the lithosphere caused by tensional stresses, or due to edge-driven convection at locations where lithospheric thickness varies laterally.

## See Also

**Igneous Processes**. **Large Igneous Provinces**. **Lava**. **Plate Tectonics**. **Rocks and Their Classification**. **Seamounts**. **Tectonics:** Propagating Rifts and Microplates At Mid-Ocean Ridges; Seismic Structure At Mid-Ocean Ridges.

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# MARS

*See*   **SOLAR SYSTEM: Mars**

# MERCURY

*See*   **SOLAR SYSTEM: Mercury**