ENCYCLOPEDIA OF SOLID EARTH GEOPHYSICS, SPRINGER, DOI:10.1007/978-90-481-8702-7, 2011

Mantle plumes

Cinzia G. Farnetani[1] and Albrecht W. Hofmann[2,3]

1. Hotspots

Plate tectonics provides a framework for interpreting volcanism at plate boundaries, namely along spreading ridges (divergent boundaries) and subduction zones (convergent boundaries). However, it does not explain intraplate volcanism, such as Hawaii, nor the excess volcanism along certain sectors of a spreading ridge, as observed in Iceland. In these regions, called hot spots, the volcanic activity can last more than a hundred million years, as indicated by the resulting age progressive volcanic chain formed during the hotspot's life-time. For example, the 6000 km long Hawaiian-Emperor volcanic chain (Figure 1) is an alignment of roughly one hundred volcanoes of progressively increasing ages, created during the last 80 Ma, as the Pacific plate drifted over the Hawaiian hot spot. The vigorous and long-lasting Hawaiian volcanism readily captured the interest of geophysicists: in 1963 Tuzo Wilson proposed that the Hawaiian chain was not caused by lithospheric fissures, but by convection currents in the mantle, and in 1971 Jason Morgan suggested that hotspots are the surface expression of mantle plumes upwelling from the Earth's lowermost mantle. Mantle plumes can be defined as localized upwelling currents of solid rocks that are hotter, and thus less dense, than the surrounding mantle.

Hotspots are traditionally characterized by some or all of the following features, although exceptions do occur: (1) An age progressive volcanic chain whose linear trend is consistent with the direction of plate motion. (2) The onset of hotspot magmatism is often marked by a Large Igneous Province (LIP), a term including continental flood basalts (CFB) and oceanic plateaus. Estimated extruded volumes of CFBs are 1-2 $\times$ 10^6 km^3, with perhaps similar (but unknown) volumes of intrusive magmatism, whereas oceanic plateaus can be one order of magnitude more voluminous (Coffin and Eldholm, 1994). According to the mantle plume initiation model (Richards et al., 1989), the transient and episodic LIP magmatism corresponds to melting of a large plume head, and the subsequent hotspot activity is associated with the long-lasting and narrow plume tail. (3) The topographic swell is a region of anomalously high topography with a lateral width of 1000 km in the direction normal to the volcanic chain (Wessel, 1993), and with an elevation of 1 km, which decreases along the chain. (4) Hotspot basalts are geochemically distinct and more diverse than mid-ocean ridge basalts.

Some hotspots, however, do not show all of the above features. For example, many volcanic chains lack a clear age progression and, according to Ito and van Keken (2007), there are only thirteen long-lived (>50 Ma) and eight short-lived (<20 Ma) age-progressive volcanic chains. Some hotspots are not associated with a LIP, for others, such as Hawaii and Louisville, the chain terminates in a subduction zone, so that the complete time record of the volcanic activity is lost. In other cases, a number of LIPs have no associated volcanic track (e.g., Shatsky, Hess) and their origin is still a matter of debate. In summary, plume magmatism is often, but not always, explained by the classical thermal plume model with a voluminous spherical head followed by a narrow columnar conduit.

2. Global hotspot distribution and hotspot fixity

Over the years, the estimated number of hotspots has varied from 20 (Morgan, 1971), to a maximum of 117 in the 1980s, whereas in recent compilations (e.g., Ito and van Keken, 2007) the number ranges between 45 and 70 (see Schubert et al., 2001 and references therein). Hotspots younger than 100 Ma are generally active, although their vigor may vary considerably; older hotspots are either waning (100<age<140 Ma) or inactive (age>150 Ma). The best-defined hotspots appear to be relatively stationary over time and are used as a reference frame to determine absolute plate motions. Most hotspots are situated in the oceans, one exception being Yellowstone (USA), whose volcanic activity can be traced back to the 16 Ma old Columbia River flood basalt. The discrepancy in the number of continental vs. oceanic hotspots has three probable reasons: First, the arrival of a mantle plume can weaken the lithosphere and enhance continental break-up (Courtillot et al. 1999). For example, the magmatic activity of the Iceland plume started with the eruption of the North Atlantic Tertiary Igneous Province (62 Ma ago) and was followed by continental rifting and the appearance of an oceanic spreading ridge after a few million years. In this respect, mantle plumes may have been influential in modifying plate boundaries. The second reason concerns the different thickness (100-150 km) between oceanic and continental lithosphere: weak plumes may not melt beneath a thick continental lithosphere, but are likely do so at lower pressure, beneath a thinner oceanic lithosphere. Third, the lack of hotspots in continents could be due to the continents accumulating over downwellings, precisely where plumes are very unlikely to ascend.

The main oceanic hotspots, and their related LIPs (Figure 2) in the Atlantic Ocean are: Iceland (0-62 Ma, LIP: the North Atlantic Tertiary Igneous Province), and Tristan da Cunha (0-125 Ma, LIP: the Paran-Etendeka Province), whereas other hotspots like the Azores (0-20 Ma), Canaries (0-68 Ma), and Cape Verdes (0-Miocene) are not clearly associated with a LIP. In the Indian Ocean the main hotspots are: la Runion (0-65 Ma, LIP: Deccan Traps), Kerguelen (0.1-120 Ma, LIP: the Kerguelen plateau), and Afar (0-30 Ma. LIP: Ethiopian Traps). In the Pacific Ocean they are: Hawaii (0-76 Ma, unknown LIP, since the chain ends in a subduction zone), Louisville (1-77 Ma, possibly 120 Ma if the associated LIP is the Ontong-Java Plateau), and Galapagos (0-possibly 90 Ma if the associated LIP is the Caribbean Plateau). There are also some hotspots (e.g., in French Polynesia) that may have a plume origin, but with an unclear age progression. Finally, the oceanic floors are littered by hundreds of thousands of volcanic seamounts that

[1]Institut de Physique du Globe, Paris, France
[2]Max-Planck-Institut für Chemie, Mainz, Germany
[3]Lamont-Doherty Earth Observatory, Columbia University New York, NY, USA

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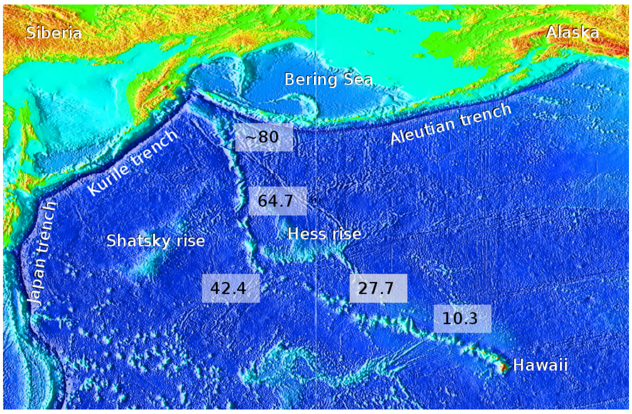

Figure 1. Sea floor topography from Smith and Sandwell (1997). Numbers indicate some ages (in Ma) along the Hawaiian-Emperor volcanic chain.

are certainly not created by plumes (e.g., Hieronymus and Bercovici, 2000; Clouard and Gerbault, 2008).

An interesting aspect of hotspots is that they are relatively stationary with respect to each other. Their respective motions of 1-2 cm/yr are much less than the plate velocities, thereby approximating a fixed-hotspot frame of reference. However, Tarduno et al., (2009) calculated the age and the paleolatitude of volcanoes belonging to Hawaiian-Emperor chain and concluded that the Hawaiian hotspot moved southward at 4-5 cm/yr during the period 80-47 Ma ago, whereas it remained relatively fixed (<2 cm/yr) afterwards. This complex behavior is probably due to the dynamical interaction between upwelling plumes and the 'mantle wind' induced by large scale mantle convection (Steinberger et al., 2004).

3. Evidence for mantle plumes

There is now wide-spread agreement on the existence of mantle plumes, although contrasting views do exist (see: Foulger and Natland (2003) and the website: www.mantleplumes.org). Several lines of evidence support the existence of mantle plumes:

First, the Earth's Rayleigh number, which governs the vigor of convection, is sufficiently high (10^6-10^8) to insure that mantle convection is time dependent and that its thermal boundary layers become repeatedly unstable. A thermal boundary layer (TBL) is a zone characterized by a high temperature gradient, since heat is transported dominantly by conduction. Fluid dynamical considerations indicate that Rayleigh-Taylor instabilities from a hot TBL generate thermal plumes (Loper and Stacey, 1983). A prominent TBL in the Earth's mantle is possibly the D" zone, which extends 100-200 km above the core-mantle boundary. The existence of another TBL, for example at 660 km depth, is still a matter of debate, but it seems unlikely, since the endothermic phase transition does not constitute a complete barrier to mantle convection. Mantle plumes are therefore expected to rise from the lowermost mantle, forming cylindrical conduits with a radius of 50-150 km. Such values are based on fluid dynamics, whereas conduit radius estimated by seismology (e.g., Montelli et al., 2004; Wolfe et al., 2009) are much broader.

Second, seismic detection of narrow plume conduits is challenging, but the first exhaustive study by Montelli et al. (2004) found that at least six plumes (Ascension, Azores, Canary, Easter, Samoa, and Tahiti) extend into the lowermost mantle, whereas others are confined to the upper mantle, and in some cases the model resolution was insufficient. Recently, an extensive ocean-bottom seismological survey of Hawaii (Wolfe et al., 2009) has shown that a low seismic velocity anomaly extends into the lower mantle. Hopefully, future progress in seismic tomography will provide us with further evidence for mantle plumes.

Third, Oceanic Island Basalts (OIBs) are geochemically distinct and more diverse than Mid-Ocean Ridge Basalts (MORBs), as reviewed by Hofmann (1997). Moreover, isotopic signatures of OIBs indicate an involvement of ancient recycled oceanic crust, as first suggested by Hofmann and White (1982), whereas noble gases indicate that plumes may also carry primordial mantle material, possibly stored in the deep mantle.

Fourth, Burke and Torsvik (2004) provide another set of evidence supporting the deep origin of mantle plumes. Their plate tectonic reconstruction over the last 200 Ma shows that the paleo-position of 90% of the 25 LIPs considered, were located, at the time of eruption, above lower mantle regions characterized, today, by low S-wave velocities. These broad regions, situated beneath the South-Central Pacific and Africa, (see Romanowicz and Gung (2002) and references therein), are likely hotter and possibly compositionally denser than the surrounding mantle and may indeed represent a long-lived source zone of plumes.

4. Geochemistry of mantle plumes

Geochemists became interested in mantle plumes when they discovered that ocean island basalts (OIBs), thought to be derived from plumes, tend to have different chemical (Schilling, 1973) and isotopic (Hart et al., 1973) compositions from mid-ocean-ridge basalts (MORBs). These differences are consistent with the plume model if the deep mantle, the inferred source of plumes, is compositionally different from the upper mantle, source of MORBs. These systematic differences were borne out by many subsequent studies of samples from virtually all available ocean islands (Hofmann, 2007), as illustrated by Figure 3a for chemical compositions and Figure 3b for isotopic compositions. Both the trace elements and the isotopic compositions indicate that the MORB source has been depleted in mantle-incompatible trace elements, i.e. those that are scavenged

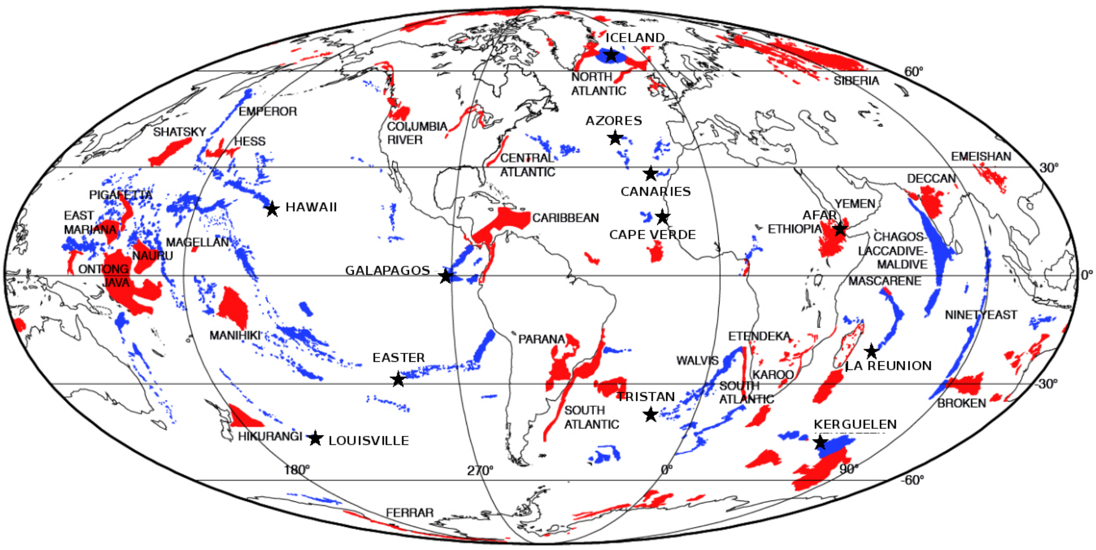

Figure 2. Distribution of hotspots (stars) and Phanerozoic LIPs. In red: LIPs (or portions thereof) generated by a transient 'plume head'. In blue: LIPs (or portions thereof) generated by a persistent 'plume tail'. Modified from Coffin and Eldholm (1994), with permission.

from the mantle by melts that ultimately form crust, leaving behind a depleted mantle. In contrast, plume-type basalts tend to be derived from less depleted mantle sources.

The above observations and conjectures led to the two-layer mantle model, in which normal melting producing MORBs and subduction-related magmas have extracted most of the incompatible elements from the upper mantle and sequestered them in the continental crust. By contrast, the lower mantle was thought to be largely undepleted or more 'primitive'. Thus, plumes, rising from the deep mantle, would sample the lower, relatively primitive reservoir, but they might entrain more depleted mantle rocks on their way up, thus producing a mixing array between primitive and depleted reservoirs (Jacobsen and Wasserburg, 1979). However, mass balance considerations for the incompatible element and isotope budget of the continental crust, the depleted upper mantle, and an undepleted lower-mantle, demanded a size of the depleted reservoir of at least 50% of the mantle, which is significantly greater than the roughly 30% mass fraction of the upper mantle. This three-reservoir model (often simply called the layered mantle model), was reinforced by the observation that many plume-derived basalts had much higher 3He/4He ratios than MORB (Farley and Neroda, 1998), where 3He is a remnant of primordial noble gases from the primitive Earth, whereas 4He is the product of subsequent decay of uranium and thorium.

In spite of its apparent geochemical persuasiveness, the layered mantle model suffered two essentially fatal blows. During the 1980s, evidence began to accumulate that showed isotopically 'enriched' (i.e. crustal-like) mantle sources in several ocean islands including Hawaii, Pitcairn, and Tristan da Cunha. This means that the endpoint of the mixing array of the type shown in Figure 3b cannot be an undepleted or 'primitive' mantle reservoir but must represent a source component that was crust-like in that it was actually enriched in incompatible elements. And although such a mixing array may pass through the isotopic locus of a primitive mantle reservoir, it clearly does not require the involvement of a primitive reservoir. This is because, when a primitive reservoir is differentiated into enriched and depleted components, any remixing of these differentiated components will produce a compositional array passing approximately through the starting point, the locus of primitive compositions. Therefore, Hofmann and White (1982) proposed that the enriched source components in mantle plumes actually come from subducted oceanic crust rather than a 'primitive reservoir'.

The second blow to the conventional layered mantle model came from seismic evidence showing tomographic images of high seismic velocities characterizing subduction slabs that penetrate the base of the upper mantle and can be traced into the lowermost mantle (e.g., van der Hilst et al., 1997). If such deep subduction occurred during major portions of Earth history, convective mixing will have destroyed the chemical separation between upper and lower mantle. The first attempt to integrate numerical convection modeling with chemical differentiation at crustal levels and recycling of ocean crust to generate enriched plume sources at the base of the mantle was made by Christensen and Hofmann (1994), and this has been followed up by more elaborate simulations.

More recently, a variety of new models have sprung up. Some invoke irregularly distributed, relatively primitive plume source regions which have been protected from convective stirring throughout Earth history and are thus able to preserve primitive geochemistry. Tolstikhin and Hofmann (2005) and Boyet and Carlson (2005) have proposed

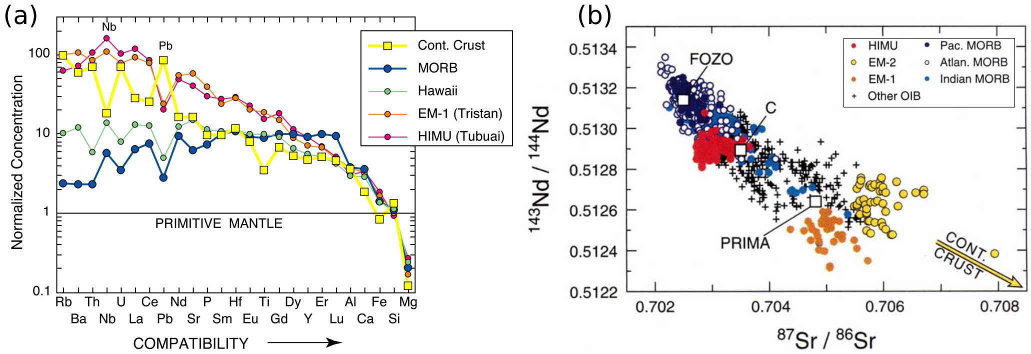

Figure 3. **(a)** Concentrations of selected trace and major elements, arranged in the order of ascending compatibility and normalized to primitive-mantle concentrations, for: average continental crust, average MORB, average Mauna Loa, Hawaii, and OIB type representing EM-1 (Enriched mantle 1) and HIMU (high-μ, where μ=238U/204Pb). The patterns for MORB and OIB differ by their enrichments, but show similar Nb and Pb anomalies, opposite to those of the continental crust. From Hofmann, 1997. **(b)** Nd and Sr isotopic compositions$^{(*)}$ of MORBs and OIBs. EM-1 (Enriched mantle 1), EM-2 (Enriched mantle 2) and HIMU. The squares indicate compositions of primitive mantle (PRIMA) and of mantle components FOZO and C (see Hofmann, 1997 and references therein). $^{(*)}$Some basic notions: 147Sm decays to 143Nd. Since Sm is more compatible than Nd, progressive melting enriches the residual rock in Sm, thus MORBs have high 143Nd/144Nd. 87Rb decays to 87Sr. Since Rb is less compatible than Sr, progressive melting depletes the residual rock in Rb, thus MORBs have low 87Sr/86Sr.

a new form of a two-layer mantle, one in which a relatively small, compositionally dense reservoir formed at the base of the mantle, in effect constituting the D layer. This irregular layer, on average about 200 km thick, has otherwise been interpreted as a 'slab graveyard'. The new geochemical model stipulates that these subducted slabs are very ancient, perhaps only a few tens of millions of years younger than the accretion of the Earth. This, in turn, requires stabilization by high intrinsic density, either because they are derived from a primordial iron-rich mafic crust, or because they were generated by downward segregation of dense partial melts in the lowermost mantle (Labrosse et al., 2007) in effect also creating an Fe-rich, dense bulk composition. Such a 'new two-layer' mantle model can account for the geochemical differences between plumes and mid-ocean ridge basalts, and it resolves at least two awkward problems with the old model: (1) it does not require that plumes have particularly high helium concentrations, which are never actually observed in plume-derived basalts; (2) it explains trace element characteristics of plume-derived basalts, which are in fact inconsistent with primitive sources. The new 2-layer model, by its nature, eliminates the need for any undifferentiated, primitive reservoir, but it creates a repository of primordial noble gases in the permanently sequestered D"-layer at the base of the mantle. From there, these noble gases can easily diffuse into the overlying actual (silicate) source reservoirs of mantle plumes.

It should be emphasized that, by its very nature, geochemical evidence is extremely unlikely to either prove or disprove the existence of mantle plumes or mantle layering, because the mixing and extraction processes that deliver plume-derived or non-plume-derived melts to the surface are not known a priori. Unfortunately, in the past, it has too often been argued that some ocean island must be derived from a deep-mantle plume because it contains high 3He/4He ratios, or near-chondritic 143Nd/144Nd ratios. Indeed, the inherent weakness of such arguments, which are clearly non-unique, has helped to discredit plume theory in some quarters. On the other hand, the plume model in combination with geochemical evidence can help to elucidate many fundamental aspects of earth evolution, including the time scale of early Earth evolution and of crustal recycling, as well as the intensity of convective stirring in the mantle.

Overall, the 'enriched' nature (i.e. enriched in highly incompatible elements relative to more compatible elements), which is evidenced by both the observed trace element and isotope abundances of the decay products of long-lived radioactive decay systems such as 87Rb-87Sr, 147Sm-143Nd, 176Lu-176Hf in plume derived rocks, is most easily explained by the subduction and recycling of enriched crustal rocks, rather than the involvement of any 'primitive' mantle material (Hofmann, 1997). Subduction of crustal materials, including ordinary ocean crust, enriched ocean crust, ocean island and seamounts, as well as some sediments and other continental material, is based on geological and geophysical observations. Thus, there is no shortage of suitable source materials in the mantle without the need for alternative enrichment processes, such as metasomatic infiltration, which might nevertheless play some role. The recycled enriched materials appear to be more prevalent in mantle plumes than in mid-ocean ridge basalts, because basaltic crust, once subducted, is somewhat denser than ordinary peridotitic mantle. Because of this, it may be segregated and stored at the base of the mantle for geologically longer periods of time, and ultimately contribute significant portions of plume source materials (Christensen and Hofmann, 1994).

5. Genesis and dynamics of plumes and superplumes

On time scales of millions of years, solid mantle rocks behave as highly viscous fluids with a viscosity of $\sim 10^{18}$-10^{22} Pa s (in comparison, the viscosity of a glacier is $\sim 10^{13}$ Pa s, and of a basaltic lava is 10-10^4 Pa s). Thus mantle dynamics is governed by a set of equations (conservation of mass, momentum and energy) for a viscous fluid where inertial effects can be neglected. The growth-rate of a Rayleigh-Taylor instability from a thermal boundary layer heated from below is controlled by the Rayleigh number:

$$Ra = \frac{\rho g \alpha \Delta T d^3}{\eta \kappa} \tag{1}$$

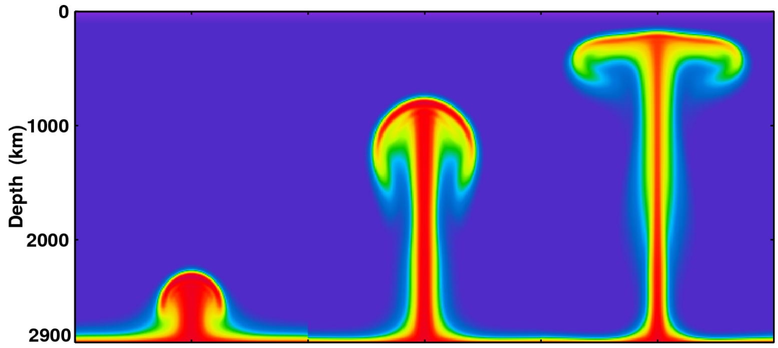

Figure 4. Numerical simulation of a thermal plume (Farnetani, 1997).

where ΔT is the temperature contrast across a layer of thickness d, density ρ, thermal expansion coefficient α, thermal diffusivity κ and viscosity η. Instabilities are enhanced by the thermal buoyancy and inhibited by viscosity and thermal diffusivity. Using reasonable values for the above physical parameters, growing instabilities of the boundary layer form diapirs. Due to the high mantle viscosity, such a diapir will separate from the TBL only when its volume becomes sufficiently large (e.g., Whitehead and Luther, 1975). The morphology of a thermal plume is controlled by the viscosity contrast between the hot fluid and the mantle above it: if the viscosity contrast is weak, plumes will have a 'spout' shape, with little difference between the radius of the leading diapir (the plume head) and the following conduit (the plume tail), whereas increasing viscosity contrast leads to a larger head and a narrower conduit. This 'mushroom' shape (Figure 4) is favored, because a hot plume is likely to be hundred times less viscous than the surrounding mantle, owing to the strong temperature dependence of viscosity.

Although fluid dynamics laboratory experiments (e.g., Whitehead and Luther, 1975; Griffiths and Campbell, 1990) and numerical simulations (e.g., Parmentier et al., 1975; Olson et al., 1993) on purely thermal plumes enabled us to gain a quantitative understanding of plume dynamics, in the 1990s it became progressively clear that the lowermost mantle is compositionally heterogeneous. D" is a region of preferential segregation and accumulation of denser subducted crust (Christensen and Hofmann, 1994), and larger-scale regions of the lower mantle may be chemically heterogeneous (Kellogg et al., 1999). This offered new and exciting avenues to explore the complex dynamics of thermo-chemical plumes, which can be defined as hot (positively buoyant) plumes that carry compositionally denser (negatively buoyant) material. Laboratory experiments by Davaille (1999) investigated a variety of regimes and found that instabilities may form dome-like structures with an oscillatory behavior (i.e., they rise and sink in response to a subtle balance between thermal and chemical buoyancies). Her experiments, together with numerical simulations of thermo-chemical convection (e.g., Tackley, 1998) provide a fluid dynamically consistent framework to interpret observations that are otherwise unexplained by purely thermal convection. For example, the commonly referred 'superplumes', situated beneath the South-Central Pacific and Africa are broad (thousands of kilometers large) lower mantle zones of low seismic velocity most likely associated with active upwelling. Although there is a debate on the thermal and-or compositional origin of 'superplumes', several lines of evidence do support their distinct composition (Masters et al., 2000). Another issue that can be readily explained by thermo chemical plumes is the discrepancy between the petrologically constrained excess temperature of plumes (100-250°C) and the estimated temperature difference across the D" region (∼1000°C). Numerical simulations (Farnetani, 1997) show that the deepest, hottest part of the thermal boundary layer does not upwell in mantle plumes if the compositional density contrast exceeds 2%. A denser zone in the boundary layer also tends to anchor the base of plume conduit, so that the flow pattern remains stable over timescales longer than the plume rise time (Jellinek and Manga, 2002). Finally, thermo-chemical plumes present a variety of shapes and surface manifestations (Farnetani and Samuel, 2005; Kumagai et al., 2007) that expand the classical plume head-tail model predictions (Figure 5).

6. Plume melting and plume strength

At spreading ridges mantle rocks rise to shallow depths in response to plate spreading, and melt by adiabatic decompression (McKenzie and Bickle, 1988). This is not the case for most plumes impinging at the base of a preexisting and unrifted continental or oceanic lithosphere. In order to melt at relatively high pressure (4-5 GPa) plumes must be either hotter than normal mantle, or compositionally more fertile than peridotitic mantle (Condie, 2001, ch.4). Plumes are probably both: their excess temperature is estimated between 100 and 250°C, (Putirka, 2005) and they can carry 20%, or more, of recycled crust in the form of eclogite (Sobolev et al., 2007). Eclogite is more fertile than peridotite because it has a lower solidus and a greater melt productivity. According to Sobolev et al., (2005), reaction between eclogite-derived liquids and solid peridotite forms a pyroxenite, whose subsequent melting may explain some compositional characteristics of Hawaiian (and many other plume-derived) lavas. Another efficient way of lowering the solidus temperature is to have fluids (H_2O, CO_2) in the upwelling plume rocks, however the evidence for fluid-rich hotspot lavas is scant. Although it is well known that melting is the most efficient way to create chemical differentiation in the mantle and that it is the unavoidable process to generate surface lavas from mantle rocks, many aspects of partial melting remain elusive, thereby limiting our quantitative understanding of the link between petrological and geochemical observations of surface lavas and the underlying mantle plume composition and dynamics.

The strength of a mantle plume can be calculated on the production rate of volcanic rocks. The method utilizes the volcano's volume and growth time, and the estimated

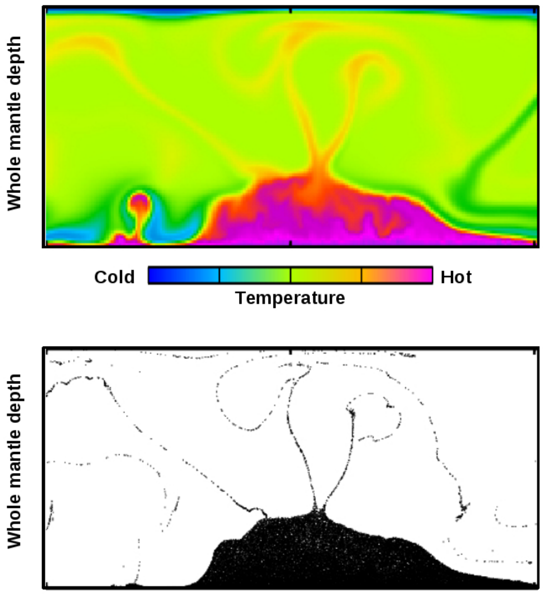

Figure 5. Numerical simulation of a thermo-chemical plume. Top: Temperature field. Bottom: Chemically denser material, represented with tracers. Animations are available on request. (Farnetani, 2003, unpublished simulations for Palais de la Découverte, Paris).

melt fraction. For example, Mauna Loa has a volume of $\sim$70000 km^3 and it grew in $\sim$1 Ma, suggesting that the Hawaiian plume melt production rate M is about 0.1-0.2 km^3/yr. Since only a fraction of the upwelling plume melts, the total plume volume flux $Q_v >> M$. Another method, based on the rate of swell formation (Sleep, 1990), enables to calculate the plume buoyancy flux $B = Q_v \Delta\rho$ where $\Delta\rho = \rho\alpha(T_{PLUME} - T_{MANTLE})$. For Hawaii, the most vigorous hotspot, $B \sim$8000 kg/s, for Iceland $B \sim$1400 kg/s, for Galapagos $B \sim$1000 kg/s. Finally, the plume heat flow is:

$$Q_h = \rho_{MANTLE} C_p Q_v (T_{PLUME} - T_{MANTLE}) \qquad (2)$$

where the specific heat is $C_p \sim$1200 J/kg K.

For Hawaii the heat flow is $Q_h \sim 0.36 \times 10^{12}$ W and it represents $\sim$16% of the global hotspot heat flow of 2.3×10^{12} W, which, in turn is a small fraction of the Earth's total heat flow of 44×10^{12} W. (For further reading see Schubert et al., 2001, chapter 11).

7. Continental Flood Basalts and continental break-up

The eruption of a large igneous province (LIP) represents a major geologic event. Over the Earth's history, episodic LIP magmatism contributed to continental growth through the emplacement of CFB and through the accretion/obduction of fragments of oceanic plateaus onto continental crust (Ben-Avraham et al., 1981). Accretion may be due to the difficulty of subducting the 20-40 km thick crust of an oceanic plateau. Three notable examples of accreted LIP fragments are found in the Solomon Island arc (from the Ontong-Java plateau), in Central America (from the Caribbean plateau), and the Wrangellia terrane outcropping in British Columbia and in SE Alaska (Figure 6). This allochthonous terrane, locally attaining a thickness of 6 km, is likely to be a fragment of a 230 Ma old oceanic plateau (Ben-Avraham et al., 1981; Richards et al., 1991; Greene et al., 2008).

Distinctive characteristics of LIP magmatism are the short duration (1-2 Ma) and the huge volumes ($1\text{-}10 \times 10^6$ km^3) of eruption. Because of the high eruption rates (>1-2 km^3/yr; for comparison Kilauea grows at 0.1 km^3/yr) and the enormous surface extent ($1\text{-}2 \times 10^6$ km^2; for comparison France's surface is $\sim 0.5 \times 10^6$ km^2), flood basalt eruptions are extraordinary volcanic events. CFBs consist of subhorizontal flows of mafic (Fe- and Mg-rich) rocks, mainly tholeiitic basalts. Individual flows can extend for hundreds of kilometers, be tens to hundreds of meters thick, and have volumes of more than 10^3 km^3. CFBs are more accessible than submerged oceanic plateaus; however they can be extensively eroded, fragmented and dispersed on different continents by plate tectonic processes. Clearly, uncertainties regarding the original volume and surface extent increase with their age, for example, for pre-Cambrian CFB only the giant swarm of mafic dikes, feeding surface volcanism, may be left (Ernst and Buchan, 2001).

The Phanerozoic CFBs are, in chronological order: The 258 Ma old Emeishan Traps (SW China) which presently cover only 0.3×10^6 km^2, probably a tenth of the estimated original surface. The 250 Ma old Siberian Traps, which might have had a surface extent of $3\text{-}4\times10^6$ km^2 and extrusive volumes of 10^6 km^3. The 200 Ma old lavas and dike swarms of the Central Atlantic Magmatic Province outcrop in once-contiguous parts of North America (e.g., the Palisades sill, NY), West Africa, and Brazil. This flood volcanism, with an estimated total volume of $3\text{-}5\times10^6$ km^3, preceded the opening of the central Atlantic Ocean. The 184-182 Ma old Karoo (Southern Africa) - Ferrar (Antarctica) traps preceded the breakup of Gondwana by 10-15 Ma. Similarly, the $\sim$135-130 Ma old magmatism of the Parana (Brazil) - Etendeka (Namibia) province led to the opening of the South Atlantic Ocean. The Deccan Traps (India) erupted 66-65 Ma ago over an area of $\sim1.5\times10^6$ km^2; the subsequent rifting split apart the Seychelles from India. The North Atlantic Tertiary Igneous Province covers a surface greater than 10^6 km^2 and has an estimated volume of 6×10^6 km^3. The earliest volcanism occurred as flood basalts in Baffin Island, West- and East-Greenland (62 Ma), and later (56-54 Ma) extended to the continental margins of Greenland, the British Isles and Norway. These rifted edges of continents constitute the so-called volcanic passive margins of up to 8 km thick seaward dipping basalt layers. The complete opening of the North Atlantic and the appearance of truly oceanic crust occurred about 53 Ma ago. Finally, the Ethiopian-Yemen traps, ~30 Ma old, have been associated with the rifting in the Red Sea and the Gulf of Aden (Courtillot et al. 1999).

Issues concerning the timing of rifting have been hotly debated: One side claims that rifting preceded, and enhanced, flood volcanism; the other side argues that rifting is not a prerequisite to flood volcanism, since it occurs only during and after volcanism. More geological observations are needed, as well as numerical simulations and/or laboratory experiments investigating how anomalously warm mantle can erode the lithosphere and create (or reactivate) weak zones susceptible to rifting.

8. LIP magmatism and environmental effects

The remarkable temporal correlation between LIP magmatism and mass extinctions of terrestrial and marine organisms suggests a cause-and-effect connection (Courtillot and Renne, 2003). The most notable example is the Siberian Traps at the Permian-Triassic boundary (Renne et al., 1995; Svensen et al., 2009). The emplacement of the Deccan Traps spans the Cretaceous-Tertiary boundary and certainly had an environmental effect (Self et al., 2006); however, the extinction event coincided with a large asteroid impact at Chicxulub, Mexico (Schulte et al., 2010). The peak of volcanism in the North Atlantic Volcanic Province corresponds to the Paleocene-Eocene Thermal Maximum event (PETM) 55 Ma ago. Svensen et al. (2004) propose that voluminous magmatic intrusions in carbon-rich sediments induced massive release of the methane buried in marine sediments, thus suggesting that volcanic and metamorphic processes associated with the opening of the north Atlantic may explain global climate events. Self et al., (2006); Ganino and Arndt (2009); Svensen et al. (2009) attempt to quantify the release of volatiles (CO_2, CH_4, SO_2) during flood volcanism and to estimate the consequences on the environment. In particular, the release of volcanic ashes and sulfuric acid aerosols can lead to darkening and cooling, SO_2 and HCl cause acid rains, whereas greenhouse gases like CO_2 and CH_4 cause global warming. Carbon has two origins: magmatic and sediment-derived. Interestingly, Ganino and Arndt (2009) show that the mass of sediment-derived CO_2 can be 4-8 times larger than the mass of magmatic CO_2, if contact metamorphism reactions occur in sedimentary rocks such as dolomite, evaporite, coal, or organic-rich shale. In other words, the environmental effect of flood basalt magmatism depends, among other factors, on the rock type in contact with magmatic sills and lava flows (Ganino and Arndt 2009; Svensen et al., 2009).

Sub-aqueous eruptions can also have an environmental effect, because they can cause variations in water chemistry and oceanic circulation. The emplacement of the (Alaska-sized) Ontong-Java plateau 120 Ma ago and of the Caribbean plateau correlate temporally with some of the major Oceanic Anoxic Events (OAEs) which occurred ~183, 120, 111 and 93 Ma ago. OAEs are associated with an abrupt rise in temperature, induced by rapid influx of CO_2. The consequent increase of organic productivity causes an increase of oxygen demand in the water column, eventually leading to oxygen depletion in the oceans (for a thorough review see Jenkyns (2010) and references therein). Furthermore, the formation of black shale (marine carbon-rich sediments) may have been favored by massive hydrothermal release of trace metals, poisonous to marine life.

9. Plume-lithosphere interaction

A quantitative understanding of plume-lithosphere interaction enables us to relate geophysical observations (e.g., the spatial-temporal evolution of surface topography, variations in lithospheric thickness, etc.) to the physical parameters and dynamics of the plume. Central issues concerning plume-lithosphere interactions include, but are not limited to:

First, estimating the dynamic topography[1] induced by the arrival of a mantle plume head. Pioneering laboratory experiments by Olson and Nam (1986) found that a thermal diapir upwelling and spreading beneath the lithosphere induces a rapid surface uplift followed by a slower subsidence. According to Farnetani and Richards (1994), uplift should precede flood volcanism by 1-2 Ma, a prediction that is sometimes validated by geological observations (e.g., for the Emeishan Traps), but not always (e.g., for the Siberian Traps). A present-day example of dynamic topography is provided by Lithgow-Bertelloni and Silver (1998), who suggest that the anomalous elevation (~1 km) of South-Central Africa has a dynamic origin, being induced by mantle upwelling from the African superplume, a broad lower mantle zone of low seismic velocity anomalies.

Second, understanding the origin of the topographic hotspot swell and the thermo-mechanical response of the lithosphere drifting over the plume. Two main mechanisms have been proposed for the Hawaiian swell: 'thermal rejuvenation' invokes heating and thinning of the lithosphere above the hotspot, whereas 'dynamic support' invokes stresses applied to the base of the lithosphere by the buoyant plume. Seismic studies by Li et al. (2004) around Hawaii found a thinning of the lithosphere from the expected 100 km (beneath the Big Island) to only 50-60 km (beneath Oahu and Kauai, 400 km downstream from the Big Island), thus suggesting a hybrid scenario, whereby 'dynamic support' prevails above the plume conduit and 'thermal rejuvenation' prevails downstream. Far from the conduit, thermal thinning of the lithosphere may be enhanced by small-scale convective instabilities, which take the form of rolls aligned with the plate motion, as observed in numerical simulations with strongly temperature dependent viscosity (Moore et al., 1998).

Third, mechanical and chemical interaction between plume magmas and the lithosphere can explain a number of geophysical and geochemical observations. For example, the occurrence of regularly spaced volcanoes in hotspot chains is probably caused by the interaction of magma transport with lithospheric flexure (Hieronymus and Bercovici, 1999). Under the load of the volcanoes the lithosphere deforms and the resulting flexural stresses have a key role in determining the locus where magma preferentially extrudes. There are also clear examples of geochemical interaction between the continental lithosphere and plume magmas 'en route' to the surface. In particular, CFB lavas erupted at the onset of flood volcanism often have anomalous compositions, reflecting contamination by the lithosphere and crust. It is

[1] Definition of dynamic topography, extracted from Lithgow-Bertelloni and Silver, (1998): *Dynamic topography is a deformation of the surface of the Earth, supported by the vertical stresses at the base of the lithosphere, that are generated by flow in the mantle below. This is in contrast to the more familiar mechanism of isostatically supported topography which is in equilibrium at the Earth's surface and would exist even in a static mantle.*

Figure 6. Photograph of 1000 m of continuous subaerial flood basalt stratigraphy in the Wrangell Mountains, Alaska. The yellow line marks the contact between Nikolai basalts and the overlying Chitistone Limestone. From Greene et al. (2008), with permission.

thus important to quantify the contribution of the continental lithosphere before inferring the plume's geochemical fingerprint.

Fourth, plume material may preferentially flow –and partially melt– in pre-existing zones of thinned lithosphere (Sleep, 1996). In other words, the thickness of the lithosphere and the slope of its base may control the spatial distribution of plume magmatism. According to Ebinger and Sleep (1998), the Cenozoic magmatism in East Africa (including the Afar, the Cameroon volcanic line to the West, and the Comoros Islands to the South) can be explained by a single plume impinging beneath the Ethiopian plateau and its subsequent lateral flow, 'channeled' in pre-existing rift-zones characterized by a relatively shallow base of the lithosphere.

10. Plume-ridge interaction

At least 21 hotspots are situated near a spreading ridge, the clearest examples being Iceland, a ridge-centered hotspot, Galapagos, and the Azores. At Iceland the Mid-Atlantic Ridge rises ∼4 km above normal and remains anomalously shallow for ∼1000 km to the North and the South. Below Iceland, the oceanic crust is 40 km thick, whereas more normal crustal thicknesses of 8-10 km are attained only ∼500 km far from the hotspot. Furthermore, compositional anomalies in trace elements and isotope ratios observed along the ridge clearly indicate plume-ridge interaction (Schilling et al., 1985). The Galapagos hotspot is situated ∼200 km off-axis form the Galapagos Spreading Center, and, also in this case, geophysical and geochemical anomalies along the ridge suggest plume-ridge interaction.

These observations raise several questions concerning: (a) how plume material flows toward the ridge, opposite to the plate motion, (b) how plume material, once 'trapped' beneath the ridge, flows parallel to the ridge axis, and (c) how the plume-ridge flow is affected by physical parameters such as the plume volume flux, the plume-ridge distance, and plate velocity. As reviewed by Ito et al., (2003), such questions have been addressed with laboratory experiments (e.g., Kincaid et al., 1995) and numerical simulations (e.g., Ribe, 1996) showing that two factors are particularly important. First, the oceanic lithosphere thickens by conductive cooling away from the ridge and its sloping base favors the flow of plume material toward the ridge (Sleep, 1996). This flow, however, becomes less efficient if dehydration strengthening increases the viscosity of the plume residuum (Ito et al., 2003). Second, the radial self-spreading of the plume head beneath the lithosphere occurs also in the direction opposite to the plate motion, so that plume material can reach the spreading ridge. For a fluid dynamical analysis of plume-ridge interaction see the comprehensive paper by Ribe et al., (2007), and references therein.

11. Future directions

This chapter on mantle plumes has taken us from the deep Earth's mantle where plumes originate, to sublithospheric depths where partial melting occurs, and up to the surface where LIPs have been emplaced in the geologic past, and hotspot lavas are being erupted today. We also explored the environmental effect of massive LIP volcanism on the oceans and the atmosphere. This range of topics shows that a full understanding of mantle plumes requires a multi-disciplinary approach. Important progress will be achieved by linking geophysical, geochemical, and geological observations, to models solving for the fluid dynamics of mantle plumes, melt migration, and physical volcanology, just to mention a few. In our view, fundamental questions for the future include the following:

1. Quantifying the effect of heterogeneities on plume dynamics and fertility. The term heterogeneity is used here in a wide sense, from small scale (1-10 km) lenses of recycled eclogite to the large-scale (100-1000 km), denser superplumes that may generate short-lived instabilities rising from their top. Future studies should model heterogeneous plumes in a fully convecting mantle, with physical properties provided by mineral physics experiments at appropriate temperature and pressure conditions. Lithologic variability affects the fertility and the geochemical characteristics of lavas (Sobolev et al., 2005), but, at present, numerical simulations are unable to model the complex petrological and geochemical evolution of the solid matrix and melts.

2. Quantifying the role of the lithosphere. This is needed to understand the extent to which lithospheric stresses and variations in lithospheric thickness and composition may affect partial melting and the ascent of magma. Furthermore, the hypothesis that thermo-elastic cracking of the oceanic lithosphere (Sandwell and Fialko, 2004) can trigger magmatism should be explored quantitatively, to assess the volumes and the composition of such magmas, and to investigate the possibility of generating non age-progressive volcanic lineaments.

3. The cause-and-effect relation between the arrival of a mantle plume and continental break-up is often invoked, but we lack a quantitative understanding of the processes involved. In other words, although there is evidence supporting the role of plumes in the creation and modification of plate boundaries, it is unclear how a large-scale system, with continents and pre-existing spreading ridges responds to the arrival of a large mantle plume. Moreover, studies of plume-lithosphere interaction should explore the possibility of dynamic instabilities, such as lithospheric delamination, fingering of low viscosity fluid, and small scale convection (see Ito and vanKeken (2007) and references therein), and their ability to generate partial melts.

4. Future studies should address in a constructive and quantitative way the current debate on the 'existence' of plumes, (see www.mantleplumes.org) or better, on the 'coexistence' of plumes with other processes that can produce surface volcanism. Progress will be achieved through improved seismic imaging of plumes and their depth of origin, and through a quantitative understanding of lithospheric processes, as explained above. We also note that much of the ongoing debate questioning the existence of plumes is based on a narrow view of what a plume should look like: the head-tail model is valid in a purely thermal, homogeneous mantle, but a variety of plume shapes are found if the Earth's lower mantle is compositionally heterogeneous (Farnetani and Samuel, 2005; Kumagai et al., 2007). Furthermore, when considering plumes in a convecting mantle, rather than in an unperturbed fluid, it is obvious that conduits cannot be fixed (Steinberger et al., 2004), an observation that does not preclude the existence of deep plumes.

5. Although we have restricted our treatment to Mesozoic-Cenozoic plumes (<250 Ma), it is likely that plumes had an important role also at earlier times, although geological evidence becomes increasingly scant with geological age (Ernst and Buchan, 2001). The early Earth, with its vigorous and highly time-dependent convection, probably had an intense plume magmatism. Hopefully, studies of volcanism on Mars and Venus can complement and improve our knowledge of early terrestrial magmatism (see Condie, 2001 ch. 3). Both Mars and Venus lack plate tectonics, so that old geologic features can be better preserved. On Mars, the Olympus Mons is an enormous shield volcano (volume 2×10^6 km^3), and the Tharsis rise, which covers 20% of the Martian surface, has been volcanically active for the last 2 Ga, suggesting the presence of huge and long-lived plumes in the Martian mantle (Harderer and Christensen, 1996).

Acknowledgments. We thank David Bercovici for a thorough review and Harsh K. Gupta, Editor, for constructive comments.

12. References

Ben-Avraham, Z., Nur, A., Jones, D., and Cox, A., 1981. Continental accretion: From oceanicc plateaus to allochthonous terranes, Science, 213: 47–54.

Boyet, M., and Carlson, R.W., 2005, Nd-142 evidence for early (> 4.53 Ga) global differentiation of the silicate Earth, Science, 309: 576–581.

Burke, K., and Torsvik, T.H., 2004. Derivation of large igneous provinces of the past 200 million years from long-term heterogeneities in the deep mantle, Earth Planet Sci. Lett., 227: 531-538.

Christensen, U.R., and Hofmann, A.W., 1994. Segregation of subducted oceanic crust in the convecting mantle, J. Geophys. Res., 99: 19867–19884.

Clouard, V., and Gerbault, M., 2008. Break-up spots: Could the Pacific open as a consequence of plate kinematics? Earth Planet. Sci. Lett., 265: 195–208.

Coffin, M.F., and Eldholm, O., 1994. Large igneous provinces: crustal structure, dimensions and external consequences, Rev. Geophys., 32: 1–36.

Condie, K. C., 2001. Mantle plumes and their record in earth history, Cambridge University Press.

Courtillot, V. E., Jaupart, C., Manighetti, I., Tapponier, P., and Besse, J., 1999. On causal links between flood basalts and continental breakup, Earth Planet. Sci. Lett., 166: 177–195.

Courtillot, V. E., Davaille, A., Besse, J. ,and Stock, J., 2003. Three distinct types of hot spots in the Earth's mantle, Earth Planet. Sci. Lett., 205: 295–308.

Courtillot, V.E., and Renne, P., 2003. On the ages of flood basalt events, Comptes Rendus Geoscience, 335: 113–140.

Davaille, A., 1999. Simultaneous generation of hot spots and superswells by convection in a heterogeneous planetary mantle, Nature, 402: 756–760.

Ebinger C.J., and Sleep, N.H., 1998. Cenozoic magmatism throughout east Africa resulting from impact of a single plume, Nature, 395: 788–791.

Ernst, R.E., and Buchan, K.L., 2001. The use of mafic dike swarms in identifying and locating mantle plumes. In Ernst, R.E., and Buchan, K.L (eds.), Mantle plumes: their identification through time, Geological Society of America, 352.

Farley, K.A., and Neroda, E., 1998. Noble gases in the Earth's mantle, Annu. Rev. Earth Planet. Sci., 26: 189–218.

Farnetani, C.G., and Richards, M. A., 1994. Numerical investigation of the mantle plume initiation model for flood basalt events, J. Geophys. Res., 99: 13813-13883.

Farnetani, C.G., 1997. Excess temperature of mantle plumes: The role of chemical stratification across D", Geophys. Res. Lett., 24: 1583–1586.

Farnetani, C.G., and Samuel, H., 2005. Beyond the thermal plume paradigm, Geophys. Res. Lett., 32, doi:10.1029/2005GL022360.

Foulger, G. R., and Natland, J. H., 2003. Is Hotspot volcanism a consequence of plate tectonics? Science, 300: 921–922.

Ganino, C., and Arndt, N.T., 2009. Climate changes caused by degassing of sediments during the emplacement of large igneous provinces, Geology, 37: 323–326 doi: 10.1130/G25325A.1.

Greene, A. R., Scoates, J. S., and Weis, D., 2008. Wrangellia flood basalts in Alaska: A record of plume-lithosphere interaction in a Late Triassic accreted oceanic plateau, Geochem. Geophys. Geosyst., 9, Q12004, doi:10.1029/2008GC002092.

Griffiths, R.W., and Campbell, I.H., 1990. Stirring and structure in mantle starting plumes, Earth Planet. Sci. Lett., 99: 66–78.

Harderer, H., and Christensen, U.R., 1996. A one-plume model of Martian mantle convection, Nature, 380:507–509.

Hart S.R., Schilling J.G., and Powell J. L., 1973. Basalts from Iceland and along the Reykjanes Ridge: Sr isotope geochemistry, Nature 246: 104–107.

Hieronymus,C. F., and Bercovici, D., 1999. Discrete alternating hotspot islands formed by interaction of magma transport and lithospheric flexure, Nature, 397: 604–607.

Hieronymus, C.F., and Bercovici, D., 2000. Non-hotspot formation of volcanic chains: control of tectonic and flexural stresses on magma transport, Earth Planet. Sci. Lett., 181: 539-554.

Hofmann, A.W., and White, W.M., 1982. Mantle plumes from ancient oceanic crust, Earth Planet. Sci. Lett., 57: 421–436.

Hofmann, A.W., 1997. Mantle geochemistry: the message from oceanic volcanism, Nature, 385: 219–229.

Hofmann, A.W., 2003. Sampling mantle heterogeneity through oceanic basalts: Isotopes and trace elements. In R.W. Carlson, Holland, H.D., and Turekian, K.K., (Eds.). Treatise on Geochemistry:The Mantle and Core, Elsevier, 61-101.

Ito, G., Lin, J., and Graham, D., 2003. Observational and theoretical studies of the dynamics of mantle plume-mid-ocean ridge interaction, Rev. Geophysics., 41, doi:10.1029/2002RG000117.

Ito, G., and van Keken, P.E., 2007. Hotspots and melting anomalies, In Bercovici, D. (Ed.) Treatise on Geophysics: Mantle Dynamics, Vol. 7, Elsevier.

Jacobsen, S.B., and Wasserburg, G.J. 1979. The mean age of mantle and crustal reservoirs, J. Geophys. Res. 84: 7411–7427.

Jellinek, A.M., and Manga, M., 2004. Links between long-lived hot spots, mantle plumes, D" and plate tectonics, Rev. Geophys., 42, doi:10.1029/2003RG000144.

Jellinek, A.M., and Manga, M., 2002. The influence of a chemical boundary layer on the fixity, spacing and lifetime of mantle plumes, Nature, 418: 760–763.

Jenkyns, H. C., 2010. Geochemistry of oceanic anoxic events, Geochem. Geophys. Geosyst., 11, Q03004, doi:10.1029/2009GC002788.

Kellogg, L.H.; Hager, B.H., and van der Hilst, R. D., 1999. Compositional stratification in the deep mantle, Science, 283: 1881–1884.

Kincaid, C., Ito, G., and Gable, C., 1995. Laboratory investigation of the interaction of off-axis mantle plumes and spreading centres, Nature, 367: 758–761.

Kumagai, I., Davaille, A., and Kurita, K., 2007. On the fate of thermally buoyant mantle plumes at density interfaces, Earth Planet. Sci. Lett. 254: 180193.

Labrosse, S., Hernlund, J.W., and Coltice, N., 2007. A crystallizing dense magma ocean at the base of the Earth's mantle, Nature, 450: 866–869.

Lithgow-Bertelloni, C., and Silver, P.G., 1998. Dynamic topography, plate driving forces and the African superswell, Nature, 395: 269–272.

Li X., Kind, R., Yuan, X., Wolbern, I., and Hanka, W., 2004. Rejuvenation of the lithosphere by the Hawaiian plume, Nature, 427: 827–829.

Loper, D.E., and Stacey, F.D. 1983. The dynamical and thermal structure of deep mantle plumes, Phys. Earth Planet. Int., 33: 305–317.

Masters, G., Laske, G., Bolton, H. and Dziewonski, A.M., 2000. The relative behavior of shear velocity, bulk sound speed and compressional velocity in the mantle: implication for chemical and thermal structure, in Earths deep interior. In S.I.Karato, A. Forte, R. C. Liebermann, G. Masters, and L. Stixrude (Eds.), Mineral physics and tomography from the atomic to the global scale, 117: 63–87, American Geophysical Union, Washington, DC.

McKenzie, D., and Bickle, M.J., 1988. The volume and composition of melt generated by extension of the lithosphere, J. Petrology, 29: 625–679.

Montelli, R., Nolet, G., Masters, G., Engdahl, E. R., and Hung, S.-H., 2004. Finite-frequency tomography reveals a variety of plumes in the mantle, Science, 303: 338–343.

Moore, W.B., Schubert, G., and Tackley, P., 1998. Three-dimensional simulations of plume-lithosphere interaction at the Hawaiian swell Science, 279: 1008–1011.

Morgan, W.J., 1971. Convection plumes in the lower mantle, Nature, 230: 42–43.

Olson, P., and Nam, I-S., 1986. Formation of seafloor swells by mantle plumes J. Geophys. Res., 91: 7181–7191.

Olson, P., Schubert, G., and Anderson, C., 1993. Structure of axisymmetric mantle plumes, J. Geophys. Res., 98: 6829–6844.

Parmenteir, E.M., Turcotte, D.L., and Torrance, K.E., 1975. Numerical experiments on the structure of mantle plumes, J. Geophys. Res., 80: 4417–4424.

Putirka, K.D., 2005. Mantle potential temperatures at Hawaii, Iceland, and the mid-ocean ridge system, as inferred from olivine phenocrysts: Evidence for thermally driven mantle plumes, Geoch. Geoph. Geosyst. 6, doi:10.1029/2005GC000915.

Renne P.R., Zhang, Z.C., Richards, M.A., Black, M.T., Basu, A.R., 1995. Synchrony and causal relations between Permian-Triassic boundary crises and Siberian flood volcanism, Science, 269: 1413–1416.

Ribe, N.M., 1996. The dynamics of plume-ridge interaction 2. Off-ridge plumes, J. Geophys. Res., 101: 16195–16204.

Ribe, N.M., Davaille, A., and Christensen, U.R., 2007. Fluid dynamics of mantle plumes In: Ritter, J.R.R., Christensen, U.R. (eds.), Mantle plumes - a multidisciplinary approach, 1-48, Springer, Berlin.

Richards, M. A., Duncan, R. A., and Courtillot, V. E., 1989. Flood basalts and hotspot tracks: Plumes heads and tails, Science, 246: 103–107.

Richards, M. A., Jones, D.L., Duncan, R. A., and DePaolo, D.J., 1991. A mantle plume initiation model for the Wrangellia flood basalt and other oceanic plateaus, Science, 254: 263–267.

Romanowicz, B., and Gung, Y.C., 2002. Mega-upwellings from the coremantle boundary to the lithosphere: implications for heat flux, Science, 296: 513–516.

Sandwell, D.T., and Fialko, Y., 2004, Warping and cracking of the Pacific plate by thermal contraction, J. Geophys. Res., 109, B10411, doi:10.1029/2004JB003091.

Schubert, G., Turcotte, D.L., and Olson, P., 2001. Mantle convection in the earth and planets, Cambridge University Press.

Schulte, P., et al., 2010. The Chicxulub asteroid impact and mass extinction at the Cretaceous-Paleogene boundary, Science, 327: 1214–1218.

Self, S., Widdowson, M., Thordarson, T., and Jay, A.E., 2006. Volatile fluxes during flood basalt eruptions and potential effects on the global environment: A Deccan perspective, Earth Planet. Sci. Lett., 248: 518–532.

Schilling, J.G., 1973. Iceland mantle plume: geochemical evidence along Reykjanes Ridge, Nature, 242: 565–571.

Schilling, J.G., Thompson, G., Kingsley, R., and Humphris, S., 1985. Hotspot-migrating ridge interaction in the South Atlantic, Nature, 313: 187–191.

Sleep, N.H., 1996. Lateral flow of hot plume material ponded at sublithospheric depths, J. Geophys. Res., 101: 28065–28083.

Sleep, N.H., 1990. Hotspot and mantle plumes: Some phenomenology, J. Geophys. Res. 95:6715–6736.

Smith, W.H.F., and Sandwell, D.T., 1997. Global sea floor topography from satellite altimetry and ship depth soundings, Science, 277: 1956–1962

Sobolev, A.V., Hofmann, A.W., Sobolev, S.V., and Nikogosian, I.K., 2005. An olivine-free mantle source of Hawaiian shield basalts, Nature, 434: 590–597.

Sobolev, A.V., Hofmann, A.W., et al., 2007. The amount of recycled crust in sources of mantle-derived melts, Science, 316: 412–417.

Steinberger B., Sutherland R., and O'Connell R.J., 2004. Prediction of Emperor-Hawaii seamount locations from a revised model of global plate motion and mantle flow, Nature, 430: 167–173.

Svensen H., Planke, S. et al., 2009. Siberian gas venting and the end-Permian environmental crisis, Earth Planet. Sci. Lett., 277: 490–500.

Svensen, H., Planke, S., Malthe-Sorenssen, A., Jamtveit, B., Myklebust, R., Eidem, T.R., and Rey, S.S., 2004. Release of methane from a volcanic basin as a mechanism for initial Eocene global warming. Nature, 429:542545.

Tackley, P.J., 1998. Three-dimensional simulations of mantle convection with a thermo-chemical basal boundary layer: D"? In Gurnis M. (ed.), The Core-mantle boundary region, 28, Geophys. Monogr. Ser. AGU, 231–253.

Tarduno J., Bunge H-P., Sleep N., and Hansen U., 2009. The bent Hawaiian-Emperor hotspot track: inheriting the mantle wind, Science, 324: 50–53.

Tolstikhin, I. and Hofmann, A.W., 2005. Early crust on top of the Earths core, Phys. Earth Planet. Inter., 148: 109–130.

van der Hilst, R., Widiyantoro, S., and Engdahl, E., 1997. Evidence for deep mantle circulation from global tomography, Nature, 386: 578-584.

Wessel, P., 1993. Observational constraints on models of the Hawaiian hot spot swell. J Geophys Res 98:16,09516,104.

Whitehead, J.A., and Luther, D.S., 1975. Dynamics of laboratory diapir and plume models, J. Geophys. Res., 80: 705–717.

Wilson, J. T., 1963. A possible origin of the Hawaiian islands, Canadian Journal of Physics, 41: 863–870.

Wolfe, C.J., Solomon, S.C., Laske, G., Collins, J.A., Detrick, R.S., Orcutt, J.A., Bercovici D., and Hauri, E.H., 2009. Mantle shear-save velocity structure beneath the Hawaiian hot spot, Science, 326: 1388–1390.