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Do mantle plumes exist?

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The physics of low Reynolds number plumes is well understood, which allows a number of testable predictions to be made about mantle plumes. They are predicted to originate from the core-mantle boundary and consist of a large head followed by a narrower tail. Millions of cubic kilometers of magma can be produced from a plume head. Rifting over a previously-emplaced plume head should produce a narrow zone of thickened oceanic crust along the rift margin. Initial eruption from a plume head should be preceded by ~1,000 m of domal uplift. Picrites are expected to dominate the first eruptive products of a new plume and should be concentrated near the centre of the volcanic province. All of these predictions are confirmed by observations, and so provide strong support for the validity of the mantle plume hypothesis.

Introduction

Tuzo Wilson (1963) was the first to suggest that aseismic ridges, like the Hawaiian-Emperor chain and the Chagos-Laccadive ridge (Figure 1), can be explained by plates moving above a stationary "hotspot" in the mantle. Morgan (1971) realized that Wilson's concept of mantle hotspots had no physical basis and suggested that Wilson's hotpots are nearly-stationary narrow plumes of hot mantle, which rise from the core-mantle boundary. Morgan also noticed that flood basalts are associated with arrival of new plumes and suggested that they might be the product of melting the head of a starting plume (Figure 2). This concept was expanded by Richards et al. (1989) who showed that eruptive rates for flood basalts, such as the Deccan Traps, are an order of magnitude higher than for their associated aseismic rides (e.g. the Chagos-Laccadive ridge), which provided strong support for the suggestion that flood basalts are produced by melting plume heads and that aseismic ridges are produced by melting plume tails. Griffiths and Campbell (1990) developed a quantitative model for plumes that Campbell and Griffiths (1990) tested against the observations that were available at the time. In spite of the excellent agreement between theory and observation, a number of papers have been published claiming that there are observations that are inconsistent with the plume hypothesis.

The physics of low Reynolds number plumes is well understood. As a consequence, the mantle plume hypothesis makes numerous testable predictions, many of which are quantitative. The purpose of this paper is to set out these predictions and to test them against the observations.

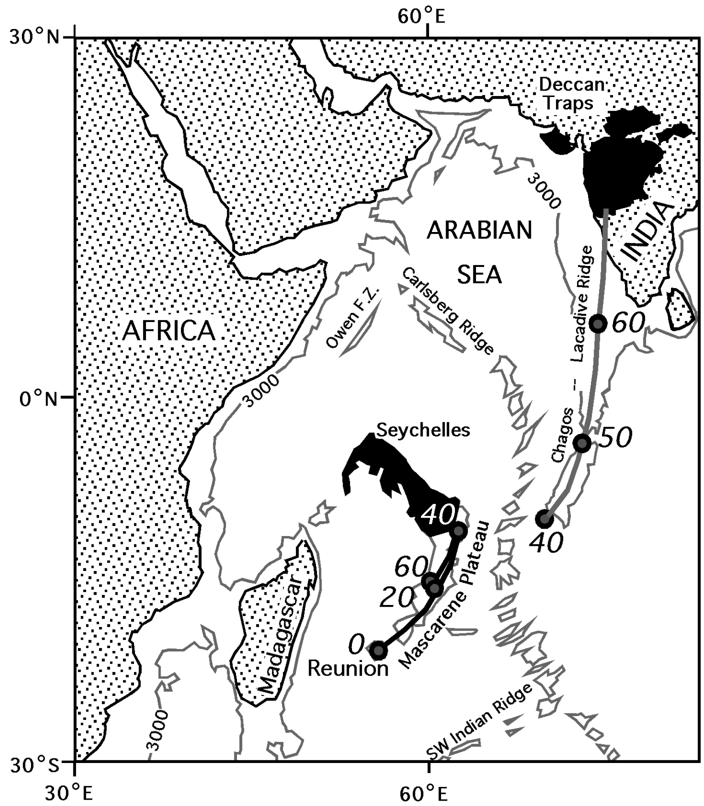

Figure 1 Map of the western Indian Ocean, showing the distribution of volcanic rocks associated with the Reunion-Deccan plume. Seychelles were part of the Deccan Traps prior to its separation, caused by the spreading on the Carlsberg Ridge. Note that the Deccan Traps are connected via the 200–300 km wide Chagos-Laccadive Ridge, across the Carlsberg Ridge spreading centre, to the Mascarene Plateau and eventually Reunion Island, the current position of the plume. (Adapted from White and McKenzie, 1989.)

The mantle plume hypothesis

Convection in fluids is driven by buoyancy anomalies that originate in thermal boundary layers. Earth's mantle has two boundary layers. The upper boundary layer is the lithosphere, which cools through its upper surface and eventually sinks back into the mantle, driving plate tectonics. The lower boundary layer is above the contact between the core and mantle. High-pressure experimental studies of the melting point of iron-nickel alloys show that the core is several hundred degrees hotter than the overlying mantle. A temperature difference of this magnitude must produce an unstable boundary layer above the core which, in turn, must give rise to mantle upwellings. This conclusion is supported by dynamo theory, which also requires a large heat flow out of the core in order to produce Earth's magnetic field. Theory, laboratory experiments, and numerical models all demonstrate that the likely form of

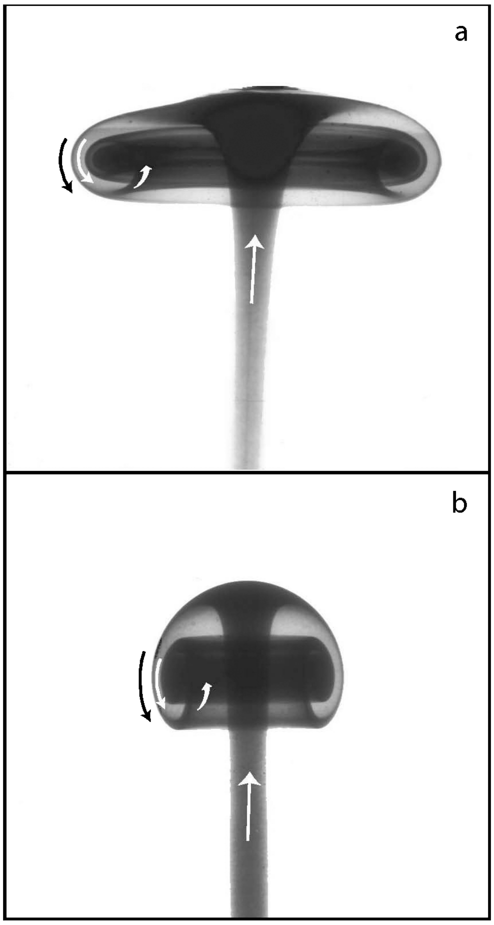

Figure 2 Photograph of a laboratory model of a starting thermal plume (A) mid-way through its ascent and (B) at the top of its ascent. The dark fluid is hot material from the plume source and the lighter fluid is cooler entrained material. The white arrows show motion within the plume and the black arrows the direction of motion in the boundary layer adjacent to the plume; the fluid in the boundary layer has been heated by conduction so that its density is approximately the same as that of the plume. (After Griffiths and Campbell, 1990).

upwelling is columnar, with a large 'head' preceding a narrow 'tail'. Thus mantle plumes are the inevitable consequence of a hot core.

If the core is hotter than the mantle, heat conducted out of the core must warm the overlying mantle and lower its density. The material in the resulting boundary layer is then lighter than the overlying mantle and begins to rise but, before it can do so at a significant rate, it must gather enough buoyancy to overcome the viscosity of the mantle that opposes its rise. As a consequence, new plumes have a large head followed by a relatively small tail. The tail or feeder conduit is comparatively narrow because the hot, upwelling fluid has a relatively low viscosity and thus requires only a narrow existing conduit up which to flow. As the head rises through the mantle, it grows for two reasons (Griffiths and Campbell, 1990). First, material in the high-temperature, low-viscosity tail rises faster than the head and feeds a constant flux of hot mantle into the head of the plume. When this material reaches the stagnation point at the top of the plume, it flows radially to give the head its characteristic doughnut shape. Second, as the plume rises, heat is conducted into the adjacent mantle. The temperature in the boundary layers adjacent to the head and tail increases, and their density decreases, so that they begin to rise with the plume. This material becomes part of the plume and is swept into the base of the head by its re-circulating motion. The head is therefore a mixture of hot material from the plume source region and cooler entrained mantle (Figure 2). Its average temperature is intermediate between the high temperature of the plume tail and the cooler entrained mantle. When the plume head reaches the top of its ascent it flattens to form a disk with a diameter twice that of the head. Note that both mechanisms responsible for the growth of the plume head as it rises through the mantle are a direct consequence of the strong temperature dependence of the mantle's viscosity. Plume heads in isoviscous computer models of mantle convection do not grow or entrain as they rise because they omit this essential aspect of the physics.

Numerical models of mantle plumes, which take into account the strong temperature dependence of the mantle's viscosity, confirm that the essential physics of the laboratory plumes are applicable to the mantle. However, they also introduce two additional considerations that are not easily modeled in laboratory experiments. First, as the plume head passes from the high-viscosity lower mantle into the lower-viscosity upper mantle, the top of the head, which is first to enter the lower-viscosity upper mantle, rises faster than the rest of the plume. This causes the head to narrow as it passes though the 670 km discontinuity, but it recovers most of its shape by the time it reaches the top of the mantle. Second, recent models of thermo-compositional plumes show that plumes, which include a compositionally dense component, may stall at the 670 km discontinuity with only part of the head, presumably a compositionally light part of the plume, ascending into the upper mantle (Farnetani and Samuel, 2005).

Predictions and observations

The plume hypothesis makes the following testable predictions.

New plumes consist of a large head followed by a narrow tail

Flood basalts and oceanic plateaus, the oceanic equivalent of flood basalts, are the first eruptive product of a new mantle plume. The volumes of basalt produced during these events are enormous and flood basalt provinces typically cover an equi-dimensional area that is 2,000 to 2,500 km across (White and McKenzie, 1989). Furthermore, eruption times are short, typically about 1 Myr, so that the eruption rates are the highest seen in the geological record. Most young flood basalts and oceanic plateaus are linked to the current position of the parent plume by a chain of volcanoes that is 200 to 300 km wide (Figure 1). Richards et al. (1989) have shown that eruptive rates for flood basalts are one to two orders of magnitude higher than that of the associated ocean island chain that connects them to the current position of the plume. This observation fits well with the starting plume hypothesis, which attributes the high eruption rates of flood basalts to melting of plume heads, and the lower eruptive rates of ocean island chains to melting of smaller plume tails.

Plumes must originate from a hot boundary layer, probably the core-mantle boundary

The obvious way of showing that plumes originate from the core-mantle boundary is to use seismic methods to trace plume tails from the top of the upper mantle to their source. However, the small diameter of plume tails (100±50 km) makes them difficult to resolve and attempts to use seismic methods to image plume tails for a long time met with little success. However, Montelli et al. (2004) pioneered a new technique which they claim can resolve plume tails. They have used this method to trace the Ascension, Azores, Canary, Easter, Samoa and Tahiti plumes to the core-mantle boundary. A number of other plumes, including Iceland and Yellowstone, are not resolved in the lower mantle. Montelli et al. note that it is more difficult to image plumes in the lower mantle than in the upper mantle. Their method shows great promise and may eventually allow unambiguous imaging of plume tails in both the upper and lower mantles.

Flattened plume heads should be 2,000 to 2,500 km in diameter

The diameter of a plume head (D) depends on the temperature difference between the plume and the adjacent mantle (its excess temperature ΔT), its buoyancy flux (Q), the kinematic viscosity of the lower mantle (υ) and its height of rise (Z), as described in equation (1)

$$D = Q^{1/5}(\upsilon/g\, \alpha\, \Delta T)^{1/5} K^{2/5} Z^{3/5} \qquad (1)$$

where g is gravitational acceleration, α is the coefficient of thermal expansion and K is the thermal conductivity. Note that the plume height of rise is raised to the power 3/5, whereas most other terms are raised to the power 1/5. It is therefore the height of rise of the plume, which in the case of Earth is the depth of the mantle, which has the dominant influence on the size of plume heads. If ΔT of plumes is assumed to be 300 ºC, and their buoyancy flux to vary between $3{\times}10^3$ and $4{\times}10^4$ N s^{-1}, the calculated diameter of a plume head, originating at the core-mantle boundary, is 1,000 to 1,200 km. The plume head should flatten to produce a disk with diameters between 2,000 and 2,400 km when it reaches the top of the mantle.

When a plume head rises beneath continental crust the associated buoyancy anomaly lifts the lithosphere, which places it under tension. This can lead to runaway extension and to the formation of a new ocean basin, though the rifting can sometimes take several million years to fully develop (Hill, 1991). During the initial stages of rifting, the hot mantle in the plume head is drawn into the spreading centre to produce thickened oceanic crust. If the line along which the continent splits lies close to the centre of the plume head, as was the case when the North Atlantic opened above the Iceland plume head, the length of thickened oceanic crust should be equal to the diameter of the plume head — 2,000 to 2,400 km long. Figure 3 shows the zone of thickened oceanic crust along the east Greenland coast and Rockall-Vøring plateaus, those associated with the break-up of the North Atlantic. Notice that the zone of thickened oceanic crust, on both sides of the North Atlantic, is ~2,400 km long.

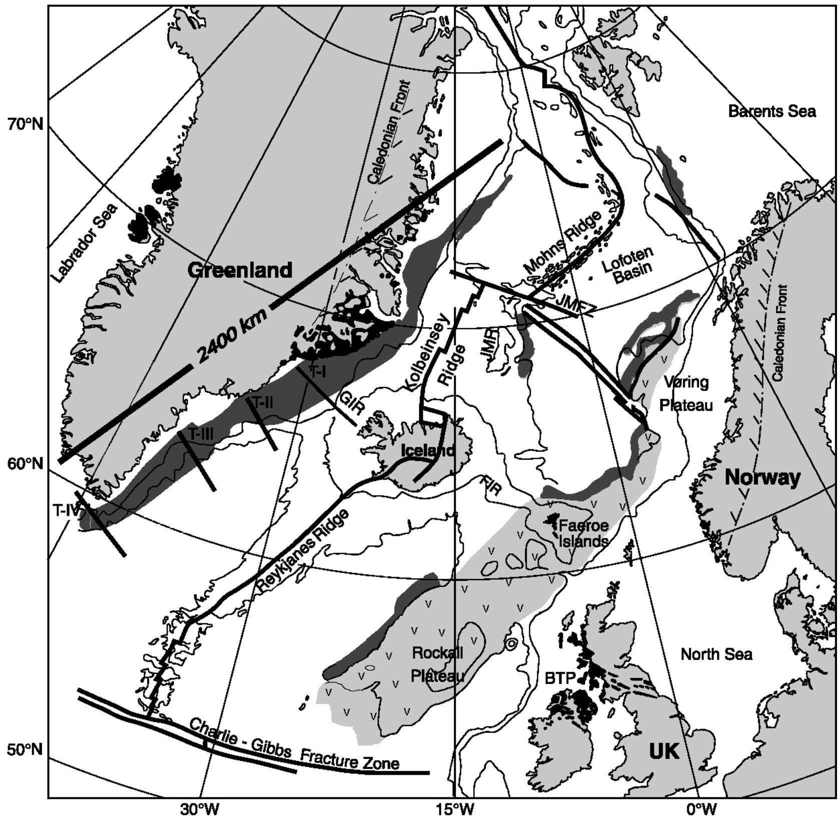

***Figure 3** Geological Map of the North Atlantic region after Hopper et al. (2003). Areas shaded black are onshore basalt exposures, which erupted from the plume head prior to opening of the North Atlantic. The dark gray areas locate offshore seaward dipping reflectors, which are inferred to be thickened oceanic crust and light gray areas with v's show the areal extent of basalts associated with the initial opening of the ocean. T-I to T-IV are the seismic transects along the East Greenland coast. Abbreviations are: BTP, British Tertiary Province; GIR, Greenland Iceland Ridge; FIR, Faeroes-Iceland Ridge; JMR, Jan Mayen Ridge; JMFZ, Jan Mayen Fracture Zone.*

Plume heads can produce millions of cubic kilometers of magma

Melting of ascending normal mantle is believed to begin 60–80 km below mid-ocean ridges. Plume heads that rise under continental lithosphere or old oceanic lithosphere are not expected to reach such shallow levels, and it was thought for some time that this would prevent significant melting of plume heads and thus invalidate the plume head model of flood basalt eruptions. However, three factors have led to a revision of this conclusion.

First, mantle plumes are not expected to have normal mantle composition. Geochemical observations (Hofmann, 1997) show that the sources of hotspots are enriched in incompatible elements, and this enrichment is widely thought to be due to the presence of extra basaltic-composition material in the plume source due to the recycling of ancient oceanic crust through the D" zone of the mantle (Hofmann and White, 1982). The old oceanic crustal material, in the form of eclogite, will melt deeper and cause more extensive melting than in normal mantle (Cordery, 1997; Campbell, 1998).

Detailed numerical modeling of the ascent of a starting plume has revealed two more factors that enhance melting (Leitch and Davies, 2001). The second factor is that the growth of a plume from a thermal boundary layer ensures that the hottest material is at the center of the plume and reaches the shallowest depth. Third, narrowing of the plume head as it passes though the viscosity drop across the 670 km discontinuity allows the head to "punch through" to shallower depths than if there were no viscosity change. Figure 4 illustrates the distribution of temperatures and melting rates within a modeled plume head at two late stages of its ascent. This plume head is assumed to have 15 vol% of eclogite.

Modeling of melting in a mantle plume head by Leitch and Davies (2001), which takes these three factors in account, typically yields melting rates of 1–10 km^3/a and total melt volumes of 1–20 million km^3 (Figure 5), depending on the assumed conditions of melting, especially the minimum depth to which the top of the plume head rises. These results cover the range of observed melting rates and volumes of most Large Igneous Provinces (LIP). Note that the Leitch and Davies (2001) model ignores the expected back reaction between the basalt produced by eclogite melting and the adjacent peridotite (Campbell, 1998), and melting of the peridotite. Ignoring the former will lead to an overestimation of the amount of melt produced and the latter to an underestimation.

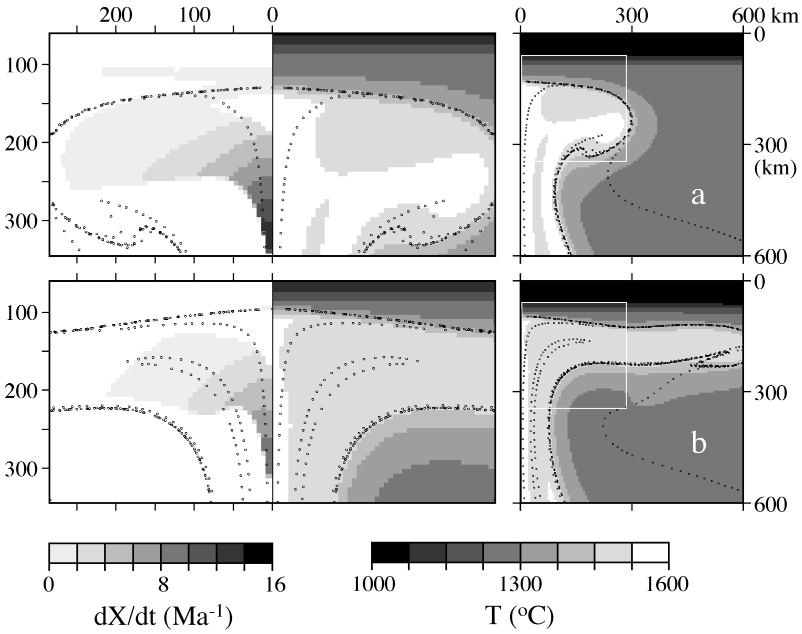

Figure 4 Distributions of melting rate and temperature in a numerical model of a plume head at two late times (see Figure 5). The middle panels are blowups of the squares outlined in the right panels. Temperature is lowered in the melting zones by the latent heat effect. The plume is assumed to have 15% eclogite, in the form of inclusions in a peridotite matrix. Peridotite melting is secondary and is not calculated, which is why the melting zone does not extend to the top of the plume on the plume axis and the expected back reaction between the basalt produced by eclogite melting and the adjacent peridotite has been ignored. From Leitch and Davies (2001).

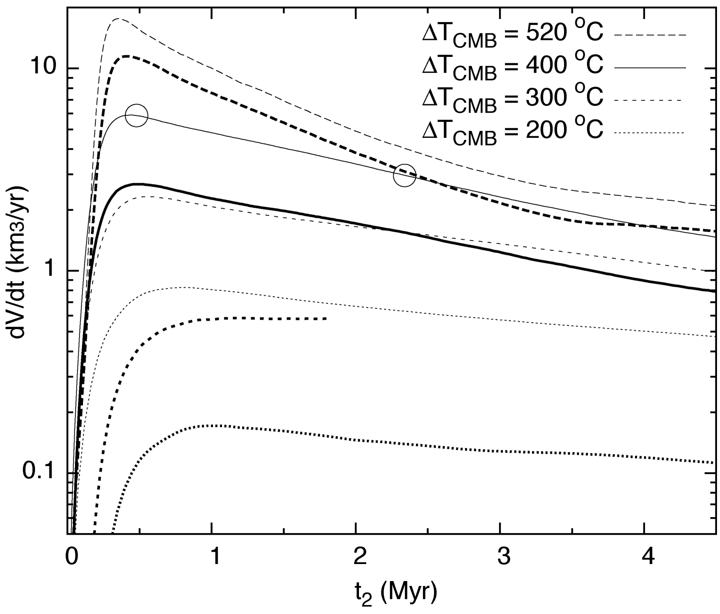

Figure 5 Melting rate versus time since the start of melting for plumes with 15% eclogite component. Thin lines assume melting starts at 60 km depth, and thick lines 90 km depth. ΔT_{CMB} is the effective temperature excess at the core-mantle boundary that generates the plume. Circles indicate times of snapshots in Figure 4. From Leitch and Davies (2001).

Rifting over a plume head should produce a narrow zone of thickened oceanic crust

When a new oceanic ocean basin opens above a previously emplaced plume head, the first oceanic crust to form is anomalously thick, for example, the margins of South America and Africa above the Parana plume, and India above the Deccan plume (White and McKenzie, 1989). When new oceanic crust forms away from the influence of plumes, it has normal thickness.

The zone of thickened oceanic crust at the rift margins is typically only 100–200 km wide (Figure 3). At first sight this may seem to be inconsistent with melting of a plume head thousands of kilometers across. However a numerical model of the process (Figure 6) shows that excess melting occurs only as long as plume head material is being pulled up into the rift (Leitch, et al., 1998). Since the flattened plume head is only 150–200 km thick, this is accomplished after only 150–200 km of seafloor spreading. Thereafter normal mantle is pulled up and the oceanic crustal thickness returns to normal. Comparisons of calculated crustal thickness profiles with observed profiles (Figure 7) confirm this.

Both heads and tails should erupt high-temperature picrites

The ΔT of mantle plumes can be estimated from the maximum MgO content of their erupted magmas because, for dry melts, there is a linear relationship between MgO content and magma temperature. As a rough rule, a 4 wt% increase in MgO in the melt equates to a 100 ºC increase in mantle potential temperature. The maximum MgO content of plume-derived picrites varies between 18 and 22 wt%, suggesting that the temperature excess for mantle plumes lies within the range 150 and 250ºC. Examples of volcanic provinces, which have been attributed to plumes, that include high-MgO picrites are: Reunion-Deccan, Parana, Iceland-North Atlantic Province, Siberian Traps, Karroo, Emeishan, Caribbean and Hawaii.

The temperature excess of a plume head is predicted to be ~300 ºC at the centre of the head and to taper to ~100 ºC towards the margin

The thickness of the oceanic crust produced when new oceanic crust opens up above a plume head is dependent on the temperature of the mantle that is drawn into the new mid-ocean ridge spreading centre. The plume hypothesis predicts that the temperature of a plume should be highest at the plume axis where the tail rises through the centre of the head. Here ΔT is expected to be 300 ±100°C. The ΔT of the remainder of the head, which is a mixture of hot material from the boundary layer source and cooler entrained lower mantle, varies with the plume buoyancy flux but must be less than the temperature at the centre. For typical plume buoyancy fluxes the average ΔT ~100°C. Hopper et al. (2003) have used seismic reflection and refraction data to determine the thickness of the first oceanic crust to form when the North Atlantic opened above the Iceland plume for four traverses from the plume axis to its margin. They obtained thicknesses 33 and 30 km for traverses T-I and T-II, close to the plume axis, and 18 and 17 km for T-III and T-IV, closer to the margin of the head (see Figure 3). The ΔT required to produce these crustal thicknesses, based on the work of McKenzie and Bickle (1988), are 350 and 100°C for the centre and margin of the head, respectively, which is consistent with the plume hypothesis.

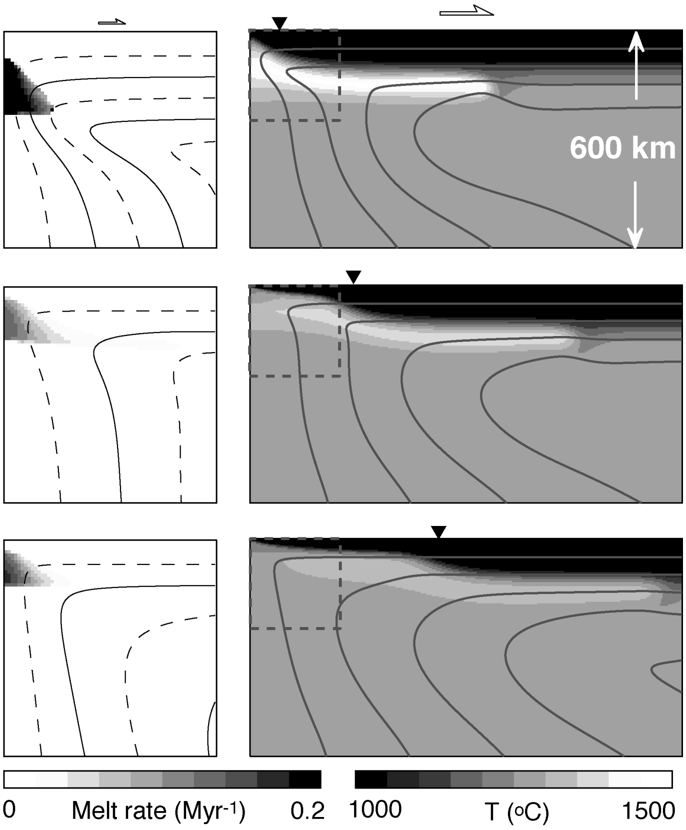

Figure 6 Time sequence from a numerical model of rifting over a previously-emplaced plume head: melting rate (left) and temperature (right). The left panels are blowups of the dashed squares in the right panels. Curves are streamlines and the triangles mark the rift margin. The plume head material is pulled up into the rift and then horizontally away, to be replaced by normal mantle. Excess melting occurs only while the hotter plume head material is being pulled through the shallow rift zone, so thickened oceanic crust is formed only at that time. From Leitch et al. (1998).

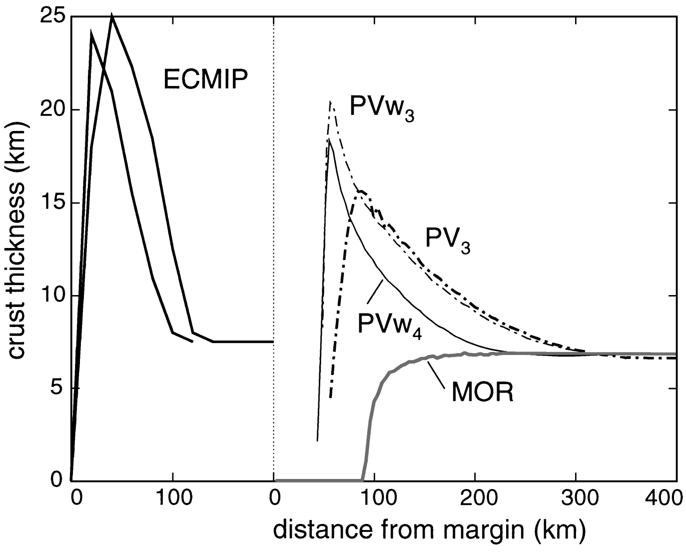

Figure 7 Calculated thicknesses of oceanic crust from rift models like that in Figure 6 (right panel) compared with measured crustal thicknesses in two profiles across the East Coast Margin Igneous Province of North America (left panel). Calculated thickness from a model with no plume head (MOR) is included for comparison. From Leitch et al. (1998).

Picrites should erupt early during flood volcanism and be most abundant near the centre of the plume head and less abundant towards the margin

The hottest material in the head is the mantle from the plume source (the dark colored fluid in Figure 2), which is 300±100 °C hotter than the entrained mantle. Although this temperature difference decreases with time, as adjacent layers exchange heat, the high temperatures of the hot layers will persist for millions to 10's of millions of years, depending on their thickness, which in turn depends on their distance from the plume axis (see Figure 2). When a plume head melts to form a flood basalt province, only the top of the head ascends to a level in the mantle where the pressure is low enough to allow melting. Note in Figure 2 that the hot layer at the top of the plume thickens towards the centre, where it grades into the tail. Provided the first magmas for a new plume don't pond and fractionate in crustal magma chambers, picrites should dominate the early melting products of plume heads and become less abundant as the cooler, second layer enters ascends to a level where it can begin to melt. Picrites should also be most abundant towards the centre of the plume head and become less abundant towards the margins.

Unfortunately, the early melting products of plume heads are rarely seen in flood basalts, especially near the centre, because they are commonly covered by later flows. However, the predicted early picrites have been documented for the Parana-Etendeka, Deccan, Emeishan, North Atlantic, Siberian Traps and Karoo flood basalts. A greater abundance of picrites at the centre of a flood basalt province, thinning towards the margins, can be seen in the Letaba picrites, along the Lobombo monocline of the Karroo, and in the deeply dissected valleys of the Emeishan flood basalt province in China.

Flood volcanism should be preceded by domal uplift of between 500 to 1000 m at the center of the dome

The arrival of the hot plume head in the upper mantle will produce domal uplift at the surface, the magnitude of which depends on its average temperature. The area of maximum uplift is predicted to have a radius of ca. 200 km and to be surrounded by a zone, with a radius of ca. 400 km, in which uplift is still significant (Figure 8).

The best-documented example of domal uplift occurs in association with the upper Permian Emeishan flood-basalt province in China (He et al., 2003). Here the carbonate beds of the underlying Maokou Formation have been systematically thinned by erosion towards the centre of the flood-basalt province. Iosopachs of the

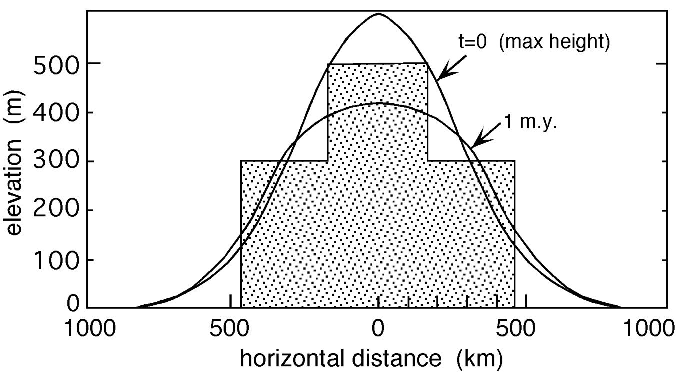

Figure 8 Uplift above a plume head, as predicted by Griffiths and Campbell (1991), compared with the uplift observed at the centre of the Emeishan flood basalt by He et al. (2003). Predicted profiles are given for maximum uplift (t = 0), when the top plume is at a depth of ~250 km, and 2 Myr later (t = 2 Myr) when flattening of the head is essentially complete. The uplift for the Emeishan flood basalt province is the minimum average value for the inner and intermediate zones as determined from the depth of erosion of the underlying carbonate rocks.

Maokou Formation show that uplift was broadly dome shaped as expected. He et al. recognize three zones: an inner zone with a radius of ca. 200 km, where uplift is estimated to be at least 500 m and could possibly exceed 1000 m, an intermediate zone, with a radius of 425 km in which the average uplift is ca. 300 m; and an outer zone of radius 800 km in which uplift is minimal. The magnitude and shape of the uplift agrees remarkably well with that predicted by Griffiths and Campbell (1991) (Figure 8) and the shape of the uplift with Farnetani and Richards (1994). Farnetani and Richards (1994) predict more uplift than Griffiths and Campbell (1991) because they assume a higher plume head temperature in their model. Other examples of uplift prior to volcanism include the Wrangellia province of northwest Canada and southwest Alaska, the Natkusiak province in northwest Canada, the Deccan flood basalt of India and the Ethiopian flood basalt. The timing of uplift has been documented for the Emeishan and Wrangellia volcanism where it started 3 to 5 Myr before volcanism.

The fixed position of hot-spots

The lower mantle is believed to have a viscosity that is about a factor of 30 greater than the average viscosity of the upper mantle, which means that convective velocities, which scale with $\upsilon^{2/3}$, should be about a factor of ten times slower in the lower mantle than they are in the upper mantle. The slow convective motion in the lower mantle pins the entry point of plumes into the upper mantle so that plumes are expected to move ~10× more slowly than plates in the upper mantle. As a consequence, their position is effectively stationary relative to plates and the path traced out by chains of plume-related islands, such as the Hawaiian–Emperor chain and Chagos–Laccadive Ridge, should follow directions dictated by plate motion. Morgan (pers. com. 2003) showed that all of the plume-related volcanic chains of the Pacific plate do follow the path predicted by plate tectonics. In addition, the age progression of the volcanoes along the different chains varies unidirectionally in a way that is consistent with the plate motion and with each other. These observations are convincingly explained by the plume hypothesis.

Thermo-compositional plumes

An obvious weakness of the thermal plume hypothesis is that it fails to explain a number of minor volcanic chains that stretch across the Pacific ocean basin and cannot be linked to LIPs. These appear to be the product of plume tails without heads. It has been suggested that the plumes responsible for these volcanic chains may originate in the mid-mantle (Courtillot et al., 2003). A recent computer model of thermo-compositional plumes, by Farnetani and Samuel (2005), provides a possible solution. As noted earlier they have shown that if the head of a weak plume contains a high proportion of a dense component in the mantle (e.g. subducted former basaltic crust), the ascent of the head can stall at the 670 km discontinuity and it can separate into low- and high-density components. Only the light component penetrates the discontinuity and gives rise to a weak plume with a lower temperate excess than the strong plumes associated with LIPs. The suggestion that plumes may have an above average mantle concentration of dense, subducted former oceanic crust is consistent with the usual interpretation of D", a heterogeneous, seismically fast layer of variable thickness (300±300 km) at the base of the mantle, as a zone where subducted slabs have concentrated. It is also consistent with the isotopic (both stable and radiogenic) and trace element characteristics of many plume-derived magmas, which also require a higher than average mantle concentration of former oceanic crust (Hofmann and White, 1982).

Conclusions

The mantle plume provides a simple explanation for all of the essential features of classic LIPs, such as the Deccan Traps and the North Atlantic Province, and all of its predictions have been confirmed by observation. Five days of intense scrutiny during the Great Plume Debate (Fort William, 28 August–1 September 2005) failed to present a serious challenge to the mantle plume hypothesis, and no credible alternative was presented to explain the principal features of LIPs, especially their high-temperature magmas and large eruptive volumes. However, this does not mean that all intraplate volcanoes and volcanic chains are due to mantle plumes. Each case must be considered on its merits.

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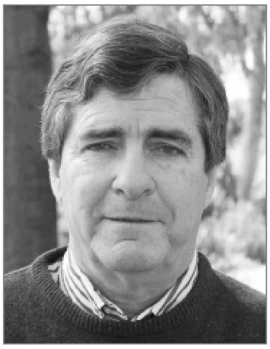

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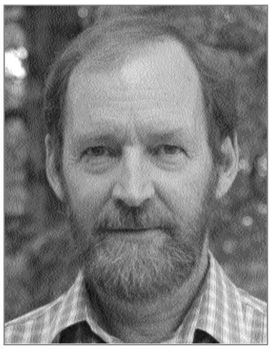

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