Earth and Planetary Science Letters, 99 (1990) 66–78
Elsevier Science Publishers B.V., Amsterdam

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Stirring and structure in mantle starting plumes

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Received October 17, 1989; revised version accepted March 16, 1990

ABSTRACT

Simple arguments show that ascending thermal plumes will entrain their surroundings as the result of coupling between conduction of heat and laminar stirring driven by the plume motion. In the initial stages of ascent of a plume fed by a continuous buoyancy flux (a starting plume) the plume consists of a large buoyant head followed by a narrow vertical conduit. Laboratory experiments reported here show that the spherical head entrains ambient material as it rises, while the axial conduit carries hot source material to the stagnation point at the cap of the plume, from where it spreads laterally into thin laminae. We develop an analysis of the effects of entrainment on the structure and dynamics of starting plumes. The analysis predicts that under conditions appropriate to the earth's mantle large volumes of cooler lower mantle will be stirred into the head of a plume by the time it reaches the top of the mantle if it originates at the core–mantle boundary. The result is a major cooling and enlargement of the head. Source material ascending in the trailing conduit will undergo little contamination or cooling until the conduit is deflected from the vertical by large scale shear associated with plate motion. This plume structure explains the close association of high-temperature melts (komatiites or picrites) with more voluminous, lower temperature basalts in Archaean greenstones and modern continental flood basalt provinces: the picrites can be produced by melting in the hot axial conduit and the basalts from the cooler bulk of the head. More generally, we put forward stirring in plumes as one plausible mechanism contributing to compositional heterogeneity in hotspot melts.

The predicted diameter of plume heads originating at the core–mantle boundary is ~ 1000 km and this is expected to enlarge to ~ 2000 km when the plume collapses beneath the lithosphere. This result is in excellent agreement with the observed extent of volcanism and uplift associated with continental flood volcanism. It also provides support for the hypothesis that at least some plumes originate at the core–mantle boundary.

1. Introduction

Basalts erupted at oceanic and continental hotspots originate from zones of melting having potential temperatures greater than normal mantle [1,2] and are therefore attributed to plumes of hot material upwelling from deep in the mantle, as earlier proposed by Morgan [3,4]. Hotspot basalts formed from the modern mantle also tend to show a characteristic enrichment of light rare earth elements relative to mid-ocean ridge basalts (MORBs). However, melts erupted at a given hotspot can vary from alkalic to tholeiitic (and to picritic in the case of flood basalts), indicating that partial melting took place over a range of temperatures. Similarly, Archaean greenstone belts, also apparently of plume origin [5], involve bimodal eruptions of high-temperature komatiitic melts and lower-temperature basalts. Geochemical variability too can be seen in both the radiogenic isotopes and the level of LREE enrichment. For example, the picrites found in modern continental flood basalt provinces are commonly more enriched in character than are the basaltic melts of intermediate temperature [6]. Isotopic ratios in flood basalts and in ocean island basalts (OIBs) are widely scattered in a manner that implies mixing of a number of end-member source compositions [7–9]. In this paper we concentrate on the possibility of entrainment of surrounding mantle into hot plumes. As well as modifying the dynamics and physical characteristics of mantle plumes, stirring may contribute to compositional variance in hotspot basalts.

Griffiths [10–12] demonstrated that when a hot buoyant volume of fluid is placed in very viscous fluid the hot material will form a rising spherical vortex which continuously heats and entrains the adjacent fluid as it ascends. This thermally buoyant diapir (a "thermal") could occur if mantle plumes break away and become isolated from their source regions, either because the departure of a large

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mass of hot material from the source region (such as the bottom boundary layer of the mantle) leaves an insufficient volume of buoyant mantle in the source region to maintain a continuing flow into the diapir, or because individual diapirs break away from an unstable plume conduit in the manner suggested by Skilbeck and Whitehead [13]. Within such an isolated thermal, recirculation driven by stresses induced by the ascent wraps the entrained mantle material into the diapir. Despite a very small thermal conduction length (i.e. a large Peclet number) a large volume of outer material can be heated and stirred into the thermal during its ascent. The stirring, in combination with conduction of heat, leaves only small temperature gradients within the thermal, effectively expelling these gradients to the perimeter of the thermal. In this case stirring is not effective in continuing the shift of compositional heterogeneity to smaller and smaller scales, since most of the material from the plume source simply becomes confined to a torus (which has a "doughnut" appearance in laboratory experiments).

The isolated diapir is the most fundamental structure with which to illustrate the mechanism of entrainment and its potential effects. However, it probably does not reflect the structure of most plume activity in the mantle. Other plume structures likely to occur in the mantle are the newly initiated (or "starting") plume and the continuous narrow conduit that follows a starting plume. Starting plumes are of interest because the growth of new plumes (and disappearance of old ones) normally occurs in unsteady three-dimensional convection at large Rayleigh numbers and can reasonably be expected in the mantle if a component of the convection is driven by an unstable bottom boundary layer. In isothermal experiments with compositionally buoyant source fluid it has been shown that a large volume of buoyant fluid must accumulate in a diapir before the plume can penetrate the overlying fluid [14,15]. Later, a continued supply from the source can be carried up into the ascending head through a narrow feeder conduit, further enlarging the head. However, there has been no discussion of the possibility of entrainment and its effects on plume structure, temperature and chemistry in the context of the mantle. This is surprising given that thermal convection is characterised by an essential coupling between advection and diffusion of heat, and the well-known occurrence of "mushroom" shaped heads on starting plumes from a heated horizontal boundary under laboratory conditions. These plume heads accumulate buoyant boundary layer fluid but also entrain large volumes of surrounding fluid as they ascend.

The ascent and melting of the large head of a new plume has been suggested as the cause of the anomalously large rate of eruption of modern flood basalts over short periods of time near the beginning of many hotspot tracks [4,16,17]. Arrival of a starting plume near the top of the mantle has also been suggested as the explanation for the large volume and small age distribution of basalts in Archaean greenstone provinces [5]. In the latter case we have already suggested that stirring of cooler mantle into the head may account for the very large lateral scale of greenstone provinces and for the close association between high-temperature komatiitic and lower-temperature basaltic melts. Another explanation of the production of flood basalts, put forward by White and McKenzie [18], involves externally imposed lithospheric extension coincident with a hot plume. In their model a steady flow ascends and cools by conduction as it spreads laterally beneath the lithosphere. However, the steady plume does not explain the dominant characteristics of continental flood basalt provinces, namely their coincidence with the initiation of hotspot tracks, the absence of volcanism prior to the flood basalt eruptions and the very short duration of anomalous rates of eruption. Hence we proceed to explore the dynamics of starting plumes. We do not address crustal problems such as melt generation and the role of externally imposed extension and rifting.

A quantitative evaluation of the effects of thermal entrainment and mixing in mantle plumes is a necessary prerequisite for a comparison of plume models with physical characteristics of hotspots and may also be essential in understanding the varied geochemistry and isotopic characteristics of their magmas. In this paper we present solutions modelling the flow in the heads of newly initiated mantle plumes as they travel from their source region to the surface. The amount of convective stirring and cooling within each plume structure and the lateral scale of plumes on reaching the top of the mantle are predicted within the limits set by uncertainties in a number of mantle properties. These calculations complement the earlier studies of isolated thermals [10–12,19]. We begin by reporting observations of the dynamics of laboratory starting plumes fed by a constant flux of buoyancy from a source, and show that entrainment into the plume heads is similar to, but leads to more effective stirring than, that in isolated thermals. In a companion paper [6] we consider data from modern flood basalt provinces and make a detailed comparison with the implications of the predicted plume structure.

2. Experiments with starting plumes

The initiation of new plumes is normally modelled in the laboratory using viscous fluids of different densities but the same temperature [14,15]. The aim of the present experiments was to investigate the effects of heat conduction on the behaviour of thermally buoyant starting plumes fed by a constant buoyancy flux and having a much smaller viscosity than their surroundings. Clear glucose syrup having a viscosity of 150 St (1.5×10^{-2} m^2 s^{-1}) at 20°C was used as the working fluid. This was placed in a Perspex tank 30 cm square to a depth of 48 cm. Some of the same syrup was mixed with a small quantity of dye and placed in an elevated reservoir, heated and gravity-fed through a 2 cm diameter tube to the base of the tank. The supply syrup was reheated by passing through a metal heat exchanger immediately before entering the tank through a 0.6 cm diameter hole. The heat exchanger was connected to a hot water bath and held at 80°C. In order to minimise heat conduction into the tank, the exchanger was plugged into the base of the tank only a few seconds before commencing a run. The flowrate was adjusted by a small stopcock and determined from readings of the head loss in the reservoir. In each run it was constant and of order 0.3 cm^3 s^{-1}. The dyed syrup entered at 80°C (viscosity 0.5 St) whereas the syrup in the tank was at 21°C (140 St).

When the flow was turned on the dyed syrup inflated a sphere adjacent to the inlet hole. After about 60 s the sphere was large enough to rise faster than the rate at which its radius was increasing. It then moved away from the base but remained connected to the source via a narrow feeder conduit. Up to this stage the flow was indistinguishable from that reported for compositionally buoyant injections [14]. Ascent of the enlarging sphere, however, soon induced recirculation within, while conduction of heat made a thin boundary layer of the surrounding syrup buoyant. Undyed surroundings were thereby entrained and wrapped around inside the plume head (Fig. 1a, b, c) in a manner similar to that seen in fixed volume injections [10]. This behaviour of hot plumes contrasts with that in isothermal experiments [12], where no obvious contamination occurs and the plume heads enlarge only as a result of the continuing source input.

The measured volume of the plume head (Fig. 2) increased with time faster than expected from the source input alone. Its rise velocity also continued to increase slowly with time (or height) despite increasing drag on the enlarging head. Once the head had risen three to six diameters from the source the bulk of the source syrup which was in the sphere at the time it moved away from the source had become confined to a doughnut shaped volume encircling the plume axis (marked "A" in Fig. 1c). However, most of the dyed syrup that was supplied after that time was carried upward in the axial conduit to the leading edge of the head (which in a reference frame moving with the head is a stagnation point). From there it spread radially and became strained into a thin surface bounded on both sides by layers of material from the surroundings. Efficient heat transfer is affected between the source and entrained syrups and the temperature anywhere in the head (but outside the axial conduit) must be significantly lower than that in the conduit. Despite the potentially large temperature difference between the bulk of the plume head and the central stem the head remains very close to spherical.

In our experiments the temperature difference gives rise to a maximum viscosity ratio of source fluid to surroundings of approximately 0.003. However, the viscosity of the injected fluid alone determines the diameter of the conduit required to carry the imposed flux [14], whereas the viscosity of the surroundings determines the rate of rise and size of the plume head. Hence the stem of mantle plumes may be more or less narrow relative to the

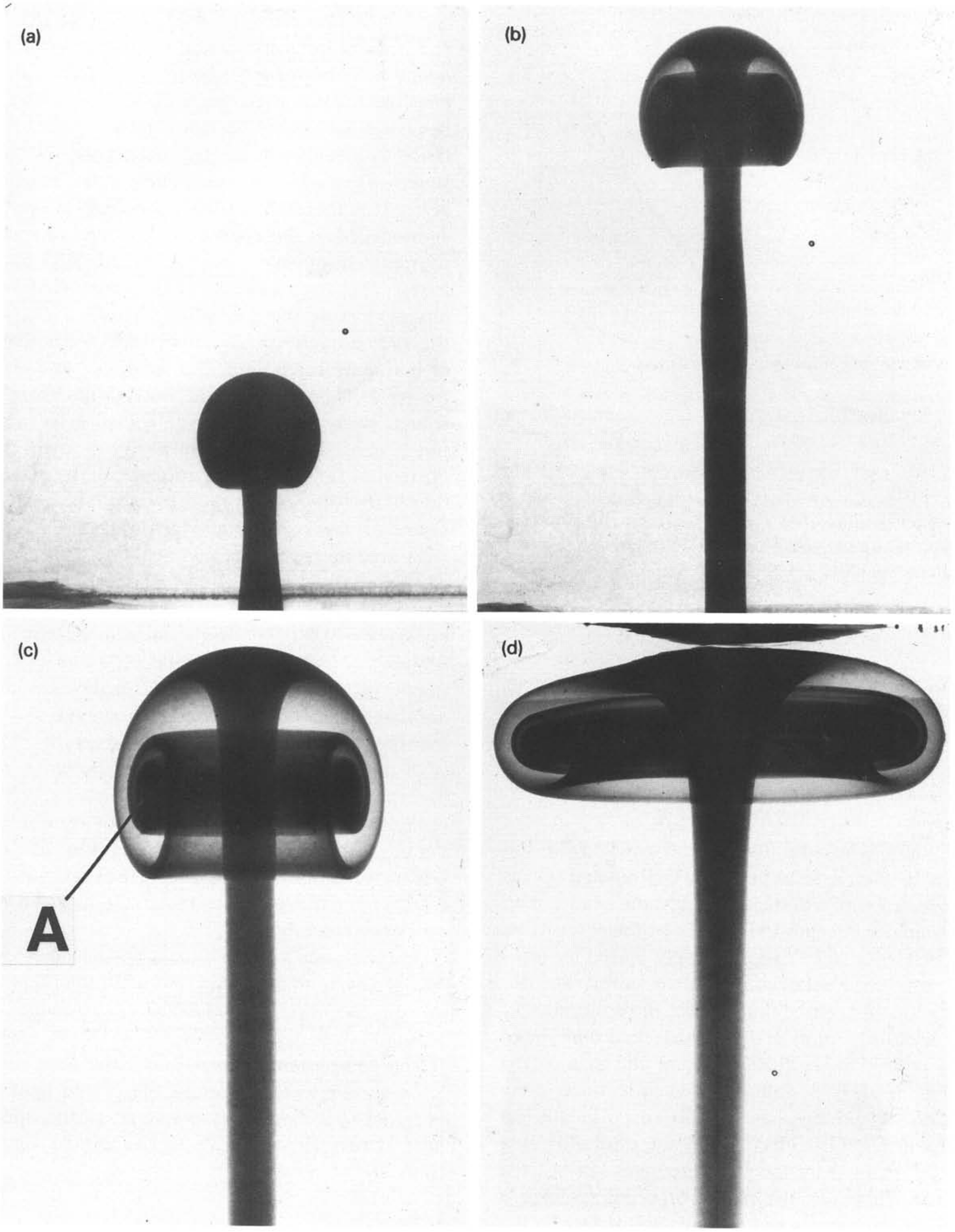

Fig. 1. Photographs of a starting plume in glucose syrup at several stages during its ascent. Times elapsed after the source is turned on are (a) 60 s; (b) 130 s; (c) 397 s; and (d) 540 s. Scale is identical in all frames and the head is 6.9 cm across in (c). The distribution of dye is axisymmetric.

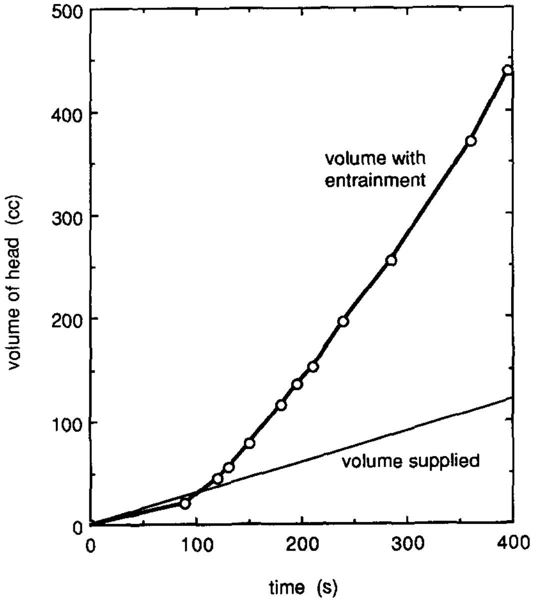

Fig. 2. The volume of the head of a starting plume as measured in a typical experiment. As a result of entrainment the volume increases much more rapidly with time than the linear relationship accounted for by the source flux.

diameter of the head than is the case in the laboratory, depending on the viscosity contrast. Smaller viscosity contrasts are achieved in isothermal experiments and lead to much narrower conduits.

Finally, when the plume head reached the top of its ascent, a near perfect axisymmetry of the plume structure was maintained as the head spread horizontally beneath the surface of the syrup (Fig. 1d). After the head had effectively ceased its vertical motion the stem continued to supply hot uncontaminated source fluid which, in the absence of any horizontal motion of the surface, accumulated in a pool close to the surface at the axis of the plume head. We note that when surface plate motion is present, as in the experiments of Richards and Griffiths [20–22], the conduit is only a little inclined by the time the head reaches the surface. However, the plume proceeds to form a band of source material beneath the surface as the flattened head is carried farther away from the source position and the plume deflection increases with time until it reaches a steady state [20].

3. A theoretical model for starting plumes

Consider plumes emanating from an anomalously hot and unstable boundary layer. With the possible exception of the horizontal spacing, the behaviour of plumes as they ascend a large distance is insensitive to the details of the early stages of growth of perturbations on the boundary layer. It is therefore sufficient to analyse the flow as modelled in the experiments, where a localised source of hot viscous fluid is turned on at a time $t = t_0$. The large amount of buoyancy that accumulates before the plume head moves away from the source models a similar diapiric accumulation of boundary layer fluid. For simplicity and clarity we let both the mass and heat fluxes from the source be constant in time. These fluxes model those carried into the plume by a convergent horizontal flow of heated, softened material in the boundary layer [23], although an unstable boundary layer source may give fluxes which vary with time in unpredictable ways. For the purposes of this analysis we assume that the plume is distant from other starting plumes and neglect effects of larger scales of motion such as those driven by surface cooling or plate motion. (We discuss the important effects of these large scale motions in another paper [29].) It is also assumed that the mantle, aside from heated material in the plume, is of uniform potential density ρ_∞, potential temperature T_∞ and kinematic viscosity ν_∞. More complicated models including, for example, a depth-dependent viscosity or a non-adiabatic density step at the mid-mantle transition zone are readily constructed after understanding the concepts presented here.

If the source volume flux is Q, then the source will at any time have supplied a volume V_s, where

$$V_s = Q(t - t_0) \tag{1}$$

If the source temperature is T_s, the temperature anomaly $\Delta T_s = T_s - T_\infty$ and the plume heat flux becomes $Q_H = \rho c Q \Delta T_s$, where c is the specific heat. Taking a potential density of the simplest form

$$\rho/\rho_\infty = 1 - \alpha(T - T_\infty) \tag{2}$$

the buoyancy flux is

$$Q_B = g\rho_\infty \alpha \Delta T_s Q = g\rho_\infty \alpha Q_H / \rho c \tag{3a}$$

where g is the gravitational acceleration. The nett buoyancy B in the plume at any time is

$$B = g\rho_\infty \alpha \Delta T_s Q(t - t_0) \tag{3b}$$

Our experiments demonstrate that the bulbous head of a plume maintains its nearly spherical shape as it rises. The motion is driven by the buoyancy force associated with the average density difference between the head and its surroundings and is not influenced by density variations within the head. Although the viscosity within the diapir must decrease with time as a result of cooling, we assume that $\nu \ll \nu_\infty$ at all times, in which case the dynamics of the head will not depend on the inner viscosity. The head therefore ascends at a speed U given by Stokes law for a relatively inviscid drop:

$$U = (1/12) g\alpha \Delta T d^2/\nu_\infty \tag{4}$$

where ΔT is the (time dependent) average temperature anomaly and $d = (6V/\pi)^{1/3}$ is the diameter of the diapir. This application of Stokes law to entraining, thermally buoyant diapirs was tested extensively in the laboratory [10] and shown to be valid. Stress-dependent rheologies are expected to lead to only quantitative modifications and not to qualitatively different behaviour.

The temperature anomaly in the head is related to the volume by conservation of heat:

$$\Delta T V = Q \Delta T_s (t - t_0) \tag{5}$$

The total volume V is of course the sum of that supplied by the source and an entrained volume of previously cooler mantle. It obeys the very simple differential equation

$$\mathrm{d}V/\mathrm{d}t = Q + e \tag{6}$$

where e is the rate of entrainment of ambient material. In (5) and (6) it is assumed that the volume of the narrow conduit is negligible compared with that of the head, although this is not quite true when the viscosity ratio, and hence the ratio of conduit to head diameters, is not extremely small.

Entrainment into the plume can be scaled as the volume of a thin boundary layer that is heated and wrapped into the head [10,11]. The thickness δ of this conductive layer is

$$\delta \approx k_1 (\kappa d/U)^{1/2}, \ \delta \ll d \tag{7}$$

and the entrainment flux is

$$e \approx k_2 U d \delta \tag{8}$$

where k_1 and k_2 are similarity constants which are assumed to be independent of time and scale. Combining (7) and (8) gives

$$e \approx k U^{1/2} d^{3/2} \kappa^{1/2} \tag{9}$$

with $k = k_1 k_2$. This single constant must be evaluated empirically and, from the experiments with thermals, lies in the range $2 \leqslant k \leqslant 8$.

We combine (4, 5, 6 and 9) into the single equation

$$\mathrm{d}V/\mathrm{d}t = Q + C\left[\kappa g \alpha \Delta T_s Q(t - t_0)/\nu_\infty\right]^{1/2} V^{1/3} \tag{10}$$

where $C = (1/2\sqrt{3})(6/\pi)^{5/6} k$, and solve to find the evolution of the starting plume. The most convenient initial condition is $V(t_0) = 0$. The empirical value for k implies that, to within a factor of two, $C \approx 2$.

Solutions to (10) are most simply found by numerical integration. However it is instructive to look first at asymptotic solutions for small and large times. For small times (10) reduces to $V = Q(t - t_0)$. Hence the plume head grows only by addition of fluid from the source. In this regime the temperature is constant at $\Delta T \approx \Delta T_s$ while the speed and distance travelled by the centre of the head increase according to

$$U \sim (g\alpha \Delta T_s/\nu_\infty) Q^{2/3} (t - t_0)^{2/3},$$
$$z - z_0 \sim (g\alpha \Delta T_s/\nu_\infty) Q^{2/3} (t - t_0)^{5/3} \tag{11}$$

As time elapses, the second term on the right-hand side of (10) becomes increasingly important and eventually dominant, so that the addition of mass to the head is primarily by entrainment. The rate of entrainment depends on the nett buoyancy in the head, and this continues to increase due to flow up the conduit. The transition occurs after a time of order t_c, where

$$t_c - t_0 \sim \left(C^2 \nu_\infty/\kappa g \alpha \Delta T_s\right)^{3/5} Q^{1/5}$$
$$= C^{-6/5} (Q^2/k^3) R^{-3/5} \tag{12}$$

Here Q^2/κ^3 is the fundamental time scale for the flow and R is analogous to a Rayleigh number:

$$R = g\alpha \Delta T_s Q^3/\nu_\infty \kappa^4 \tag{13}$$

At times $t - t_0 \gg t_c - t_0$ the solution to (10) becomes

$$V \approx (4C/9)^{3/2}\kappa^{15/4}Q^{-3/2}R^{3/4}(t - t_0)^{9/4} \qquad (14)$$

It follows from (4, 5) that

$$\Delta T/\Delta T_s \approx (4C/9)^{-3/2}\kappa^{-15/4}Q^{5/2}R^{-3/4} \times (t - t_0)^{-5/4} \qquad (15)$$

and

$$U \approx (1/8)(6/\pi)^{2/3}C^{-1/2}\kappa^{11/4}Q^{-3/2}R^{3/4} \times (t - t_0)^{1/4} \qquad (16)$$

while integration of the ascent speed (16) implies that the height of rise of the center of the head increases as $z \sim (t - t_0)^{5/4}$.

In terms of the position of the diapir its volume increases as $V \sim z^{9/5}$, its diameter as $d \sim z^{3/5}$ and its speed as $U \sim z^{1/5}$. The average temperature anomaly presented by the head of the plume at large times decreases simply as the inverse of plume height:

$$\Delta T/\Delta T_s \approx (27/80)(6/\pi)^{2/3}C^{-2}(Q/\kappa)z^{-1} \qquad (17)$$

Note that the temperature anomaly is a direct measure of the dilution of the source material by entrainment. Remarkably, the amount of dilution and cooling predicted by (17) after the head has ascended through a layer of given depth is independent of the mantle viscosity: larger viscosities slow the plume ascent, leading to greater entrainment through conduction of heat but also a correspondingly greater input from the source when measured as a function of plume position.

Stirring during the ascent of the plume leads to a regular pattern of zonation of any compositional differences that exist between the source and the surrounding material. The zonation in the upper portions of the head, especially after lateral spreading beneath the surface, consists of a number of quasi-horizontal laminae alternating between source and entrained ambient material (Fig. 1). Apart from a slightly higher temperature near the core of the doughnut (A on Fig. 1c), these laminae effectively will be at the same temperature. The thickness δ of the laminae of source material, as defined in Fig. 3, is determined by a balance of the source flux Q (supplied by the feeder stem to the stagnation point at the leading edge of the diapir) and the rate of recirculation within the diapir. The circulation velocity relative to the centre of the head is $U/2$, giving

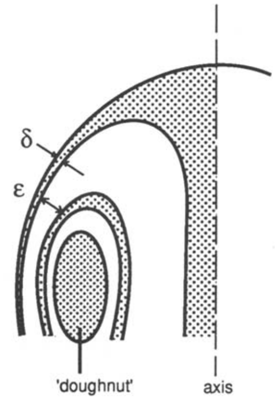

Fig. 3. A sketch of a section through the internal structure of a plume head showing the definitions of length scales used in the text. The "doughnut" contains the material supplied from the source during the earliest development of the plume.

$$\delta \approx 2Q/\pi\, dU \qquad (18)$$

The corresponding thickness of the entrained layers at large times can be estimated from the ratio of volumes:

$$e \approx \delta(V - V_s)/V_s \qquad (19)$$

The asymptotic solution for $t \gg t_c$ (14, 16) gives $\delta \sim t^{-1} \sim z^{-4/5}$ and $\epsilon \sim t^{1/4} \sim z^{1/5}$. Thus the spiral lamina of source composition becomes thinner with time while the regions of entrained material become thicker.

The model and asymptotic solution successfully predicts the behaviour of laboratory starting plumes described above, particularly an early period of accumulation of buoyancy with very little diapiric ascent, the growth of a very large head (which is dominated by entrainment at large times) and the slow increase in ascent speed at large times. The thickness of laminae within the head can also be compared. For the experiment pictured in Fig. 1 the volume flux, diameter and speed measured when the head was centred 40 cm from the source (Fig. 1c) give, from (18), $\delta \approx 0.18$ cm (or $\delta/d \approx 1/38$). The corresponding thickness from (19) (but taking into account the volume of the conduit) is $\epsilon \approx 0.54$ cm (or $\epsilon/\delta \approx 3.0$). Both values are in satisfactory agreement (within 30%) with the measured thicknesses.

4. Starting plume solutions under mantle conditions

The large effects of entrainment observed at the source fluxes and fluid properties used in the experiments do not immediately carry over to conditions appropriate for the mantle. It is essential to make quantitative estimates for the behaviour of mantle plumes before drawing conclusions about the role of entrainment in that case. This requires an evaluation of a number of mantle parameters and numerical solution of (10). We adopt commonly used values for material properties in the deep mantle: $\kappa = 10^{-6}$ m^2 s^{-1}, $\alpha = 2 \times 10^{-5}$ K^{-1}, $\rho_\infty = 4 \times 10^3$ kg m^{-3} and $c = 750$ J kg^{-1} K^{-1}. For the average viscosity of the modern *lower* mantle we use $\mu_\infty = 10^{22}$ Pa s ($\nu_\infty = 2.5 \times 10^{18}$ m^2 s^{-1}) [24]. However, if the temperature of the Archaean mantle was of the order of 100 °C hotter than it is today [25], the viscosity is likely to have been significantly smaller. Hence the implications of a viscosity for the lower mantle as small as $\mu_\infty = 10^{21}$ Pa s are also explored. The latter value will also serve to estimate plume behaviour in the modern upper mantle [24].

Only two parameters remain: the source temperature anomaly and the source volume flux. If the source is a thermal boundary layer at the core–mantle boundary (CMB) the temperature anomaly of interest is dependent upon the temperature difference between the core and the bulk of the lower mantle. This difference has been estimated from thermodynamic models of the earth but remains poorly constrained. From seismic data for the D″ layer and an assumption that it contains no compositional density differences, Stacey and Loper [23] predicted a drop of 860 °C across this layer. They also show that continued withdrawal will occur only from a layer of highest temperature and lowest viscosity, a layer much thinner than the thermal boundary layer. On this basis the source material entering the base of the plume may be close to 700 or 800 °C hotter than the mantle through which it will pass. However, this result assumes there is no compositional stratification within the D″ layer. If there exists even a very thin compositionally stabilised interfacial layer in either the mantle or the outer core, it would insulate the unstable portions of the thermal boundary layer and could readily lead to plume temperature anomalies only a small fraction of the difference between core and mantle. If a lower part of the D″ layer is stabilised, plume anomalies may be much less than 800 °C. On the other hand, potential temperature anomalies of at least 200 °C (or 450 °C) are delivered to the top of the modern (or Archaean) mantle by plumes. Thus we will consider source anomalies in the range $\Delta T_s = 200$–800 °C.

The volume flux Q can now be derived from the plume buoyancy flux Q_B according to (3). Davies [26] and Sleep [27] have evaluated the buoyancy fluxes (or more precisely, the mass anomaly fluxes) for a large number of currently "active" hotspots using the rate of creation of seafloor topography and obtain values in the range $0.3 \leq Q_B/g \leq 8$ Mg s^{-1}. These become $3 \times 10^3 \leq Q_B \leq 8 \times 10^4$ N s^{-1}. The largest flux is that for the Hawaiian hotspot, and is almost three times greater than the next largest. The smallest are inferred from seafloor swells with heights of only about 300 m. The corresponding volume fluxes are $20 \leq Q \leq 500$ m^3 s^{-1} if $\Delta T_s = 200$ °C or $5 \leq Q \leq 125$ m^3 s^{-1} if $\Delta T_s = 800$ °C. From (3a) the Hawaiian flux also corresponds to a heat flux $Q_H = 3 \times 10^{11}$ W. There are no constraints on the fluxes in plumes that may have been active prior to 200-Myr ago. Hence we consider a range of fluxes only slightly broader than that inferred for identifiable present day hotspots. Note that a key prediction of the model is that, as a result of entrainment, temperature anomalies are smaller and mass fluxes larger in the upper mantle than at the source (for a prescribed buoyancy or heat flux).

Solutions of (4, 5, 10) (Fig. 4) show that starting plumes having the largest volume flux and largest temperature anomaly (i.e. the largest buoyancy flux) ascend the most rapidly, accumulate the greatest volume and are least diluted by entrainment. At the largest buoyancy flux considered (10^5 N s^{-1}), heads can penetrate the depth of the modern mantle in as little as 50 Ma, or the depth of the Archaean mantle in less than 10 Ma. These plume heads will not be deflected far from their vertical ascent by large scale overturning of the mantle associated with plate motion and could remain connected to their source by a stable feeder conduit [13,14,20]. Strong plumes are therefore likely to give rise to dramatic events such as the formation of flood basalts and Archaean

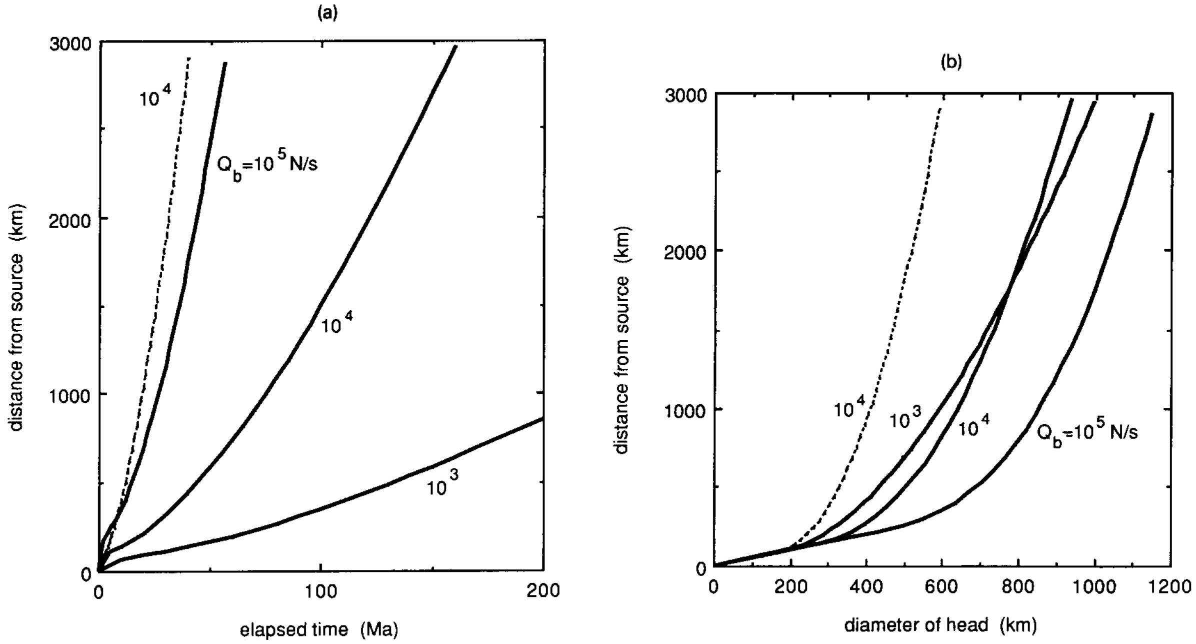

Fig. 4. The height of rise (a) and diameter (b) of plume heads having a source anomaly $\Delta T_s = 400\,^\circ$C in a mantle of viscosity 10^{22} Pa s (solid lines) or 10^{21} Pa s (dashed). The higher viscosity may describe the modern lower mantle; the lower viscosity is most appropriate for the Archaean lower mantle or the modern upper mantle. Plumes with small buoyancy fluxes ($Q_B < O\,(10^4$ N s^{-1})) are unlikely to reach the surface before breaking up into a sequence of diapirs in the presence of plate motion.

greenstone belts. Weaker plumes fed by buoyancy fluxes less than $\sim 10^4$ N s^{-1} are unable to approach the surface within a time of order 10^8 years. These will tend to be deflected by large scale overturning of the mantle [20] to such an extent that the conduit may well become unstable [13,14], disconnecting the plume head from its source long before it reaches the upper mantle. Thus weak plumes may break up into a sequence of smaller diapirs instead of ascending as one large head and, if they reach the lithosphere at all, there may be no major event at the beginning of the consequent hotspot track. Still weaker plumes are greatly cooled by entrainment and will never approach the surface. This predicted cut-off in the strength of observable plumes is consistent with the smallest buoyancy fluxes (300 N s^{-1}) inferred for hotspots [27].

As demonstrated by experiments, plume heads grow large before departing their source region. As can be seen in Fig. 4b, our model predicts that modern mantle plumes must reach 400–600 km in

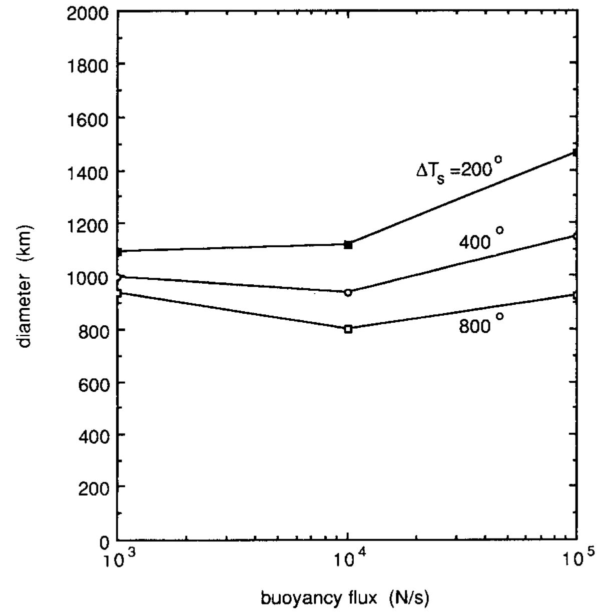

Fig. 5. The predicted diameters of starting plumes from the CMB before they begin to spread beneath the lithosphere. Mantle viscosity 10^{22} Pa s.

diameter before commencing to rise. A further doubling of the diameter (an order of magnitude in volume) is achieved if they ascend through 2800 km. Thus plume heads that reach the lithosphere while still receiving a constant influx from their source region are predicted to be extremely large: 800–1200 km in diameter in the modern mantle and 600–800 km in a warmer Archaean mantle. In Fig. 5 it is shown that the diameters are almost independent of the buoyancy flux. These dimensions imply that for plumes from the CMB it is the viscosity of the lower mantle only that concerns us: the head will be influenced by a smaller upper mantle viscosity only in its final stages of ascent and when spreading beneath the lithosphere. Furthermore, on reaching the top of the mantle the head of a plume from the CMB will predominantly consist of entrained *lower* mantle material. Our calculations also indicate that plumes originating at the 650 km transition zone rather than the CMB, and rising through an upper mantle having a viscosity of 10^{21} Pa s, will reach diameters of only 300 km on ascending 650 km. Indeed, they will begin to spread beneath the lithosphere almost as soon as they leave their source. In that case little entrainment will occur.

Additional laboratory experiments (to be reported elsewhere) show that when plume heads flatten and spread beneath a rigid surface they approximately double their diameter before vertical and horizontal velocities become extremely small compared with the initial ascent speed. Hence a spherical head 1000 km in diameter should lead to a pancake thermal anomaly about 2000 km across.

Despite the enormous size of the heads, stirring is effective in shifting compositional heterogeneity to relatively small scales. The model predicts laminae thicknesses (18, 19) of $\delta \approx 5$–10 km for the source material and $\epsilon \approx 20$–40 km for the entrained mantle once the head has risen 2800 km. If the head doubles its diameter by spreading beneath the lithosphere, these layer thicknesses will have decreased to 1–3 km and 1–5 km, respectively. We believe the ratio of $\delta/\epsilon \approx 0.25$ has important implications for the geochemistry of continental flood basalts [6].

The mean temperature in the head decreases most rapidly soon after the head moves away from the source. For example, for a buoyancy flux of

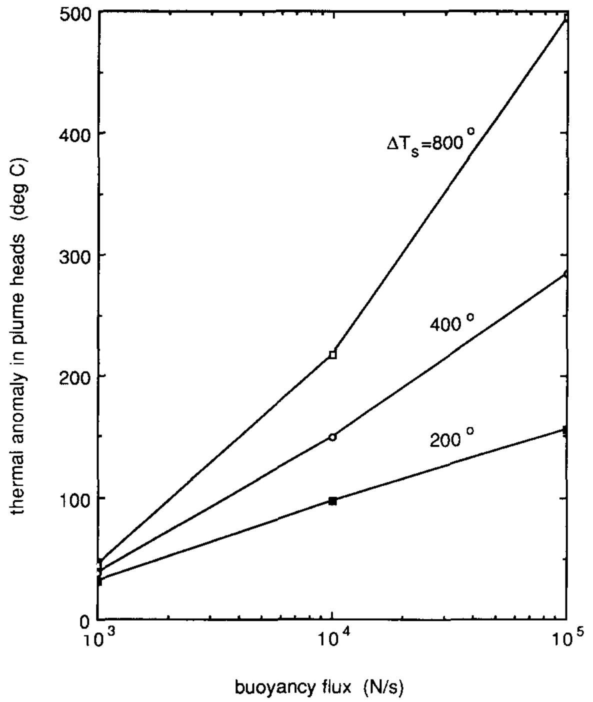

Fig. 6. The mean temperature anomaly in starting plume heads as they reach the top of the mantle. Most known hotspot tracks give buoyancy fluxes of order 10^4 N s^{-1}.

10^4 N s^{-1} an initial anomaly of 400°C falls to 250°C when the head is centred at 700 km from the source (assuming there is no upper boundary at that level) and to 150°C at 2800 km (see Campbell et al. [5] for plots of $\Delta T(z)$). The temperature anomalies induced in the upper mantle (Fig. 6) vary strongly with both buoyancy flux and source temperature. Recall that the degree of cooling (and chemical dilution) does not depend upon mantle viscosity. The predictions do, however, have uncertainties corresponding to those in the empirical entrainment constant C and in physical properties such as the thermal diffusivity. We estimate that our calculated rates of entrainment and dilution factors are reliable only to within a factor of two.

It was on the basis of the cooling of the head, compared with relatively little cooling in the supply conduit, that Campbell et al. [5] proposed a model for production of Archaean komatiites in close association with vast volumes of lower-temperature basalts: komatiites could form by melting of the hot, uncontaminated axis of the plume and the basalts from melting of the cooler plume head.

From the MgO contents it was inferred that the komatiite melts had potential temperatures of approximately 1875 ° C, or 600 ° C greater than that of modern normal (MORB) mantle, while the more voluminous basaltic magmas were generated at temperatures approximately 150 ° C above the modern MORB source. The potential temperature of normal Archaean mantle may also have been ~ 50–100 ° C hotter than the MORB source [25], making the mean temperature of the mantle in the plume head 50 ° to 100 ° C greater than that of normal Archaean mantle. Our model implies that the potential temperature of the high-temperature melts is almost as large as that of the plume source. We therefore require an Archaean plume source temperature anomaly $\Delta T_s \approx 500 \pm 100$ ° C. Given the uncertainties a mantle potential temperature for the plume head 50–100 ° C above normal Archaean mantle is then entirely consistent with the probable range predicted by the model for buoyancy fluxes of order 10^4 N s^{-1} or smaller (see Fig. 6).

In the case of modern flood basalts there is a smaller difference in potential temperatures between the mantle sources for basalts (1300–1400 ° C) and picrites (1500–1600 ° C), and the ambient temperature is well constrained (by MORB composition) to be 1280 ° C. The absence of high-temperature komatiites may be a result of a generally cooler modern mantle. However, the modern plume sources may also have had smaller temperature anomalies than the plumes that gave rise to Archaean greenstone belts. Our starting plume model requires modern plume source temperature anomalies $\Delta T_s \approx 300 \pm 100$ ° C, for which the entrainment calculations (Fig. 6) predict plume head temperature anomalies at the top of the mantle of the order of 50 ° C for the smallest "observed" buoyancy fluxes and 250 ° C for the largest (Hawaiian) flux. These results cover the range of melt temperatures noted above.

Although our concern in this paper is to demonstrate a general principle rather than attempt precise predictions, we note that higher source temperatures can be accommodated by the plume model without increasing the temperatures of the axial stem or head. So far we have assumed no heat loss or contamination along the feeder stem. Contamination of the conduit is predicted to be negligible when it is vertical [28] but becomes possible once the plume becomes deflected by large-scale shear [20,22]. Contamination of the conduit will reduce the average temperature of material rising at the plume axis and hence the maximum melt temperature, without reducing the heat flux supplied to the head. There will be little effect on the other model predictions apart from a small increase in the volume of entrained material in the head. Assuming a small conduit tilt the starting plume model suggests plume source anomalies no greater than 600 ° C in the Archaean and 400 ° C in the modern mantle. These values are somewhat less than Stacey and Loper's [23] estimate of the thermal anomaly in the D″ boundary layer above the core, but we believe the approach we have used to obtain the temperature of the source layer is more direct than that used by Loper and Stacey.

5. Conclusions

Our first aim has been to point out that in laboratory models of mantle starting plumes a large amount of surrounding fluid can be stirred into the plume as it ascends. Calculations indicate that similar entrainment is unavoidable in mantle plumes if they ascend sufficiently far from their source region. Such stirring is the normal consequence of the coupling between conduction of heat and advection driven by thermal buoyancy. This coupling ensures that mixing of heat between source and entrained materials is almost complete, reducing the temperature of the plume as it ascends. Any compositional differences, however, are only stirred, not mixed, due to the negligible rate of chemical diffusion and lead to chemical zonation within the hot plume. We suggest that an effective compositional mixing may occur later, in melts derived from the plume, if ascent of the plume carries the zoned material through its solidus. Under these conditions melt may be extracted from a range of depths and the composition of the extruded melt will vary, reflecting the relative melt volumes extracted from each of the compositional zones. The implications of these stirring and mixing mechanisms for the geochemistry of continental flood basalts are discussed by Campbell and Griffiths [6].

Second, we have shown that a simple parameterisation of entrainment can be used to predict the properties of mantle plumes as they approach the top of the mantle. The effects of stirring in the mantle are predicted to be major: they will determine the size, velocity, temperature and composition of the heads of starting plumes. If a plume originates at the core–mantle boundary, the heads are likely to have diameters close to 1000 km when they reach the upper mantle. These will collapse rapidly beneath the lithosphere to produce large thermal anomalies 2000 km or more across. Plumes originating at the core–mantle boundary can therefore account for the lateral extent of volcanism and uplift (2000–2500 km) associated with continental flood volcanism [6]. In contrast, plume heads originating from the base of the upper mantle are predicted to reach diameters of only 300 km and to entrain relatively little overlying mantle before being halted by the lithosphere. On spreading these cannot achieve the lateral dimensions observed for flood volcanism. This supports the hypothesis that plumes responsible for flood basalts originate at the core–mantle boundary.

Finally, the amount of entrainment in plume heads, judged from the magma compositions, can also be used to identify the source of mantle plumes. Plume heads originating at the upper mantle–lower mantle boundary entrain little of the overlying mantle as they rise and consist dominantly of hotspot source (OIB-type mantle) material. In this respect they contrast with plume heads originating from the CMB, which contain large temperature differences and a high percentage of entrained mantle. We believe that the success of the starting plume model in explaining the mixed trace-element geochemistry of flood basalts and the association of high- and low-temperature melts in continental flood basalts [6] and Archaean greenstone belts [5] provides further supporting evidence for the hypothesis that those plumes originate near the CMB.

Acknowledgements

We wish to thank Geoff Davies, Robert Hill and Mark Richards for stimulating discussions and for their collaboration in related work.

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