Tails of two plume types in one mantle

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ABSTRACT

Observations related to flood basalts suggest the existence of mantle plumes with large heads and thin trailing tails (cavity plumes). Seismic data suggest the existence of mantle plumes with thick tails (diapir plumes). The conditions required for diapir versus cavity plume generation are different, and in a chemically homogeneous mantle both types are not predicted to coexist. We show, however, that if a variable thickness chemical layer exists at the base of the mantle, consistent with seismic observations, then the coexistence of morphologically distinct plumes types is expected. The chemical layer governs temperature, and thus viscosity variations, in the thermal boundary layer from which mantle plumes rise. A locally thick layer leads to small viscosity variation instabilities and hence to diapir plumes. A locally thin chemical layer allows for large viscosity variations across the active portion of the lower mantle thermal boundary layer and, hence, for cavity plume formation. A chemical layer that can move in response to changing flow patterns allows for the potential that plumes can morphologically transition over their lifetimes. An expectation that the morphology and thermal structure of mantle plumes should vary according to the thickness of a chemical layer is consistent with correlations between seismic observations of chemically distinct material at the core-mantle boundary, the varied morphology of mantle thermal anomalies, and the inferred diversity in hotspot buoyancy fluxes and excess temperatures.

INTRODUCTION

Tomographic studies suggest that mantle plumes inferred to underlie many of the Earth's hotspots have diverse shapes in the lower mantle. Whereas "fat" low-velocity anomalies underlie weak hotspots such as the Azores, "thin" axisymmetric anomalies underlie stronger hotspots such as Tahiti and Louisville, and no seismic anomaly is observed beneath Hawaii (e.g., Montelli et al., 2006). The morphology of mantle plumes is governed by the magnitude of the viscosity variations within the unstable portion of the thermal boundary layer from which they ascend (cf. Olson and Singer, 1985). Where the viscosity of the plume is much less than the viscosity of the background mantle, such an upwelling takes the form of a cavity plume with a large spherical head and a thin trailing conduit. The thin conduit reflects the thickness of the lowest viscosity active part of the thermal boundary layer (i.e., the velocity boundary layer), which feeds the upwelling plume. For a cavity plume to form, this velocity boundary layer must be thinner than the thermal boundary layer from which the upwelling ascends (Stacey and Loper, 1983). This condition is met where vertical temperature gradients lead to viscosity variations on the order of 10 or larger (Thayalan et al., 2006). In contrast, where viscosity variations in the unstable portion of the hot boundary layer are small (i.e., on the order of 1), a diapir plume will form, characterized by a cylindrical stem with a diameter twice the thickness of the thermal boundary layer capped by a head only slightly larger in radius than the stem (cf. Yuen and Schubert, 1976).

The association of mantle plumes with midplate volcanism goes back to the 1970s (e.g., Morgan, 1971). Following the recognition that high excess temperatures are required to explain the eruption rate and composition of many hotspots and large igneous provinces (Davies, 1988; Sleep, 1990; Schilling, 1991), the mantle plume–hotspot connection was refined through extensive fluid dynamical experiments (e.g., Griffiths and Campbell, 1990, 1991) and numerical studies (e.g., Sleep et al., 1988; Olson et al., 1993). In particular, the occurrence of flood basalts and their associated hotspot tracks were thought to indicate that cavity plumes better fit observational constraints on hotspot origin than do diapir plumes (e.g., Duncan and Richards, 1991; Davies, 1990; Campbell and Griffiths, 1990; Loper, 1991).

The expectation of thin plume tails made their lack of seismic detection initially unsurprising (Nataf, 2000). More recently, seismic detection of some broad plume tails has been reported (Montelli et al., 2006; Boschi et al., 2007). To account for these new observations, plume theories have been refined once again to explain fat plumes (e.g., Korenaga, 2005). At the same time, debate continues as to whether any hotspots are associated with mantle plumes (e.g., Foulger et al., 2005). To those who disagree with the association of hotspots and mantle plumes, for well-established plume theories to be adjusted such that diapir plumes are now what are required seems arbitrary in the sense that all the observational support for the existence of cavity plumes seems to now be shelved. We take the view that the main issue for plume theorists is whether cavity and diapir plumes can both dynamically exist within the mantle of the Earth.

A key for understanding the dynamics of plume formation in the Earth is recognizing that the core-mantle boundary region is chemically heterogeneous, particularly where mantle plumes are thought to originate (Fig. 1) (e.g., Jellinek and Manga, 2004). The idea that a chemical layer at the base of the mantle can absorb a portion of the temperature drop from the core to the mantle is not a new one (Sleep, 1988; Nakagawa and Tackley, 2004). Seismic evidence suggests that a basal chemical layer is highly variable in thickness (e.g., Lay and Garnero, 2004). The association of chemical layer location to subduction zones further suggests that the basal layer may not be static, but can be displaced by changing mantle flow patterns (McNamara and Zhong, 2004). A variable thickness chemical layer can lead to a range of plume shapes (Farnetani and Samuel, 2005). By considering the effects of strongly temperature dependent viscosity, we take this conclusion one step further to argue that a variable thickness chemical layer can allow for the coexistence of dynamically distinct plumes types. We show that a varying basal chemical layer thickness leads to temperature and viscosity variations that permit the coexistence of cavity and diapir plumes in an otherwise thermally well stirred mantle. Moreover, we argue that such a picture explains the disparate plume excess temperatures and buoyancy fluxes inferred for many of the Earth's hotspots.

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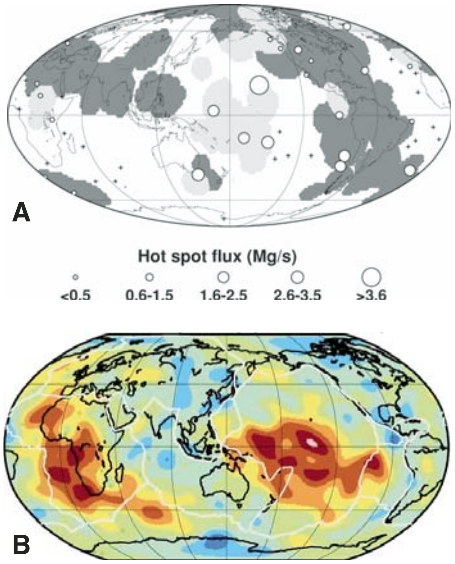

Figure 1. A: Hotspots (open circles) and ultralow-velocity zone material (ULVZ—light gray regions) at core-mantle boundary. Dark gray regions indicate where ULVZ is <5 km thick (Williams et al., 1998). Diameters of open circles correspond to plume buoyancy fluxes estimated from dynamic topography (Davies, 1988; Sleep, 1990). B: Global shear velocity perturbations at core-mantle boundary (Ritsema et al., 2004). Red indicates large regions of low shear velocity interpreted to indicate variably thick, laterally extensive layers of dense, chemically distinct material (Garnero and McNamara, 2008).

NUMERICAL SIMULATIONS

We consider mantle convection with a temperature-dependent viscosity. On its own, a temperature-dependent rheology cannot generate cavity plumes. Strongly temperature dependent viscosity leads to stagnant lid convection in which the surface layer absorbs the majority temperature drop across the system. This leads to a small temperature, hence viscosity, drop across the lower thermal boundary layer, which favors diapir plume formation (Nataf, 1991). For a planet with plate tectonics, this does not rule out the existence of cavity plumes (Lenardic and Kaula, 1994; Jellinek et al., 2002). Subduction cools the interior mantle, which allows for an enhanced temperature drop across the lower boundary layer and for cavity plume formation. Stirring of the upper boundary layer into the interior mantle, akin to subduction, can be incorporated into our simulations by imposing a fixed surface velocity or by employing a visco-plastic rheology (Moresi and Solomatov, 1998). We consider both approaches.

Along with temperature-dependent viscosity and the allowance for subduction, the final element of our simulations is a variable thickness, thermally conducting basal layer. This layer is an analog for chemical piles inferred to occur at the core-mantle boundary. In our simulations we fix the thermal conductivity of the layer to be the mantle value. More complex models could consider, for example, an effective thermal resistance that might depend also on whether the layer is partially molten (Lay et al., 2004; Labrosse et al., 2007) or convecting. In detail, such thermal effects can be parameterized in terms of an effective thermal thickness and compared directly to our results. Our principal objective is, however, to show only that a variable thermal resistance layer can allow for the coexistence of diapir and cavity plumes. To show this most clearly we have designed the models only to include the necessary complexities. The neglect of added complexities (e.g., depth-dependent material properties) does, however, set limits in terms of modeling validity. Specifically, the quantitative chemical thickness variations needed for plume transitions will be effected by these added complexities.

Figure 2 shows results from a representative simulation from a modeling suite. A finite element code (Moresi and Solomatov, 1998) is used to solve the equations of mass, momentum, energy, and composition conservation for an infinite Prandtl number and Boussinesq convection (e.g., Lenardic and Kaula, 1996). A basal chemical layer is imposed over one-half of the modeling domain. Subduction is achieved by prescribing horizontal surface velocity at the top the domain. The imposed velocity is always less than the root mean square (rms) system velocity for equivalent parameter condition, stagnant lid cases. For the simulation shown, 128×128 finite elements cover any 1×1 patch of the domain. The bottom heated Rayleigh number, based on the viscosity of the mantle at the system base, is 10^8. The activation temperature is set to provide a six order of magnitude viscosity drop across the mantle.

In Figure 2, the non-dimensional interior temperature is 0.56. (for all simulations, temperature is non-dimensionalized by the total temperature variation across the mantle). Where no basal chemical layer is present, the associated temperature increase across the basal thermal boundary layer leads to a factor of 437 viscosity variation, which allows cavity plumes to form. In more detail, provided the internal temperature corresponds to viscosity variations in the hot thermal boundary layer of a factor >30–40, cavity plumes can locally form. This was determined by varying the surface velocity for a large suite of simulations (a lower surface velocity led to a hotter interior mantle). The basal viscosity variation in the absence of a chemical layer is determined by the internal temperature and the total viscosity variation across the system, i.e., the activation temperature. In the presence of a chemical layer, however, the temperature variations within the unstable thermal boundary layer are modulated, depending on the thickness of the layer. In Figure 2 the chemical pile absorbs a significant

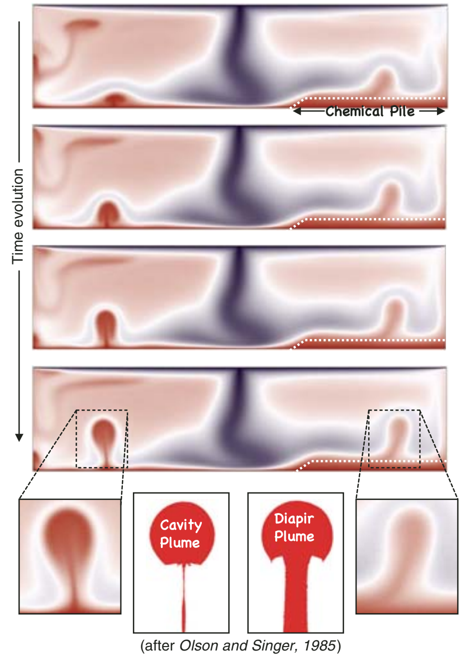

Figure 2. Thermal field evolution from numerical simulation with fixed, variable thickness basal chemical layer. Thickness of chemical layer over the right portion of this simulation is 0.1 of full mantle depth. Viscosity variation between active thermal upwellings and mantle is on the order of 100 in regions free of the chemical layer and on the order of 10 or less for those that form above the layer. Enlargements at the bottom, along with insets (from Olson and Singer, 1985), show how this allows for morphologically distinct upwellings to coexist.

portion of the total temperature drop from the system base to the interior, leading to relatively small temperature, hence viscosity, variations across the active portion of the thermal boundary layer above the pile. Quantitatively, the viscosity variation across the active lower thermal boundary above the chemical pile is less than a factor of 10 and diapir plumes form.

Figure 3 shows a simulation from a second suite. Rather than an imposed surface velocity, a visco-plastic rheology is employed (Moresi and Solomatov, 1998), which permits weak zones to form in regions of high stress. The weak zones are model analogs for weak plate boundaries. They allow the otherwise cold, and therefore high-viscosity, upper boundary layer to partake in convective overturn and cool the interior mantle. Figure 3 shows a reference simulation with no chemical layer. A suite of models is run in which a variable thickness chemical layer is imposed at the base of the reference model. The basal Rayleigh number is 10^8. The activation temperature is set to provide a seven order of magnitude viscosity drop across the mantle.

The transition between cavity and diapir structures is quantified by tracking the thickness of the basal thermal and velocity boundary

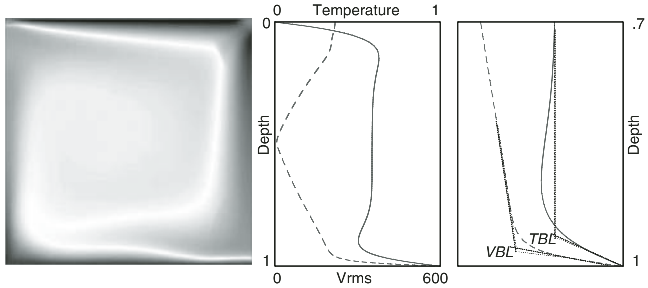

Figure 3. Thermal field from a simulation with visco-plastic rheology (far left). Medium gray tones in upper corners show high-strain-rate regions that are associated with dynamically generated weak zones. Middle graph shows horizontally averaged thermal profile and root mean square (rms) velocity profile across entire model domain. Right graph shows the profiles near the base of the model domain. vbl—velocity boundary layer; tbl—thermal boundary layer; U— horizontal velocity.

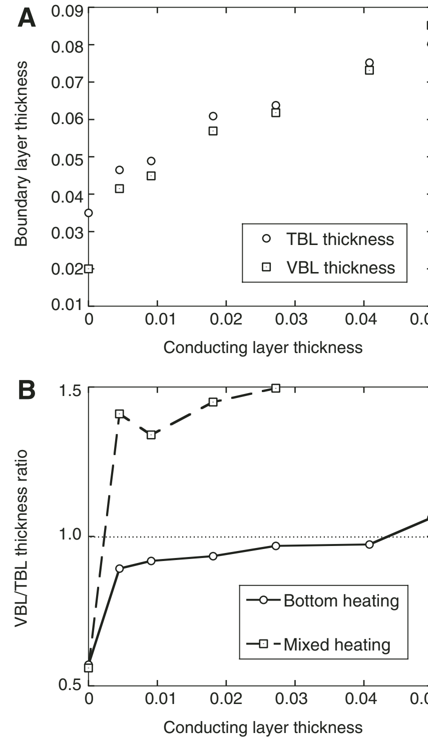

Figure 4. A: Velocity and thermal boundary layer (vbl, tbl) thickness from models of the type in Figure 3. B: Velocity to conducting layer thickness divided by mantle thickness for two simulation suites. For mixed heating suite, internal to basal heating Rayleigh number ratio is 1. With no chemical layer, this leads to a non-dimensional internal mantle temperature of 0.703 and factor of 120 viscosity variation across lower thermal boundary. Suite with heating ratio of 2 also allowed for transition (non-dimensional internal mantle temperature of 0.762 and factor of 46 viscosity variation for the reference case). Greater degrees of internal heating require larger activation temperatures to allow for transition.

layers (Fig. 3). The boundary layers are associated with the diffusion of heat and momentum, respectively, from the system base into the interior, and transition from diapir to cavity plume structure occurs when the velocity layer becomes embedded within the thermal boundary layer (Stacey and Loper, 1983; Thayalan et al., 2006). For convergence testing, the numerical resolution was varied from 64 × 64 to 256 × 256 finite elements over a 1 × 1 domain (with three different mesh densities between). The thinness of the velocity layer drove the need for high-density meshes. For simulations with a chemical layer, the active portions of the boundary layers are tracked (i.e., the portions above the top of the chemical layer).

Figure 4A shows boundary layer thicknesses from a suite of bottom-heated simulations. For the case with no chemical layer, the non-dimensional mean internal mantle temperature was 0.588, which leads to a factor of 766 viscosity variation across the lower thermal boundary layer. The velocity boundary layer was half the thickness of the thermal layer. As the chemical layer thickness was increased, the temperature and, hence, viscosity, drop across the active portion of the basal boundary layer decreased, leading to a transition from cavity to diapir plume structure. The transition occurs at a larger chemical layer thickness than it does for equivalent parameter suites that impose subduction via a fixed surface velocity, because for visco-plastic suites, the surface velocity decreases with increasing chemical layer thickness. This occurs because the chemical layer lowers the temperature drop across the mantle, which reduces the total buoyancy force available to drive convective motions. Despite this added complexity, the general conclusion that increased chemical layer thickness can cause a transition from cavity to diapir plumes is robust.

In Figure 4B we plot the ratio of the velocity boundary layer thickness to the thermal boundary layer thickness. A transition from cavity to diapir plumes will occur where this ratio is >1. Figure 4B also shows that our general conclusion holds if internal heating is included. The main consequence of internal heating is to increase the background mantle temperature, and thus the key issue is the extent to which this effect reduces the temperature and viscosity variations in the hot thermal boundary layer. Compared with the basally heated cases, a transition from cavity to diapir plumes in the presence of internal heating occurs for a thinner chemical layer because of the reduced viscosity variations.

DISCUSSION AND CONCLUSIONS

The principal message from our results is that a variable thickness chemical layer at the base of the Earth's mantle can lead to the coexistence of morphologically distinct plume types. Many Pacific and African hotspots are correlated with a dense layer at the core-mantle boundary (Fig. 1). Although global resolution of this layer is data limited, it appears to be characterized by significant lateral thickness variations. Our results show that such a nonuniform thickness will have a profound effect on the structure and heat-transfer properties of proposed mantle plumes. Putirka (2008) estimated excess temperatures for the mantle plumes associated with 28 hotspots using an olivine-liquid geothermometer, and found significant variability that correlates with inferred plume buoyancy fluxes indicated in Figure 1. Whereas excess temperature is ~300 °C for the strong Hawaii hotspot, excess temperatures are ~200 °C for Tahiti, Samoa, Kerguelen, Reunion, and Louisville, and closer to 100 °C for the weak Azores and Cape Verde hotspots.

Assuming that mantle viscosity decreases by an order of magnitude per 100 °C of excess temperature (Kohlstedt et al., 1995), the inferred range in plume excess temperatures implies an order of magnitude of three range of viscosity variations. In particular, the viscosity variations expected for the weaker hotspots are on the order of 10, less than the factor of 30–40 required for cavity plumes to form. Thus, consistent with existing plume theory, we expect weak hotspots to be associated with diapir plumes and strong hotspots to be associated with cavity plumes. Montelli et al. (2006) presented a catalog of deep mantle plumes as imaged using finite-frequency tomography. Consistent with our expectation for a diapir structure, weak hotspots with excess temperatures close to 100 °C such as Azores or Cape Verde are characterized by broad low S-wave and P-wave velocity anomalies at the core-mantle boundary and upper mantle. Hotspots characterized by excess temperatures of ~200 °C, including Tahiti, Samoa, Reunion, and Kerguelen, are, in contrast, characterized by relatively narrow, axisymmetric anomalies. Finally, for Hawaii, which has an excess temperature approaching 300 °C, a very narrow axisymmetric anomaly is observed in the upper mantle, and no S-wave or P-wave velocity anomaly is observed at the base of the mantle. Although this observation in the deep mantle may be related to data limits, an alternative interpretation is that the plume conduit diameter is below the detection limit, consistent with expectations for a plume conduit that is ~1000 times less viscous than the surrounding mantle. Resolution of a Hawaiian conduit in the upper mantle may be related simply to the spread of the thermal anomaly related to the cooling of upwelling mantle.

The correlations between chemical layer thickness and inferred plume size and strength are not perfect, nor do we expect them to be. Other factors such as subducting slabs and shear associated with background mantle flow will affect plume structures at the base of the mantle. What is key is that, to first order, our model-based expectations are consistent with observations related to several hotspots. This gives confidence that this modeling approach can provide the foundation for developing both a conceptual and a quantitative link between surface observations (e.g., hotspot temperatures) and deep Earth seismic observations.

In summary, our work shows that diverse plume morphologies are an expected natural consequence of the interaction between mantle convection in the Earth and a dense chemical layer inferred to occur at the core-mantle boundary. Such a picture is consistent with recent varied tomographic images of mantle plumes and with the reported ranges in hotspot buoyancy fluxes and plume excess temperatures. As a final remark, much of the current debate over the mantle plume hypothesis is stimulated by an apparent discord between the recent catalog of tomographic images of plumes by Montelli et al. (2006) and early plume-hotspot theories that argued for exclusively cavity plumes. Our work suggests that both pictures are easily reconciled if the chemically heterogeneous structure of the core-mantle boundary region is considered.

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