Implications of lower-mantle structural heterogeneity for existence and nature of whole-mantle plumes

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ABSTRACT

Recent seismological studies demonstrate the presence of strong deep-mantle elastic heterogeneity and anisotropy, consistent with a dynamic environment having chemical anomalies, phase changes, and partially molten material. The implications for deep-mantle plume genesis are discussed in the light of the seismological findings. Nearly antipodal large low–shear velocity provinces (LLSVPs) in the lowermost mantle beneath the Pacific Ocean and Africa are circumscribed by high-velocity regions that tend to underlie upper-mantle downwellings. The LLSVPs have sharp boundaries, low $V_S/V_P$ ratios, and high densities; thus, they appear to be chemically distinct structures. Elevated temperature in LLSVPs may result in partial melting, possibly accounting for the presence of ultra-low-velocity zones detected at the base of some regions of LLSVPs. Patterns in deep-mantle fast shear wave polarization directions within the LLSVP beneath the Pacific are consistent with strong lateral gradients in the flow direction. The thermal boundary layer at the base of the mantle is a likely location for thermal instabilities that form plumes, but geodynamical studies show that the distribution of upwellings is affected when piles of dense chemical heterogeneities are present. The location of lowermost mantle plume upwellings is predicted to be near the boundaries of the large thermochemical complexes comprising LLSVPs. These observations suggest that any large mantle plumes rising from the deep mantle that reach the surface are likely to be preferentially generated in regions of distinct mantle chemistry, with nonuniform spatial distribution. This hypothesis plausibly accounts for some attributes of major hotspot volcanism.

Keywords: core-mantle boundary, $D''$, plumes, thermochemical piles, hotspots

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INTRODUCTION

The depth of origin of the source of long-lived hotspot volcanism has been of great interest to geological scientists for decades (e.g., Morgan, 1971). This question intersects nearly all Earth science disciplines, and hence continues to attract active debate. The most common interpretation is that thermal plumes rise from an internal mantle thermal boundary layer and sustain hotspot activity. As long as heat is flowing into the base of the mantle from the core—an apparent requirement for long-term maintenance of the geodynamo (e.g., Buffett, 2002)—a thermal boundary layer should be present at the base of the lower mantle. This boundary layer is commonly invoked as the source of deep mantle plumes (see Lay, 2005), consistent with the early notions advanced by Morgan. Certainly, cylindrical plumes commonly initiate from the basal boundary layers in numerical and experimental convection experiments with basal heating or basal injection of fluid (e.g., Davies, 1990; Olson and Kincaid, 1991; Farnetani and Richards, 1994; van Keken, 1997; Farnetani and Samuel, 2005; Lin and van Keken, 2005), carrying heat and any unique isotopic signatures from the boundary layer to the surface. However, demonstrating that such plumes rise ~2900 km from the core-mantle boundary (CMB), traversing the Earth’s entire mantle, has proven challenging. Many discussions of this problem invoke very simple notions of the lower-mantle boundary layer, at odds with recent seismic findings. Our goal is to place the question of deep-mantle plume genesis in the context of current seismological and geodynamical ideas about lower-mantle structure and processes. We will avoid the issue of connecting specific hotspot observations to plumes or alternate explanations, as that is addressed in detail elsewhere in this volume (e.g., Sleep, this volume). Our focus is on the implications of seismically defined lower-mantle structures for the occurrence and characteristics of any plumes that do rise from the lowermost mantle.

A number of fields (e.g., seismology, geodynamics, and geochemistry) have presented arguments and some evidence for deep mantle plumes (e.g., Ji and Natalf, 1998; Lin and van Keken, 2005; Montelli et al., 2006; Wen, 2006), but the issue is still under debate, and an increasing number of studies find that some hotspots do not require origins in the lower mantle (e.g., Cserepes and Yuen, 2000; Foulger and Pearson, 2001; Foulger et al., 2001; Courtillot et al., 2003). In this article, we focus primarily on the elastic structure of the deep mantle derived by seismic methods and the dynamics of plume initiation, stability, fixity, and longevity in the presence of the large-scale chemical heterogeneity suggested by the seismic results, including the fact that hotspots are typically only found away from regions of subduction. Over the past several years, a variety of deep-mantle structural characteristics have emerged from high-resolution imaging with broadband seismic data. These structures include chemically distinct provinces beneath the Pacific Ocean and Africa, thin ultra-low-velocity zones (ULVZs) at the CMB, deep-mantle seismic wave anisotropy, and variable occurrence and topography of the D” seismic velocity discontinuity. This article considers these deep-mantle findings, exploring their implications for the possibility of lowermost mantle plume origination. Key seismological observations are summarized in the next section, followed by a section that considers the chemistry and dynamics of these features. This description provides a framework for considering the implications for any deep mantle plumes that may rise to the surface.

LOWERMOST MANTLE SEISMIC VELOCITY STRUCTURE

It has been known for decades that relatively high seismic velocities in the deep mantle tend to underlie past or present subduction zones, whereas lower-than-average wavespeeds are commonly found beneath the Pacific Ocean, the southern Atlantic Ocean, and Africa (e.g., Dziewonski, 1984; Hager et al., 1985). The distribution of lower-mantle velocities is quite consistent among recent tomographic S-wave velocity (Vp) models, but there is less consistency among P-wave velocity (Vs) models (Fig. 1). The differences between large-scale lower-mantle Vs and Vp heterogeneity have led to the inference that the origin of the velocity perturbations is not solely thermal (e.g., Masters et al., 2000). Unfortunately, until the Vs maps are better resolved (as indicated either by agreement between results from different studies or by demonstration that a particular study has produced the most robust results), it is difficult to confidently separate chemical and thermal effects based on the patterns of heterogeneity at the present time.

It has recently been demonstrated that the expected primary lower-mantle mineral—(Mg,Fe)SiO3, magnesium silicate in perovskite structure (Pv)—should undergo a phase transformation at pressure-temperature conditions within a few hundred kilometers above the CMB (Murakami et al., 2004; Oganov and Ono, 2004; Tsuchiya et al., 2004; Lay et al., 2005). Pv transforms into a post-perovskite structure (pPv) that is predicted to be accompanied by a Vs increase of several percent, but little change in Vp. This difference in response could be one cause of decoupling of variations in Vs and Vp in the lowermost mantle. The Vs/Vp ratio should be highest in high–shear velocity regions because the phase transition will occur at shallower depths in cool regions that have higher seismic velocities to begin with. One challenge in seeking this behavior is uncertainty in the reference level for measuring velocity anomalies; for example, the increase in temperature in the thermal boundary layer above the CMB will tend to lower seismic velocities, with more pronounced effects on S-wave velocity than for P-wave velocities. As seen in Figure 1, most seismic models tend to have means of zero at a given depth, which affects inferences about relative velocity behavior significantly.

The circum-Pacific band of high shear velocities apparent in Figure 1 is plausibly linked to occurrence of pPv in regions with relatively low temperatures below present-day and historic subduction zones. If this link is the case, the large low–shear ve-
locity provinces (LLSVPs) beneath Africa and the Pacific might have no pPv or only a very thin layer of it, and may be relatively warm. The Pv-pPv phase boundary has a large positive Clapeyron slope that would allow large lateral variations of thickness of a layer of pPv within the boundary layer to be caused by large-scale thermal variations. The LLSVPs are basically isolated from locations where subduction has occurred over the past 200 m.y., which is commonly invoked as an indication of control on the deep seismic heterogeneity by large-scale mid-mantle convection coupled to the shallow subduction history. The relatively low shear velocities in LLSVPs can thus be attributed to a combination of relatively high temperature and lack of pPv, but there are indications that there is also a chemical anomaly present in the LLSVPs.

Several free-oscillation studies have found evidence for lateral variation in large-scale lowermost-mantle density distribution (Ishii and Tromp, 1999, 2004; Kuo and Romanowicz, 2002; Trampert et al., 2004). Although debate continues on this topic (Romanowicz, 2001; Masters and Gubbins, 2003), indications are that a density increase is associated with the strongest V\textsubscript{S} reductions located in LLSVPs (e.g., Ishii and Tromp, 1999; Trampert et al., 2004). Simultaneous analysis of V\textsubscript{S} and V\textsubscript{P} behavior further suggests that LLSVPs have bulk sound velocity anomalies (increases) that are anticorrelated with the low–shear velocity anomalies (e.g., Masters et al., 2000). These observations suggest that LLSVPs are chemically distinct from the surrounding mantle. This possibility immediately complicates the interpretation of these regions, because chemical differences can also affect the occurrence of the pPv phase change, and some compositional effects, such as iron enrichment, tend to reduce shear velocity as much or more than high temperature does at high pressures. The presence of iron or aluminum can also affect the phase transition pressure and sharpness (see Lay et al., 2005), although the magnitude of such effects is being debated. Sorting out these tradeoffs requires more detailed structural information than provided by tomography alone.

Portions of LLSVPs have been characterized at relatively short scale lengths (e.g., study regions spanning 500–1000 km laterally) using forward modeling of body wave travel times and waveforms. For example, a significant number of LLSVP margins (Fig. 2) show strong evidence for an abrupt lateral transition over a few hundred kilometers or less between the LLSVP and surrounding mantle. The sharpness of the LLSVP margins supports the notion that there is a chemical contribution because thermal gradients should be more gradual. Additionally, weak reflections from a velocity decrease in the upper portion of the LLSVP beneath the Pacific may indicate a chemical boundary (Lay et al., 2006) or a phase boundary within chemically distinct LLSVP material (Ohta et al., 2007). Additional internal layering within the LLSVP beneath the Pacific has been inferred from seismic wavefield reflectivity resolved by stacking a large number of seismic data (Lay et al., 2006). In Lay et al. (2006), the northern portion of the LLSVP beneath the Pacific Ocean is found to have a sharp velocity increase overlying a sharp decrease that

Figure 1. Tomographically derived P-wave (left column) and S-wave (right column) velocity perturbations at the base of the mantle. Red and blue colors indicate lower and higher velocities than global averages, respectively. Color scales are not uniform for the different models; the peak-to-peak value is indicated in the lower right of each map (blue number). Model names are given in the upper right and correspond to the following studies: MK12WM13 (Su and Dziewonski, 1997), B10L18 (Masters et al., 2000), SPRD6 (Ishii and Tromp, 2004), KH00 (Kárason and van der Hilst, 2001), TXBW (Grand, 2002), BD00 (Becker and Boschi, 2002), S20RTS (Ritsema and van Heijst, 2000), Z01 (Zhao, 2001), S362D1 (Gu et al., 2001), HWE97 (van der Hilst et al., 1997), and SAW24B16 (Mégnin and Romanowicz, 2000). More information comparing many of these models is found in Becker and Boschi (2002). Hotspot locations (from Steinberger, 2000) are shown as red-filled circles. Plate boundaries are magenta lines, but convergent boundaries are shown in blue.
ULVZs appear to exist preferentially beneath low-hotspots are inferred to be underlain by low--shear velocity provinces (LLSVPs; thick black lines) using waveform and/or travel time analyses. Lower-cased letters in boxes indicate specific studies for different regions: a (He et al., 2006), b (Luo et al., 2001), c (Bréger and Romanowicz, 1998), d (To et al., 2005), e (Ford et al., 2006), f (Ni and Helmberger, 2001), g (Wen et al., 2001), h (Ni et al., 2002), i (Ni and Helmberger, 2003a,b), j (Wang and Wen, 2004). Solid lines indicate regions with sharp lateral boundaries, and dashed lines indicate regions where the LLSVP edges are only loosely resolved by travel time analysis.

Lowermost-mantle seismic wave anisotropy may also offer clues to deep-mantle chemistry and dynamics; as suggested by seismic studies (e.g., see Kendall and Silver, 1998; Lay et al., 1998; Kendall, 2000), mineral physics calculations (e.g., Stixrude, 1998; Karki et al., 1999; Mainprice et al., 2000; Wentzcovitch et al., 2004; Hirose, 2006; Hirose et al., 2006), as well as by geodynamics experiments (e.g., McNamara et al., 2001, 2002, 2003). Several seismological studies have mapped geographical changes in the fast propagation direction of deep-mantle shear waves (Russell et al., 1998; Garnero et al., 2004; Wookey et al., 2005; Rokosky et al., 2006). In one case, geometrical patterns in fast propagation directions have been interpreted as being related to lowermost-mantle boundary layer convective currents that may involve flow into a boundary layer upwelling, possibly related to a plume that rises to the Hawaiian hotspot (Russell et al., 1998). Given that there is a first-order correlation between the distribution of surface hotspots and the locations of the LLSVPs (e.g., Thorne et al., 2004; any correlation is less apparent for P-wave velocity heterogeneity) at the base of the mantle (Fig. 1), it is reasonable to seek any evidence for plumes extending through the mantle above these regions.

Direct seismic imaging of any deep mantle plumes is very difficult, primarily owing to the expected small dimension of the plume conduit (e.g., <500 km) compared to the long seismic wave propagation paths (typically >5000 km; see, e.g., Nataf, 2000; Dahlen, 2004). Most tomographic efforts have not directly imaged vertically continuous deep mantle plumes or their relationship to LLSVPs, as the minimum lateral wavelength of resolvability is commonly >1000 km. One notable exception is the study by Montelli et al. (2004), in which several surface hotspots are inferred to be underlain by low $V_p$ values extending down to the CMB (this observation is currently under active debate; see Dahlen and Nolet, 2005; de Hoop and van der Hilst, 2005). The seismological community may eventually converge on models that either support or refute the existence of whole-mantle low-velocity plume conduits. However, deep-mantle seismic plume detection may be almost impossible if plume temperature does not significantly exceed the surrounding mantle (e.g., Farnetani, 1997; Farnetani and Samuel, 2005), giving too small an elastic velocity signature. If the velocity perturbations are, in fact, strong enough, there is some hope to image plume features if wavepath coverage is dense enough (e.g.,
alyzed layer or sinking after a previous ascent. Like this material was either a strat-
ically distinct material in the dense dome. This entrainment may, from the tops of these large domes, entrain some of the chem-
and D). This modeling demonstrated that it is dynamically fea-
sential density anomaly. This increase to the dense material, so such a model is not favored
that the presence of long-lived, stable, dense piles of chemically distinct material (e.g., Christensen and Hofmann, 1994; Tackley 1998, 2002; Kellogg et al., 1999; Jellinek and Manga, 2002, 2004; Ni et al., 2002; McNamara and Zhong, 2004a,b, 2005; Nakagawa and Tackley, 2005; Tan and Gurnis, 2005). In these models, piles of dense material maintain a near-neutral (slightly negative) buoyancy, and as a result, they are passively swept aside by downwelling flow and are focused beneath upwelling regions. Piles tend to form large, ridge-like structures that have thermal plumes originating from their peaks that entrain a small fraction of the more-dense material.

McNamara and Zhong (2005) performed numerical thermochemical calculations in a 3-D spherical geometry with Earth’s recent plate history imposed as surface boundary conditions (120 m.y. over eleven stages of plate motions, as provided by Lithgow-Bertelloni and Richards 1998). The calculation employed the Boussinesq approximation with constant thermal properties; however, depth-dependent thermal conductivity was explored and found to have only a minimal effect on the resulting thermal and chemical structures. A depth- and temperature-dependent rheology that included a thirty-fold increase across the transition zone was used. The initial condition included a flat, 255-km-thick more-dense layer and a steady-state temperature field derived from an axisymmetric thermochemical calculation. The imposed plate history acted to guide the formation of downwellings in historical subduction regions, which resulted in the focusing of the lower-mantle dense material into piles beneath Africa and the Pacific. These are the same regions characterized by the observed LLSVPs (see Fig. 3A and D). This modeling demonstrated that it is dynamically feasible that global flow patterns derived from the history of sub-
duction can focus a dense component into thermochemical structures that, to first order, resemble the present-day LLSVP configuration.

Our preferred interpretation of LLSVPs is that they are large, dense thermochemical piles stabilized by upwelling cur-
ents that are downwelling-induced return flow. The temperature within and around the pile depends on several uncertain factors, like thermal conductivity and the degree of viscous heat-
is also expected to be denser than surrounding material at lower-mantle conditions and thus may account for the dense LLSVPs, assuming that MORB has accumulated progressively in the lowermost mantle (Hirose et al., 1999; Ohta et al., 2007).

The detailed structure (e.g., topography) of the sides and top of chemically distinct LLSVP material can play a significant role in the style and morphology of local upwelling currents and plume initiation (e.g., Jellinek and Manga, 2002, 2004; McNamara and Zhong, 2005). Numerical calculations show upward convective return flow guided by the LLSVP margins. Basal heating and internal flow of the LLSVP cause the boundaries between the surrounding mantle and LLSVP to be particularly hot (see Fig. 3B). If partial melt is indeed the origin of ULVZ structure, we would expect the edges of LLSVP structure to have the highest occurrence of ULVZ structure. As previously mentioned, the geographical distribution of ULVZ structure at present is not known in great enough detail to document such a spatial correlation. It is noteworthy, however, that two recent high-resolution studies detailing ULVZ structure are both near (and within) LLSVP margins: a double-array stacking study of ScS (a core-reflected S-wave) beneath the northern margin of the LLSVP beneath the Pacific Ocean (Avants et al., 2006; Lay et al., 2006), and a multiple vespagram analysis of ScP (an S-wave that converts to a P-wave upon reflection at the CMB) beneath the southwest margin of the same LLSVP (Rost et al., 2005, 2006). The strongest lateral gradients in tomographically derived $V_S$ structures are near the margins of the LLSVPs (consistent with pile “edges”), and these regions of strong gradients are found to statistically correlate with surface hotspot locations (Thorne et al., 2004).

These findings are consistent with the conceptual model put forth in Figures 3C and 4. Large thermochemical piles are deflected away from downwellings by subduction currents and are swept to concentrate beneath upwelling return flow. LLSVP piles may thus be the key to the long-term history of subduction. LLSVP topographical features near their margins guide upwelling and serve as sites of thermal boundary layer instabili-

Figure 3. Continents, plate boundaries, and hotspots are shown on maps at the top of four boxes that represent the area of the whole globe, and the volume of the mantle from the surface to the core-mantle boundary (CMB). (A) Compositionally distinct, dense piles from the geodynamic calculation of McNamara and Zhong (2005). (B) The locations of the hottest temperatures in the mantle for the calculation of panel A are shown. Isotemperature contour is 0.98 for the calculation that spans temperatures from 0 to 1. The hottest temperatures are within the piles and are typically near the edges. (C) A temperature cross-section is shown, along with the piles from panel A in a transparent gray, and the hottest CMB temperatures of panel B are included (faint red stripes within the piles). Pile topography guides plume upwellings. (D) Shear velocity heterogeneity from Ritsema and van Heijst (2000) filtered to maximum spherical harmonic degree $l = 8$, with iso-velocity contours at -0.3% (red) and 0.5% (blue). Dense piles in the geodynamic calculation of panel A are geographically distributed similarly to the low velocities (red) in panel D.
ties. Ascending plumes from the LLSVP margins may carry distinct chemical tracers from the deep mantle and CMB. OIB geochemistry for major hotspots favors recycled slab material as a significant source (e.g., Hofmann, 1997), which is consistent with past and/or ongoing subduction of slabs to the base of mantle and concentration of slab materials into the piles (e.g., Christensen and Yuen, 1984; Hager et al., 1985; Hutko et al., 2006). Thus, LLSVPs can be viewed as a by-product of whole-mantle convection, with physical segregation of dense material in the boundary layer. This process could occur today even if slabs temporarily go stagnant in the transition zone because of the difficulty of penetrating the 670-km phase boundary (Mitrovica and Forte, 1997) before they avalanche into the lower mantle.

Of course, not all slab material has to penetrate into the deep mantle, and the LLSVPs may be comprised of slab material subducted long ago.

Seismological evidence for reflections down to 1000 km beneath southwest Pacific subduction zones is consistent with the penetration of MORB-bearing material into the lower mantle (Rost et al., 2007). Sequestration of dense MORB material (e.g., Hirose et al., 1999; Tan and Gurnis, 2005; Ohta et al., 2007) may account for the chemically distinct nature of the LLSVP material. This concept certainly requires geochemical assessment, as it is the only approach to establishing the temporal isolation of the LLSVP reservoir.

CONCLUSIONS

Recent deep-mantle seismological findings give rise to the hypothesis that any deep mantle plumes will originate near the margins of LLSVPs at the base of the mantle. Chemically distinct and dense LLSVP piles may be organized underneath large-scale upwellings associated with return flow from subduction-induced downwellings. The margins of the LLSVP at the CMB are the hottest locations in the mantle and may contain partial melt at the CMB that is imaged as ULVZ structure. The LLSVP and ULVZ structures may contain important isotopic signatures that become entrained in plumes that rise from boundary layer instabilities on the LLSVP margins. The recent data and models do not demonstrate that whole-mantle plumes exist. However, the emerging understanding of lower-mantle structure and processes does provide guidance as to where and why any plume rising from the deep mantle will originate, and how they may sample thermally and chemically distinct source regions other than right at the CMB.

ACKNOWLEDGMENTS

This work was supported in part by the U.S. National Science Foundation under grants EAR-0125595, EAR-0453884, EAR-0453944, EAR-0510383, and EAR-0456356. Editors Gillian Foulger and Donna Jurdy and reviewers Anne Hofmeister and Dion Heinz made numerous helpful suggestions that improved the article. The authors thank Thorsten Becker for posting the tomography models from Becker and Boschi (2002) on his Web page and for software to expand the spherical harmonics, and Jeroen Ritsema for plotting software. Figures 1–3 were made with the aid of Generic Mapping Tools (Wessel and Smith, 1998).

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Manuscript Accepted by the Society 31 January 2007