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 Vs/Vp ratios, and high density; 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 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 non-uniform spatial distribution. This plausibly accounts for some attributes of major hot-spot volcanism.

1. 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 layer 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 hot spots 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 Nataf, 1998; Lin and van Keken, 2005; Wen, 2006; Montelli et al., 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 paper, 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 last 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 at the CMB, deep mantle seismic wave anisotropy, and variable occurrence and topography of the D" seismic velocity discontinuity. This paper 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 provides a framework for considering the implications for any deep mantle plumes that may rise to the surface.

2. 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, while lower-than-average wave speeds are commonly found beneath the Pacific Ocean and 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 (V_S) models but there is less consistency among P-wave velocity (V_P) models (Figure 1). The differences between large scale lower mantle V_S and V_P 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 V_P 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, $(Mg,Fe)SiO_3$, magnesium silicate in perovskite structure (*Pv*), should undergo a phase transformation at pressure-temperature conditions within a few hundred km above the core-mantle boundary (CMB) (Murakami et al., 2004, Tsuchiya et al., 2004; Oganov and Ono, 2004; Lay et al., 2005). *Pv* transforms into a post-perovskite structure (*pPv*) that is predicted to be accompanied by a V_S increase of several percent, but little change in V_P. This could be one cause of decoupling of variations in V_S and V_P in the lowermost mantle. The Vs/Vp ratio should be highest in high shear velocity regions because the phase transition will occur at shallower depth 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 affect on S wave velocity. As seen in Figure 1, most seismic models tend to be zero-meaned 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 relatively low temperature regions below present day and historic subduction zones. If this is the case, the large low shear velocity 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 where subduction has occurred over the past 200 million years, 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). While debate continues on this topic

(Romanowicz, 2001; Masters and Gubbins, 2003), indications are that a density increase is associated with the strongest V_S reductions located in LLSVPs (e.g., Ishii and Tromp, 1999; Trampert et al., 2004). Simultaneous analysis of V_S and V_P behavior further suggests that LLSVPs have bulk sound velocity anomalies (increases) that are anti-correlated 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 immediately complicates the interpretation of these regions, because chemical difference 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 large pressure. 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 trade-offs 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 (Figure 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 since thermal gradients should be more gradual. Additionally, weak reflections from a velocity decrease in the upper portion of the LLSVP in the Pacific may indicate a chemical boundary (Lay et al., 2006) or a phase boundary within chemically distinct LLSVP material (Ohta et al., 2006). Additional internal layering within the LLSVP beneath the

Pacific has been inferred from seismic wavefield reflectivity resolved by stacking of 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 was attributed to forward and reverse transformations of Pv to pPv (as predicted by Hernlund et al., (2005)). This region is also underlain by a mild ULVZ (Avants et al., 2006). Lateral depth variations of the mapped perovskite phase boundaries within the LLSVP are consistent with a lateral increase in temperature toward the LLSVP margin in the central Pacific region.

In addition to evidence for sharp boundaries of the two LLSVPs, several forward modeling studies have advocated a chemical origin for isolated seismic heterogeneities, at small to intermediate scales (500-1000 km) (e.g., Wysession et al., 1994; Bréger and Romanowicz, 1998), as well as at small scales (1-10's of km) that scatter seismic waves and contribute to high frequency coda energy (e.g., Hedlin and Shearer, 2000; Earle and Shearer, 2001). Some of the small scale scattering features at the CMB have been attributed to partial melt of deep mantle material (e.g., Vidale and Hedlin, 1998; Wen and Helmberger, 1998, Rost et al., 2005), owing primarily to magnitude of the requisite velocity reductions and a 3:1 Vs:Vp velocity reduction ratio. It is difficult to attribute large deep mantle velocity reductions (e.g., > 10%) to any expected deep mantle materials in absence of some level of melt; in fact, the possibility of partial melt in the deep mantle has been advocated as an explanation for occurrence of thin ultra-low velocity zones (ULVZ) right above the CMB (Williams and Garnero, 1996; Revenaugh and Meyer, 1997). Ultra-low velocity zones appear to exist preferentially beneath low V_S regions, including the two LLSVPs (e.g., Garnero et al., 1998; Williams et al., 1998).

This is hard to quantify since the global distribution of ULVZ structure is not presently well constrained; less than half of the surface area of the CMB has been even qualitatively characterized (Thorne and Garnero, 2004). However, the CMB in some isolated spots has been analyzed in great detail, suggesting partial melt in small domes, small dense zones with a flat top, and even small pockets right beneath the CMB with anomalous properties (Wen and Helmberger, 1998; Helmberger et al., 1998, 2000; Wen, 2000; Rost and Revenaugh, 2001; Rost et al., 2005). These structures all point to complex processes occurring down to small scales, undoubtedly related to high temperatures at the base of the mantle thermal boundary layer.

Lowermost mantle seismic wave anisotropy may also offer clues to deep mantle chemistry and dynamics; this is suggested from seismic studies (e.g., see Lay et al., 1998; Kendall and Silver, 1998; Kendall, 2000), mineral physics calculations (e.g., Stixrude, 1998; Karki et al., 1999; Mainprice et al., 2000; Wentzcovitch et al., 2004; Hirose et al., 2006b), as well as 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 hot spot (Russell et al., 1998). Given that there is a first-order correlation between the distribution of surface hot spots and the locations of the LLSVPs (e.g., Thorne et al., 2004; any correlation is less apparent for P velocity heterogeneity) at the base of the mantle (Figure 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 plume conduit dimension (e.g., < 500 km) compared to the long seismic wave propagation paths (typically > 5000 km) (see for example, 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 hot spots are inferred to be underlain by low V_P velocities extending down to the CMB (this 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 of 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., Tilmann et al., 1998), but at present this does not appear to have been done convincingly.

Less direct approaches, such as correlation studies of surface hot spots distributions and deep mantle velocity patterns have been pursued over several decades (e.g., Morgan, 1971; Hager et al., 1985; Thorne et al., 2004), but such analyses do not unequivocally constrain or require the existence of whole mantle plumes.

3. Lower mantle chemistry and dynamics

Several conceptual models have been developed in recent years to explain the observed LLSVPs. It proves dynamically difficult to account for the huge low-velocity anomalies beneath Africa and the Pacific in an isochemical mantle, even in models that impose a large scale pattern of downwelling by employing geologically recent plate velocities as surface boundary conditions (e.g., Bunge et al., 1998; McNamara and Zhong, 2005). In dynamical models that lack a thermochemical component, plumes tend to organize into clustered networks of thin upwellings (plume clusters) that form away from downwelling regions (e.g. McNamara and Zhong, 2005; Schubert et al., 2005). Although current research is assessing whether regions of plume clusters may resemble the large, low-velocity anomalies in the lowermost mantle when viewed through the blurred "eyes" of seismic tomography (Ritsema et al., 2006), it appears that thermochemical models of mantle convection provide the best explanation for the existence of the LLSVPs.

Thermochemical conceptual models that strive to explain the dynamics related to LLSVPs beneath Africa and the Pacific typically fall into two categories. Both invoke a large volume of anomalously-dense mantle material; however, they differ in terms of the relative buoyancy and geologic longevity of the chemical anomaly.

The superplume hypothesis typically describes the large, low-velocity anomalies as being due to the presence of large, upward-doming plumes of the more-dense material (e.g. Davaille, 1999; Forte and Mitrovica, 2001; Davaille et al., 2002; Davaille et al;

2005). These models are characterized by the denser component having a net positive buoyancy (the thermal buoyancy exceeds the negative buoyancy associated with the intrinsic density anomaly) such that it becomes unstable and forms large superplumes that are currently rising in the mantle. It has been shown (e.g., Davaille et al., 2002) that these structures may rise and sink many times before ultimately being well-mixed into the background mantle. Smaller-scale thermal plumes can originate from the tops of these large domes, entraining some of the chemically distinct material in the dense dome. This may, in turn, explain the anomalous chemistry observed at ocean island basalts (OIBs) (e.g., Hofmann, 1997, Fitton, this volume). One aspect of this model is that the present instance in Earth's history has the more-dense material actively rising as opposed to other times in the past in which this material was either a stratified layer or sinking after a previous ascent.

Another, similar superplume phenomenon observed in geodynamical modeling involves the presence of long-lived, stable superplumes (McNamara and Zhong, 2004a). If the anomalously dense mantle component has a higher intrinsic viscosity (~100x) than the less-dense material, it may form large dome structures that migrate laterally across the lower mantle. These structures can maintain a vertical height, and they do not experience the rising and sinking observed in laboratory studies (e.g., Davaille et al., 2002). However, it is difficult to put forth a mineral physics explanation that would provide the necessary intrinsic viscosity increase to the dense material, so such a model is not favored here.

Although the basic superplume model remains a viable hypothesis, here we focus on the second category of thermochemical mantle hypotheses, which involve the presence of long-lived, stable dense piles of chemically distinct material (e.g., Christensen, 1994; Tackley 1998, 2002; Kellogg et al., 1999; Jellinek and Manga, 2002, 2004; Ni et al., 2002; McNamara and Zhong, 2004a, 2004b, 2005; Nakagawa et al., 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 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 Myr over 11 stages of plate motions as provided by Lithgow-Bertelloni et al., (1998)). The calculation employed the Boussinesq approximation with constant thermodynamic 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 which included a 30x increase across the transition zone was used. The initial condition included a flat, 255km 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 the observed LLSVPs (see Figure 3a, 3d). This modeling demonstrated that it is dynamically feasible that global flow patterns derived from the history of subduction 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 currents 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 heating. If we assume that deep mantle piles have some temporal stability, then it is reasonable to assume they are denser than surrounding mantle (e.g., at a minimum, 2-5% denser). This is consistent with the suggestion of increased density in these locations from some seismic studies (Trampert et al., 2004; Ishii and Tromp, 2004). The possibility of Fe-enrichment in D″ would decrease shear velocity and result in density elevation, so the chemical heterogeneity could be residual material from core-formation process or accumulation of core-mantle reaction products. Subducted mid-ocean ridge basalt (MORB) 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., 2006)

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 of the LLSVP and internal flow of the LLSVP cause the boundaries between the surrounding mantle and LLSVP to be particularly hot (see Figure 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 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 beneath the SW margin of the same LLSVP (Rost et al., 2005, 2006a). 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 Figure 3c and Figure 4. Large thermochemical piles are deflected away from downwellings by subduction currents, and swept to concentrate beneath upwelling return flow. LLSVP piles may thus be the key to long-term history of subduction. LLSVP topographical features near their margins guide upwelling and serve as sites of thermal boundary layer instabilities. 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., Hoffmann, 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 due to 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 SW Pacific subduction zones is consistent with the penetration of MORB-bearing material into the lower mantle (Rost et al., 2006b). Sequestration of dense MORB material (e.g., Hirose et al., 1999; Tan and Gurnis, 2005; Ohta et al., 2006) may account for the chemically distinct nature of the LLSVP material. This concept certainly requires geochemical assessment, as this is the only approach to establishing the temporal isolation of the LLSVP reservoir.

4. 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.

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Figure Captions

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, 2001), KH00 (Kárason and van der Hilst, 2001), TXBW (Grand, 2002), BD00 (Becker and Boschi, 2004), 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 (2004). Hotspot locations (from Steinberger, 2000) are shown as red circles. Plate boundaries are magenta lines, but convergent boundaries are shown in blue.

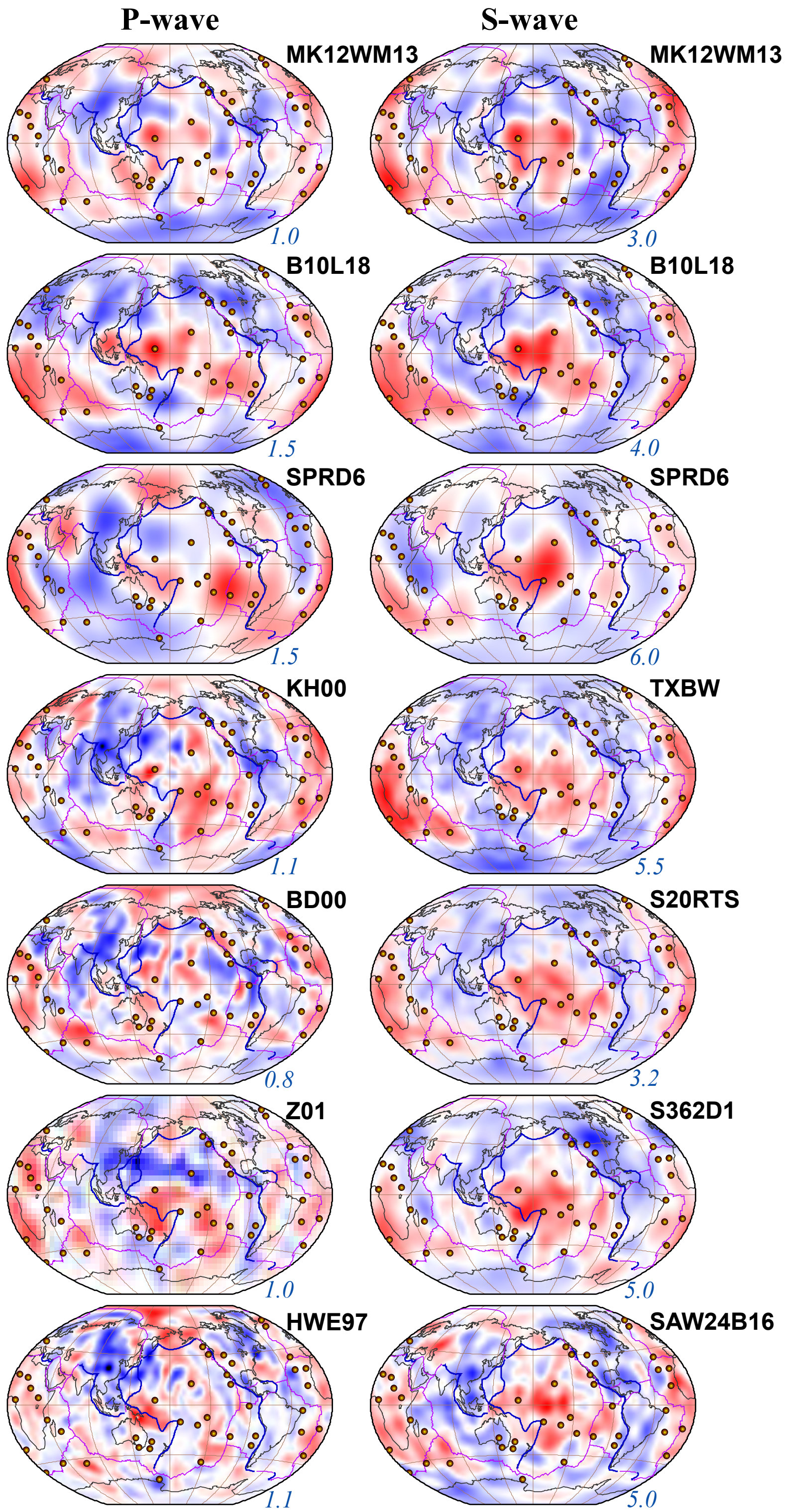
Figure 2. Map showing lowermost mantle V_S perturbations from model TXBW (Grand, 2002) (color scale is the same as in Figure 1) along with locations where seismic studies have inferred distinct edges to the LLSVPs (thick black lines) using waveform and/or travel time analyses. Lower case letters in boxes indicate specific studies for different regions, and correspond to: (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 (2003ab), and (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.

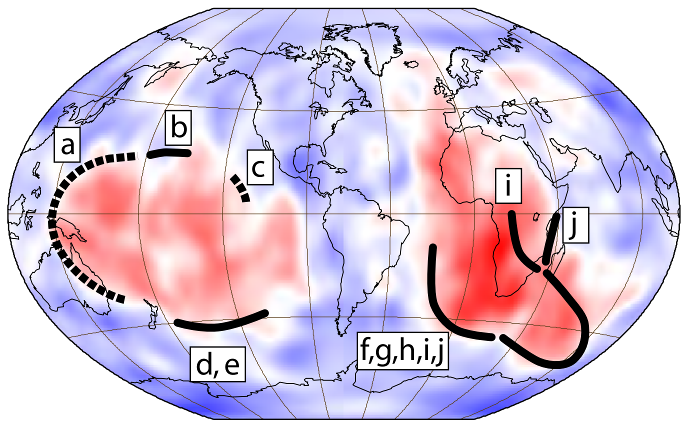
Figure 3. Continents, plate boundaries, and hotspots are shown on maps at the top of four boxes that represent the area of the whole globe, and the volume of the mantle from the surface to the CMB. (a) Compositionally distinct dense piles from the geodynamic calculation of McNamara and Zhong (2005). (b) The locations of the hottest temperature in the mantle for the calculation of (a) are shown. Iso-temperature contour is 0.98 for the calculation that spans temperatures from zero to one. The hottest temperatures are within the piles and typically near the edges. (c) A temperature cross-section is shown, along with the piles from (a) in a transparent gray, and the hottest CMB temperatures of (b) 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 $\ell=8$, with iso-velocity contours at -0.3% (red) and 0.5% (blue). Dense piles in the geodynamic calculation (a) are geographically distributed similar to the low velocities (red) in (d).

Figure 4. A close-up of the thermochemical dense pile beneath the Pacific Ocean in Figure 3a is shown. A cross-section from the surface to the CMB displays temperature variations, with the yellow line denoting the boundary of the chemically distinct material in the pile. We identify the thermal-chemical anomaly as the LLSVP. Part of the CMB surface is shown in front of the cross-section, along with an iso-temperature contour (at 0.98, as in Figure 3). The lower panel shows the same cross-section, but only the pile,

with a more expanded color scale (colors span T=0.7 to T=1.0). Convective motions are indicated by the arrows. The hottest zones may invoke partial melt of LLSVP material, either at the CMB (denoted as ULVZ in the figure), or in some isolated locations further up within the LLSVP.



Garnero, McNamara, Lay
[Figure 1, P4 monograph]



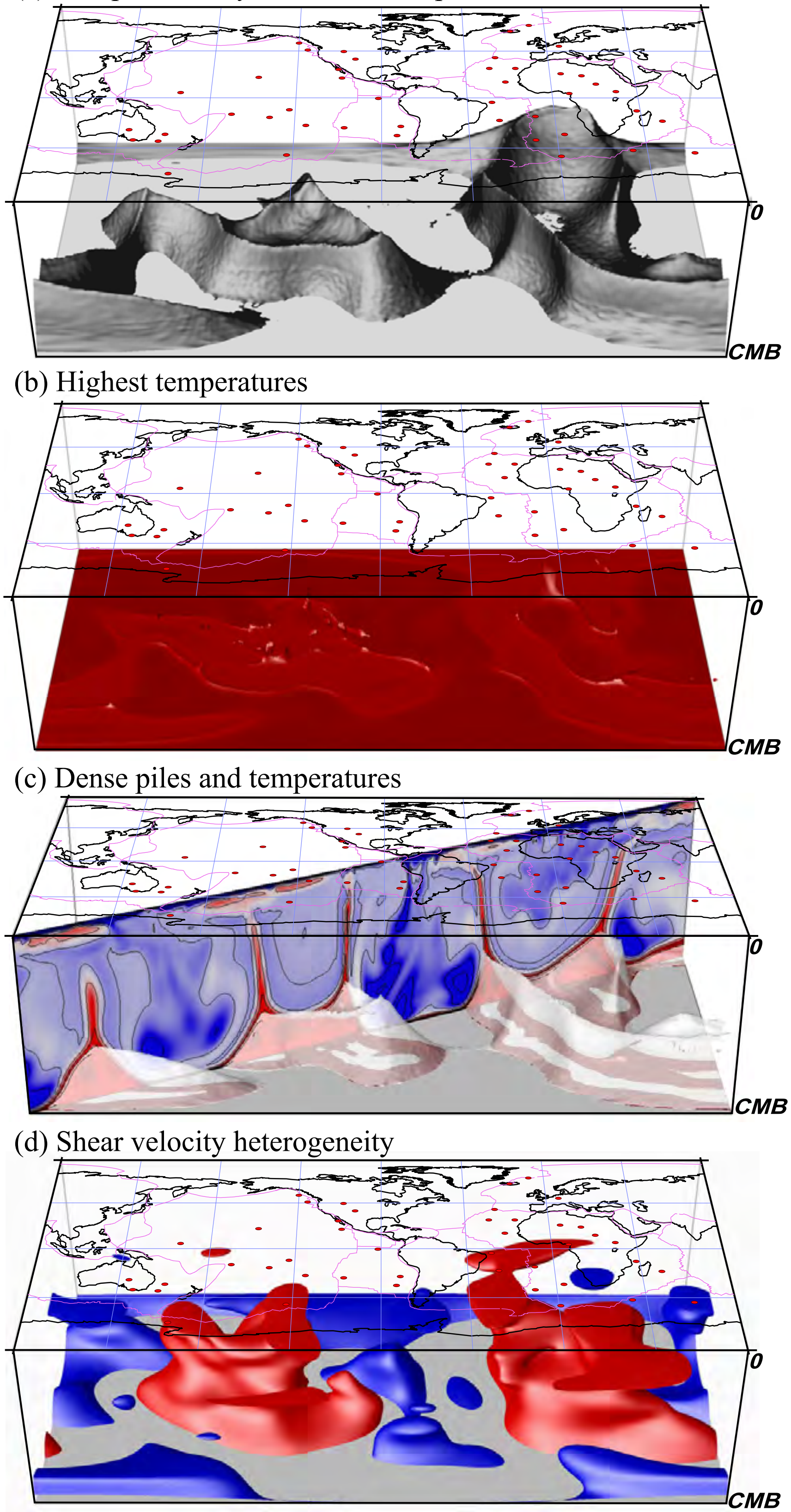
Garnero, McNamara, Lay
[Figure 2, P4 monograph]

(a) Compositionally distinct, dense piles

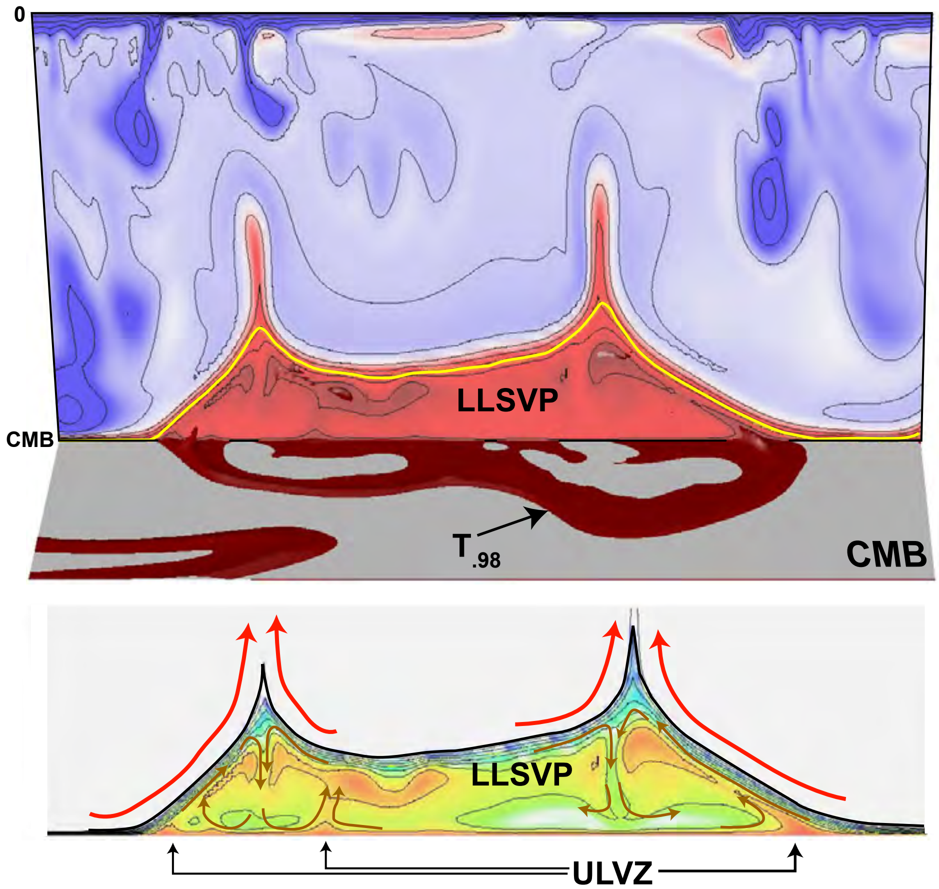
(b) Highest temperatures

(c) Dense piles and temperatures

(d) Shear velocity heterogeneity



Garnero, McNamara, Lay
[Figure 3, P4 monograph]



Garnero, McNamara, Lay
[Figure 4, P4 monograph]