The deep mantle thermo-chemical boundary layer: The putative mantle plume source

Thorne Lay*
Earth Sciences Department, University of California–Santa Cruz, Santa Cruz, California 95064, USA

ABSTRACT

Hypothetical narrow, cylindrical thermal plumes from great depths in the Earth, which are commonly invoked to account for relatively stationary and persistent concentrations of volcanism at the surface, are intimately linked to the notion of instabilities and upwellings of a deep thermal boundary layer in the mantle. If the mantle is undergoing whole-mantle convection with no internal stratification, the primary thermal boundary layer within the interior that may serve as a plume source is located above the core-mantle boundary. If the mantle convection regime is globally or locally stably stratified by compositional layering, thermal boundary layers should exist on the margins of the compositional contrasts, and any such boundary layers could serve as plume sources. Seismological, mineral physics, and geodynamic evidence favors the existence of a complex thermo-chemical boundary layer in the lowermost few hundred kilometers of the mantle, the so-called D″ region, involving extensive, if not global, chemical stratification. Although chemical stratification of the mid-mantle has not been ruled out, the lack of direct evidence for global existence of thermal boundary layers in the mid-mantle means that the lowermost mantle region is generally invoked as the source of thermal plumes that give rise to melting and chemical anomalies associated with hotspot volcanism. Observational constraints on this putative mantle plume source are summarized here, along with consideration of the attendant implications for plume existence.

Keywords: D″ region, core-mantle boundary, superplumes, thermal plume, mantle heterogeneity.

INTRODUCTION

The mantle plume hypothesis (Morgan, 1971, 1972; Anderson and Natland, this volume) postulates the existence of narrow cylindrical upwellings that drain hot material from an internal thermal boundary layer in the mantle, giving rise to massive pressure-release melting within and below the lithosphere, thereby accounting for hotspot volcanism. Although various aspects of the shallow melting and dynamic system predicted for thermal plumes can be explained by alternate mechanisms, all notions of thermal plumes share the need for a deep thermal boundary layer from which the plumes originate. Characterization of any mantle thermal boundary layers at depth thus has direct relevance to assessment of the plume hypothesis.

The existence of thermal boundary layers in a planet has two main prerequisites: heat must flow across a persistent boundary,

*E-mail: tlay@es.ucsc.edu.

and there must be density or viscous stratification of the system that permits superadiabatic thermal gradients to develop and persist adjacent to the boundary (e.g., Stacey and Loper, 1983; Christensen, 1984; Olson et al., 1987). These requirements are certainly satisfied in the Earth at the surface and at the core-mantle boundary. Dramatic compositional and viscosity contrasts across these two boundaries combine with the ongoing cooling of the mantle and the core to give rise to thermal boundary layers at the top and bottom of the mantle. If the mantle convection regime is further dynamically stratified as a consequence of chemical or rheological layering, thermal boundary layers with superadiabatic thermal gradients will have developed across the stratified boundaries as heat has sought its way to the surface. Advocates of the mantle plume hypothesis thus invoke boundary layer instabilities either from the core-mantle thermal boundary layer or from some mid-mantle thermal boundary layer associated with stratification of the mantle flow regime (e.g., Kellogg et al., 1999).

The presence of a global seismic velocity discontinuity near 650 km depth, combined with the change in seismic velocity gradients above and below the discontinuity, prompted early consideration of chemical layering of the mantle at the base of the transition zone (e.g., DePaolo and Wasserburg, 1976). Recognition that the 650 km discontinuity can be explained well by a phase transition in (Mg,Fe)$_2$SiO$_4$ (with possible contribution from a garnet to perovskite transition) removed the need for a chemical contrast at this depth (e.g., Anderson, 1967). The endothermic phase change at 650 km probably impedes radial transport of material and may be accompanied by a rheological transition with important dynamic effects (e.g., Dziewonski, this volume). No globally extensive seismic velocity discontinuity has yet been established at depths between the base of the transition zone and the core-mantle boundary, and although it is plausible that density stratification does exist and is simply difficult to resolve with seismic waves (Anderson, 2002), there is as yet no strong observational basis for asserting that there are global thermal boundary layers in the mantle anywhere other than near the surface and the core-mantle boundary (e.g., Castle and van der Hilst, 2003). Thus advocates of thermal plumes usually invoke deep mantle origins, although upwellings from the transition zone or from a mid-mantle discontinuity have not actually been ruled out.

Seismic wave analysis has revealed that there is significant structural heterogeneity near the base of the mantle, with strong radial seismic velocity gradients (or discontinuities) and strong lateral velocity variations. The lowermost 200–300 km of the mantle, designated by seismologists as the D” region, has the most pronounced variations, and these are more strongly manifested in shear wave velocities (Vs) than in compressional wave velocities (Vp). These seismic wave velocity variations provide evidence that a major thermo-chemical boundary layer exists on the mantle side of the core-mantle boundary, extending at least several hundred kilometers upward into the lower mantle (e.g., Lay et al., 2004). Of particular importance are two large-scale regions of low shear velocity in the deep mantle below the central and southern Pacific Ocean and below the southern Atlantic Ocean, Africa, and the southern Indian Ocean (Dziewonski et al., 1993). These regions have horizontal dimensions of 2000–3000 km and shear velocity anomalies (dVs) averaging –3% to –5%. The common assumption that low shear velocity in the mantle is a consequence of relatively hot, buoyant material has led to designation of these features as “superplumes,” an unfortunate term that has dynamic implications that are actually at odds with some of what is known about these regions and with the conventional fluid dynamic definition of a thermal plume. I will call these regions large low-velocity provinces, a dynamically neutral description appropriate for the uncertain nature of these structures.

The two large low-velocity provinces in the lower mantle have strong seismic velocity expressions in the D” region, but it appears that some portions of them extend upward (or are overlain by separate low-velocity material) 800–1000 km above the core-mantle boundary, significantly exceeding the conventional thickness associated with D”. Weaker velocity anomalies may extend to even shallower depths. Evidence for chemical contrast between large low-velocity provinces and surrounding mantle, summarized later, suggests that large low-velocity provinces are massive chemical heterogeneities in the lower mantle, and as such they may be enshrouded in thermal boundary layers. Thus even though large low-velocity provinces may not be buoyantly ascending or may be unable to penetrate up to the surface, they may still give rise to thermal plumes shed from their boundary layers. If they are long-lived they may significantly affect the overall thermal transport in the deep mantle.

In addition to large low-velocity provinces, there are more localized regions of relatively low velocity in the lower mantle, some of which underlie surface hotspots and some of which do not. With as yet poorly resolved lateral dimensions of 200–500 km, those low-velocity regions with significant near-vertical continuity are commonly interpreted as plumes rising through the mantle (e.g., Goes et al., 1999; Bijwaard and Spakman, 1999; Nataf, 2000; Montelli et al., 2004; Zhao, 2004), with the explicit assumption that low shear velocity indicates hot, buoyant material.

Combining the general expectation of thermal boundary layers above the core-mantle boundary and on the margins of major chemical boundaries, the deep sources for plumes are commonly envisioned, of late, as shown in Figure 1. “Classic” plumes are viewed as narrow conduits with diameters of 50–200 km rising directly from the thermal boundary layer at the core-mantle boundary, some with sufficient plume flux to ascend 2900 km through the mantle conveying excess temperatures that induce large-scale pressure-release melting at locales such as Hawaii and Iceland (e.g., Stacey and Loper, 1983; Davies, 1988; Sleep, 1990). If D” is a distinct chemical layer, such plumes may rise from the top of that layer rather than directly from the core-mantle boundary (e.g., Tackley, 1998). Hotspot regions of the south Pacific and Africa are, perforce, attributed to plumes ris-
ing from the large low-velocity provinces in the deep mantle beneath them (e.g., Jellinek and Manga, 2002; Romanowicz and Gung, 2002). Yet other hotspots are attributed to shallow plate tectonic processes (e.g., Courtillot et al., 2003; Anderson, this volume). Entrainment of material from the core, D″, or large low-velocity provinces in ascending plumes from the deep mantle is commonly advocated as an explanation for geochemical anomalies in hotspot magmas (e.g., Hofmann and White, 1982; Hart et al., 1992; Hofmann, 1997; Albarede and van der Hilst, 2002), although the level of confidence with which these can be tied to any deep boundary layer is not high.

Given the array of possibilities for a deep mantle source of thermal plumes, it is worthwhile to consider what is actually known about the lowermost mantle structure to provide a basis for assessing conceptual notions such as those in Figure 1. Fortunately several recent review articles summarize the multidisciplinary contributions to our current understanding of lowermost mantle structure (e.g., Lay et al., 1998a; Garner, 2000; Lay and Garner, 2004; Lay et al., 2004). As those references provide comprehensive reading lists and citations, the goal here will be to cast summary observations and constraints on the structure and processes near the core-mantle boundary in the context of the debate (e.g., Anderson, this volume) about the existence and nature of thermal plumes in the mantle.

ATTRIBUTES OF THE CORE-MANTLE THERMO-CHEMICAL BOUNDARY LAYER

Thermal Structure

The temperature drop across the surface thermal boundary layer is well established as being ~1200–1300° (e.g., Green and Falloon, this volume), but there is great uncertainty about the temperature contrast across the core-mantle boundary layer. The latter contrast must be inferred based on extrapolations from the few relatively robust temperature tie-points in the deep interior constrained by experiments, the temperatures of upper mantle phase transitions in olivine, and the estimated freezing point of core alloy at the inner core boundary. One can estimate the temperature contrast across the core-mantle boundary when these tie-point estimates, themselves subject to substantial uncertainty due to imprecisely known compositional effects, are extrapolated down or up along mantle and core adiabats, respectively, under the assumption that there are no regions of superadiabatic gradients anywhere other than near the core-mantle boundary. This procedure yields a temperature contrast across the core-mantle boundary on the order of 1000–2000° (e.g., Williams, 1998; Boehler, 2000), comparable to the lithospheric temperature drop. The core-mantle boundary itself is nearly isothermal as a result of the very low viscosity of core material, with a temperature somewhere between 3300 K and 4800 K, and any instability in the overlying mantle thermal boundary layer could thus bring very hot material upward into the mantle. The observation that excess temperatures associated with hotspot magmas are usually found to be less than 200° (see Anderson and Natland, this volume; Green and Falloon, this volume; Presnall and Gudfinnsson, this volume) raises immediate questions about the viability of thermal plumes rising directly from the core-mantle boundary due to the large excess temperature estimates for the boundary layer. Given secular cooling of the core, the temperature contrast of ancient plumes would be even higher, and Archaean komatiites are often interpreted in this light; however, questions are now arising about the excess temperatures actually associated with these melts (e.g., Parman and Grove, this volume). Albers and Christensen (1996) and Farnetani (1997) have explored the excess temperature issue,
with the latter establishing that unless the core-mantle boundary temperature estimates are in gross error, it is far easier to account for the relatively modest temperature excess of hotspot melts if a much weaker boundary layer serves as the source of a thermal plume.

If the lower mantle is chemical layered—for example, if D″ is a compositionally distinct layer—the estimated temperature contrasts across the upper boundary of D″ and the core-mantle boundary can be reduced by a factor of two, helping to resolve the excess temperature problem (Farnetani, 1997). Dynamic instabilities of a thermal boundary layer above a dense, compositionally distinct layer tend to draw in material from the upper portion of the thermal boundary layer, not sensing the full temperature increase. This same argument also holds for thermal boundary layers at shallower depths than D″. A key issue that then emerges is the longevity of a chemical contrast in the face of steady entrainment (e.g., Sleep, 1988; Tackley, 2000). Relatively small density increases of 1%–3% allow chemical layers to persist for a long time for entrainment models that account for strong pressure effects on material properties such as thermal expansion (e.g., Zhong and Hager, 2003). Pressure effects also serve to increase the temporal and spatial scales of lower mantle features so that very large long-lived features are favored at the base of the mantle.

Absolute temperature issues aside, the best argument for a thermal boundary layer at the base of the mantle is that the heat flux through the core-mantle boundary needed to sustain the geodynamo is estimated to be from 1 to 15 terawatts (TW), which is a very uncertain number but almost always a positive value (except in a few exotic models), implying heatflow from the core into the mantle (e.g., Buffett, 2002; Lay et al., 2004). Given estimates of a global surface heat flux on the order of 42–44 TW (e.g., Kellogg et al., 1999), there is a correspondingly large uncertainty in the degree to which the mantle is heated from below versus heated from within, with the former essential for genesis of thermal plumes by boundary-layer instabilities. Lower values of core-mantle boundary heat flux are easier to reconcile with an old inner core and with reasonable initial Earth temperatures. If chemical layering in the mantle produces any mid-mantle thermal boundary layers one can obtain lower core-mantle boundary heat flux estimates in some of the calculations. Reassessment of the estimates of global surface heat flux (e.g., Hoffmeister and Criss, this volume) may lower the heat flux values for the surface and the core-mantle boundary as well.

**Large Low-Velocity Provinces**

Seismological constraints on the temperature field in D″ are even more indirect than those from geodynamic and geomagnetic arguments, with lack of constraint on the compositional effects and on the high-pressure thermal expansion leading to large uncertainties (e.g., Trampert et al., 2001; Julian, this volume). Perhaps the most robust statement is that seismology reveals the existence of large-scale lateral variations in the seismic velocity structure near the base of the mantle (Fig. 2; see also Ritsema, this volume), with a large degree 2 component in the shear wave structure. Recent global seismic tomography models for D″ are discussed in detail by Lay et al. (2004), but for our immediate purpose there are two key attributes of the D″ boundary layer relevant to the plume question.

The first is that two large low-velocity provinces, as mentioned previously, are evident in all of the shear wave models, separated by a circum-Pacific ring of relatively high-velocity material. The primary wave (P-wave) velocity expression of these two large low-velocity provinces is much more muted, but there is less agreement among Vp models, and Vp fluctuations are only ~20%–50% of those in the Vs models. As the resolution of seismic tomography improves, internal coherence of the large low-velocity provinces may diminish, but the large-scale pattern will persist (see Dziewonski, this volume). The predominance of long-wavelength structure runs counter to expectations for a very hot, inviscid thermal boundary layer beset by many instabilities giving rise to plumes: some process imposes large-scale structure on the boundary apart from the intrinsic boundary layer temperature structure. Possibilities include disruption of the hot thermal boundary layer by cooled downwelling material in the lower-mantle circum-Pacific high-velocity ring or the existence of large-scale chemical heterogeneities embedded in the boundary layer. If there is a thermal contribution to the large-scale heterogeneities, attendant variations in radial temperature gradients will impose a spatially varying heat flux boundary condition on the core flow regime (e.g., Glatzmaier et al., 1999), which may play an important role in reversals of the magnetic field.

The second important attribute of the D″ boundary layer to be gleaned from these low-resolution tomography models is that in some well-sampled regions, such as under the central Pacific, there are strong decorrelations between Vp and Vs variations, or at least much stronger relative perturbations in Vs versus Vp. This indicates that thermal variations are not the sole explanation for the large-scale patterns. Inversions for bulk sound velocity variations (which depend on bulk modulus and density, but not on rigidity) emphasize the differences in behavior of Vp and Vs and suggest some anticorrelation with Vs patterns on large scales (Fig. 3). This behavior cannot be readily accounted for by temperature alone, and although there are many technical challenges to bulk sound velocity estimation, most seismologists agree that compositional heterogeneity is probably important on large scales in D″ (e.g., Masters et al., 2000). The differences in global sampling for P-waves and shear waves (S-waves) in the deep mantle is significant, but even localized determinations of velocity structure, as for the central Pacific region illustrated in Figure 3, support a difference in the behavior of Vp and Vs that can be accounted for only by bulk modulus and shear modulus perturbations of opposite sign. This appears to be the case for both the Pacific and African large low-velocity provinces, although complex relationships between Vp and Vs are documented in other regions as well (e.g., Wysession et al., 1999; Saltzer et al., 2001).
Attributes of the large low-velocity provinces are summarized in Figure 4, and the many seismic observations associated with these structures are discussed at length by Lay and Garnero (2004). The lateral gradients on the edges of these structures are very abrupt, resolved by detailed waveform and traveltime analyses (not by global tomography) to be on the scale of tens of kilometers (Ni et al., 2002). For some portions of the large low-velocity provinces, sharp lateral gradients persist into the mid-mantle, 800–1000 km above the core-mantle boundary. There are ongoing debates about the internal velocity structure of the large low-velocity provinces, but most researchers agree that the Vs reductions are much stronger than Vp reductions. The strong lateral gradients and anomalous Vp/Vs observations strongly suggest a distinct chemistry of large low-velocity provinces. At the high pressures near the base of the mantle, the sensitivity of seismic wave velocity to temperature variations is significantly suppressed relative to that in the shallow mantle (e.g., Chopelas, 1996; Julian, this volume), so very large temperature increases would be required to account for the Vs reductions, yielding very hot large low-velocity provinces. However, the apparent need for a compensating bulk modulus increase (to explain why Vp is not proportionally low) requires a chemical change, which alone may suffice to reduce the rigidity with no thermal anomaly (candidate materials are not yet constrained). While pushing the limits of normal mode analysis, there is evidence that large low-velocity provinces have higher than normal density (e.g., Ishii and Tromp, 1999, 2004).

Taking the information presently available at face value, one could argue either that large low-velocity provinces are massive, hot upwellings with distinct chemistry that gives rise to their density and Vp/Vs characteristics (“superplumes”) or that they are chemically distinct piles of dense material that may or may not have any thermal anomaly, but do have steep lateral gradients sustained by the shear flow of the surrounding high-viscosity
deep mantle. Numerous investigations of the dynamics of chemical plumes have been spurred by these observations (e.g., Tackley, 2000; McNamara and Zhong, 2004), but a key question places all interpretations in doubt: what is the relative contribution of thermal versus chemical heterogeneity to the seismic velocity anomalies and to any positive or negative buoyancy? This issue is wide open, but in the context of considering scenarios such as that in Figure 1, we can accept with confidence that there are large chemical anomalies in the deep mantle concentrated near the core-mantle boundary, possibly with localized extensions upward beyond the conventional limits of the D″ region. Whether or not there are strong thermal boundary layers surrounding these regions that are capable of detaching and rising as plumes depends on whether large low-velocity provinces have relatively high, normal, or low temperatures. If the structures are piles of chemically distinct material with no major heat flux through their boundaries, shear flows in the surrounding mantle could still transport hot material from the core-mantle boundary layer.

Figure 3. (A) Map of bulk sound velocity ($V_b = (V_p^2 - 4/3 V_s^2)^{0.5}$) variations in the D″ region from the tomographic model Sb10118 of Masters et al. (2000). Comparison with the shear and compressional velocity models in Figure 2 shows anticorrelation of $V_b$ and $V_s$ in D″ in the central Pacific, where relative reductions in $V_s$ are more than 2.7 times those in $V_p$. The small box indicates the region of a detailed analysis of S- and P-wave triplications by Russell et al. (2001). (B) Models SPAC and PPAC obtained by Russell et al. (2001) for the central Pacific, along with a reference structure (PREM-H), a modification of the PREM model of Dziewonski and Anderson (1981), with D″ having the same velocity gradients as in the overlying lower mantle rather than reduced gradients as in PREM. This is viewed as an empirical homogeneous, adiabatic reference model relative to which effects of thermal and chemical heterogeneity in the boundary layer should be assessed. The sizes of the velocity discontinuities at 2661 km depth are indicated, along with the presence of a 10 km–thick ultra-low-velocity zone for models SPAC and PPAC. From Lay et al. (2004).

Figure 4. Top: Two massive provinces with thick regions of low $S$ velocity exist in the lower mantle: under the southern Pacific and under the southern Atlantic, Africa, and the southern Indian Ocean. Bottom: A south-to-north (left-to-right) cross-section sketch of the latter feature is shown, with a 200–300 km–thick layer in D″ having an abrupt decrease in $V_s$ of 1–3% overlying a layer with average $V_s$ decreases of 3%–5%. Under Africa this low-velocity body appears to extend as much as 800 km above the core-mantle boundary (CMB), and it may have an ultra-low-velocity zone (ULVZ) just above the CMB. It has sharp lateral boundaries in D″ and in the mid-mantle. The low-S-velocity province under the central Pacific is not yet as fully imaged, but appears comparable in size and in the strength of the velocity reductions.

Should this be expressed as $V_b$ (with “$V$” in ital and “$b$” subscripted?)
up along the margins of the large low-velocity provinces, with instabilities arising from the complex flow geometry (e.g., Jellinek and Manga, 2002). Thorne et al. (2004) report a statistical correlation between the location of surface hotspots and strong lateral gradients near the margins of low-velocity regions in the lowermost mantle (not only for the two large low-velocity provinces), which may provide some support for the latter notion. While the statistical significance of spatial correlations of this type is open to question, it is noteworthy that hotspots tend to not be centered over the core regions of large low-velocity provinces, which may bear upon their dynamic status.

Ultra-Low-Velocity Zones

It is important to recognize that the seismically characterized large low-velocity provinces are not equivalent to the “ultra-low-velocity zones” detected just above the core-mantle boundary (e.g., Garnero et al., 1998; Thorne and Garnero, 2004). The latter features occur rather extensively (Fig. 5) and involve very strong reductions in both Vp (up to 10%) and Vs (up to 30%), and there is evidence that they have strong (10%) density increases. The ultra-low-velocity zones are typically only a few kilometers to a few tens of kilometers thick, juxtaposed against the core-mantle boundary, and have sharp upper boundaries that reflect short-period seismic waves. The velocity anomalies in Vs and Vp have a ratio of 2.5 to 3, which is compatible with the presence of partial melting (e.g., Williams and Garnero, 1996; Lay et al., 2004), with likely iron partitioning into partially segregated melts possibly accounting for excess density. As seen in Figure 5, the properties of ultra-low-velocity zones are remarkable in terms of their velocity perturbations, and their location next to the core-mantle boundary, inarguably the hottest place in the mantle, strongly suggests a thermal influence associated with the thermal boundary layer there. However, there are interesting issues, such as, why, if there is dense melt, it has not completely drained to the core, and whether these regions represent core material that has penetrated into the mantle. The dynamics and thermal transport of ultra-low-velocity zones are just beginning to be explored numerically.

Figure 5 indicates that ultra-low-velocity zones occur in many regions of the core-mantle boundary layer. Williams et al. (1998) present a correlation analysis suggesting that the distribution of surface hotspots is better correlated with this distribution of ultra-low-velocity zones than with that of the large-scale low-velocity regions of D″. The significance of such correlations can be questioned because some ultra-low-velocity zones appear to actually have very localized three-dimensional configurations, involving mounds rather than thin lenses (e.g., Helmberger et al., 1998; Wen and Helmberger, 1998), so their horizontal dimensions may be much smaller than indicated by the 1D velocity profiles. Plumes rising from ultra-low-velocity zones should have large temperature excesses and should quickly drain out the localized structures, so it seems difficult to invoke ultra-low-velocity zones as sources for major plumes. If dense melts separate out from upwelling boundary layer instabilities, the ultra-low-velocity zones could be consequences rather than causes of thermal plumes.

The likelihood that partial melting or a melt phase is involved in ultra-low-velocity zones raises the possibility that the lower mantle is near the eutectic (the lower mantle assemblage is very likely to have a eutectic) of the bulk mantle composition (e.g., Akins et al., 2004). The effects of partial melt on seismic velocities are much more pronounced than those of subsolidus temperature variations, and this fact motivates consideration of large low-velocity provinces as regions with very small melt fractions, perhaps as a consequence of their chemical anomaly.
Ultra-low-velocity zones might then be viewed as relatively high-melt-fraction portions of D″, while large low-velocity provinces have a low melt fraction, with both differing chemically from overlying mantle. As it is unclear whether melts in D″ will be buoyant or will sink, the importance for the plume issue of any partial melting in ultra-low-velocity zones or large low-velocity provinces remains unclear. Ultra-low-velocity zones may represent partial melting of chemically distinct material and need not be related to the bulk lower-mantle eutectic at all. Not all regions of large low-velocity provinces are underlain by ultra-low-velocity zones (the northern and western portion of the Pacific large low-velocity province is, and so is the northernmost region of the African large low-velocity province; see Figures 5 and 4), and some ultra-low-velocity zones may underlie regions of high velocity in D″, so no generally valid interpretation of the connection between these low-velocity features has emerged.

Seismic Velocity Discontinuities

The strong lateral gradients at the margins of large low-velocity provinces appear to extend right down to the core-mantle boundary, either as vertical or as steeply dipping boundaries (e.g., Wen et al. 2001; Ni and Helmberger, 2003). While the possibility exists that the boundaries are caused by lateral transitions in partial melt fraction, it is more likely that they involve gradients in composition. In regions of D″ well removed from the large low-velocity provinces, there is evidence for regionally extensive seismic velocity stratification, with seismic velocity discontinuities involving increases in Vp and Vs ~200–300 km above the core-mantle boundary (e.g., Wysession et al., 1998; Lay and Garnero, 2004). Vs increases of 2%–3% have been widely observed (Fig. 6), with Vp increases tending to be weaker and more variable. The strongest discontinuities are observed in circum-Pacific regions where the large-scale structure of D″ has high Vs (Fig. 2); however, there is evidence for both Vp and Vs increases beneath the central Pacific, near the northern margin of the Pacific large low-velocity province (e.g., Russell et al., 2001). The latter observations complicate efforts to account for the D″ velocity discontinuities as thermal or chemical anomalies associated with subducted oceanic slab (e.g., Grand, 2002), and they have prompted notions either of a phase change or of chemical layering of D″.

The possibility of a phase change near the top of D″ has recently been boosted by the discovery of a postperovskite phase transition in MgSiO₃ perovskite (e.g., Murakami et al., 2004), which appears to have potential to account for the D″ seismic velocity discontinuities (Tsuchiya et al., 2004a). Theoretical calculations suggest a large positive Clapeyron slope for this phase transition (Tsuchiya et al., 2004b), which would result in large topography on the transition boundary and also in increased potential for dynamic instability of the boundary layer in regions where the postperovskite phase is present (Matyska and Yuen, 2004). There is a weak correlation between shallow depths of the D″ discontinuities and higher-velocity regions in large-scale tomographic models. To the extent that the volumetric velocity variations are an indicator of thermal variations, this is consis-
tent with a large positive Clapeyron slope for a phase change (e.g., Sidorin et al., 1999). However, the patchy occurrence of D̆ discontinuities favors the existence of the phase change primarily in relatively cool regions of high seismic velocity. Thus one might not expect any phase change to exist or for it to be deeper in the mantle in hot areas (large low-velocity provinces?), which is at odds with the observations of discontinuities under the central Pacific several hundred kilometers above the core-mantle boundary (e.g., Russell et al., 2001). It is also curious that the margins of large low-velocity provinces do not yet clearly indicate any transition to regions with a positive velocity increase at the top of D̆. What little evidence exists for structure at the tops of large low-velocity provinces tends to favor small Vs decreases at about the same depth as Vs increases in other regions (e.g., Wen 2001; Ni and Helmberger, 2003). Some combination of chemical heterogeneity and a phase change must be invoked to account for the full suite of observations, and at present there are substantial uncertainties in doing so (see Lay et al., 2004).

The other likely explanation for D̆ discontinuities involves chemical heterogeneities, either of primordial nature or possibly replenishing accumulations of oceanic crustal components delaminated from slabs in the deep mantle. The occurrence of strong D̆ discontinuities beneath regions of subduction over the past 100 million years and beneath regions of relatively high Vs in the mid-mantle is often invoked in favor of an oceanic slab connection. There are only a few regions where tomographic images show high velocities extending continuously across the mantle, with the region below the Gulf of Mexico perhaps the best example (e.g., Grand, 2002), but raypath sampling in the tomography studies is quite limited and may accentuate apparent continuity in finely parameterized tomographic models (see Dziewonski, this volume), so connections between D̆ structure and shallow subduction remain tenuous. The correlation between regions of historical subduction and mantle heterogeneity at depths of 800–1000 km are relatively strong (e.g., Scrivner and Anderson, 1992; Wen and Anderson, 1995, 1997), but this correlation weakens significantly with depth.

Because the interpretation of deep mantle seismic velocity structures alone is ambiguous, additional observations are needed. Seismologists have detected seismic anisotropy in many regions of D̆ (Fig. 7) and are seeking to establish a probe of the flow and strain system in the boundary layer and its relationship to flow in the mid-mantle (e.g., Lay et al., 1998b; Moore et al., 2004). While progress is being made in mapping the existence and properties of anisotropy in D̆, the most general current observation is that regions of relatively high shear velocity, where strong D̆ discontinuities are observed, tend to have the most pronounced anisotropy, and the orientation is close to being vertical transverse isotropy. This observation can be accounted for by dynamic models in which low temperatures and high stresses associated with mid-mantle downwellings drive underlying regions of D̆ into a dislocation deformation regime in which lattice preferred orientation in (Mg,Fe)O can develop over large scales (e.g., McNamara et al., 2002). It remains to be determined whether a postperovskite phase transition in the deep mantle plays a significant role in the anisotropy in the boundary layer, but there does appear to be some correspondence between the D̆ velocity discontinuities and the onset of detectable anisotropy.

**DISCUSSION AND CONCLUSIONS**

Given the many degrees of freedom that exist for the properties and configuration of hypothesized chemical heterogeneities, phase changes, partial melting, and thermal structure in D̆, it is possible to account for the diverse seismic observations with
various scenarios, but difficult to formulate definitive tests of
each hypothesis. Most discussions of the properties in the low-
ermost mantle invoke a hybrid scenario involving a partially
mixed boundary layer disrupted by downwelling slab material,
with a phase change and large piles of chemical heterogeneities
(Fig. 8). This scenario is motivated by and can be reconciled
with all of the seismic observations discussed in this paper (see
Lay and Garnero, 2004), and one can then superimpose thermal
plumes originating either within the thermal boundary layer at
the core-mantle boundary or above large low-velocity provinces,
resulting in the “vision” depicted in Figure 1. Direct demonstra-
tion of the existence of small-plume structures is at the frontier
of deep Earth seismic imaging efforts, and caution is warranted
with respect to all published interpretations of small-scale hetero-
geneties. This scenario requires some mechanism for reducing
the excess temperature of plumes rising directly from the core-
mantle boundary, and it suggests that chemical differences may
exist between plumes rising from large low-velocity provinces
and those rising from direct contact with the core (as may be
argued in the case of the Dupal anomaly, e.g., Wen et al., 2001).

An alternate scenario for the core-mantle boundary layer,
considered by Lay et al. (2004), involves reconciling the full
suite of D″ observations with a globally chemically stratified
lowermost mantle (Fig. 8B). This approach, motivated by the
first-order chemical stratification of the planet, suggests that
lateral variations of partial melting within the chemically dis-
tinct layer account for the diversity of seismic observations.
This concept would involve moderate thermal boundary layers at the
core-mantle boundary and at the top of the compositional layer.
Instabilities arising from the upper boundary layer could then
ascend to the surface as in Figure 1, presumably with all plumes
involving entrainment of similar compositional heterogeneities

Figure 8. Scenarios for the thermo-chemical boundary layer in the D″ region based on seismic ob-
servations. (A) The hybrid thermo-chemical boundary layer scenario for the deep mantle, in which sub-
ducting slabs penetrate to the core-mantle boundary, providing thermal and chemical anomalies that
will eventually rise back up in the mantle while hot dense chemical anomalies of either primordial or
slab-related origin are swept into large piles under upwellings. Either a phase change or a radial gra-
dient in structural fabric exists as well, giving rise to reflections from the top of D″. (B) The global
thermo-chemical boundary layer scenario, in which the lowermost mantle is a dense chemically dis-
tinct layer, possibly of primordial nature, which remains unmixed but thermally coupled to overlying
flow. Variable heatflow out of the chemical layer occurs in response to the configuration of mantle
flow, leading to lateral variation in thermal structure across the boundary layer. Topography is induced
on the chemical layer by the mid-mantle flow as well. Either of these scenarios may be overlain by
thermal boundary layer instabilities yielding plumes as in Figure 1, but at present the direct observa-
tional basis for them is weak. Modified from Lay et al. (2004).
from the D" layer and having smaller excess temperatures than would plumes arising directly from the core-mantle boundary. However, such plumes are not a necessary consequence of this scenario, and the majority of seismic observations can be accounted for without instabilities in the upper boundary layer.

Overall, the lowermost mantle appears to have substantial chemical and thermal heterogeneity, and this complex thermo-chemical boundary layer may plausibly serve as a source of thermal instabilities that rise through the mantle. However, a sober assessment of our current understanding and observations is that there are many unknowns with regard to the nature of this deep boundary layer, and it is quite plausible that there are, in fact, no thermal plumes rising from the region. Thus much more work will be required before we can attach much confidence to cartoon notions such as that of Figure 1.

ACKNOWLEDGMENTS

This paper was motivated in part by discussions at the 2004 CIDER (Cooperative Institute for Deep Earth Research) workshop, which was supported by the National Science Foundation (NSF) under Grant PHY99-07949. This research was supported by NSF grant EAR-0125595. Ed Garnero collaborated with the author on several studies from which graphics and ideas have been drawn. Don Anderson, Gillian Foulger, Bruce Julian, and Lianxing Wen provided helpful comments. Contribution no. 479, Center for the Study of Imaging and Dynamics of the Earth, Institute of Geophysics and Planetary Physics, University of California–Santa Cruz.

REFERENCES CITED


