Core-Mantle Boundary Structures and Processes

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Seismological and geodynamical observations have established the presence of a major thermo-chemical boundary layer (TCBL) in the lowermost mantle. This boundary layer plays a critical role in regulating heat flow through the core-mantle boundary, thereby influencing the dynamo-generating core flow regime. It also plays an important role in the mantle convection system, possibly serving as a source of boundary-layer instabilities and as a reservoir for long-lived geochemical heterogeneities. Two end-member conceptual models for the TCBL have emerged, both reconcilable with current observational constraints: a global, stably-stratified, chemically distinct layer may exist in the lowermost 250 km of the mantle (the global TCBL model), or this region may be a partially mixed boundary layer involving a composite of downwelling thermo-chemical anomalies such as oceanic lithospheric slabs or eclogitic oceanic crustal components and ancient dense chemical anomalies dynamically concentrated into large agglomerations beneath upwellings (the hybrid TCBL model). For the global TCBL model, laterally varying partial melt fractions within the layer are required to account for various seismological observations, and large dynamic topography on the upper boundary of this layer is expected: there is evidence for both of these attributes of the TCBL. The hybrid TCBL model requires additional complexity such as a phase transition or structural fabric transition to account for various seismological observations: some mineralogical candidates have been proposed. The outstanding challenge, requiring multi-disciplinary advances, is to discriminate between these competing conceptual models, as they differ in implications for thermal history, chemical processing, and dynamical behavior of the TCBL.

INTRODUCTION

The past two decades have witnessed tremendous progress in seismological, geodynamical, geomagnetic, and mineral physics efforts to quantify deep Earth processes. Our understanding of structures and processes in the lowermost mantle has advanced accordingly, and there is now widespread agreement that some form of major thermo-chemical boundary layer (TCBL) exists on the mantle side of the core-mantle boundary (CMB), extending at least several hundred kilometers upward into the lower mantle. Boundary layers play critical roles in heat transport and dominant length scales of thermal convection systems, so there is a concerted effort to understand the lowermost mantle boundary layer. The end-member conceptual models now being considered for the TCBL at the base of the mantle are reviewed here along with

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their observational foundations, and the directions of future multi-disciplinary research required to advance our understanding of CMB structures and processes are defined.

The CMB lies about 2900 km below Earth's surface, with this interface between mantle silicate and oxide rocks and molten iron alloy core materials being the primary internal compositional contrast within the planet. With density, viscosity, convective flow, and compositional contrasts comparable to or exceeding those at the surface of the Earth, the CMB separates the two major dynamical regimes of the interior; this makes it a place where chemical heterogeneities might be expected to accumulate. As such, from the early stages of core-formation and evolution of the primary chemical stratification of the planet to the current mantle convection system driving plate tectonics and the core flow regime generating the magnetic field, the boundary layers on either side of the CMB have played key roles in the chemical and dynamical evolution of the Earth. The multi-disciplinary advances in observational constraints on structures and processes near the CMB have been eagerly greeted by all disciplines engaged in understanding deep Earth processes. An assessment of our current state of knowledge (and ignorance) of this remote region is provided here, augmenting recent reviews by *Lay et al*. [2003], *Garnero* [2000], *Lay et al*. [1998a], and many papers in *Gurnis et al*. [1998].

CONCEPTUAL MODELS OF A DEEP THERMO-CHEMICAL BOUNDARY LAYER

By 1949 seismological observations were sufficient to establish that the lowermost mantle exhibits inhomogeneity relative to the overlying lower mantle, primarily manifested as decreased gradients in seismic velocities with depth. The bottom 200 km of the mantle were designated the D" region [*Bullen*, 1949], and it has become common to associate this region with a thermal and/or chemical boundary layer above the CMB. Subsequent radially averaged Earth models, such as the Preliminary Reference Earth Model (PREM) of *Dziewonski and Anderson* [1981], have incorporated reduced velocity gradients in the deepest 150 km of the mantle to accommodate the global departure of D" velocity structure from that of the overlying mantle, where velocity gradients are generally compatible with homogeneous self-compression. The presence of a thermal boundary layer (TBL) in D" caused by heat fluxing from the core into the mantle has long been postulated, and efforts have been made to infer the properties of such a boundary layer based on the reduced velocity gradients in models like PREM [e.g., *Stacey and Loper*, 1983]. However, as established by studies over the past half century that demonstrate increased seismic wave travel time fluctuations for paths traversing D", there appears to be substantial heterogeneity in this boundary layer on a wide variety of scale-lengths; a simple TBL interpretation is not sufficient to account for all seismically inferred properties of the region, nor is there a meaningful 'average' structure for the region to guide any physically viable interpretation of the boundary layer.

In the decades following the plate tectonics revolution, the importance of boundary layers to mantle and core dynamic systems became increasingly evident. It is well-recognized that the behavior of the plate tectonics system is largely governed by the surface thermo-chemical boundary layer: strong temperature-dependent viscosity effects and chemical differentiation play key roles in development of oceanic plates, while strong compositional variations play a key role in sustaining continental masses at the surface. The notion of hot plumes ascending from the interior to feed long-lived hotspot volcanic systems raised interest in the possible role of deeper boundary layers as the source region for plume genesis. Chemical anomalies and variance associated with many hotspot magmas relative to mid-ocean ridge basalts suggest the notion of deep boundary layers containing unmixed, isolated reservoirs within the mantle [cf., *Hofmann*, 1997; *Kellogg et al*., 1999]. While most of the ensuing work on internal boundary layers focused on the possibility of compositional and dynamical stratification of the upper and lower mantles, substantial interest was directed toward the D" region as a possible boundary layer source for thermal instabilities and chemical heterogeneities. The location of the D" boundary layer at the CMB, across which there has been a long history of chemical transport and where there is a massive density contrast, makes D" a logical site to accumulate chemically anomalous dregs from the mantle and dross from the core [e.g., *Anderson*, 1998]. Emerging notions of mantle-wide distributed chemical heterogeneities sampled by upwellings from deep boundary layers [e.g., *Davies*, 1990; *Helffrich and Wood*, 2001] sustain this interest in the possible existence of a deep mantle boundary layer, even if the mantle is not chemically stratified.

Advances in geophysical disciplines have yielded the current state of knowledge of the mantle dynamic system summarized in Figure 1a. The dynamical system near the surface is well-characterized as being comprised of a thermo-chemical boundary layer that is partially mixing, with production and recycling of oceanic lithosphere, gradual addition to the chemical heterogeneities of the buoyant and enduring continental crust, and diverse scales of upwelling beneath ridges and hotspot volcanoes. Substantial complexity of the phase equilibria of the upper mantle is recognized, particularly in the shallow mantle where abundant volatiles are likely to be present. But there is broad agreement that (a) global seismic velocity increases near depths of 410 and 660 km are likely due to phase transitions in the $(Mg,Fe)_2SiO_4$ system; (b) most com-

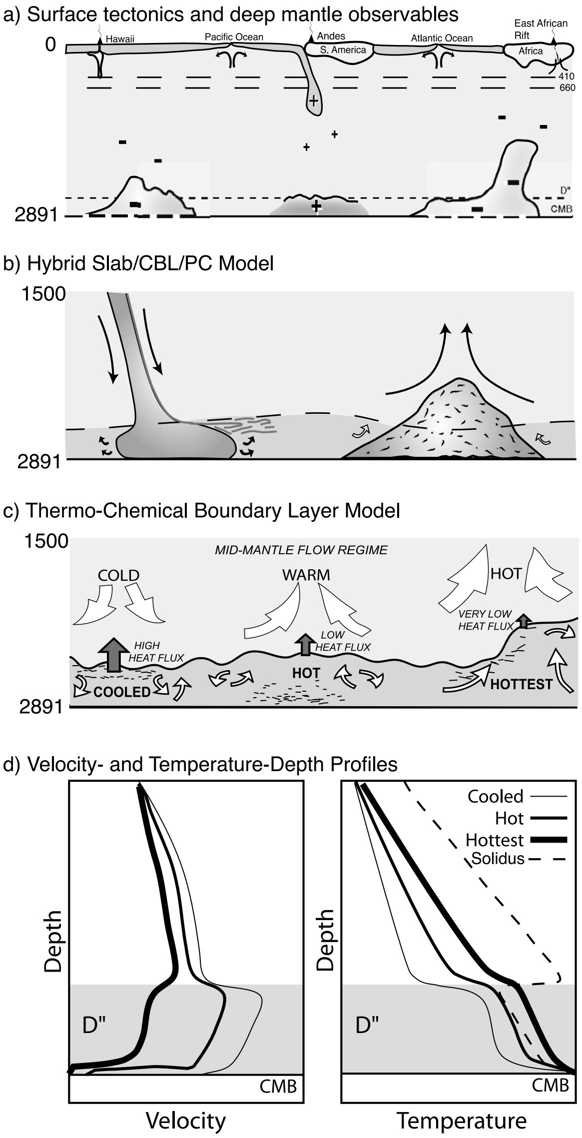

Figure 1. (a) Schematic cartoon of well-resolved attributes of the mantle convection system emphasizing the boundary layer near the surface comprised of the chemically differentiated continents, the chemically and thermally cycling oceanic slabs, and primary upper mantle phase transitions. The mid-mantle has large-scale patterns of seismic velocity heterogeneity with faster velocities (+ signs) tending to underlie regions of recent subduction of oceanic lithosphere, some of which may penetrate below the transition zone, and slow velocities (- signs) under the central Pacific and south Atlantic/Africa regions. The CMB boundary layer has large scale patterns of stronger heterogeneity which involves strong radial increases or decreases in velocity about 250 km above the CMB, with a predominant degree 2 pattern. (b) The hybrid thermo-chemical boundary layer (TCBL) concept for the deep mantle, in which subducting slabs penetrate to the CMB, providing thermal and chemical anomalies that will eventually rise back up in the mantle flow, while hot dense chemical anomalies are swept into large piles under upwellings. Either a phase change or radial gradient in structural fabric exists as well. (c) The global TCBL model, in which the lowermost mantle is a dense chemically distinct layer, possibly of primoridal nature, which remains unmixed, but thermally coupled to overlying flow. Variable heat flow 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. (d) Schematic elastic velocity profiles across the D" region for regions of the TCBL that have relatively hot or cool thermal structures. For the global TCBL model, it is assumed that the eutectic solidus is intersected by the thermal profiles over varying depth extent depending on the regional temperature level.

mon upper mantle minerals undergo a dissociative phase transition to $(Mg,Fe)SiO_3$ in perovskite structure and (Mg,Fe)O between 660 and 800 km or so; and (c) $(Mg,Fe)SiO_3$ and (Mg,Fe)O are the primary components of a relatively homogeneous lower mantle (with additional Ca-perovskite and other minor components). Dynamical and seismological constraints favor the notion of flux of at least some oceanic slab material into the upper portions of the lower mantle, with seismic images in some regions suggestive of deep penetration of advectively thickened masses of cold oceanic lithosphere to depths of 2000 km or deeper [e.g., *van der Hilst and Kárason*, 1999; *Kellogg et al*., 1999; *Fukao et al*., 2001; *Grand*, 2002]. The mid-mantle may have chemical stratification, but as yet, there is no clear detection of global layering, so we focus on the lowermost mantle boundary layer.

Global seismic tomography has established that large-scale patterns of mid- and deep-mantle heterogeneity have some correlation with the shallow mantle structures and circulation: large regions of relatively high seismic velocity mantle (presumably lower temperature, higher density, and thus, descending) tend to underly regions of substantial oceanic lithosphere subduction over the past 200 million years or so [e.g. *Lithgow–Bertelloni and Richards*, 1998], while large regions of relatively low seismic velocity mantle (presumably higher temperature, lower density, and thus, ascending) are located beneath regions replete with surface hotspot volcanism. These large-scale seismic velocity patterns increase in strength in the deepest mantle; the D" region exhibits predominant large-scale heterogeneity at spherical harmonic degree 2. The highest velocity areas of D" tend to be accompanied by fairly strong (1–3%) increases in seismic velocity with depth about 250 km above the CMB, while the lowest velocity areas have comparable decreases in velocity near the same depth. Overall lateral variations in seismic velocity for large scales in most tomographic models are on the order of ±3% for *S* velocity and ±1% for *P* velocity within the D" region, about a factor of 2 to 3 larger than for large-scale variations at mid-mantle depths. The increase in velocity heterogeneity and the existence of large-scale structures in D" support the probability that this regions serves as a major boundary layer within the interior.

Efforts to reconcile the general observations of Figure 1a (and their attendant details as discussed below) with characteristics of a boundary layer at the base of the mantle have resulted in different conceptual models for the nature of D". For the purpose of focusing discussion, we define two end-member scenarios for the boundary layer, and relate them to pertinent observational details in the next section.

The first is what we will call the hybrid TCBL model, involving large-scale chemical heterogeneities embedded in a partially mixed boundary layer that has many similarities to the surface thermo-chemical boundary layer (Figure 1b). In this hybrid model, subducting slabs descend to the D" region, cool large-scale regions beneath downwellings, and help to physically sweep aside dense lowermost mantle chemical heterogeneities [e.g., *Wysession*, 1996; *Grand*, 2002; *Tan et al*., 2002], which subsequently concentrate beneath upwellings. The upwellings include thermal boundary layer instabilities that give rise to plumes that ascend to the upper mantle. The pile of dregs can resist total entrainment if dense enough, but some chemical anomaly will be conveyed by the plume nonetheless. This model requires an additional aspect such as either a phase change or a strong vertical gradient in structural fabric imparted by shear flow to provide the abrupt radial seismic velocity increases at the top of the D" region observed in cooled areas.

An alternate scenario [*Lay et al*., 2003] that we call the global TCBL model invokes the notion of a global chemically distinct layer in D" that remains relatively stable beneath overlying mantle upwellings and downwellings. The chemical composition is not constrained, but could involve differences in relative amounts of iron, calcium, aluminum, and/or silica relative to the overlying mantle. The composition could represent primordial differences associated with heterogeneous accretion [e.g., Ruff and Anderson, 1980], accumulation of ancient dense subducted products [e.g., *Anderson*, 1998], or products of chemical reactions between the core and mantle [e.g., *Knittle and Jeanloz*, 1989; *Goarant et al*., 1992; *Dubrovinsky et al.*, 2001]. Substantial topography over a wide spectrum of wavelengths may be imposed on the layer by overlying mantle flow, including downwelling slabs in the mid-mantle and possibly plumes from the upper thermal boundary layer of D" (Figure 1c). Lateral variations in the overlying thermal system modulate heat flow out of the layer, resulting in large-scale lateral temperature gradients in the boundary layer that are thermally coupled to the mid-mantle, giving apparent continuity of seismic velocity heterogeneities. The lateral variations in temperature cause the boundary layer to either exceed or remain below the eutectic solidus at different lateral positions in the layer. The eutectic is probably reduced from that of the overlying mantle due to the distinct composition of the layer. This leads to laterally varying partial melt fraction within the layer. The melt itself must be effectively neutrally buoyant, remaining distributed across the layer but possibly with increasing melt fraction with depth. In turn, the variations in temperature and melt cause large-scale seismic velocity variations within the layer and variable velocity increases or decreases at the upper boundary of the layer (Figure 1d).

Both conceptual models explicitly involve lateral temperature variations in D", long-lasting chemical heterogeneity of the region, and seismological and dynamical complexity on a wide-variety of scales. However, there is potential for substantial differences between the strongly and partially stratified TCBL scenarios in, for example: (a) heat transport efficiency; (b) origin and nature of the chemical heterogeneities; (c) the role the TCBL plays relative to surface phenomena such as hotspots and the fate of slabs; (d) the thermal evolution of the system including the history of inner core growth; and (e) the extent of partial melting in the ancient lower mantle. The key observations underlying these competing end-member models will now be outlined, along with discussion of how each model may accommodate the observations. Future directions of research needed to discriminate between the models are then discussed.

OBSERVATIONAL CONSTRAINTS ON THE TCBL

Available probes of the structures and processes near the CMB have quite limited resolution, thus, characterization of the boundary layer must draw upon multiple lines of evidence. Given that seismological, geodynamical and geomagnetic information does not directly reveal the thermal structure of the deep mantle, even the existence of a TBL must be deduced indirectly. For example, the reduced seismic velocity gradients used to define the D" region could be a manifestation of chemical heterogeneity such as increasing iron content with depth rather than purely an effect of a superadiabatic temperature increase. When efforts are made to extrapolate experimentally constrained tie-points on temperature from the inner-core boundary and from upper mantle phase transitions in $(Mg,Fe)_2SiO_4$, the estimated superadiabatic temperature increase across the D" region is on the order of 1000–2000K [e.g., *Williams*, 1998; *Boehler*, 2000; *Anderson*, 2002]. This is comparable to the temperature contrast across the lithosphere, and favors the existence of a major TBL above the CMB. However, if superadiabatic thermal gradients are present shallower in the mantle, perhaps in the transition zone, mid-mantle or at the top of a global TCBL, this estimated temperature increase could be significantly reduced.

Another line of evidence is that the energetics of Earth's geodynamo require ongoing loss of heat from the core to the mantle, with estimates ranging from 2–10 TW annually [e.g., *Buffet* 2002, 2003; *Labrosse*, 2002]. These estimates are also somewhat dependent on assumptions about mantle stratification, the presence of which could reduce the estimates. While the uncertainties in both lines of evidence remain large, there is broad agreement, essentially now a paradigm, that a TBL with a significant overall temperature contrast exists at the base of the mantle. This raises the potential for TBL instabilities rising from the CMB, either within a stably stratified layer or as part of the larger mantle convection system. For the global TCBL model, one TBL is at the base of the chemically distinct layer (at the CMB), with a second TBL expected at its top. Such stratification gives large uncertainties in the actual temperature drop across the CMB, but does not negate the requirement of some heat fluxing through the CMB. For the hybrid TCBL model, the TBL at the base of the mantle is a more prominent feature of mantle convection because it serves as the lower boundary layer of the deep mantle convective system, but heat flow is still modified laterally by thermal and chemical heterogeneity.

Lateral Variations in the Boundary Layer

The case for lateral variations in the boundary layer, whether of thermal or chemical nature, is most compelling from the arena of global seismic tomography. There is now substantial convergence in large-scale mapping of seismic velocity heterogeneity in the deep mantle, particularly amongst shear wave models. Plate 1 presents comparisons of several recent global models for shear velocity variations (dVs) and compressional velocity variations (dVp) within the lowermost 250 km of the mantle; additional model comparisons are provided by *Garnero* [2000]. Large-scale shear velocity variations of ±3% are dominated by relatively high velocities beneath the circum-Pacific, with relatively low velocities under the central Pacific and south Atlantic/Africa [e.g., *Grand,* 2002; *Gu et al.,* 2001; *Ritsema and van Heijst*, 2000; *Mégnin and Romanowicz*, 2000; *Masters et al.,* 2000; *Kuo et al.*, 2000; *Castle et al.*, 2000]. The predominant degree 2 pattern in shear velocities is readily evident in Plate 1. The lowermost mantle portions of global compressional velocity models with ±1% velocity fluctuations tend to be less consistent, as apparent in Plate 1. Consistent features between *P* velocity models include fast regions beneath eastern Asia and Middle America, and slow regions under the South Pacific and the southern Atlantic [e.g, *Kárason and van der Hilst*, 2001; *Zhao*, 2001; *Fukao et al*., 2001; *Boschi and Dziewonski*, 1999, 2000; *Bijwaard et al*., 1998; *Vasco and Johnson*, 1998]. There is generally good correlation between dVp and dVs structures at very long wavelength [e.g*., Masters et al.*, 2000], although there are some well-sampled regions where the velocity variations decorrelate, such as within the Central Pacific and beneath North America (Plate 1). Several simultaneous inversions of *P* and *S* wave data have been performed with the intent of isolating bulk sound velocity variations from shear velocity variations [e.g., *Robertson and Woodhouse*, 1995; *Su and Dziewonski*; 1997, *Kennett et al*., 1998; *Masters et al*., 2000], but as yet there is little agreement amongst bulk sound velocity models.

The general expectation that shear velocity will be more sensitive to thermal variations than compressional velocity suggests that at least some of the large-scale regional pattern is the result of lateral temperature variations in the boundary layer of the order of several hundred degrees [e.g., *Forte and Mitrovica*, 2001; *Trampert et al*., 2001]. However, both large- and small-scale regions are found where P and S velocity anomalies do not correlate [e.g., *Saltzer et al*., 2001; *Wysession et al*., 1999]. There are also regions where shear velocity variations are positively correlated with, but much stronger than, compressional velocity variations (as in the south Pacific) [*Masters et al*., 2000; *Lay et al.,* 2003]. These observations require that any thermal variations be augmented by or competing with chemical or partial melting effects. For the hybrid TCBL model, high seismic velocity regions are associated with cooled regions where slabs have descended to the CMB, while low seismic velocity regions are hot piles of chemical dregs concentrated, but largely resisting entrainment under upwellings. The global TCBL model accounts for the large-scale patterns of seismic velocity by lateral variations in temperature (hence, in partial melt volume), resulting from thermal coupling with the overlying mid-mantle convection system. In either case, one expects lateral variations in thermal gradient above the CMB, affecting both mantle and core dynamics.

An indirect line of evidence favoring a thermal contribution to the seismic velocity heterogeneity comes from the nature of Earth's magnetic field, for which a few strong flux bundles in the northern and southern hemispheres appear to sustain relative stationarity beneath high seismic velocity regions in D" [e.g., *Gubbins*, 1998]. The very low viscosity of the core ensures that the CMB is nearly isothermal; however, the probable existence of lateral temperature variations within the D" region will result in lateral variations in the thermal gradient above the CMB, introducing a variable heat flow boundary condition on the core convection regime. This variable heat flux boundary condition can drive thermal winds in the core [e.g., *Bloxham and Gubbins*, 1987; *Zhang and Gubbins*, 1993] while possibly stabilizing large magnetic flux concentrations, and perhaps even influencing preferred paths of virtual geomagnetic poles (VGP) during reversals [cf., *Gubbins*, 1998]. The complexity of quantifying mantle-core thermal

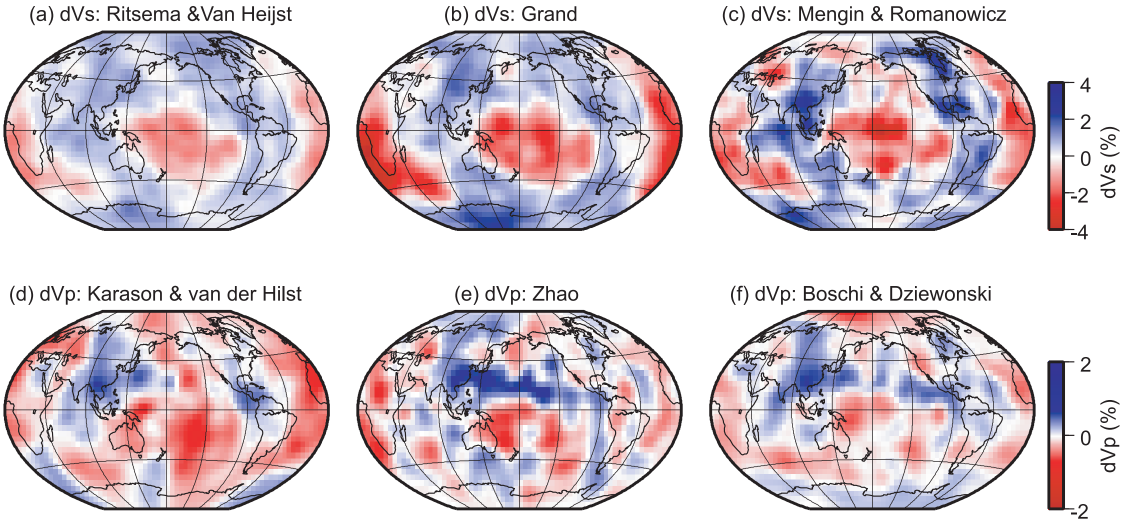

Plate 1. Maps of shear velocity (a–c) and compressional velocity (d–f) variations in the lowermost portions of several global mantle models. Shear velocity variations are all on a common scale, for the models of *Ritsema and van Heijst* [2000], *Grand* [2002], and *Mégnin and Romanowicz* [2000]. Compressional velocity variations are a different common scale for the models of *Kárason and van der Hilst* [2001], *Zhao* [2001], and *Boschi and Dziewonski* [2000].

interactions is enhanced by the possibility that chemical variations in D" may affect both thermal and electrical conductivity in the mantle [e.g., *Buffett*, 1992], as well as by the possible effects of CMB topography (which remains poorly determined), so this continues to be an area of active research [cf., *Buffett*, 1998].

Geomagnetic observations also provide one probe of a possible boundary layer on the core side of the CMB; particularly the possibility of a stably stratified outermost core layer [*Gubbins et al*., 1982; *Braginsky*, 1993]. Thermal buoyancy in such a thermally stratified layer would compete with effects of compositional buoyancy in the deeper core associated with expulsion of light alloying components upon solidification of core material at the inner core boundary [e.g., *Lister and Buffet*, 1998]. Such a TBL could defy seismological detection, but there have been several studies that suggest the presence of slightly anomalous compressional velocity gradients in the outermost 50–200 km of the core based on *SmKS* phases [*Lay and Young*, 1990; *Souriau and Poupinet*, 1991; *Garnero et al*., 1993b; *Tanaka and Hamaguchi*, 1993], and this remains an open issue even in the face of increasing complexity being recognized to exist in the mantle-side boundary layer [*Garnero and Lay*, 1998]. This is an important area for study given that the existence of any inhomogeneous structure in the outer core (generally assumed to be negligible) could trade-off with models for inner core structure. On a much finer scale, core-side boundary layer structure could involve ponding of buoyant light-alloying components under topographic highs in the CMB and development of finite rigidity in a very thin (<5 km) underplating layer [e.g., *Garnero and Jeanloz*, 2000; *Buffet et al*., 2000]. Possible observation of very localized structure less than 0.2 km thick has been presented by *Rost and Revenaugh*, [2001], sustaining interest in the possibility of a thin mushy layer of sediments accumulating on the CMB. As yet there is very little constraint on any core-side thermal/chemical boundary layer, so the remainder of this article will focus on mantle-side structure.

Local Stratification of the Boundary Layer

Complex locally layered seismic structures have also been detected on the mantle side of the CMB at both the top and bottom of the D" region. The shallower structure, typically from 150 to 350 km above the CMB, was first detected by array analysis of *P* waves [*Wright and Lyons*, 1979; *Wright et al*., 1985; *Weber and Davis*, 1990] and by analysis of profiles of *S* waves [*Lay and Helmberger,* 1983; *Young and Lay*, 1987b; *Young and Lay*, 1990]. *Wysession et al.* [1998] review the many subsequent studies that characterize this feature as a rapid increase in seismic velocity with depth, over a depth extent of 0–30 km, with 2–3% shear velocity increase and 0.5–3% compressional velocity increase. The depth of the velocity increase varies laterally over both large (>500 km) and short (<100 km) spatial scales [e.g., *Kendall and Shearer*, 1994; *Weber et al*., 1996; *Lay et al*., 1997]. This feature is often called a discontinuity, but the sharpness and lateral continuity of the increase remains important research topics. Figure 2b indicates regions where the most compelling observations (from detailed waveform analyses) of the shear velocity increase are found, relative to large scale patterns in shear velocity (Figure 2a, Plate 1). Generally, the regions with strong velocity increases are imaged by tomographic analyses as having higher than average shear velocity, as under Middle America, eastern Eurasia, and India; however, the Pacific has evidence for relatively small (0.5–1.5%) velocity increases in areas that are low velocity in the global tomographic models [*Russell et al.*, 2001].

There are intermittent or isolated regions of the lowermost mantle that are fairly well sampled by seismic waves where any shear velocity increase appears to either be very small or not present [e.g. *Weber and Davis*, 1990; *Kendall and Nangini*, 1996; *Garnero and Lay*, 2003]. When considering all regions sampled, the statistical correlation between shear and compressional velocity increases and large-scale tomographic patterns is actually quite low [*Wysession et al.*, 1998]. This requires further investigation and assessment of the reliability of isolated detections based on waveform complexity. This is particularly true for *P* waves, as array processing of large-numbers of observations is required to confidently detect, or rule-out, small velocity increases of 0.5% or so, especially if distributed over some tens of kilometers radially. This is the case even at grazing incidence where the phases are amplified by triplication effects [e.g. *Reasoner and Revenaugh*, 1999]. The relative infrequency of clear short-period reflections from the top of D" for steeply incident waves [e.g., *Persch et al.*, 2001; *Castle and van der Hilst*, 2003] tends to favor either strong lateral variations in the D" discontinuity or obscuring effects such as a gradational transition zone or small-scale topography on the feature. Procedures embedding assumptions of one-dimensional reference models and horizontal reflectors may give incorrect estimates of actual structures.

The global TCBL model interprets the observed *P* and *S* velocity increases as the upper boundary of the chemically distinct layer, which is expected to be the site of a TBL as well. The topography on the boundary is expected as a result of dynamic loading by mid-mantle flow and possibly by internal convection of the boundary layer, with depressed discontinuity depths below downwellings, and elevated discontinuities under upwellings, both with high variability. Lateral variations in observability of the discontinuity caused by the intrinsic chemical contrast of D" are explained as the result of both topography and gradient of the thermal-chemical contrast at the top of the boundary, compounded by lateral variations in degree of partial melt within the layer, which has a profound affect on seismic velocities. Cooled regions under downwellings are expected to be sub-solidus, and thus have the

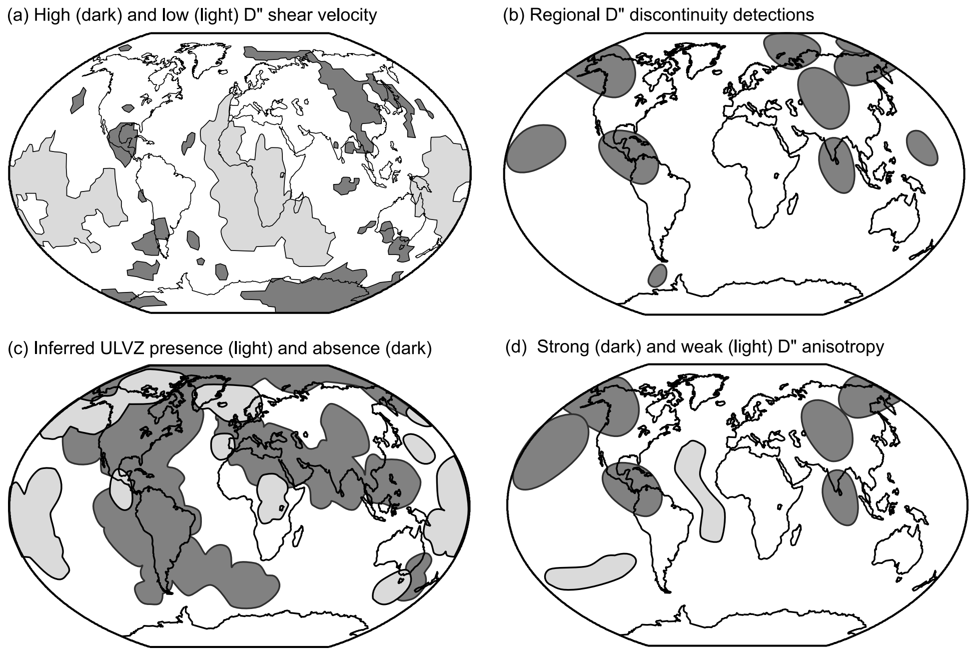

Figure 2. (a) Large-scale shear-velocity variations in the lowermost 250 km of the mantle from the global model of *Grand* [2002]. Dark shaded regions have shear velocity anomalies 1.2% or more faster than average, while light shaded regions have shear velocity anomalies 1.2% or more slower than average. (b) Shaded zones highlight regions where rapid shear velocity increases of 1.5–3% have been observed at depths near 250±50 km above the CMB. (c) Distribution of ultra-low velocity zone observations (light shade) and non-observations (dark shade). (d) Locations where relatively strong (>1%) vertical transverse isotropy have been detected in D" (dark shading) and where weak or non-existent anisotropy has been observed (light shading).

strongest positive velocity increases due to the absence of any competing partial melt component. Warmer regions under upwellings should have higher degrees of melt and hence weaker velocity increases, or possibly even decreases.

The hybrid TCBL model accounts for the velocity increases at the top of D" as the result of thermal anomalies of slab materials combined with an unspecified phase change to sharpen the velocity increase [e.g., *Sidorin et al*., 1999]. The notion of a global phase change to account for the discontinuity has been around for a decade [*Nataf and Houard*, 1993], but a clear candidate for the phase change has not been established amongst the predominant Mg-perovskite and ferropericlase minerals expected in the lower mantle [see *Wysession et al*., 1998]. Recent work by *Badro et al*. [2003] has demonstrated a possible change in the spin state of (Mg,Fe)O, from high-spin to low-spin for lower mantle pressures. This could favor iron enrichment in this mineral and iron depletion of silicate perovskite in the lowermost mantle, with possible attendant viscosity and thermal transport effects, but probably only a minor seismic velocity or bulk density effect. Another candidate effect for producing a radial increase in velocity in the hybrid model is a gradient in fabric into the boundary layer, with the stress increase and cooling of the mid-mantle downwelling allowing dislocation creep processes to develop lattice preferred orientation (LPO) in the (Mg,Fe)O component. *McNamara et al.* [2001] have presented models where suitable conditions are predicted below slabs, even if they do not descend all the way to the CMB. Anisotropy in D" is discussed further below.

Complex Mantle-Core Transition Zone

Evidence for very strong velocity contrasts just above the CMB dates back to work on spectra of core reflections and diffracted wave velocities in the 1960's and 1970's [see a review by *Young and Lay*, 1987a], but much more compelling evidence for a transition zone at the CMB emerged from studies of *SPdiffKS* waves [e.g., *Garnero et al*., 1993; *Garnero and Helmberger*, 1996], *PcP* precursors [e.g. *Mori and Helmberger*, 1995; *Revenaugh and Meyer*, 1997]; and *PKP* precursors [e.g., *Wen and Helmberger*, 1998; *Vidale and Hedlin*, 1998]. These studies, and many since (see *Garnero et al*. [1998] for a review) demonstrate that in some regions, 10–40 km thick zones have 5–10% low compressional velocities and 15–30% low shear velocities right above the CMB, either in horizontally extensive layers or concentrated into blob-like domes. The magnitude of the velocity reductions and the large ratio between shear and compressional velocity variations favor an interpretation of these so-called ultra-low velocity zones (ULVZs) as regions of significant (6% to 30%) partial melt volume [*Williams and Garnero*, 1996]. The spatial extent of ULVZ detections and non-detections is indicated in Figure 2c. Non-detections are difficult to appraise, as the structure may be too thin or too gradational to give rise to clear waveform complexity, but there are regions where there is at least no evidence supporting significant ULVZ presence [e.g. *Castle and van der Hilst*, 2000; *Persch et al.*, 2001]. Figure 2 demonstrates a general correlation between major ULVZ regions and large-scale patterns of low shear velocity in D", although there are some exceptions such as under Central America. The interpretation that ULVZ material is partially molten is shared by the two end-member TCBL models, but in the hybrid model localized chemical anomalies may account for the finite extent of the ULVZ zones. For the global model it is also possible that lateral chemical heterogeneities contribute to the spatial pattern, but it may simply be that the ULVZ is seismically detectable in the hottest regions of the layer, where the steep thermal gradient in the CMB TBL exceeds the solidus over tens of kilometers rather than over only a few kilometers. In cool regions with a steep thermal gradient over a very narrow depth range, it may be very difficult to detect any ULVZ even with short-period reflected waves.

The dramatic velocity reductions invoked to explain the ULVZ observations motivate consideration of partial melting in the lowermost mantle, the presence of which has profound implications for seismic velocity heterogeneity, viscosity structure of the boundary layer, and chemical processing in the boundary layer [*Lay et al*., 2003]. Experimental constraints on end-member perovskite and ferropericlase mineralogies of the deep mantle suggest that each may have much higher melting temperatures in D" than the likely upper bound on CMB temperatures [*Zerr and Boehler*, 1993, 1994]; however, for a multi-component eutectic system a much lower solidus temperature is likely. *Zerr et al.* [1998] and *Boehler* [2000] extrapolated a pyrolite composition solidus to CMB conditions, estimating a melting temperature of 4300 K, within the upper bound of estimates of CMB temperatures. Additional chemical heterogeneity, associated with subduction products, core-mantle reaction products, or ancient chemical stratification could lower the melting temperature within D" even further (Figure 1), accounting for laterally varying melt fraction within the layer and/or partial melting in the basal TBL to form the ULVZ. The few available constraints on liquid-solid partitioning of iron in silicates at high pressure [see *Knittle*, 1998], favor iron concentration into the melt phase, which provides a density effect that can stabilize melts in the boundary layer, or possibly lead to their accumulation in the ULVZ due to negative buoyancy [see *Lay et al*., 2003]. At this point, partial melting of either the lowermost or all of the hottest regions of D" must be considered a possibility. Detailed investigation of seismic attenuation properties of the ULVZ could help to constrain the structure.

Low Velocity Provinces

Improvements in global distribution of seismometers have enabled increasing focus on detailed structure in the low velocity regions under the Central Pacific and Africa [e.g., *Tanaka*, 2002; *Wen et al*., 2001; *Bréger et al*., 2001]. Both regions appear to have strong lateral gradients, with shear velocity variations involving abrupt 3–5% reductions and low velocity zones several hundred kilometers in thickness. Extensive regions of the South Atlantic have a velocity decrease at the top of the low velocity region, with 1–3% drop near 200–250 km, at about the same average depth as the 2–3% increases in circum-Pacific regions [e.g., *Ni and Helmberger*, 2003; *Wen* 2002]. The low velocity region extends upward 500–800 km into the mid-mantle, still with sharp lateral gradients, under Africa [*Ni et al.*, 2002], suggesting large topography on the low velocity body. The Atlantic/Africa region appears to have shear/compressional velocity ratios compatible with partial melting [e.g., *Simmons and Grand*, 2002; *Tkalcic and Romanowicz*, 2001], while the central Pacific region may involve anomalous ratios requiring chemical effects as well as partial melting [e.g., *Masters et al*., 2000; *Lay et al.,* 2003]. In the hybrid TCBL model, the large low velocity regions under the southern Pacific and southern Atlantic/Africa are considered chemical superplumes; where hot, dense chemical heterogeneities have piled up under upwellings (Figure 1b). The chemical anomaly accounts for part of the velocity decrease and the sharp edges of the structure. In the global TCBL model, these are the thickest, hottest regions of the boundary layer, with the most extensive fractional melting across the chemical layer, resulting in relatively abrupt seismic velocity decreases across the layer interface (Figure 1d). Relatively small melt fractions (0.5–2%) can produce the strong velocity reductions needed to account for these structures [*Lay et al*., 2003]. In both cases the large topography is induced by mantle flow above hot, but dense material.

Boundary Layer Anisotropy and Scattering

Seismic anisotropy has been demonstrated to exist in the D" region for quite some time (see *Lay et al.* [1998] and *Kendall* [2000] for reviews). The general observations support widespread occurrence of anisotropy compatible with vertical transverse isotropy (VTI) (Figure 2d), primarily below circum-Pacific, relatively high shear velocity regions, with weak anisotropy beneath regions of moderate shear velocity anomalies. Intermittent azimuthal anisotropy is found beneath the low velocity central Pacific. Anisotropy appears to have increased strength and spatial coherence in the boundary layer relative to the overlying mid-mantle. Recognizing that the boundary layer is likely to have relatively strong shear flows and lateral temperature variations, a mix of possible anisotropy mechanisms have been proposed, ranging from shearing of partial-melt components in horizontal or vertical flows [e.g., *Kendall and Silver*, 1998; *Russell et al.,* 1998] to stress-induced LPO in (Mg,Fe)O in low temperature regions where dislocation glide mechanisms are active [e.g., *Yamazaki and Karato*, 2002; *McNamara et al.*, 2001, 2003]. Both end-member TCBL models can accommodate either form of anisotropy as a result of boundary layer shearing of chemical heterogeneities, partial melt blobs, or mineral alignments. Observations of coupling between the shear velocity increase at the top of D" and an onset of anisotropy [e.g., *Matzel et al.,* 1996; *Garnero and Lay*, 1997] suggest that either acquisition of preferred fabric is responsible for the discontinuity or that the chemical change giving rise to the discontinuity produces favorable conditions for the development of the fabric. Further characterization of D" anisotropy through more detailed seismic analyses is needed to establish its role in the TCBL.

A further line of evidence pertaining to structures and processes near the CMB is the presence of very small-scale seismic heterogeneity, manifested in the scattered wavefield accompanying deeply penetrating seismic phases. Scattering of short-period *P* waves by structural heterogeneity, including possible CMB topography, indicates that scale-lengths of a few to tens of kilometers have velocity fluctuations of a few to ten percent [e.g., *Bataille and Lund*, 1996; *Cormier*, 2000; *Earle and Shearer*, 1997; *Hedlin and Shearer*, 2000]. In some cases the scattering may arise from the ULVZ [e.g., *Vidale and Hedlin*, 1998; *Wen and Helmberger*, 1998 *Niu and Wen*, 2001; *Rost and Revenaugh*, 2003], and in a few cases it can be imaged by scattering migrations [e.g., *Thomas et al.*, 1999; *Rost and Thomas*, 2003]. The overall spectrum of heterogeneity of the boundary layer is not well determined yet, but it appears to be relatively red, with substantial power at long wavelengths and moderate power at short wavelengths, possibly with an anisotropic spatial distribution [e.g., *Cormier*, 1999]. The two end-member TCBL models both involve thermal and chemical heterogeneities that can account for a reddened heterogeneity spectrum, with the global model having the added degrees of freedom provided by strong velocity effects of distributed fractional melts. Both models also admit the possibility of CMB topography on a variety of scale-lengths, which could account for much of the scattering in the region due to the strong density and velocity contrasts involved. Determining CMB topography has proved very challenging and there is little agreement amongst recent models [e.g., *Garcia and Souriau*, 2000; *Sze and van der Hilst*, 2003]. Improved characterization of the spatial pattern of small-scale heterogeneities could help to constrain specific causal mechanisms within the boundary layer.

DISCUSSION AND CONCLUSIONS

Table 1 summarizes the primary observational constraints on the D" region and the various ways that these constraints have been built into the two competing models we discuss in this paper. With so few hard-constraints on viable chemistry, melting, and dynamical structures in the boundary layer, the end-member models are sufficiently flexible to accommodate most observations with reasonable degree of plausibility. Particularly challenging for the global model is to account for the reversal in sign of the velocity contrast across the top of the layer from positive in high velocity areas to negative in low velocity areas. However, the dramatic effect of small amounts of partial melt, suitably distributed in the boundary layer material, provides a possible mechanism. The absence of a universal sharp reflector is also a challenge for this model, but the demonstrated presence of strong topography, along with the viability of a gradational transition zone with superimposed effects of a TBL may reconcile this constraint. The hybrid model struggles to account for the presence of rapid velocity increases at the top of D", as such are not readily produced in thermal models of subducting slabs [e.g., *Sidorin et al.*, 1999], and this leads to the need for an additional effect, such as a phase change (of unknown type), or perhaps an

Table 1. Observational Characteristics of the CMB TCBL and How End-Member Models Account for Them

Feature	Characteristics	Hybrid TCBL Model	Global TCBL Model
High Heat Flux Values Across CMB	Thermal model based estimates of CMB heat flux of as much as 10 TW	Hot thermal boundary layer instabilities	Layering would lower heat flux estimates
Mantle/Magnetic Field Correlations	Strong localized quasi-stationary magnetic flux concentrations under regions of high seismic velocity	Slab cooling produces long-term heat flux boundary condition	Thermally modulated layer produces long-term heat flux boundary condition
Large-scale Vs and Vp Heterogeneity	±3-4% dVs, ±1% dVp, degree 2 dominance ±1% bulk sound velocity circum-Pacific high velocity C. Pacific/S. Atlantic low velocity strongest in lowermost 300-400 km	Thermal origin in large-scale flow regime, slabs cause high velocity, chemical heterogeneity in slow regions, superplumes	Partial melt variations within chemical layer temperatures modulated by thermal coupling chemical layer has bulk sound velocity anomaly
D" Velocity Increases	+1.5-+3% in Vs, +0.5-+3% in Vp 0-30 km thick transition zone in high velocity regions, weaker in low velocity regions, depth varies from 130 to 300+ km above CMB	Slab related with a phase change, gradient in fabric, gradient in heterogeneity spectrum	Intrinsic to chemical contrast, varying partial melt fraction modulates, dynamic topography on chemical layer
Simple ScP Phases	Widespread observations of steeply incident CMB reflections with no high frequency precursors and simple waveforms	Pristine areas between slab piles and low velocity chemical blobs	Fuzzy upper boundary of chemical layer due to topography and/or chemical/thermal gradients, ULVZ very thin in cooled regions
ULVZ	5-10% Vp reductions, 10-30% Vs reductions, 5-40 km thickness sometimes sharp onset (< few km) best developed in low velocityregions, but present in high velocity regions	Partial melt in basal boundary layer in regions with piles of chemical heterogeneities,	Partial melt in basal boundary layer due to laterally modulated temperature profile across chemical layer
Low Velocity Blobs	3-5% slow shear velocity over 250-800 km thick zones, with sharp (-1 to –3%) decreases at top and laterally	Chemical heterogeneities, hot, dense piles under up-wellings	Hottest, highest melt fraction regions of TCBL thick under upwellings
D" Anisotropy	Large scale 1-2% in boundary layer, primarily with VTI symmetry in high velocity regions. variable strength/symmetry in low velocity regions. absent from average regions.	LPO in MgO, SPO in chemical/melt, lamellae from sheared chemical anomalies	LPO in MgO SPO in chemcial/melt lamellae from sheared chemical anomalies
D" Scattering	1-10% rms heterogeneity in boundary layer, 10-100 km scale lengths related to ULVZ, change in spectrum relative to mid-mantle	CMB topography, chemical/melt variations, thermal explanation not likely without melt, slab crustal components	CMB topography chemical/melt variation small scale convection

onset of anisotropy in the boundary layer due to the stresses and cooling of the downwelling slab, as in the models of *McNamara et al*. [2003].

The evidence for chemical heterogeneity above and beyond thermal effects, including partial melting, is rather compelling. The strong gradients laterally and radially into the low velocity zone beneath the south Atlantic/Africa are hard to account for thermally without some chemical contribution. While abundant dense partial melt could play some role, sustaining strong lateral gradients in upwelling material for hundreds of kilometers appears problematic. The anomalous bulk sound velocity measurements found for the central Pacific also appear to require a compensating bulk modulus perturbation even if partial melting accounts for the strong shear velocity reductions in the region. This could still be partially accounted for by anisotropy, so further constraints on the cause and orientation of anisotropy are needed. Small-scale fluctuations are unlikely to be due to thermal heterogeneity unless the temperatures are right at the solidus (as proposed by the global TCBL model). Of course, partial melting itself should give rise to chemical heterogeneity, such as iron fractionation into the melt, so one cannot truly separate melting from chemical heterogeneity.

Dynamical models for global and hybrid TCBL structures comprised of primordial components, core-mantle chemical reaction products, or segregated subduction products have been explored quite extensively in the past few years using two- and three-dimensional mantle convection codes. Some of the key issues are the evolution of a TCBL and the viability of chemical heterogeneities denser than normal mantle material accumulating at the base of the mantle and surviving entrainment by mid-mantle flow. Most calculations indicate that a density contrast of 3–6% relative to the overlying mantle is required to sustain a coherent global layer [e.g., *Christensen*, 1984; *Sleep*, 1988; *Kellogg and King*, 1993; *Kellogg*, 1997; *Sidorin and Gurnis*, 1998; *Montague and Kellogg*, 2000, *Montague et al*., 1998]. The requisite density contrast for stability of the layer may actually be much lower, on the order of 0.5–1% when allowance is made for compressibility effects and strong temperature dependence of thermal expansion (the buoyancy number must be computed for the local, reduced values, which is not usually done in the literature), along with effects such as reduction of viscosity in the boundary layer caused by temperature-dependent viscosity [e.g., *Schott et al*., 2002]. If the density increase of the chemical heterogeneities is too low to maintain a coherent layer, the material will be concentrated into patches under upwellings [e.g., *Davies and Gurnis*, 1986; *Hansen and Yuen*, 1989, *Manga and Jeanloz*, 1996; *Tackley*, 1998; *Davaille*, 1999; *Gonnermann et al*., 2002]; a means by which an initially global TCBL situation could have evolved into a hybrid TCBL today. Various dynamical models differ in the extent to which subducting slabs reach the deepest mantle, whether crustal components can separate from the slab, and the extent of thermal and viscous disruption and induced topography of the boundary layer that takes place [e.g., *Christensen and Hofmann*, 1994; *Sidorin and Gurnis*, 1998; *Tackley*, 2000; *McNamara et al*., 2001; *Coltice and Ricard*, 1999].

There have also been numerous geodynamic explorations of the possible role of D" as a source of thermal plumes, and their potential to bring up chemical heterogeneities from the deep boundary that can account for geochemical anomalies in ocean island basalts [*Hofmann and White*, 1982; *Albarede and van der Hilst*, 1999]. The presence of chemical heterogeneity in the boundary layer can affect the stability and distribution of thermal plumes rising from within or above the boundary layer [e.g., *Kellogg and King*, 1993; *Farnetani*, 1997; *Jellinek and Manga*, 2002] as well as the overall heat transport across the boundary layer [e.g., *Namiki and Kurita*, 2003]. The extent to which entrained boundary layer materials are mixed in ascending plume shear flows has also been examined [e.g., *Farnetani and Richards*, 1995]. Shear flow in the boundary layer likely plays a major role in the development of seismic wave anisotropy [*McNamara et al*., 2001; *Lay et al.,* 1998], as well as possibly contributing to ULVZ formation and/or growth by shear heating [*Steinbach and Yuen*, 1999]. At present, the calculations support the viability of plumes sampling chemical heterogeneities either within or from the top of the boundary layer and bringing them to the surface, but this has not yet been demonstrated to occur.

Advancing our understanding of structures and processes near the CMB will require observational, laboratory and modeling advances across several disciplines. Figure 3 highlights some of the major boundary layer features that require improved observational constraints; including aspects of D" discontinuities, D" anisotropy, large-scale low velocity zones, ULVZs, CMB topography, and attendant dynamical issues. It is also clearly of great importance to establish whether there is any density increase in the lowermost mantle along with whether there are density anomalies associated with large low velocity provinces. Preliminary work with normal modes suggests that low velocity regions may have anomalously high density [e.g., *Ishii and Tromp*, 1999], but the resolution of normal mode approaches remains limited [e.g., *Romanowicz*, 2001; *Kuo and Romanowicz*, 2002]. Many of the topics noted in Figure 3 require improved seismological imaging and modeling, which includes better 2- and 3-dimensional wave propagation approaches as well as more advanced array methods for characterizing subtle features in the wavefields. There are tremendous limitations imposed by the geometry of sources and receivers that need to be addressed by innovative data collection strategies.

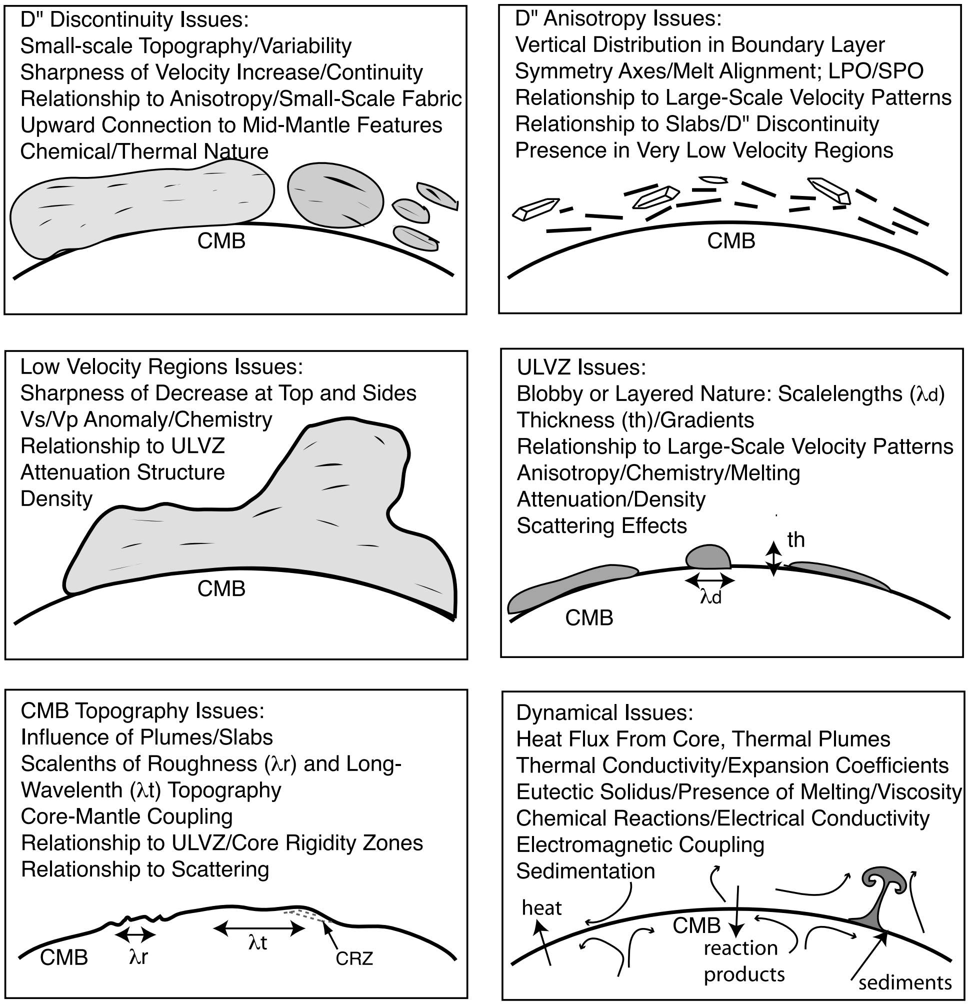

Figure 3. Summary of principal research topics associated with major features of the CMB TCBL including the D" discontinuity, D" anisotropy, large-scale low velocity regions, ULVZ, CMB topography, and associated dynamical issues. Progress on each area requires advances in seismology, mineral physics, geodynamics, and possibly geomagnetism.

While numerically challenging at present, dynamical calculations need to account for three-dimensional, compressible flow with the possibility of small-scale partial melting. The strong viscosity reduction upon melting and the multi-scale nature of partial melting present formidable challenges. Quantification of boundary layer shear flows that might account for anisotropy by mineral or melt/chemical inclusion alignment is also needed. Enhanced mineral physics experimental constraints on the stability and iron spin-state of silicate perovskite and ferropericlase for D" conditions are needed. Exploration of the eutectic melting behavior of plausible high pressure lower mantle assemblages is also needed. All of these needs are at the frontiers of current technologies, and concerted effort will be required to resolve the issue of a strongly stratified or dynamically disrupted TCBL above the CMB. Establishing the current configuration of deep mantle structure is prerequisite to extrapolating back in time to an earlier, hotter Earth system, and to understanding the thermal evolution of the core [e.g., *Buffett*, 2003], the extent of partial melting of the ancient mantle, and variations in the role played by D" through time.

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