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Hawaii, Boundary Layers and Ambient Mantle—Geophysical Constraints

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Recent high-resolution seismic observations and geodynamic calculations suggest that mid-plate swells and volcanoes are plausibly controlled by processes and materials entirely in the upper boundary layer (<220 km depth) of the mantle rather than by deep-seated thermal instabilities. The upper boundary layer (BL) of the mantle is fertile enough, hot enough and variable enough to provide the observed range of temperatures and compositions of mid-plate magmas, plus it is conveniently located to easily supply these. Seismic data show that the outer ~220 km of the mantle is heterogeneous, anisotropic and has a substantially superadiabatic vertical temperature gradient. This is the shear, and thermal, BL of the upper mantle. It is usually referred to as the 'asthenosphere' and erroneously thought of as simply part of the well-mixed 'convecting mantle'. Because it supports both a shear and a thermal gradient, the lower portions are hot and move slowly with respect to the surface and can be levitated and exposed by normal plate tectonic processes, even if not buoyant. The nature of BL anisotropy is consistent with a shear-induced laminated structure with aligned melt-rich lenses. The two polarizations of shear waves travel at different velocities, V_{SV} *and* V_{SH}*, and they vary differently with depth.* V_{SH} *is mainly sensitive to the temperature gradient and indicates a high thermal gradient to 220 km depth.* V_{SV} *is mainly sensitive to melt content. The depth of the minimum isotropic shear velocity,* V_s*, under young plates occurs near 60 km and this rapidly increases to 150 km under older oceanic plates, including Hawaii; 150 km may represent the depth of isostatic compensation for swells and the source of tholeiitic basalt magmas. The high-velocity seismic lid thickens as the square-root of age across the entire Pacific, but the underlying mantle is not isothermal; average sub-ridge mantle is colder, by various measures, than mid-plate mantle. Ambient mantle potential temperature at depth under the central Pacific may be ~200°C higher, without deep mantle plume input, than near spreading ridges. This is consistent with bathymetry and seismic velocities and the temperature range of non-ridge magmas. Some of the thinnest and, in terms of traditional interpretations, hottest transition zones (TZ; ~410–650 km depth) are under hotspot-free areas of western North America, Greenland, Europe, Russia, Brazil and India. The lowest seismic velocity regions in the upper mantle BL are under young oceanic plates, back-arc basins and hotspot-free areas of California and the Pacific and Indian oceans. Cold slabs may displace hotter material out of the TZ but geophysical data, and geodynamic simulations, do not require deeper sources. Magmas extracted from deep in a thick conduction layer are expected to be hotter than shallower oceanic ridge magmas and more variable in temperature. Mid-plate magmas appear to represent normal ambient mantle at depths of ~150 km, rather than very localized very deep upwellings. Shear-driven upwellings from the base of the BL explain mid-plate magmatism and its association with fracture zones and anomalous anisotropy, and the persistence of some volcanic chains and the short duration of others. The hotter deeper part of the surface BL is moving at a fraction of the plate velocity and is sampled only where sheared or displaced upwards by tectonic structures and processes that upset the usual stable laminar flow. If mid-plate volcanoes are sourced in the lower half of the BL, between 100 and 220 km depth, or below, then they will appear to define a relatively fixed reference system and the associated temperatures will increase with depth of magma extraction. Lithospheric architecture and stress control the locations of volcanoes, not localized thermal anomalies or deep mantle plumes.*

KEY WORDS: *asthenosphere; boundary layer; Hawaii; lithosphere; mantle temperature; plumes; Transition Zone*

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INTRODUCTION

It is still debated whether plates and subducted slabs drive mantle flow from above (Elsasser, 1969; Hager & O'Connell, 1979, 1981) or whether the plates are passive passengers that simply record the surface motions of deep mantle convection (Tackley, 1998, 2006; Phillips & Bunge, 2005; Nakagawa *et al.*, 2008). The contrasting views are that (1) the dominant forces that drive plate tectonics derive from cooling and subducting plates, and other body forces operating on the outer shell, and (2) large-scale motions of the mantle are driven by internal and core heat sources and plate tectonics is simply the most visible manifestation of mantle convection. What distinguishes these models is the nature of the coupling between the plates and the deeper mantle, the level of shear stress at the base of the plates, and the thickness and anisotropy of the boundary layer (BL; Fig. 1). The BL-driven flow model is consistent with a number of tectonic, geophysical and geochemical observations, including the coherent motions of large plates, dips of slabs, large-scale upper mantle anisotropy and the existence of shallow recycled and trapped components (e.g. Hager & O'Connell, 1979, 1981; Kay, 1979; Tanimoto & Anderson, 1984; Tommasi *et al.*, 1996, 2006; Spakman & Wortel, 2004; Doglioni *et al.*, 2005; Simon *et al.*, 2008). Although mantle convection and plate tectonics can be regarded as two aspects of the same coupled system (Tackley, 1998) they can also be regarded as far-from-equilibrium self-organized thermodynamic systems that derive energy, material and information from each other (e.g. Anderson, 2007*a*).

In the top-down or BL model, plates and slabs organize mantle convection, as well as themselves. The mantle is the source and sink of matter, most of which is shallowly recycled, and of energy, but the motions of the boundary layer drive motions in the interior. In laboratory Rayleigh–Bernard, heated-from-below convection simulations, with constant fluid properties, the outside world provides the source and sink of energy and the fluid is the self-organizing system. In the case of the mantle, most of the heat that is conducted through the surface, and lost to space, is internally generated. External sources (core heat) and sinks (secular cooling) play less of a role, and pressure plays a larger role, than in laboratory and most computer simulations. When the physical and thermal properties depend strongly on temperature and pressure, the upper boundary layer is more active, and the lower boundary layer is less active (more sluggish), than in the usual Boussinesq approximation. Parts of the upper boundary layer are resistant to subduction and recycling and can therefore build up and preserve significant geochemical (including isotopic) anomalies.

The past several years have seen a considerable improvement in the resolving capability and self-consistency of geodynamic simulations and in the ability to model

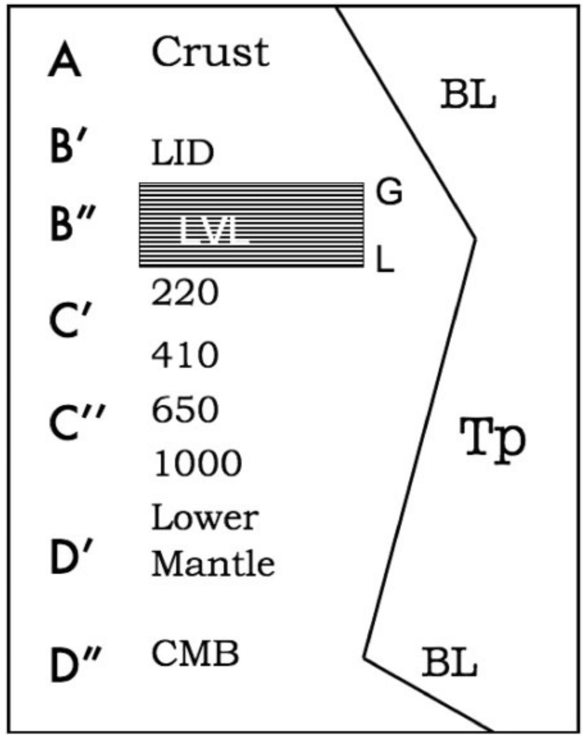

Fig. 1. Nomenclature of the mantle. Region B (Gutenberg, 1959) includes the lid and the laminated boundary layer. The lower mantle (Region D) starts below ~900 km. The Transition Region was defined originally as the mantle between 410 and ~900 km. The Transition Zone (TZ) is the region between 410 and 650 km. The low-velocity anisotropic layer (LLAMA) extends from the Gutenberg (G) discontinuity to the Lehmann (L) discontinuity. A schematic variation of the potential temperature geotherm (T_p) with depth is shown, along with the upper mantle and lower mantle Boundary Layers (BL). The T_p geotherm is the actual geotherm minus the adiabatic gradient. BL geotherms follow the conduction gradient, which is high enough to cause V_s to decrease with depth (e.g. Anderson, 1965; Stixrude & Lithgow-Bertelloni, 2007). Over most of the mantle the geotherm is subadiabatic, meaning that T_p in D" can be less than in the upper mantle.

boundary layer scale phenomena (e.g. Coltice *et al.*, 2007; Schuberth *et al.* 2009; Adam *et al.*, 2010; Ballmer *et al.*, 2010; Conrad *et al.* 2010; Faccenna & Becker, 2010; Schmandt & Humphreys, 2010). This modeling shows that many features that have been attributed to the deep mantle are actually shallowly rooted, consistent with high-resolution seismic imaging and geochemistry (e.g. O'Reilly & Griffin, 2006; O'Reilly *et al.*, 2009) and with the petrological model of Kay (1979). This study explores the implications of these new geodynamic and seismological results and revisits the issues of mantle anisotropy and ambient mantle temperature.

Lateral advection of mass

The lateral motion of a plate sets up a velocity gradient in a boundary layer between the surface and the deep

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mantle that results in laminar flow and seismic anisotropy. Such a shear boundary layer is not the same as the lithosphere, asthenosphere or 'the convecting mantle'. Although the layer is being sheared and deformed, heat is transferred mainly by conduction. This laterally advecting mantle (LAM) is underlain by the weaker asthenosphere in which both heat and mass are transferred by convection. The layer of lateral advection of mass, and anisotropy (LLAMA; Fig. 1), is both a conduction and a shear boundary layer, with anisotropic solid-like properties. It is not called 'lithosphere' or 'plate' here because the primary characteristics that define it are not strength, rigidity or, in the case of McKenzie & Bickle (1988), temperature. The base of this layer appears to be deeper than 200 km and may correspond to the Lehmann discontinuity at ~220 km depth.

Mantle anisotropy is usually attributed to solid-state deformation and orientation of olivine. The seismic low-velocity layer (LVL) is sometimes attributed to high temperature gradients with no partial melting (e.g. Gutenberg, 1959; Stixrude & Lithgow-Bertelloni, 2005; Priestley & McKenzie, 2006). Mid-plate magmatism is usually attributed to mantle upwelling (e.g. a narrow radially zoned, vertical or tilted, cylindrical upwelling driven by thermal buoyancy) rather than to a process, as are other forms of magmatism. Recent studies, discussed below, challenge all of these attributions and lend support to a boundary layer model that involves shear deformation between the plate and the underlying 'partially molten' mantle, the formation of melt-rich shear bands and shear-driven non-buoyant upwellings. It will be shown that seismic data are consistent with a thicker (~220 km) BL and a higher basal temperature than in standard models of petrology. Bathymetry data are consistent with lateral temperature gradients and high, ~1600°C, mid-plate potential temperatures (see below).

BACKGROUND

Basics of mantle structure

Beno Gutenberg discovered a minimum in shear-wave velocity (V_s) at a depth of 150 km in the upper mantle (Gutenberg, 1959), a feature that has been repeatedly confirmed for over 50 years. The BL at the top of the mantle and the Gutenberg low-velocity layer (LVL; Fig. 1) hold the key to a number of petrological and geodynamic problems; however, in recent years these have received much less attention from geodynamicists and geochemists than the core–mantle boundary (CMB) region, including the D" layer at the base of the mantle. Ironically, the features of D" that have been quoted as arguments for it being a plausible geochemical reservoir, such as heterogeneity, anisotropy, possible presence of melt, and a high thermal gradient, also apply to the upper BL. D", however, has been assumed to have a higher potential temperature and to define a more stable reference system (compared with the plates) than any part of the upper mantle. In fact, both the surface of the Earth and the surface of the core are plausibly interpreted as free-slip boundaries, implying that their associated BLs are not stable reference systems.

Mantle nomenclature (Fig. 1)

There is no generally agreed upon name or thickness of the upper boundary layer of the mantle; the terms 'lithosphere', 'lid' and 'plate' are not appropriate, as we shall see. Gutenberg (1959) referred to the region of the mantle between the Moho and about 200 km depth as Region B, and the 750 km thick region between 200 and 950 km depth as Region C (Fig. 1). Bullen (1947) named the region between 410 and ~900 km depth 'the Transition Region between the upper and lower mantle'. Region D', the main part of the lower mantle, extends from ~900 km to 2700 km and D" is the ~200 km thick CMB region. D" is the only term still used widely today and it receives considerable attention as a possible geochemical reservoir and recycling bin. Part of the reason for this focus on D", rather than the shallow mantle, is the perception that the whole upper mantle is homogeneous and can provide only depleted mid-ocean ridge basalts (MORB), and that D" is isolated from 'the convecting mantle'. To focus attention on the upper mantle boundary layer region, I will revive Gutenberg's nomenclature and refer to the mantle region above ~220 km depth, the depth of a prominent discontinuity in the global seismic reference model PREM (Dziewonski & Anderson, 1981), as Region B. It contains the more loosely defined lithosphere.

Regions B and D" are the upper and lower BLs of the mantle and they are of roughly equal thickness, but differ considerably in volume and accessibility. Although we know that there are large temperature increases across each of these BLs, we do not know, surprisingly, if the average potential temperature in D" is greater than the temperature at the base of B. In a fluid with constant properties, heated from below and cooled from above, the upper and lower BLs play equivalent roles in convection. This is far from the case for the mantle. Internal heating and the effects of pressure on melting points and physical properties not only break the symmetry between the top and bottom of the mantle but can lead to a completely different form of convection, driven and organized from the top. Although a high vertical thermal gradient and lateral mobility are what characterize horizontal BLs, lateral temperature and density gradients also occur and these plus other body and boundary forces are responsible for driving advective motions.

Discontinuities and gradients

High-frequency seismic waves interact with a sharp mantle discontinuity that occurs at a depth between 50 and 120 km, depending on the age and nature of the plate.

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This is the Gutenberg discontinuity (G), which represents an abrupt drop in seismic velocity and the boundary between the seismic lid and the LVL. The seismic lid is not the same as the lithosphere (Anderson, 2007*a*). Region B terminates at the Lehmann, or L, discontinuity (Lehmann, 1959; Dziewonski & Anderson, 1981; Rost & Weber, 2001) at ~220 km depth.

The LVL appears to be composed of a series of low-rigidity sills that may be melt-rich (Kawakatsu *et al.*, 2009). This causes the LVL to adopt a form of anisotropy known as transverse isotropy that has the symmetry of a hexagonal crystal with a near-vertical *c*-axis (Fig. 2). The lid plus the LVL (e.g. Gutenberg's Region B) constitute the conduction layer and the upper BL of the mantle. Cooling of this laterally advecting composite layer, plus spreading and migration of ridges, sinking and roll-back of slabs and delamination of over-thickened crust, are what drive mantle convection in the top-down model. The outer shell may have an overall westward drift component (Doglioni *et al.*, 2005). Radioactivity, secular cooling and gravity are the ultimate energy sources.

The robust features of modern upper mantle structures include the high-velocity seismic lid, the anisotropic and attenuating LVL bounded by the G and L discontinuities, and the Transition Zone (TZ) bounded by two major first-order discontinuities, discovered in the 1960s, at average depths of 410 and 650 km (Fig. 1). The axis of the LVL (minimum V_s) is generally near 150 km depth except near spreading ridges, where it shoals significantly. These observations are not new (see Anderson, 1965, for an early review); they have been repeatedly confirmed and refined over the years. What is new is that we now know the lateral heterogeneity in each region and this reinforces the importance of Gutenberg's Region B, which includes the lid and the LVL, for mantle petrology.

The negative shear velocity gradient in B implies a superadiabatic temperature gradient to depths of the order of 60 km under ridges and 150–220 km elsewhere, at least for a homogeneous solid mantle. When anisotropy is taken into account, the decreasing velocity of SH waves (V_{SH}) indicates that the thermal gradient in the mantle is superadiabatic to 220 km depth. The presence of thin near-horizonal melt-rich layers does not change this conclusion (see below). The lid and the LVL are of variable thickness and these variations account for most of the lateral changes in subcrustal seismic wave delay times observed in teleseismic travel-time studies. The TZ is bounded by temperature-dependent phase boundaries, so it is a different kind of BL from Regions B and D". It is also a plausible geochemical filter and reservoir. The main focus in this study is, however, on the upper BL.

G and L are plausibly interpreted as melt-in and melt-out boundaries, or fluid-rich–fluid-poor boundaries. They also delineate the most anisotropic part of the mantle. The top of the LVL occurs between about 50 and 110 km depth beneath oceans and islands, and at depths of 100 ± 20 km under continents (Thybo, 2006; Rychert & Shearer, 2009). The seismic velocities in the LVL differ from region to region (e.g. Tan & Helmberger, 2007). This may, in part, be due to a lateral temperature gradient in which temperatures at depth are lower under ridges than under older plates (e.g. Hillier & Watts, 2004, 2005).

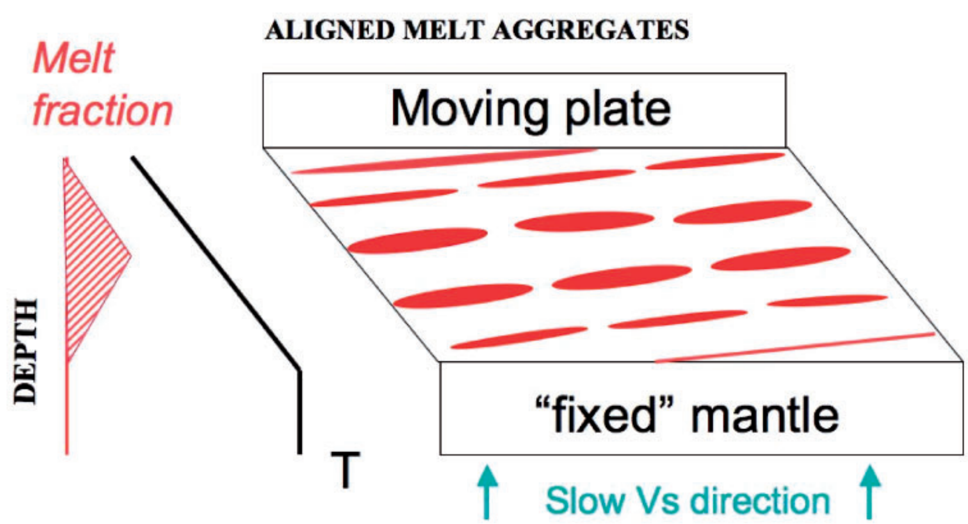

Fig. 2. The LLAMA model. A moving plate shears the underlying partially molten mantle; low-rigidity material ('melt-rich') segregates into fine-grained shear zones that dip gently in the direction of plate motion. Because most of the driving forces are in the plate–slab system (the boundary layers) there is little motion in the 'fixed' mantle. The theory of wave propagation in laminated media with thin low-rigidity layers shows that seismic waves with subhorizontal polarizations and propagation directions are mainly sensitive to the solid matrix and temperatures whereas vertically propagating waves are slowed down by the low-rigidity lamellae and are sensitive to the melt content.

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Lid, lithosphere, low-velocity layer and asthenosphere

The lid and LVL are different concepts from the lithosphere and asthenosphere, although the lid–LVL interface (G) is often called the lithosphere–asthenosphere boundary (LAB). The heterogeneity and anisotropy of the LVL is consistent with it being in a BL, but inconsistent with it being the same as the classical asthenosphere, which should be relatively isothermal (laterally), adiabatic (vertically) and homogeneous (e.g. Schuberth *et al.*, 2009), and which probably extends to greater depth than the LVL. The asthenosphere may decouple plate motions from the interior (e.g. Doglioni *et al.*, 2005). The mantle between L and 410, labeled C' in Fig. 1, is comparatively homogeneous (e.g. Kustowski *et al.*, 2008).

The extreme heterogeneity and anisotropy of B means that it must be understood in detail before inferring deeper structure, as deeply penetrating body waves sample it at least once and usually twice before arriving at a seismometer. Most mantle samples either originate in, evolve in, or pass through this region prior to being sampled. Nevertheless, this important part of the mantle has been largely ignored in recent geodynamic (e.g. Farnetani & Samuel, 2005), geochemical (e.g. Konter *et al.*, 2008) and seismological studies (e.g. Montelli *et al.*, 2004; Wolfe *et al.*, 2009). In older papers, it was referred to as 'the depleted mantle' or as 'the convecting mantle' and was considered to be homogenized by vigorous convection.

Semantics (Fig. 3)

Lithosphere was originally a rheological concept that involved long-term strength. Strength is a function of mineralogy, applied stress, duration of load, grain size, temperature and volatile content. In petrology, the term has taken on a variety of other connotations, including implications about isotope composition, temperature gradient, melting point, seismic velocity and thermal conductivity. 'Lithosphere' is, by some definitions (e.g. McKenzie & Bickle, 1988), non-convecting and characterized by a conductive (or locally advective) geotherm. This is actually the conduction, or thermal boundary, layer.

'Lithosphere' and 'asthenosphere' have also been assigned distinctive major element, trace element and isotope characterists (Kay, 1979; Ellam & Cox, 1991; Ellam, 1992; Haase, 1996; Phipps Morgan, 1997; O'Reilly & Griffin, 2006; Griffin *et al.*, 2008; O'Reilly *et al.*, 2009). Lithospheric mantle is often considered to be infertile (i.e. low in basaltic components such as CaO, FeO and Al_2O_3, and high in MgO) compared with the 'convecting asthenospheric mantle' but may be locally refertilized (metasomatized) by melts infiltrating from the asthenosphere (O'Reilly & Griffin, 2006) The latter is considered to be isotopically depleted (i.e. low in such ratios as $^{87}Sr/^{86}Sr$, $^{206}Pb/^{204}Pb$, $^{3}He/^{4}He$, etc.) compared with the sources for mid-plate magmas. Geochemically, the lithosphere–asthenosphere boundary is defined not by physical properties but in terms of geochemical and isotopic characteristics.

The term 'lithosphere' is sometimes used when what is actually meant is 'seismic lid' (Fig. 3); that is, a high seismic velocity, thermal boundary layer, comprising both crustal material and relatively refractory peridotite. Here the plate, lid and boundary layer concepts are distinct and cannot be lumped into the term 'lithosphere', convenient as that term is. Terms such as 'lithospheric roots' and 'lithospheric discontinuities' are in common use in the literature and are used, when necessary, as structural terms with no implications regarding their major and trace element geochemistry, age or thermal gradient. The focus of this study is on the upper boundary layer of the mantle, Gutenberg's Region B in Fig. 1, and it is important to distinguish this from 'lithosphere', 'lid' and 'convecting' or 'depleted mantle'.

AMBIENT OCEANIC UPPER MANTLE

In a cooling half-space, isotropic shear-wave velocities, V_s, are predicted to pass through a minimum value at depths of about 60 km under 10 Ma plates and about 150 km under 100 Ma plates (Stixrude & Lithgow-Bertelloni, 2005). These predictions are well satisfied by the seismic observations (e.g. Shapiro & Ritzwoller, 2002) except that (1) values of V_s at the minima (the axis of the LVL) are lower than predicted by the calculations, and (2) at depths below ~200 km, near-ridge mantle has higher average shear velocities than at the same depth under older plates, suggesting that it may be colder.

The heterogeneity and anisotropy of the upper mantle have now been worked out in great detail (e.g. Ekström & Dziewonski, 1998; Webb & Forsyth, 1998; Shapiro & Ritzwoller, 2002; Maggi *et al.*, 2006; Tan & Helmberger, 2007; Kustowski *et al.*, 2008; Nettles & Dziewonski, 2008). The discussion below is consistent with a large number of recent studies of which those listed above are representative.

Minima in the isotropic shear-wave velocity (V_s) and the vertically polarized shear-wave velocity (V_{SV}) as a function of depth, and the maximum in anisotropy, occur under mature parts (>50 Ma) of the Pacific plate at a nearly constant depth of ~140–150 km. For ages <50 Ma, these depths decrease from ~90 km to 60–70 km at the East Pacific Rise (EPR).

The deepening of the LVL axis below older plates is consistent with half-space cooling calculations but, in contrast to predictions, V_s in near-ridge mantle increases rapidly with depth below the axis of the LVL and does not converge to the same values as under older plates. In fact, average near-ridge mantle is seismically faster, at depth,

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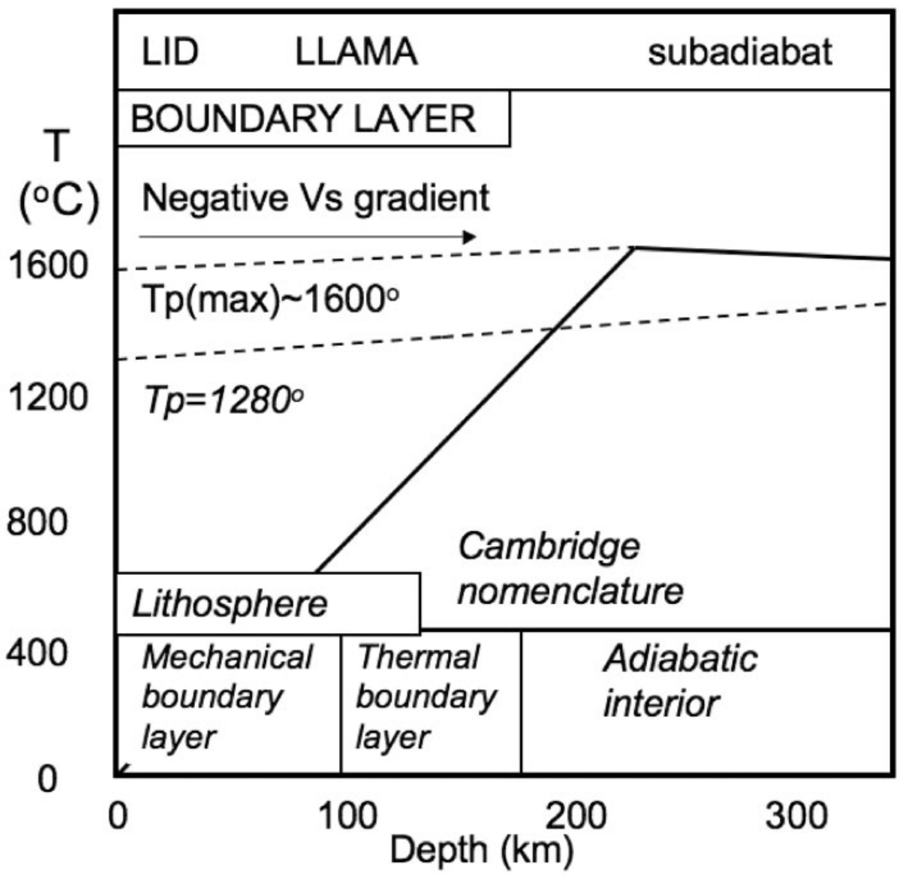

Fig. 3. In the Cambridge model (McKenzie & Bickle, 1988) the 'lithosphere' encompasses most of the depth interval in which the geotherm is conductive and includes the mechanical boundary layer, or plate, and the top part of 'the thermal boundary layer' of that model. The interior sub-'lithosphere' part of the mantle is adiabatic, implying absence of radioactive heating and secular cooling. In the top-down model of this study, the surface boundary layer includes the high-velocity seismic lid and the laminated low-velocity layer that contains aligned melt arrays (LLAMA). The boundary layer is both a conduction and a shear layer. A negative V_s gradient implies a superadiabatic gradient. Two potential temperature curves (1280 and 1600°C) are shown.

than average 100 Ma mantle, which implies 50–100°C lower temperature, if entirely due to temperature. The seismic velocities at ~200 km depth under older plates vary by amounts that imply temperature variations that are consistent with those inferred from bathymetry. Hillier & Watts (2004) obtained, from north Pacific bathymetry, $T = 1522 \pm 180$°C at 115 km depth. Temperature variations below 100 km depth, inferred from both seismology and bathymetry, are an order of magnitude higher than given by Priestley & McKenzie (2006), who forced convergence with the ridge geotherm.

There are also local high-velocity and lid thickness anomalies in the BL that correlate with melting and isotope anomalies. These have been attributed to refractory Archean-age domains in the plate that perturb mantle flow (e.g. O'Reilly *et al.*, 2009). These may also be the high ^{3}He/^{4}He domains that occur in the oceanic mantle near fracture zones and thick sections of the lid (Anderson, 1998, 2007*a*; Anderson & Natland, 2007).

The axis of the LVL

The negative shear velocity gradient extends to about 60 km under ridges and to at least 150 km depth under mature oceanic plates. The most obvious and traditional explanation for a gradual decrease of seismic velocity with depth is a thermal gradient of 6–10°C km^{-1} (e.g. Anderson, 1965; Whittington *et al.*, 2009), which far exceeds the adiabatic gradient.

Explanations for the minimum value of V_s include partial melting, an approach to the melting point, presence of volatiles, anelasticity and small grain size. A more abrupt decrease in V_s near 65 km depth could occur if the peridotite is carbonated (Presnall & Gudfinnsson, 2010); part of

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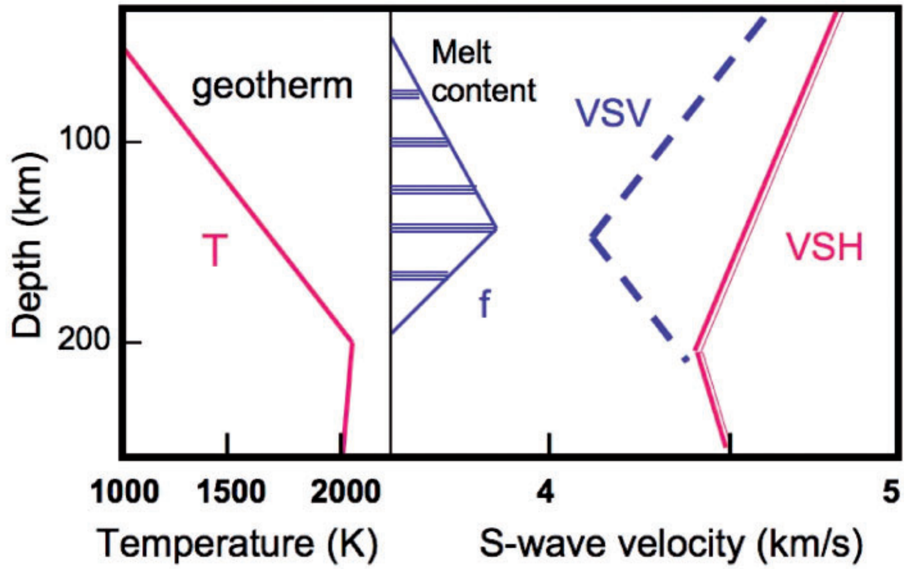

Fig. 4. Schematic geotherm, melt content and simplified V_{SV} and V_{SH} vs depth (after Shapiro & Ritzwoller, 2002). Horizontally propagating SH-wave velocities, V_{SH}, are mainly controlled by the velocities of solid olivine-rich layers and their T and P derivatives. The velocities V_{SV} of SV waves are mainly controlled by the low-velocity, melt-rich or low-rigidity lamellae.

the decrease of V_s with depth in this case may be due to increasing melt content. Arguments against partial melting as an adequate explanation for the observed low seismic velocities assume that the mantle is isotropic and homogeneous and that melts are in thermodynamic and textural equilibrium with a static matrix. The LVL may contain melt but this does not imply that it is necessarily an equilibrium partial melt. Thus, the term 'partially molten low-velocity zone' may be misleading.

The model proposed by Kawakatsu *et al.* (2009), which involves aligned melt-rich lamellae, suggests that the nature and origin of the LVL must be reconsidered. This model calls into question the usual practice of equating the LVL with the asthenosphere, particularly as small amounts of melt do not cause large changes in rheology. It may also explain chemical disequilibrium between melt and mantle peridotite and the presence of localized high-flux conduits (e.g. Spiegelman & Kenyon, 1992). Oriented lamellae explain why the lowest shear velocities are associated with the highest anisotropy. Vertically polarized shear-wave velocities, V_{SV}, are lower than horizontally polarized shear-wave velocities, V_{SH}, and go through a minimum at shallower depths than V_{SH} (Fig. 4).

Lateral variations in ambient mantle

The sub-ridge mantle, at the depth of magma extraction, is remarkably isothermal (Niu & O'Hara, 2008; Presnall & Gudfinnsson, 2010). The temperature range along the East Pacific Rise corridor inferred from seismology is also small, only $\pm 20^\circ$C (Melbourne & Helmberger, 2002); it may be twice this along the slower spreading Mid-Atlantic Ridge (Sheehan & Solomon, 1991; but see Presnall & Gudfinnsson, 2010). The very rapid increase of V_s with depth below 60 km under spreading ridges indicates that the thermal gradient cannot be high, and may even be negative. The upper mantle below 200 km depth appears to be colder under ridges than elsewhere.

The trend of bathymetry vs age (subsidence rate away from ridges) is too gradual to be explained by simple cooling plate models with realistic parameters (Hillier & Watts, 2004; Korenaga & Korenaga, 2008). Observed depth anomalies with respect to cooling plate models are too deep at ridge crests and too shallow at older ages. The most straightforward explanation that is consistent with the seismic data is that ridge mantle is colder than the average mantle under older plates. Subsidence in the South Pacific has been interpreted as due in part to cooling of the oceanic plate and in part to a lateral temperature gradient in the underlying mantle (Hillier & Watts, 2004).

The flattening of bathymetry for older plates has been interpreted in terms of cessation of cooling, reheating, or a horizontal isotherm at ~100 km depth. However, the seismic lid continues to increase in thickness across the entire Pacific (Zhang & Tanimoto, 1993; Maggi *et al.*, 2006) and there is no indication of lower seismic velocities in either the lid or the LVL, or high heat flow, in regions that are shallower than expected. The lateral density gradient (from bathymetry) and velocity gradient (from tomography) may be due to lithology or temperature. The

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tomography data suggest that this lateral gradient occurs deeper than ~200 km depth but it may also extend to shallower depths.

At depths greater than 200 km most of the Pacific mid-plate mantle has lower seismic velocities than near-ridge and Nazca plate mantle (Ritsema *et al.*, 1999; Maggi *et al.*, 2006), but there is otherwise little correlation with the age of the plate. At 250 km depth there are very few areas in the Pacific that have shear velocities as high as under young plates, with the notable exception of the mantle around Hawaii. Localized high-velocity anomalies have been attributed to stranded fragments of lithosphere and a possible shallow source for geochemical anomalies (e.g. O'Reilly *et al.*, 2009).

If these variations in physical properties are in fact due to temperature variations then MORB temperatures cannot be used as an upper bound on ambient mantle temperature elsewhere. Mid-plate magmas may reflect the potential temperature, T_p, of ambient mantle at the depth of magma extraction. The usual interpretation, of course, is that volcanic islands lie above localized hotspots that have radii not much bigger than the islands themselves. The most obvious way of explaining the observed range in magma temperatures is to vary the depth of magma extraction in the high thermal gradient region of the upper mantle (Figs 1–4).

HAWAII

From the usual petrological perspective, mid-ocean ridge basalts represent partial melts of ambient upper mantle and ocean island basalts (OIB) are derived from somewhere else. It is therefore instructive to compare the mantle under ridges, mid-plate locations and hotspots from a geophysical perspective. Based on plume models that have been created to explain the superficial and petrological evidence, Hawaii should be the largest geophysical anomaly; for example, in heat flow, lid thickness, seismic velocity, and depths of discontinuities and isostatic compensation. This is not a requirement of alternative mechanisms for mid-plate magmatism, including the one developed here. Predictions of the plume hypothesis (e.g. Campbell & Kerr, 2007) are that hot mantle should spread out laterally for large distances beneath the surface volcanoes, that heat flow and magma volumes should be high, that upper mantle seismic velocities should be low, and that the lithosphere should be weak and thin. Ridges, of course, are more extreme than Hawaii in all these respects. Comparisons between Hawaii and the adjacent mantle, or with hotspot-free mid-plate locations, do not support these predictions. The seismic anomalies associated with most hotspots extend to the same sorts of depths as do ridges and cratons and do not extend far from the volcanoes (Ritsema & Allen, 2003; Pilidou *et al.*, 2005; Priestley & McKenzie, 2006; Adam *et al.*, 2010) but, in plume theory, Hawaii is predicted to be a more significant anomaly. Geochemical anomalies in the Atlantic have been attributed to the presence of high-velocity remnant lithospheric fragments isolated by disruption of ancient continents during rifting (O'Reilly *et al.*, 2009).

Authors of travel-time studies that use only island stations ordinarily attribute 'anomalies' to the local mantle and assume that the unsampled adjacent mantle is 'normal' (Appendix A). The same reasoning is used in petrogenetic models of ocean island basalts; inaccessible parts of the sub-plate mantle are assumed to be MORB source-like. In other words, anomalous mantle is confined to the region being sampled. Seismic studies that use surface reflected phases show that oceanic reflection points away from 'hotspots' can be more 'anomalous' than island stations, including Hawaii (Butler, 1979; Kustowski *et al.*, 2008).

Seismic velocity structure under Hawaii

The most obvious anomalies associated with the Hawaiian chain are topographic and petrological. Magma volumes and compositions are variable and they correlate with the locations and trends of large fracture zones (FZs; e.g. Basu & Faggart, 1996; Van Ark & Lin, 2004). Very low magma volumes are associated with the Mendocino FZ and very high magma volumes are associated with the Molokai FZ. The orientation of FZs is an important consideration in the interaction of moving plates with the underlying mantle (e.g. Yamamoto *et al.*, 2007; Yamamoto & Phipps Morgan, 2009; Conrad *et al.*, 2010).

The following discussion is based on studies that are recent, and that use large amounts of data and a variety of seismological techniques (e.g. Collins *et al.*, 2002; Ritzwoller *et al.*, 2002, 2003, 2004; Rost & Weber, 2002; Li *et al.*, 2004, 2008; Lawrence & Shearer, 2006; Maggi *et al.*, 2006; Priestley & McKenzie, 2006; Wolbern *et al.*, 2006; Deuss, 2007; Laske *et al.*, 2007; Kustowski *et al.*, 2008; Tauzin *et al.*, 2008; Visser *et al.*, 2008; Priestley & Tilmann, 2009). The basic conclusions, however, are mainly confirmations and refinements of earlier studies (e.g. Best *et al.*, 1974; Zhou & Wang, 1994; Woods & Okal, 1996; Ekstrom & Dziewonski, 1998; Katzman *et al.*, 1998; Priestley & Tilmann, 1999). Most of the studies listed above used absolute times and velocities, and some accounted for anisotropy; these are essential considerations if Hawaii is to be put into the context of global and regional datasets (see Appendix A).

The global reference model PREM and a high-resolution trans-Pacific profile, or corridor, that avoids Hawaii and other mid-plate volcanism are used as standards of comparison (Tan & Helmberger, 2007). A recent high-resolution seismic study of California permits comparison of a

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continental tectonic region with Hawaii (Schmandt & Humphreys, 2010).

Based on these studies, the following conclusions may be drawn.

(1) The seismic velocities under Hawaii are comparable with or higher than under unperturbed middle-aged oceanic plates.
(2) Absolute travel times to Hawaii from circum-Pacific events and the inferred elastic and anelastic structure of the upper mantle beneath the Hawaiian swell are inconsistent with locally high temperatures or extensive melting.
(3) Shear velocities under the Hawaiian swell reach a minimum value near 150 km depth and the base of the LVL is near 200 km, similar to normal oceanic values and some continental shields.
(4) The 410 and 650 km discontinuities, on average, are at normal depths, implying normal mantle temperatures.
(5) Transition Zone thicknesses under California, and many other regions, are less than under Hawaii and under young ridges, implying temperature excesses of $\sim$200°C relative to the latter.
(6) Travel time and seismic velocity anomalies in the vicinity of Hawaii are not particularly extreme by global standards. Regions that have larger travel time delays or lower seismic velocities include western North America, the Lau Basin, Tonga, Israel, Arabia, Tibet, New Zealand, eastern Australia, Japan, and the SW, NE and eastern Pacific. If seismic velocity variations at 250 km depth are mainly controlled by temperature then parts of the western Pacific, and the Indian and South Atlantic oceans and between Hawaii and the Americas are hotter than under Hawaii. On the other hand, near-ridge mantle, on average, is cold by this measure and at depth (>200 km) it does not have seismic velocities similar to the mean under mid-plate locations.

In addition, heat flow, flexure, earthquake depth and electrical conductivity data imply normal unperturbed mantle beneath the Hawaiian swell (e.g. Stein & Stein, 1992; Watts & Zhong, 2000; McKenzie *et al.*, 2005). The mantle under the central Pacific may have higher absolute temperatures than standard petrological models and than sub-ridge mantle, but Hawaii does not stand out as a localized low-velocity or high-temperature anomaly in any geophysical parameter or in any well-constrained seismological study.

None of the above conclusions contradict a recent study of relative arrival times of nearly vertical teleseismic waves recorded on a temporary seismic array installed around the Hawaiian swell (Wolfe *et al.*, 2009), which showed that half of the arrivals over the duration of the experiment and over this small area were delayed relative to the other half. No absolute arrival times or velocities were measured, depth resolution was poor and no comparisons with global data or with other regions were made. The regional and global context of Hawaii (see below) shows the unreliability of conclusions based only on relative times and near-vertical rays (see also Appendix A).

The Hawaiian lid

The average depth to the G-discontinuity under the Big Island is 110 km (Li *et al.*, 2000), which is greater than the average value of 95 ± 4 km beneath Precambrian shields and platforms (Rychert & Shearer, 2009). The thickness and seismic velocity of the lid under and around Hawaii, and the depth of compensation of the Hawaiian swell (Van Ark & Lin, 2004), are comparable with continental shield values (Bechtel *et al.*, 1990). The V_s in the lid under the Hawaiian swell reaches values as high as 4·8 km s^{-1}, similar to values under the Canadian shield (Laske *et al.*, 2007). These are surprising results, at least in the context of the plume hypothesis. The unperturbed lid under the most active volcanoes suggests that they are fed by systems of narrow dikes and fissures that may extend to $\sim$150 km depth. The plate is younger to the north of the Molokai FZ and it may therefore be thinner because it has cooled less, not because it has been thermally eroded. Nevertheless, the lid under Hawaii is anomalously thick and may contain ancient refractory and ancient components (e.g. Simon *et al.*, 2008), including high ^{3}He/^{4}He components.

High seismic velocities under Hawaii

A large number of surface-wave paths have now been studied that crisscross the Pacific and, in particular, the region containing the Hawaiian chain and swell. In a large area surrounding Hawaii the essentially raw dispersion data are almost precisely PREM (i.e. average Earth). Other areas of the Pacific are slower than the region surrounding Hawaii (e.g. Ritzwoller *et al.*, 2002, 2004; Maggi *et al.*, 2006; Visser *et al.*, 2008).

It can be argued that the lateral resolution of these studies (500–1000 km) cannot detect a plume under Hawaii, even if one existed. Hot active upwellings, if they exist, should spread out in low-viscosity zones and under plates and should influence a large area of the mantle (Campbell & Kerr, 2007; Yamamoto *et al.*, 2007; Korenaga & Korenaga, 2008; Yamamoto & Phipps Morgan, 2009). Seismic data, however, have restricted the radius of influence surrounding mid-plate volcanoes and ruled out a strong Morgan–Campbell type plume under Hawaii and a plume-fed asthenosphere.

There are associations of shallow high seismic velocity domains with mid-plate magmatism in the Atlantic and Indian oceans (O'Reilly *et al.*, 2009). These may be related to ancient, and large, lithospheric fragments that have

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been stranded in the shallow mantle, and that perturb the local mantle flow.

Hawaii vs California

Schmandt & Humphreys (2010) used seismic velocities between depths of 60 and 200 km in California to infer potential temperatures (T_p) of ~1600°C plus ~1% melt, or a lower T_p and more melt. California is part of a large tomographic low-velocity anomaly that is similar to others that occur away from hotspots in global tomographic studies. The tectonic history of western North America is favorable to the idea of a disrupted boundary layer and access to its deeper hotter parts. Suitably oriented large offset fracture zones on a moving plate may play a similar role (e.g. Conrad *et al.*, 2009; Yamamoto & Phipps Morgan, 2009).

Seismic velocities and TZ parameters beneath California are consistent with hotter mantle than under Hawaii. For example, Pasadena has one of the thinnest TZs in the world (Lawrence & Shearer, 2006). Mantle T_p inferred from TZ thicknesses under California range from 1380 to 1530°C, depending on modeling assumptions (Ritsema *et al.*, 2009). The inferred temperatures in the LVL under the North Pacific, far from Hawaii, is 1522 ± 180°C (Hillier & Watts, 2005). The implication is that ambient temperatures of the mantle at depths of >100 km are higher than inferred at mid-ocean ridges and that these are not just localized hotspots in the mantle. These inferences about absolute temperature are less reliable than the simple observation of relative seismic velocities and TZ thicknesses. However, none of these suggest that the mantle under Hawaii is particularly hot by global standards.

Melt-rich lamellae under Hawaii?

What is seismologically unique about the mantle surrounding Hawaii, in addition to its thick lid with high average shear velocities, is its extreme anisotropy below ~100 km (Ekstrom & Dziewonski, 1998). Anisotropy of this kind has been explained by Kawakatsu *et al.* (2009) as due to a series of melt-rich lamellae in the upper ~200 km of the mantle. This, combined with plate-driven flow, has the potential to explain mid-plate magmatism and the associated geophysical observations, including large vertical travel-time delays, without a localized high-temperature anomaly.

In the sheared mantle beneath a moving plate, fluids are predicted to redistribute into networks of shear zones (Holtzman *et al.*, 2010; Kohlstedt *et al.*, 2010) oriented at 5–15° to the shear plane, dipping down in the direction of plate motion. Nearly vertical S waves that pass through the core (SKS) and through this laminated structure are predicted to be slowed down compared with less vertical S waves, as observed (Wolfe *et al.*, 2009); horizontal SH (Love) waves are predicted to be relatively fast, as observed.

Summary

From a geophysical point of view, in the context of regional and global data, Hawaii does not stand out in any way that is commensurate with its status as the world's largest, hottest and longest-lived hotspot. It has a thick high-velocity lid, unperturbed heat flow, deep compensation and is far from having the lowest upper mantle seismic velocities or thinnest TZ. If Hawaii is not anomalous, by global or Pacific standards, but sub-ridge mantle is, then mid-ocean ridges have to be reconsidered as the petrological reference state. The Pacific plate may overlie, and Hawaiian volcanoes may sample, ambient shallow mantle. In other words, Hawaii is not a localized anomaly in a MORB-like mantle.

MID-OCEAN RIDGES

The shallow structure of the EPR has been mapped in some detail (e.g. Conder *et al.*, 2002; Dunn & Forsyth, 2003). The relatively high shear velocities found below ~200 km depth beneath many young oceanic plates, and near Hawaii (e.g. Shapiro & Ritzwoller, 2002; Maggi *et al.*, 2006; Visser *et al.*, 2008), and other hotspots (O'Reilly *et al.*, 2009), are unexpected. Why ridges, on average, appear to overlie cold or geochemically distinct mantle is an intriguing question. Ridges may migrate towards regions of cold mantle, or high degrees of melt extraction may refrigerate the mantle. On the other hand, there are a variety of mechanisms for raising the shallow mantle temperature beneath rapidly moving or long-lived, thick, insulating plates. High seismic velocities may also be due to compositional effects such as high pyroxene/garnet ratios or low volatile or FeO contents in the mantle peridotite.

Most discussions of mantle temperature and composition assume that ridges are passive samplers of mantle that is representative of the whole upper mantle. If ridge mantle is not representative of ambient mid-plate mantle, the hotspot problem is turned on its head. The question is not why are hotspots hot but why is mantle below ~200 km depth under young oceanic plates, on average, cold, or at least different? It is important to note that the lateral resolution of most tomographic studies is 500–1000 km, and 'near' must be understood in this context. Only a few ridges have been sampled at high resolution.

How deep does the 'thermal anomaly' beneath ridges extend? A high V_s gradient extends from 100 to 200 km depth under young plates (e.g. Shapiro & Ritzwoller, 2002), which suggests a low or negative vertical thermal gradient, or some compensating gradient in grain size or fabric that is unique to the mantle under spreading ridges. Relatively high shear velocities extend from 200 km to at

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least 400 km depth although some near-ridge segments are underlain by near-average shear velocities.

Niu & O'Hara (2008) have argued that the variation of the depths of ridges combined with MORB geochemical variations requires density differences to extend throughout the upper mantle. These may be, in part, due to chemistry. Temperature effects would show up in the depths of mantle phase change discontinuities. The 410 km discontinuity is elevated under the ridges sampled by Schmerr & Garnero (2007), implying that the mantle is colder than average near 400 km depth. Melbourne & Helmberger (2002) showed that the TZ acts as an effective seismic waveguide along the EPR, including segments near hotspots, which implies that there can be little variation in TZ thickness, velocity or temperature in the mantle under this ridge. The average TZ thickness for global and trans-Pacific paths is 242–245 km (Lawrence & Shearer, 2006; Deuss, 2007; Tan & Helmberger, 2007), but is 250–255 km for paths near the EPR (Shen *et al.*, 1998*a*; Melbourne & Helmberger, 2002). Courtier *et al.* (2007) claimed that TZs under ridges are thicker and colder than under mid-plate volcanoes. On the other hand, TZ thicknesses under hotspots differ little from global averages and do not correlate, among themselves, with inferred magma temperatures.

BOUNDARY LAYER DYNAMICS

The uppermost and lowermost 200 km of the mantle are obvious boundary layers (Fig. 1), and the TZ may be a more subtle one (Montagner, 1998). Boundary layers are fundamental concepts in mantle dynamics as they drive convection. In the upper and lower BLs lateral advection is important. BLs are also often invoked as geochemical reservoirs as they serve as debris basins and recycling bins. From a seismological point of view, BLs are recognized by their heterogeneity, anisotropy, negative shear-wave velocity gradients and, possibly, melt content. In petrology, the terms 'lithosphere' and 'the convecting mantle' are often used to refer to the BL and the mantle below the BL, respectively.

The simplest theory for plate tectonics and mantle convection involves a cooling upper BL that drives itself and drives the underlying mantle by viscous drag (e.g. Elsasser, 1969; Harper, 1978; Hager & O'Connell, 1979, 1981). The potential energy of cooling plates and sinking slabs provides the immediate driving force, but gravitational and other body forces may also cause overall motion of the outer shell relative to the interior (e.g. Doglioni *et al.*, 2005). Mass balance is provided by matter displaced outwards and upwards by sinking slabs, by sluggish large-scale upwellings, and by counterflow in a low-viscosity channel or boundary layer at the base of the system.

In the plate tectonic–shallow BL and top-down hypotheses, mantle flow and upwellings are induced by plate motions, horizontal temperature and viscosity gradients, steps in plate thickness, shear between the plate or slab and the surrounding mantle, by delamination, slab rollback, and by plates spreading apart. In contrast to the plume hypothesis, the mantle, by and large, is passively responding to plate architecture and to plate and slab motions, and, perhaps, to overall motion of the outer shell over the interior (Doglioni *et al.*, 2005). The most heterogeneous, active and mobile parts of the mantle are in Region B.

Laterally advecting mantle

Oceanic bathymetric anomalies are not necessarily the surface expression of large-scale vertical motions or localized thermal anomalies in a homogeneous fluid (e.g. Niu & O'Hara, 2008; Yamamoto & Phipps Morgan, 2009). For example, topographic and geophysical data for the South Pacific favor horizontal flow and shallow support of topography over vertical flow from sub-BL depths (Hillier & Watts, 2004).

What distinguishes top-down models from other models of mantle convection is the nature of the coupling between the plates and the deeper mantle, the level of shear stress at the base of the plate, and the thickness and anisotropy of the BL. Resistive drag sets up a velocity gradient between the plate and the deep mantle, generating laminar flow and seismic anisotropy. The top-down model explains (or adopts) a number of plate tectonic and geophysical observations, including the coherent motions of plates, abrupt plate boundaries, dips of slabs and large-scale upper mantle anisotropy (e.g. Hager & O'Connell, 1979, 1981; Tanimoto & Anderson, 1984).

A dynamic boundary layer model

Flow in which a uniform plate, or shell, moves with constant velocity over a viscous medium creates a uniform seismic anisotropy. Laminar flow ordinarily keeps the deeper hotter parts of the surface boundary layer from being sampled. This flow is upset by fracture zones, delamination, subduction, ridge–trench collisions, and the breakup of supercontinents and the formation of ocean basins. When the otherwise laminar mantle flow induced by plate drag is perturbed or encounters thickness perturbations in the plate, it has a component of downwelling or upwelling. Igneous provinces typically occur at plate boundaries and at structural boundaries within the plate, and in regions of extension. It is probably no coincidence that the largest outpouring of magma along the entire Hawaiian chain occurs at the Molokai FZ, a major age boundary and zone of weakness in the Pacific plate, and where differential drag forces may allow buoyant magmas to magma-fracture the plate. In other places, differential drag can reduce extensional stresses and shut down magmatism.

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Boundary layer dynamics

The emphasis on whole-mantle and low-resolution convection models has diverted attention away from the possibility that mid-plate volcanoes may be the result of shallow processes, which until recently have not been modeled at high resolution or by thermodynamically self-consistent techniques. Recent geodynamic calculations show, surprisingly, that mid-plate swells and volcanism, and even large volcanoes, can be explained by physical property variations, and motions, entirely confined to the shallow mantle (Ballmer *et al.*, 2007, 2009, 2010; Conrad *et al.*, 2010), even if one places an impermeable isothermal boundary at 240 km depth (Adam *et al.*, 2010). Most of the thermal and chemical heterogeneity and melting occur above 200 km (i.e. in the boundary layer). Shallow mantle processes and compositions can also explain the petrology and chemistry of mid-plate magmas (e.g. Kay, 1979; Niu & O'Hara, 2008; Pilet *et al.*, 2008).

The lateral variations of seismic velocity and anisotropy in the BL can also account for the kind of teleseismic delay patterns that are often attributed to lower mantle plumes (e.g. Gao *et al.*, 2003, 2004; West *et al.*, 2004; Schmandt & Humphreys, 2010).

The success of boundary layer dynamics in reconciling high-resolution seismic and geodynamic data for the South Pacific and western North America, and the recognition that lateral temperature gradients and large-scale horizontal flow and shear-driven upwellings may be more important than local hotspot- and buoyancy-driven flow (e.g. Hillier & Watts, 2004; Conrad *et al.*, 2010) suggest that a different approach to mid-plate magmatism may be timely.

Shear-driven upwellings

Conrad *et al.* (2010) evaluated a mechanism for generating intraplate upwelling, named 'Shear-Driven Upwelling' (SDU), that does not require thermal buoyancy. SDU can also explain intraplate volcanism. The driver for SDU is the relative motion between the plate and the underlying mantle; upwelling flow is excited by lateral heterogeneity of various types rather than by thermal buoyancy. SDU is fundamentally different from upwellings associated with thermal convection, edge convection and buoyant diapirs. It explains the association of intraplate volcanism with fracture zones and edges of cratons, and with anomalous mantle anisotropy.

Shearing redistributes melt into networks of melt-rich shear zones (e.g. Kohlstedt & Holtzman, 2009; Holtzman *et al.*, 2010; Kohlstedt *et al.*, 2010) that serve to make the upper mantle seismically anisotropic. This explains the Rayleigh wave–Love wave discrepancy, shear-wave splitting and the variation in delay times of S and SKS waves, and, in principle, can be used to infer the relative motions of plates over the underlying mantle.

THE LAMINATED MANTLE

Lubricated lid and aligned melt anisotropy

Kawakatsu *et al.* (2009) observed abrupt shear-wave velocity reductions of 7–8% at depths that correlate with the age of the oceanic plate. These features correspond to the G discontinuity. The fact that shear-wave anisotropy below the G discontinuity is also about 7% indicates that sub-horizontal low-rigidity lamellae occur throughout the LVL. A model with melt segregated into horizontal layers that lubricate plate motion is consistent with the data. The LVL is composed of Aligned Melt-rich Arrays (AMA) encased in refractory peridotite (Fig. 2). Such a structure forms by shearing and spontaneous segregation of magma into fine-grained shear zones. Experiments show that the major shear bands in such a lithology dip down in the direction of plate motion (Holtzman *et al.*, 2010). The dip direction is the fast axis for P waves and S waves polarized in the plane of the lamellae. Shear bands in the mantle are expected to be fine-grained, volatile-rich and to have low melting points, and therefore to have low rigidity. Modeling of the velocity drop and the anisotropy shows that the rigidity of the amorphous or melted assemblages (AMA) may be an order of magnitude less than the rigidity of the refractory bands.

The Kawakatsu *et al.* (2009) model (see also Fuchs *et al.*, 2002; Anderson, 2005, 2006) consists of laminated lithologies and aligned melt arrays (e.g. LLAMA). Both gravity and shear play a role in the stratigraphy. The overall negative shear velocity gradient between G and L shows that LLAMA is in the conduction boundary layer. The strong anisotropy suggests that it is also the shear boundary layer between the plate and the mantle below 220 km. Because of its heterogeneity, and proximity to the melting point and the surface, it is a plausible geochemical reservoir that is larger than D" and the continental 'lithosphere' combined, which are the usual candidates for non-MORB reservoirs.

The seismic properties of a structure that is composed of laminated lithologies and melt arrays (LLAMA) can, in principle, constrain both the thermal gradient and the melt content.

Structural anisotropy of a laminated mantle

The long-wavelength directional rigidities, N and L, of a two-lithology laminated mantle can be written in terms of the component rigidities, G_{LLA} and G_{MA}:

$$N = (1-f)G_{\mathrm{MA}} + fG_{\mathrm{LLA}}$$
$$L = G_{\mathrm{LLA}}G_{\mathrm{MA}}/[fG_{\mathrm{LLA}} + (1-f)G_{\mathrm{MA}}].$$

LLA are the fine-grained liquid-like or amorphous arrays that contain volatiles and low-melting components or other low-rigidity materials; MA are the intervening

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mineral arrays or matrix, and f is the volume fraction of LLA.

The rigidity N controls the velocity of propagation, V_{SH}, of SH waves, traveling and polarized in the plane of the laminations. The rigidity L controls the SV-wave velocity ($V_{SV} < V_{SH}$), normal to the lamellae. SH and SV waves are polarized such that the vibration is parallel to or normal to the laminations, respectively.

In the limit of thin (low f) and low-rigidity LLA

$$N \sim G_{MA}$$

$$L \sim G_{LLA}/f.$$

N, which controls V_{SH}, is approximately the same as the matrix rigidity and is independent of the low-rigidity lamellae properties. The decrease of V_{SH} with depth, therefore, is evidence for a thermal boundary layer (TBL), a region of high thermal gradient (Fig. 4). The TBL, by this criterion, extends to a depth of 220 km. L, which controls V_{SV}, is sensitive to the low-rigidity or melt-rich lamellae. A decrease of V_{SV} with depth implies an increase in melt content, if the low-rigidity lamellae are melt-rich.

The lid-to-LVL transition is marked by a large decrease in rigidity, typically between 12 and 18% (Gaherty *et al.*, 1996; Kawakatsu *et al.*, 2009). The directional rigidity differences fall in the same range and peak at about 150 km depth in the Pacific (e.g. Ekstrom & Dziewonski, 1998; Shapiro & Ritzwoller, 2002).

The magma-absent layers may be refractory high-melting lithologies (e.g. dunite or harzburgite), which collect at the top of the mantle because they are buoyant. They may even contain fragments of ancient shallow mantle. The intervening layers have low melting points and rigidities and can be considered as pyroxenite, magma-mush lenses or sills, metasomatic lamellae or as fine-grained shear zones. The LLA are large-scale versions of the veins in metasomatized 'lithosphere'.

The low-rigidity lamellae do not all have to have the same composition; it is the large contrast in rigidity that controls the anisotropy and the seismic reflection and transmission coefficients. The thickness of the lamellae may be tens of meters to kilometers. Scattering of seismic waves in the upper mantle is effective at wavelengths between 3 and 100 km and is asymmetric, favoring flat oblate features (Baig & Dahlen, 2004; Shearer & Earle, 2004).

The low-rigidity lamellae (AMA) can be modeled either as continuous or discontinuous bands (Anderson, 1965, 1989, 2007a) or as aligned melt arrays (Tommasi *et al.*, 2006). Both the anisotropy of LLAMA and the rigidity drop at G are consistent with 1–1·5% melt in large aspect ratio flat lenses or lamellae that have rigidities about 10% of the matrix rigidity.

Seismic velocity, anisotropy and anelasticity are usually analyzed at the crystal and grain scale. LLAMA attributes these properties, in part, to the macroscopic structure, which is not evident at the xenolith, thin-section and hand-specimen scales. At the microscopic scale, one still expects to see crystal orientation and interstitial melt segregation.

Application to the mantle

It has long been known that average shear velocities decrease with depth in the upper 150 km of the mantle (e.g. Gutenberg, 1959; Anderson, 1965). The velocities of horizontally polarized (V_{SH}) and vertically polarized (V_{SV}) shear waves decrease with depth in the boundary layer, but at different rates, and they have their minima at different depths. This allows one to disentangle the effects of temperature and melt content.

Figure 4 shows a typical shear velocity profile for the mid-Pacific mantle. V_{SV} reaches a minimum at about 150 km depth, which could be either the depth of maximum melt accumulation or the depth of maximum equilibrium partial melting. According to Presnall & Gudfinnsson (2010) this may be the source of oceanic tholeiitic basalts (e.g. Iceland and Hawaii). V_{SH} continues to decrease to about 220 km, consistent with PREM (Dziewonski & Anderson, 1981). This is logically taken as the bottom of the surface boundary layer.

Steady-state and cratonic mantle geotherms have steep conductive gradients to depths of >200 km (e.g. McKenzie *et al.*, 2005). The seismic data suggest that the mid-plate oceanic geotherm is conductive to similar depths, in spite of the youth of the overlying plate. However, the minimum V_s under plates less than ~20 Myr old may occur at depths of order 60 km; the velocity gradient thereafter is higher than under older plates (e.g. Shapiro & Ritzwoller, 2002).

THE CAMBRIDGE MODEL

The Cambridge model of mantle petrology is based on a series of important and influential papers by McKenzie & Bickle (1988), McKenzie (1989) and Watson & McKenzie (1991). It has been extended by Priestley & Tilmann (1999, 2009), McKenzie *et al.* (2005), Pilidou *et al.* (2005), Priestley & McKenzie (2006) and Tilmann & Dahm (2008). McKenzie & Bickle (1988) proposed a sub-solidus potential temperature (T_p) of 1280°C ± 20°C for 'ambient mantle', for the base of the plate and for the whole upper mantle beneath the plate (Fig. 3). Mid-ocean ridges, being passive, sample 'ambient mantle' but mid-plate volcanoes do not. McKenzie *et al.* (2005) adopted $T_p = 1315$°C for ridge mantle and ruled out higher upper mantle temperatures by assuming that (1) seismic crust equals igneous crust, (2) crust forms by 100% melt extraction from a homogeneous mantle that is equally fertile at all depths and all locations, (3) ridges are fed vertically, and (4) sub-ridge mantle is representative of sub-plate mantle

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elsewhere. Higher mantle temperatures can be tolerated if the crust is not entirely igneous, if magma extraction is not 100% efficient, if there are less fertile portions of the mantle than the melting region sampled by ridges or if shallow lateral transport at ridges is more important than elsewhere.

Ambient mantle temperatures were based on MORB temperatures inferred from experimental petrology and a model of a vigorously convecting homogeneous subsolidus constant viscosity fluid (the so-called convecting mantle). In the model the mantle at ~100 km depth is artificially kept at a constant temperature and the temperature gradient in the mantle below the depth of melt extraction beneath the ridge is adiabatic. Melting occurs only as a result of lithospheric thinning, plate separation, or hot upwelling jets, and the mantle cannot retain melt after it forms. McKenzie & Bickle (1988) did not refer to any Morgan paper or the mantle plume hypothesis and their jets are different from Wilson–Morgan plumes. For one thing, jets do not spread out as they approach the surface, as an active upwelling should do, and do not have long-lasting thermal signatures (Pilidou *et al.*, 2005; Priestley & McKenzie, 2006). Melting at depths greater than 100 km, in the Cambridge model, requires potential temperatures significantly higher than MORB potential temperatures; mid-plate magmas require maximum temperatures more than 200°C higher than the assumed ambient temperatures. In the Cambridge model the LVL is entirely subsolidus, oceanic crust is entirely igneous and it represents complete melt extraction from a homogeneous mantle. The underlying mantle is essentially devoid of heat-producing elements and seismic velocities are functions of temperature and grain size only.

Problems with the cooling plate model

As McKenzie & Bickle (1988) pointed out, there is no physical basis for their constant thickness cooling plate model. It requires either zero conductivity below ~100 km depth or constant removal of the deeper hotter portions of the 'plate', or thermal boundary layer (TBL). In the latter case, there is, temporarily at least, material in the TBL that has a potential temperature higher than the 1280°C value adopted in the cooling plate model, and this may be brought to the surface at fracture zones by shear-driven upwellings (Conrad *et al.*, 2010). On the other hand, the whole basis of the constant thickness plate model is the observation, or inference, that seafloor bathymetery flattens out after about 70 Myr of cooling. However, bathymetry is controlled by density and density is not controlled solely by temperature. In the McKenzie & Bickle (1988) model, the cooling mantle is homogeneous and the deeper parts of the cooling boundary layer are of the same composition as the shallow parts and as the adjacent mantle. In principle, the lower part of the 'plate' can be partially molten or neutrally buoyant, even if it has cooled from some initial condition. The thickness of the seismic lid continues to increase out to the full age of the Pacific basin; there is no flattening. The large number of seamounts and plateaux in the older Pacific may mask the true subsidence signal. Although the constant thickness cooling plate model is widely accepted, it is contradicted both by the lid thickness data and the occurrence of the minimum in V_s occurring well below the plate thickness inferred by McKenzie & Bickle (1988).

Inertia can be safely ignored in geodynamic models. The term 'hot jet' implies, at least to some critics, something with inertia, but this is not an essential attribute of a jet. The term has also been used, in geodynamics, for the central hot core of a plume and for super-fast upwellings.

How do jets differ from plumes?

The McKenzie & Bickle (1988) jet hypothesis is independent of the Morgan mantle plume hypothesis. The jets in the Cambridge model come from an unspecified hot source below the adiabatic interior or the so-called 'convecting mantle'. Some researchers use 'jet' instead of or interchangeably with 'mantle plume', or use it for the hot central core of a plume (e.g. Nisbet, 1987; Campbell & Kerr, 2007; Adam *et al.*, 2010). Larsen & Yuen (1997) used the term 'jet' for ultrafast upwellings. A 'jet' in fluid dynamics is simply a stream of fluid moving at a higher velocity than the surrounding fluid.

'Plume' was initially used in geodynamics for both rising and sinking features, including the mantle rising, either actively or passively, under ridges. In fluid dynamics 'plume' means any buoyancy-driven vertical motion; 'mantle plume' has other implications, such as depth of origin, temperature, source fixity, dimensions, etc. (e.g. Morgan, 1971; Anderson & Natland, 2005, 2007; Campbell & Kerr, 2007). Boundary layer convection consists of cells with isentropic interiors enclosed by thermal boundary layers. Buoyancy forces are concentrated in narrow rising and descending plumes, and these drive both mantle convection and the plates. In mantle plume theory, there are no concentrated downwellings; settling of the whole mantle compensates the flux in upwelling plumes (Morgan, 1971). Plate tectonic theory focuses on the horizontal advection of thin plates and the sinking of narrow slabs. Upwellings caused by internal heating or displacement by slabs are broad; there are no concentrated upwellings from below the surface BL. Smaller-scale upwellings are passive or secondary. Key issues are whether the conduction part of the BL extends to great enough depth to reach temperatures of ~1600°C and, if so, where and how are these temperatures and materials sampled.

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The petrological lithosphere

Heat is removed from the interior of a convecting fluid through the surface by conduction through the TBL. McKenzie & Bickle (1988) referred to this as 'lithosphere' and used 'TBL' only for the lowermost portion (Fig. 3). They also used 'plate' and 'lid' for the conducting part of the upper mantle. The concept of the lithosphere–asthenosphere boundary (LAB) has been associated with the base of the plate, the depth at which heat transport changes from conductive to advective, the depth to the horizontal isotherm, a critical temperature, and the base of the high-velocity seismic lid. In the Cambridge model these are all the same, by assumption; the plate thickness, sub-plate temperatures and temperature gradients are fixed (~100 km, T_P ~1300°C and adiabatic, respectively). It is assumed that seismic velocity and crustal thickness are proxies for temperature, and that long-term rheological and heat transport properties can be inferred from seismic velocities. The LAB is also assumed to be characterized by a geochemical and fertility transition that is recognizable in both magmas and peridotites.

Many modern mantle petrology papers still use the above assumptions in their interpretations (e.g. Haase, 1996; Herzberg *et al.*, 2007; Herzberg & Gazel, 2009). The most critical assumption is that mid-plate magmas are from localized thermal and chemical anomalies and do not represent ambient shallow mantle. For some time, a T_P near 1280°C was accepted by petrologists as characterizing ambient or average mantle, 'the convecting mantle' and the mantle below ~100 km depth ('the adiabatic interior'). Magmas that had inferred temperatures or depths greater than the ~1300°C isotherm formed, by definition and assumption, below the 'lithosphere' or 'plate' (e.g. Haase, 1996). Data that implied higher temperatures for ambient mid-plate mantle away from hotspots were systematically discounted (e.g. Hillier & Watts, 2005).

Stein & Stein (1992) and Lee *et al.* (2009) inferred significantly higher potential temperatures, ~1400°C, for ambient mantle. Herzberg & Gazel (2009) proposed a T_P range of 1280–1400°C for MORB and ambient mantle and a range of 1370–1600°C for OIB; cold 'hotspot' magmas were attributed to secular cooling of hot plume mantle. Presnall & Gudfinnsson (2010) pointed out that even the most recent petrological estimates of T_P cover a large range: 1243–1488°C for MORB, 1361–1637°C for Iceland and 1286–1722°C for Hawaii.

Watson & McKenzie (1991) adopted a potential temperature (T_P) of 1558°C in the center of a 130 km wide vertical cylinder as a model for the mantle under Hawaii. Priestley & McKenzie (2006) could find no seismic evidence for this and argued that the feature is below the resolution of their data, implying no lateral spreading or mushroom-shaped head. This is inconsistent with an active thermal upwelling under Hawaii, which would naturally spread out beneath the plate. Courtier *et al.* (2007) proposed a T_P of 1431–1462°C for 'ridge-influenced hotspots' and 1453–1501 °C for mid-plate hotspots. The maximum depths inferred for mid-plate lavas closely follow the ~1600°C half-space cooling isotherm, not a horizontal 1300°C isotherm (Haase, 1996). Obviously, with these interpretations, the boundary between ridge and hotspot mantle temperatures is arbitrary, as is the assumption that ridges sample ambient mantle whereas mid-plate magma temperatures reflect local hotspots in the mantle. Heat flow, subsidence and sedimentary data also do not confirm that 'hotspots' are localized thermal anomalies (e.g. Clift, 2005).

Hillier & Watts (2005) noted that T_P based on bathymetry of the Pacific plate is much higher than expected (i.e. higher than MORB). They used crustal thickness as an alternative constraint. Conversely, Niu & O'Hara (2008) argued that crustal thickness, as determined by seismology, cannot be used to constrain T_P and they used bathymetry instead. There is thus a certain amount of circular reasoning involved in the decisions as to what temperatures to adopt for ambient mantle.

Presnall & Gudfinnsson (2010) determined a clear separation between the temperatures of mid-ocean ridge and Hawaiian magmas. They concluded that the P–T conditions for MORB extraction are less than ~1·5 GPa and 1280°C; Iceland and other near-ridge hotspots are the same. Their P–T conditions for extraction of Hawaiian magmas are ~3–6 GPa and ~1300–1600°C. In boundary layers, temperatures increase rapidly with depth and this is part of the reason for low MORB temperatures and variable OIB temperatures.

The boundary layer model (LLAMA) for mid-plate magmatism is based on the seismic properties of Region B and the high-velocity lid, the anisotropy of the low-velocity zone and the decreasing shear velocities with depth (Figs 2–4). In contrast to the usual conventions (e.g. McKenzie & Bickle, 1988), the lid and the conduction layer are not equivalent to lithosphere, the LVL is not the asthenosphere, and the underlying mantle is not required to be on an adiabat. LLAMA is the laterally mobile (advecting and deforming) part of the upper mantle but it transmits heat from the interior mainly by conduction. LLAMA is a laterally extensive, shallow reservoir. Evidence for such a reservoir is widespread but is often attributed to long-distance lateral transport through the shallow mantle from the nearest hotspot (e.g. Duggen *et al.*, 2010) rather than as a ubiquitous feature of the shallow mantle.

LLAMA

The LLAMA concept marries recent developments in seismology, petrology, mineral physics and tectonophysics, and very recent high-resolution geodynamic modeling.

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The laminated structure is created and maintained by the motion of the plate over the hotter and more deformable interior. The melt-rich or potentially melt-rich lamellae are organized by the shear and segregated into fine-grained shear bands; the orientation of the banded structure controls the fast and slow directions of seismic waves. The decrease of seismic velocity with depth is controlled both by the high thermal gradient and the variation of the number and thickness of the Aligned Melt-rich Arrays (AMA) (Fig. 2). The anisotropy of LLAMA, in most places, is consistent with uniform shear. This laminar flow is interrupted at steps and fracture zones in the plate (Fig. 5), and at plate boundaries, and this is where anomalous mantle fabrics and volcanism is expected. Shear-driven upwellings bring deeper hotter material from depth towards the surface. Fracture zones, mid-plate volcanoes and complex anisotropy are related and have nothing to do with the core–mantle boundary.

Relation to Archean tectonics

LLAMA is distinct from, but related to the 3-L LLLAMA concept (Large Laterally Linked Archean Magma Anomalies) of Nisbet (1987). That model implies that the outer layers of the mantle are stabilized by petrological factors such as density and melt segregation rather than by purely physical thermal constraints upon which simple fluid dynamic convection models are based. The outer shell of the present Earth (LLAMA) is composed primarily of buoyant harzburgite, with both ancient fragments and low-melting sills that are organized by shear, rather than by chaotic convection. 'LLAMA' captures both the structure (Lithologically Laminated and Anisotropic Melt Arrays) and the dynamics (Lubricated Lateral Advection of MAntle) of the boundary layer. Ancient Os and the isotopic signatures can be isolated and preserved in shallow refractory and depleted domains. The most ancient, most isolated and best-preserved isotopic domains in the mantle may be exposed by the processes of continental breakup.

Geochemical implications

Evidence from oceanic basalts, mantle xenoliths and abyssal peridotites suggests that old (including Archean) materials are embedded in oceanic plates, particularly under oceanic islands but also in mid-ocean ridge settings (e.g. Simon *et al.*, 2008). Ultra-refractory (thus buoyant) peridotites may represent fragments of recycled ancient 'lithosphere' that were reincorporated into modern plates, and preserved by their size, strength and buoyancy. They thus represent floating mantle reservoirs.

Old refractory lithospheric roots deeper than ~200 km may become sheared off (delaminated) by lateral motions of the boundary layer rather than by dense Rayleigh–Taylor instabilities as usually assumed; harzburgite-rich blobs are buoyant and, if large enough, are gravitationally resistant to recycling into the deeper mantle (e.g. O'Reilly & Griffin, 2006). They are primarily entrained in the advecting flow and stay in the boundary layer as thick high seismic velocity anomalies. Because they are cold, volatile-poor, refractory and buoyant, they are in effect isolated highly resistant inclusions in the plate that resist melting and subduction. In these respects they are similar to cratonic roots and the so-called sub-continental lithospheric mantle (SCLM) reservoir, and they may also carry ancient isotope signatures, including high $^3He/^4He$ ratios.

Predictions

The most ancient surviving parts of the surface boundary layer are expected to be associated with the most ancient cratons and with the breakup of long-lived supercontinents. The associated passive upwelling levitates and exposes materials that have been isolated for long periods of time. The LLAMA model predicts the long-term shallow survival of high $^3He/^4He$ ratios despite convective mixing in the deep mantle. Extremely high $^3He/^4He$ ratios are commonly attributed to the presence of 'undegassed' (high ^3He concentrations) and undifferentiated material preserved deep in, or below, the Earth's 'convecting mantle', in long-lived convective eddies or

Fig. 5. Flow lines in the shear boundary layer induced by a moving plate, with thickness variations or fracture zones.

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stagnation points or in high-viscosity blobs. The effects of diffusion at high mantle temperatures and of convective mixing or chaotic mixing make this unlikely.

The upper parts of the surface boundary layer are protected from convective homogenization by location, buoyancy, and low temperature. The refractory parts, in particular, can preserve ancient $^3He/^4He$ ratios, without being gas-rich, simply by being U- and Th-poor (Anderson, 1998). The evolution of helium isotopic ratios in the mantle depends on the $^3He/U$ ratio, not on the ^3He content. High $^3He/^4He$ ratios can reflect the long-term survival of cold low U/He refractory domains in the shallow mantle; survival is facilitated by buoyancy, low temperatures, low diffusion rates, high melting points, high viscosity and high strength. Buoyancy, strength and lateral advection protect these refractory domains from convective mixing.

Although high $^3He/^4He$ ratios are commonly attributed to long-term survival in a ^3He-rich, undegassed, deep, hot, primitive reservoir, the isotope decay equations are equally consistent with survival in a ^3He-poor, U-poor, cold, shallow refractory depleted reservoir, such as the refractory lamellae in LLAMA or in the sub-cratonic 'lithosphere'. High $^3He/^4He$ ratios do not imply high ^3He concentrations, or a deep undegassed source.

The isotopic compositions and locations of lavas in Baffin Island and West Greenland, at the site of the breakup of an ancient supercontinent, suggest that their source is an ancient accessible shallow ^3He-poor reservoir, exposed by processes of plate tectonics (final opening of the Atlantic), rather than a deep mantle reservoir fortuitously protected for 4·5 Gyr in a convective eddy (e.g. Jackson *et al.*, 2010). The highest $^3He/^4He$ ratios are associated with the final stages of continental separation, not the initial stages as predicted by the plume hypothesis.

GEOTHERM

Bathymetric constraints on mantle temperature

Oceanic swells were initially attributed to high temperatures, rejuvenation and delamination, and were predicted to have thin lids and high heat flow. It is now clear that variations in bathymetry have a large component of non-thermal and isostatic support (Katzman *et al.*, 1998; Hillier & Watts, 2004; Van Ark *et al.*, 2004; Niu & O'Hara, 2008; Yamamoto & Phipps Morgan, 2009; Leahy *et al.*, 2010) and may be of shallow origin (McNutt, 1998; Adam *et al.*, 2010). In the context of LLAMA, it has been proposed that swells and mid-plate magmatism are the result of greater thicknesses of refractory lithologies and shear-driven upwelling (e.g. Conrad *et al.*, 2010) from the hot base of the boundary layer. Mid-plate magmatism, fracture zones, swells, lithosphere steps and mantle anisotropy, or lack thereof, are all related. They are not accidentally juxtaposed and they do not imply long-distance lateral transport of plume material from some remote hotspot (e.g. Duggen *et al.*, 2010). O'Reilly *et al.* (2009) summarized evidence that high-velocity domains in the Atlantic Ocean affect the underlying mantle flow and control the location and chemistry of mid-plate magmas, some of which carry xenoliths with Archean isotope signatures.

The subadiabat (Fig. 1)

Seismic data indicate that mantle temperatures, in general, can be superadiabatic to depths much greater than the depths of MORB extraction, consistent with half-space cooling calculations. On the other hand, the effects of slab cooling and radioactive heating are expected to cause the geotherm to be sub-adiabatic below the surface boundary layer (Jeanloz & Morris, 1987; Sinha & Butler, 2007). The solidi of volatile-free mantle lithologies increase with pressure so the most likely place to find low melting points and high temperatures is in the boundary layer, above the subadiabatic and slab-cooled portions of the geotherm. The top of the LVL may be controlled by melting of carbonated peridotite or eclogite, in which case it is the decrease in the solidus temperature with depth that is important. The density and mobility of CO_2 suggests that it also will collect in the BL and contribute to shallow melting.

Temperatures in the D" boundary layer

Standard assumptions in geodynamic modeling are that the mantle geotherm between boundary layers defines an adiabat and that physical properties are not strong functions of pressure. A critical requirement of the plume hypothesis is that T_p in the D" source is higher than anywhere in the surface BL and that this additional heat can be rapidly delivered to the surface in narrow conduits. Because of plate motions and the effects of pressure on viscosity, expansivity and thermal conductivity, narrow vertical plumes are an unlikely mode of convection at depth, and this is confirmed by high-resolution and thermodynamically self-consistent simulations of mantle flow (e.g. Schuberth *et al.*, 2009). Jagged, sloping 'walls' have been identified deep in the lower mantle by high-resolution seismic imaging (Sun *et al.*, 2010) and have been called 'plume-like'; however, as they are not vertical they cannot have high buoyancy. They do not seem to interact with the TZ or the surface and may therefore be features that are confined to the deep mantle. In addition, long-lived thermal plume boundaries will be diffuse.

Because of secular cooling and internal radioactivity the average mantle geotherm between boundary layers is subadiabatic (e.g. Jeanloz & Morris, 1987) so it is not obvious that the above conditions apply. These effects can lower the inferred temperature in D" by about 400°C compared with models with an adiabatic interior (Sinha & Butler,

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2007). T_p in at least the upper part of D" will therefore be less than in the shallow mantle.

For a given temperature of the core there is a direct tradeoff between the total temperature increases across the top and bottom boundary layers of the mantle (Fig. 1). If the base of the mantle has a high intrinsic density, then the temperature rise across the lower BL must be much greater than the temperature rise across the upper BL to make up for the subadiabatic gradient, the low coefficient of thermal expansion and the lowered buoyancy. Self-consistent computer simulations show that the maximum excess temperatures achieved near the base of the mantle (mean plus fluctuations) do not necessarily exceed the highest temperatures achieved in the upper BL, and when they do, they do not do so over a thick enough interval to create buoyant upwellings. The hottest part of D" is next to a free-slip boundary and this does not provide a stable reference system for fixed hotspots. This, plus the small difference between inferred MORB and OIB temperatures and the possibility that the base of D" may be iron-rich, has led to the suggestion that only the colder parts of D" may spawn plumes. Farnetani & Richards (1994) and Farnetani & Samuel (2005) and other advocates of the D" reservoir overlook the possibility that it is the upper BL, not the lower one, that is responsible for mid-plate volcanism. It is the assumptions, not the data, that rule out this possibility.

SUMMARY

Although the attention of geochemists and geodynamicists has been focused on whole mantle vs layered mantle convection issues, and on deep mantle reservoirs (Farnetani & Richards, 1994; Farnetani & Samuel, 2005; Phillips & Bunge, 2005; Nakagawa *et al.*, 2008) the structure, dynamics and composition of the upper ~200 km of the mantle are emerging as key elements in the petrology and evolution of the mantle (Tackley & Stevenson, 1993; Hieronymus & Bercovici, 1999; Lynch, 1999; Elkins-Tanton & Hager, 2000; Pearson & Nowell, 2002; Raddick *et al.*, 2002; Gao *et al.*, 2003, 2004; West *et al.*, 2004, 2009; Hales *et al.*, 2005; Presnall & Gudfinnsson, 2005, 2008; Hirano *et al.*, 2006; Anderson, 2007*a*, 2007*b*, 2007*c*; Ballmer *et al.*, 2007, 2009, 2010; Pasyanos & Nyblade, 2007; Pilet *et al.*, 2008).

Shear between the plate and the mantle affects both the mantle flow and plate stress. Based on bathymetry and seismic data alone, the ambient mantle at depth in the boundary layer can be hundreds of degrees hotter than the shallow mantle sampled at spreading ridges. Mid-plate volcanoes are, therefore, small-scale samplers of ambient local mantle rather than localized hotspots, and they are expected to have higher and more variable temperatures than shallow and near-ridge magmas.

The upper 220 km of the mantle, Region B, is not uniform in age, stress state, composition, fertility or melting point, and the amount of available magma is not simply a function of absolute temperature. The rarity of high-temperature magmas is related to the difficulty in levitating hot material from >150 km deep in the boundary layer. Perturbations in mantle viscosity and plate thickness can cause local upwellings (Fig. 5). Shear-driven upwellings in mantle wedges, mantle displaced by sinking slabs and through slab windows, and forces associated with continental breakup may also levitate deeper, hotter, older material to the surface.

In the boundary layer model, volcanism, plate architecture and mantle flow are intimately related. The motion of a uniform thickness plate over a homogeneous mantle causes laminar advection in the BL; heterogeneous mantle and plates with steps can perturb this simple flow. The laminated-shear boundary layer mechanism (LLAMA) explains the coexistence of mid-plate magmatism, fracture zones, swells and perturbations of mantle anisotropy. Fracture zones are not just a convenient way for magmas to move out of the mantle; they can be responsible for perturbations in mantle flow. There would be no magmas from the hot base of the BL if plates were uniform and there were no fracture zones or steps to perturb the mantle flow; there would be little mid-plate magmatism other than that accessed by thermal contraction and bending of plates.

Relation to the marble cake mantle

LLAMA contains fertile, metasomatic, enriched, depleted, refractory and ancient components, which are segregated into lamellae, by shear, rather than chaotically stirred into a marble cake. Low-rigidity components segregate into low-angle shear bands, dipping down in the direction of plate motion. In addition to the sheared lamellae there can be large resistant lumps entrained in the shallow flow.

Far away from fracture zones and other tectonic boundaries, mantle dynamics may consist of sluggish, laminar flow. Boundary layers and the adjacent mantle are dynamic, laterally advecting systems; upwellings are secondary. Geochemical heterogeneity along the global mid-ocean ridge system, as sampled by MORB, may mimic a marble cake mantle, statistically (e.g. Meibom & Anderson, 2004; Armienti & Gasperini, 2010), but the lamellae are formed by unidirectional lateral shear and there is no continuous or repeated folding. Geochemical markers, reflecting recycling, melt segregation and metasomatism, coexist and may do so, in part, in a self-similar fashion, but cross-lamellae mixing is limited and the statistics of the heterogeneity are not isotropic. Lateral shearing organizes the heterogeneity such that lateral and vertical scales differ considerably and the melt-rich lamellae are

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constantly re-forming. Natural length scales are the thickness of the BL, the thickness of the lamellae and fracture zone spacings. Self-similar, isotropic, chaotic convection (marble cake) has no characteristic length scale.

DISCUSSION

Can mid-plate volcanoes and volcanic chains be produced by boundary layer dynamics?

The proposed D" source for mantle plumes, and the continental lithosphere, for large igneous provinces, are examples of BL reservoirs. The idea that the surface BL may be a source for mid-plate magmas has been contested (Arndt & Christensen, 1992), ruled out by assumption (McKenzie & Bickle, 1988) or simply overlooked (e.g. Duggen *et al.*, 2010). The usually invoked MORB-source reservoir is 'the convecting mantle' or the asthenosphere, although Yamamoto *et al.* (2007) argued for a plume-fed asthenospheric reservoir.

The largest boundary layer on Earth, by volume, is the surface one, Gutenberg's Region B. If the entire upper BL of the mantle were as mobile as the plate, as cold as the lithosphere *sensu stricto* and as thin as 100 km, it would not be a strong candidate as a reservoir for mid-plate volcanoes or volcanic chains. In the Cambridge model, the BL is half the thickness of the seismologically constrained conduction-shear layer.

A straightforward application of thermodynamics and laboratory calibrations to the estimation of temperature from seismic velocity, phase change depths and bathymetry gives mantle temperatures and temperature variations that are, or can be, much higher than assumed or allowed in the Cambridge model. High inferred temperatures are often discarded as being unreasonable (e.g. Hillier & Watts, 2005; Priestley & McKenzie, 2006; Wolbern *et al.*, 2006; Ritsema *et al.*, 2009). Although there are many factors that affect seismic parameters, and there are reasons to be cautious, it has been common practice to discard data and interpretations solely on the basis of preconceptions or disagreement with the Cambridge model. Even with these biases in the interpretations geophysical estimates of ambient mantle temperatures are generally 50–100°C higher, and lateral variations are much higher, than in the model of McKenzie & Bickle (1988) (e.g. Anderson, 2000).

The following discussion addresses the kinds of questions that can be raised about the plausibility of a shallow, <220 km, laterally extensive mantle reservoir for mid-plate volcanism and the viability of the top-down LLAMA mechanism. Such a reservoir is often invoked (e.g. Sleep, 2007; Duggen *et al.*, 2010, and references therein) but it is treated as a ~1000–5000 km long lateral extension of a vertical plume, or a plume-fed asthenosphere (Yamamoto *et al.*, 2007), rather than a permanent semi-isolated part of the upper mantle. There are shallow and plate tectonic mechanisms for the generation of lateral temperature gradients, large volumes of basalt (e.g. Coltice *et al.*, 2007), and temporal changes in magma temperatures.

What are the geophysical manifestations of boundary layer dynamics?

Features in the lower part of the BL and in the upper mantle proper are relatively 'fixed' in comparison with plate motions. Volcanic chains are therefore parallel to, and have rates set by, upper plate velocities. The motion of a plate over a viscous asthenosphere can result in a variety of motions in the intervening shear BL, ranging from laminar flow to rolls oriented parallel to plate motion to shear-induced upwellings that, along with decompression melting or melt release, can spawn 'hotspot' and 'hot-line' volcanism (e.g. Ballmer *et al.*, 2009, 2010; Conrad *et al.*, 2010).

Progressive delamination, shear-driven upwellings, volcanic loading, plate bending, thermal contraction and extension along pre-existing fracture zones are plate tectonic processes that can produce both short- and long-lived volcanic chains from shallow sources (e.g. Tackley & Stevenson, 1993; Katzman *et al.*, 1998; Hieronymus & Bercovici, 1999; Lynch, 1999; Raddick *et al.*, 2002; Foulger *et al.*, 2005*a*, 2005*b*; Foulger & Jurdy, 2007; Natland & Winterer, 2005; Conrad *et al.*, 2010, and many earlier publications).

High-resolution mantle flow simulations from several groups show that BL dynamics, with no heat or material input from below 240 or 400 km, explains the bathymetry, tomography and magmatism in the hotspot-rich region of the South Pacific (Adam *et al.*, 2010; Ballmer *et al.*, 2010). The same shear that explains mantle anisotropy can elevate even non-buoyant material from deep in the BL.

With only 1–2% melt fraction can the surface boundary layer provide the volumes of melt contained in large igneous provinces?

Removal of a 1% melt fraction from LLAMA would give a surface layer more 2 km thick over the whole globe. A typical large igneous province (LIP) covers an area of $10^6\,km^2$ and erupts about $10^6\,km^3$ of basalt. A prism of mantle $10^6\,km \times 200\,km$ contains $2 \times 10^8\,km^3$ and will fit within the boundary layer under an LIP. Five per cent of this represents $10^7\,km^3$, which is 10 times larger than a typical LIP. If the low-rigidity lamellae in LLAMA are fertile, and represent 1–2% of the volume, then they can provide the observed volumes of even LIPs. Only a small fraction of this would have temperatures as high as the maximum temperatures recorded by oceanic island magmas. Large igneous provinces can also be the result of supercontinent breakup or the displacement of mantle material out of

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the TZ by subducted slabs (e.g. Anderson, 2005, 2007*a*; Coltice *et al.*, 2007).

It is not obvious that D", which is more remote and four times smaller in volume, can provide such volumes in such a short period of time. Considering the subadiabatic gradient that occurs over most of the mantle and the effect of pressure on thermal properties, it is also not obvious that the potential temperature in D" is high enough, over a large enough depth interval, to create a narrow buoyant upwelling that will exceed the T_p in Region B when it reaches the surface.

What is the volume of the upper mantle boundary layer compared with other proposed geochemical reservoirs?

LLAMA extends from about 60 to 220 km and represents 5·8% of the mantle by volume, if it is a global layer. A 200 km thick layer at the top of the mantle contains about four times the volume of a similar layer at the core–mantle boundary (CMB). The continental lithospheric mantle (CLM) and D" are small-volume (*c.* 2·5% of the mantle each) reservoirs that are prominent in discussions about the source of continental and mid-plate oceanic magmatism.

What kinds of magma volumes and eruption rates can LLAMA produce?

It has been argued intuitively that large eruptive volumes require fixed, long-lived source regions in the deep mantle and that appropriate conditions cannot be attained in the convectively homogenized upper mantle (e.g. Konter *et al.*, 2008). Removal of the deep part of the BL by delamination or shearing can create magma volumes of the order of a million cubic kilometers in 1 Myr (Elkins-Tanton & Hager, 2000). A typical delamination event might remove 30 km from the base of the surface BL. For a conductive geotherm this can be up to 300°C hotter than the overlying material. The ambient mantle that replaces it will also be 300°C hotter.

Delamination is usually considered to be a Rayleigh–Taylor density instability, but it can also be due to shearing. The most likely places for convective instabilities and SDU in the oceanic mantle would be beneath suitably oriented fracture zones. Sometimes it is assumed that a delamination event is triggered by a plume but this is unnecessary in the LLAMA model.

Plate thickness and age variations are found at fracture zones, rifts and cratonic edges. Some of these are attributed to isolated fragments of old continental lithosphere because of their inferred isotope chemistry (e.g. O'Reilly *et al.*, 2009). Even if some of the LLAMA lamellae are below the solidus, shear-driven upwellings at the downstream-facing edge of an appropriately oriented FZ, craton boundary or lithospheric step can produce melts at rates comparable with Hawaiian or flood basalt eruption rates (Yamamoto & Phipps Morgan, 2009; Conrad *et al.*, 2010).

Isn't LLAMA just 'metasomatized lithosphere'?

Decompression melting of laminated lithologies with distinct solidi (e.g. metasomatic veins, eclogite, silica-deficient garnet pyroxenite, peridotite, etc.) is similar to the problem analyzed by Pilet *et al.* (2008). Melting during levitation is enhanced by the presence of non-melting components; latent heat refrigerates the refractory lamellae, which are buoyant, strong and relatively cold, and can therefore retain ancient isotopic signals, including otherwise mobile noble gas atoms (Anderson, 1998).

Doesn't the geochemistry of OIB require deep sources?

The reservoirs for mid-plate magmas and their various components are plausibly in the shallow mantle (Natland, 1989; Anderson, 1994, 1998, 2000, 2005; Gallagher & Hawkesworth, 1994; Foulger *et al.*, 2001, 2005*a*, 2005*b*; Doglioni *et al.*, 2005; Foulger & Jurdy, 2007; Winterer & Natland, 2007; Adam & Bonneville, 2008; Adam *et al.*, 2010; Faccenna & Becker, 2010). Most of the geochemical attributes of mid-plate magmas originate in the crust and shallow mantle. Region B is semi-isolated from the so-called convecting mantle by its buoyancy, low temperatures in the upper part, and by its mainly lateral motions.

LLAMA contains both ultra-refractory and ultra-enriched lithologies, consistent with inferences regarding the shallow oceanic mantle (Anderson, 1989; Pilet *et al.*, 2008; Simon *et al.*, 2008; O'Reilly *et al.*, 2009). Metasomatized shallow mantle explains much of the geochemistry of mid-plate magmas (e.g. Pilet *et al.*, 2008). Ultra-depleted harzburgites and mantle wedge and other metasomatized material, some ancient, should float to the top of the mantle. The shallow mantle under volcanic islands—or continents—is not MORB-source mantle; however, the presence of non-MORB source is not a criterion for deep mantle sources.

Many oceanic island xenoliths are too refractory to have formed at the temperatures that are currently hypothesized for the shallow mantle (Simon *et al.*, 2008). The principal geochemical attributes appear to predate the oceanic crust and were already locked into shallow mantle rocks that were incorporated into the plate beneath the islands. This includes the helium, partly carried in fluid inclusions inside refractory olivine crystals. The helium carrier is a long-term resident of the shallow mantle, which is cold and retentive of helium (e.g. Anderson, 1998).

Don't age-progressive volcanic chains require deep stationary sources?

The advective velocity of the deep part of a shearing boundary layer is much less than the surface velocity. For plausible parameters, the velocity at 150 km depth is one-tenth the surface velocity (e.g. Tommasi *et al.*, 1996). The relative fixity of hotspots therefore implies that they

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originate below about 100 km depth, either in or below the boundary layer (Fig. 5). Similarly, the base of D" may be moving faster than the top.

Isn't the upper mantle completely homogenized by convection?

Konter *et al.* (2008) have argued that persistence of magmatism and of geochemical fingerprints is unlikely to be attained in the dynamic upper mantle and thus that these require a deep mantle source. They argued that OIB sources are anchored deep in the mantle, isolated from homogenization by mantle convection. They specifically pointed to the islands in the South Pacific as evidence for hot plumes that have thinned the lithosphere under the islands. Unfortunately, these arguments are not valid (e.g. Ballmer *et al.*, 2007; Adam & Bonneville, 2008; Adam *et al.*, 2010) and do not apply to LLAMA and BL advection. In general, BLs are semi-isolated from the so-called convecting mantle and this applies to Region B as well as to D". The mantle is convecting but this does not imply that it is being homogenized by vigorous, or chaotic, stirring.

Because mid-ocean ridges circle the globe, why are MORB not representative of mean mantle T_p*?*

The original argument for equating MORB with an ambient upper mantle source considered stationary symmetrically spreading ridges, passively sampling the underlying homogeneous mantle that was rising from depth to fill the gap. The T_p at a ridge, however, does not constrain the temperature elsewhere or even the mantle below the depth of magma extraction; the material at that depth may have flowed laterally as well as vertically or may have been overridden by a migrating ridge. The magma extraction depth for migrating ridges may even be in a previously formed boundary layer. Considering that the shallow mantle can flow laterally and that ridges can migrate freely, and that there is evidence for lateral heterogeneity in the BL, it is not obvious that shallow sub-ridge mantle should be identical to the sub-plate mantle under Hawaii. However, this is the key assumption in the Cambridge model.

Are Transition Zone thicknesses (TZTs) consistent with an ambient mantle that is hotter than in standard models of mantle petrology?

The Transition Zone is an imperfect thermometer (Anderson, 1967, 2007*b*; Gu & Dziewonski, 2002; Gilbert *et al.*, 2003; Lawrence & Shearer, 2006; Deuss, 2007; Tauzin *et al.*, 2008) and its thickness variations do not support the deep mantle plume or whole mantle convection hypotheses. Attempts to infer temperature differences between hotspot and normal mantle, or to turn the depths of the 410 and 650 km discontinuities into equivalent temperatures, give mixed results [see discussion following Deuss (2007) in *Geological Society of America Special Paper* **430**]. Discrepancies of more than 100°C are common between laboratory, theoretical and seismological results (Stixrude & Lithgow-Bertelloni, 2005; Wolbern *et al.*, 2006; Courtier *et al.*, 2007). The highest temperature results are often dismissed as unrealistic unless they occur where expected.

Courtier *et al.* (2007) presented a plot of hotspot magma temperatures vs TZT, which shows a weak positive correlation; in other words, the coldest TZ lies beneath the hottest magma source. However, the correlation coefficient is only +0·3, which is not significant. If inferred MORB temperatures are added to the plot, the correlation turns negative, as it should if anomalous sub-ridge mantle extends as deep as 650 km and is cold, but the correlation coefficient is only −0·3 and the trend does not agree with theoretical and laboratory calibrations of Clapyron slopes. On the other hand, if mid-plate magmas are derived from within or just below the surface BL, their compositions and inferred temperatures should correlate with plate age, which they do.

The actual depths of the 410 and 650 km discontinuities are not usually anti-correlated as would be the case for whole mantle vertical plumes, either hot or cold. Below hotspots, the 410 and 650 km discontinuities vary independently, and petrologically inferred T_p and TZT are uncorrelated, implying that hotspots have shallow sources, or they are athermal (Tauzin *et al.*, 2008; see discussion in Deuss, 2007). If mid-plate volcanoes are underlain by locally hot mantle this excess temperature does not extend to depths as great as 650 km. This lack of correlation does not rule out the possibility that sub-plate mantle in general, except near ridges, is hotter than generally assumed.

Where are the 'hottest' regions of the mantle?

Hawaii does not overlie a region of the mantle with the lowest seismic velocities or the thinnest TZ. These attributes are often used, elsewhere, as evidence for plumes but they are seldom compared with values in non-hotspot areas, and are not unambiguous indicators of temperature. In terms of low seismic velocities, TZ thicknesses, and 410 depths, hotspots do not define the 'hottest' mantle (Lawrence & Shearer, 2006; Tauzin *et al.*, 2008; Ritsema *et al.*, 2009). The range of T_p inferred from global TZ thicknesses is ±200°C and is ±100°C over a more restricted area that includes only parts of the northern Pacific and adjacent continents. Some of the thinnest (highest inferred temperature) TZs are in California, eastern North America and hotspot-free areas of the Pacific. In the LLAMA model, the thick lid under Hawaii and the large offset Molokai FZ are responsible for localizing the magmatism.

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What is the evidence for subplate temperatures away from ridges?

In the Cambridge model ambient mantle is assumed to have the same potential temperature as the shallow mantle at ridges. Priestley & McKenzie (2006) claimed that the temperature at 150 km depth is within 20°C of 1400°C throughout the entire Pacific, but they required their geotherms to be asymptotic to the MORB adiabat. In other words, temperatures are forced to converge as they approach 1400°C. Their constrained temperatures are more than 300°C lower than would be inferred from the same shear velocity without an enforced cutoff (e.g. Stixrude & Lithgow-Bertelloni, 2005; Schmandt & Humphreys, 2010). Ambient mantle temperatures may be hundreds of degrees hotter than assumed by McKenzie & Bickle (1988) and subsequent workers.

What are normal mantle temperatures?

The bathymetry of most of the Pacific, particularly the western part, is considered 'anomalous', which means shallower or 'hotter than it should be' (e.g. Korenaga & Korenaga, 2008). When these regions are filtered out, there is very little area left from which to infer 'normal' depths and 'unperturbed' mantle temperatures. Nevertheless, the best-fitting cooling plate model based on filtered and reprocessed bathymetry data yields mantle temperatures for 'normal' regions of the North Pacific that are ~200°C higher than in the model of McKenzie & Bickle (1988) (Hillier & Watts, 2005). A similar increase is required by seismic velocities in the LVL if the results are not constrained to approach the ridge adiabat at depth.

The P–T–age trajectories calculated by Haase (1996) for mid-plate magmas track the 1500–1600°C cooling half-space isotherms, rather than the predicted 1300°C horizontal isotherm. The maximum depths of melting, and inferred temperatures, increase with plate age. These observations are consistent with mid-plate magmas being extracted from within and near the base of the boundary layer and with the ambient mantle being ~200°C hotter than generally assumed.

Is the mantle nearly isothermal below 100 km depth?

If the mantle below the plate really is isothermal and homogeneous, as usually assumed, then the seismic velocities should reflect this. The peak-to-peak amplitude of V_s heterogeneity in the upper 200 km of the mantle exceeds 7%, implying temperature variations of >700°C, unless melting intervenes (Schmandt & Humphreys, 2010). Large lateral variations in shear velocity extend to depths of 220 km in the Pacific (e.g. Ekstrom & Dziewonski, 1998; Shapiro & Ritzwoller, 2002). The peak-to-peak amplitude of lateral V_s heterogeneity between 150 and 200 km depth in the Pacific mantle exceeds 3·5%. Ritzwoller *et al.* (2004) inferred lateral temperature variations of more than ±100°C in the Pacific mantle at depths of 150 km. Priestley & McKenzie (2006) inferred much lower variations but they constrained the deeper part of the geotherm and disallowed high temperatures. The range of T_p inferred from Transition Zone thicknesses over a part of the northern hemisphere that includes the North Pacific (e.g. Ritsema *et al.*, 2009) is 1450 or 1353 ± 100°C, depending on modeling assumptions.

Why is the evidence that mid-plate mantle is ~200°C hotter than MORB mantle overlooked?

Best-fitting models for updated and 'corrected' bathymetry data imply a mean mantle temperatures that is 170°C higher than assumed by McKenzie & Bickle (1988) (Hillier & Watts, 2005). This was considered by those workers to be 'unreasonably hot' because 'it is much hotter than estimates of normal mantle temperature from mid-ocean ridge basalts'. They then constrained the basal temperature of the plate to be ~1350°C. The data had already been reprocessed to avoid regions that were suspected of being hot.

The slow subsidence rates of near-ridge plates imply that sub-ridge mantle is denser, possibly colder, than the mantle under older plates (Hillier & Watts, 2004). This is consistent with seismic velocity data that show that near-ridge mantle below 200 km depth has higher shear velocities than mid-plate upper mantle. Thus, geophysical data confirm that MORB-source mantle and mid-plate mantle temperatures differ by ~200°C but the higher temperatures are widespread and reflect ambient mantle under plates, whereas ridge temperatures appear to be localized under spreading centers. Hawaii is not a localized thermal anomaly; it is a small-scale sample of ambient mid-plate mantle.

If ambient mantle is ~200°C hotter than MORB, why are high T_p magmas so rare?

The laminar flow induced by plate motions ordinarily prevents the deeper hotter parts of the BL from being sampled. This flow is upset by fracture zones, delamination and convergence. In the LLAMA model, high temperatures reflect conditions in the lower parts of BLs and mantle wedges; hot and variable temperature BL material is displaced upwards by descending slabs or driven upwards by plate shear (Fig. 5).

For example, model T_p derived from petrology for dry back-arc basin basalts ranges from 1350 to 1500°C (Kelley *et al.*, 2006). One expects the mantle under back-arc basins to be cooled by the underlying slab but these temperatures exceed MORB temperatures and overlap hotspot temperatures.

What about constant thickness oceanic crust?

The thickness of the oceanic crust, plus the assumption that the mantle is homogeneous, has been used to provide

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temperature and degree-of-melting constraints in some petrological models (e.g. McKenzie & Bickle, 1988; Hillier & Watts, 2005). The assumption that thick seismic crust equals low-density igneous crust equals high temperature equals hot jet fails when considering the elevations and subsidence histories of Iceland and the Ontong–Java Plateau (e.g. Menke, 1999; Gudmundsson, 2003; Korenaga, 2005), and these remain as major conundrums of the hot jet hypothesis. Much of the crust in slow-spreading areas is thin, as appropriate for cold mantle, but is often composed of serpentinized peridotite; thus, the assumption that seismic crust can be equated to the products of mantle melting fails for both deep and elevated areas of the ocean.

Does the recent identification of low-velocity features in the deep mantle not prove the existence of high upwelling plumes?

A tilted, broad, low-velocity anomaly apparently extends from the CMB beneath the SE Atlantic Ocean into the upper mantle beneath eastern Africa, more than 45° away to the NE (Ritsema *et al.*, 1999). A different study placed a tilted narrow low-velocity feature at the CMB under the southern tip of Africa that narrows upward and towards the NE (Sun *et al.*, 2010). TZ studies do not confirm these features as whole mantle thermal anomalies.

CONCLUSIONS

The dynamics of melt and shear localization in partially molten aggregates, and the role of shearing in the upper mantle boundary layer (Tommasi *et al.*, 1996, 2006; Katz *et al.*, 2006; Kohlstedt & Holtzman, 2009; Kohlstedt *et al.*, 2010; Yoshino *et al.*, 2010) resolve some long-standing controversies in mantle physics and petrology regarding the role of magma in influencing seismic velocities and plate tectonics, the origin of mantle anisotropy, and the origin of the large travel-time delays that have been attributed to deep mantle plumes (when anisotropy and the shallow mantle are ignored).

The mainly horizontal flow that characterizes plate tectonics is disrupted at lithospheric discontinuities such as fracture zones, plate boundaries and continental edges. These disruptions are evident in the fabric of the shallow mantle as changes in the anisotropy and volcanic activity, both intermittent and long-lived.

Geophysical data are consistent with temperatures as high as ~1600°C in the upper mantle. Bathymetry data imply mean temperatures >1500°C at low-velocity zone depths, with excursions of >150°C, consistent with back-arc basin basalts and basalts from other tectonic regions where deeper mantle might be displaced upwards. Surface subsidence rates and seismic velocities at depth are consistent with colder than average upper mantle under ridges and higher temperatures under older plates. Lateral temperature gradients below 200 km may contribute to observed subsidence rates and magma temperatures.

In the boundary layer model, Hawaiian and other mid-plate magmas are also relatively hot because they are extracted from tens of kilometers deeper in the mantle than MORB.

Geophysical data are inconsistent with a localized high-temperature anomaly under Hawaii, but are consistent with ambient mid-plate temperatures ~200°C higher, below 150 km depth, than in standard models. The Hawaiian thermal anomaly is with respect to ridges, not with respect to the surrounding mantle. If anything, the local seismic mantle around Hawaii, and other hotspots (e.g. O'Reilly *et al.*, 2009), is a high-velocity anomaly. This does not conflict with teleseismic body-wave studies that use only near-vertical relative arrival times (e.g. Wolfe *et al.*, 2009). The data in such studies simply confirm that half of the arrival times, over the limited area of the study, are slower than the other half.

Melt segregation and a veined or laminated mantle are created by the shear between the plate and the rest of the upper mantle. This zoning may explain geochemical evidence previously attributed to radially zoned thermal upwellings. Swells, fracture zones, volcanoes and variations of mantle anisotropy can all be related to the dynamics and embedded heterogeneities, some ancient, in the surface boundary layer.

Shearing and largely lateral advection are fundamental and important concepts in LLAMA. They are responsible for shear-driven upwellings and, counterintuitively, for the relative fixity of hotspots. Upwellings are secondary aspects of lateral flow; they do not have to initiate from below the BL and they do not have to be driven by their own buoyancy. By contrast, long distance lateral flow, or upside-down drainage, is an amendment, but a critical one, to the vertical plume hypothesis.

Magma volume depends on the fraction of melt-rich lamellae and is not a unique function of absolute temperature. Seismic velocity and anisotropy also depend on the volume fraction of melt-rich lamellae. Melt release occurs when laminar flow in the boundary layer is disrupted, or the plate is breached; melt is focused by the melt-free lamellae.

Ancient buoyant refractory domains are part of the permanent fabric of tectonic plates and the advecting adjacent mantle; these show up in the xenolith population and are evident in the chemistry of basalts erupted through or beside them.

CONCLUDING REMARK

Despite its many shortcomings, the mantle plume hypothesis is widely accepted, partly because alternative mechanisms, until very recently, have lacked the quantitative modeling that demonstrates their feasibility. Tackley

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(2006), for example, asked 'if plumes are not the answer, what is?' Important studies regarding boundary layer dynamics and chemistry have appeared since that question was asked and since this paper was submitted for publication (e.g. Humphreys & Niu, 2009; O'Reilly *et al.*, 2009; West *et al.*, 2009; Adam *et al.*, 2010; Armienti & Gasperini, 2010; Ballmer *et al.*, 2010; Conrad *et al.*, 2010; Faccenna & Becker, 2010; Paulick *et al.*, 2010; Schmandt & Humphreys, 2010). Taken together, these studies reinforce the conclusions reached here that deep mantle plumes are not required to explain the geophysical and petrological data and they provide tectonic, or plate, mechanisms for mid-plate magmatism.

The LLAMA model is distinct from other explanations of mid-plate magmatism. It is made plausible by only recently obtained ability to model boundary layer and plate thickness scales of mantle dynamics. Prior to these advances, mantle geochemistry and convection discussions focused on whole mantle vs layered convection models. Here I have shown that mantle studies that incorporate composition, anisotropy and high-resolution imaging of the shallow mantle into seismological and geodynamic modeling are able to explain data that have previously been interpreted in terms of deep thermal upwellings.

Gutenberg's Region B (the region of the mantle between the Mohorovičić and Lehmann discontinuities), which is often ignored in geochemical and geodynamic speculations and in body-wave tomography, is more important in mantle geochemistry and petrology than Bullen's Region D". The arguments used in this study, and very recent high-resolution fluid dynamic simulations, counter the arguments that are commonly used against the idea of shallow and heterogeneous reservoirs, plate tectonic and athermal mechanisms for mid-plate magmatism, and the presence of melt in the upper mantle (e.g. DePaolo & Manga, 2003; Campbell & Kerr, 2007; Herzberg *et al.*, 2007; Laske *et al.*, 2007; Sleep, 2007; Konter *et al.*, 2008; Hirschmann, 2010). LLAMA is distinct from both the low-resolution isotropic global body-wave tomographic models and the mechanisms (propagating cracks, delamination, small-scale convection, fertile blobs and the Cambridge cooling plate model) that have been used to criticize various mechanisms for mid-plate magmatism.

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APPENDIX A

The Editor of *Journal of Petrology* and reviewers of this paper requested that I discuss recent reports in *Science* (Montelli *et al.*, 2004; Wolfe *et al.*, 2009) that appear to image mantle plumes and to contradict numerous other seismological studies and the main thesis of this paper. These studies, as applied to Hawaii, are examples of Vertical Tomography (VT) in which one attempts to map the mantle below a seismic array with essentially vertical seismic rays that travel large distances before and after they enter the region of interest. By equating low relative seismic velocity to high absolute temperature and low density, images under Brazil, Britain, Hawaii, the Azores, Iceland and elsewhere have been interpreted by the Carnegie group (e.g. VanDecar *et al.* 1995; Wolfe *et al.* 1997, 2002, 2009; Yang *et al.*, 2006; Shen *et al.*, 1998*b* and others) as narrow thermal plumes or decapitated plume heads (e.g. VanDecar *et al.* 1995; Arrowsmith *et al.* 2005). These studies, based on travel times of body waves, are good for mapping the deep mantle, and the region near the core–mantle boundary where the rays are nearly horizontal and provide good coverage. This type of data, however, is not appropriate for mapping the shallow mantle under the receivers. The shallow mantle is best mapped using a combination of surface waves and body waves that interact with upper mantle discontinuities, as discussed in the main text.

Wolfe *et al.* (2009) used only relative times in their study area around Hawaii and determined that half of the arrivals were slower than the other half but did not compare their data with any global reference model. There therefore can be no contradiction with studies that constrain absolute velocities. Montelli *et al.* (2004) had no constraint on the upper 300 km, the subject of this paper. Any shallow anomaly is smeared along the ray. Neither study took the strong shallow anisotropy into account or compared their results with random or non-hotspot places in the Pacific.

Earlier claims of whole mantle plumes (e.g. Bijwaard & Spakman, 1999) based on these methods were clearly artifacts of data selection and processing, orientation and cropping of cross-sections, and the color schemes used to present the results (e.g. Keller *et al.*, 2000; Foulger *et al.* 2001). These early studies all used only near-vertical ray paths, ignored upper mantle anisotropy and assumed that the shallow structure, the subject of this paper, did not affect inferences about deep structure. Other applications of this method to Hawaii and Iceland include studies by Wolfe *et al.* (1997) and Lei & Zhao (2006), who also interpreted their results in terms of hot deep mantle plumes. Higher resolution, higher accuracy and less ambiguous techniques are discussed in the main text. Examples of these, including evidence that the mantle under Hawaii has PREM-like or faster than average seismic velocities, include studies by Ekstrom & Dziewonski (1998), Katzman *et al.* (1998), Ritzwoller *et al.* (2004), Levshin *et al.* (2005), Maggi *et al.* (2006), Priestley & McKenzie (2006), Lebedev & van der Hilst (2008), Kustowski *et al.* (2008), and Ferreira *et al.* (2010).

The VT technique has been used to propose the existence of vertical or tilted low-velocity cylinders or blobs under Britain, Germany, France, Brazil, Scandinavia, Iceland, China, eastern Australia, the North Pacific, and Yellowstone, as well as under isolated oceanic islands.

Limitations of vertical body-wave tomography

The limitations of VT, and the plume-like and slab-like artifacts that it produces, are well known and extensively documented (e.g. Green, 1975; Zhou *et al.*, 1990; Masson & Trampert, 1997; Keller *et al.*, 2000; Foulger *et al.*, 2001). Evans & Achauer (1993) and West *et al.* (2004) gave examples of how teleseismic body waves, used alone, can make it appear, convincingly, that shallow structures extend into the deep mantle. Masson & Trampert (1997) and Priestley & Tilmann (2009) demonstrated the method's limitations for studying the shallow mantle and its inability to retrieve some types of structures, even in the target volume. Global travel-time models, for example, are unable to retrieve the global mid-ocean ridge system, which has been dominant in other forms of tomography for more than 25 years (e.g. Nataf *et al.* 1984). In the case of Hawaii, the mantle above about 200–300 km is inaccessible by the method, but the unknown structure will be smeared into the target volume. The method is often supplemented with Occam's Razor; the inverted structure is required to be smooth and to not deviate much from a preferred model. Upper mantle anisotropy has a large effect on near-vertical rays but is ignored in both the applications of the method and discussions of its limitations.

Relative vs absolute travel times

The VT travel-time method as implemented by the Carnegie group does not use absolute delay times or velocities, so the results cannot contradict evidence for high or average absolute seismic velocities at depth under Hawaii and other hotspots. What they demonstrate,

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unsurprisingly, is that half of their arrivals are later than the other half. However, their color images suggested to them that Hawaii is the result of an upwelling high-temperature plume from the lower mantle. This kind of visual interpretation of false color images involves assumptions about relations between relative arrival times, absolute arrival times, seismic velocities, absolute temperatures and buoyancies. Wolfe *et al*. (2009) stated that they could not rule out the possibility that there may be shallow complexities (e.g. fracture zones, lithospheric steps, anisotropy) that would fit the data equally well as a deep smooth plume structure, but their inclination was to choose the smoother deeper interpretation and to interpret the relative arrivals that are later than the local average as evidence for high absolute temperature.

Global travel time models

The Princeton group (Montelli *et al*., 2004) claimed to have pioneered a new technique that resolves plume tails where other methods have not. They argued that their evidence for plumes is strong, despite their inability to detect mid-ocean ridges and the 'surprising' absence of plume heads in their models. The poor resolving power and uniqueness of travel-time data, and artifacts of their smoothing procedures, have raised doubts about the claims of the Princeton group [see van der Hilst & de Hoop (2005) and Boschi *et al*. (2006)]. The locations of low-velocity features in body-wave travel-time tomography, which are interpreted as upwelling plumes, correlate with the data distribution and are probably artifacts. The high-amplitude low-velocity anomalies occur mainly for small anomalies, whose resolution by the data used is questionable.

Given the ray path coverage of teleseismic body waves, it is possible to obtain fictitious narrow anomalies even from synthetic input data calculated without such structures. In fact, it is almost impossible not to. A particularly dramatic example is illustrated in the study by Lei & Zhao (2006) for Hawaii. If only teleseismic and near-vertical rays are used in the inversion one is guaranteed to retrieve only the low-velocity parts, in attenuated and smeared form, and one concludes that there is a low-velocity near-vertical cylinder centered under the seismic array. Similar features result if the unresolved shallow mantle is anisotropic.

The elusive Hawaiian plume

The Princeton travel-time model is superseded by the MIT model (Li *et al*., 2008), which is constrained by many more data. Nevertheless, the MIT model also fails to recover well-known features of the upper mantle, such as the global mid-ocean ridge system, which are prominent in well-constrained tomographic models. The Princeton, Carnegie and MIT models all disagree about the location of the elusive Hawaiian plume, and disagree with Wolbern *et al*. (2006), who showed that the mantle above 400 km is laterally heterogeneous and exhibits S-delays that vary as much as the delays attributed by the other groups to deep thermal plumes. These results highlight the fundamental limitations of methods that use body-wave travel-times alone.

Effect of anisotropy

It has been known since the 1960s that the uppermost mantle is anisotropic with $V_{SH} > V_{SV}$. Ignoring anisotropy is known to bias depth distribution of heterogeneity (Anderson & Dziewonski, 1982). This kind of anisotropy slows down nearly vertical shear waves more than shear waves at other angles of incidence, and this alone may invalidate the results of many studies. The mantle under the central Pacific has unique and extreme anisotropy. A seismic array installed over an anisotropic upper mantle will map out a vertical low-velocity cylinder with the approach used in many studies. The usual correction of near-vertical travel-times to the vertical, for ease of interpretation, makes things worse.

The state-of-the-art in seismology

Seismology has progressed far beyond the simple isotropic models that are derived from VT. Examples of the state-of-the-art in global, corridor, and regional imaging include the studies by Trampert & Woodhouse (1995), Ekstrom & Dziewonski (1998), Katzman *et al*. (1998), Ritzwoller *et al*. (2004), Levshin *et al*. (2005), Maggi *et al*. (2006), Priestley & McKenzie (2006), Song & Helmberger (2007*a*, 2007*b*, 2007*c*), Kustowski *et al*. (2008), Lebedev & van der Hilst (2008), Sun *et al*. (2009, 2010) and Ferreira *et al*. (2010). Those studies that included Hawaii showed that the upper mantle around Hawaii is either not anomalous or has high V_s, and that there are many non-hotspot areas that have much lower V_s than Hawaii, including western North America.

Sun *et al*. (2010) imaged a sharp narrow feature in the deep mantle under southern Africa that appears to extend upward from a D" anomaly. They called this feature a 'plume'. The sharpness of the feature precludes it from being a long-lived thermal plume. High-velocity features imaged in the deep mantle were called 'slabs'. There is no evidence, however, that these features pass through the TZ or that the 650 km discontinuity is perturbed by their passage. They are either athermal or are not rising or sinking. Present data suggest that both the upper and lower boundary layers are complex and dynamic regions, but there is no evidence that they interact or send material deep into the interior of the mantle. In spite of intensive mapping by a variety of techniques there is no evidence that the small-scale, 100–200 km, features found in the deep mantle extend upwards into the TZ or to the surface, or from the surface into the deep mantle, including under Africa (Pasyanos & Nyblade, 2007).

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