Deep mantle plumes and convective upwelling beneath the Pacific Ocean

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A B S T R A C T

Earth’s mantle is thought to convect as a whole, with material flow across the upper mantle phase transitions of the mineral olivine at 410 and 660 km depth. However, the details of convection, especially mantle upwelling and plumes, are not well constrained. Here, we study seismic shear wave reflections from the underside of these temperature and composition dependent phase boundaries, which we resolve to be relatively flat beneath most of the Pacific, except under subduction regions and volcanic hotspots. The phase boundaries are closer together beneath the Hawaiian hotspot and also in a larger region of the South Pacific that is flanked by a number of volcanic hotspots. This region overlies the southern part of the large-scale low shear velocity province in the lowermost mantle. A large plume head or cluster of several plumes originating in the lowermost mantle and impinging upon the South Pacific 660 km discontinuity is consistent with observed phase boundary topography and subduction patterns. This feature may be related to large volume volcanic eruptions, such as the Cretaceous Ontong Java Plateau flood basalt, which have been proposed to originate in the South Pacific.

1. Introduction

Convection in Earth’s interior drives tectonic forces that shape the surface of the planet. It is widely accepted that oceanic crust and lithosphere are created at mid-ocean ridges and consumed back into the planet at subduction zones, a process generally agreed to be fundamentally linked to mantle convection. Seismic tomography shows that anomalously high seismic velocities presumed to be cold subducted material (or “slabs”) descend through the phase transitions of the mineral olivine near 410 and 660 km depth (van der Hilst, et al., 1997). Inversion modeling using gravity and post-glacial rebound data (Mitrovica and Forte, 1997) suggests a strong viscosity increase accompanies the 660 km deep phase transition which may impede or slow the descent of some slabs (Ribe, et al., 2007). However, most slabs appear to have an ultimate fate of being subducted deep into the lower mantle, in some places to the core mantle boundary (CMB) at ~2891 km depth (Grand, et al., 1997). Upwelling mantle flow and plumes are less well understood, but should occur away from downwelling slabs. Most of Earth’s hotspot volcanoes are situated above assumed return flow, giving rise to suggestions that the core–mantle boundary may be the source of mantle plumes that lead to surface hotspot volcanism (Morgan, 1971). Deciphering the dynamics and evolution of Earth’s interior requires knowledge of mantle thermal structure, which can be inferred from tomographically derived seismic wave speeds; however, this is often accomplished assuming an isochemochemical deep mantle, conflicting with recent work (Trampert, et al., 2004).

Earth’s mantle beneath the Pacific Ocean is an ideal location to investigate mantle structure: it includes the largest single tectonic plate on Earth (the Pacific Plate) with active or recently active subduction zones defining a significant portion of the ocean’s perimeter, and a number of hotspots and fast spreading mid-ocean ridges occupying its interior. Furthermore, a large number of seismic sources recorded by a growing number of increasingly available receivers surround the Pacific, enabling dense seismic wave sampling of the region. Numerous hotspots form linear chains of successively older volcanoes, thought to be the surface expression of stationary plumes of hot ascending material. For some hotspots, such as Hawaii, geochemical (Courtillot, et al., 2003; Putirka, 2005) and geophysical (Li, et al., 2000; Montelli, et al., 2006; Courtier et al., 2007b) evidence support a hot, deep mantle origin as the source of this intraplate volcanism. However, the source of magmas for many hotspots remains either uncertain or unresolved (Courtillot, et al., 2003). Whether or not plume “plumbing” is restricted to cylindrical conduits, vertical or contorted, remains in debate. Alternate possibilities for hot spot magma sources based on geochemical (Albarede and van der Hilst, 2002) or geodynamical considerations (Tackley, 2008) include isolated blobs in the lower mantle or upper mantle reservoirs (Anderson, 2006). Thus, improving our understanding of the thermal and chemical structure of the mantle beneath the Pacific will help constrain the nature of plumes within the mantle, as well as the style and mode of mantle convection within the Earth, which is necessary before the evolution of Earth’s interior can be confidently deciphered.

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Lateral variations in upper mantle temperature and/or chemistry will perturb the depths of the solid-to-solid phase transitions in the dominant mantle mineral olivine. The exothermic transition of olivine \((\alpha)\) to wadsleyite \((\beta)\) at 410 km depth (Katsura and Ito, 1989), and the endothermic dissociation of ringwoodite \((\gamma)\) into Mg-perovskite \((\text{pv})\) and magnesiowüstite \((\text{mw})\) near 660 km depth (Ito and Takahashi, 1989), each produce a discontinuous increase in seismic wave velocity and density. The mantle transition zone (MTZ) is defined as the depth shell between these two major seismic discontinuities. Each discontinuity is of finite thickness that can be modified by passive upwelling flow (e.g., Solomatov and Stevenson, 1994). The Clapeyron slope (or the slope of the change in pressure associated with a change in the temperature at a phase transition boundary) of the \(\alpha\) to \(\beta\) phase transition at 410 km depth is positive, and the slope of \(\gamma\) to \(\text{pv} + \text{mw}\) transition at 660 km depth is negative, resulting in the 410 and 660 km phase boundaries shifting in opposite directions in response to the same thermal anomaly. For example, a region of high temperature thins the MTZ. Conversely, low-temperatures will thicken the MTZ. Thus, excluding large chemical heterogeneities in the mantle, MTZ thickness variations provide an estimate of mantle temperature perturbations.

2. Seismic dataset

Here, we seismically examine the MTZ beneath the Pacific using transversely polarized shear waves that reflect from the underside of the 410 and 660 km discontinuities, as well as the surface, midway between an earthquake and receiver. The S-wave that reflects from Earth’s surface, denoted SS, arrives several hundred seconds after the discontinuity reflections (Fig. 1a). The discontinuity reflections, called SS precursors, depend on the discontinuity structure at the reflection location; perturbations in discontinuity depth incur deviations in their travel times. In contrast to other techniques for observing the discontinuities (see an extensive review by Shearer, 2000), SS waves provide a nearly complete sampling of the Earth, making them an ideal seismic probe of upper mantle discontinuities at both global and regional scales. To construct our dataset, we selected earthquakes with a moment magnitude \(M_w \geq 5.8\) to ensure a high-energy SS arrival, source depths \(\leq 75\) km to avoid interference with near-source surface reflections (i.e., “depth phase” arrivals), and source–receiver geometries in the epicentral distance range of 80–165°. This resulted in a dataset of over 130,000 SS waveforms recorded at broadband seismic stations, that densely sample the mantle beneath the Pacific Ocean in unprecedented detail. A stringent signal-to-noise ratio criterion was employed, resulting in retention of 17,000 of the highest quality data (Fig. 1b). This regional dataset beneath the Pacific is similar in size to the entire datasets of past global studies (Flanagan and Shearer, 1998; Gu and Dziewonski, 2002) and takes advantage of EarthScope’s recently deployed USArray, as well as nearly 20 years of broadband instrument seismograms available through the Incorporated Research Institutions for Seismology. The data sample a region constituting 20%
of Earth’s surface area, and were instrumental in providing the first-ever dense sampling of the southern Pacific. Broadband data permits investigation of discontinuity topography at the shortest possible scale lengths, e.g., ≤1000 km lateral scales, as these instruments retain sensitivity to seismic energy at shorter wavelengths.

3. Seismic stacking method

SS precursors typically have amplitudes that are 1–10% of the SS wave, requiring the stacking of many seismograms in order to bring coherent precursory arrival energy out of the background noise (Shearer, 1990). A variety of innovative approaches for studying SS precursors exist (Deuss and Woodhouse, 2002; An, et al., 2007; Schmerr and Garnero, 2007; Houser, et al., 2008; Lawrence and Shearer, 2008). Here, we implement a stacking procedure detailed in (Schmerr and Garnero, 2006; Schmerr and Garnero, 2007), similar to the methodology of (Flanagan and Shearer, 1998), with several special considerations that allow us to assign travel time variations exclusively to the precursors, eliminating a number of effects that can contaminate or obfuscate the resulting discontinuity topography. Stacking is done with both data and reflectivity synthetic seismograms (Fuchs and Müller, 1971), where synthetics were made for each source and receiver combination to test the effects of waveform on the stacked results. Fig. 1c shows stacking results for both the Pacific dataset and corresponding PREM synthetic seismograms. To investigate regional discontinuity structure, the dataset is organized into 484 geographic bins each with a 1000 km radius. Bin locations were originally defined every 500 km on a regularly spaced grid, and are subsequently adjusted according to the average SS bouncepoint location in each bin. The exact locations of the 484 bins used in this study are displayed in Fig. 1b and Supplemental Fig. S1. The stacks of the SS precursors off the 410 km and 660 km phase boundaries correspond to the bin location map.

We stack S410S and S660S precursors along their predicted time-distance moveouts (i.e., their “slowness”, or inverse of the seismic ray parameter) for each precursor, namely, the PREM (Dziewonski and Anderson, 1981) slownesses of S400S and S670S (Fig. 2). We use the 1-D velocity model PREM as a reference background for our study. Note that in the PREM model, the major discontinuities are at depths of 400 and 670 km (i.e., not 410 and 660 km), and hence analyses of stacks made from synthetic seismograms in this paper result in retrieving discontinuity depths at the PREM values, e.g., as in Fig. 1c. Stacking near the observed precursory slowness produces sharper, better-defined peaks for each precursor (Supplemental Fig. S2), stacking on an intermediate slowness does not significantly change the arrival times of the precursors, but slightly defocuses each precursor waveform in the stack. Thus, in this paper, we independently stack on each precursory slowness to enhance the precision of our retrieved discontinuity depths.

A primary point of concern for stacking is a variety of other seismic phases that arrive in the time window of interest and constructively or deconstructively interfere with the amplitudes of the SS precursors. One example includes topside reflections off the discontinuities, such as s660Sdiff and s410Sdiff, which directly interfere with S410S and S660S, respectively. Also, the precursors to ScSScS, i.e., Scs660ScS, cross over in time with the S410S arrival; Fig. 2c shows the time-distance behavior, i.e., the moveout, of these arrivals. We exclude particular epicentral distance and timing windows (Schmerr and Garnero, 2006) to avoid

![Fig. 2. The seismic wave field surrounding SS stacked in 0.5° epicentral distance bins for the Pacific dataset. a) The number of seismograms in each stack. b) Stacked seismic amplitudes where seismograms are aligned in travel time and amplitude to that of the SS arrival (shown at zero time), with the first positive swing of SS aligned on zero (positive amplitudes are blue and red amplitudes are negative). Underside reflections off the 410 and 660 km discontinuities are seen as the nearly horizontal bright arrivals between 100 and 200 s before SS. c) Predicted travel times for a variety of seismic phases near SS and the precursor wave field. Note that while this is the transverse component of motion, several vertically polarized seismic phases such as PPS and PSKS are visible.](image-url)
contamination of S410S and S660S travel time perturbation estimates introduced by these additional phases (this approach is further detailed in the Supplemental methods and Figs. S3 and S4).

To further avoid contamination from non-precursory energy, we weight each seismogram in the stack based upon a signal-to-noise ratio (SNR) measurement of the SS pulse and the precursory wavefield (Schmerr and Garnero, 2006). This gives seismograms that have noise in the precursory wavefield a lower weight than those with no noise at the expected arrival time of the precursory wavefield. This is justified, as the precursors are weak arrivals compared to the SS pulse, often near the amplitude level of the noise in the seismogram. Without weighting, seismograms possessing large degrees of noise in the precursory wavefield dominate over those with very impulsive SS arrivals and relatively high SNR values. Assigning a high weight for earthquakes possessing impulsive SS signals ensures the formation of well-defined precursory arrivals from higher quality seismograms.

Our stacking approach employs a bootstrap-resampling algorithm (Efron and Tibshirani, 1986) evaluated at the 95% confidence level (2σ) that quantifies the variability (and hence reliability) of every stack and resulting discontinuity depth estimate. We apply our bootstrap histogram approach (Schmerr and Garnero, 2007) to detect multiple discontinuities and complex behavior in the stacked waveforms. In each stack, we take 300 random bootstrap resamplings of the data (Efron and Tibshirani, 1986), allowing replacement, and use the resulting set of stacked seismograms to evaluate the 2σ confidence interval for amplitudes that fall above zero at each discrete point in time. Stacked amplitudes that satisfy this requirement are termed "robust." In general, the robustness of each stack is strongly dependent upon the number of seismograms used in each stack (see Supplemental methods and Supplemental Fig. S5). In addition to amplitude, we measure the travel time of the maximum amplitude falling within ±15 s of the predicted precursory travel time on each of the random bootstrap resampled stacks. This allows the construction of a histogram of the travel time estimates of a given precursor (for every bin). These times are converted to depth by introducing theoretical reflectors within the PREM model above, below, and throughout the transition zone (and interpolating travel times for depths between these layers (Schmerr and Garnero, 2006)), then comparing the observations to the predictions. The histogram approach allows the detection of multiple discontinuities, if present (e.g., Deuss and Woodhouse, 2002), and also provides a measure of the spread in the depth of the discontinuity and presumably the noise in the stack.

Owing to the low amplitude nature of the precursors, it is necessary to use a reference phase to align and stack the data. This reference phase is the SS arrival, as it has a raypath, slowness, and waveform that is the SS arrival, as it has a raypath, slowness, and waveform that is most consistent with the expected arrival time of the precursory wavefield. This is justified, as the precursors are weak arrivals compared to the SS pulse, often near the amplitude level of the noise in the seismogram. Without weighting, seismograms possessing large degrees of noise in the precursory wavefield dominate over those with very impulsive SS arrivals and relatively high SNR values. Assigning a high weight for earthquakes possessing impulsive SS signals ensures the formation of well-defined precursory arrivals from higher quality seismograms.

We observe only a 1–2 km difference in MTZ thicknesses corrected for different shear wave tomography models and observe a strong correlation between our computed corrections and the precursory travel times. The differential depth between the 410 km and 660 km precursory arrivals (i.e., the MTZ thickness) is largely free from the effects of mantle heterogeneity, and thus largely free of travel time effects, unless strong heterogeneity is isolated within the MTZ, and/or localized just below the 660 km discontinuity (where S660S is isolated from 5410S and SS).

4. Discontinuity topography

Depth deviations of the 410 km and 660 km discontinuities and the thickness of the MTZ beneath the Pacific are shown in Fig. 3 (the corresponding standard deviations are presented in Supplemental Fig. S6). To study frequency dependence of our result, we explore two different filters of the data:ailapass from 15–50 s or "short" period in Fig. 3a, and a bandpass from 25–50 s or "long" period in Fig. 3b. At both frequencies, the majority of the MTZ beneath the Pacific has a thickness within ±5 km of the global average of 242 km (Fig. 3, white region), which is consistent with temperatures at or near the adiabat, and a relatively homogeneous chemical composition beneath these areas. Thus the thickness of 242 km is an appropriate average value for the Pacific Ocean. These "global average" regions predominantly underlie the abyssal plains of the Pacific Ocean, and constitute over 75% of our study area. Only 5% of our study region has a thickened MTZ (blue) and is predominantly associated with subduction processes, similar to observations in past SS precursor studies (Flanagan and Shearer, 1998; Gu and Dziewonski, 2002); the remaining 20% is thinned (warm colors), and is geographically associated with surface regions having hotspot volcanism; other SS precursor studies have also found thinned MTZ beneath hotspots (Courtier et al., 2007b; Deuss, 2007).

4.1. Thickened mantle transition zone

A thickened MTZ beneath subduction having an elevated 410 km and deepened 660 km discontinuity is consistent with cold lithosphere subducting through the isochemical phase changes of olivine. The presence of chemical heterogeneity associated with subduction is not unexpected (Schmerr and Garnero, 2007), and may result in the phase boundaries defocusing from expectations based solely on thermal anomalies, but still result in a thickened MTZ. We therefore focus on MTZ thickness for these regions. At higher frequencies, we observe a thickened MTZ beneath the Cascadian subduction zone (15–20 km), the New Hebrides subduction zone (5–20 km), the New Zealand subduction zone (5–10 km), and the Aleutian subduction zone (5–10 km), though this latter subduction zone lies at the edge of our study region. For comparison with past studies that use data recorded at long period instruments, we note the thickened MTZ beneath the New Zealand and Cascadian subduction zones vanishes in stacks of data that were high-pass filtered at 25 s period (Fig. 3). Stacks of data retaining shorter periods in the SS data thus bring smaller-scale discontinuity features into focus.

The Tonga–Kermadec subduction zone is the most anomalous — it is partially underlain by a thin MTZ (Fig. 3). However, the Tonga–Kermadec and New Hebrides subduction zones have rolled back to the east over the past 50 million years to their present location (Schellart et al., 2006), and appear to have swept into a large province of South Pacific MTZ thinning located to the east. Several cross-sections through this region (Fig. 4) show that the 660 km phase boundary does indeed abruptly deepen at the intersection of the Tonga–Kermadec slab (profiles A–A' and C–C'), consistent with the slab being relatively cold, but interacting with a hotter region. This is further complicated by a bias in the upper mantle travel time corrections for SS from underpredicting the slab's presumed high velocity signature, which is not fully imaged by the tomography model in cross-section C–C' (in Fig. 4).
This would also result in the 660 km boundary being imaged shallower than its actual depth. High velocities associated with the slab are observed in the P-wave tomography models of this region (Li, et al., 2008). If we introduce a high velocity slab with a shear wave velocity anomaly of 3–4% into the upper mantle, we will deepen the 410 km boundary by 6–10 km, and the 660 by 12–15 km, effectively bringing the thinned region of the MTZ in line with the anomaly further east, and thickening the MTZ to the west, beneath subduction. This is demonstrated in the cross-sections A–A′ and C–C′ of Fig. 4 where the blue dotted line shows our computed correction for the missing slab structure when mapped onto discontinuity depth (and MTZ thickness). A similar analysis of the upper mantle corrections for regions of thinned mantle transition zone did not identify the presence of a consistent bias from under-resolved structure (see Supplemental methods).

4.2. Thinned mantle transition zone

A major feature of the MTZ beneath the Pacific is several regions of thinning (Fig. 3). To the southeast and beneath Hawaii, the 410 km discontinuity is deepened to 421 km in a localized down-warping no more than 500–1000 km wide, detectable only at dominant periods ≤15 s, and isolated to only bins immediately to the east (see stacks in Supplemental Fig. S2). The 660 km discontinuity is elevated to 645–650 km over a much broader region, extending 1500–2000 km to the south-southeast of Hawaii. A thinned MTZ is also detected by other SS precursor studies (Flanagan and Shearer, 1998; Gu and Dziewonski, 2002; Schmerr and Garnero, 2006), receiver function analyses (Li, et al., 2000), and ScS reverberations (Courtier et al., 2007a) that image discontinuity depths close to the Big Island of Hawaii. The observed transition zone thinning is consistent with the presence of a warm material rising from the lower mantle beneath Hawaii through the MTZ. Over the past half-decade, several tomography studies have detected a vertically continuous, reduced seismic wave speed anomaly extending from the CMB to the surface beneath the Hawaiian hotspot (Lei and Zhao, 2006; Montelli, et al., 2006). In the cross-sections A–A′ and B–B′ in Fig. 4, lowered shear wave velocities associated with our thinned Hawaiian anomaly are apparent, though there is considerable variation between tomography models on the exact location of the shear wave speed anomaly. However, when wave speed is averaged within a 500 km radius cylinder of material underlying the hotspot, the top and bottom of which are bounded by the 410 and 660 km discontinuities, the MTZ is found to have lowered velocities. A deep 410 km and shallow 660 km discontinuity associated with
lowered MTZ seismic velocities is well explained by a warm plume passing through the MTZ (Putirka, 2005), ultimately reaching the surface as the source for the Hawaiian hotspot. The lateral scale of this anomaly is larger than estimates based on flux considerations, e.g., 100–500 km (Sleep, 2004), which may be due to the sensitivity of SS precursors, which sample structure over a large area and hence laterally smear finer-scale features (Chaljub and Tarantola, 1997; Lawrence and Shearer, 2008). Filtering our data to 25 s dominant period, comparable to long period studies (Flanagan and Shearer, 1998), significantly reduces the amplitude of the imaged Hawaiian and the South Pacific anomalies (Fig. 3), supporting this interpretation. Not all hotspots in our study area are associated with thinned MTZ, suggesting (1) the source of volcanism is not deep-seated (Anderson, 2006), (2) the plume conduit does not pass through the MTZ directly beneath all hotspots, (3) high excess temperature plumes induce decomposition of ringwoodite into majorite (Hirose, 2002; Deuss, et al., 2006) obfuscating the presence of the plume, and/or (4) the associated plumes are too thin to detect with our method, e.g., of order 100 km across or less.

The largest anomaly of this study is the MTZ thinning by 10–25 km in the southern Pacific, in a large province roughly 4000–5000 km long and 2000–3000 km wide, beginning to the south-southwest of the Pacific Superswell and ending near the Darwin Rise (Fig. 3a). The entire province is aligned with the dominant motion of the Pacific plate. The broad scale thinning is largely due to upwarping of the 660 km discontinuity by up to 25 km, excluding the largest anomalies associated with the undercorrection for the subducting Tonga–Kermadec lithosphere. This MTZ anomaly is flanked by a number of volcanic hotspots, including Louisville, Macdonald, Tahiti, and Cook, and underlies the Samoan hotspot. While the small-scale details in tomography models differ between studies, lowered shear wave velocities are robustly found in both the deep mantle and within the MTZ for the South Pacific (Ritsema, et al., 2004), and are correlated with this region of thinned MTZ (cross-section C–C’ in Fig. 4). The hotspots flanking this anomaly are characterized as having deep-rooted origins from seismic, geologic, and geochemical evidence (Courtillot, et al., 2003; Montelli, et al., 2006). This is consistent with
excess temperature responsible for the large-scale transition zone thinning we find in the South Pacific.

5. Sources of discontinuity topography

Scientists have long speculated on the relationship between hotspot volcanism and Earth’s deep mantle. Early work noted that hotspots are situated away from subduction (Morgan, 1971; Hager, et al., 1985) and over large low shear velocity provinces (LLSVPs) in the lowermost mantle beneath the Pacific and southern Africa. More recent work has identified that hotspots are far more likely to overlie the edges of the LLSVPs than the center (Thorne, et al., 2004). This result is consistent with a chemically distinct origin to LLSVPs, whereby the perimeter structure of an LLSVP can serve to guide and root mantle upwellings and plumes (Garnero and McNamara, 2008). The thermochemical LLSVPs, or “piles”, are predicted to be the hottest mantle rock on Earth, and are possibly partially molten near their margins and base (Garnero and McNamara, 2008). The observed MTZ thinning predominantly overlies the LLSVP in the Pacific (Fig. 5). The most anomalous thinning, which occurs beneath Hawaii and the southern portion of the large South Pacific anomaly, is situated above the north and south margin, respectively, of the Pacific LLSVP. This is consistent with hot thermal upwellings and plumes guided by the edges of the chemically distinct LLSVP. Shear wave velocity reductions are more voluminous up into the lower mantle from the core–mantle boundary on the southern part of the LLSVP (Fig. 5), consistent with greater plume upwelling flux there that gives rise to the large 660 km phase boundary upwarping. Our observations connect these two features in the mantle transition zone.

The Pacific Ocean is largely surrounded by subduction, with the notable exception of the Pacific–Antarctic Ridge that defines the southern boundary of the Pacific plate. It is therefore expected that mantle plumes could more easily form within the central and southern Pacific, as predicted in geodynamical calculations that employ Earth’s subduction history as surface boundary conditions (McNamara and Zhong, 2005; Garnero and McNamara, 2008). The bottom of the transition zone marks a mantle viscosity contrast, in which the lower mantle has 1–2 orders of magnitude higher viscosity than the upper mantle (Lithgow-Bertelloni and Gurnis, 1997; Mitrovica and Forte, 1997). This

Fig. 5. The relationship of MTZ thickness to shear wave velocity heterogeneity in the mantle. a) Tomographically derived (Ritsema, et al., 2004) shear velocity perturbations at a various depths within the mantle beneath our study region. The contour interval is 0.5%. b) Cross-sections through the tomographically derived shear velocities in panel a, and MTZ thickness. The locations of the cross-sections are given in the small globe underlying each panel and correspond to those presented in Fig. 4. The relative MTZ thickness beneath our study region is shown on each cross-section (dotted black lines), though the magnitude of the anomaly is greatly exaggerated to allow visualization of the thickness variations.
is expected to cause necking in mantle plumes, leading to significantly
narrowed conduits in the upper mantle (van Keken and Gable, 1995).
Furthermore, the endothermic phase transition associated with the
660 km may act to resist upwelling flow of hot material, perhaps leading
to a small degree of bulging or ponding of plume material beneath the
transition (Davies, 1995).

Plume narrowing due to the viscosity jump and possible accumu-
lization of material due to the endothermic phase change are consistent
with our results for Hawaii and the southern Pacific which display
upwarping on the 660 km discontinuity on a much broader lateral scale
than the depression associated with 410 km. If the large region of
thinned transition zone beneath the southern Pacific is associated with
plume upwelling, it is uncertain whether this is the result of a large,
single plume head (Richards, et al., 1989) or a cluster of smaller plumes
(Kelly and Bercovici, 1997; Schubert, et al., 2004) which could be
responsible for thermally perturbing the entire region. Alternatively,
these results may indicate the presence of thermochemical super-
plumes (Davaille, et al., 2005; Farnetani and Samuel, 2005; Samuel and
Bercovici, 2006; Kumagai, et al., 2007), the tops of which may presently
 reside at or near the 660 km, providing the source region for smaller-
scale thermal plumes that ascend through the 410 km.

To test the hypothesis of a thermal origin for our topographic
variations, we correlate our MTZ thickness measurement with several
tomography models. For a purely thermal origin to seismic velocity
heterogeneity, low shear wave velocities correspond with higher
mantle temperature and conversely high shear wave velocity with
lower mantle temperature. In regions with a vertically continuous
thermal anomaly across the MTZ, the thermally induced changes in
seismic velocities within the MTZ should correlate with the thickness of
the MTZ. Past global studies have found a weak correlation of MTZ
thickness with MTZ shear velocities (Plagannan and Shearer, 1998; Deuss,
2007), though the data coverage in past studies is sparser than in our
study, and the resulting resolution is longer wavelength. We also find
a weak positive correlation of MTZ thickness and velocity perturbations,
suggesting that beneath the Pacific, thermal variations are the primary
mechanism responsible for the topographic variations seen on the MTZ
discontinuities. With the exception of Tonga, subduction zones exhibit
a thickened transition zone, implying colder temperatures in the MTZ.
Hotspots are more complex; if we measure the average MTZ thickness
within 500 km of hotspots that are classi-

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Incorporated Research Institutions for Seismology, the EarthScope
USArray, and the Canadian National Seismograph Network. Data were
collected using the Standing Order for Data software (Owens, et al.,
2004). The TauP toolkit (Crotwell, et al., 1999) and Seismic Analysis
Code were used as part of this work.

Appendix A. Supplementary data

Supplementary data associated with this article can be found in the
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6. Conclusions

These observations provide geographic locations of phase boundary
topography at a variety of scales. Large-scale deflections of the 660 km
discontinuity are stronger and broader than that observed for the
410 km boundary at the same locations, a result consistent with the
presence of a strong viscosity change and phase transition at 660 km
depth. The thinned MTZ regions appear coupled to upwelling plume
processes, and are surrounded by large areas of remarkably flat phase
boundaries that presently do not underlie any active tectonic processes.
While these conclusions do not uniquely mandate a specific morphol-
ogy of plume upwellings, they are well explained by plume structures
with a geographical distribution tied to deep mantle thermochemical
pile locations and a flux that is dependent on variable deep mantle
pressure forces cause by the relative preponderance of subduction in
the northern circum-Pacific.

Similar processes may be related to large igneous provinces in other
locations. For example, the deep mantle African LLSVP is also flanked
by hotspots at Earth’s surface, particularly in the southern hemisphere.
LIPs have been previously related to deep mantle piles (Burke, et al.,
2008); high resolution MTZ imaging offers hope to assess relative
plume flux differences in these regions, especially where traditional
tomographic techniques do not have adequate resolution to directly
image plumes. If the MTZ can be used as a witness of recent past large
eruptive processes, then the opportunity exists for consideration of
correlations of mass extinction events and large flood basalt eruptions
(Bralower, 2008).