A Sporadic Low-Velocity Layer Atop the Western US Mantle Transition Zone and Short-Wavelength Variations in Transition Zone Discontinuities

B. Schmandt

Ken Dueker
University of Wyoming, dueker@uwyo.edu

S. M. Hansen

J. J. Jasbinsek

Z. Zhang

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A sporadic low-velocity layer atop the western U.S. mantle transition zone and short-wavelength variations in transition zone discontinuities

B. Schmandt
Department of Geological Sciences, University of Oregon, Eugene, Oregon 97403, USA
(bschmand@uoregon.edu)

K. G. Dueker and S. M. Hansen
Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming 82071, USA

J. J. Jasbinsek
Physics Department, California Polytechnic State University, San Luis Obispo, California 93407, USA

Z. Zhang
Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming 82071, USA

[1] Teleseismic receiver function analysis of data from six dense arrays in the western U.S. is used to investigate mantle transition zone (MTZ) discontinuities and the prevalence of a low-velocity layer atop the 410 km discontinuity (410-LVL). Negative polarity Ps arrivals indicative of a low-velocity layer with a top 25–60 km above the 410 are identified in 8–11 out of 18 stacks of receiver functions from highly sampled back azimuth corridors. The 410-LVL is interpreted as partial melt resulting from upwelling of hydrated mantle across a water solubility contrast at the 410. The 669 km mean depth of the 660 km discontinuity (660) and the magnitude of 660 topography suggest variable hydration, locally approaching saturation, in addition to >150 K lateral temperature variations beneath five arrays. Mean amplitudes of P410s and P660s increase monotonically with period from 2 to 10 s; however, greater variations are observed in the frequency dependence of P410s compared to P660s implying 410 thickness is more heterogeneous. Variable 410 thickness is attributed to changes in hydration modulating the width of the olivine- to-wadsleyite transition interval. Frequency dependence of P660s amplitudes suggests a broad velocity gradient consistent with multivariate phase changes in the olivine and garnet systems. Sporadic detection of the 410-LVL, the magnitude and length scales of MTZ discontinuity topography, and inferred variations in hydration support the occurrence of vigorous small-scale convection in the western U.S. mantle. Comparison of receiver functions with body wave tomography suggests small-scale convection driven by sinking slab segments and lithospheric instabilities contributes to the intermittent nature of the 410-LVL.

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1. Introduction

Discontinuities in the radial seismic structure of the Earth at nominal depths of 410 km and 660 km are generally attributed to phase transformations in the olivine-dominated mantle assemblage [Ringwood, 1975], and the depth interval bounded by these two velocity discontinuities is referred to as the mantle transition zone (MTZ). In this context, seismic constraints on the depth and sharpness of the discontinuities can be used to estimate thermodynamic variables that control the absolute depth (pressure) and transition interval thickness (pressure range) of the reactions. Seismically detected variations in MTZ discontinuities have most often been interpreted as thermal anomalies [Helffrich, 2000], but in recent years there is increasing evidence that variations in bulk and volatile composition may strongly influence mantle discontinuity structure within and near MTZ depths. Investigations of radial seismic velocity gradients challenge purely thermal interpretations of MTZ topography associated with subducted slabs [Schmerr and Garnero, 2007; Cao and Levander, 2010] and yield contentious results regarding correlations between MTZ discontinuities and hot spots/plumes [Deuss, 2007; Courtier et al., 2007; Tauxin et al., 2008; Das Sharma et al., 2010]. In addition, prior results support the existence of additional smaller velocity discontinuities associated with olivine and non-olivine phase transformations [Shearer, 1990; Revenaugh and Jordan, 1991; Simmons and Gurrula, 2000; Lawrence and Shearer, 2006a], are consistent with hydration of nominally anhydrous minerals at MTZ depths [van der Meijde et al., 2003], and sporadically detect a negative velocity gradient 20–80 km above the MTZ that is frequently attributed to partial melt [Revenaugh and Sipkin, 1994; Song et al., 2004; Jasbinsek and Dueker, 2007; Vinnik and Farra, 2007; Courtier and Revenaugh, 2006, 2007; Jasbinsek et al., 2010; Vinnik et al., 2010; Schaeffer and Bostock, 2010; Tauxin et al., 2010]. Improved resolution of these more complex velocity discontinuities and their lateral variations is potentially diagnostic of fundamental characteristics of mantle dynamics including: prevalence and length scales of mass transfer across the MTZ, mantle hydration levels and abundances of minerals other than olivine, and the physical and chemical processes stimulated by cold slab or hot plume fluxes across the MTZ.

The importance of the MTZ in advancing understanding of mantle dynamics is highlighted by the recently introduced Transition Zone Water Filter (TZWF) model [Bercovici and Karato, 2003], which offers potential to resolve long-standing debate as to how mantle convection supplies geochemically distinct mantle melts to mid-ocean ridges (MORB) and hot spots (OIB), variably permits slab penetration deep into the lower mantle [van der Hilst et al., 1991; Grand et al., 1997] and produces surface heat flow and convective form consistent with internal heat production dominating over basal heating [Tackley, 2002; Zhong, 2006]. A key element of the TZWF model is invoking chemical filtering of the upper mantle via partial melting of sufficiently hydrated wadsleyite upwelling across a water solubility contrast at the 410 as opposed to mechanical decoupling of upper and lower mantle circulation. Diffuse upwelling of hydrated wadsleyite across the 410 into an upper mantle with an approximately order of magnitude lower water solubility [Kohlstedt et al., 1996] is thought to create a layer of hydrous partial melt [Inoue, 1994] that remains perched atop the 410 because its density is intermediate with respect to the MTZ and the overlying upper mantle [Stolper et al., 1981; Ohtani et al., 1995; Sakamaki et al., 2006]. The TZWF model predicts that incompatible elements of the upwelling mantle are preferentially partitioned into the melt to produce a chemically filtered upper mantle consistent with the relatively anhydrous incompatible-element-depleted composition characteristic of MORB. Importantly, this model also predicts that high temperature buoyantly rising plumes will have higher water content and more incompatible-element-enriched composition as a result of temperature dependent water solubility and relatively rapid plume ascent rates minimizing chemical filtering upon upwelling across 410. Thus, plumes could deliver relatively incompatible-element-enriched and hydrated OIB melts to hot spots without physically isolated mantle reservoirs [Bercovici and Karato, 2003]. However, the validity of all assumptions underlying the TZWF is not conclusively established [Karato et al., 2006]. In particular, uncertainties remain regarding the gravitational stability and thermodynamics of a hydrous melt layer atop the 410 [Hirschmann et al., 2006; Sakamaki et al., 2006; Leahy and Bercovici, 2007; Youngs and Bercovici, 2009] and whether the transition zone is sufficiently hydrated for the TZWF to operate globally [Huang et al., 2005; Yoshino et al., 2008; Karato, 2011]. Progress in testing the TZWF hypothesis hinges on achieving improved resolution of the physical and chemical properties of MTZ materials and the processes associated with mass transfer across the MTZ. The TZWF prediction of a melt layer atop the 410 is an ideal target for observational seismology and valuable constraints are...
emerging from regions where sufficient seismic sampling exists.

Seismological inferences of hydrous partial melt atop the 410 based on detection of a low-velocity layer, hereafter referred to as 410-LVL, predate introduction of the TZWF model [Revenaugh and Sipkin, 1994; Vinnik et al., 1996; Bostock, 1998; Vinnik and Farra, 2002], and in recent years efforts to investigate the 410-LVL are accelerating. Detection of 410-LVL has been reported in eastern Asia [Revenaugh and Sipkin, 1994], Siberia [Vinnik and Farra, 2002], Arabia [Vinnik et al., 2003], western U.S. [Song et al., 2004; Fee and Duerer, 2004; Jasbinsek and Duerer, 2007; Vinnik et al., 2010; Jasbinsek et al., 2010], northern Mexico [Gao et al., 2006], eastern U.S. [Courtier and Revenaugh, 2006], eastern Australia [Courtier and Revenaugh, 2007], and northwestern Canada [Bostock, 1998; Schaeffer and Bostock, 2010]. A recent receiver function study of 152 globally distributed stations reports evidence of a 410-LVL beneath 59% of stations and no correlation with the overlying tectonic setting [Taucin et al., 2010]. Estimates of the shear velocity decrease at the top of the 410-LVL are similar in magnitude (3–8%) to the velocity increases across the 410 and 660; however, the depth to the top of the 410-LVL varies dramatically (320–390 km) over short lateral distances, compared to depth variations of the 410 or 660. Large 410-LVL thickness variations likely contribute to lack of detection in global stacks of long-period receiver functions [Lawrence and Shearer, 2006a]. The mounting evidence for widespread presence of a melt layer above the 410 is in accord with TZWF predictions, but many seismic estimates of the layer thickness (20–90 km) greatly exceed the 1–24 km steady state thickness that Bercovici and Karato [2003] first derived based on a range of MTZ water concentrations and a background upwelling rate of 1 mm/yr. Uncertainties in upwelling rates, hydration levels, melt fraction, melt density (hence buoyancy), and permeability of the partially molten layer can make a thicker melt layer permissible, but in general geochemical and geodynamic calculations favor a melt layer that is thinner than many seismic estimates of 410-LVL thickness [Karato et al., 2006].

Thickening of the hypothesized melt layer atop the 410 beyond its density crossover [Youngs and Bercovici, 2009] or convectively perturbing the melt layer by subduction-induced return flow [Faccenna et al., 2010] could cause upward instabilities (rising diapirs) from the hydrous melt layer and in some circumstances account for seismic detection of a heterogeneous and surprisingly shallow top of the 410-LVL. Vinnik and Farra [2007] suggest a correlation between the presence of 410-LVL and upwelling plumes, and Jasbinsek et al. [2010] similarly suggest that upward instabilities of the 410 melt layer may be the origin of steeply dipping low-velocity “pipes” tomographically imaged within the western U.S. mantle [Sine et al., 2008]. The lack of seismic results indentifying a thin 410-LVL (<15 km) may reflect that such a thin layer is absent or rare or it may be a result of inadequate vertical resolution owing to interference from signals associated with the adjacent 410 and the fact that little teleseismic energy is observed at frequencies >1 Hz. Thus, from existing seismological constraints the 410-LVL is both a common and heterogeneous feature and whether it is a manifestation of a hydrous melt layer created by the TZWF stands as a plausible and provocative hypothesis for further testing.

Regardless of the veracity of the TZWF model, it is clear that a more complex role for the MTZ in thermochemical mantle convection is emerging and that higher resolution seismic mapping of lateral variations in mantle layering will provide valuable constraints on the thermal and chemical processes active in the MTZ. In this paper, we capitalize on several (sub)arrays that provide exceptionally high-fold sampling of selected portions of the western U.S. upper mantle to constrain the depth and frequency-dependent character of P-to-S (Ps) scattering from upper mantle discontinuities, determine whether MTZ topography is common over short wavelengths, and further constrain 410-LVL prevalence and its correlation with 3-D mantle velocity structure [e.g., Schmandt and Humphreys, 2010].

2. Data and Methods

2.1. Data

From the extensive archives of broadband data recorded by temporary and permanent stations in the western U.S. six small aperture (<200 km) arrays were selected that provide exceptionally dense sampling as a result of spatially dense deployments and/or long recording duration (Figure 1 and Table 1). The Wallowa, Mendocino, Sierra Nevada, and Tucson arrays are all composed of a combination of temporary and permanent stations, not all of which operated contemporaneously. We use the 12 westernmost stations of the temporary RISTRAS1.5 array combined with three nearby USAArray stations and herein refer to the
entire group as the RISTRA1.5 array. Our “Anza array” is a spatially compact eight-station subset of the long running Anza network located in southern California. These six arrays were primarily selected for their data density within an aperture similar to first Fresnel zone width of 2–10 s period body waves in the MTZ. Secondary consideration was given to achieving broad sampling of the western U.S. Regardless of whether all the stations comprising each array operated contemporaneously the high-fold sampling of the teleseismic wavefield is of great value for locally constraining upper mantle stratification.

We focus our analysis on three narrow back azimuth corridors which contain the majority of teleseismic events observed in the western U.S. with Mb > 5.6 and epicentral distance from 35 to 95 degrees. All arrays have sources distributed across most of the 35–95 degree distance range in the southeast and northwest corridors, however in the southwest corridor sources are only available from approximately 75–95 degrees (Figure 2).

The seismogram segments are centered on the P wave arrival and are 400 s in duration with a 20 Hz sampling frequency. After removing the mean and trend, the traces are padded with zeros, tapered, and filtered from 0.01 to 1.5 Hz. The three-component seismograms are then rotated into the P, SV, and SH coordinate system [Vinnik, 1977]. Traces with high signal-to-noise on both the P and SV components are visually selected.

### Table 1. Array Characteristics

<table>
<thead>
<tr>
<th>Array</th>
<th>Stations</th>
<th>Station Years</th>
<th>Max Aperture (km)</th>
<th>Min Aperture (km)</th>
<th>Dates</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wallowa</td>
<td>19</td>
<td>37</td>
<td>125</td>
<td>115</td>
<td>2006–2009</td>
</tr>
<tr>
<td>Mendocino</td>
<td>18</td>
<td>34</td>
<td>135</td>
<td>105</td>
<td>2005–2009</td>
</tr>
<tr>
<td>Sierra Nevada</td>
<td>30</td>
<td>29</td>
<td>155</td>
<td>75</td>
<td>2005–2008</td>
</tr>
<tr>
<td>Anza</td>
<td>8</td>
<td>98</td>
<td>50</td>
<td>40</td>
<td>1997–2010</td>
</tr>
<tr>
<td>Tucson</td>
<td>7</td>
<td>26</td>
<td>115</td>
<td>80</td>
<td>1997–2010</td>
</tr>
<tr>
<td>Ristra1.5</td>
<td>15</td>
<td>27</td>
<td>220</td>
<td>30</td>
<td>2004–2009</td>
</tr>
</tbody>
</table>
sampling of radial velocity gradients in three “corridors” (Figure 1). For each earthquake, a single source wavelet is assumed to be incident at each array. The multichannel deconvolution method used assumes the intramodal scattering (e.g., P-to-P) impulse response is near minimum phase [Bostock, 2004; Baig et al., 2005; Mercier et al., 2006; Hansen and Dueker, 2009] to estimate the three-component impulse response due to intermodal scattering from horizontal interfaces beneath each station. In brief, the method forms an initial estimate of the source wavelet amplitude spectra by stacking the P component spectra and then fitting a smooth spline to the stacked spectra. This procedure assumes that the source wavelet spectra are smooth with respect to the receiver function spectra. The initial source spectra are used as constraint equations in a least squares inversion to form a separation of the source spectra from the three-component receiver function spectra. Using the assumption that the P-to-P scattering response is minimum phase (an implicit assumption in all receiver function deconvolution methods), the Kolmogorov operator and algebraic rearrangement of the phase equations shows that the source phase and the phase of the transverse receiver function can be reconstructed [Bostock, 2004].

This blind deconvolution method is particularly well suited for dense small-aperture arrays due to the highly coherent incident teleseismic source wavelets. To take advantage of the high-fold sampling provided by redundant source-receiver geometries, the sources for each station are binned by ray parameter (0.0035 s/km increments) and back azimuth (10° increments) and a single impulse response is found for all the sources in each bin. Small variations in the bin dimensions are found to recover consistent results. The mean number of events per bin varies from 1.5–4.4 for all 18 corridor stacks from the six arrays, and there are about 20–35 bins per back azimuth corridor.

Invoking the minimum phase character of the P-to-P impulse response [Bostock, 2004], rather than using the vertical or P component as a proxy for the source wavelet, is theoretically more sound as it honors the potential contribution of intramodal P scattering and the effectiveness of the method with observational data is demonstrated by successful recovery of intramodal P scattering from the Moho beneath stations in Canada [Mercier et al., 2006]. Hansen and Dueker [2009] previously applied our implementation of this multichannel spectral deconvolution method and readers are referred to that text for further details.

Broadband estimates of the impulse response functions, hereafter referred to as receiver functions, are filtered into five frequency bands to reveal the consistency and frequency-dependent character of scattered arrivals. A common long period corner of 70 s is used for all bands and the short period corners are 2 s, 3 s, 4.5 s, 7 s, and 10 s. While P, SV, and SH component receiver functions were determined, this paper focuses on the SV receiver functions as the other components show little evidence for coherent scattering from MTZ structure that is consistent between multiple arrays.

2.3. Depth Domain Corridor Stacks and Culling

Receiver functions times are mapped to depth and stacked for the southeast, southwest, and northwest back azimuth corridors for each of the six arrays. To obtain more accurate absolute depth estimates and mitigate stack incoherence caused by 3-D velocity heterogeneity the P and S raypaths for the receiver functions are traced through the high-resolution P and S tomography models of Schmandt and Humphreys [2010], which incorporate traveltime data from all the arrays used in our receiver function study. Typical direct S teleseismic traveltime residuals within these small-aperture arrays have root mean square (RMS) values of 0.4–0.8 s, so this correction is most important for shorter period stacks (2–3 s) and has little effect on the longest period (10 s) stacks. After correction for 3-D velocity heterogeneity the RMS of S traveltime residuals within the six arrays varies from 0.11–0.23 s, which is less than one eighth of the shortest period in our receiver function stacks. Two steps of culling are applied to the depth domain receiver functions. First, receiver functions are culled by visually rejecting outliers in the distribution of SV receiver function RMS in each corridor, which eliminates 7–15% of the total number of traces. This is accomplished by visually adjusting a nominal threshold of rejecting the 10% of traces with the highest RMS. A slightly greater fraction of traces are rejected if the RMS distribution for a corridor has a “fat tail” and slightly fewer traces are rejected if the distribution is more compact. In the second step of culling, each SV depth domain receiver function is cross correlated with its corridor stack and traces with correlation coefficients more than two standard deviations below the mean correlation coefficient are rejected. This step removes another 6–13% of the original number of traces. The standard deviations of the culled receiver function stacks are calculated via boot-
Figure 3
strapping with each stack resampled 100 times [Efron and Tibshirani, 1986].

3. Results

3.1. Receiver Function Stacks

[15] Identification of robust Ps arrivals is accomplished by considering several factors: 1) amplitude of an arrival relative to two standard deviations of the stack, 2) consistency across multiple frequency bands, 3) Ps amplitude relative to sidelobe estimates from P receiver function stacks when an opposite polarity Ps arrival is proximal; sidelobe amplitude estimates are 17–25% of main peak amplitude for all stacks, and 4) moveout analysis of Ps timing as a function of ray parameter. Moveout analysis for southwest corridor stacks is poorly constrained because sources are only observed over a small distance range at far offsets (Figure 2).

3.1.1. Wallowa Array

[16] The Wallowa array receiver function stacks and phasing analysis all show clear P_{410}s and P_{660}s arrivals (Figure 3). The southeast corridor shows a negative polarity arrival at ~375 km depth with absolute amplitude similar to or greater than P_{410}s. We consider this arrival as evidence for a 410-LVL and hereafter label it P_{Ws} [Bostock, 1998]. Previous research regarding 410-LVL in this region found evidence for a regional-scale 410-LVL from modeling of triplicated waveforms and receiver functions [Song et al., 2004], but recent receiver function results from a dense broadband array found only isolated evidence for 410-LVL beneath the northwest U.S. [Eagar et al., 2010].

3.1.2. Mendocino Array

[17] Corridor stacks from the Mendocino array have large two standard deviation envelopes due to higher noise levels and fewer useable events than our other arrays (Figure 4). The higher noise level is consistent with the proximity of the array to oceanic noise sources [McNamara and Buland, 2004]. We rely on consistency across multiple frequency bands and moveout analysis to identify robust Ps phases. P_{660}s is observed in all three corridors, but surprisingly a P_{410}s arrival is not observed in the northwest corridor. A robust P_{Ws} arrival is detected at 390 km depth in the southwest corridor and a possible P_{Ws} arrival is found at 350 km depth in the southeast corridor. Vinnik et al. [2010] report indications of 410-LVL in S receiver functions from stations just south of the Mendocino array.

3.1.3. Sierra Nevada Array

[18] The Sierra Nevada array receiver function stacks show clear P_{410}s and P_{660}s arrivals in all corridors (Figure 5). A robust P_{Ws} arrival is only detected in the southwest corridor at 370 km depth. Previous S receiver function results from four long-recording stations near the Sierra Nevada find 410-LVL arrivals in half of their corridor stacks [Vinnik et al., 2010]. Also, the southwest corridor stacks find a negative arrival at about 570 km depth.

3.1.4. Anza Array

[19] Considering that the Anza array has the highest data fold (Table 1), the stacks are surprisingly energetic throughout the 300–750 km depth range (Figure 6). P_{410}s and P_{660}s are found in all corridors, except P_{410}s in the lowest frequency stack from the southeast corridor. A P_{Ws} arrival is observed at 380–400 km in the southwest corridor, and variations of the P_{Ws} arrival with frequency are likely a result of interference with the P_{410}s side-lobe. The southeast stacks are particularly energetic at 350–500 km depth hence we refrain from interpreting the negative pulse above P_{410}s as robust evidence of 410-LVL. Negative pulses above P_{410}s are observed in the northwest corridor.
Figure 4
however this observation is also ambiguous due to substantial variations in pulse characteristics across the frequency bands. Additionally, a robust negative polarity arrival with P58s moveout is found near 530 km depth in the northwest corridor.

3.1.5. Tucson Array

[20] All corridor stacks from the Tucson array show clear P410s and P660s arrivals except in the two lowest frequency stacks for the southeast corridor that show negligible amplitude P410s arrivals (Figure 7). Tucson is the only array with robust P58s arrivals in all three corridors with a remarkable 35–40 km depth variation between corridors. Additionally, a negative arrival with P58s moveout is observed at 500 km depth in the southwest corridor.

3.1.6. RISTRA1.5 Array

[21] P410s and P660s are found in all corridors from the RISTRA1.5 array with the exceptions of a weak P410s in the two shortest period southeast corridor stacks and a weak P660s in the two shortest period southwest corridor stacks (Figure 8). P58s is only observed in the southeast corridor and this detection is spatially continuous with previous detections of 410-LVL using the adjacent Ristra-1 array [Jasbinsek et al., 2010]. A robust negative polarity arrival is observed at 570 km depth in the southeast corridor, and another more ambiguous negative arrival is found at ~600–620 km in the southwest corridor.

3.1.7. S_H Receiver Functions

[22] The S_H receiver function stacks generally show lower amplitudes and greater variation than the S_V receiver functions (Figure 9). A straightforward explanation for the relative lack of coherent arrivals in the S_H receiver functions is that most instrument orientations and our three component rotations are accurate, and that coherent dipping and anisotropic discontinuities are not consistently present at MTZ depths beneath our arrays. That there are arrivals with amplitude greater than two standard deviations in many of the S_H receiver function stacks suggests dipping and anisotropic discontinuities could be present, but they must vary laterally to explain the lack of consistency between S_H receiver function stacks from different arrays.

3.2. MTZ Thickness and Topography

[23] Estimates of MTZ thickness have two components of error. The first is the estimation of the arrival time of the peak amplitude of P410s and P660s phases. This error is estimated via bootstrap resampling of the arrival times and when converted to depth the standard error is <1 km. The second source of error is inaccuracy of the 3-D P and S tomography models used to map receiver function time to depth. Given that all of our arrays are within USArray (70 km nominal station spacing), the tomography models [Schmandt and Humphreys, 2010] also use data from all the non-USArray stations in our study, and MTZ thickness estimates only depend on the path-integrated MTZ velocity variations this source of error is considered negligible. The mean MTZ thickness calculated from all corridor stacks is 249 km. The maximum variation in MTZ thickness across all arrays is 43 km and the maximum variation within one array is 34 km for the RISTRA1.5 array (Figure 10).

[24] The mean depth of the 410 and 660 are 420 km and 669 km and maximal variations in the 410 and 660 depths are 23 km and 27 km, respectively. These discontinuity depths are deeper than global values [Kennett et al., 1995; Lawrence and Shearer, 2006a], but similar to a recent USArray receiver function study that found mean depths of 415–420 km and 665–673 km depending on the velocity model used to map time to depth [Cao and Levander, 2010].

3.3. Frequency Dependence of P410s and P660s Amplitude

[25] The mean stack amplitudes of P410s and P660s increase monotonically with increasing short
Figure 5
period filter corner from 2 s to 10 s (Figure 11). This observation is consistent with Ps conversions from a finite-width velocity gradient rather than a step function increase in velocity with depth. Greater scatter is observed in the frequency dependence of P410s (Figure 12) compared to P660s (Figure 13) suggesting greater heterogeneity in the sharpness of the 410. Interference between PWS and P410S may contribute to some of the observed scatter in the frequency dependence of P410s amplitude. Interestingly, Singh and Kumar [2009] studied frequency dependence of P410s and P660s amplitudes beneath the eastern Himalaya and southern Tibet and found very different results: they found greater P410s amplitude compared to P660s and stronger frequency dependence for P410s amplitude than for P660s amplitude.

To obtain simple constraints on 410 and 660 velocity gradient thickness, the rule of thumb that significant P-to-S converted energy will be observed if the gradient thickness is less than or equal to half of the incident P wavelength is used [Bostock, 1999]. For the five frequency bands presented here the incident P wavelength at MTZ depths varies with increasing period from 20 to 100 km. Thus, the slope of mean P660s amplitude increasing with period implies a velocity gradient thickness of at least 50 km. Unlike P660s, the slope of mean P410s amplitude with period decreases beyond 4.5 s suggesting gradient thickness of about 25 km. It is not clear whether P410s amplitude has reached a plateau at 10 s period, hence it is possible that gradient thickness could be >25 km. However, this is a simplified view of discontinuity structure and it is important to consider that the gradient need not be uniform over its entire thickness. For example, observations of significant Ps conversions for short periods could be satisfied by a combination of a wide gradient with an embedded sharper gradient [e.g., Melbourne and Helmberger, 1998]. Detailed waveform modeling would be required to determine the fine-scale structure of the 410 and 660 velocity gradients.

### 3.4. Scattering From the 410-LVL

Out of 18 corridor stacks eight (44%) show robust PWS arrivals indicative of a 410-LVL, seven stacks find a robust absence of PWS, and three stacks are ambiguous (Figure 14). These statistics are in contrast to similarly high-fold receiver function studies in the western U.S. interior, which detected PWS in 88% of corridor stacks beneath the northern Rocky Mountains [Jasbinsek and Dueker, 2007] and 77% beneath the southern Colorado Plateau and Rio Grande Rift [Jasbinsek et al., 2010]. By including the three ambiguous stacks the detection level rises to 61%, which remains lower than the two Cordilleran interior 410-LVL occurrence statistics. Prior receiver function results from regions closer to the west coast are in better agreement with our 44–61% detection of PWS.

Vinnik et al. [2010] stacked S wave receiver functions from well sampled back azimuth corridors at single stations and clusters of 2–4 stations in California and found 410-LVL arrivals in 51% of their stacks. In the northwest U.S., Eagar et al. [2010] report only isolated detections of negative polarity arrivals just above P410s. Thus, from the existing distribution of western U.S. receiver function studies the 410-LVL appears to be less prevalent near the western plate margin compared to more interior regions (Figure 15), but it is certainly possible this trend is an artifact of small sample size.

The mean amplitude of PWS is slightly less than P410s (Figure 11) and the depth of PWS varies from 355 to 400 km. The depth of PWS arrivals in the corridor stacks from the Tucson array, which is the only array with robust PWS detections in multiple corridors, has a maximum variation of 35–40 km. Such large changes in depth over short lateral distances could result in bias toward nondetection of PWS in studies stacking receiver functions from all back azimuths. The mean depth of our PWS arrivals is 384 km, which is similar to PWS detections from other

**Figure 5.** S wave receiver functions from the Sierra Nevada array. (a) Cumulative stacks of RFs from all three corridors with five short corner periods (2, 3, 4.5, 7, and 10 s) increasing from left to right. Bootstrap-derived two standard deviation envelopes are plotted (black dashed line) around each trace and the number of traces in the stack is indicated at the top of the plot. Black horizontal lines demark the mean depths of P410s and P660s for all frequency bands in corridors with robust arrivals. (b–d) Corridor stacks from the southeast, southwest, and northwest. (e–h) Moveout analysis for the indicated corridor or cumulative stack (Figure 5e). All five frequency bands are averaged for moveout analysis, effectively favoring the lower frequency data. The amplitudes of P38 conversions from a horizontal interface at depth should be maximized where the depth on the y axis is equal to the phasing depth on the x axis. Amplitude maxima near the 410, 660, and top of the 410-LVL are indicated (pluses).
Figure 6
P receiver function studies in the western U.S. [Jasbinsek and Dueker, 2007; Jasbinsek et al., 2010; Eagar et al., 2010]. However, S receiver functions from California find the top of the 410-LVL at a shallower mean depth of 350 km [Vinnik et al., 2010].

[29] Following Schaeffer and Bostock [2010], receiver functions from corridors with PpWs arrivals were stacked for reverberations from the 410 (Pp410s) and the top of the 410-LVL (PpWs) in hope of yielding constraints on Vp/Vs within the layer. In general, coherent energy with the appropriate moveout for PpWs was not observed, with the exception of a probable PpWs arrival from the southeast corridor of the RISTRA1.5 array. Further analysis of PpWs and Pp410s arrivals from the RISTRA1/1.5 arrays will be the subject of future research.

4. Discussion

4.1. MTZ Temperature, Hydration, and Discontinuity Topography

[30] Experimental mineral physics estimates of Clapeyron slopes for the olivine-to-wadsleyite and postspinel phase transformations are used to constrain the range of lateral variations in mantle temperature and hydration consistent with observed 410 and 660 topography, respectively. Anhydrous in situ X-ray diffraction experiments estimate Clapeyron slopes of 3.6 to 4.0 MPa/K for olivine-to-wadsleyite [Morishima et al., 1994; Katsura et al., 2004] and −0.4 to −2 MPa/K for postspinel [Katsura et al., 2003; Fei et al., 2004; Litasov et al., 2005]. Based on multiple internal pressure standards, Fei et al. [2004] suggest −1.3 MPa/K as a best estimate of the postspinel Clapeyron slope. The addition of two weight percent H2O was found to increase the magnitude of the postspinel Clapeyron slope from −0.5 MPa/K to −2.0 MPa/K [Litasov et al., 2005]. Aside from effects on the Clapeyron slope, near saturated water content (2–3 wt%) is suggested to decrease the depth of olivine-to-wadsleyite by approximately 30 km [Smyth and Frost, 2002] and increase the depth of the postspinel transition by approximately 15 km [Boffan-Casanova et al., 2003; Litasov et al., 2005].

[31] Using anhydrous Clapeyron slopes of 3.6 to 4.0 MPa/K for olivine-to-wadsleyite and −1.3 MPa/K for postspinel, the 23 km and 27 km maximum variations of P410s and P660s depth correspond to lateral temperature variations of 210–230 K and 815 K, respectively. Noteworthy is that the lower bounds of postspinel Clapeyron slope from two studies [Katsura et al., 2003; Litasov et al., 2005] of −0.4 to −0.5 MPa/K would correspond to unrealistically large temperature variations of >2000 K. If experimental estimates finding such low magnitude postspinel Clapeyron slopes are correct [Katsura et al., 2003; Litasov et al., 2005], nonthermal influences would be required to explain even modest 660 topography; e.g., 5 km of relief would correspond to a 400–500 K temperature variation.

[32] The magnitude of 410 topography observed can be reasonably accounted for via anhydrous thermal variations or variations in mantle hydration, or both. In contrast, the estimate of 815 K temperature variation associated with 660 topography is large even in the presence of subducted slabs. Furthermore, using the anhydrous postspinel Clapeyron slope of −1.3 MPa/K [Fei et al., 2004], five of six arrays (excluding Sierra Nevada) would have intercorridor temperature variations of at least 225 K. Hence, the magnitude of 660 topography and mean depth of 669 km lead us to suggest mantle hydration has as an important influence via increasing the magnitude of the postspinel Clapeyron slope (thereby decreasing the implied thermal variations) and increasing the depth of the 660.

[33] The RISTRA1.5 array provides the strongest evidence for hydration contributing to 660 topography. Receiver functions from the array sample across the edge of a tomographically imaged slab crossing the 660 (Figure 16), and the southwest corridor stack has the deepest 660 (687 km) and greatest transition zone thickness (274 km) in this

Figure 6. Sv receiver functions from the Anza array. (a) Cumulative stacks of RFs from all three corridors with five short corner periods (2, 3, 4.5, 7, and 10 s) increasing from left to right. Bootstrap-derived two standard deviation envelopes are plotted (black dashed line) around each trace and the number of traces in the stack is indicated at the top of the plot. Black horizontal lines demark the mean depths of P410s and P660s for all frequency bands in corridors with robust arrivals. (b–d) Corridor stacks from the southeast, southwest, and northwest. (e–h) Moveout analysis for the indicated corridor or cumulative stack (Figure 6e). All five frequency bands are averaged for moveout analysis, effectively favoring the lower frequency data. The amplitudes of PpWs conversions from a horizontal interface at depth should be maximized where the depth on the y axis is equal to the phasing depth on the x axis. Amplitude maxima near the 410, 660, and top of the 410-LVL are indicated (pluses).

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Figure 7
study. The portion of the Farallon slab now beneath the eastern Great Basin is thought to have subducted prior to 30 Ma and was <30 Myr old when it reached the trench [Lithgow-Bertelloni and Richards, 1998; Engebretsen et al., 1985]. Conductive heating calculations suggest the temperature contrast between the slab and adiabatic mantle is <375 K [Turcotte and Schubert, 2002]. Assuming water concentration is near saturated locally the depression of the 660 can be explained reasonably with ~300 K cooler temperature. Thus, this is our preferred interpretation, in agreement with Cao and Levander [2010]. The amount of water that subducted slabs can carry into the MTZ is still debated [Schmidt and Poli, 1998; Rüptke et al., 2004; Komabayashi et al., 2004; Ohtani et al., 2004; Hacker, 2008; Green et al., 2010]. If the water content in the slab is not sufficient to alter the absolute depth of the 660 and postspinel Clapeyron slope as outlined above, then a different and currently unknown compositional contribution to 660 topography would be required to explain the observed magnitude of 660 depression.

[34] The thinnest MTZ is observed in the southeast corridor stacks from the Mendocino array with $P_{410}$ at 430 km and $P_{660}$ at 660 km, which is about a 10 km deviation from the mean depths of both discontinuities. This observation of relatively thin MTZ correlates with a strong low-velocity anomaly that extends through the 410 but not the 660 and is adjacent to the bottom and southern edge of the subducting Gorda slab as imaged by body wave tomography (Figure 15) [Sigloch et al., 2008; Burdick et al., 2009; Roth et al., 2008; Schmandt and Humphreys, 2010; Obrebski et al., 2010; Tian et al., 2011]. Hence, we favor relatively high temperatures (150–200 K) coincident with the low-velocity anomaly intersecting the 410 and dry mantle near the 660 to explain these results.

[35] We are unable to uniquely determine the relative contributions of temperature and water content to our MTZ topography results. However, the large magnitude of temperature variations implied by the anhydrous postspinel Clapeyron slope and mean depth of 669 km make a strong case that contributions from both temperature and water are necessary. Furthermore, inference of high water content in the western U.S. transition zone is plausible given the 150 Myr history of continuous subduction beneath the region [Engebretsen et al., 1985] and that diffusion effectively transfers water from sinking slabs to the surrounding MTZ [Richard et al., 2006].

4.2. A Broad 660 Gradient and Mantle Phase Transformations

[36] The remarkably uniform frequency dependence of $P_{660}$ amplitude for all corridor stacks (Figure 13) is in accord with a number of studies suggesting that phase transformations at similar pressure and temperature conditions in the olivine and garnet systems both contribute to the velocity increase with depth near 660 km [Weidner and Wang, 1998; Simmons and Gurrola, 2000; Hirose, 2002; Deuss et al., 2006; Lawrence and Shearer, 2006a; Schmerr and Garnero, 2007]. The transition of majorite garnet to perovskite occurs over a broader pressure interval (1–1.5 GPa) than the postspinel transition (~0.5 GPa), and depending on local temperature and Al concentration it may begin tens of kilometers shallower or deeper than the postspinel transition and ilmenite may become stable as an intermediate phase [Weidner and Wang, 1998; Hirose, 2002]. Regional variability in the character of PP and SS precursors from depths near 660 km provides global evidence for such mineralogic complexity near the base of the upper mantle [Deuss et al., 2006]. The combination of a relatively sharp postspinel phase transition and a variably overlapping or adjacent broader phase transition in the garnet system is consistent with our $P_{660}$ results suggesting total velocity gradient thickness of at least 50 km. A recent P wave triplication study suggests similar thickness of the 660 (50–70 km) beneath northeastern China [Wang and Niu, 2010]. Some seismic studies report detections of multiple sharp discontinuities near 660 km depth [Niu and Kawakatsu, 1996; Simmons and Gurrola, 2000; Ai et al., 2003] as opposed to a broad gradient. We find only isolated evidence for multiple sharp discontinuities in the form of distinct positive polarity

Figure 7. Sv receiver functions from the Tucson array. (a) Cumulative stacks of RF’s from all three corridors with five short corner periods (2, 3, 4.5, 7, and 10 s) increasing from left to right. Bootstrap-derived two standard deviation envelopes are plotted (black dashed line) around each trace and the number of traces in the stack is indicated at the top of the plot. Black horizontal lines demark the mean depths of $P_{410}$ and $P_{660}$ for all frequency bands in corridors with robust arrivals. (b–d) Corridor stacks from the southeast, southwest, and northwest. (e–h) Moveout analysis for the indicated corridor or cumulative stack (Figure 7e). All five frequency bands are averaged for moveout analysis, effectively favoring the lower frequency data. The amplitudes of $P_{58}$ conversions from a horizontal interface at depth should be maximized where the depth on the y axis is equal to the phasing depth on the x axis. Amplitude maxima near the 410, 660, and top of the 410-LVL are indicated (pluses).
Figure 8
arrivals in our shorter period stacks. For example, multiple peaks in $P_{660}$ are observed in the southwest and, to a lesser extent, the northwest corridor stacks from the Anza array, which was included by Simmons and Gurrola (2000) (Figure 6). The northwest corridor from the Tucson array also shows multiple peaks near $P_{660}$ in the shorter period stacks (Figure 7). The facts that $P_{660}$ is observed in shorter period stacks (2–3 s) and that its amplitude strongly increases with period suggest the 660 may be composed of a relatively sharp gradient (~10–20 km thick) embedded within a broader gradient (~50 km thick).

4.3. Thickness and Heterogeneity of the 410

The anhydrous olivine-to-wadsleyite transition occurs over a 7–13 km thick interval [Katsura et al., 2004], but hydration may broaden the transition interval to as much as 30–40 km [Wood, 1995; Smyth and Frost, 2002; Hirschmann et al., 2005; Litasov et al., 2006]. The observation that our mean $P_{410}$ amplitude increases with period and that the increase in amplitude lessens for the two longest periods suggests gradient thickness of at least 25 km. Thus, $P_{410}$ amplitudes are inconsistent with anhydrous conditions and potentially explained by 500–1200 ppm H$_2$O in the deep upper mantle, depending on the partition coefficient of H$_2$O between olivine and wadsleyite [Wood, 1995; Smyth and Frost, 2002; Hirschmann et al., 2005]. In contrast to the remarkably consistent frequency dependence of $P_{660}$ amplitude, individual stacks show greater heterogeneity in the frequency dependence of $P_{410}$ amplitude particularly at the longer periods (Figure 12). If thickness of the 410 is controlled by H$_2$O concentration, this suggests lateral variations in hydration over length scales of a few hundred km are prevalent in the western U.S. mantle.

4.4. Implications for the TZWF Hypothesis

Unambiguous detection of $P_{660}$ arrivals that require a 410–LVL, and evidence that the thickness of the 410 and the magnitude of 660 topography are consistent with hydration of the MTZ all provide support for the TZWF hypothesis. Our results also present a view of the 410–LVL as a strongly heterogeneous region owing to positive and negative detections and up to 35–40 km of change in $P_{660}$ depth between different corridor stacks from individual arrays.

We suggest the global prevalence of 410–LVL may be even greater than observed by Tauzin et al. [2010] if the short-wavelength 410–LVL heterogeneity found beneath the western U.S. is common in other regions. In addition, the remarkable heterogeneity of the 410–LVL exhibited by the global survey of Tauzin et al. [2010] and at smaller scales in this study likely explains why the widespread presence of this feature was only recently recognized. Assuming the 410–LVL is a manifestation of melt produced by upwelling of hydrated mantle across the 410, the melt is expected to spread laterally

Figure 9. $S_H$ receiver function stacks. The $S_H$ component receiver function stacks are labeled by array, filtered with a short corner period of 4.5 s, and arranged with southeast, southwest, and northwest corridor from left to right.
Leahy and Bercovici, 2007]. Thus, robust positive and negative detections separated by ∼300 km suggest recent/ongoing disruption of the melt layer by convective flow: e.g., downward entrainment into the MTZ [e.g., Leahy and Bercovici, 2007] and/or upward diapiric instabilities of the melt layer. Upward instabilities could be stimulated by return flow in response to segmented subduction [e.g., Faccenna et al., 2010] or thickening of the melt layer to its density crossover [Youngs and Bercovici, 2009].

[40] Vigorous small-scale convection in the western U.S. upper mantle is supported by short-wavelength topography of MTZ discontinuities found herein and by previous studies [Gilbert et al., 2003; Cao...
[41] While it is difficult to account for widespread existence of the 410-LVL with processes other than partial melting [e.g., Revenaugh and Sipkin, 1994; Bercovici and Karato, 2003; Jasbinsek and Dueker, 2007; Tazuin et al., 2010], low seismic velocity is still indirect evidence for the melt layer predicted by the TZWF. Therefore, the recent finding from P wave receiver function reverberation analysis that the 410-LVL beneath northwestern Canada has anomalously high Poisson’s ratio (~0.4) provides an important further constraint that partial melt is the physical origin of the layer [Schaeffer and Bostock, 2010]. Additionally, magnetotelluric constraints on mantle conductivity beneath the Tucson area are found to be consistent with a 5–30 km thick layer of hydrous melt atop the 410 [Toffelmier and Tyburczy, 2007], which is intriguing considering that out of the six arrays in our study only the Tucson array detected the 410-LVL in all three corridor stacks. However, Toffelmier and Tyburczy [2007] did not find electromagnetic evidence for melt in other regions where seismic studies have subsequently detected 410-LVL including California, Hawaii, and western Europe [Vinnik et al., 2010; Tazuin et al., 2010].

[42] Alternative explanations for 410-LVL observations (aside from the TZWF) include ponding of hydrous melt diapirs from the 410-LVL [e.g., Youngs and Bercovici, 2009] and potentially drive episodes of intraplate volcanism [Iwamori, 1992; Vinnik and Farra, 2007; Richard and Iwamori, 2010; Jasbinsek et al., 2010].

Figure 11. Mean amplitudes of P410s, P660s, and PWS arrivals. The mean amplitudes of P410s (green) and P660s (red) arrivals for all 18 corridor stacks in each frequency band are plotted. The mean absolute value of PWS arrivals is derived from only the eight corridors with robust PWS detections (see Figure 14).

Figure 12. Frequency dependence of P410s amplitude. The amplitude of P410s in the cumulative (black), southeast (red), southwest (green), and northwest (blue) corridor stacks is plotted for each array. The mean of all corridors (excluding cumulative array stacks) is included for reference.
volatile rich melts that were created above the 410 but below the melt density crossover and then sank to a gravitationally stable position atop the 410 \cite{Revenaugh and Sipkin, 1994; Courtier and Revenaugh, 2007; Dasgupta and Hirschmann, 2006}. However, this alternative may require deep upper mantle melts that are unreasonably dense \cite{Courtier and Revenaugh, 2007}. The possibility of a subsolidus compositional origin for the 410-LVL, such as subducted ocean crust, is unlikely. It would be fortuitous for such chemical heterogeneity to be distributed in a subhorizontal layer atop the 410 beneath 59% of global stations \cite{Tauzin et al., 2010}, subducted ocean crust is expected to sink to 600–800 km depth before becoming neutrally buoyant \cite{Anderson, 1979; Christensen and Hofmann, 1994; Lee and Chen, 2007}, and we are not aware of an abundant solid composition in the mantle with low seismic velocity and density intermediate to the MTZ and overlying upper mantle. Hence, we favor a TZWF origin for the 410-LVL and acknowledge that potential inconsistencies between simplified predictions of the TZWF hypothesis and seismic observations, namely 410-LVL thickness and dramatic lateral variability, are yet to be resolved.

4.5. The 410-LVL and Regional Mantle Setting

Globally 410-LVL observations do not correlate with overlying tectonic setting \cite{Tauzin et al., 2010}. Yet, a diverse range of small-scale mantle structure may underlie similar tectonic settings at the surface and correlations with mantle setting may offer valuable insight into melt layer dynamics. The exceptional seismic sampling of USAArray provides an opportunity to more directly assess correlations

![Figure 13](image-url)  
**Figure 13.** Frequency dependence of $P_{660S}$ amplitude. The amplitude of $P_{660S}$ in the cumulative (black), southeast (red), southwest (green), and northwest (blue) corridor stacks is plotted for each array. The mean of all corridors (excluding cumulative array stacks) is included for reference.

![Figure 14](image-url)  
**Figure 14.** Detections of 410-LVL. The 18 corridor stacks with a short corner period of 4.5 s are arranged with stacks containing robust $P_{W,S}$ arrivals on the left, ambiguous $P_{W,S}$ arrivals in the middle, and absent $P_{W,S}$ arrivals on the right. The subset of stacks with robust $P_{W,S}$ arrivals are arranged with depth of $P_{W,S}$ increasing from left to right. Identification criteria for robust $P_{W,S}$ arrivals are discussed in section 3.1. The mean depths of 410 and 660 are plotted for reference (horizontal gray lines).
between 410-LVL presence and mantle setting via traveltime tomography images of 3-D mantle velocity structure (Figures 15 and 16) [Schmandt and Humphreys, 2010] (similar structure resolved by Sigloch et al. [2008], Burdick et al. [2009], Roth et al. [2008], Schmandt and Humphreys [2010], Obrebski et al. [2010], and Tian et al. [2011]). The 410-LVL is unlikely to be resolved by teleseismic traveltime tomography as the method has poor sensitivity to thin subhorizontal layers, but strong lateral variations associated with convective activity including subduction, lithospheric instabilities, the Yellowstone plume, and small-scale asthenospheric upwellings are well resolved. We combine our 410-LVL detections with selected previous results [Jasbinsek and Dueker, 2007; Jasbinsek et al., 2010] because the similar data density and source-receiver geometry of these studies provides a good basis for comparison to provide greater coverage of the western U.S. mantle. There is not a simple one-to-one correspondence between 410-LVL detections and 3-D velocity structure (Figure 15), but we do find noteworthy relationships regarding the sporadic presence of 410-LVL.

Figure 15. Mantle setting and 410-LVL occurrence. (a) The cumulative 410-LVL detection results from this study, Jasbinsek and Dueker [2007], and Jasbinsek et al. [2010] are overlain on a P wave tomogram of velocity perturbations between 350 and 410 km depth (updated version of tomography from Schmandt and Humphreys [2010]). Approximate locations of Ps conversion points at 410 km depth beneath each array are enclosed by contours: 410-LVL detection (black), ambiguous detection (gray), and 410-LVL absence (white). The surface location of the cross section in Figure 15b and the cross section beneath the Ristra array shown in Figure 16 are labeled (black lines). (b) A vertical cross section through the tomography model from northern California to western Nebraska. The black triangles at the surface shows the location of our high-fold Ps sampling beneath the Mendocino array from this paper and the Lodore and Laramie arrays from Jasbinsek and Dueker [2007].
Beneath the three dense temporary arrays in the northern Rocky Mountains and the linearly continuous Ristra and RISTRA1.5 arrays the 410-LVL is nearly ubiquitous [Jasbinsek and Dueker, 2007; Jasbinsek et al., 2010]. The absence of 410-LVL at the northwest and southeast ends of the Ristra line can be reasonably interpreted as a result of the slab currently imaged in the MTZ recently descending through 410 and ongoing descent of a lithospheric instability [Gao et al., 2004; Song and Helmberger, 2007] (Figure 16) through the 410, respectively. This leaves one corridor stack near the Utah/Colorado border (−109°, 40.5°) with no receiver function detection of 410-LVL and no tomographic evidence that the absence is only a local disruption of an otherwise pervasive layer.

Aside from the RISTRA1.5 array, the distribution of 410-LVL detections is generally more intermittent beneath the other five arrays in this study, and these arrays are located closer to the western plate margin. Both the Wallowa array that samples the mantle inboard of the Cascadia subduction zone in the northwestern U.S. and the arrays sampling the relatively slab-free window of the southwestern U.S. yield positive and negative detections of 410-LVL, which precludes the pos-
sibility of a simple relationship between 410-LVL occurrence and subducted slab proximity.

[46] It is tempting to consider connections between low-velocity anomalies imaged just above the 410 by tomography and the possibility of upward instabilities of the 410 melt layer [Jasbinsek et al., 2010] of the form produced by some geodynamic models [Youngs and Bercovici, 2009]. Yet, the aggregate results do not reveal a consistent correlation with tomographically imaged low-velocity “pipes” or “blobs” above the 410. Thus, we will just note two intriguing locations and acknowledge that we have insufficient evidence to suggest a causal relationship. First, 410-LVL detection in the southeast corridor from the Wallowa array correlates with a strong ~100 km wide low-velocity anomaly in the tomography. This low-velocity anomaly is bounded by a dipping (~45°) high-velocity fragment of Farallon slab to the south and a subvertical “curtain-like” high-velocity anomaly to the north that Schmandt and Humphreys [2011] suggest is an older segment of the Farallon slab stagnant in the upper mantle since a 55 Ma accretion event in the northwest U.S. (Figure 15). Second, the tomograms that include USArray and RISTRA1/1.5 data, confirm the presence of a low-velocity “pipe” extending from the top of the MTZ/410-LVL to near the base of the lithosphere beneath the Navajo volcanic field in the center of the Colorado Plateau [Jasbinsek et al., 2010; Sine et al., 2008].

5. Conclusions

[47] Lateral variations in the depths of transition zone discontinuities and changes from 410-LVL presence to absence occur commonly in the western U.S. mantle over distances of about 300 km. The mean depth of the 660 and magnitude of topography on this discontinuity are consistent with variable water concentrations that locally approach saturation (e.g., RISTRA1.5). In addition to contributions from hydration, the magnitude of 660 topography suggests lateral variations in temperature of >150 K between different back azimuth corridors at five out of six arrays. Frequency dependence of $P_{410}$ and $P_{660}$ amplitudes suggest the total thicknesses of velocity gradients at the 410 and 660 are ~25 km and >50 km, respectively, with greater heterogeneity in the sharpness of 410. Inference of a thick and heterogeneous 410 is consistent with variable high water concentrations. The close proximity of multivariate phase transitions in the garnet and olivine systems provides an explanation for the thickness of the 660. Clear evidence of a 410-LVL is found in 8–11 out of 18 corridor stacks, and positive and negative detections are commonly observed in different corridor stacks from individual arrays. Beneath the linear RISTRA1/1.5 arrays, transitions from 410-LVL presence to absence correlate with a subducted slab fragment and a lithospheric instability imaged by traveltime tomography. This suggests local disruption of a more widespread 410-LVL via regional convective flow. However, correlations between the presence of 410-LVL and 3-D structure imaged by tomography are not observed beneath other arrays. Sporadic detection of 410-LVL, the magnitude and length scales of MTZ topography, and inference of lateral variations in mantle hydration all support the occurrence of vigorous small-scale convective flow in the western U.S. mantle.

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