Pervasive Upper Mantle Melting beneath the Western US

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Abstract

We report from converted seismic waves, a pervasive seismically anomalous layer above the transition zone beneath the western US. The layer, characterized by an average shear wave speed reduction of 1.6%, spans over an area of $\sim 1.8 \times 10^6$ km² with thicknesses varying between 25 to 70 km. The location of the layer correlates with the present location of a segment of the Farallon plate. This spatial correlation and the sharp seismic signal atop of the layer indicate that the layer is caused by compositional heterogeneity. Analysis of the seismic signature reveals that the compositional heterogeneity can be ascribed to a small volume of partial melt (0.5 ± 0.2 vol% on average). This article presents the first high resolution map of the melt present within the layer. Despite spatial variations in temperature, the calculated melt volume fraction correlates strongly with the amplitude of P-S conversion throughout the region. Comparing the values of temperature calculated from the seismic signal with available petrological constraints, we infer that melting in

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the layer is caused by release of volatiles from the subducted Farallon slab. This partially molten zone beneath the Western US can sequester at least 1.2×10^{17} kg of volatiles, and can act as a large regional reservoir of volatile species such as H or C.

Keywords: Transition zone; Mantle Melting; Volatile Cycle; Seismic Anomaly, LVL

1. Introduction

The mantle transition zone plays a unique role in controlling the Earth's volatile cycle. Nominally anhydrous silicate minerals in the transition zone can dissolve substantially larger quantities of H in their crystal structure, compared to the major mantle minerals above and below (Bolfan-Casanova, 2002; Kohlstedt et al., 1996). The gradient in H solubility across the transition zone has been posited as a source for volatile-induced melting atop and below the transition zone in regions of upwelling or downwelling (Bercovici and Karato, 2003; Schmandt et al., 2014). In addition, a recent study of melting of slab carbonates suggests that carbonate phases in subducting slabs can lead to the generation of carbonatitic melts near the base of the upper mantle (Thomson et al., 2016). These constraints from laboratory experiments, as well as constraints on melt density (e.g. Ghosh et al., 2007), indicate the likelihood of a partially molten layer atop the transition zone. Several features of such a layer provide important information regarding the transport and storage of volatiles in and around the transition zone.

Owing to the compositional contrast arising from melting, seismic signature of a partially molten layer should be marked by a sharp boundary, unlike

19 thermal anomalies that can have diffuse boundaries. In addition, thickness
20 and spatial extent of the layer, magnitude of seismic wave speed reduction,
21 and spatial correlation with tectonic features such as subduction can pro-
22 vide additional insight into the origin and nature of melting within the layer.
23 Spatial correlation between the partially molten layer and cold regions of the
24 transition zone can indicate the possibility of volatile-induced melting, as the
25 solidus of dry mantle peridotite is likely higher than the temperature within
26 the zone. In studying such partially molten layers, it is crucial to quantify
27 the amount of melt in the layer as the melt content and its spatial variations
28 can provide indirect evidence for processes associated with the origin, trans-
29 port, and storage of the melt. While a number of previous studies reported
30 the presence of melting atop the transition zone, detailed regional maps of
31 partial melt, derived from seismic observations, still remain scarce.

32 Several previous studies reported the occurrence of low seismic velocity
33 layers (LVLs) 350 km below the surface (Courtier and Revenaugh, 2007; Gao
34 et al., 2006; Revenaugh and Sipkin, 1994; Song et al., 2004; Tauzin et al.,
35 2010; Vinnik and Farra, 2007). The sharp reduction of velocity at the onset
36 of the LVL is sometimes referred to as the 350 discontinuity (Vinnik and
37 Farra, 2007). Recent studies suggest that the LVL can be present on a
38 global scale (Tauzin et al., 2010; Vinnik and Farra, 2007), with the distance
39 above the 410 discontinuity changing laterally from 20 km to as much as
40 90 km over a few hundred kilometers. Correlations of these variations with
41 hot (Hier-Majumder et al., 2014; Vinnik and Farra, 2007) or cold (Courtier
42 and Revenaugh, 2007; Hier-Majumder and Courtier, 2011; Song et al., 2004)
43 tectonic environments have remained elusive (Tauzin et al., 2010), suggesting

44 that the variations in position cannot be explained by temperature alone.

45 In the Western US, studies reporting an LVL are based either on P-S re-
46 ceiver functions and P-wave triplication (Song et al., 2004), P-S receiver func-
47 tions (Fee and Dueker, 2004; Jasbinsek and Dueker, 2007; Jasbinsek et al.,
48 2010; Schmandt et al., 2011), or S-P receiver functions (Vinnik et al., 2010).
49 The LVLs have been found beneath the border between Oregon and Wash-
50 ington (Song et al., 2004), Yellowstone (Fee and Dueker, 2004; Jasbinsek and
51 Dueker, 2007), the northern Rocky Mountains (Jasbinsek and Dueker, 2007),
52 the southern Colorado Plateau and the Rio Grande Rift (Jasbinsek et al.,
53 2010), and under California (Vinnik et al., 2010)(Figure 1a).

54 Due to the lack of coverage and absence of extensive analysis involving
55 rock physics and melt microstructure, however, these studies were unable
56 to quantify the spatial expanse and local variations in the melt content in
57 the LVL. Using limited coverage underneath the Coral Sea and Hawaii, the
58 LVLs were estimated to contain approximately 1 vol% melt (Hier-Majumder
59 and Courtier, 2011; Hier-Majumder et al., 2014). The seismic data in these
60 two studies, however, were too sparse to create a detailed regional map of
61 melting. Such detailed regional maps of melting are crucial in understanding
62 the global volatile cycle, as they allow correlation between the structure
63 and geometry of the melt zone and the tectonic environment, potentially
64 identifying processes involved in melt generation, metasomatism, and melt
65 storage.

66 In this study, we address the issue of a detailed regional map of melting
67 underneath the Western US, using high resolution seismic data. The seis-
68 mic signature of the LVL was derived from 820 seismometers of the dense

69 broad-band US Transportable Array (Figure 1b). We applied the P-to-S re-
70 ceiver function (RF) technique to the records of 932 teleseismic earthquakes
71 giving a set of 65,000 RFs (Tauzin et al., 2013). The RF technique uses
72 compression-to-shear (P-S) converted seismic waves to detect sharp shear-
73 wave velocity changes beneath stations. With such a dense seismic array,
74 the RF technique enables the detection of thin layers in the transition zone
75 over a semi-continental scale and with a high lateral resolution.

76 In the following sections we outline our findings for the Western US.
77 We discuss the methods of analysis of the RF data and the rock physics
78 analysis in Section 2, present our key findings in Section 3, and discuss the
79 implications for the regional volatile cycle in Section 4.

80 **2. Methods**

81 *2.1. Receiver function data analysis*

82 *2.1.1. Data*

83 In this study, we used 3-component broad-band records of passive seis-
84 micity at stations deployed during the US Transportable Array experiment
85 between January 2004 and November 2009 (Tauzin et al., 2013). Waveforms
86 were obtained from the IRIS Data Management Center for 932 teleseismic
87 earthquakes, occurring at depths shallower than 350 km, with epicentral
88 distances between 40° and 95° , and magnitudes of at least 5.5. These earth-
89 quakes were recorded during the two first deployments of the Transportable
90 Array covering the western half of the US at 820 sites (Figure 1b).

91 To extract the signal of waves converted from P-to-S (P-S) at seismic
92 boundaries beneath the receivers, we built receiver functions (RFs) by de-

93 convolving the records of the P-wave rotated along the radial component
 94 by the records rotated along the vertical component. The original records
 95 are low-pass filtered at 5 s period, then deconvolved using an iterative time
 96 domain deconvolution method (Ligorria and Ammon, 1999). In this deconvolu-
 97 tion method, the applied Gaussian function has a width $L = 1.125$ s at half
 98 the maximum amplitude to respect the vertical resolution of converted waves
 99 at TZ depth ($\lambda/2$ where λ is the wavelength of the shear-wave). Quality con-
 100 trol were made, with a selection based on the ratio of the RMS amplitudes of
 101 the signal after the P-wave and of the noise before the P-wave (Tauzin et al.,
 102 2013). The data set consists of 64,578 RFs and provides a good coverage in
 103 P-to-S piercing points at transition zone discontinuities. A precise map of
 104 this coverage is shown in the study from Tauzin et al. (2013). This coverage
 105 is highlighted by the area that is not shaded in Figures 3 and 4. The data
 106 collected in this study have been obtained from the mobile Transportable Ar-
 107 ray and do not include the data from the Caltech Regional Seismic Network,
 108 explaining a gap in coverage in the extreme South of California (Figures 3
 109 and 4).

110 *2.1.2. Imaging procedure*

111 Our imaging procedure is based on common conversion point (CCP)
 112 stacking (Tauzin et al., 2013; Wittlinger et al., 2004). Assuming locally
 113 horizontal interfaces, rays corresponding to S-legs for P-S conversions are
 114 back-propagated with their theoretical azimuth and incidence from each re-
 115 ceiver to depth in the mantle. The back-propagation is computed using the
 116 IASP91 velocity model (Kennett and Engdahl, 1991) after an Earth flatten-
 117 ing transformation (Muller, 1985). For every time sample on the RFs the

118 amplitude is projected to its theoretical spatial location along the ray in the
 119 1D medium. The medium is discretized with blocks of 1 km-vertical and
 120 10 km-lateral dimensions. In these blocks, coherent stacking of the energy
 121 from RFs of many rays incoming from different directions gives the loca-
 122 tions of seismic interfaces. To benefit from data redundancy and increase
 123 the signal-to-noise ratio, the CCP amplitudes within a distance from the
 124 vertical plane along the profile are projected and averaged onto this plane to
 125 obtain a 2D depth-distance CCP section (e.g. Figure 2a). These CCP sec-
 126 tions are smoothed using a Gaussian weighting with the length of semi-axes
 127 corresponding to the lateral extent of the Fresnel zones of converted waves
 128 at transition zone depths (Wittlinger and Farra, 2007).

129 We constructed two types of seismic sections, dedicated to describe the
 130 2D structure at different length-scales. To obtain a section with high signal
 131 ratio, such as in Figure 2, we projected the amplitudes within ± 200 km from
 132 the vertical plane along the profile. For building higher resolution maps of
 133 the discontinuity structure over the western US, we reduced the lateral extent
 134 of projection of the CCP signal to ± 100 km from the vertical plane of the
 135 profile (see Tauzin et al., 2013, for details).

136 The geographical distribution and variation in depths of the major seismic
 137 boundaries (the 410, 660 and the LVL) were obtained from picking their
 138 signal on 2D CCP seismic sections along orthogonal profiles at intervals of
 139 0.5° in longitude and latitude. In that case, the distance of projection is ± 100
 140 km around the profile. We derived the maps of the interfaces by combining
 141 the information provided by the North-South and East-West seismic cross
 142 sections (Tauzin et al., 2013). From the picked topography of the 410 and 660

143 boundaries, we estimated variations in the TZ thickness from the reference,
 144 δh (Figure 3a), which is a more robust proxy for the temperature within the
 145 TZ than the absolute depth of discontinuities (Tauzin and Ricard, 2014a).
 146 We also picked the negative signal associated with the P-S conversion at the
 147 top of the LVL (Figure 4a). The depth difference from the 410 discontinuity
 148 provides an estimate of the LVL thickness (Figure 4b).

149 The vertical resolution achieved by the data, *i.e.* the ability to separate
 150 vertically two seismic interfaces, is $\lambda/2 \approx 10$ km, where $\lambda = C T$ is the
 151 wavelength of the shear-wave given the minimal period T of the data and C
 152 the velocity ($T = 5$ s and $C = 4.5$ km/s). This vertical resolution ensures
 153 the reliable detection of a 20 to 90 km-thick LVL atop the 410 discontinuity
 154 (Figure 1a-b). The lateral resolution, given by $\sqrt{\lambda z / \cos^2 i_c}$ where z is the
 155 depth of analysis and i_c the incidence angle (Cerveny, 2005; Wittlinger and
 156 Farra, 2007), is of the order of 100 km at transition zone depths. Finally,
 157 the maximal vertical extension of a detectable velocity gradient is given by
 158 $\lambda_P/4 \approx 10$ km, λ_P being the P-wavelength (Bostock, 1999).

159 Along the A-A' profile, we applied a bootstrap resampling approach
 160 (Efron and Tibshirani, 1990) to provide an estimate of the standard errors
 161 on the depths and amplitudes of seismic signals. We constructed $N = 50$
 162 bootstrap samples by choosing randomly 67% of the original RF data set and
 163 duplicating 33% of it to complete the data set. We then applied the CCP
 164 stacking approach on these N samples to obtain an average seismic section
 165 and a standard error on the seismic amplitudes (Figure 2a-b). Clearly, along
 166 the A-A' profile in Figure 2, amplitudes on the RF section are reliable up to
 167 the lateral distance of +500 km along the profile. Further East, the strongly

oscillating seismic amplitudes are the result of noise, as demonstrated by the large standard errors in Figure 2b.

We also obtained for the A-A' profile the errors on δh , the LVL thickness, and P-S conversion amplitudes. We repeated the picking operation over the N bootstrap samples, then measured an average and a standard deviation. Figure 2c-d gives the result of such a picking on the N bootstrap samples (gray lines), the average (solid lines), and the 95% confidence levels (dashed lines) for the depths and seismic amplitudes of the 410-km discontinuity and the top of the LVL. The temperature distributions obtained from the bootstrap samples of δh along the A-A' profile are also shown in Figure 5b.

2.1.3. Computing the shear-wave velocity in the LVL

In addition to the depth of interfaces, we are also interested in the seismic shear-wave velocity at ~ 350 km depth within the LVL, V_S^{obs} . We use the method outlined in Hier-Majumder et al. (2014) to estimate this seismic velocity. In this method, we calculate V_S^{obs} from the P-S conversion amplitudes A_{PS} , determined from the transmission coefficients at discontinuities. First, we define the measured amplitude ratio of the conversion at the top of the LVL and at the 410 discontinuity, R_{norm}^{LVL} as,

$$R_{norm}^{LVL} = \frac{A_{PS}^{LVL}}{A_{PS}^{410}} \leq 0. \quad (1)$$

The map in Figure 4(c) depicts the distribution of this ratio over the western US. The value of this ratio is plotted with uncertainties for the A-A' profile in Figure 5e.

Variations in the amplitudes of conversion, A_{PS}^{LVL} , among sites can result from site specific conditions, and may not necessarily reflect the variations

caused by properties of the LVL. Using the ratio normalizes this effect, as both amplitudes should be equally influenced by conditions specific to one site. The ratio of the amplitudes is then used to determine the observed shear wave speed atop the LVL from wave-speeds at 410 km and 350 km in the PREM model,

$$V_S^{obs} = V_S^{norm} (1 - R_{norm}^{LVL}) + V_S^{410} R_{norm}^{LVL} \quad (2)$$

where V_S^{410} and V_S^{norm} are absolute shear-wave speeds from the PREM model (Dziewonski and Anderson, 1981) at the depths of 410 and 350 km, respectively. Notice that $R_{norm}^{LVL} \leq 0$, thus the calculated V_S^{obs} , is not a weighted average of the reference velocities V_S^{norm} and V_S^{410} in equation (2). In Section 2.3, we describe the way this observed shear-wave speed is used to calculate the unknown melt volume fraction at each site. The map in Figure 6(a) depicts the distribution of this velocity, V_S^{obs} , over the western US.

2.2. Temperature calculation

The temperature of the mantle plays an important role in the seismic signature of the low velocity layer, by controlling the bulk and shear moduli of the solid. Lateral variations in the temperature can trade-off with the signature of the mantle melting. It is, therefore, crucial to account for spatial variations in temperature.

We use the method outlined in Tauzin and Ricard (2014a) to calculate the temperature from the thickness of the TZ below each site of observation of the LVL. This empirical method extracts from the apparent δh a relationship between the 410 and 660 Clapeyron slopes, and takes into account in a self-consistent manner the effect of temperature on boundary topography and

214 seismic velocities. The advantage of this approach is that it does not rely on
 215 *a priori* corrections from independent tomographic models.

216 First, we assume a reference mantle temperature T_{ref} which corresponds
 217 to the top of the transition zone coincident with the global depth of 410
 218 km. Since both olivine polymorph transitions at 410 and 660 km depths are
 219 sensitive to temperature, any deviation of the TZ thickness from a reference
 220 thickness indicates a temperature anomaly in the TZ. In the western US,
 221 the TZ is in average 252-km thick and shows ± 25 km lateral variations
 222 (Figure 3a). Corresponding temperature variations normally depend on the
 223 Clapeyron slopes of the 410 and 660 phase transitions. The seismic analy-
 224 sis of Tauzin and Ricard (2014a) revealed a linear relationship between the
 225 Clapeyron slopes ($\gamma_{660} = -0.64 \gamma_{410} - 1.17$), allowing to parameterize the
 226 temperature- δh relationship only with one of the Clapeyron slopes. We thus
 227 expressed the temperature under each site as,

$$T = T_{ref} + \Delta T(\gamma, \delta h), \quad (3)$$

228 where δh is the TZ thickness variation, γ is the Clapeyron slope of the
 229 410 phase transition, and the exact expression of the temperature anomaly
 230 $\Delta T(\gamma, \delta h)$ is given by Tauzin and Ricard (2014a). To test the sensitivity of
 231 our results to the prescribed Clapeyron slope, we calculated T for 9 differ-
 232 ent values of γ varying between 0.5 to 4.5 MPa/K. In this article, we show
 233 the results for a Clapeyron slope of +3 MPa/K, as suggested by Tauzin and
 234 Ricard (2014a).

235 2.3. Rock physics analysis

236 Small amounts of partial melting can exert significant influence on the
 237 effective physical properties of rocks. The extent of this influence depends on
 238 the volume fraction of the melt; geometry of the melt; and physical properties
 239 of the solid matrix and the melt, which, in turn, are controlled by temperature
 240 and composition of both phases. In this analysis, we use the method outlined
 241 in the study of Hier-Majumder et al. (2014).

242 The principal component of this analysis involves predicting a reference
 243 seismic wave speed $V_S^{ref}(T, C)$ based on the composition of the solid reference
 244 mantle C and the temperature T at each LVL site. We use the data from
 245 Xu et al. (2008), for a bulk composition containing 40% basaltic component.

246 The map in Figure 6(b) shows the distribution of this velocity, V_S^{ref} , in the
 247 western US. The variations in V_S^{ref} are primarily caused by the variations
 248 in the temperature, as depicted by the map of TZ thickness variation in
 249 Figure 3(a). We also define the melt anomaly function $\xi(\theta, \phi, \rho_m, \rho_s)$,
 250 which depends on the dihedral angle, θ , densities of the solid and the melt
 251 (ρ_s and ρ_m , respectively), and the unknown melt volume fraction, ϕ . We
 252 then constrain ξ such that V_S^{ref} matches the observed seismic wave speed
 253 V_S^{obs} , within a preset tolerance, $\epsilon = 1 \times 10^{-9}$, such that,

$$V_S^{obs} - \xi V_S^{ref} = \epsilon. \quad (4)$$

254 Since V_S^{ref} depends on temperature and solid composition, while ξ depends
 255 on melt related parameters, it is useful to define the anomaly ΔV_S , which,
 256 assuming $\epsilon \approx 0$ in equation (4), is given by

$$\Delta V_s = \frac{V_S^{obs} - V_S^{ref}}{V_S^{ref}} = \xi - 1. \quad (5)$$

While the observed amplitude ratio, R_{Norm}^{LVL} depends on temperature, solid composition, and melting, this anomaly isolates the effect of melting. In the absence of melting, marked by $\xi = 1$, $\Delta V_s = 0$. In the presence of melting, $\xi > 1$, and $\Delta V_s < 0$.

Using the formulation for ξ prescribed in Hier-Majumder et al. (2014), we can rewrite equation (4) as,

$$V_S^{obs} - V_S^{ref} \sqrt{\frac{(1 - \phi)(1 - (1 - \psi(\theta, \phi))^n)}{\left(1 + \phi \left(1 - \frac{\rho_m}{\rho_s}\right)\right)}} = \epsilon, \quad (6)$$

where $\psi(\theta, \phi)$ is the dihedral angle and melt fraction dependent contiguity, the fractional area of intergranular contact. The exponent n also depends on the contiguity ψ . We use the formulation of von Bagen and Waff (1986), to express contiguity as a function of melt volume fraction for a given dihedral angle. Equation (6) thus becomes a nonlinear, implicit equation in the unknown ϕ .

We solve the nonlinear equation (6) for ϕ , using a modified Newton-Raphson method for each of the 583 sites using the numerical model MuMaP (Hier-Majumder et al., 2014). To account for variations in the Clapeyron slope, dihedral angle, and reference potential temperature, we carried out a total of 46,507 analyses over 583 locations. We carried out a series of numerical experiments for 5 different reference mantle temperatures between 1400 K and 1800 K, in increments of 100 K, and Clapeyron slopes of the olivine-to-wadsleyite phase transition ranging from +0.5 and +4.5 MPa/K, in increments of 0.5 MPa/K, for a dihedral angle of 25°, resulting in a total

278 of 26,235 analyses. In addition, we carried out another set of simulations
 279 for the same range of temperatures, a Clapeyron slope of +3 MPa/K, and
 280 7 values of dihedral angle ranging from 10° to 40° , in increments of 5° , re-
 281 sulting in another additional 19,822 analyses to study the variation in the
 282 calculated melt volume fraction as a function of the dihedral angle. The
 283 resulting variation in calculated melt fraction as a function of dihedral angle
 284 is plotted in Figure 6(b). The uncertainty in our calculations arising from
 285 assumptions about the Clapeyron slope and the dihedral angle is discussed
 286 in the following subsection.

287 Global correlation between petrological and seismic thermometers sug-
 288 gests a negative linear correlation between TZ thickness and mantle potential
 289 temperature (Courtier et al., 2007). For the regional average TZ thickness
 290 of 252 km in this study, this correlation yields a reference mantle potential
 291 temperature of 1554 K for the region. In this article, we chose to report the
 292 results for a reference mantle temperature of 1600 K, Clapeyron slope of +3
 293 MPa/K and a dihedral angle of 25° . Our analyses yield the melt volume
 294 fraction for the 583 locations in the region (Figure 7).

295 *2.4. Uncertainty analysis*

296 We distinguish two types of uncertainties associated with our results. The
 297 first type arises from seismic observations (variations in the transition zone
 298 thickness δh , and the amplitudes of P-S conversions R_{norm}^{LVL}), while the second
 299 type is associated with the rock physics analysis (the Clapeyron slope γ , the
 300 dihedral angle θ , the bulk mantle composition C , the melt density ρ_m). A
 301 similar analysis by Hier-Majumder et al. (2014) reveals that the influence of
 302 bulk mantle composition and melt density on the predicted melt fraction is

insignificant. Consequently, we focus on estimating the uncertainty arising from the remaining 4 parameters. This uncertainty can be calculated by numerical propagation of errors.

Propagated numerically, the uncertainty in the final measurement, α , can be expressed as a function of uncertainties α_i in the values of the parameter x_i ($x_i = \theta, \gamma, \delta h, R$), and the derivative of the average melt fraction with x_i as,

$$\alpha^2 = \sum_i \alpha_i^2 \left\langle \frac{\partial \phi}{\partial x_i} \right\rangle_{j \neq i}^2, \quad (7)$$

where $\langle q \rangle$ is the arithmetic mean of the quantity q . We calculated the derivatives $\langle (\partial \phi) / (\partial x_i) \rangle$ numerically from the data for each of these 4 variables, keeping the other 3 constant, as shown in Supplementary Figure 1. We present the values of the derivatives and the uncertainties in Table 1 in the supplementary material. We use the uncertainty in the Clapeyron slope from a compilation of laboratory measurement by Tauzin and Ricard (2014b). The uncertainty in dihedral angle is relatively poorly constrained. We chose to use a value of 5° , similar to the variations observed in carbonate melt -olivine ($25^\circ - 30^\circ$) systems by Minarik and Watson (1995).

Our estimates of the standard errors on seismic parameters have been obtained from bootstrap resampling (see section 2.1.2). We provide the average standard errors α_i for the A-A' profile and derivatives $\left\langle \frac{\partial \phi}{\partial x_i} \right\rangle_{j \neq i}$ in Table 1 of the supplementary material. Generalizing the bootstrap approach to the whole western US region in a future work will provide a basis for interpreting lateral variations of melt volume fraction given their uncertainties.

3. Results

We report the layer over a region of $\sim 1.8 \times 10^6$ km² in Western US, below the Cascade Ranges, the Rocky Mountains and Yellowstone caldera, and the topographic lows of the Snake River plain and Columbia plateau (Figure 4a). Given the frequency content of the RF data, the top of the LVL is sharp, characterized by a velocity gradient over a vertical distance smaller than 10 km. As shown on either side of the dashed line in Figure 4b, this interface is located at a height of 42 ± 6 km above the transition zone with significant lateral variations, from 45-70 km in the southwest to 20-35 km in the northeast. The amplitude of the P-S conversion shown in Figure 4c is an indirect measure of the shear-wave velocity drop of -0.2% to -5% across the region, with a median value of -1.6%.

Multiple lines of evidence demonstrate a clear spatial correlation between the LVL and the subducted former Farallon plate. Increased thickness of the transition zone over a large area ($\sim 800 \times 1000$ km²) west of the NNW-SSE boundary (dashed line in Figure 3a) coupled with the high P-wave velocity (Figures 3b and 5a) outline the cold slab. In addition, the calculated mantle temperature (Figure 5b) suggests that the transition zone underneath the western side of the Snake River plain is nearly 150 ± 80 K colder than the surrounding mantle. In addition, there is a good agreement in this region between tomography models and predicted location of the subducted Farallon plate from flow models constrained by plate motions (e.g. Liu and Stegman, 2011). Therefore, the elongated fast and cold body, stalled within the transition zone from the California-Nevada border to the Snake River plain and Yellowstone region (Figures 3b and 5a), is likely a remnant of the subducted

350 Farallon plate.

351 The LVL is likely caused by compositional anomaly rather than elevated
352 mantle temperature. Overlying the 410 discontinuity, the pervasive negative
353 signal marking the sharp drop in velocity atop the LVL directly overlies the
354 cold slab fragment (Figure 5a, c). The LVL is the thickest and marked by the
355 strongest velocity contrast at a distance of -250 km along the profile (Figure
356 5d-e), where the transition zone is the coldest (Figure 5b). In addition,
357 the LVL thickness drops gradually as the plate approaches the Yellowstone
358 Caldera and the associated warm plume. At West, where the Farallon-Juan
359 de Fuca plate is the youngest and the temperature is cold (95 ± 25 K below
360 the Basin and Range in Nevada), the LVL is absent (Figs. 4 and 5).

361 The partially molten region is widespread in spatial extent, marked by a
362 small average amount of 0.5 ± 0.2 vol% melt (Figure 7) for a reference mantle
363 temperature of 1600 K. The spatial distribution of the highest melt fraction
364 shows a distinctive linear trend roughly parallel to the western boundary of
365 the North American plate (Figure 7a). Within the region of coverage, the
366 drop in the melt volume fraction toward the NE at the edge of the stalled
367 plate in the transition zone coincides with the decrease in thickness of the
368 LVL, as shown in the cross section along the A-A' profile in Figure 5. The
369 map of ΔV_s , the melt fraction dependent wave speed anomaly as defined in
370 equation (5), is plotted in Figure 7(b). The anticorrelation between the the
371 melt volume fraction and ΔV_s is clear from visual comparison of the two
372 maps in Figures 7(a) and (b). Histograms of the two quantities are shown in
373 panel (c) annotating their median values. It is worth remembering that ΔV_s ,
374 defined in equation (5), only isolates the anomaly associated with melting,

375 as the effects of the temperature and bulk solid composition are incorporated
 376 in V_S^{ref} .

377 The amplitude of P-S conversion shows an anticorrelation with the cal-
 378 culated melt volume fraction, with a Pearson correlation coefficient of -0.75
 379 in figure 7(d). Such an anticorrelation is expected since the melt fraction is
 380 calculated from the amplitude of conversion and a higher magnitude of of
 381 conversion is likely to give rise to a higher melt volume fraction. The scatter,
 382 in this raw data, however, arises from variations in temperature between dif-
 383 ferent stations. If the temperature at each location were the same, the data
 384 points in this plot would follow a smooth curve. This is shown in supplemen-
 385 tary Figure 1(c), where these smooth curves are plotted for three different
 386 values of δh , the proxy for temperature. When the effect of temperature is
 387 removed, a smooth curve is also obtained for ΔV_S , which isolates the effect
 388 of melting, as shown in Figure 7(e).

389 In the following section, we discuss the implications of our results for
 390 volatile transfer and storage in the mantle and the long term stability of the
 391 melt within the TZ.

392 4. Discussion

393 4.1. How robust are seismic observations?

394 An apparent LVL signal atop the 410-km discontinuity could result from
 395 various spurious signals on migrated CCP sections; (1) uncorrelated noise
 396 on raw and deconvolved seismograms, (2) sidelobes due to deconvolution,
 397 (3) interference of seismic phases, and (4) multiple reverberations within the
 398 uppermost mantle structure. While these issues are discussed in Tauzin et al.

399 (2013), we present some key arguments below.

400 (1) With the high density US Transportable Array data, uncorrelated
401 noise is not expected to give coherent signals over large-scale RF sections,
402 and our uncertainty analysis from bootstrap resampling (section 2.1.2) clearly
403 demonstrates that the LVL signal magnitude exceeds the noise level. (2) Side-
404 lobes may be introduced by the deconvolution (Ligorria and Ammon, 1999)
405 but are expected to be symmetric and on both sides of the main structural
406 discontinuities. We demonstrated that these features appear neither on real
407 nor on synthetic data (Tauzin et al., 2013). (3) In addition, our synthetic
408 tests did not indicate any interference effects from seismic phases (such as
409 PP or PcP) on the recovery of the CCP signal (Tauzin et al., 2013). (4)
410 Reverberations in the shallow structure may be mistaken for direct conver-
411 sions and wrongly interpreted as a true structure. Two reasons, however,
412 indicate that it is not the case. First, to obtain flat converters, such as the
413 LVL, requires flat and strong interfaces at shallow depths, which is unlikely
414 given the present knowledge of the upper mantle structure in Western US.
415 Second, at the scale of the western US, the precursory arrival to the P410s
416 wave displays slowness similar to the conversion at the 410, (see Figure 2 in
417 the supplementary material) demonstrating that this arrival is due to a true
418 conversion at the top of the LVL.

419 4.2. *Source of volatiles*

420 Volatiles from the subducted Farallon plate must play an important role
421 in melting in the LVL. We note that the temperature of the slab surface for
422 both hot and cold slabs at depths of 350 km (Turcotte and Schubert, 2001)
423 are colder than the solidus of mantle peridotite by 600-800 K, as shown

424 in Figure 8(a) and (b). To induce melting in the mantle at the observed
425 depth, therefore, the presence of a volatile phase, namely H and/or C is
426 crucial. While the seismic signal at such small melt fractions is insensitive
427 to the melt composition (Hier-Majumder et al., 2014), a few indirect lines of
428 evidence can point to the potential source of the melt (Figure 8(c)).

429 One evidence for melting induced by slab-derived C comes from a recent
430 experimental study, suggesting that the carbonates in the subducting slab
431 largely escapes devolatilization in the upper mantle leading to the gener-
432 ation of carbonatitic melt above the transition zone (Figure 8a; Thomson
433 et al., 2016). In the strongly reducing mantle environment, however, the car-
434 bonatitic melt will rapidly react with the surrounding solids (Rohrbach and
435 Schmidt, 2011), leaving behind a small amount of residual, unreacted melt.
436 The small average melt volume fraction in our study is compatible with the
437 signature of residual C-depleted melt resulting from such redox freezing in
438 the mantle.

439 A second possible source of volatiles (Figure 8(c)) can arise from dehydra-
440 tion of high pressure mineral phase E, a dense hydrous silicate mineral, in the
441 transition zone (Ohtani et al., 2004). Coupled with mantle upwelling due to
442 Yellowstone plume or small scale convection between the LVL and the slab,
443 such dehydration can trigger melting, observed in the LVL (Bercovici and
444 Karato, 2003; Richard and Bercovici, 2009). The observed small quantity of
445 melt, in this scenario, is likely explained by a modest amount of H released
446 upon dehydration, which reduces the solidus temperature modestly (Figure
447 8(b)). The small volume fraction of melt will remain bound to triple grain
448 junctions in the matrix and will likely be advected with the matrix following

449 the regional mantle flow pattern.

450 The melting zone reported here has a large storage capacity for volatiles.
451 Considering a concentration of 1 to 10 wt% of H₂O or CO₂ in the melt,
452 the melt layer underneath the Western US (1.8×10^6 km² areal span and
453 40 km thick, density of 3500 kg/m³) can store up to $1.2 \times 10^{16-17}$ kg of
454 H₂O or CO₂ in the LVL. If the observed melt is a CO₂ depleted residual
455 melt from mantle metasomatism, then this estimate serves as a lower bound
456 to the amount of CO₂ that can be sequestered to the LVL upon melting.
457 For example, a global LVL of similar thickness and 500 km width, associated
458 with all subductions zones (global length of 37,000 km; Reymer and Schubert,
459 1984) and containing 1 vol% carbonatitic melt (prior to redox freezing) with
460 50 wt% CO₂ (Thomson et al., 2016), can sequester 1.3×10^{19} kg of CO₂
461 in the upper mantle prior to metasomatism, a substantial portion of the
462 mantle carbon reserve (10^{19-20} kg; Dasgupta and Hirschmann, 2010). The
463 presence of such a deep CO₂ reservoir can explain the low H:C ratio of the
464 bulk silicate Earth compared to the surface abundances (Hirschmann and
465 Dasgupta, 2009). While our results do not permit direct identification of the
466 volatile species associated with the LVLs, they demonstrate that the melt
467 in the LVL, owing to its large volatile storage capacity, can act as a larger
468 global reservoir of volatiles than previously assumed.

469 4.3. *Stability of the melt within the LVL*

470 An important consideration following the generation of the melt is the
471 retention of the melt in the source region. We argue that independent of the
472 nature of the volatile species leading to melting, percolation of the observed
473 small melt volume fraction will be inefficient for two reasons. First, volatile-

rich melts should be nearly neutrally buoyant at the base of the upper mantle (Ghosh et al., 2007), reducing the primary driving force for percolation. Second, even in the presence of a density contrast between the melt and the matrix, mobility of the melt will be substantially reduced by the small melt fraction and low permeability, resulting in geologically significant residual time of the melt within the LVL (Hier-Majumder and Courtier, 2011).

To calculate the efficiency of percolation of residual melt out of the melting zone, we followed the analytical solution outlined by Hier-Majumder and Courtier (2011). In this model, we consider a 1-D melting column of constant melt volume fraction underneath each station location. The melt migration model assumes the matrix is motionless atop and at the bottom of the layer, isolating only density contrast-driven compaction within the layer. The velocity of melt and the matrix are coupled by conservation equations for mass and momentum (Bercovici et al., 2001). The analytical solution for the nondimensional velocity of the melt is given by,

$$v'_m = - \left[\frac{(1 - \cosh z_0) \sinh z}{\sinh z_0} + \cosh z - 1 \right] \frac{3R}{4} \phi_0 (1 - \phi_0)^2, \quad (8)$$

where z is the nondimensional height of the column, R is the dimensionless density contrast, ϕ_0 is the melt volume fraction recorded at the location, and $z_0 = 1/\sqrt{\phi_0(1 - \phi_0)}$. We use a value of $R = 0.05$ in arriving at the analytical solution, which represents a 5% density contrast between the melt and the matrix, compared to the density of the matrix. While petrological measurements indicate that volatile-rich silicate melts are likely to be neutrally buoyant at the base of the upper mantle (Ghosh et al., 2007), this calculation demonstrates even if the driving force was present, melt percolation will still be inefficient. Equation (8) indicates that the percolation velocity of the melt

498 will be zero for a neutrally buoyant melt, for which $R = 0$.

499 We dimensionalize the velocity in the melt column using the scheme,

$$v_m = \frac{4\rho_m g}{3c} v'_m, \quad (9)$$

500 where $\rho_m = 3300\text{kg/m}^3$ is the matrix density, $g = 10\text{m/s}^2$ is gravity, and
 501 the frictional resistance to melt percolation (Pas/m^2) can be expressed as
 502 (Hier-Majumder, 2011)

$$c = \mu_f \frac{72\pi}{a^2 \phi_0^2}. \quad (10)$$

503 In equation (10), $\mu_f = 1\text{ Pas}$ is the viscosity of the melt and $a = 1\text{ mm}$
 504 is the grain size. To create the map of melt velocity in Figure 9, we take
 505 the maximum value of the absolute magnitude of velocity within the vertical
 506 column.

507 The dimensional permeability, k , at each point is given by the relation,

$$k = \frac{\mu_f}{c}. \quad (11)$$

508 The results for the melt percolation velocity and permeability at each loca-
 509 tion are shown in Figure 9. The mean velocity of the melt is $\sim 30\mu\text{m/yr}$,
 510 indicating that it will take nearly 1 Ga for melt to be completely extracted
 511 from a melt layer of 40 km thickness. In this calculation, we do not consider
 512 the effect of surface tension at grain boundaries, which will render melt ex-
 513 traction even more ineffective (Hier-Majumder et al., 2006). The time scale
 514 of 1 Ga for melt stability in the LVL is similar to the estimated residence
 515 time for deep carbon (Hirschmann and Dasgupta, 2009), suggesting that the
 516 melting zone can act as a potential deep carbon reservoir.

517 5. Conclusion

518 In summary, our results identify a pervasive LVL underneath the western
519 US spanning over an area of $1.8 \times 10^6 \text{ km}^2$. The LVL shows significant varia-
520 tions in thickness in this region with some of the thickest regions associated
521 with the coldest parts of the mantle. It also shows a clear spatial correlation
522 with the stalled Farallon plate. The average melt content of the LVL is 0.5
523 $\pm 0.2 \text{ vol\%}$, with significant spatial variations. The sites with the strongest
524 seismic anomaly register the highest melt volume fraction. These evidences
525 suggest that melting induced by slab-derived volatile species plays a crucial
526 role in the origin of the LVL.

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Figure 1: Low-velocity layer observations from dense seismic arrays. (a) A map summarizing previous observations of the LVLs across Western US. These observations have been obtained from several independent studies using small-aperture seismic networks. (b) The larger-aperture seismic network used in this study is the Transportable Array (black triangles), covering the western half of the US with an average station spacing of 70 km. The Caltech Regional Seismic Network has not been processed, explaining a gap in coverage in the extreme South of California. The seismic profile discussed in this study is labeled A-A' and marked with a black line. Important physiographic features of the western US are labeled, such as the Cascadia ranges (CR), the Yellowstone Snake River plain (YSRP), the Yellowstone caldera (Y), the Columbia plateau (CP), the Sierra Nevada (SN), and the Basin and Range province (B&R). Major Quaternary active volcanoes (red triangles) are either arc-related, due to the present subduction of the Juan de Fuca plate below the Cascadia ranges, or possibly hotspot-related in the Snake River Plain and Yellowstone regions.

Figure 2: The result of applying common conversion point stacking and bootstrap resampling on the RFs of the A-A' profile. (a) The average seismic section. (b) One standard deviation giving the 65% confidence level on seismic amplitudes. A robust interpretation can be conducted up to +500 km along the profile. (c) The result of bootstrap resampling for the depth of the 410-km discontinuity and the top of the LVL with the $N = 50$ samples (gray lines), the averages (blue and red solid lines), and the 95% confidence levels (blue and red dashed lines). (d) Same as (c) except that we analyze the P-S conversion amplitudes at the 410-km discontinuity and at the top of the LVL.

(a) TZ thickness variation

(b) Tomography

Figure 3: The transition zone structure. (a) An estimate of the variation in transition zone thickness below the seismic array (Tauzin et al., 2013). (b) A cross-section at 410 km depth through the P-wave tomographic model of Burdick et al. (2010). The area that is not covered by the seismic array is shaded in gray. The black dashed line that is roughly parallel to the coastline delimits distinctive patterns of thickened TZ and fast velocity anomalies from other regions of normal mantle in the northeast.

Figure 4: The 350-km deep low-velocity layer below the western US. (a) The LVL (black dots) spreads out across most part of the region, including the topographic highs of the Cascadia ranges (CR), the Rocky Mountains (RM) and the Yellowstone caldera (Y), but also the topographic lows of the Yellowstone Snake River plain (YSRP) and Columbia plateau (CP). At the southwest, the LVL is absent below the Sierra Nevada (SN) and the Basin and Range province (B&R). Major Quaternary active volcanoes are indicated with red triangles. The seismic structure along the A-A' profile is described in detail in Figure 5. (b) A map of the lateral variations of the LVL thickness. (c) Map of the ratio of the amplitude of the P-S conversion at the LVL relative to the 410. In bottom panels, the area with no seismic data coverage is shaded in gray. The black dashed line that is roughly parallel to the coastline delimits distinctive patterns of thick LVL and high-amplitude of conversion at the uppermost interface of the LVL, from other regions of thinner LVL and lower amplitude of conversion in the northeast.

Figure 5: The seismic structure of the LVL. (a) The superimposition of a cross-section through the P-wave tomographic model of (Burdick et al., 2010) on our RF seismic section along the A-A' profile indicates that the slow regions are mostly located in the uppermost part of the mantle, below the Sierra Nevada (SN), the Basin and Range (B&R), and the Yellowstone Snake River plain (YSRP). An elongated fast velocity body represents a fragment of the Farallon plate (F), stalled within the transition zone. In the northeast, at the top of the lower mantle, a slow velocity anomaly has been interpreted as the Yellowstone plume (YP) (Obrebski et al., 2011), piercing through the transition zone, and reaching the Yellowstone Caldera in surface (Y on the topographic profile). (b) Measured temperature anomalies indicate a succession of hot (red) and cold (blue) thermal regimes in the transition zone. (c) A zoom over our RF seismic section depicts the fine structure of the LVL atop the 410-km discontinuity. Negative amplitudes (blue) mark a shear-wave velocity decrease, such as atop the LVL, whereas positive amplitudes (red) mark a shear-wave velocity increase, such as at the 410 discontinuity. (d) The LVL is the thickest at the western side of the YSRP and gradually decreases toward the Yellowstone region. (e) P-S conversion amplitudes at the LVL follow roughly the same trend.

Figure 6: (a) Map of the observed seismic wave speed, V_S^{obs} , calculated from the amplitude of P-S conversion using equations (2). (b) Map of the reference shear wave speed from the model of Xu et al. (2008), for a reference mantle potential temperature of 1600 K and a bulk peridotite composition containing 40% basaltic component. The lateral variation in V_S^{ref} is caused by variation in temperature calculate from the variations in transition zone thickness as shown in Figure 3(a).

Figure 7: (a) Map of the melt volume % derived at the 583 sites of observation of the LVL. In all panels, the reference mantle temperature is 1600 K, the Clapeyron slope of olivine-Wadsleyite transition is +3MPa/K and the melt dihedral angle is 25°. (b) Map of ΔV_S in % values. (c) Histogram of melt volume % and ΔV_S from the maps in panel (a) and (b). The median value of each variable is annotated in the plot. (d) Plot of the normalized amplitude of P-S conversion as a function of calculated melt volume %. The Pearson correlation coefficient of the data points is -0.75. (e) Plot of ΔV_S as a function of melt volume fraction.

Figure 8: Plots of peridotite solidi and key reactions. In both plots the dry peridotite solidus is constructed from Hirschmann (2000) for pressures less than 10 GPa and Hirschmann et al. (2009) for pressures above 10 GPa. The slab geotherms were calculated for subduction velocities of 10 (cold) and 7 hot cm/yr, respectively (Turcotte and Schubert, 2001, Ch. 4-29). The carbonatite solidus in panel (a) is taken from Thomson et al. (2016). The wet peridotite solidus and the dehydration of dense hydrous magnesium silicate (DHMS) fields are taken from Ohtani et al. (2004). This cartoon outlines a few possible hypotheses on devolatilization reactions and their consequences in and around the transition zone. The observed melt in the LVL can arise from decarbonation melting atop the transition zone (Thomson et al., 2016) or subsolidus dehydration of dense hydrous silicate minerals followed by small scale convection above the stalled slab (Ohtani et al., 2004; Richard and Bercovici, 2009). In the small scale convection model, the small plumes are melt-free, hydrated, low-viscosity, low-density aggregates of solid. In the first model, a redox freezing zone (Rohrbach and Schmidt, 2011), which may not be seismically visible, should surround small amounts of residual melt. Dehydration of hydrous ringwoodite, dragged along the slab triggers dehydration melting beneath the transition zone (Schmandt et al., 2014).

Figure 9: Results from the melt migration calculation. (a) A map of melt percolation velocity at each site. The velocity at each site represents the maximum velocity in a column described by equation 8. (b) A map of permeability at each point calculated using the melt fraction determined from the P-S receiver function analysis.