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Seismically deduced thermodynamics phase diagrams for the mantle transition zone

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Abstract

Seismic discontinuities at 410 and 660 km depth are usually attributed to solid phase changes within the olivine component of the mantle. The Clapeyron slopes γ_{410} and γ_{660} , *i.e.* the thermal dependence of the depths of reactions, have been shown experimentally to be of opposite signs. Yet, their values are not well constrained by laboratory measurements. Seismological observations have not been more precise due to the difficulty to separate the competing effects of background wave-velocities and of temperature on the topography of discontinuities. In this study we use conversion imaging of interfaces under western US. We propose a new approach to derive a seismological estimate of the Clapeyron slopes with respect to γ_{410} for the major and minor phase changes of the transition zone. We obtain $\gamma_{660} \approx -3 \text{ MPa K}^{-1}$ for $\gamma_{410} \approx +3 \text{ MPa K}^{-1}$. We construct “seismic phase diagrams” of the transition zone that can be directly compared with experimental phase diagrams. We also apply a “Z- Γ ” transform to better constrain the Clapeyron slopes

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γ of the minor phase changes. Although tenuous, signals in seismic phase diagrams suggest that minor phase transitions, both in the olivine and the non-olivine component of the mantle, have visible seismic expressions. They can tentatively be described as follows. The ‘410’ is overlaid at low temperature by an interface corresponding to a decrease of velocity with depth and $\gamma \approx +3 \text{ MPa K}^{-1}$. The ‘660’ widens at high temperature and is preceded at low temperature by an interface, the ‘620’, with $\gamma \approx +7 \text{ MPa K}^{-1}$. A ‘520’ is suggested with $\gamma \approx 2\text{-}3 \text{ MPa K}^{-1}$. These last two interfaces correspond to velocity increases with depth. At last, near 590 km depth, an interface may be associated with a velocity reduction showing a weak dependence to temperature ($\gamma \sim 0 \text{ MPa K}^{-1}$).

Keywords: Seismic discontinuities, receiver functions, mantle transition zone, phase transitions, Clapeyron slopes

1. Introduction

In the mantle, two sharp seismic discontinuities bounding the transition zone (TZ) are usually attributed to pressure-induced solid phase changes of the olivine $(\text{Mg,Fe})_2\text{SiO}_4$ mineral. These reactions involve the transitions from olivine to wadsleyite (ol $\rightarrow$ wd) at 410 km depth (the ‘410’), and ringwoodite to perovskite+ferropericlasite (rw $\rightarrow$ pv+fp) at 660 km depth (the ‘660’). Seismological observations of the depths of discontinuities can be used to probe the mantle pressure-temperature (P, T) conditions. This requires the knowledge of their Clapeyron slopes $\gamma = dP/dT$, *i.e.* the thermal dependence of the pressure at which the phase changes occur.

Laboratory experiments agree on the opposite signs of the two Clapey-

ron slopes. The slope γ_4 of the ol $\rightarrow$ wd transformation at 410 km depth has been measured in a +1.5 to +4 MPa K $^{-1}$ range (Akaogi et al., 1989; Katsura et al., 2004, see Supplement table S.1). Reports of γ_6 for the endothermic rw $\rightarrow$ pv+fp reaction at 660 km depth are even more scattered with values ranging from -4 MPa K $^{-1}$ to -0.2 MPa K $^{-1}$ (Ito et al., 1990; Litasov et al., 2005b). The reaction at 410 km depth should occur at higher pressure (*i.e.*, greater depth) in a hotter mantle while the ‘660’ should occur at lower pressure (*i.e.*, lower depth) under the same condition, leading to anti-correlated topographies.

A long standing debate exists among seismologists on the degree of anti-correlation of the topographies of the ‘410’ and ‘660’ discontinuities. The main source of information comes from converted/reflected body-wave imaging (see Shearer, 2000, for a review). The topography of the ‘410’ has been observed smaller than that of the ‘660’ (Shearer, 1991; Helffrich, 2000). The unique seismological study to provide direct constraints on olivine Clapeyron slopes (Lebedev et al., 2002) found $\gamma_4 = +2$ MPa K $^{-1}$ and $\gamma_6 = -3.3$ MPa K $^{-1}$ below east Asia and Australia. In some cases, seismological observations are even in disagreement with the experimental expectation of anti-correlated topographies for the ‘410’ and the ‘660’ discontinuities. Gu et al. (1998) and Houser et al. (2008) found a slight positive global correlation between them. This observation is further evidenced below hotspots (Deuss, 2007; Tauzin et al., 2008) or in the Pacific (Houser and Quentin, 2010) where the apparent $\gamma_4/\gamma_6 > 0$ has been attributed to a transition in the non-olivine component of the mantle, from majorite-garnet to perovskite in MgSiO $_3$, occurring below the ‘660’ with a positive Clapeyron slope (Hirose, 2002).

37 However uncertainty remains on the reliability of the corrections for up-
 38 permost velocity structure. Interface mapping relies on the analysis of travel-
 39 times of body waves converted or reflected at discontinuities. These travel-
 40 times not only depend on the depth of conversion/reflection but also on the
 41 velocity heterogeneities encountered along the ray paths above the discon-
 42 tinuities. This results in a trade-off between the apparent depths of dis-
 43 continuities and the velocity heterogeneities above the TZ. As depicted in
 44 Fig. 1a, shallow velocity anomalies negatively correlated with the temper-
 45 ature in the transition zone apparently enhance the ‘410’ topography and
 46 reduce the ‘660’ topography (*e.g.* the case of a hot lithosphere on top of a
 47 hot mantle, Fig. 1a, left, or reciprocally, right). Shallow velocity anomalies
 48 positively correlated with the temperature in the transition zone could lead
 49 to an overestimate of the ‘660’ topography compared to that of the ‘410’ (*e.g.*
 50 a thin crust on top of a hotter mantle, Fig. 1b left, or reciprocally, right).
 51 In general, the absence or the inaccuracy of velocity corrections obscure the
 52 anti-correlation between the discontinuity depths (*e.g.* Stammer and Kind,
 53 1992). In extreme cases where the contribution from shallow velocity het-
 54 erogeneities exceeds the contribution from the temperature (*i.e.*, all 4 cases
 55 in Fig. 1), the topographies appear correlated and the apparent Clapeyron
 56 slope ratio γ_4/γ_6 becomes positive, contrary to experimental expectations.
 57 This could be especially the case for up-going P waves converted into shear
 58 waves under seismological stations (P-to-S conversions) whose travel-times
 59 are strongly velocity-dependent (Li et al., 2003).

60 In this study, we propose a method to separate the competing effects
 61 of uppermost velocity structure and temperature on the topography of TZ

Figure 1: Behavior of discontinuities with thermal anomalies in the transition zone and strong velocity heterogeneities in the lithosphere. Topographies of opposite signs expected from mineral physics are shown with plain lines whereas seismological observations, influenced by the uppermost velocity structure, are shown with dashed lines. In the absence of shallow corrections, an apparent correlation between the interface topographies may be observed. (a) Case of a slower lithosphere on top of a hotter mantle (left) and reciprocally (right). (b) Case of a faster lithosphere (*e.g.* due to a thin crust) on top of a hotter mantle (left) and reciprocally (right).

discontinuities. This method makes possible the determination of the ratio γ_4/γ_6 with little information on the background velocity structure. From this estimate, we can correct the structure of the TZ from the effect of velocities, estimate the temperature, and build diagrams that may be used as a proxy for thermodynamics phase diagram of solid-solid transitions. Then we develop and apply a new method, the “Z- Γ stacking”, to test the presence of unknown phase transitions in the mantle and measure their relative Clapeyron slopes.

2. Seismic observations under Western US

We make use of a dataset of P-to-S conversions obtained from the Transportable Array component of USArray. This dataset is described in detail in Tauzin et al. (2013). Using receiver functions, the mantle structure of the western half of the US (Fig. 2a) has been characterized in the 5-75 s period range. Seismic images of mantle discontinuities have been produced by a migration method, stacking the receiver functions by common conversion point (CCP) (Dueker and Sheehan, 1997; Wittlinger et al., 2004). The imaged volume is thus called a “CCP volume”. By picking in this volume the seismic signal of conversions at the ‘410’ and ‘660’, high resolution maps of their apparent topographies (δZ_4 and δZ_6) have been obtained on a 0.5° grid in latitude and longitude (Fig. 2b,c). The observed average depths of the ‘410’ and ‘660’ over the area are $\langle Z_4 \rangle = 424$ km and $\langle Z_6 \rangle = 676$ km. The corresponding peak-to-peak depth variations are 55 km for the ‘410’ and 49 km for the ‘660’ (with RMS 9.8 km and 8.8 km, respectively). The migration being performed using a spherical model, IASP91 (Kennett and Engdahl, 1991), the topography maps do not take into account lateral variations of

Figure 2: (a) Surface topography with state borders. (b) Apparent topography δZ_4 of the ‘410’ seismic discontinuity. (c) Apparent topography δZ_6 of the ‘660’ discontinuity (both panels from Tauzin et al., 2013).

86 velocities within the mantle. We thus expect shallow velocity heterogeneities
 87 to significantly affect the apparent topography of TZ discontinuities.

88 Fig. 3 shows that this is indeed the case. Contrary to mineral physics
 89 expectations, the ‘410’ and the ‘660’ topographies have a positive correlation
 90 of +0.60 (Fig. 3a). This correlation is partly the result of shallow velocity
 91 heterogeneities in the mantle overlying the TZ. The depth-velocity trade-off
 92 is obvious by looking at average δv_p anomalies above the TZ taken from
 93 the tomographic model of Burdick et al. (2010), versus the ‘410’ topography

in Fig. 3b. The correlation is negative, -0.68, as a slower lithosphere tends to increase the apparent depths of the discontinuities (see Fig. 1a left, and Fig. 1b, right). This trade-off reduces slightly to -0.45 when examining the ‘660’ topography (Fig. 3c). Similar observations from SS-precursors have already been discussed by Houser et al. (2008).

3. Clapeyron slopes of the ‘410’ and the ‘660’

3.1. Seismic imaging of discontinuities

We assume that at first order, the “depth variation” δZ of a phase transition depends on its Clapeyron slope, γ and on the temperature variations δT (i.e. the temperature minus the adiabatic reference geotherm). The potential effects of compositional heterogeneities (e.g., water content or variability in Mg/Fe ratios, Litasov et al. (2005a); Irifune et al. (1998)) are considered as second order effects. Depth and temperature are therefore related by

$$\delta Z = Z - \langle Z \rangle = \frac{\gamma}{\rho g} \delta T = \Gamma \delta T \quad (1)$$

where g is Earth’s gravity, ρ the density above the discontinuity and Γ the Clapeyron slope in m K^{-1} (while γ is in Pa K^{-1} , see Helffrich and Bina (1994)). The absolute depth of discontinuity Z and its average $\langle Z \rangle$, can be directly estimated from seismic imaging using for instance P-to-S conversions (Tauzin et al., 2013) or SS-precursors (Houser et al., 2008). In the case of our P-to-S observations, the travel times of converted waves Pds relative to the direct P depend on the conversion depth Z , the ray parameter p (assumed to be identical for the two waves), the S and P velocities v_s , v_p :

$$t_{Pds}(p, Z) = \int_0^Z [\sqrt{v_s^{-2} - p^2} - \sqrt{v_p^{-2} - p^2}] dz \quad (2)$$

Figure 3: (a) Observed ‘660’ depth Z_6 versus the ‘410’ depth Z_4 . The oblique dashed line represents perfectly correlated topographies. (b) Average tomographic velocity anomalies δv_p above the ‘410’ versus the ‘410’ topography δZ_4 . (c) Average tomographic velocity anomalies δv_p above the ‘660’ versus the ‘660’ topography δZ_6 . The tomographic model is from Burdick et al. (2010). Correlation coefficients are indicated at the top. The two topographies appear positively correlated contrary to mineralogical expectations. Their negative correlations with the shallow velocity suggest a large effect of the shallow structure on the apparent depths of discontinuities.

115 Supposing v_p and v_s perfectly known, this equation enables the projection of
 116 the amplitude measured at time t_{Pds} on a seismogram at the “true” depth
 117 Z (this is called time-to-depth conversion). A more sophisticated but simi-
 118 lar approach for interface imaging is to backpropagate the seismic wavefield
 119 recorded at the surface, with the expectation that the transposed wavefield
 120 will focus at and therefore will highlight the sources of seismic scattering in
 121 the subsurface (this is called migration; *e.g.*, Rondenay, 2009).

122 However in these methods, the imperfect knowledge of the shallow veloc-
 123 ity structure leads to an imperfect recovery of the interface positions. Instead,
 124 one recovers an apparent topography:

$$\delta Z = Z - \langle Z \rangle = \Gamma \delta T + \delta h \quad (3)$$

125 The correction term δh integrates the effects on travel-times of 3D hetero-
 126 geneities $\delta v_{s,p}$ ($\delta v = v - v_{ref}$) encountered along the ray paths and not taken
 127 into account by the reference model. Instead of (2) what is really used is:

$$t_{Pds}(p, Z) = \int_0^{Z+\delta h} [\sqrt{v_{sref}^{-2} - p^2} - \sqrt{v_{pref}^{-2} - p^2}] dz \quad (4)$$

128 where v_{sref} and v_{pref} are velocities in the Earth’s reference model used for
 129 imaging. The deviations of velocities $\delta v_{s,p}$ from the reference model are thus
 130 mapped into δh . Using (2), the latter identity can be used to express δh ,
 131 assuming a vertical incidence ($p = 0$):

$$\delta h = - \frac{\int_0^Z [\delta v_s / v_{sref}^2 - \delta v_p / v_{pref}^2] dz}{[1/v_{sref}(Z) - 1/v_{pref}(Z)]} \quad (5)$$

132 (practically, the P and S ray paths are not vertical so (5) is an approxima-
 133 tion). In general δv_s and δv_p are correlated, and $|\delta v_s / v_{sref}^2| > |\delta v_p / v_{pref}^2|$,

so that δh is anti-correlated with the seismic velocities above the TZ, *i.e.* deeper interfaces $\delta h > 0$ are inferred for a slower lithosphere $\delta v_{s,p} < 0$ (see Fig. 1 and Fig. 3b,c).

The effect of δh can in principle be estimated from tomographic imaging. However, the travel-time tomography is often resolved with longer wavelengths than what would be needed to accurately correct the local information associated with converted/reflected body waves, especially at global scale.

To alleviate the problem of mapping velocity variations into interface topographies, various authors have discussed their results in term of TZ thickness (*e.g.*, Chevrot et al., 1999; Lawrence and Shearer, 2006; Tauzin et al., 2008). Computing a difference of topographies cancels the major corrections due to the shallow structure but leaves those due to the velocity perturbations in between the two interfaces. For example, because of local temperature variations, any possible interface within the TZ, appears shifted with respect to the ‘410’, by the distance, see (5),

$$\delta h_T(Z) = a(Z) \delta T = -K \frac{\int_{Z_4}^Z [1/v_{s_{ref}} - 1/(Rv_{p_{ref}})] dz}{[1/v_{s_{ref}}(Z) - 1/v_{p_{ref}}(Z)]} \delta T \quad (6)$$

where the proportionality factor computed at the ‘660’ is $a(< Z_6 >) = a_6 = 0.03 \text{ km K}^{-1}$ (using $K = \partial \log v_s / \partial T = -7 \cdot 10^{-5} \text{ K}^{-1}$ (Stixrude and Lithgow-Bertelloni, 2005) and $R = \partial \log v_s / \partial \log v_p = 2$ (*e.g.*, Schmandt and Humphreys, 2010; Becker, 2011)). The contributions of the lateral temperature variations to the apparent thickness of the TZ are therefore non negligible (3 km for 100 K) and a hotter TZ appears thicker without corrections.

156 *3.2. Fundamental hypotheses*

157 For the ‘410’ and the ‘660’ discontinuities, we suppose that the apparent
158 topographies are given by:

$$\begin{aligned}\delta Z_4 &= \Gamma_4 \delta T + \delta h \\ \delta Z_6 &= \Gamma_6 \delta T + \delta h + \delta h_T(< Z_6 >) = (\Gamma_6 + a_6) \delta T + \delta h = \Gamma'_6 \delta T + \delta h\end{aligned}\tag{7}$$

159 In the last equality we group together the two δT -dependent terms and define
160 an apparent Clapeyron slope $\Gamma'_6 = \Gamma_6 + a_6$. Equations (7) assume that the
161 temperature δT in the TZ does not vary with depth but uniquely laterally.
162 We will discuss this approximation in section 3.5.

163 The expectation from high pressure mineral physics is that $\Gamma_6/\Gamma_4 < 0$.
164 This effect alone imposes a negative correlation between δZ_4 and δZ_6 (see
165 (7)). On the contrary, the contribution of δh is opposite (see (7)). As a_6
166 and Γ_4 have the same sign, the temperature variations within the TZ also
167 tends to increase the correlation between the interfaces. The slight positive
168 correlation observed between the ‘410’ and ‘660’ topographies at a global
169 scale (Gu et al., 1998; Houser et al., 2008) could thus be partly explained by
170 an under-estimation of the velocity corrections.

171 *3.3. The Clapeyron slope ratio and temperature in the TZ*

172 Writing $\Gamma_4 \delta Z_6 - \Gamma'_6 \delta Z_4$ (see (7)), we get the contribution from the
173 shallow heterogeneities as a function of an apparent Clapeyron slope ratio
174 (Γ'_6/Γ_4)

$$\delta h = \frac{1}{1 - \Gamma'_6/\Gamma_4} \delta Z_6 - \frac{1}{\Gamma_4/\Gamma'_6 - 1} \delta Z_4\tag{8}$$

175 Our method considers the contribution δh as a noise to the real signal,
176 that must be minimized, *i.e.*, we choose the Clapeyron ratio Γ'_6/Γ_4 that

177 explains the ensemble of observations in terms of temperature variations
 178 and shallow corrections, but requires the smallest shallow corrections. We
 179 therefore look for the ratio Γ'_6/Γ_4 that minimizes the quantity:

$$\chi^2 = \langle \delta h^2 \rangle \quad (9)$$

180 We can show that an analytical solution to $\partial\chi/\partial(\Gamma'_6/\Gamma_4) = 0$ is:

$$\frac{\Gamma'_6}{\Gamma_4} = \frac{\langle \delta Z_6 \cdot (\delta Z_6 - \delta Z_4) \rangle}{\langle \delta Z_4 \cdot (\delta Z_6 - \delta Z_4) \rangle} \quad (10)$$

181 The expression (10) can also be understood from a different standpoint.
 182 The shallow corrections and the temperature in the transition zone are likely
 183 uncorrelated on average. The shallow corrections are mostly due to crustal
 184 anomalies and fossil structure of the lithosphere. Their correlation with the
 185 thermal TZ structure is very low in global tomography models (see *e.g.*,
 186 Becker and Boschi, 2002). We may therefore assume that the correlation
 187 between δh and δT is zero, *i.e.* $\langle \delta T \cdot \delta h \rangle = 0$. We discuss further the
 188 validity of this assumption in the section 3.5. In this case, one has simply
 189 from (7)

$$\begin{aligned} \langle \delta Z_4 \cdot \delta Z_4 \rangle &= \Gamma_4^2 \langle \delta T \cdot \delta T \rangle + \langle \delta h \cdot \delta h \rangle \\ \langle \delta Z_6 \cdot \delta Z_6 \rangle &= \Gamma_6'^2 \langle \delta T \cdot \delta T \rangle + \langle \delta h \cdot \delta h \rangle \\ \langle \delta Z_6 \cdot \delta Z_4 \rangle &= \Gamma_4 \Gamma_6' \langle \delta T \cdot \delta T \rangle + \langle \delta h \cdot \delta h \rangle \end{aligned} \quad (11)$$

190 Subtracting the last equation to the first two and removing the term $\langle$
 191 $\delta T \cdot \delta T \rangle$ by division, leads again to the relation (10). In conclusion, choosing
 192 the Clapeyron slope ratio according to (10) simultaneously minimizes the
 193 overall shallow corrections and their correlations with the TZ structure.

Figure 4: The minimization of the shallow corrections (see (9)) occurs for $\Gamma'_6/\Gamma_4 = -0.56$.

194 The minimization process allows us to compute the ratio of the apparent
 195 Clapeyron slopes, Γ'_6/Γ_4 but not the two mineralogical Clapeyron slopes γ_6
 196 and γ_4 , independently. If one of the Clapeyron slope is known, *e.g.* at the
 197 ‘410’, then the Clapeyron slope of the ‘660’ is obtained from

$$\gamma_6 = \frac{\rho_6}{\rho_4} \frac{\Gamma'_6}{\Gamma_4} \gamma_4 - \rho_6 g a_6 \quad (12)$$

198 where ρ_4 and ρ_6 are the reference densities above the ‘410’ and ‘660’.

199 From $\delta Z_4 - \delta Z_6$ (see (7)), we obtain the TZ temperature as:

$$\delta T = \frac{1}{\Gamma_4} \frac{1}{(1 - \Gamma'_6/\Gamma_4)} (\delta Z_4 - \delta Z_6) \quad (13)$$

200 The temperature field within the TZ is thus linearly related to the vari-
 201 ations in TZ thickness. The thermal contributions to the ‘410’ topography,
 202 i.e., $\Gamma_4 \delta T$, is obtained by our method but we must assume the independent
 203 knowledge of one of the Clapeyron slope (*e.g.* Γ_4) to obtain δT .

204 3.4. Application to western US

205 In the following, we use our observations of apparent topographies of TZ
 206 discontinuities below western US to estimate the ratio of the ‘410’ and ‘660’

207 Clapeyron slopes. We plot in Fig. 4 the root mean square function $\chi(\Gamma'_6/\Gamma_4)$
 208 (9) which goes through a minimum at $(\Gamma'_6/\Gamma_4) = -0.56$, given, as expected,
 209 by (10). We find no minima for positive values of (Γ'_6/Γ_4) . According to (12),
 210 we therefore predict a relation between the mineralogical Clapeyron slopes

$$\gamma_6 = -0.64 \gamma_4 - 1.17 \quad (14)$$

211 where we used $a_6 = 0.030 \text{ km K}^{-1}$, $\rho_4 = 3542 \text{ kg m}^{-3}$, and $\rho_6 = 3991 \text{ kg m}^{-3}$.
 212 For $\gamma_4 = +3.0 \text{ MPa K}^{-1}$, we obtain $\gamma_6 = -3.1 \text{ MPa K}^{-1}$ and $\gamma_4/\gamma_6 = -1.03$.

213 We show in Fig. 5 the result of applying (13) with $\Gamma'_6/\Gamma_4 = -0.56$ and
 214 taking $\gamma_4 = +3.0 \text{ MPa K}^{-1}$, close to the average of experimental values in
 215 table S.1. As expected from equation (13), the pattern of thermal anomalies
 216 is proportional to the TZ thickness (Tauzin et al., 2013). The choice of γ_4
 217 determines the range of thermal variations δT , from -259 K below northeast
 218 Oregon to +216 K below Yellowstone, with a 65 K RMS (Fig. 5). This
 219 temperature is weakly but positively correlated with what can be estimated
 220 from tomography in the TZ (correlation of +0.23 with Burdick et al. (2010)'s
 221 model at mid TZ).

222 In Fig. 6a, we plot a map of the depth corrections δh computed from
 223 (8) and $\Gamma'_6/\Gamma_4 = -0.56$. These corrections should represent the contribution
 224 of shallow velocity heterogeneities to the interface topographies. A clear
 225 NE-SW dichotomy is observed with δh values ranging from -25 km below
 226 northern Idaho to +17 km in the southwest of Utah, with a RMS of 8.5 km.
 227 This dichotomy dominates the initial maps of topographies (Fig. 2b,c).

228 The corrections δh can be compared with those, δh_{tomo} computed with the
 229 P-wave model of Burdick et al. (2010) (Fig. 6b). To estimate δh_{tomo} , a shear-
 230 wave model δv_s is required (5). We scale δv_s from δv_p using $\delta \log v_s / \delta \log v_p =$

Figure 5: Map of temperature anomalies δT within the TZ obtained from (13) and $\gamma_4 = +3.0 \text{ MPa K}^{-1}$.

231 2 (*e.g.*, Romanowicz and Cara, 1980; Schmandt and Humphreys, 2010). This
 232 value of $\delta \log v_s / \delta \log v_p$ has no impact on the high correlation between δh and
 233 δh_{tomo} , +0.65. The amplitudes of δh_{tomo} are generally lower than predicted
 234 with our method (Fig. 6a) but the NE-SW dichotomy is well recovered. This
 235 confirms the dominant effect of shallow velocities on the apparent topogra-
 236 phies of TZ discontinuities.

237 The corrections δh and $\delta h + a_6 \delta T$ must be removed from the observed to-
 238 pographies δZ_4 and δZ_6 to obtain the “true” topographies (7), perfectly anti-
 239 correlated by construction and due to temperature variations alone (Fig. 6c
 240 and d). The estimated Clapeyron slope ratio γ_6 / γ_4 being ~ -1.0 , the relative
 241 amplitudes of thermal topographies are similar. In term of RMS amplitudes
 242 the initial topographies for the ‘410’ and ‘660’ are of 9.8 and 8.8 km, from
 243 which we removed a shallow correction related to δh of 8.5 km, a temperature
 244 correction at ‘660’ related to $a_6 \delta T$ of 1.9 km, leaving thermal topographies
 245 of 5.3 and 4.9 km respectively.

Figure 6: (a) Map of the depth corrections δh obtained with the apparent Clapeyron slope ratio $(\Gamma'_6/\Gamma_4) = -0.56$. (b) Map of the corrections δh_{tomo} (5) computed with the δv_p model of Burdick et al. (2010) and $R = \partial \ln v_p / \partial \ln v_s = 2$. (c) Map of the '410' thermal topography $\Gamma_4 \delta T$. (d) Map of the '660' thermal topography $\Gamma_6 \delta T$. The thermal topographies have been computed with $\gamma_4 = +3.0 \text{ MPa K}^{-1}$.

246 3.5. Critical assessment

247 Our method makes use of two hypotheses: a low correlation between the
 248 shallow corrections and the temperature in the TZ and a TZ temperature
 249 varying only laterally. These hypotheses allow us to correct the direct obser-
 250 vations of interface topographies without using a tomographic model.

251 Our assumption of a weak correlation between the shallow correction and
 252 the temperature in the TZ is in relative agreement with Burdick et al. (2010)’s
 253 model where the shallowest 400 km (a proxy for shallow corrections) are
 254 weakly correlated (-0.17) with the tomographic anomalies at 500 km (a proxy
 255 for the TZ structure). In Supplement S.1, we discuss how this correlation
 256 $C(\delta h, \delta T)$ may affect our results. When $C(\delta h, \delta T)$ varies from -0.2 to +0.2,
 257 we show that the ratio Γ'_6/Γ_4 derived from the data changes from -0.19 to
 258 -1.23. The assumption of an uncorrelation between shallow corrections and
 259 the TZ temperature is therefore crucial. Using the correlation value suggested
 260 by Burdick’s mantle model would change our estimate for Γ'_6/Γ_4 from -0.56 to
 261 -0.23, or using (14), our estimate of γ_6 from -3.1 to -2.0 MPa K⁻¹. Note that
 262 crustal heterogeneities, which are not explicitly given in the mantle model of
 263 Burdick et al. (2010) should rather decrease than increase $|C(\delta h, \delta T)|$. With
 264 $C(\delta h, \delta T) = -0.17$ the RMS thermal topography at ‘410’ slightly increases
 265 from 5.3 to 6.2 km, that at ‘660’ decreases from 4.9 to 4.3 km, and the RMS
 266 of the corrections δh does not change much with $C(\delta h, \delta T)$.

267 In Supplement S.2, we also discuss the assumption of a uniform tem-
 268 perature in the TZ. We show that as soon as the RMS of the difference of
 269 temperature between the two interfaces is lower than that of the lateral vari-
 270 ations, the retrieved Γ'_6/Γ_4 is not significantly affected. The fact that the

vertical difference of temperature in the 250 km thick TZ is not larger than the horizontal temperature differences below western US seems reasonable.

Therefore, the accuracy of our method is mostly affected by the assumption of an uncorrelation between surface corrections and the TZ temperature and not much by the sampling of our dataset (we checked by bootstrap resampling that similar results are obtained by using only subsets of our data) or by the assumption of a vertically uniform TZ. Of course, if tomographic models were extremely precise they could be used to correct the observations or to check our hypotheses. Unfortunately, we consider that this appears difficult even with high quality models like that of Burdick et al. (2010).

First, tomographic inversions involve smoothing and regularization that limit their accuracy at the short length-scale at which interface topographies are mapped by receiver functions. Second, velocity variations close to an interface, can be interpreted as due to temperature variations in a medium with flat interfaces or as due to interface undulations in a medium without other velocity anomalies. In the first case we would assume that $\delta v_p/v_p = (\partial \log v_p / \partial T) \delta T \approx -3.5 \cdot 10^{-5} \delta T$. In the second case, an interface topography $\delta Z = \Gamma \delta T$ seen with a resolution λ would appear as an anomaly $\delta v_p/v_p = -(\Delta v_p/v_p)(\Gamma/\lambda) \delta T$ where $\Delta v_p/v_p$ is the velocity jump across the interface ($\Delta v_p/v_p \approx 4\%$ and 6% at the ‘410’ and ‘660’). We would obtain $\delta v_p/v_p \approx -5 \cdot 10^{-5} \delta T$ near the ‘410’ and $\delta v_p/v_p \approx +10 \cdot 10^{-5} \delta T$ near the ‘660’ for a resolution length of $\lambda = 40$ km. Of course a combination of the two effects is likely. The above numerical applications imply that we may not even know the sign of the temperature anomaly near the ‘660’ and that accurate temperature estimate at ‘410’ is unlikely. It seems to us that unless

tomographic models are obtained by simultaneous inversion of both velocities and interfaces (like *e.g.* in Gu et al., 2003; Houser et al., 2008; Lawrence and Shearer, 2008, but with a better regional resolution), the estimate of TZ temperatures from tomographic models with flat interfaces will introduce more, or at least as much uncertainties as our simpler approach.

4. Phase changes in the transition zone

4.1. Seismically deduced phase diagrams

Now that we have estimated the velocity corrections and the TZ temperature we can directly use the waveforms to construct a representation of the thermodynamic phase diagram seen from seismology and to search for other potential phase changes.

In Fig. 7a, we stack all the seismic signals $s_i(Z)$ within intervals of apparent depths δZ_4 of the ‘410’. The ‘410’ and the ‘660’ appear as well-resolved streaks. The slopes of δZ_4 and δZ_6 , both positive in this diagram, are controlled by shallow heterogeneities δh (see (7)). Their slight difference is due to the weak expression of temperature.

A more meaningful observation of the interfaces is obtained by correcting the traces from the shallow heterogeneities, hence by considering $s_i(Z - \delta h_i)$ where δh_i is given by (8) and (10). This is done in Fig. 7b, here as a function of the thermal topography of the ‘410’, $\Gamma_4 \delta T$. The effect of the corrections is drastic on the ‘660’, straightening it up to an opposite direction to that of the ‘410’.

All the interfaces in the TZ are also displaced by $a(Z) \delta T$ (see (6)). The best representation of the effects of the temperature on phase changes is

Figure 7: (a) Diagram of the seismic signals $s_i(Z)$ stacked within 2 km intervals of the apparent depth of the ‘410’, δZ_4 . Positive amplitudes (in red) mark velocity increases with depth whereas negative amplitudes (in blue) are associated with velocity reductions. The black line is the relation $Z = \langle Z_4 \rangle + \delta Z_4$. (b) Diagram of $s_i(Z - \delta h_i)$, corrected from shallow heterogeneities. The black line is the relation $Z = \langle Z_4 \rangle + \delta Z_4 - \delta h = \langle Z_4 \rangle + \Gamma_4 \delta T$. (c) Seismic phase diagram $F(Z, \delta T)$ (using $\gamma_4 = +3 \text{ MPa K}^{-1}$ and stacking the seismic signal by intervals of 20 K). (d) Our interpretation. ol: olivine; wd: wadsleyite; rw: ringwoodite; fp: ferropericlase; pv: perovskite; mj: majorite; ak: akimotoite. LVL: Low velocity layer atop the ‘410’. The phase diagram for transformations at the bottom of the transition zone is taken from Hirose (2002) for pyrolite. The dashed ellipse locates the weak signal from a possible wd→rw transition and the dotted ellipse the signal of the negative ‘590’.

320 obtained by plotting the function $F(Z, \delta T)$ (see Fig 7c) obtained by stacking
 321 $s_i(Z - h_i - a(Z) \delta T_i)$ as a function of δT_i

$$F(Z, \delta T) = \langle s_i(Z - h_i - a(Z) \delta T_i, \delta T_i) \rangle \text{ for } \delta T_i \approx \delta T. \quad (15)$$

322 The function $F(Z, \delta T)$ represents our best estimate of the position of the
 323 interfaces as a function of temperature, what we call a “seismic phase dia-
 324 gram”. Note that the construction of this diagram requires the choice of the
 325 ‘410’ Clapeyron slope to properly scale δT . Here we choose $\gamma_4 = +3.0$ MPa
 326 K^{-1} . In this “seismic phase diagram” the phase changes at ‘410’ and ‘660’
 327 are straight lines with slopes Γ_4 and Γ_6 . Any other phase transformation ap-
 328 pears as a line of slope Γ . The ol→wd and rw→fp+pv are clearly visible but
 329 the wd→rw transition is very faint (encircled by a dashed ellipse in Fig 7d).
 330 Other signals in the diagram can either be associated with phase transitions
 331 if they align on linear trends, or with “noise” if they do not, due for instance
 332 to compositional effects.

333 This approach can also be used to correct seismic sections for the mantle
 334 (see Fig. S.2). As an example, the initial section for a profile at 41°N
 335 is corrected from the shallow heterogeneity and the temperature induced
 336 velocity changes and leads to a profile where the signal should be only due
 337 to the true interface undulations.

338 Tauzin et al. (2013) reported the presence of seismic discontinuities around
 339 350, 590 and 620 km depths in the mantle below western US. These features
 340 are more local than the ‘410’ and ‘660’ but appear anyway in the diagram
 341 of Fig 7c. For cold regions of the mantle (< -150 K of thermal anomaly), a
 342 negative signal is observed atop the ‘410’ around 380 km depth (the ‘350’).
 343 This same temperature domain is characterized by multiple positive peaks

344 above the ‘660’ and by what we can associate with a positive ‘620’ disconti-
 345 nuity. The negative signal of the ‘590’ is diffuse: it can be found across the
 346 whole temperature range with higher amplitudes in the hottest environments
 347 (dotted contour in Fig 7d). We checked by making seismic phase diagrams
 348 restricted to the eastern or western regions relative to the Nevada/Utah bor-
 349 ders, that the characteristics of our observations in Tauzin et al. (2013) are
 350 preserved. We observe the ‘350’ in both eastern and western regions with
 351 stronger amplitudes in the west, the negative ‘590’ mostly in the east, and
 352 the positive ‘620’ in both areas. This confirms that these minor interfaces
 353 are robust in their seismological observation.

354 The minor interfaces seem also consistent with our model of phase changes
 355 and vertically coherent temperature in the TZ. They are not noise or ran-
 356 dom scatterers because they are seen coherently in quite specific domains of
 357 temperature in the diagram. Indeed, they do not appear as coherent as we
 358 stack the data as a function of the apparent ‘410’ topography (see Fig 7a).
 359 Their origin is thus likely related to the temperature in the TZ.

360 4.2. *Z- Γ stacking for the Clapeyron slope of unknown phase transitions*

361 This seismic phase diagram suggests a simple way to interrogate the data
 362 for the presence of phase transitions in the mantle. We simply stack all
 363 the observations, corrected from the velocity perturbations, along the lines
 364 of slopes Γ . This method is akin to the slant stack τ - p method (Rost and
 365 Thomas, 2002) or the more general Radon transform (Gu and Sacchi, 2009)
 366 that sum up signals, not at constant depth (or time) but along a specific

367 integration path. We thus build a Z - Γ diagram $G(Z, \Gamma)$,

$$G(Z, \Gamma) = \int F(Z - \Gamma \delta T, \delta T) dT \quad (16)$$

368 This function (see Fig 8) presents two maxima at $Z = 424$ km and $\Gamma =$
 369 $+0.081$ km K⁻¹ for the ‘410’ and at $Z = 677$ km and $\Gamma = -0.079$ km K⁻¹
 370 for the ‘660’. Their Clapeyron slope ratio is close to the estimate of section
 371 3.4 (here we compute the maximum of an average signal while in section 3.4
 372 we compute the average of the positions of maxima: this explains the small
 373 difference). As reported in Tauzin et al. (2013) the amplitudes of the signals
 374 converted at the ‘410’ and ‘660’ are comparable ($\sim 1\%$ of the P signal) while
 375 radial models (PREM or IASP91) would predict a 50% larger signal for the
 376 conversion at the ‘660’ which suggests that radial models overestimate the
 377 velocity jump at the ‘660’. Other maxima in the $G(Z, \Gamma)$ diagram potentially
 378 indicate the presence of additional phase transitions in the mantle near the
 379 depth Z and with a Clapeyron slope Γ .

380 The additional minor seismic discontinuities reported in Tauzin et al.
 381 (2013) and in Fig 7c are also present as weak local extrema in the Z - Γ domain
 382 (small stars). The ‘350’ is recovered at 387 km depth with an apparent
 383 Clapeyron slope $\Gamma = +0.092$ km K⁻¹ and the ‘590’ at 573 km with $\Gamma =$
 384 -0.010 km K⁻¹. The two discontinuities have absolute amplitudes within 20
 385 to 30% of the ‘410’ or ‘660’ amplitudes (*i.e.* $\sim 0.3\%$ the P-wave amplitude).
 386 A peak at $Z = 530$ km and $\Gamma = +0.064$ km K⁻¹ could correspond to the
 387 wd $\rightarrow$ rw transition but the signal is weak with an amplitude of only 0.1%
 388 the P-wave (*i.e.* 14% the conversion at the ‘660’). The ‘620’ appearing
 389 as a conspicuous feature in Fig 7c is not reliably constrained with our Z - Γ
 390 stacking approach because it is close to the strong signal from the ‘660’.

Figure 8: Z - Γ diagram. The amplitudes are red (positive) for velocity increases and blue (negative) for velocity reductions. The stars correspond to the couples $(\langle Z \rangle, \Gamma)$ expected for the ‘410’ and ‘660’ and inferred for minor extrema that are found around 350, 520, 590 and 620 km depths.

391 A summary of our results is given in table 1. In the second and third
392 columns Γ and γ assume $\gamma_4 = +3 \text{ MPa K}^{-1}$. The depths reported in the
393 fourth column correspond to the potential depth of the interface in a refer-
394 ence mantle with zero thermal anomalies. For instance the ‘620’ which only
395 appears for $\delta T \approx -200 \text{ K}$ is mentioned at a depth of 656 km in table 1 be-
396 cause of its large Clapeyron slope, $+0.174 \text{ km K}^{-1}$ (‘620’ $\approx 656 - 0.174 \times 200$
397 km). For simplicity we mostly discussed our results assuming $\gamma_4 = +3.0$
398 MPa K^{-1} but as explained in section 3 we can only derive linear relations
399 between Clapeyron slopes. In the last column of table 1 we report these re-
400 lations with respect to γ_4 (the third column is in agreement with the last one
401 when $\gamma_4 = +3 \text{ MPa K}^{-1}$). Ascribing uncertainties to these slopes is difficult.
402 However from the horizontal sizes of the extrema in Fig 8 the uncertainties
403 are probably as large as 0.02 km K^{-1} ($\approx 0.7 \text{ MPa K}^{-1}$).

Table 1: Summary of potential phase transitions observed in the TZ.

Discontinuity	Γ km K ⁻¹	γ MPa K ⁻¹	depth km	amplitude % 410	Possible transition	γ MPa K ⁻¹
'410'	+0.081	+2.8	424	100	ol→wd	γ_4
'660'	-0.079	-3.1	677	110	rw→pv+fp	-0.64 γ_4 -1.17
'350'	+0.092	+3.2	387	-30	LVL	1.0 γ_4 +0.25
'520'	+0.064	+2.5	530	+14	wd→rw	1.0 γ_4 -0.50
'590'	-0.010	-0.4	573	-22	?	0.15 γ_4 -0.83
'620'	+0.174	+6.8	656	+54	mj→ak	2.59 γ_4 -0.98

5. Conclusion

In this study, we used the competing actions of temperature and velocities on the apparent topography of TZ discontinuities to quantify their respective effects neglecting the possible presence of compositional variations. The expected anti-correlation of the '410' and '660' is largely hidden by the effects of shallow velocity anomalies and the RMS thermal topographies (5.3 and 4.9 km) are lower than that of the corrections (8.5 km).

We suggest that the Clapeyron slope ratios of the two major interfaces of

412 the TZ can be obtained by minimizing the shallow corrections or by assuming
 413 the absence of correlation between these corrections and the TZ tempera-
 414 ture. The knowledge of the correlation between the TZ temperature and the
 415 shallow corrections derived from tomography may be used to improve this
 416 estimate (see Supplement S.1) but the exercise is not easy. This correlation
 417 is an important parameter that impacts our ability to measure the exact
 418 Clapeyron slopes of the ‘410’ and ‘660’. By comparison the other sources
 419 of uncertainty, *e.g.*, the lateral averaging in the migration process or the
 420 assumption that the TZ temperature only varies laterally, seem to be less
 421 crucial.

422 The presence of minor seismic discontinuities have been reported below
 423 western US by *e.g.*, Simmons and Gurrola (2000), Song et al. (2004), Vinnik
 424 et al. (2010) Schmandt et al. (2011), and more recently Tauzin et al. (2013).
 425 These interfaces are very weak and more visible at low temperatures. Two
 426 interfaces seem to correspond to a velocity reduction, two interfaces to a
 427 velocity increase. We suggest a new method to construct the phase diagram
 428 directly from the seismic observations and, through a Z - Γ transform, we
 429 estimate the Clapeyron slopes of the potential minor phase changes.

430 The sharp reduction of velocity at 350 km depth can hardly be explained
 431 in term of a transition toward a higher pressure assemblage as it is associated
 432 with a shear-wave velocity drop ($\sim 30\%$ of the conversion amplitude at the
 433 ‘410’). It has been proposed to mark the onset of a low-velocity layer atop
 434 the ‘410’ due to dehydration-induced partial melting (Revenaugh and Sipkin,
 435 1994). We preferentially find this layer in cold environments. It may be
 436 induced by water rather released from subduction than upwelling from the

437 deeper mantle like in the transition zone water filter model of Bercovici and
438 Karato (2003). We find for this discontinuity a Clapeyron slope $\gamma = +3.2$
439 MPa K⁻¹ close to $\gamma_4 = +2.8$ MPa K⁻¹, which suggests that the two interfaces
440 may be intimately related.

441 The weak interface at ‘590’ (Tauzin et al., 2013) does not seem to vary
442 much with temperature, with a small Clapeyron slope of -0.4 MPa K⁻¹. It
443 is also associated with a velocity reduction (-20% of the reflection amplitude
444 at the ‘410’). Its origin remains mysterious.

445 The interfaces at 520 and 620 km depth being associated with increases of
446 velocity can be more easily related to pressure-induced phase changes. The
447 ‘520’ is likely associated with the wadsleyite-ringwoodite transformation. Its
448 Clapeyron slope is negative and in agreement with laboratory experiments
449 (Suzuki et al., 2000). This signal is however very weak, only 10% of the
450 conversion amplitude at the ‘410’, and its detection is not very reliable. The
451 wd→rw transition is expected to be broader and with a smaller velocity jump
452 than that of the ol→wd transition so is less likely visible to converted phases
453 (Bina, 2003).

454 The ‘620’ discontinuity is found above the ‘660’ in the coldest part of
455 the seismic phase diagram. Due to its close proximity to the ‘660’ signal
456 we cannot reliably constrain its Clapeyron slope though it appears positive
457 and greater than that of the ‘410’ in Figs 7 and 8. Tauzin et al. (2013)
458 tentatively associated this discontinuity with the onset of stability field of
459 akimotoite MgSiO₃ below the triple border between Washington, Oregon and
460 Idaho. The presence of akimotoite owing to cold slabs was also suggested by
461 Schmandt et al. (2012) on the basis of a positive v_s gradient above the ‘660’.

462 The akimotoite is the product of a phase change of majorite in the garnet
463 component occurring for low temperatures at similar pressure ranges as the
464 $\text{rw} \rightarrow \text{pv} + \text{mw}$ transition (Hirose, 2002; Akaogi et al., 2002; Ishii et al., 2011).

465 The phase diagrams that we construct from seismic observations assume
466 that the effects of metastability (see *e.g.*, Kirby et al., 1996) or compositional
467 variations (see *e.g.*, Ricard et al., 2005) are negligible. Water, for instance,
468 would promote the uplift of the ‘410’ and the thickening of the TZ (Wood,
469 1995; Smyth and Jacobsen, 2006) and may bias our temperature estimates.
470 Other factors have also been suggested to add complexity to the temperature
471 response of the TZ. The Clapeyron slopes γ for any phase transition may
472 vary slightly with pressure and temperature. The ‘410’ broadens at low
473 temperature and eventually bifurcates into two different transitions at lowest
474 temperature when olivine transforms to ringwoodite rather than wadsleyite
475 (Bina, 2003). These complexities do not seem to affect our seismic phase
476 diagram where the ‘410’ transition remains simple (albeit preceded by the
477 sharp ‘350’ velocity reduction at low temperature). The ‘660’ seems to be
478 thicker both at low and high temperature. This behavior is in agreement
479 with experimental phase diagrams (Hirose, 2002; Bina, 2003) where in cold
480 situation the akimotoite phase may form before the ringwoodite while at high
481 temperature the garnet phase transforms after the ringwoodite.

482 In conclusion, our method for computing the seismic phase diagrams and
483 estimating the Clapeyron slopes through the “ Z - γ stacking” approach, is
484 easy to implement. It can be applied at a much larger scale, *e.g.*, on global
485 datasets of P-to-S conversions (Tauzin et al., 2008), or on other global seismic
486 data such as precursors of S waves reflected halfway between sources and

487 receivers. We also plan to develop the same approach in different frequency
488 bands which should help to further refine the gradients of velocities in the
489 TZ and our knowledge of the phase transitions in the mantle.

490

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492

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497 **References**

498 Akaogi, M., Ito, E., 1993. Refinement of enthalpy measurement of MgSiO_3
499 perovskite and negative pressure-temperature slopes for perovskite-
500 forming reactions. *Geophys. Res. Lett.* 20, 1839–1842.

501 Akaogi, M., Ito, E., Navrotsky, A., 1989. Olivine-modified spinel-spinel tran-
502 sitions in the system $\text{Mg}_2\text{SiO}_4\text{-Fe}_2\text{SiO}_4$: Calorimetric measurements, ther-
503 mochemical calculation, and geophysical application. *J. Geophys. Res.* 94.

504 Akaogi, M., Ross, N., Mc Millan, P., Navrotsky, A., 1984. The Mg_2SiO_4
505 polymorphs (olivine, modified spinel and spinel). Thermodynamic proper-
506 ties from oxyde melt solution calorimetry, phase relations and models of
507 lattice vibrations. *Am. Mineral.* 69, 499.

508 Akaogi, M., Tanaka, A., Ito, E., 2002. Garnet-ilmenite-perovskite transitions
509 in the system $\text{Mg}_4\text{Si}_4\text{O}_{12}\text{-Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$ at high pressures and high temper-

510 atures: phase equilibria, calorimetry and implications for mantle structure.
 511 Phys. Earth Planet. Inter. 132, 303–324.

512 Akaogi, M., Tanaka, A., Ito, E., 2007. Low-temperature heat capacities,
 513 entropies and enthalpies of Mg_2SiO_4 polymorphs, and $\alpha\beta\gamma$ and post-spinel
 514 phase relations at high pressure. Phys Chem Minerals 34, 169–183.

515 Becker, T., 2011. On recent seismic tomography for the western united states.
 516 Geochem. Geophys. Geosys. 13, 1–11.

517 Becker, T., Boschi, L., 2002. A comparison of tomographic and geodynamic
 518 mantle models. Geochem. Geophys. Geosys. 3, 1–48.

519 Bercovici, D., Karato, S., 2003. Whole mantle convection and transition-zone
 520 water filter. Nature 425, 39–44.

521 Bina, C., 2003. Seismological constraints upon mantle composition. In: Carl-
 522 son, R. (Ed.), Treatise on Geochemistry, vol. 2. Elsevier Science Publishing,
 523 pp. 39–59.

524 Bina, C., Helffrich, G., 1994. Phase transition Clapeyron slopes and transi-
 525 tion zone seismic discontinuity topography. J. Geophys. Res. 99, 15,853–
 526 15,860.

527 Burdick, S., Hilst, R., Vernon, F., Martynov, V., Cox, T., Eakins, J., Karasu,
 528 G., Tylell, J., Astiz, J., Pavlis, G., 2010. Model update january 2010: Up-
 529 per mantle heterogeneity beneath north america from travelttime tomogra-
 530 phy with global and usarray transportable array data. Seis. Res. Lett. 81,
 531 689–693.

- 532 Chevrot, S., Vinnik, L., Montagner, J., 1999. Global-scale analysis of the
533 mantle Pds phases. *J. Geophys. Res.* 101, 20,203–20,219.
- 534 Chopelas, A., 1991. Thermal properties of β -Mg₂SiO₄ at mantle pressures
535 and pressures derived from vibrational spectroscopy: Implications for the
536 mantle at 400 km depth. *J. Geophys. Res.* 96, 11817–11829.
- 537 Chopelas, A., Boehler, R., Ko, T., 1994. Thermodynamics of γ -Mg₂SiO₄
538 from Raman spectroscopy at high pressure: the Mg₂SiO₄ phase diagram.
539 *Phys. Chemistry of Minerals* 21, 351–359.
- 540 Deuss, A., 2007. Seismic observations of transition zone discontinuities be-
541 neath hotspot locations. In: Foulger, G., Jurdy, D. (Eds.), *Plates, Plumes,*
542 *and Planetary Processes.* Geological Society of America.
- 543 Dueker, K., Sheehan, A., 1997. Mantle discontinuity structure from mid-
544 point stacks of converted P to S waves across the Yellowstone Hotspot
545 Track. *J. Geophys. Res.* 102, 8313–8327.
- 546 Fei, Y., Van Orman, J., Li, J., van Westrenen, W., Sanloup, C., Minarik,
547 W., Hirose, K., Komabayashi, T., Walter, M., Funakoshi, K., 2004. Ex-
548 perimentally determined postspinel transformation boundary in Mg₂SiO₄
549 using MgO as an internal pressure standard and its geophysical implica-
550 tions. *J. Geophys. Res.* 109.
- 551 Gu, Y., Dziewonski, A., Agee, C., 1998. Global de-correlation of the topogra-
552 phy of transition zone discontinuities. *Earth Planet. Sci. Lett.* 157, 57–67.

553 Gu, Y., Dziewonski, A., Ekstrom, G., 2003. Simultaneous inversion for man-
554 tle shear velocity and topography of transition zone discontinuities. *Geo-*
555 *phys. J. Int.* 154, 559–583.

556 Gu, Y., Sacchi, G., 2009. Radon transform methods and their applications
557 in mapping mantle reflectivity structure. *Surv. Geophys.* 30, 327–354.

558 Helffrich, G., 2000. Topography of the transition zone seismic discontinuities.
559 *Rev. Geoph.* 38, 141–158.

560 Helffrich, G., Bina, C., 1994. Frequency dependence of the visibility and
561 depths of the mantle discontinuities. *Geophys. Res. Lett.* 21, 2613–2616.

562 Hirose, K., 2002. Phase transitions in pyrolitic mantle around 670-km depth:
563 Implications for upwelling of plumes from the lower mantle. *J. Geophys.*
564 *Res.* 107.

565 Houser, C., Masters, G., Flanagan, M., Shearer, P., 2008. Determination and
566 analysis of long-wavelength transition zone structure using SS precursors.
567 *Geophys. J. Int.* 174, 178–194.

568 Houser, C., Quentin, W., 2010. Reconciling Pacific 410 and 660 km discon-
569 tinuity topography, transition zone shear velocity patterns, and mantle
570 phase transitions. *Earth Planet. Sci. Lett.* 296, 255–266.

571 Irifune, T., Kubo, N., Isshiki, M., Yamasaki, Y., 1998. Phase transformations
572 in serpentine and transportation of water into the lower mantle. *Geophys.*
573 *Res. Lett.* 25, 203–206.

- 574 Ishii, T., Kojitani, H., Akaogi, M., 2011. Post-spinel transitions in pyro-
575 lite and Mg_2SiO_4 and akimotoite-perovskite transition in MgSiO_3 : Precise
576 comparison by high-pressure high-temperature experiments with multi-
577 sample cell technique. *Earth Planet. Sci. Lett.* 309, 185–197.
- 578 Ito, E., Akaogi, M., Topor, L., Navrotsky, A., 1990. Negative pressure-
579 temperature slopes for reactions forming MgSiO_3 perovskite from calorime-
580 try. *Science* 249, 1275.
- 581 Ito, E., Takahashi, E., 1989. Postspinel transformations in the system
582 Mg_2SiO_4 - Fe_2SiO_4 and some geophysical implications. *J. Geophys. Res.* 94,
583 10,637–10,646.
- 584 Ito, E., Yamada, H., 1982. Stability relations of silicate spinels, ilmenite
585 and perovskites. In: Akimoto, S., Manghnanni, M. (Eds.), *High-pressure*
586 *research in mineral physics*. Center for Academic Publishing, Tokyo, pp.
587 405–419.
- 588 Katsura, T., Ito, E., 1989. The system Mg_2SiO_4 - Fe_2SiO_4 at high pressures
589 and temperatures: precise determination of stabilities of olivine, modified
590 spinel and spinel. *J. Geophys. Res.* 94, 15,663–15,670.
- 591 Katsura, T., Yamada, H., Nishikawa, O., Song, M., Kubo, A., Shinmei,
592 T., Yokoshi, S., Aizawa, Y., Yoshino, T., Walter, M., Ito, E., Funakoshi,
593 K., 2004. Olivine-wadsleyite transition in the system $(\text{Mg,Fe})_2\text{SiO}_4$. *J.*
594 *Geophys. Res.* 109, 1–12.
- 595 Katsura, T., Yamada, H., Shinmei, T., Kubo, A., Ono, S., Kanzaki, M.,
596 Yoneda, A., Walter, M., Ito, E., Urakawa, S., Funakoshi, K., Utsumi, W.,

597 2003. Post-spinel transition in Mg_2SiO_4 determined by high P-T in situ
 598 X-ray diffractometry. *Phys. Earth Planet. Inter.* 136, 11–24.

599 Kennett, B., Engdahl, E., 1991. Travel times for global earthquake location
 600 and phase identification. *Geophys. J. Int.* 105, 429–465.

601 Kirby, S., Stein, S., Okal, E., Rubie, D., 1996. Metastable mantle phase
 602 transformations and deep earthquakes in subducting oceanic lithosphere.
 603 *Rev. Geoph.* 34, 261–306.

604 Lawrence, J., Shearer, P., 2006. A global study of transition zone thickness
 605 using receiver functions. *J. Geophys. Res.* 111.

606 Lawrence, J., Shearer, P., 2008. Imaging mantle transition zone thickness
 607 with SdS-SS finite-frequency sensitivity kernels. *Geophys. J. Int.* 174, 143–
 608 158.

609 Lebedev, S., Chevrot, S., van der Hilst, R., 2002. Seismic evidence for olivine
 610 phase changes at the 410- and the 660-kilometers discontinuities. *Science*
 611 296, 1300–1302.

612 Li, X., Kind, R., Yuan, X., Sobolev, S., Hanka, W., Ramesh, D., Gu, Y.,
 613 Dziewonski, A., 2003. Seismic observation of narrow plumes in the oceanic
 614 upper mantle. *Geophys. Res. Lett.* 30, 1334.

615 Litasov, K., Ohtani, E., Sano, A., Suzuki, A., 2005a. Wet subduction versus
 616 cold subduction. *Geophys. Res. Lett.* 32, 1–5.

617 Litasov, K., Ohtani, E., Sano, A., Suzuki, A., Funakoshi, A., 2005b. In situ
 618 x-ray diffraction study of post-spinel transformation in a peridotite mantle:

619 Implication for the 660-km discontinuity. *Earth Planet. Sci. Lett.* 238, 311–
620 328.

621 Revenaugh, J., Sipkin, S., 1994. Seismic evidence for silicate melt atop the
622 410-km mantle discontinuity. *Nature* 369, 474–476.

623 Ricard, Y., Mattern, E., Matas, J., 2005. Synthetic Tomographic Images of
624 Slabs from Mineral Physics. In: van der Hilst, R., Bass, J., Matas, J.,
625 Trampert, J. (Eds.), *Earth’s Deep Interior: Structure, Composition, and*
626 *Evolution*. Vol. 160. American Geophysical Union, Washington, D.C., pp.
627 283–300.

628 Romanowicz, A., Cara, M., 1980. Reconsideration of the relation between S
629 and P station anomalies in North America. *Geophys. Res. Lett.* 7, 417–420.

630 Rondenay, S., 2009. Upper mantle imaging with array recordings of converted
631 and scattered teleseismic waves. *Surv. Geophys.* 30, 377–405.

632 Rost, S., Thomas, C., 2002. Array seismology: Methods and applications.
633 *Rev. Geoph.* 40, 1–27.

634 Schmandt, B., Dueker, K., Hansen, S., Jasbinsek, J., Zhang, Z., 2011. A spo-
635 radic low-velocity layer atop the western U.S. mantle transition zone and
636 short-wavelength variations in transition zone discontinuities. *Geochem.*
637 *Geophys. Geosys.* 12, 1–26.

638 Schmandt, B., Dueker, K., Humphreys, E., Hansen, S., 2012. Hot mantle
639 upwelling across the 660 beneath Yellowstone. *Earth Planet. Sci. Lett.*
640 331–332, 224–236.

- 641 Schmandt, B., Humphreys, E., 2010. Complex subduction and small-scale
642 convection revealed by body-wave tomography of the western united states
643 upper mantle. *Earth Planet. Sci. Lett.* 297, 435–445.
- 644 Shearer, P., 1991. Constraints on Upper Mantle Discontinuities From Obser-
645 vations of Long-Period Reflected and Converted Phases. *J. Geophys. Res.*
646 96, 18,147–18,182.
- 647 Shearer, P., 2000. Upper mantle seismic discontinuities. In: et al., S. K. (Ed.),
648 *Earth’s Deep Interior: Mineral Physics and Tomography From the Atomic*
649 *to the Global Scale*, *Geophys. Monogr. Ser. Vol. 117*. AGU, Washington,
650 D.C., pp. 115–131.
- 651 Simmons, N. A., Gurrola, H., 2000. Multiple seismic discontinuities near the
652 base of the transition zone in the Earth’s mantle. *Nature* 405, 559–562.
- 653 Smyth, J., Jacobsen, S., 2006. Nominally anhydrous minerals and Earth’s
654 deep water cycle. In: Jacobsen, S., van der Lee, S. (Eds.), *Earth’s deep*
655 *water cycle*. Vol. 167. AGU monograph, pp. 1–9.
- 656 Song, T., Helmberger, D., Grand, S., 2004. Low-velocity zone atop the 410-
657 km seismic discontinuity in the northwestern United States. *Nature* 427,
658 530–533.
- 659 Stammler, K., Kind, R., 1992. Comment on “Mantle Layering from ScS
660 Reverberations, 2, the Mantle Transition Zone” by Justin Revenaugh and
661 Thomas H. Jordan. *J. Geophys. Res.* 97, 17,547–17,548.
- 662 Stixrude, L., Lithgow-Bertelloni, C., 2005. Mineralogy and elasticity of the

663 oceanic upper mantle: Origin of the low-velocity zone. *J. Geophys. Res.*
664 110.

665 Suito, K., 1977. *High Pressure Research*. Academic. N. Manghani and S. Aki-
666 moto.

667 Suzuki, A., Ohtani, E., Morishima, H., Kubo, T., Kanbe, Y., T., K., Okada,
668 T., Terasaki, H., Kato, T., Kikegawa, T., 2000. In situ determination of
669 the phase boundary between Wadsleyite and Ringwoodite in Mg_2SiO_4 .
670 *Geophys. Res. Lett.* 27, 803–806.

671 Tauzin, B., Debayle, E., Wittlinger, G., 2008. The mantle transition zone as
672 seen by global pds phases : no clear evidence for a thin transition zone
673 beneath hotspots. *J. Geophys. Res.* 113.

674 Tauzin, B., van der Hilst, R., Wittlinger, G., Ricard, Y., 2013. Multiple
675 transition zone seismic discontinuities and low velocity layers below west-
676 ern United States. *J. Geophys. Res.*

677 Vinnik, L., Ren, Y., Stutzmann, E., Farra, V., Kiselev, S., 2010. Observa-
678 tions of S410p and S350p phases at seismograph stations in california. *J.*
679 *Geophys. Res.* 115, 1–12.

680 Wittlinger, G., Vergne, J., Tapponnier, P., Farra, V., Poupinet, G., Jiang,
681 M., Su, H., Herquel, G., Paul, A., 2004. Teleseismic imaging of subducting
682 lithosphere and moho offsets beneath western Tibet. *Earth Planet. Sci.*
683 *Lett.* 221, 117–130.

684 Wood, B., 1995. The effect of H_2O on the 410-kilometer seismic discontinuity.
685 *Science* 268, 74–76.

686 S. Supplements

687 S.1. Effect of a correlation between δh and δT

688 Contrary to the situation in section 3.3, we assume here that there is a
 689 correlation C between the temperature in the TZ, δT , and the corrections
 690 due to shallow velocity anomalies, δh . Thus $C = \langle \delta T \cdot \delta h \rangle / (\Delta T \Delta h)$ is
 691 non zero (where ΔT and Δh are the RMS of δT and δh). The equations (11)
 692 become

$$\begin{aligned}
 \langle \delta Z_4 \cdot \delta Z_4 \rangle &= \Gamma_4^2 \Delta T^2 + \Delta h^2 + 2\Gamma_4 C \Delta T \Delta h \\
 \langle \delta Z_6 \cdot \delta Z_6 \rangle &= \Gamma_6'^2 \Delta T^2 + \Delta h^2 + 2\Gamma_6' C \Delta T \Delta h \\
 \langle \delta Z_6 \cdot \delta Z_4 \rangle &= \Gamma_4 \Gamma_6' \Delta T^2 + \Delta h^2 + (\Gamma_4 + \Gamma_6') C \Delta T \Delta h
 \end{aligned} \tag{S.1}$$

693 These three equations can be solved for the three unknowns ΔT , Δh and
 694 Γ_6' when Γ_4 and C are given. The algebra is cumbersome but straightforward.
 695 The solution Γ_6'/Γ_4 as a function of C is plotted in Fig. S.1a. The solution
 696 discussed in the main text corresponds to the case $C = 0$ (black star). Our
 697 estimate remains correct when the correlation between the shallow correction
 698 and the TZ temperature is weak (Fig. S.1a). We found also that the RMS
 699 of the shallow correction δh (not shown here) is minimum when $C = 0$ and
 700 vary only by $\approx 2\%$ for $-0.2 < C < 0.2$. Also, the thermal topography at
 701 ‘410’ decreases rather linearly with C .

702 S.2. Effect of a vertical temperature gradient in the TZ

703 We assume now that the temperatures at the top and bottom of the TZ
 704 differ by θ so that the temperature at the ‘660’ is $\delta T + \delta\theta/2$ and at the ‘410’
 705 is $\delta T - \delta\theta/2$. Assuming that δh , δT and $\delta\theta$ (of RMS $\Delta\theta$) are uncorrelated,

Figure S.1: (a) Recovery of the Clapeyron slope ratio when varying the correlation C between δh and δT . The solution discussed in the main text is $\Gamma'_6/\Gamma_4 = -0.56$ for $C = 0$ (black star). (b) Sensitivity of the Clapeyron slope ratio when varying the RMS of the difference in temperature at the ‘410’ and ‘660’. The solution discussed in the text is $\Gamma'_6/\Gamma_4 = -0.56$ for $\theta = 0$ K (black star).

the equations (11) become

$$\begin{aligned}
\langle \delta Z_4 \cdot \delta Z_4 \rangle &= \Gamma_4^2 \Delta T^2 + \Delta h^2 + \Gamma_4^2 \Delta \theta^2 / 4 \\
\langle \delta Z_6 \cdot \delta Z_6 \rangle &= \Gamma_6'^2 \Delta T^2 + \Delta h^2 + \Gamma_6'^2 \Delta \theta^2 / 4 \\
\langle \delta Z_6 \cdot \delta Z_4 \rangle &= \Gamma_4 \Gamma_6' \Delta T^2 + \Delta h^2 - \Gamma_4 \Gamma_6' \Delta \theta^2 / 4
\end{aligned} \tag{S.2}$$

Again, these three equations can be solved for the three unknowns ΔT , Δh and Γ'_6 when Γ_4 and $\Delta \theta$ are given. The solution Γ'_6/Γ_4 as a function of $\Delta \theta$ is plotted in Fig. S.1a where we see that Γ'_6/Γ_4 is not drastically affected by the assumption used in the main text of $\Delta \theta = 0$.

Figure S.2: Example of depth corrections for a west-east profile at 41°N. (a) Initial CCP section. Positive amplitudes (in red) mark increases of velocities with depth whereas negative amplitudes (in blue) are associated with reductions of velocities. The observed depths of the ‘410’ and ‘660’ from Tauzin et al. (2013) are indicated with black dashed lines. (b) Depths corrections δh for shallow velocity heterogeneities. (c) Depth corrections $a(Z)\delta T$ for velocities within and below the TZ. (d) Depth corrected CCP section. The corrected depths for the ‘410’ and ‘660’ are indicated with black dashed lines.

Table S.1: Experimental and seismological measurements for the Clapeyron slopes of olivine transformations.

	Clapeyron slope (MPa K ⁻¹)	Reference
Olivine → wadsleyite γ_4	+3.5	Suito (1977)
	+3.0	Akaogi et al. (1984)
	+2.5	Katsura and Ito (1989)
	+1.5	Akaogi et al. (1989)
	+2.8	Chopelas (1991)
	+3.0	Bina and Helffrich (1994)
	+2.0*	Lebedev et al. (2002)*
	+4.0	Katsura et al. (2004)
	+3.1	Akaogi et al. (2007)
Ringwoodite → Perovskite+Magnesiowustite γ_6	-2.0	Ito and Yamada (1982)
	-2.8	Ito and Takahashi (1989)
	-4.0	Ito et al. (1990)
	-2.6	Akaogi and Ito (1993)
	-2.0	Bina and Helffrich (1994)
	-2.5	Chopelas et al. (1994)
	-3.3*	Lebedev et al. (2002)*
	-2 to -0.4	Katsura et al. (2003)
	-1.3	Fei et al. (2004)
	-0.8 to -0.2	Litasov et al. (2005b)
	-2.6	Akaogi et al. (2007)
	-1.0	Ishii et al. (2011)

*Measurements obtained from seismology.