Supplementary Material for

**Pervasive seismic-wave low velocity zones within stagnant plates in the mantle transition zone: thermal or compositional origin?**

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**S1 Extended method section**

**S1.1 Data processing**

We use the waveforms of 531 teleseismic earthquakes recorded at 232 three-component broad-band seismic stations (Figure 1) located on the Korean peninsula, in Japan (F-Net), and from a temporary array in northeast China (NECESSArray). The earthquakes have magnitude ranging between 5.5 and 7.2, and were recorded at epicentral distances between 30° and 90°. We extract the P-to-S converted wavefield with the receiver function (RF) method.

We use the iterative time-domain deconvolution (Ligorria and Ammon, 1999) on records rotated along the radial and vertical directions with higher corner frequency at 0.2 Hz. For borehole stations in Korea the orientation of the sensor was determined by minimizing the projection of the P-wave amplitude on the transverse component. We compute the root-mean-square (rms) distribution over a 25 s-long window after the P-wave and over the whole set of
RFs used to build the seismic profiles, and get the median and standard deviation. The RF waveforms with rms within 1.5 standard deviations from the median rms value were finally selected. This operation removes highly oscillating waveforms associated with reverberations in thick sedimentary basins or above subducting plates. The final RF dataset is composed of 41,814 waveforms and provides a good coverage of P-to-S piercing points across the 410 and 660-km discontinuities below the area (Figure S1).

S1.2 Back-projection of the RF signal

Receiver functions with a range of incidence angles and back-azimuths are combined using a common conversion point (CCP) stacking procedure to form two-dimensional (2D) images of the seismic reflectivity structure. We exploit an efficient back-projection method for the P-S phases from the time domain to the depth domain (Wittlinger et al., 2004). Our approach follows the steps described in Tauzin et al. (2013): (i) ray-tracing to compute compressional and shear-wave ray coordinates in a one-dimensional Earth’s model, (ii) computing the travel-times of converted waves relative to the P-wave arrival, (iii) conversion of the amplitudes measured on the RFs from the time domain to the depth domain, (iv) projection of these amplitudes onto a 3D spatial grid encompassing the vertical plane through the profile, (iv) averaging the projected amplitudes into a 2D vertical plane along the profile, (v) smoothing the 2D image with a 2D ellipsoidal function, and finally (vi) masking (in gray) the parts of the image with no ray coverage. We refer to the work of Tauzin et al. (2013) for details of the application of this approach.

We use the *iasp91* velocity model (Kennett and Engdahl, 1991) for ray-tracing and computation of travel-times. As discussed in section S3 below, the use of a three-dimensional model from linearized tomographic inversion does not change much the overall reflectivity
patterns observed around and within the MTZ. Usually, these models do not achieve the resolution necessary for accurately correcting the MTZ topography obtained from receiver functions (e.g. Tauzin and Ricard, 2014), and instead of mapping the structure of the model into our images, we make the choice of showing the raw images back-projected using \textit{iasp91}.

For grouping the RFs by common conversion points, we use bins of 5 km along the direction of the profile. To increase the signal-to-noise ratio in the large-scale images in figures 2-6, the bin dimension in the direction orthogonal to the profile is 500 km (±250 km). We reduce this orthogonal bin dimension to ±100 km when mapping the geographical location of seismic boundaries with higher resolution (see section S1.4). The grid size is 1 km in the vertical direction. Afterwards, we smooth the 2D CCP images with a Gaussian function, with semi-axes at half the maximum amplitude of 25 km in the horizontal and 5 km in the vertical direction. We cannot therefore expect to distinguish interfaces or objects separated by less than 50 km and 10 km in these directions. In Figure S2, we illustrate the effect of applying the smoothing operation on the A-a profile. The seismic structure remains clearly identifiable even without this smoothing operation. The spatial resolution in the direction perpendicular to the profile, however, depends on the distance of projection of stacked amplitudes on the profile.

In the seismic cross-sections (Figures 2-6), we plot the amplitude of stacked RFs, which is related to the P-to-S transmission coefficient, itself a function of the vertical shear-wave velocity contrast (see inset in Figure 1). Positive amplitudes shown in red correspond to a downward increase of velocity ($\Delta v_s > 0$) and negative amplitudes (blue) to a velocity reduction ($\Delta v_s < 0$).

\textbf{S1.3 Estimating the noise level}
Natural or artificial noise can be thought of as a stationary stochastic process that is uncorrelated on records of earthquakes taken at different times. In contrast the structure below the receivers is not expected to change through time. We have therefore used a bootstrap resampling approach (Efron and Tibshirani, 1990) to estimate the time-variability in the seismic record (the noise), and thereby determine uncertainties for the stacked seismic amplitudes. This approach was applied for data from the western US in a preliminary way by Hier-Majumder and Tauzin (2017). We constructed 100 RF bootstrap samples (Efron and Tibshirani, 1990). We applied CCP stacking for these RF samples to obtain an average seismic section and a standard error on the seismic amplitudes. The 68% and 95% confidence intervals, given by one and two standard deviations respectively, include 68% and 95% of the recovered models.

S1.4 Mapping the apparent LVZs

We built maps of the geographical distribution of reflective zones (see Figures 7 and 8). Following Tauzin et al. (2013), these maps are obtained from CCP stacking along orthogonal profiles at intervals of 0.5° in longitude and latitude, with a projection distance of ±100 km, and combining the information provided by the North-South and East-West seismic cross sections. We do not attempt here precise picking for the depth of interfaces, because of the variable and possibly multi-modal nature of the zones in some regions. Instead we average the CCP amplitudes in two distinct depth windows, from 325 to 400 km depth for the shallower Z1, and from 550 to 650 km depth for the deeper Z2. The 100 bootstrap amplitude samples provide at each geographical location the average amplitude and corresponding standard deviation (errors) for each of the two depth intervals. The lateral variations of seismic amplitudes, within their uncertainties, provide an indication as to whether a negative zone is present or not (Figures 7 and 8).
S2 Comparison of CCP stacked RF images with tomographic models

We compare our RF images with three global tomographic models and one regional model (Schaeffer and Lebedev, 2013; Debayle and Ricard, 2012; Obayashi et al., 2013; Gorbatov and Kennett, 2003). These four models give consistent images of the structure of the upper-mantle and transition zone (Figures S11-S12). Despite being global models from long-period surface-waves, Schaeffer and Lebedev (2013)’s and Debayle and Ricard (2012)’s models have a nominal resolution of ~600 km in the upper 200 km (equivalent to spherical harmonic degree ~60) and are consistent up about degree 30 (horizontal wavelengths of about 1300 km) in parts of the deeper mantle. The S body-wave travel-time tomographic model from Gorbatov and Kennett (2003) provides better lateral resolution in the upper mantle, but puts fewer constraints on MTZ structure due to the limitations of station coverage at the surface. The global P-wave model from Obayashi et al. (2013) provides a good compromise between resolution and coverage. However, this latter model is not representative of the shear-wave velocity structure seen by receiver functions.

S3 On the origin of possible artifacts in CCP sections

Spurious signals in CCP sections derived from receiver functions can arise from various reasons, in some cases without any relation with the Earth’s mantle structure. The most common origins are seismic noise on raw and deconvolved seismograms, sidelobes due to signal processing artifacts, in particular from deconvolution, interference of seismic phases, contamination by multiple reverberations within the uppermost mantle structure, or approximations in the physics underlying the imaging principle.

In Figures S13 and S14, we provide an analysis of the noise level for two of our seismic
images along the E-e and D-d profiles, and more generally over the whole region when
mapping Z1 and Z2. Stations from the permanent network in Korea, and to a lesser extent
from F-Net in Japan, show less signal variability than the stations from NECESSArray in
China. This is expected because NECESSArray is a temporary array of portable stations.
Significant spatial variations are also observed within this network, which covers a range of
geological environments. In particular, the NECESSArray shows a greater noise level in the
Songliao sedimentary basin, possibly due to incoherent stacking of waveforms contaminated
by shallow reverberations. We plot in Figures 7 and 8 the signals when the amplitudes exceed
the 68% confidence level. In Figure S13 we illustrate the effect on our seismic images of
masking the amplitudes below the 95% confidence level. The major features discussed in
section 3 of the main article are robust because their stacked amplitudes exceed the noise
level.

Because the apparent zones of reflectivity with negative amplitude have in some regions no
clear lateral coherence or are observed below sedimentary basins, some authors (e.g. Gao et
al., 2010; Liu et al., 2015; Cottaar and Deuss, 2016) are cautious when they come to the
interpretation of the negative signals. The signals could indeed result from the interference of
shallow reverberations in basins or in the shallow upper mantle. A way of checking the origin
of seismic arrivals is to build a vespagram, showing the slowness (~incidence angle) of the
seismic arrivals as a function of time for the receiver functions (Figure S15). In figure S15,
direct conversions are expected with a positive slowness (pP-pPds) whereas multiple
reverberations are expected with a negative slowness (pP-pPPds; pP-pPSds). We show in
Figure S15 these diagrams for three zones where Z1 and Z2 are observed: zone I in Korea,
zone II below the Songliao basin in northeast China, and zone III sampling the deep cluster of
earthquakes below the Japan Sea (Figures 7 and 8). In all three zones, the negative arrivals
associated with Z1 and Z2 have slowness similar to the conversions at the 410-km and 660-km discontinuities (Figure S15). We cannot exclude that multiple reverberations contribute to the observed negative signals, but the signature of apparent LVZs appears mainly dominated by direct conversions. This is also valid for the Songliao sedimentary basin (Figure S15b) despite the observed complex seismic signature (Figures 4, 5, 7).

There are several physical assumptions underlying our imaging principle. One of them is the use of a one-dimensional Earth’s model (iasp91) for ray-tracing and back projection of the signal of the individual receiver function traces from the time domain to the spatial domain. We show in Figure S16 that using a (global) three-dimensional model for the back-projection only induces changes of a few kilometers in the average depths of the TZ boundaries (e.g. the 410 is uplifted by 3 km with the model from Debayle and Ricard (2012) compared with iasp91), but does not change much the overall pattern of stacked amplitudes, in particular for the negative zones. Our imaging principle does not take into account the physical process of scattering (Wu and Aki, 1985) where wave paths can be deviated by localized heterogeneities in the medium properties. Due to this approximation, the P-P and P-S diffracted energy recorded on the RFs is only approximately back-projected to the spatial domain. A portion of the signal of the apparent LVZs may therefore arise from incoherent mapping. We believe however that many contiguous diffraction points would be required to explain our results across such a large scale. 3D pre-stack Kirchhoff depth migration (Cheng et al., 2017) offers the possibility of significant improvement of imaging and mapping of anomalous reflectivity zones in the MTZ, with suitable base velocity models.

Potential problems may arise in our imaging from projecting 3D structure on to 2D profiles. We should not expect these problems to arise when taking profiles perpendicular to a 2D
structure, such as the dipping direction of a subducting slab. However in the case of profiles in other directions, using large bin dimensions in the direction orthogonal to the profile may introduce spurious structure related to the 3D structure. We tested this effect in three ways. First, we made a comparison of mapping the 410-km discontinuity using automatic picking, bin size of 5x200 km, and using regular profiles in orthogonal E-W and S-N directions (Tauzin et al., in preparation). Second, we varied the bin dimension in the direction orthogonal to the profile from ±150 km to ±25 km, and tested the recovery of the structure (Figure S3). Third, we also tested the effect of stacking using circular bins instead of rectangular bins (Figure S3). The first test of mapping the 410 demonstrates if the effect of projecting 3D structures away from the profile affects the recovery of the 410 structure. We found a remarkable consistency between the maps, with the exception of a few places, possibly due to the reverberations in the Songliao basin, and the poorer coverage at the extreme sides of the studied area. In the second test, varying the bin dimension in the direction orthogonal to the profile does not affect much the observed structure, although it degrades significantly the signal to noise ratio with decreasing dimensions (Figure S3). Finally, we found that using circular bins instead of rectangular bins give similar images along the tested profile (Figure S3). This suggests that the effect of projecting the 3D structure far away from the profile does not bias much the recovered structure along a 2D profile. Finally, in a synthetic test (Figure S4-S8), we also tested the possibility that the negative zones arise from interference of seismic phases other than conversions, or by sidelobes introduced by the deconvolution. We created for the source-receiver pairs in our database full-waveform synthetic seismograms in a spherical 1-D structure using a reflectivity algorithm (Fuchs and Muller, 1971) with source parameters from the global CMT catalogue (http://www.globalcmt.org/). These synthetic seismograms are then processed in the same
way as the real data. We simulate wave-propagation using the *iasp91* model (Kennett and Engdahl, 1991). No clear negative signal appears above the 410 (Figure S4-S8). Faint negative signals appear at the base of the MTZ. These weak features result likely from sidelobes and interference of phases in regions where the distribution in data slowness is heterogeneous. It is for example notable that these effects increase where data coverage is weaker. A modest component (one third of the amplitude observed on the data) of the negative reflectivity at the base of the transition zone is therefore likely to be associated with the characteristics of illumination by distant earthquakes, but still a significant portion of it remains to be explained.

**S4 Thermal model of subduction zones and recovery of the thermal gradient from CCP stacking**

We use the finite difference code TEMSPOL (Negredo et al., 2004) for calculating the temperature anomaly in the deep subduction zone of North Honshu. The geometry of plate subduction (dip angle), the plate subduction velocity, and the age of subducting plate are prescribed as given by Syracuse and van Keken (2010). These parameters as well as other parameters for the model domain, the adiabatic and conductive temperature gradients, and the heat transfer equation for the finite difference calculation are provided in table S1. We checked that the geometry of the resulting thermal model fitted the Wadati-Benioff zone observed in North Honshu at the location defined by Syracuse and van Keken (2010).

In our simple test in Figure 10, we supposed that teleseismic converted waves sample a one-dimensional horizontal stratification along the strike of the slab. We extracted a temperature profile at the specific location where the slab is the coldest at 500 km depth. A one-dimensional profile of thermal anomaly is obtained from the difference between the
temperature along the adiabatic temperature gradient and the slab geotherm. We then convert
this profile of temperature anomalies into shear-wave velocity anomalies relative to a
reference velocity model. The reference velocity model is *iasp91*. To estimate the shear-wave
velocity anomalies, we use the partial derivatives of $v_s$ with respect to temperature provided
by the self-consistent thermodynamical modeling from Stixrude and Lithgow-Bertelloni
(2005), $K = -7.0 \times 10^{-5}$ K$^{-1}$. We test also a value at the higher limit of the values estimated for
minerals of the mantle transition zone, $K = -12.0 \times 10^{-5}$ K$^{-1}$. The $v_p$ model is scaled from the $v_s$
model with $R = d \ln v_s / d \ln v_p = 2$. The density is kept at the *iasp91* values. We then
computed synthetic seismograms and receiver functions for each source-receiver pairs in our
database with the reflectivity algorithm (Fuchs and Muller, 1971). We process these receiver
functions the same way as the data to obtain the synthetic profiles shown for the slab models
in figure 10.

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**Table S1** Parameters for the finite difference calculation.

**References for the Supplementary Material**


Supplementary Fig. 1 (a) Map of coverage in piercing points at the 410-km (red crosses) and 660-km (black crosses) discontinuities. (b) Density of conversion points at the 410-km discontinuity in bins of 0.5°x0.5° in latitude and longitude (same grid as for mapping seismic interfaces).
Supplementary Fig. 2 Examples of the effect of smoothing on the seismic image obtained for the A-a profile. (a) Raw CCP stacked image. Positive (red) and negative (blue) amplitudes correspond to downward increase/decrease of shear-wave velocities. Saturation at ±3% the P-wave amplitude (as in Figures 2-4). (b) Superimposition of the seismicity from the USGS (black dots) and of our interpretation. Following Kawakatsu and Yoshioka (2011), the negative signal below the 410-km discontinuity and the deepest cluster of earthquakes (contoured in black) could be associated with metastability of olivine upon transformation into wadsleyite. (c) A map locating the A-a profile with the seismic stations (black triangles), volcanoes related to the arc of subductions (small red triangles) and intra-plate (large red triangles). (d) Smoothed version of the image in (a) and (b). Question marks highlight zones of scattering with apparent low-velocities within or surrounding the MTZ.
Supplementary Fig. 3 (a) Effect of changing the rectangular bin size in the direction orthogonal to the profile for the structure of the 660-km discontinuity along the E-e profile. The bin has a dimension of 5 km along the profile, and an orthogonal dimension given as the double of the value in the bottom left of each panel. (b) Comparison with CCP stacking in circular bins taken every 5 km along the profile, and with a radius indicated at the bottom left of each panel.
Supplementary Fig. 4 (a) CCP seismic section obtained for the A-a profile from the simulation of wave-propagation (Fuchs and Muller, 1971) in the iasp91 (Kennett and Engdhal, 1991) velocity structure. The radial shear-wave velocity profile is shown in (b). (c) Observed seismic section.
Supplementary Fig. 5 (a) CCP seismic section obtained for the B-b profile from the simulation of wave-propagation (Fuchs and Muller, 1971) in the iasp91 (Kennett and Engdhal, 1991) velocity structure. The radial shear-wave velocity profile is shown in (b). (c) Observed seismic section.
Supplementary Fig. 6 (a) CCP seismic section obtained for the C-c profile from the simulation of wave-propagation (Fuchs and Muller, 1971) in the iasp91 (Kennett and Engdhal, 1991) velocity structure. The radial shear-wave velocity profile is shown in (b). (c) Observed seismic sections.
Supplementary Fig. 7 (a) CCP seismic section obtained for the D-d profile from the simulation of wave-propagation (Fuchs and Muller, 1971) in the iasp91 (Kennett and Engdhal, 1991) velocity structure. The radial shear-wave velocity profile is shown in (b). (c) Observed seismic sections.
Supplementary Fig. 8 (a) CCP seismic section obtained for the E-e profile from the simulation of wave-propagation (Fuchs and Muller, 1971) in the iasp91 (Kennett and Engdhal, 1991) velocity structure. The radial shear-wave velocity profile is shown in (b). (c) Observed seismic sections.
Supplementary Fig. 9 A series of roughly West-East cross-sections using the RF data from the NECESSArray. (a) Map locating the four profiles. (b-e) Cross-sections with (top) RF
image and (bottom) seismic tomography (Debayle and Ricard, 2012). We superimpose the signal from RF CCP stacking, emphasizing negative waveforms in black.
Supplementary Fig. 10 Same as in Supplementary Figure 5, except that the images are constructed along profiles in a South-North direction.
Supplementary Fig. 11 Cross-sections along A-a through the shear-wave tomographic models of (a, b) Gorbatov and Kennett (2003) and (c, d) Schaeffer and Lebedev (2013). In (b) and (d), we superimpose the signal from RF CCP stacking, emphasizing negative waveforms in black.
Supplementary Fig. 12 Cross-sections along B-b (a), C-c (b), D-d (c) and E-e (d) profiles through the shear-wave tomographic models of Gorbatov and Kennett (2003) (left), Schaeffer and Lebedev (2013) (middle), and the P-wave tomographic model from Obayashi et al. (2013). We superimpose the signal from RF CCP stacking, emphasizing negative waveforms in black.
Supplementary Fig. 13 (a) Bootstrap errors (95% confidence level) obtained for the seismic section along the E-e profile. The regions of errors (in % of the P-wave amplitude) are shown with gray patterns. The uniform intermediate gray background indicates no ray coverage. The seismicity is shown with black dots. (b) CCP cross-section (similar to Figure 6) except that we report only the stacked amplitudes above the level of uncertainty. The color scale saturates at ±3% the P-wave amplitude with positive amplitude in red and negative amplitude in blue. (c) Same as in (a) but for the D-d profile. (d) Same as in (b) but the D-d profile.
Supplementary Fig. 14 (a) Map of the average seismic amplitudes measured within the 325-400 km depth range. (b) Corresponding uncertainties (68% confidence level). (c) Map of the average seismic amplitudes measured within the 560-660 km depth range. (d) Corresponding uncertainties (68% confidence level). To build Figure 7, we mask the geographical locations where seismic amplitudes (a, c) do not exceed the 68% confidence level (b, d).
(a) Zone I. (Korea)

(b) Zone II. (Songliao)

(c) Zone III. (Japan Sea)
Supplementary Fig. 15 Testing the presence of multiple reverberations in three of the zones of observation of LVZ1 and LVZ2. We plot the stacked RF sections as function of the epicentral distance in bins of 0.5° distance (left panel) and the slant-stack diagram of corresponding individual RFs. The slant-stack diagram (or vespagram) shows the differential slowness ($pP-pPds > 0$) between the P-wave and the P-to-S conversions as a function of the travel-time ($t$) after the P-wave. The reference travel-time and slowness ($t = 0, p = 0$) in these diagrams are from the P direct arrival. (a) Diagrams from RFs sampling LVZ1 and LVZ2 in zone I in Korea (33-38°N; 125-130°W). (b) Diagrams for the zone II below the Songliao basin in northeast China (40-48°N; 120-127°W). (c) Diagrams for RFs sampling the deep cluster of earthquakes in zone III below the Japan Sea.
Supplementary Fig. 16 Testing the effect of corrections for the 3D velocity structure. (a) Original image obtained by backprojecting the RF signal using the IASP91 velocity model (Kennett and Engdahl, 1991). (b) Image obtained after back-projection using the 3D shear-wave velocity model of Debayle and Ricard (2012) (DR2012). The \( v_p \) model is scaled from the \( v_s \) anomalies relative to the PREM model and using \( R = \Delta \ln v_s / \Delta \ln v_p = 2 \). The 410 is in average at 421 km compared to 423 km in IASP91, mainly due to the fact that the upper mantle in DR2012 is slower than IASP91 (see (c)). (c) A comparison of the shear-wave velocity profile in the IASP91 model (red) and over the region [21-51°N; 111-149°E], in the DR2012 model (black). The crust is taken from 3SMAC (Nataf and Ricard, 1996). Our CCP stacking approach extracts a one-dimensional velocity model by sampling DR2012 along the
rays, computes the associated time-to-depth relationship, and back-projects the time signal into the spatial domain.