Seismic discontinuities in the mantle transition zone and at the top of the lower mantle beneath eastern China and Korea: Influence of the stagnant Pacific slab

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\textbf{A B S T R A C T}

We have analyzed broadband data to identify and determine mantle discontinuities in the mantle transition zone (MTZ) beneath eastern China, where the Pacific slab is stagnant. The depths of the 410 and 660 km discontinuities are generally shallower and deeper, respectively, than the global averages in and near the Pacific slab beneath eastern China. The MTZ is thicker in the slab than the global average. This observation is consistent with the thermally controlled olivine to wadsleyite transformation for the 410 km discontinuity and the post-spinel transformation for the 660 km discontinuity. Other discontinuities appear from 690 to 750 km in the lower mantle part of the Pacific slab. Recent mineralogical experiments indicate that the most plausible interpretation of these deep discontinuities is that they represent the ilmenite to perovskite transformation in a cold environment, such as that in the Pacific slab.

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\section{1. Introduction}

The two main global discontinuities in the mantle are located at depths of around 410 and 660 km (called the “410” and “660” in this paper), though their depths vary slightly in different tectonic zones (e.g., Flanagan and Shearer, 1998). The “410” and “660” are considered to be due to the olivine to wadsleyite and post-spinel phase transformations, respectively (e.g., Ito and Takahashi, 1989). The pressure of both transformations is thermally controlled because the olivine to wadsleyite and post-spinel phase transformations are exothermic (positive Clapeyron slope) and endothermic (negative Clapeyron slope) reactions, respectively (e.g., Bina and Helffrich, 1994). In a cold (hot) environment such as a subduction zone, the “410” and “660” should be elevated (depressed) and depressed (elevated), respectively, which should generate temperature-related topography in the mantle transition zone (MTZ).

However, recent mineralogical studies suggest that phase transformations related to garnet, the other major component of the mantle, are involved, in addition to the olivine-related post-spinel transformation, at depths from 600 to 750 km (e.g., Vacher et al., 1998), and should therefore appear as multiple seismic discontinuities in that depth range. Recent seismological studies have detected such multiple discontinuities (Simmons and Gurrula, 2000; Deuss et al., 2006; Tibi et al., 2007; Schmerr and Garnero, 2007; Andrews and Deuss, 2008). For subduction zones in particular, the relationship between subducted slabs and multiple phase transformations has been investigated in order to understand the fate of subducted slabs. Niu and Kawakatsu (1996) revealed a complicated 660 km discontinuity at one seismic station (MDJ) in northeastern China, and from a small array of data, Ai et al. (2003) found multiple discontinuities near the “660” in the same region, where the Pacific slab is imaged as high velocity anomalies that lay horizontally (called a stagnant slab). Chen and Ai (2009) also found multiple discontinuities near the “660” beneath the south-eastern part of North China Craton, where high velocity anomalies are located. The stagnant slab is, however, spreading in the MTZ beneath a wide area of eastern China, where multiple discontinuities have never been reliably reported except in northeastern China. Among other previous studies, Shen et al. (2008) was the only systematic receiver function study to determine the “410” and “660” depths beneath the entire Chinese region. Their result showed a thick MTZ beneath eastern China, which is a signature of the cold MTZ around the stagnant slab. However, they did not
obtain clear signals due to the multiple discontinuities below the “660,” while they suggested a thin low-velocity layer above the “660.”

The aim of the present study is to investigate the complex structure of the MTZ that underlies the entire region of eastern China using receiver function analysis (e.g., Langston, 1997; Owens and Crosson, 1988). We analyzed shorter-period Ps converted waves at a larger number of permanent broadband stations in eastern China and the Korean peninsula than were used by Shen et al. (2008), which enabled us to detect multiple discontinuities and to study the interaction of the stagnant slab and mantle discontinuities.

2. Data and methods

We used all the permanent broadband stations in the eastern Chinese region, which includes northeast China, north China, and Korea. We analyzed data from the National Seismograph Network of China (NSNC), the Chinese Digital Seismic Network (CDSN), the Incorporated Research Institution for Seismology (IRIS) of the United States of America, and the Ocean Hemisphere Project Network (OHP) of Japan. We used a total of 21 stations in the research for this paper (Fig. 1a): 15 NSNC stations, 4 IRIS/CDSN stations, 1 IRIS station, and 1 OHP station in Korea. The stations were spread over all of eastern China and the Korean peninsula, where the stagnant slab is stalling in the MTZ. We selected earthquakes that took place between 2004 and 2006, whose magnitude ranged from 6.0 to 7.5 and whose epicentral distance ranged from 30° to 90° (Fig. 1b). Before the receiver function analysis was conducted, we visually inspected all seismograms and discarded data with low signal-to-noise ratios. Finally, more than 30 records were obtained from each station (Table 1). A band-pass filter with corner periods from 1 to 50 s was applied to the selected seismograms to suppress noise, partic-

![Fig. 1](image-url)

**Table 1**

Depths of the “410,” “660,” and “720” discontinuities and MTZ thickness.

<table>
<thead>
<tr>
<th>Station</th>
<th>“410” depth (km)</th>
<th>“660” depth (km)</th>
<th>MTZ thickness (km)</th>
<th>“720” depth (km)</th>
<th>Numbers of data</th>
</tr>
</thead>
<tbody>
<tr>
<td>BJT</td>
<td>421 ± 3</td>
<td>686 ± 3</td>
<td>265 ± 3</td>
<td>–</td>
<td>88</td>
</tr>
<tr>
<td>BNX</td>
<td>404 ± 4</td>
<td>680 ± 7</td>
<td>276 ± 8</td>
<td>–</td>
<td>40</td>
</tr>
<tr>
<td>CN2</td>
<td>408 ± 4</td>
<td>668 ± 2</td>
<td>260 ± 4</td>
<td>–</td>
<td>52</td>
</tr>
<tr>
<td>CNS</td>
<td>411 ± 11</td>
<td>671 ± 6</td>
<td>260 ± 14</td>
<td>–</td>
<td>66</td>
</tr>
<tr>
<td>DL2</td>
<td>–</td>
<td>656 ± 4</td>
<td>–</td>
<td>60</td>
<td></td>
</tr>
<tr>
<td>HEH</td>
<td>410 ± 5</td>
<td>662 ± 8</td>
<td>251 ± 10</td>
<td>–</td>
<td>75</td>
</tr>
<tr>
<td>HHC</td>
<td>423 ± 1</td>
<td>666 ± 10</td>
<td>244 ± 10</td>
<td>–</td>
<td>100</td>
</tr>
<tr>
<td>HIA</td>
<td>415 ± 1</td>
<td>667 ± 2</td>
<td>253 ± 3</td>
<td>–</td>
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</tr>
<tr>
<td>HNS</td>
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<td>675 ± 1</td>
<td>253 ± 2</td>
<td>721 ± 9</td>
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</tr>
<tr>
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<td>658 ± 5</td>
<td>–</td>
<td>58</td>
<td></td>
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<tr>
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<td>678 ± 2</td>
<td>265 ± 11</td>
<td>–</td>
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</tr>
<tr>
<td>MDJ</td>
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<td>664 ± 12</td>
<td>237 ± 15</td>
<td>700 ± 20</td>
<td>82</td>
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<td>677 ± 2</td>
<td>260 ± 6</td>
<td>748 ± 10</td>
<td>55</td>
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<tr>
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<td>258 ± 6</td>
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<tr>
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<td>672 ± 9</td>
<td>–</td>
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<tr>
<td>SLY</td>
<td>–</td>
<td>668 ± 11</td>
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<tr>
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<td>406 ± 3</td>
<td>659 ± 8</td>
<td>253 ± 8</td>
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<td>672 ± 10</td>
<td>255 ± 10</td>
<td>719 ± 7</td>
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<tr>
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<td>668 ± 10</td>
<td>262 ± 10</td>
<td>–</td>
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<td>–</td>
<td>242 ± 6</td>
<td>–</td>
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In this study, we employed the velocity spectrum stacking (VSS) method (Gurrola et al., 1994). To equalize the source effect and extract Ps phases, we computed the receiver functions by performing a frequency-domain deconvolution of a vertical-component record from a radial-component record for each event at each station. All receiver functions were plotted for each station for

Fig. 2. Receiver functions at HIA are shown as examples. Predicted arrival times are also shown.

particularly the long-period noise observed in the horizontal-component records.

Fig. 3. VSS plot computed from all receiver functions at station HIA. (Top) Amplitudes of stacked receiver functions normalized to 20% of the direct P-wave amplitudes and shown with respect to Ps conversion depths and the average shear wave velocity perturbations above the conversion depths. Two plus signs indicate peaks of the “410” and “660” obtained using velocities taken from Ritsema and van Heijst (2000). (Bottom) Stacked amplitudes versus depth curve for zero velocity correction. Two plus signs indicate converted signals from the “410” and “660.” The dashed curves indicate a 95% confidence level computed by the bootstrap method.

Fig. 4. All stacked and depth-migrated receiver functions obtained by this study. A band-pass filter between 4 and 50 s was used in computing receiver functions. Dashed curves indicate a 95% confidence level. We identified the “720” signals (dots) as later phases to P660s with amplitudes above a 95% confidence level. We did not choose seismograms showing a continuous oscillation around the MTZ (such as WZH, HHC, and HEH) even when they showed apparent later arrivals.

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Fig. 5. Depths of mantle discontinuities. Deviations from the global averages (418 and 660 km for the “410” and “660”, respectively, by Flanagan and Shearer, 1998) are shown. (a) “410”, (b) “660”, and (c) MTZ thickness. Black and white circles denote shallower and deeper discontinuities, respectively. Circles are plotted at average positions of Ps pierce points for each station. P-velocity tomograms at each depth (Fukao et al., 2001) are shown in the background. The correlation between MTZ thickness and P-velocity anomalies is shown in (d). Straight lines in (d) indicate lines predicted from Clapeyron slopes of $-0.5$ MPa/K, $-1.0$ MPa/K, $-2.0$ MPa/K, and $-3.0$ MPa/K for the post-spinel transformation.

comparison with the theoretical Ps time curves expected for conversion depths of “410” and “660”; an example is shown in Fig. 2. The peak alignments of the receiver functions near the expected P410s and P660s curves are obvious, though they are not large enough to analyze individually. To constrain the “410” and “660” depths, we employed the VSS method for stacking with moveout correction. In this process, the time axis of the receiver functions is converted to a depth axis. The moveout correction is computed for 1 km grids of discontinuity depths from 0 to 800 km and 0.1% grids of shear wave velocity perturbations from $-2.5\%$ to $+2.5\%$, with respect to the Iasp91 model (Kennett and Engdahl, 1991). In stacking receiver functions, a clear enhanced Ps-converted signal should be obtained with a moveout computed at the true conversion depth and true average shear velocity perturbation (Fig. 3). There is a strong trade-off between the discontinuity depth and the average velocity perturbation above the discontinuity, which thus cannot be determined independently (e.g., Gurrola and Minster, 1998). We therefore relied upon existing three-dimensional shear velocity models to determine the discontinuity depths. We used CRUST2.0 (Bassin et al., 2000) for the crustal structure and the global shear velocity model of Ritsema and van Heijst (2000) below the Moho. The average shear velocity perturbations above the discontinuities beneath the stations used in the present study ranged from $-0.1\%$ to $+1.0\%$ with respect to the Iasp91 model, and these values were used to determine the discontinuity depths. The average S-velocity anomalies of $+1.0\%$ above 410 and 660 km increased our estimates of the depths of “410” and “660” by $+5$ and $+8$ km, respectively. The discontinuity depths and their confidence levels were computed using the bootstrap method (Efron and Tibshirani, 1986), which randomly resamples (500 times) the receiver functions prior to stacking. See Gurrola et al. (1994) and Suetsugu et al. (2004, 2005) for more details.

3. Results

3.1. Depths of the “410” and “660”

First, we obtained the discontinuity depths by stacking the long-period receiver functions at each station. The functions were obtained using a Gaussian low-pass filter with a corner period of 4 s, which is comparable to the period range used by Shen et al. (2008). The stacked and depth-converted receiver functions are shown for
The “410” is generally elevated by 10–20 km where fast P-velocity anomalies were found and is within ±10 km of the average outside the high-velocity area (Fig. 5a), which suggests that the “410” depths in the studied region are controlled mostly by thermal anomalies associated with the subducted Pacific slab. An exception to this is MDJ, where the “410” is depressed by 9 km in the high-velocity area. The P410s signal is broadened or double-peaked at MDJ, which makes the depth estimate more uncertain than that for other stations. Shen et al. (2008) did not obtain “410” depths at MDJ, perhaps because the P410s signal is too weak (Fig. 2 in their paper). The “660” is generally depressed by 10–20 km in the high-P-velocity anomalies (Fig. 5b), which is consistent with Shen et al. (2008). Stations DL2 and INCN show nearly normal “660” depths. These stations are located in areas where high P-velocity anomalies are less pronounced ([35°N, 120°E] and [35°N, 125°E]), which suggests that the stagnant slab may have a gap and is not a continuous structure. The station BJT gives a depth of 686 km (26 km greater than the global average) for the “660”, despite the fact that the Ps conversion points are located near the edge of the fast anomalies. Although the deep “660” at BJT has been reported in previous studies (Niu and Kawakatsu, 1998; Shen et al., 2008), its interpretation remains unclear. The MTZ thickness (the “660” depth minus the “410” depth) is presented in Fig. 5c, and was more accurately determined than the “410” and “660” since an estimate of the MTZ thickness is less affected by possible errors in velocity correction above the “410.” P-velocity anomalies averaged in the MTZ are shown in the background of Fig. 5c, and in most of the studied region they are faster than the globally averaged MTZ velocity. The MTZ thickness is also thicker than the global average, which is in good agreement with the results of Shen et al. (2008).

3.2. Discontinuities below the “660”

As shown in Fig. 4, the “660” peaks and their codas are often broadened or composed of multiple peaks, the first of which are used in the “660” depth plot. We identified significant signals arriving after the “660” (indicated by dots in Fig. 4) as those with amplitudes above the 95% confidence interval. We did not consider later arrivals that are part of a long oscillation around the “660” to be significant, since they may represent a computational artifact (ringing) due to deconvolution. To see whether the broadened signals were composed of multiple signals, we computed the stacked depth-converted receiver functions with a short period filter (Gaussian low-pass filter with a corner period of 2 s). We employed the same shear velocity perturbations as those in the long-period stack (4 s) for the velocity corrections. The velocity perturbations in the P660s Fresnel zones (about 50 and 100 km in diameter at periods of 2 and 4 s, respectively) are almost the same, since the shortest wavelength of the shear velocity model of Ritsema and van Heijst (2000) used for velocity correction is 2000 km, which is much longer than the Fresnel zones. In the short-period stack (Fig. 6) we can see more isolated peaks at depths of 690–750 km than are shown in the long-period stack (Fig. 4). The long tails after the “660” signal at SNY and NNC in the long-period stack (Fig. 4) are considered to be a combination of the “660” signal and a later wave, as shown in Fig. 6. Depths corresponding to isolated signals, which are listed in Table 1, were determined by the bootstrap method. We call the discontinuity that generates the later wave the “720,” according to its average conversion depth. At MDJ, the “660” is accompanied by a long tail, which may also be the arrival of the “720”, though the separate signals are not clear. Fig. 7 shows the locations of the “720” conversion, from which we can observe a correlation between the conversion locations and the high-velocity anomalies at the top of the lower mantle that may represent the bottom part of the stag-
nent Pacific slab. The discontinuities located by the present study are plotted on vertical cross sections in Fig. 8. The “660” and “410” are detected either inside or outside of the high-velocity slab. The “660” depths are deeper inside the high-velocity slab than outside the slab. On the other hand, the “720” is located only inside the high-velocity slab, which implies that the “720” is closely related to the subducted slab. The “720” is not visible in the stacks of Shen et al. (2008). There are differences between the present study and that of Shen et al. (2008), e.g., the receiver function methods used in the present study and the selection of data and frequency, but it is not clear why we arrived at different degrees of visibility of the “720”.

4. Discussion

As shown in Fig. 5, a spatial correlation exists between the P-velocity tomograms and the discontinuity depths. To see this correlation quantitatively, we plotted MTZ thickness versus P-velocities in the MTZ, as shown in Fig. 5d. We can observe a positive trend (thicker MTZ at a higher P-velocity) with a correlation coefficient of 0.72. We compared the observed trend with that predicted from recent mineralogical results. Assuming the temperature dependence of the P-velocity to be $-3.0 \times 10^{-4} \text{km/s/K}$ for minerals in the MTZ (Irifune et al., 2008) and assuming the Clapeyron slope of the olivine to wadsleyite transformation to be 2.5 MPa/K (Katsura and Ito, 1989), we varied the Clapeyron slope of the post-spinel transformation from $-0.5$ to $-3.0$ MPa/K according to the experimental data (Ito and Takahashi, 1989; Irifune et al., 1998; Akaogi and Ito, 1993; Bina and Helffrich, 1994; Katsura et al., 2003; Fei et al., 2004; Litasov et al., 2005). The observed trend is consistent with those expected from the Clapeyron slope gentler than $-2.0$ MPa/K for the post-spinel transformation, which was obtained by experiments under dry conditions (Katsura et al., 2003; Fei et al., 2004). Fukao et al. (2009) ascertained the Clapeyron slope beneath the Philippine Sea region to be $-2.5$ MPa/K from P-velocity perturbation and the “660” depth in the lower part of the stagnant Pacific slab (Obayashi et al., 2006; Niu et al., 2005), a value which is slightly steeper than the slope of $-0.5$ to $-2.0$ MPa/K in the present study. This difference in slope may indicate that the MTZ beneath eastern China is relatively dry compared to that beneath the Philippine Sea, since the Clapeyron slope of the post-spinel transformation is gentler in a dry condition than it is in a wet condition (Litasov et al., 2005). The kinetic (non-equilibrium) effect could also be significant in a cold and dry condition (e.g., Wang et al., 1997; Kubo et al., 2002), a factor which is not addressed in the present study but which will be the subject of future studies.

Shen et al. (2008) observed negative Ps signals at a depth of around 600 km beneath many Chinese stations, which would represent a downward velocity reduction. Shen and Blum (2003) also found negative signals from 570 to 600 km beneath southern Africa. They suggest that these signals originate from a low-velocity zone at the base of the MTZ, which they attribute to the accumulated basaltic layer in the MTZ. We also see negative signals in the stacked receiver functions at some stations. They are distributed in a wide range of depths from 450 to 650 km and at stations located both inside and outside the Pacific slab. We cannot conclude that they are actual Ps signals from a low-velocity zone in the MTZ because they do not have spatial coherence. These negative signals may be due to reverberation phases from shallower discontinuities or may simply be a computational artifact of the deconvolution process. To identify the origin of these signals, further analysis is needed that is based upon more data from the denser seismic networks that are now being deployed. Our discussion will next focus upon the complex discontinuities around the “660”.

The complex discontinuities around the “660” have been extensively studied in the last decade. While most previous studies have

![Cross sections of P-velocity tomogram and the “410” (green), “660” (red), and “720” (yellow) discontinuities along two great circles shown in (c). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of the article.)](image-url)
reported multiple discontinuities around the “660” in subduction zones, some have reported similar multiples away from subduction zones (e.g., Simmons and Gurrola, 2000; Deuss et al., 2006; Andrews and Deuss, 2008). Niu and Kawakatsu (1996) first found complex discontinuities below the “660” beneath MDJ by using the receiver function method. They determined discontinuities at depths of 740 and 780 km in addition to the “660” at a depth of 670 km. Our own findings showed that the “660” signal at MDJ has a long tail, which may correspond to the 740 km discontinuity reported by Niu and Kawakatsu (1996). Ai et al. (2003) also employed the receiver function method to study mantle discontinuities and found multiple discontinuities at depths from 650 to 750 km beneath northeastern China (40–46°N, 126–132°E) using broadband array data. Their result is consistent with our observation of the “720” at SNY and MDJ. Ai and Zheng (2003) analyzed dense broadband array data from Shandon Province in eastern China (36°–37°N, 117–120°E) with the receiver function method. They found a discontinuity at around 700 km below the “660” in one area (34–36°N, 116–122°E). Chen and Ai (2009) reported similar deep discontinuities in the same area, based upon more array data than was used by Ai and Zheng (2003). Their result is consistent with the “720” found at TIA and HNS in the present study. Ai et al. (2003), Ai and Zheng (2003), and Chen and Ai (2009) interpreted the multiple discontinuities as a garnet-related phase transformation. A discontinuity that is deeper below the “660” has been detected in other subduction zones. Tibi et al. (2007) analyzed broadband data from island and ocean-bottom seismic stations in Mariana with the receiver function method. They found a discontinuity at depths from 740 to 770 km near the Mariana slab, which they attributed to the ilmenite to perovskite phase transformation. Schmerr and Garnero (2007) analyzed the precursors to SS waves in order to study mantle discontinuities in the South America subduction zone and found multiple discontinuities around the “660” in the vicinity of the subducted Nazca plate. One possible interpretation of the high visibility of the “720” signals in the subduction zones is the presence of a Ps wave that was converted from a later P wave due to multipathing caused by the subducted slabs. However, if the later P wave is present, it should also produce a Ps converted wave at the “410”, which should arrive after P410s, but this was not observed. For this reason we conclude that the “720” signal represents the Ps waves that were converted at the “720”. The discontinuities below the “660” can be attributed to the non-olivine components of the slab material, since there is no olivine-related phase transformation below the “660” (except for the post-perovskite transformation at the bottom of the mantle). Vacher et al. (1998) indicated that a garnet-related phase transformation should occur around the “660” in a low-temperature environment (1000 K), such as found in cold slabs. The garnet to ilmenite phase transformation should take place from 608 to 664 km for pyrolite composition, and the ilmenite to perovskite transformation should take place from 709 to 731 km. It is more likely that this latter transformation would be detected by the seismic waves used in the present study (wavelengths of 20–40 km) than the former, because the ilmenite to perovskite transformation is relatively sharp. In normal temperature conditions (1550–1600 K), a garnet-related transformation should occur at pressures close to the post-spinel transformation (Vacher et al., 1998), which could not be isolated seismologically. The depths of the discontinuities located by the present study (690–750 km) and their confinement in the lower mantle part of the cold slab indicate that the most likely origin of the deep discontinuities is the ilmenite to perovskite transformation. The depression of the ilmenite to perovskite transformation in the cold slab should cause a positive buoyancy of the slab in addition to that due to the depression of the “660”, which should contribute to slab stagnation in the MTZ.

5. Conclusions

We have estimated the depths of the mantle discontinuities from 400 to 750 km using broadband data from the entire region of eastern China and Korea, which enabled us to study the relationship between the discontinuity depths and the stagnant Pacific slab. Our results are summarized as follows:

(1) The depths of the “410” and “660” are generally shallower and deeper, respectively, than the global averages in and near the Pacific slab beneath eastern China. The MTZ is thicker in the slab than the global average. These observations are consistent with those reported by Shen et al. (2008). These observations are consistent with the thermally controlled olivine to wadsleyite transformation of the “410” and the post-spinel transformation of the “660.”

(2) There is another discontinuity at a depth from 690 to 750 km in the lower mantle part of the Pacific slab. Recent mineralogical experiments indicate that the most plausible interpretation of this deep discontinuity is the ilmenite to perovskite transformation.

(3) We observed negative signals between the “410” and “660”, as was also reported in previous studies. The origin of this negative phase remains unknown. The deployment of the denser broadband seismic network which is now being constructed should be useful for making detailed topography maps of the discontinuities identified above and for examining the significance and origin of these negative signals.

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