New constraints on structures of the mantle transition zone beneath the Trans-north China orogen and western north China craton revealed by receiver functions

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We calculated a total of 109,654 receiver functions from 848 teleseismic events recorded by two dense seismic arrays with 672 stations in the trans-north China orogen (TNCO) and western North China Craton (NCC). After moveout correction based on the 3-D crustal and mantle models, the common conversion point (CCP) stacking method was used to image the structures of the mantle transition zone (MTZ). The topographic reliefs of the 410-km and 660-km discontinuities exhibit significant spatial variations across the study area. The average 10 km depression of the 410-km discontinuity is observed under the regions from the Datong volcano to the Yin mountain, which may result from the hot mantle materials. The about 10 km uplift of the 410-km discontinuity and 15 km MTZ thickening are likely to be caused by the lithospheric delamination beneath the Abaga and Dalinor volcano regions. The almost flat topography of the two discontinuities and normal MTZ thickness of 250 km under the Ordos block and its southwestern regions indicate that the original structures of the MTZ are well maintained. Our obtained structures of the MTZ provide new constraints on the mantle dynamic process and tectonic evolution beneath the TNCO and western NCC.

1. Introduction

The mantle transition zone (MTZ) is the layer bounded by two global-scale velocity interfaces, 410-km and 660-km discontinuities (Shearer and Flanagan, 1999), and its structure has been widely used to understand the upper mantle dynamic processes (Liu et al., 2015; Zhang et al., 2016; Zuo et al., 2020). According to high-pressure and high-temperature experiments, 410-km discontinuity is associated with the transition of α-β phase olivine (Katsura and Ito, 1989), while 660-km discontinuity is caused by the transition from ringwoodite to perovskite plus magnesiowüstite (Ito and Takahashi, 1989). The depths of the 410-km and 660-km discontinuities vary with the phase transition temperature corresponding to positive and negative Clapeyron slopes. The shallower 410-km and deeper 660-km discontinuities (i.e. thicker MTZ) are often accompanied by cold regions in the MTZ, whereas thinner MTZ is expected to appear in hotter areas. In addition, previous studies suggested that a certain amount of water content can elevate the 410-km discontinuity and deepen the 660-km discontinuity (Litavin et al., 2005; Pearson et al., 2014). Therefore, the variations of the two seismic discontinuities, as well as the thickness of MTZ, can provide detailed information to constrain the deep mantle evolution, such as the delaminated lithosphere, subducted stagnant slab and hot mantle upwelling within the MTZ.

The North China Craton (NCC) is located in the eastern Eurasian plate and consists of the western NCC, trans-north China orogen (TNCO) and eastern NCC (Zhao et al., 2001). The central Asian orogenic belt, Qilian orogen and Qinling-Dabie orogen surrounded the NCC during the Paleozoic and Mesozoic in the north, southwest and south, respectively (Fig. 1). As the oldest craton in China, the NCC has experienced severe lithospheric rejuvenation from the late Mesozoic to Cenozoic (Wu et al., 2019; Zhu et al., 2011). In contrast to the stable craton characterized by the low heat flow (~40 mW/m²) and thick lithosphere (more than 180...
km), the eastern NCC is dominated by higher heat flow (greater than 65 mW/m²), thinner lithosphere (less than 100 km) and a series of magmatic activities (Chen, 2010; Zhu et al., 2011). Accordingly, the subduction and rollback of the Paleo-Pacific plate were proposed to understand the deep mantle process and dynamic mechanism of the lithospheric thinning under the eastern NCC (Wu et al., 2019; Zhu et al., 2011). Recent studies revealed that the TNCO and parts of the western NCC also suffered lithospheric modifications (Tang et al., 2021; Zhang et al., 2019). However, the knowledge on dynamic model of geological evolution beneath the TNCO and western NCC is still not sufficient enough because of the sparse seismic stations and spatially limited petrological sampling.

The large-scale high velocity anomalies imaged by P wave traveltime and waveform tomography were interpreted as the subducted Pacific stagnant slab inside the MTZ beneath the NCC (Huang and Zhao, 2006; Li et al., 2008; Liu et al., 2017; Tao et al., 2018). It was advocated that the dehydration of the stagnant slab would release the fluids, induce the unsteady mantle flow and cause the overlying lithospheric thinning by the erosion and metasomatism (Chen, 2010; Zhu et al., 2011). In addition, the lower mantle plume mechanism was proposed to explain the lithospheric modification beneath the NCC (Lei, 2012). Obviously, these hypotheses are controversial. To demonstrate the mantle dynamic process of the NCC, it is essential to further obtain the high-resolution structures of the MTZ. However, seismic tomography is not sensitive to the discontinuities of the Earth’s interior. In contrast, receiver function imaging is an effective method to investigate the topographic

![Map of tectonic setting and seismic stations used in this study.](image-url)

Fig. 1. Map of tectonic setting and seismic stations used in this study. The red and black triangles indicate the seismic stations of the ChinArray phase II and III deployments, respectively. The blue triangle represents station 62,409 as an example of showing the receiver functions in Fig. S1. The red volcanic symbols denote the Datong (DTV), Abaga (ABV) and Dalinor (DLV) volcanoes. The thick gray line is the north–south gravity line (NSGL). CAOB: central Asian orogenic belt; YM: Yin mountain; AB: Alxa block; HG: Hetao graben; QO: Qilian orogen; OB: Ordos block; LM: Lüliang mountain; TM: Taihang mountain; BBB: Bohai bay basin; SGT: Songpan-Ganzi terrane; QDO: Qinling-Dabie orogen; SSG: Shanxi-Shanxi graben; SB: Sichuan basin. Inset shows the location of the study area with the blue rectangular. WNCC: western north China craton; TNCO: trans-north China orogen; ENCC: eastern north China craton; CAOB: central Asian orogenic belt; PHSP: Philippine Sea plate; PP: Pacific plate.
variations of the 410-km and 660-km discontinuities. Previous studies using receiver function have revealed that the two discontinuities showed significant lateral variation beneath the eastern NCC (Sun et al., 2020; Zhang et al., 2016; Zuo et al., 2020). Their proposed dynamic models based on the thickness of the MTZ agreed that the subduction of the Paleo-Pacific plate was the main cause of the lithospheric thinning under the eastern NCC. The detailed topographic reliefs of the two discontinuities beneath the TNCO and western NCC are still absent due to the lack of dense seismic array data. New constraints on the structures of the MTZ need to be revealed to discuss the geological evolution and dynamic mechanism of the mantle lithosphere beneath the TNCO and western NCC.

The ChinArray project (ChinArray-Himalaya, 2011) organized by the Institute of Geophysics, China Earthquake Administration, has deployed two dense broadband seismic arrays in the TNCO and western NCC. In this study, we collected teleseismic waveform data recorded by these two seismic arrays (Fig. 1) and performed a common conversion point (CCP) stacking of the receiver function to obtain the refined structures of the MTZ under the TNCO and western NCC. Then, they are followed a discussion of some implications on dynamic mechanism of lithospheric deformation and volcanic activities in the study area.

2. Data and methods

2.1. ChinArray phase II and III deployments

The second and third phase deployments of the ChinArray program (ChinArray-Himalaya, 2011) were composed of more than 1000 temporary broadband seismic stations with an average station spacing of ~ 40 km. Each seismograph is equipped with a Guralp CMG-3ESP or CMG-3ESP seismometer and a Reftek 130 data acquisition system. The second and third phases of ChinArray deployment rolled from the western NCC to the TNCO and provided an excellent opportunity to reveal the structures of MTZ in this area. In this study, we selected a total of 672 seismic stations (Fig. 1), including 385 stations of the second phase deployed from October 2013 to April 2016 and 287 stations of the third phase deployed between May 2016 and January 2019.

By visually examining all the data from earthquakes with a magnitude greater than 5.0 and an epicentral distance between 30° and 90°, we chose a total of 848 earthquakes that were shown in Fig. 2. Although most of these teleseismic events are located inside the western Pacific subduction zone and the Java trench, all distances and azimuths are well covered (Fig. 2).

2.2. Receiver functions

To estimate receiver functions from the three-component seismic recordings, we first calculated the true orientation of each station with the method of the P wave particle motions (Niu and Li, 2011) to eliminate the potential misorientation of the seismometers (Zeng et al., 2020). A 50 s to 5 Hz zero-phase filter was also used to remove both long period and high frequency noise of the three-component seismograms. Then, we rotated the two horizontal components of the seismograms into the radial (R) and transverse (T) directions based on the great arc ray paths connecting the events and stations. Next, we further projected the R and vertical (Z) components of the P waves to the principal

Fig. 2. Earthquake locations used in this study. The blue triangle represents the center of the study area. Note that although some events are located inside 30° and outside the 90° circle, the epicentral distance of all the seismograms is between 30° and 90°. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
directions (longitudinal, P, and in-plane transverse, SV) based on the covariance matrix (Niu et al., 2007). Then, receiver functions were computed from the data projected into this coordinate system (Niu et al., 2007; Vinnik, 1977). Here, the “water-level” deconvolution technique (Ammon, 1991; Clayton and Wiggins, 1976) was used to calculate receiver functions:

\[ RF(\omega) = \frac{V(\omega) \cdot P(\omega)}{\max\{|P(\omega)| \cdot \gamma \cdot |P_{max}(\omega)|\}} e^{-|\gamma|} \] (1)

Here, \( \gamma \) and \( \alpha \) are two constants that denote the “water level” and the corner frequency of the Gaussian low pass filter. We set them to be 0.01 and 1.5 (\( \sim 0.75 \) Hz), respectively. We use \( P(\omega) \) and \( V(\omega) \) to present the spectra of the P and SV components in the time window (50 s before and 150 s after the P wave). After selecting the receiver functions with high signal to noise ratio (SNR), a total of 109,654 receiver functions from the 672 stations were used in this study, with a station average of about 163 receiver functions. As an example shown in Fig. S1, the receiver functions were plotted as a function of epicentral distance at station 62,409 located in the Ordos block (Fig. 1).

2.3. CCP gathering

The CCP stacking technique (Dueker and Sheehan, 1997; Niu et al., 2005) was employed to image the topography of the 410-km and 660-km discontinuities and the thickness of the MTZ. Firstly, the study area (30’N to 46’N, 99’E to 119’E) was divided into parameterized grids of 0.2’ \times 0.2’ . The circular raypath with a radius of 1’ was used to gather the receiver functions and a total of 8181 (81 \times 101) caps were obtained. According to the raytrace of the Pds conversion phase, we calculated the relative arrival time of the P-to-S conversion waves (Pds) with respect to the direct P arrival, for an assumed conversion depth \( d \), and the corresponding converted location of each receiver function. The tables contained the Pds moveouts and conversion points were constructed at a set of grids \( (d) \), interval defined as 1 km, in the depth range of 200 to 800 km. Then, we gathered the receiver functions in each cap, which is based on the computed Pds conversion coefficient and the Pds conversion points located inside the cap. The hit counts in each cap at the 410-km and 660-km discontinuities are demonstrated in Fig. 3.

1-D IASP91 velocity model (Kennett and Engdahl, 1991) was adopted to compute the 1-D Pds moveouts in the raytracing Pds. To improve the accuracy in estimating the Pds moveouts, we further used the crustal and mantle corrections with the crustal thickness model (Li et al., 2014; Xu et al., 2021; Xu et al., 2018a) and the 3-D EARA2014 model (Chen et al., 2015), respectively, to reduce or eliminate the lateral variations in crustal and mantle structures. We followed the steps of Liu et al. (2015) to determine the time residual of the Pms with respect to the IASP91 model. More details about this crustal correction are described in Liu et al. (2015). To calculate the mantle corrections on the Pds moveouts, we recomputed the traveltime by projecting the 1-D ray paths into the blocks (50 km \times 50 km \times 10 km) of the 3-D ERA2014 model. Then, we added the segment traveltime and obtained the 3-D Pds moveouts along the ray paths. In addition to the Pms time residuals, the new tables of the 3-D Pds moveouts and their geographic location of the conversion points from the 1-D IASP91 model were reconstructed for summing the receiver functions.

The \( N^\theta \)-root stacking technique (Muirhead, 1968) was applied to stack the receiver functions. Here, \( n(t) \) indicates the \( k^{th} \) receiver function summed in the \( i^{th} \) cap and \( t_d \) denotes the Pds arrival time at depth \( d \). Then, the \( N^\theta \)-root stacking, \( R_i(d) \), can be expressed by:

\[ R_i(d) = y_i(d)/|y_i(d)|^{N-1} \] (2)

\[ y_i(d) = \frac{1}{K} \sum_{k=1}^{K} w_k \exp \left\{ \frac{1}{a^2} \left[ \sum_{n=1}^{N} \exp \left( -x_n \right) \right] \right\} \] (3)

Here \( K \) is the total number of receiver functions gathered at the \( i^{th} \) cap, and \( w_k \) is a Gaussian weight function:

\[ w_k(d) = \exp \left( -\frac{x_k^2}{a^2} \right) \] (4)

Here, \( x_k \) presents the distance from the cap center to the conversion point for the \( k^{th} \) receiver function and \( a \) is the Gaussian width parameter that is the same as the cap radius. In contrast to the linear stack \( (N = 1) \), we set \( N = 4 \) to suppress the irrelevant noise. Fig. 4 shows three profiles of stacked receiver functions based on the 3-D Pds moveout tables along the section lines shown in Fig. 3a.

2.4. Reliability analysis

The reliability of the CCP stacking results depends mainly on the quality and quantity of receiver functions and the reference velocity model used in raytracing. We first employed the method proposed by Chen et al. (2010) with the cross-correlation coefficient (\( > 85 \% \)) for each pair of receiver functions to remove the noisy receiver functions. After the strict selection, we further removed the receiver functions with low SNR and coherence by visually examination. With the above processing, we obtained a high-quality dataset of 109,654 receiver functions. In Fig. S1, we can see clearly the direct P phase, P-to-S converted
phase from the Moho (Pms) and its reverberation phase (PpPs), and P-to-S converted phases from 410-km (P410s) and 660-km (P660s) discontinuities.

In addition, we adopted the Nth-root stacking technique in performing the CCP stacking at each cap. Taking account of both amplitude and coherence of the conversional signals, the 4th-root stacking, compared to the linear stacking, at the high coverage density of the conversion points ensured the robustness of the CCP imaging (Fig. S2). The crustal thickness model derived from receiver functions (Li et al., 2014; Xu et al., 2021; Xu et al., 2018a) and 3-D EARA2014 model inverted by adjoint tomography (Chen et al., 2015) were also used to improve the accuracy in migrating the receiver function data. An inaccurate reference model will introduce significant deviations of the Pds moveouts and their geographic location of conversion points due to the lateral heterogeneity of the upper mantle velocity structure (Guan and Niu, 2018; Liu et al., 2015). For example, increasing the depth of the 410 km discontinuity by 10 km or reducing the velocity in the reference model by 3.5% can lead to an increase of 1.1 s in the traveltime of the receiver function at the 410 km discontinuity (Yang and Zhou, 2001). Zuo et al. (2020) further suggested that the differences of the CCP stacked depths based on the models of the EARA2014 (Chen et al., 2015) and PWEA18 (Tao et al., 2018) were only within one standard deviation. By comparing 1-D migration, approximately 10% higher amplitudes can be found in the conversion phases of P410s and P660s estimated from the 3-D moveouts (Fig. S3), which indicates that the EARA2014 model is close to the true velocity structures beneath this study area and leads to stronger Pds conversion phase in the CCP stacking. Fig. S4 further shows the improvements of 3-D migration in both amplitudes of conversion phases and depths of 410-km and 660-km discontinuities. Therefore, we did not perform Pds moveouts with different 3-D velocity models to demonstrate the uncertainties of the 410-km and 660-km discontinuity depths, which is indeed the corresponding limitations of this study.

The number of stacked receiver functions in each cap also plays a key role in the reliability of the obtained 410-km and 660-km discontinuity depths. In this study, we only performed the CCP stacking at caps with the sampling number of the conversion points more than 200 at depths of 410-km and 660-km (Fig. 3). Except for the edge of the study area, most of regions show the dense ray piercing points with more than 1000, which ensures the robustness of the observed results. In order to reduce the measurement uncertainties of the 410-km and 660-km discontinuities, we picked the maximum amplitudes and the average of the positive amplitudes of the P410s and P660s conversion phases to represent the depths of the two discontinuities, respectively. The small standard deviation of the depths at the 410-km and 660-km discontinuities (Fig. S5) indicated that the results of the two measurements have good consistency and reliability.

3. Results

Fig. 4 shows three profiles of the CCP gathered receiver functions using the 3-D crustal and mantle models. The variations of the 410-km and 660-km discontinuities are revealed by strong positive amplitudes of P410s and P660s conversion phases from each profile. Fig. 5 shows the topography of the 410-km and 660-km discontinuities as well as the MTZ thickness based on both the 1-D IASP91 model and 3-D EARA2014 model, which is picked by the depths of the maximum amplitudes of the P410s and P660s conversion phases within the respective depth ranges of 390–430 km and 630–690 km. Both uplift and depression of two discontinuities are marked by the rectangles, circles and ellipses in Fig. 5. We also estimated the average values for the anomalous regions in Table S1.

The average depths of the two discontinuities are 414.5 ± 6.4 km and 669.1 ± 7.5 km for the 1-D IASP91 model (Fig. S6a-b), respectively. We find that the two discontinuities estimated using the 1-D IASP91 model are significantly deeper than normal, especially for the regions from the Bohai bay basin to the Datong volcano. After the 3-D velocity corrections, the average depths of the two discontinuities become shallow with 412.7 ± 4.8 km and 663.5 ± 2.9 km (Fig. S6c-d), respectively. The standard deviations for the 3-D EARA2014 model are simultaneously smaller than that for the 1-D IASP91 model (Fig. S6e). The depressed absolute depths of the two discontinuities are corrected by more than 20 km under the Bohai bay basin to the Datong volcanic region, which suggests that the 3-D EARA2014 model is dominated by low velocity anomalies beneath the Bohai bay basin to the Datong volcanic region. The corrected depth of the 660-km discontinuity even appeared to be closed to the normal depth of 660 km, showing relatively flat topography (Fig. S6). These observations indicate that the strong lateral velocity heterogeneities exist in the upper mantle of the TNCO and western NCC.

The 410-km discontinuity depressed more than 10 km in area A, marked by the square, representing mainly the Yin mountain and its surrounding regions (Table S1, Fig. 4a and Fig. 5d). In contrast, the depth of the 660-km discontinuity is almost consistent with the global average of 660 km beneath this area (Fig. 5e). Under the Abaga volcano and Dalinor volcano, circled by the area B, the striking uplift of the 410-km discontinuity is up to 10 km (Table S1, Fig. 4a and Fig. 5d). The 660-km discontinuity exhibits about 10 km of fluctuation in this area. The two discontinuities of the Datong volcanic region, labelled with ellipse C, present a quite different distribution pattern, depressing simultaneously for approximately 10 km (Table S1, Fig. 4b and Fig. 5d-e). The slight uplift and depression of the 410-km and 660-km discontinuities
with about 5 km are observed beneath the western Bohai bay basin region (area D) (Table S1, Fig. 4c and Fig. 5d-e). However, the southern TNCO and western NCC are dominated by the nearly flat 410-km and 660-km discontinuities (Fig. 4 and Fig. 5d-e).

The MTZ thickness is more accurate than the absolute depths of the 410-km and 660-km discontinuities, thanks to eliminating the upper mantle heterogeneity of the velocity structure for either the 3-D EARA2014 model or the 1-D IASP91 model in migration. Therefore, maps of the MTZ thickness obtained using the 3-D EARA2014 model and the 1-D IASP91 model are roughly similar (Fig. 5c,f), except for that in the area A. We notice that areas B and D are characterized by the dramatically thinker MTZ shown in Fig. 5c and f. In contrast, the thinner MTZ of area A can be only found in Fig. 5f, which indicates that the MTZ of area A is featured by the lower velocity anomaly in the 3-D EARA2014 model relative to the 1-D IASP91 model. The MTZ thickness in most other regions shows a uniform distribution with the global average of 250 km. Another noteworthy feature is that the topographic relief of the MTZ thickness presents more positive correlation with that of the 410-km discontinuity based on the 3-D velocity model (Fig. 5d,f).

The depth sections of the P-wave velocity perturbation (Tao et al., 2018) were plotted in Fig. S7 to compare with the CCP imaging along the sections of AA', BB' and CC' shown in Fig. 3a. We find that the depression and uplift of the 410-km discontinuity correspond well to the low and high velocity anomalies located in the areas A and B.
respectively. Besides, the deeper 660-km discontinuity matches the high velocity anomaly inside the MTZ. Throughout Fig. 4, there is no positive amplitude at depth of around 520 km, which suggests that the 520-km discontinuity is not continuous and difficult to observe. The possible double-branched 660-km discontinuity mentioned by Wang and Niu (2011) also cannot be imaged in this study.

4. Discussion

4.1. Comparisons with previous studies

The MTZ structures beneath the NCC have been investigated by many previous studies (e.g., Chen and Ai, 2009; Chen et al., 2015; Tao et al., 2018; Zuo et al., 2020). In the eastern NCC, earlier works (Chen and Ai, 2009; Wang and Niu, 2011; Zuo et al., 2020) estimated the depths of the 410-km and 660-km discontinuities as well as the MTZ thickness from the CCP stacking of the receiver functions. Zhang et al. (2016) and Sun et al. (2020) also obtained the MTZ structures but paid more attention to the results in northeastern China. Across the TNCO and western NCC, the 410-km and 660-km discontinuities can be found at the average depths of 412.7 km and 663.5 km, respectively, estimated from the 3-D migration. These two values suggest a different difference in estimates of Zuo et al. (2020), with the average depths of 416.9 km and 676.3 km. The deeper 660-km discontinuity, more than 10 km, may be related to the cold Pacific slab in the MTZ beneath the eastern NCC.

In contrast to the dense seismic arrays deployed by the ChinArray project in this study, the results mentioned above only used the seismic network consisting of a few temporary stations or the permanent stations of the China Earthquake Administration. The uneven distribution of stations with larger station spacing resulted in relatively low lateral resolution, which may be the main reason for the difference between previous results and our measurements. For example, the structures of the 410-km and 660-km discontinuities in the area A were almost absent from previous studies due to the lack of seismic station coverage and mainly demonstrated by tomographic images (e.g., Chen et al., 2015; Tao et al., 2018; Xu et al., 2018b). The depths of the two discontinuities under the area B could not be resolved well by the results of Sun et al. (2020) and Zuo et al. (2020), Liu et al. (2015) and Zhang et al. (2016) revealed the topographies of the two discontinuities in parts of the area B and suggested similar characteristics to our results. Both the depth of the 660-km discontinuity and the thickness of the MTZ beneath the Bohai bay basin (area D) measured by Sun et al. (2020) are more than 10 km larger than that in our images (Fig. 5e-f). Our observed thickness of the MTZ in the Ordos block is roughly consistent with the earlier results (Chen and Ai, 2009; Wang and Niu, 2011), but with better resolution for the TNCO and western NCC. Overall, we can provide new constraints on the structures of the MTZ beneath the TNCO and western NCC based on the more refined topographic reliefs of the two discontinuities in this study.

4.2. Mechanisms of the intraplate volcanoes

A series of volcanic activities have occurred in the west of the north–south gravity line since the Neogene (Fan et al., 2015; Xu et al., 2005). Most of these volcanoes (e.g., Abaga volcano, Dalinor volcano and Nuomin River volcano) are located in the Central Asian orogenic belt and its northern regions (Fan et al., 2015). However, the Datong volcano is one of the most concerned volcano in the NCC. At present, the Datong volcano located in the northern TNCO has attracted extensive attention due to its significant surface geomorphic characteristic and controversial formation mechanisms (Guo et al., 2014; Lei, 2012; Ma et al., 2019). As the best preserved and most spectacular Quaternary volcanic group in the NCC, the Datong volcano, consisting of more than 30 volcanic cones, was formed during the early Pleistocene and late Pleistocene, 0.74–0.20 Ma, (Fan et al., 2015; Xu et al., 2005). The eastern and western Datong volcanic fields are dominated by different mineralogy, tholeiitic basalts in the east and the alkaline basalts in the west, respectively.

Tomographic images indicated that the Datong volcano originated from the asthenospheric upwelling (Xu et al., 2018b; Yao et al., 2020). However, we still cannot conclude its deeper origin, such as the mantle plume (Lei, 2012) and the upwelling flow triggered by the stagnant Pacific slab in the MTZ (Ma et al., 2019). Lei (2012) observed a Y-shaped low velocity anomaly and suggested the lower mantle plume contributed to the Datong volcano, which is supported by recently published results (Kimura et al., 2018). More other seismic images showed that the low velocity anomalies beneath the Datong volcano only extend into the MTZ, but not the lower mantle (e.g., Huang and Zhao, 2006; Liu et al., 2017; Tao et al., 2018; Xu et al., 2018b). Ma et al. (2019) speculated that the Datong volcanism is caused by the upwelling flow triggered by the westward push of the stagnant slab based on their tomographic results. Thus, it is essential to obtain more measurements to discuss the origin of the Datong volcano.

The topographic variations of the 410-km and 660-km discontinuities play an important role in revealing the hot asthenospheric upwelling, and have been widely used to research the volcanic activities (Liu et al., 2015; Sun et al., 2020; Zhang et al., 2016). In this study, we estimate the depth of the 410-km and 660-km discontinuities from the receiver function CCP stacking. The topographic relief of the two discontinuities under the Datong volcano region simultaneously deepens almost 10 km but is dominated by a normal MTZ thickness of approximately 250 km(Fig. 5d-f and Table S1). These estimates are in conflict with the hypotheses of either the lower mantle plume, leading to a deep 410-km discontinuity and shallow 660-km discontinuity, or the cold stagnant Pacific slab, forming opposite depths of the two discontinuities, within the MTZ. In addition, it is worth noting that the depressed 410-km discontinuity extends northwestward from the Datong volcano region to the Yin mountain (area C to A, Fig. 5d). Previous studies suggested that the MTZ could have a strong water storage capacity and play a key role in terrestrial magmatism (Bercovici and Karato, 2003; Pearson et al., 2014). If this is the case, the water-rich MTZ may induce the partial melting and enrich the hot materials near the 410-km discontinuity. According to the S wave velocity perturbation estimated from full-waveform inversion (Chen et al., 2015), we followed the equations of Meier et al. (2009) to calculate the temperature variation and water content relative to the global average inside the MTZ (Table S1). The average water content of +0.2 wt% can be found in the regions from the Datong volcano region to the Yin mountain (area C to A). Moreover, using the scaling parameter of $dV_P/dT = -4.8 \times 10^{-4} \text{km s}^{-1} \text{C}^{-1}$ (Deal et al., 1999), the negative 0.8–0.9 % P wave velocity anomaly (Chen et al., 2015) at the 410-km discontinuity under these regions is equivalent to a positive temperature anomaly of about 170–190 K, producing more than 15 km depression of the 410-km discontinuity for a Clapeyron slope of +2.5 MPa/K (Bina and Helffrich, 1994; Katsura and Ito, 1989). This overestimation of about 5 km may be mainly due to the existence of the water content at the 410-km discontinuity. The elongated low velocity anomaly beneath the Datong volcano was observed from shallow downward to the 410 km depth of area A and C (Tao et al., 2018; Xu et al., 2018b). Thinner lithosphere with the thickness of less than 100 km under the area A and C further imply that the hot mantle upwelling could severely modify the overlying lithosphere (Zhang et al., 2019). These geophysical observations also agrees well with the geochemical interpretation that the lava of the Datong volcano originated from the deep mantle asthenosphere (Fan et al., 2015; Xu et al., 2005). Therefore, the variations of the 410-km discontinuity beneath the area A and C are the isogenesis that may be induced by the hot mantle materials. Combined with these results mentioned above, we propose that the origin of the Datong volcano may be derived from the hot mantle materials rooted near the 410-km discontinuity beneath the local areas of A and C (Fig. 6).

In addition, the apparent uplift of the 410-km discontinuity, more than 10 km, and thickening of the MTZ, approximately 15 km, can be
found in the Abaga volcano and Dalinor volcano regions (Fig. 5d,f and Table S1), which agrees well with the measurements of Liu et al. (2015) and Zhang et al. (2016). Assuming the Clapeyron slope of +2.5 MPa/K (Bina and Helffrich, 1994; Katsura and Ito, 1989), the 10 km uplift of the 410-km discontinuity corresponds to a 100 K temperature decrease. Associating with the high velocity anomalies shown with the background in Fig. S7a and low temperature in the MTZ (Table S1), this pattern is completely inconsistent with that of classic volcanoes caused by a hot mantle plume, such as the Hawaii volcano with the significantly low velocity in the MTZ (Lei and Zhao, 2006) and depressed 410-km and uplift 660-km discontinuities (Agius et al., 2017). The seismic tomographic images suggested that the western edge of the subducted Pacific plate (Fig. S7a) with obvious high velocity anomaly did not reach the Abaga volcano and Dalinor volcano regions (Liu et al., 2017; Tao et al., 2018). The uplift of the 410-km discontinuity as well as the high velocity anomaly beneath Abaga volcano and Dalinor volcano regions are unlikely to be caused by the cold Pacific slab. Chen (2010) and Zhang et al. (2019) revealed that the lithosphere has been thinned to approximately 100 km beneath the Abaga volcano and Dalinor volcanic regions. The ongoing lithospheric delamination was proposed by various seismic observations under these two volcanic regions (Chen et al., 2017; Zuo et al., 2020). Therefore, we speculate that the delaminated lithosphere maybe sink into the MTZ, resulting in the observed high velocity anomaly and uplift of the 410-km discontinuity. It is this sinking and locally retained lithosphere that may have formed the significantly different structures beneath the Abaga volcano and Dalinor volcano. The hot mantle materials caused the delamination further percolated upward through the weak overlying lithosphere and led to the eruption of the Abaga volcano and Dalinor volcano.

4.3. Dynamic model beneath the NCC

Geodynamic process plays an important role in revealing the lithospheric deformation and tectonic evolution. A reasonable dynamic model is helpful for us to understand the lithospheric transformation and rejuvenation of the NCC (Tang et al., 2021; Zuo et al., 2019). Recent studies on the lithospheric thickness showed that the mantle lithosphere of the NCC has been modified based on the gradual thinning of the lithosphere from west to east (Chen, 2010; Zhang et al., 2019). Seismic tomography and xenoliths suggested the large-scale low velocity anomalies in the shallow asthenosphere and widely distributed mantle-derived volcanic activities in the eastern NCC (Li et al., 2008; Tang et al., 2021; Tao et al., 2018; Xu et al., 2018b; Xu et al., 2005). Accordingly, many models provided in previous studies concentrated on the upper mantle structures of the eastern NCC (Tang et al., 2021; Zhu et al., 2011). A common understanding on the rejuvenation of the eastern NCC is related to the subduction and rollback of the Paleo-Pacific plate that the dehydration of the stagnant slab resulted in the fluid metasomatism, induced the unsteady mantle flows above the subduction zone and caused the overlying lithospheric thinning (Tang et al., 2021; Zhu et al., 2011). However, little attention has been paid to the cratonic evolution in the TNCO and western NCC. The corresponding tectonic process under the TNCO and western NCC is still unclear. The variations of the 410-km and 660-km discontinuities can characterize the temperature and water content of the MTZ (Meier et al., 2009; Pearson et al., 2014; Suetsugu et al., 2010), which is an indicator to constrain the structures of the MTZ and its above asthenosphere and lithosphere. In this study, the topographic relief of the 410-km and 660-km discontinuities provides another perspective to help us discuss the lithospheric modification and mantle evolution beneath the TNCO and western NCC. The variation of the 410-km and 660-km discontinuities presents obviously different patterns under the northern and southern TNCO. In contrast to the two discontinuities that are almost close to the global average beneath the southern TNCO, the depressed 410-km and 660-km discontinuities can be found under the northern TNCO (Fig. 5d-e and Fig. S7b). Taking into account the estimates of the temperature variation and water content (Table S1), we have proposed the existence of the hot mantle upwelling beneath the northern TNCO. Results of low velocity anomalies (Xu et al., 2018b; Yao et al., 2020) and lithospheric thickness (Zhang et al., 2019) further confirmed that this hot mantle upwelling is responsible for the lithospheric modification beneath the northern TNCO. Moreover, about 10 km thickening of the MTZ occurs in the western Bohai bay basin (area D). This thickening arises from the elevated 410-km discontinuity and deepened 660-km discontinuity with an average of approximately 5 km (Table S1), respectively. The magnitudes of these anomalies are much smaller than those observed in eastern NCC and northeastern China caused by the cold Pacific slab, with more than 10 km uplift of the 410-km discontinuity, 20 km depression of the 660-km discontinuity and 30 km thickening of the MTZ (Liu et al., 2015; Zhang et al., 2016; Zuo et al., 2020). The high velocity anomalies related to the cold stagnant slab were revealed by previous tomographic images (Liu et al., 2017; Tao et al., 2018), which...
is far from the area D (Fig. 5c). The topographic reliefs of the two discontinuities and Pacific stagnant slab are spatially uncorrelated, providing no evidence for the cold stagnant slab contributing directly to the uplift of the 410-km discontinuity and depression of 660-km discontinuity. The anomalies of the two discontinuities observed in area D are likely to be due to the local high velocity anomalies inside the MTZ (Fig. 5). In contrast to the significant fluctuation of the 410-km and 660-km discontinuities in the regions of A-D, the almost flat topographies of the two discontinuities and normal MTZ thickness of about 250 km are dominated in the western NCC (Fig. 5d-e). This phenomenon means that the original structures of the MTZ in these regions are well maintained without any distinct deep mantle tectonic activities.

According to the above discussion and previous published results, we propose an upper mantle dynamic model shown in Fig. 6 to understand the structures of the MTZ under the TNCO and western NCC and its implication for the lithospheric modification. Specifically, it seems to indicate the scenarios: the Pacific subduction is unlikely to cause the topographic anomalies of the 410-km and 660-km discontinuities under the TNCO and western NCC; the water-rich MTZ contributed to the hot mantle materials and resulted in the depression of the 410-km discontinuity beneath the areas from the Datong volcano to the Yin mountain; the upwelling of these hot mantle materials modified the overlying lithosphere, caused the lithospheric thinning beneath the northern TNCO, and then further intruded into the weak lithosphere to feed the lava eruption of the Datong volcano during the early Pleistocene and late Pleistocene; the southern TNCO and the WNCC still preserve their original structures of the MTZ, which helps to maintain their thicker lithosphere.

5. Conclusions

To better understand the deep mantle structures beneath the TNCO and western NCC, we used 109,654 high-quality receiver functions to image the topography of the 410-km and 660-km discontinuities with the CCP stacking method. The 3-D crustal and mantle models were employed to improve the accuracy of the Pds moveouts, which reduced the effects of upper mantle velocity heterogeneity. Our results reveal the detailed topographic relief of the 410-km and 660-km discontinuities, which provides new constraints on the structures of the MTZ beneath the TNCO and western NCC. The depression of the 410-km discontinuity under the regions from the Datong volcano to the Yin mountain may be related to the hot mantle materials at the 410-km discontinuity. The upwelling of this hot mantle materials leads to the overlying lithospheric modification and thinning beneath the northern TNCO. The discontinuities of the the 410-km and 660-km are almost close to the global average beneath the southern TNCO and western NCC, which suggests that their original structures of the MTZ are retained.

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Appendix A. Supplementary material

Supplementary data to this article can be found online at https://doi.org/10.1016/j.jseaes.2023.105554.

References
