Ambient noise Love wave tomography of China

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Abstract: We first report on the Love wave tomography of China based on ambient noise cross-correlations. We used 3 years of continuous waveform data recorded by 206 broadband seismic stations on the Chinese Mainland and 36 neighboring global stations and obtained Love wave empirical Green’s functions from cross-correlations of the horizontal components. The Love wave group velocity dispersion measurements were used to construct dispersion maps of 8- to 40-s periods, which were then inverted to obtain a three-dimensional horizontally polarized S-wave (SH) velocity structure. The resolution was approximately 4° × 4° and 8° × 8° for eastern and western China, respectively, and extended to a depth of approximately 50 km. The SH model was generally consistent with a previously published vertically polarized S-wave (SV) model and showed large-scale features that were consistent with geological units, such as the major basins and changes in the crustal thickness across the north-south gravity lineament. The SH and SV models also showed substantial differences, which were used to examine the subsurface radial anisotropy. We define the radial anisotropy parameter as
\[ \psi = \frac{2(V_{SH} - V_{SV})}{(V_{SH} + V_{SV})}. \]
At a shallow depth, we observed significant radial anisotropy under major basins, which may be related to thicker sedimentary layers. At the mid to lower crust, most of the Chinese continent showed strong positive radial anisotropy (SH > SV). Central and southern Tibet showed strong positive anisotropy, whereas the radial anisotropy was relatively weak at the northern and eastern margins, which suggests a change in deformation style from the plateau interior to its margins. The North China craton showed prominent positive radial anisotropy, which may be related to decratonization and strong extension since the Mesozoic Era. Love waves are less well retrieved than Rayleigh waves from ambient noise cross-correlations. Increasing the duration of the cross-correlation data beyond 4 to 8 years may not aid in retrieving Love waves of longer periods, for which improved methods need to be explored.

Keywords: ambient noise tomography; Love wave; radial anisotropy; China

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1. Introduction

The Love wave is a type of surface wave that is polarized in the horizontal direction. Its propagation speed is related to the horizontally polarized S-wave (SH) velocity. The dispersion (frequency dependence of the wave speed) of the Love wave can be inverted to obtain the subsurface SH wave velocity structure. The traditional surface wave tomography method utilizes the waveform data excited by earthquakes and is limited by the spatial distribution of the earthquake sources. Ambient noise tomography (ANT) is a relatively recent method that uses interferometry of the Earth’s ambient vibrations between seismic stations and does not rely on energetic earthquakes.

The theoretical basis for ANT is that the cross-correlation of a diffuse field can yield a Green’s function between a pair of receivers (Lobkis and Weaver, 2001; Weaver and Lobkis, 2004; Shapiro and Campillo, 2004). To date, the great majority of ANT studies have focused on Rayleigh waves. Shapiro et al. (2005) measured dispersions of the Rayleigh wave empirical Green’s functions (EGFs) obtained from ambient noise cross-correlations to obtain the surface wave group velocity structure of California, USA. The ANT method has since been widely demonstrated and applied (e.g., Yao HJ et al., 2006, 2008; Zheng SH et al., 2008; Lin FC et al., 2010; Yang YJ et al., 2010; Lin FC and Ritzwoller, 2011). The method is applicable to tomography on a continental scale, such as on the European continent (Yang YJ et al., 2007), in the United States (Bensen et al., 2008), and in China (Zheng SH et al., 2008; Sun XL et al., 2010). Rayleigh wave dispersion tomography can be used to invert three-dimensional (3-D) vertically and radially polarized S-wave (SV) velocity structures. For example, 3-D SV velocity structures were obtained in the eastern margin of the Tibetan Plateau (TP; Yao HJ et al., 2008), North China (Zheng Y et al., 2011), South China (Zhou LQ et al., 2012), and the northeastern margin of the TP (Li HY et al., 2012, 2014). With the accumulation of seismic data, Bao XW et al. (2015) and Shen WS et al. (2016) used the continuous waveform data from more than 1,000 permanent seismic stations and hundreds of portable seismic stations in China to obtain a high-resolution SV velocity structure of the crust and uppermost mantle in China.
Application of the noise correlation method to obtain the EGFs of Love waves has been much more limited. Obtaining reliable Love wave EGFs from noise correlations is more difficult and requires much more computation time. Published ANT studies on Love waves in China have included the TP region (Huang H et al., 2010), the Sichuan Basin (Li HY et al., 2010; Xu XM et al., 2015; Zheng DC and Wang J, 2017), North China (Fang LH et al., 2013; Fu YV et al., 2016), and Northeastern China (Guo Z et al., 2016). The Love wave tomography results of Fang LH et al. (2013) provided the boundary of the destruction of the North China lithosphere. Zheng DC and Wang J (2017) showed the lateral heterogeneity of sediments in the Sichuan Basin. Li HY et al. (2010) and Xu XM et al. (2015) studied the distribution of low-velocity layers in the North-South seismic belt, and He WG et al. (2015) studied those in the Qinling area. In general, Love wave tomographic studies based on earthquake data are much scarcer than Rayleigh wave studies. However, Peng YJ et al. (2002) performed Love wave tomography on the Chinese continent by using earthquake data from 33 broadband digital stations available in China at the time.

In this study, we performed ANT of Love waves on the Chinese continent, which had not been carried out previously. The study area was roughly 72°E–140°E, 0°–52°N. We collected seismic waveform data recorded by 206 permanent stations in China and 36 nearby global stations. We obtained Love wave group velocity maps of 8- to 40-s periods and the 3-D SH velocity structure of the top 50 km below the surface of the Chinese Mainland. The 3-D SH wave velocity structure was generally consistent with large-scale tectonic structures of the Chinese Mainland and the 3-D SV structure. The discrepancy between our SH velocity structure and a published SV wave velocity structure revealed several consistent features of the crustal radial anisotropy of the Chinese continent.

2. Data and Method
The China Seismic Network includes more than 1,000 permanent seismic stations. Our focus was on large-scale structures. Thus, to reduce the computation time and data storage, we selected a subset of stations (a total of 206) without losing the large-scale spatial coverage by downsampling the dense station distribution in the eastern part of China (Figure 1). We used the horizontal components of 3 years of continuous waveform data (from January 2013 through December 2015). The flat response bandwidth of these instruments was up to 120 s. In addition, we obtained the waveform data from 36 permanent international stations in the surrounding regions that are part of the International Research Institution of Seismology (IRIS) consortium. We extracted the Love wave EGFs from cross-correlations of the horizontal components. The data processing method was similar to that used by Lin FC et al. (2008), which is described here only briefly. For the China Seismic Network stations, the original broadband waveform data were downsampled to 1 Hz, and the continuous data were cut into daylong files. For the global stations,
the LHE and LHN component data (which are at 1 Hz) were downloaded from the IRIS at one day length per file.

To obtain the Love wave EGFs, we needed to perform cross-correlations between the tangential (T) components of the two stations, which required rotating the horizontal components (east, E; and north, N) to the T components according to the azimuth and back azimuth of the station pair. Lin FC et al. (2008) proposed first calculating the cross-correlations of the E-E, E-N, N-E, and N-N components separately. The T-T cross-correlation was then obtained from the combination of these cross-correlations according to the azimuth and back azimuth of the station pair. This procedure reduced considerably the computing time and disk storage by avoiding the computation and storage of long, continuous T components from the horizontal components for each station pair.

The data preprocessing method before cross-correlations was similar to that described by Bensen et al. (2007) but required simultaneous preprocessing of the E and N components. For each of the E and N component traces, the mean, trend, and instrument response were removed and the trace was filtered by a bandpass between 5 and 150 s. A running absolute mean normalization was applied simultaneously to the E and N components to suppress the influence of strong short-term signals such as earthquakes. The weight of the running absolute normalization was defined by the inverse of the larger value of the absolute averages of the E and N components in the same time window (80 s). A frequency domain whitening was then applied simultaneously to the E and N components to improve the bandwidth of the final EGFs. The spectral whitening weight for each frequency was the average of the spectral amplitudes of the E component over a small frequency window (the window length was 40 frequency samples, and the spectra of the E and N components were similar). For each possible station pair, we calculated the cross-correlations of the E-E, E-N, N-N, and N-E components, respectively. We then rotated them to the T component according to the azimuth and back azimuth of the great circle between the station pair (Lin FC et al., 2008, equation 1) to obtain the daily EGF of the Love wave. Finally, we added all the daily EGFs to obtain the final stacked EGF for the station pair. The nonuniform distribution of the noise sources makes the EGF non-symmetric between the positive and negative lags. The “symmetric” component was obtained by superimposing the positive and negative lags, which reduced the effects of nonrandom distribution of the noise sources and improved the signal-to-noise ratio (SNR). The SNR in this study was defined as the peak amplitude in a signal window divided by the root-mean-square of the trailing segment (same definition as in Bensen et al., 2007).

Figure 2 shows a record section of the Love wave EGFs from the T-T cross-correlations between station CD2 (Chengdu, China; Figure 1) and other stations. The Love wave is clearly observable at a speed of approximately 3.3 km/s to a distance of approximately 1,000 km. The traces beyond 1,000 km deteriorate in SNR. Figure 3 shows a comparison of the cross-correlations of the T, R, and Z components for the same station pair. The cross-correlations of the Z and R components are dominated by Rayleigh wave signals, and the T-component cross-correlation is dominated by the Love wave. We can observe a clear dispersion of the Love and Rayleigh waves. Overall, the Love wave is faster than the Rayleigh wave for a similar period. For the Rayleigh wave, the Z-Z cross-correlation has a higher SNR ratio than does the R-R cross-correlation. Because this is generally true, the Z-Z cross-correlation is commonly used to retrieve the Rayleigh wave. The Love wave from the T-T cross-correlation has a lower SNR than does the Rayleigh wave (from Z-Z), which is also generally the case (see the Discussion section below).

After obtaining the Love wave EGFs of the station pairs, we used a multiple-band filtering technique to measure the dispersions of the Love wave group velocities in a manner similar to the Rayleigh wave measurements (Bensen et al., 2007). After obtaining the group velocity dispersion curves, we applied data quality control, which included the following criteria: (1) The Love wave SNR was greater than 10.0. The SNR was defined as the maximum absolute value within a signal window of 2–5 km/s divided by the root-mean-square value of the trailing noise. (2) The interstation distance exceeded 2 times the wavelength. (3) In tomography, travel time measurements with deviations greater than twice the standard deviation of the residuals were discarded.

Figure 4 also shows the distribution of the number of dispersions used for tomography at the 8- to 40-s period band and the path density for a period of 20 s after data quality control. The dispersion paths in the 10- to 30-s periods that were usable for tomography after quality control accounted for 23–35% of the total. The usable data were much less than those for Rayleigh wave ANT. For example, Zheng SH et al. (2008) reported 50–80% usable ray paths after quality control for Rayleigh wave group velocity tomography in the 10- to 30-s periods from cross-correlations of 18 months of continuous waveforms. They showed that visible Rayleigh wave signals might appear on a station pair at distances of up to 5,000 km, whereas the SNR of the Love wave signal deteriorated beyond 1,000 km (Figure 2).

We used the fast marching surface wave tomography (FMST) method (Rawlinson and Sambridge, 2004a, 2004b, 2005) to invert for the Love wave group velocity map for each period. The FMST uses a fast marching method to calculate the theoretical travel time. This method considers the effect of off-great-circle propagation and is suitable for a research area with strong lateral heterogeneity, such as the margins of the TP. The FMST program uses a subspace inversion method to adjust the velocity structure through an iterative process so that the residuals between the observed travel time and the theoretical travel time are gradually reduced to a stable state. The inversion process minimizes the objective function $S(\mathbf{m}) = (g(m) - d_{\text{obs}})C^-1_d(g(m) - d_{\text{obs}})^T + \epsilon(m - m_0)C^-1_m(m - m_0)^T$. Here, $d_{\text{obs}}$ indicates the observed travel time, $g(m)$ indicates the predicted travel time, $C^-1_d$ is the a priori data covariance matrix, $C^-1_m$ is the prior model covariance matrix, $m$ represents the modified model, and $m_0$ represents the initial model (updated for each iteration). The first item in the objective function involves fitting the observation travel time and the theoretical travel time. The latter is a damping term that prevents the model from large changes, where $\epsilon$ is the damping factor. The initial velocity model had little effect on the inversion results according to our tests; hence, the average of the observed group velo-
city was used as the initial velocity model. Finally, the inverted velocity models were spatially smoothed.

Before the velocity structure inversion, we first tested the spatial resolution of the tomography by using the same ray path coverage (Figure 5). We divided the study area into two-dimensional uniform velocity grids of 0.1° spacing and conducted checkerboard tests based on the ray paths actually used in each period. Two different checkerboards are given here, with grid sizes of 4° × 4° and 8° × 8° in half-wavelength, respectively. The perturbed value of the checkerboard was ±10%. The predicted travel times of the ray paths were calculated from the assumed velocity model. These travel times were used as the input data to invert for the velocity model (Figure 5). The resolution of the eastern part of the Chinese Mainland reached 4° × 4°. The resolving power of the western region was relatively weaker, with distinguishable large-scale velocity anomalies of 8° × 8°. The spatial resolution obtained by the checkerboard tests was consistent with the ray path density distribution, in which the ray density in the east (95° boundary) was substantially greater than in the west (Figure 4b).

With the Love wave group velocity maps in different periods, we used a linear method (Herrmann, 2013) to invert for the shear wave velocity structure to fit the dispersion curve of each grid point. We use multiple iterations (25 in total) to fit the observed dispersion curve. A good initial velocity model is helpful in accelerating the convergence of the linear inversion and to constrain the structure at greater depths where the Love wave dispersion loses sensitivity. Here, we used the SV velocity model of Bao XW et al. (2015) as the initial reference model. The top 10 km consisted of 4 thin layers with a thickness of 2.5 km, and the 10- to 100-km depth consisted of 18 thin layers of 5 km thick. After obtaining the layered shear wave velocity structures of all the grid points, we combined them into a 3-D SH wave velocity model.

The group velocities of the Love waves at different periods are sensitive to the subsurface SH velocities at different depths.
Figure 6 shows the Love wave group velocity sensitivity kernels calculated for the AK135 model (Laske et al., 2013). In general, the short-period Love wave is sensitive to shallow depths, and the long-period Love wave is sensitive to both shallow and deep structures. The Love wave velocity of the 10-s period was sensitive to depth structures greater than 20 km, and the Love wave of the 40-s period could sense deeper velocity structures. The velocity structure discussed in this paper had a resolution to a depth of approximately 50 km.

3. Results

3.1 Love Wave Group Velocity Dispersion Maps

Figure 7 shows the Love wave group velocity maps for the 10-, 20-, 30-, and 40-s periods, respectively. We lacked the ray path coverage to obtain dispersion maps at periods longer than 40 s. The lateral variation of the group velocity was strong at all periods. The main features of the high-speed and low-speed distribution at each period were generally consistent with those of the Rayleigh wave noise tomography (Zheng SH et al., 2008; Bao XW et al., 2015; Shen WS et al., 2016). The low-speed anomalies of the 10-s period were found in major basins (Tarim, Sichuan, Qaidam, Ordos, Bohaiwan, and Songliao), which correlated well with the thick sediment distributions of the Crust1.0 model (Laske et al., 2013; Figure 8, left panel). At the longer periods of 30 and 40 s, a clear transition zone could be seen between the high- and low-speed anomalies, roughly along the well-known north-south gravity lineament (NSGL). This result is consistent with the increase in crustal thickness from east to west across the NSGL (Figure 8, right panel). The TP generally had lower speeds than did the stable Yangtze craton in the southeastern Chinese continent at all periods.

3.2 3-D SH Wave Velocity Model

The 3-D SH wave velocity model obtained from inversion of the Love wave dispersion maps is shown in Figure 9 at four depth slices (5 to 45 km). Figure 10 shows the dispersions and velocity structures at four representative tectonic regions, including the TP, Sichuan Basin, South China, and North China (locations marked in Figure 9a). The velocity structure was almost the same as the initial model below a depth of 50 km because the Love group velocity dispersion of 8- to 40-s periods lost sensitivity at greater depths (Figure 6).

We estimated the errors of the SH model by using the following
bootstrap method (Efron and Tibshirani, 1994; Tichelaar and Ruff, 1989); (1) We randomly selected dispersion measurements, allowing repeat sampling. We constructed a new set of dispersion measurements with the same number of station pairs as in the original data set (but containing repeat pairs). (2) We followed the same procedure as described above to construct the dispersion maps and to invert for the 3-D SH model. (3) We repeated these steps 100 times. (4) The errors of the SH model were calculated from the resulting 100 models and ranged up to approximately 0.09 km/s, with larger errors in the west and smaller errors in the east (Figure 11). The error maps were consistent with the ray path density (Figure 4) and resolution maps (Figure 5).

Figure 5. Checkerboard resolution tests for different periods. The input models are alternating patterns of ±10% peak perturbation with half-wavelengths of 4° × 4° (top left) and 8° × 8° (top right), respectively. The inversion results for different periods (labeled) were obtained by using the same ray coverages as the real data in this study.
The maps of SH velocity at depth slices (Figure 9) showed considerable variation, which correlated with the geological units. At 5 km (Figure 9a), the major basins (Tarim, Sichuan, Qaidam, Ordos, Bohaiwan, and Songliao) showed low speeds, reflecting their thick sedimentary deposits. The Yangtze Block and eastern part of the TP showed a relatively high velocity. At 15 km (Figure 9b), the lateral variation was weaker. The Tarim Basin showed distinct differences between the northern and southern parts. At 30 km (Figure 9c), the map showed a clear contrast from the east to the west, with fast velocities east of the NSGL for areas in the mantle and slow velocities for areas in the mid-crust of the TP. At 45 km (Figure 9d), the contrast between east and west was clearer because the eastern part is in the mantle, whereas the TP and most of western China are still in the crust.

The depth profile A-A' (Figure 12) shows the velocity structure beneath the TP, the Sichuan Basin, and the Yangtze Block. The SH and SV wave velocity structures were generally consistent, showing relatively large differences only below the Sichuan Basin. This section showed that the low-velocity depth of the deep underground gradually thinned from west to east and that the low-velocity thickness had a good positive correlation with the crustal thickness. Section B-B' passing through the Yangtze Block and the Bohaiwan Basin shows that the SH wave is mostly consistent with the SV wave but that a large difference occurs below the North China region, which indicates strong radial anisotropy (see below).

4. Discussion

4.1 Convergence of Long-Period Love Wave EGFs

The dispersion maps we obtained from the ANT of Love waves were at relatively short periods (8 to 40 s) compared with those of Rayleigh waves, which extend up to 70 s (Zheng SH et al., 2008). This was because the SNRs of Love waves for most pairs were not high at longer periods. We examined whether the SNRs of the Love wave EGFs could be improved by increasing the duration of the continuous data used in the cross-correlations (Figure 13). We selected two stations (Enshi, ENH; and Xi'an, XAN) with a long re-
According to history and compared the SNRs of Love and Rayleigh waves at periods of 42 and 60 s, respectively. We obtained cross-correlations for one year of data separately and stacked the annual cross-correlations (Figure 13). It was apparent that the curves of the SNRs as a function of increasing stacking duration were slightly different depending on the order of stacking. We performed two stacks with different orders and plotted the results after 1, 2, 4, 8, and 16 years of stacking (Figure 13). One stacking was in chronological order from the early years to the later years (Figure 13a). Another method was to sort the annual stacks of the 42-s Love wave in order of increasing SNRs from lower to higher values and then obtain the total stacks (for both periods and for both Love and Rayleigh waves) according to that order (Figure 13b).

The results for the SNRs showed the following: (1) The SNR of the Rayleigh wave was almost always greater than that of the Love wave for either period. (2) The SNR of the Rayleigh wave at 60 s was lower than that at 42 s, but the difference did not change greatly when the duration was greater than 4 years. (3) The SNR of the Love wave at 60 s was lower than that at 42 s and considerably lower when the duration was greater than 4 to 8 years. (4) The SNR increased rapidly in general (for either period and for either Rayleigh or Love waves) as the duration increased from 1 to 8 years. (5) After 8 years, the SNR of the Rayleigh wave at 60 s continued to increase as the duration increased, whereas the SNR of the Rayleigh wave at 42 s reached a plateau. The SNR of the Love waves increased slowly, reached a plateau, and sometimes decreased.

Thus, increasing the amount of continuous data may not improve the retrieval of the relatively long-period Love waves from the am-

**Figure 8.** Maps of (left) sedimentary layer thickness from Crust1.0 (Laske et al., 2013) and (right) crustal thickness from Bao XW et al. (2015).

**Figure 9.** Inversion results of 3-D SH wave velocity at depths of (a–d) 5, 15, 30, and 45 km, respectively.
bient noise correlation. The tests above suggest little benefit from using more than 4 to 8 years of continuous data.

It is particularly worth noting that the SNR for the Love wave could sometimes decrease as the duration increased to more than 8 years (Figure 13a). This surprising result was due to two primary factors. First, when a new annual stack with poor SNRs was added to the total stack, it might not have improved the SNR of the total stack. As shown in Figure 13b, the SNR of the total stack continued to improve, although very slightly, when the better annual stacks were added. Second and more importantly, although the amplitude of the Love wave continued to increase as the duration increased, the amplitude of the trailing segment (after the Love wave) also increased. As a result, the SNR of the total stack could decrease. Essentially, the trailing segment was also part of the Green’s function, containing coherent information from the wave propagation. In fact, the coda of EGFs have been studied, such as in the construction of Green’s functions (so-called C; Stehly et al., 2005; Zhang J and Yang XN, 2013) or in coda interferometry (Benguier et al., 2008). Thus, the traditional definition of the SNR by using the trailing segment (Bensen et al., 2007) may not be the best way to judge the quality of the EGF. An alternative would be to use the preceding segment (before the surface wave); however, this segment is often contaminated by noncausal signals. This issue is worth exploring in the future.

4.2 S-Wave Radial Anisotropy

We obtained the 3-D SH wave velocity structure of the crust and uppermost mantle in China by using the ANT method for the first time. The main features of the 3-D SH velocity model showed correlations with geological units (discussed above) and similarities to those of the SV velocity structure (Bao XW et al., 2015; Figure 4). This consistency validates the use of ANT for Love waves. In contrast, the difference between the SH velocity structure and the previously published SV velocity structure may reflect the radial anisotropy of the subsurface structure. The subsurface radial anisotropy causes a difference in shear wave velocities between the horizontally polarized (SH) and vertically polarized shear wave (SV), which has long been recognized, such as in the Preliminary Reference Earth Model (Dziewonski and Anderson, 1981).

Here, we compare our 3-D SH velocity model obtained from the ANT of Love waves with the 3-D SV model of Bao XW et al. (2015), which was obtained from Raleigh waves by using both earthquake and ambient noise data. Before we discuss the results (Figure 14), we need to bear in mind that the two models have different data coverages and model resolutions; thus, we caution that some of the discrepancy may not be real. We focus only on large-scale structures of larger anomalies, which we believe are more robust. We define the radial anisotropy parameter as

$$\psi = 2\left(\frac{V_{SH} - V_{SV}}{V_{SH} + V_{SV}}\right).$$
No previous publication on the crustal radial anisotropy of the Chinese continent has used ambient noise. However, some radial anisotropy results for relatively small areas have been published, such as for Northeast China (Guo Z et al., 2016), North China (Cheng C et al., 2013; Fu YV et al., 2016), the TP (Shapiro et al., 2004), the eastern margin of the TP (Xie JY et al., 2013), and the northeastern margin of the TP (Li L et al., 2016). Our results showed strong but variable radial anisotropy at different depths (Figure 14). At 5 km, the anisotropy was not strong. Major basins (Ordos, Sichuan, Bohaiwan, Songliao) showed significant (2–8%) and positive (SH > SV) anisotropy. The anisotropy in the Tarim Basin is more complex, with both positive and negative values between the southern and northern parts. At 45 km, the overall strength of the radial anisotropy was relatively weak.

At an intermediate depth (15–30 km; Figure 14b, 14c), most of the study region showed strong positive radial anisotropy. The central and southern TP showed positive anisotropy. These results agreed with those from an earlier study by Shapiro et al. (2004), who explained the anisotropy by the preferred orientation of mica crystals resulting from crustal thinning of the mechanically weak mid-crust. At the northern and eastern margins, both studies showed relatively weak radial anisotropy. This change in the radial anisotropy pattern suggests a change in the style of deformation from the western and southern TP to the northern and eastern margins as the plateau grew. This result, however, differs from that of Xie JY et al. (2013), who reported strong positive radial anisotropy in the northeastern margin of the TP. In the future, this discrepancy needs to be examined further with a denser array of stations. Strong anisotropy was found under the Tarim and Junggar Basins and Tianshan Mountains, which may be related to thick sedimentary layers or Tianshan deformation.

The most prominent radial anisotropy was found in North China at the mid-lower crust. The radial anisotropy below the Songliao Basin was similar to that reported by Guo Z et al. (2016), who explained it as a crustal extension. Positive radial anisotropy has been found below the Ordos and Bohaiwan Basins, but the strength in the Bohaiwan Basin is much stronger. Cheng C et al. (2013) also reported significant differences in radial anisotropy below the Ordos and Bohaiwan Basins. They interpreted this difference as a distinct tectonic process between the Ordos and Bohaiwan Basins. They proposed that the western part of the North China Craton has not undergone decratonization, whereas the eastern part has experienced significant craton destruction (Menzies and Xu YG, 1998; Chen L et al., 2009; Zhu RX et al., 2011). Fu YV et al. (2016) also reported significant positive radial anisotropy in the lower crust of the Bohaiwan Basin, which they believe is a lateral flow of crustal material caused by the extension of the Mesozoic crust.

5. Conclusions

In this study, we performed ANT for Love waves on the Chinese continent by using EGFs from cross-correlations of the horizontal components of 3 years of continuous data. To save computation time and focus on the large-scale structure in this initial effort, we selected 242 stations on the Chinese Mainland and surrounding regions, which were distributed relatively uniformly. We obtained Love wave group velocity dispersion maps of 8- to 40-s periods and a 3-D SH velocity model from the surface to a depth of 50 km. The resolution was approximately 4° × 4° and 8° × 8° for eastern and western China, respectively. The 3-D SH model showed large-scale features that were consistent with geological units, such as major basins and changes in crustal thickness across the Chinese continent.
The SH model and the previously published SV model based on Rayleigh wave tomography (Bao XW et al., 2015) were generally consistent. However, the SH and SV models also showed substantial differences, which we used to examine the subsurface radial anisotropy. At 5 km, the anisotropy was not strong. Major basins (Tarim, Sichuan, Qaidam, Bohaiwan, and Songliao) showed significant (2–8%) and positive (SH > SV) anisotropy. The anisotropy in the Tarim Basin was more complex, with both positive and negative values between the southern and northern parts. At 45 km, the overall strength of the radial anisotropy was relatively weak. At an intermediate depth (15–30 km; Figure 14b, 14c), most of the study region showed strong positive radial anisotropy. The central and southern TP showed positive anisotropy, consistent with thinning of the mechanically weak mid-crust. At the northern and eastern margins, the radial anisotropy was relatively weak, which may suggest a change in the style of deformation from the western and southern TP to the northern and eastern margins as the plateau grew. Strong anisotropy was found under the Tarim and Junggar Basins and Tianshan Mountains, which may be related to thick sedimentary layers or Tianshan deformation. Prominent positive radial anisotropy was found beneath the North China craton at the mid-lower crust. The strength under the Bohaiwan Basin was much stronger than that under the Ordos Basin. These observations could be explained by decratonization and extension of the North China craton since the Mesozoic Era.

The SNRs of the EGFs of Love waves were generally much lower than those of Rayleigh waves. The maximum period at which reliable Love wave dispersion measurements could be obtained for the construction of dispersion maps was 40 s, considerably shorter than the Rayleigh wave limit (approximately 70 s; e.g., Bao XW et al., 2015). Nonlinear stacking methods, such as the phase-
weighted stacking or time-frequency stacking algorithm (Schimmel and Paulssen, 1997; Baig et al., 2009), may help improve the Love waves for longer periods. Merely increasing the amount of continuous data over 4 to 8 years may not improve retrieval of the relatively long-period Love waves from the ambient noise correlation. Ingenious methods need to be explored for retrieving long-period Love waves, which are important for imaging deeper sub-surface structures.

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**Figure 13.** Comparison of signal-to-noise ratios (SNRs) of EGFs as a function of the data duration between Love and Rayleigh waves at two periods (42 and 60 s). The cross-correlation stations are IC.ENH and IC.XAN, which have a long recording history. The annual stacks of cross-correlations were calculated first. The total stacks had durations of 1, 2, 4, 8, and 16 years, respectively. (a) The total stacks were calculated by using the annual stacks in chronological order (from early years to later years). (b) The annual stacks were sorted first by using the SNRs, from lower to higher SNR values. The total stacks were then calculated by adding the annual stacks with increasing SNR values.

**Figure 14.** Maps of radial anisotropy at depths of 5, 15, 30, and 45 km. The anisotropy is defined as $\psi = 2(V_{SH} - V_{SV})/(V_{SH} + V_{SV})$. The $V_{SH}$ and $V_{SV}$ values are from this study and Bao XW et al. (2015), respectively.

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group velocity was inverted by using FMST software (http://res.anu.edu.au/~nick/surfomo.html). The linear inversion technique was conducted by using the CPS330 software package (http://www.eas.slu.edu/eqc/eqccps.html). Most of the figures in this paper were plotted with GMT software (http://www.soest.hawaii.edu/gmt/). This research was supported by the Natural Science Foundation of China (grants 41774069 and 41774056).

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