Crustal shear-wave velocity structure of northeastern noise Tibet revealed by ambient seismic noise and receiver functions

Zhenbo Wu\textsuperscript{a, b}, Tao Xu\textsuperscript{a, c, e}, José Badal\textsuperscript{d}, Huajian Yao\textsuperscript{e, f}, Chenglong Wu\textsuperscript{a, b}, Jiwen Teng\textsuperscript{a}

\textsuperscript{a} State Key Laboratory of the Lithosphere Evolution, Institute of Geology and Geophysics, Chinese Academy of Sciences, Bejing 100029, China
\textsuperscript{b} University of Chinese Academy of Sciences, Beijing 100049, China
\textsuperscript{c} National Geophysical Observatory at Mengcheng, Anhui, China
\textsuperscript{d} Laboratory of Seismology and Physics of Earth’s Interior, School of Earth and Space Sciences, University of Science and Technology of China, Hefei 230026, China
\textsuperscript{e} CAS Center for Excellence in Tibetan Plateau Earth Sciences, Beijing 100029, China
\textsuperscript{f} Laboratory of Seismology and Physics of Earth’s Interior, University of Zaragoza, Pedro Cerbuna 12, 50009 Zaragoza, Spain

\begin{abstract}
The Tibetan plateau is formed by the persistent convergence between the Indian and Eurasian plates. The northeastern Tibetan plateau is undergoing young deformation that has been noticed for a long time. We conduct a passive-source seismic profile with 22 stations in NE Tibet in order to investigate the crustal shear-wave velocity structure and its relationship with tectonic processes. In this paper we obtain the Rayleigh-wave phase velocity dispersion data among all station pairs within the period bandwidth of 5–20 s from the method of ambient noise cross-correlations. Phase velocity variations correlate well with surface geological boundaries and tectonic features, for instance, low phase velocity beneath the Songpan–Ganzi block and the Guide basin. We also compute P-wave receiver functions based on the selected teleseismic events with similar ray parameters, and perform the joint inversion of surface wave dispersion data and receiver functions to obtain the 2-D crustal shear-wave velocity structure along the profile. The inversion results show that low shear-wave velocities beneath the Songpan–Ganzi block are widespread in the middle-to-lower crust. In together with high crustal \( V_p/V_s \) ratios and high temperature suggested by the P-wave velocities obtained from the active-source seismic study, we suggest that the low velocity zone beneath the Songpan–Ganzi block is probably attributed to partial melting. Across the North Kunlun fault, there is no crustal LVZ found beneath the Kunlun block. This structural difference may have already existed before the collision of the two blocks, or due to limit of the northward extension for the crustal LVZ across the North Kunlun fault.
\end{abstract}

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1. Introduction

Uplift of the Tibetan plateau is the most spectacular tectonic event during the Cenozoic created by the collision of India and Eurasia. Fundamental questions persist concerning the initiation of the convergence (e.g., Molnar and Tapponnier, 1975; Rowley, 1996; Aitchison et al., 2011; Sun et al., 2012; Zhang et al., 2012; Hu et al., 2015) and the mechanism of lithospheric deformation (e.g., Tapponnier and Molnar 1976; England and Houseman, 1989; Royden et al., 1997; Replumaz et al., 2014; Chen et al., 2015). Many models have been proposed to explain its dynamic responses to collision and its consequent deformation patterns. To this respect, three models have received wide attention, that is, the rigid block extrusion (Tapponnier and Molnar 1976; Tapponnier et al., 1982), the thin-viscous-sheet model (England and Houseman, 1986, 1989), and the crustal channel flow model (Royden et al., 1997).

The northeastern (NE) Tibetan plateau, viewed as a young outgrowth of its evolution and deformation (Meyer et al., 1998), has been a factor of many studies (e.g., Galvè et al., 2002; Vergne et al., 2002; Clark et al., 2010; Duvall and Clark, 2010; Karplus et al., 2013; Tian and Zhang, 2013; Xia et al., 2011; Deng et al., 2015). Deformation mechanisms that work here at present may resemble what happened in the central plateau and participated in the formation of the Tibetan plateau. This is one of the motivations for this study, which will help us understand the earlier deformation of central Tibet. On the other hand, some studies in the southeastern (SE) Tibetan plateau have found that low shear-wave velocities exist in the middle-to-lower crust (Yao et al., 2008, 2010; Liu et al., 2014). We are concerned about whether this phenomenon also exists in the NE Tibetan plateau. It will be very helpful to understand how the eastward expansion of the Tibetan plateau material is bifurcated by the rigid Sichuan Basin on the basis of comparing deformation patterns in the NE and SE Tibetan plateaus. Many works have been done from different perspectives...
to discuss this issue. There may exist a mechanically weak lower crust that accounts for crustal channel flow in the surrounding regions of the Sichuan Basin (Clark and Royden, 2000). Resistivity models obtained from magnetotelluric data show evidences for penetration of partial melting crust across the Kunlun Fault into northern Tibet (Pape et al., 2012).

Galvé et al. (2002) collect the wide-angle reflection–refraction data by active-source seismic survey. The obtained P- and S-wave velocities support predominant felsic composition for the crust and suggest that only the upper crust has been thickened to the north of the Kunlun fault; to the south of this fault the thicker crust is composed by two layers, which could be the superposition of the originally thin crust of the Bayar Har Terrane on the lower crust of the domain to the north. However, Liu et al. (2006) suggest that crustal thickening mainly happens in the lower crust in the NE Tibetan plateau, based on the Darlag–Lanzhou–Jingbian seismic refraction profile.

Seismic interferometry technique using ambient seismic noise has rapidly become an important method to investigate the Earth structure at different scales. In fact, surface wave tomography based on the ambient noise method has provided essential constraints on crustal structure at different scales. In fact, surface wave tomography based on the ambient noise method has provided essential constraints on crustal structure at different scales. In fact, surface wave tomography based on the ambient noise method has provided essential constraints on crustal structure at different scales. In fact, surface wave tomography based on the ambient noise method has provided essential constraints on crustal structure at different scales. In fact, surface wave tomography based on the ambient noise method has provided essential constraints on crustal structure at different scales. In fact, surface wave tomography based on the ambient noise method has provided essential constraints on crustal structure at different scales.
In order to avoid any ambiguity arising from propagation effects caused by different ray paths and unequal anisotropic properties from different azimuths, we select teleseismic events located in the West Pacific subduction zone near Japan to compute the receiver functions, these events have similar back azimuths and ray parameters (Fig. 2).

4. Method

Deep structure along this seismic survey line has been studied in two ways: the wide-angle reflection–refraction (Zhang et al., 2011) and the receiver function imaging (Xu et al., 2014). However, the steeply incident body waves used in the wide-angle reflection–refraction are less sensitive to the shallow heterogeneity than dispersion curves of short-period surface waves, and usually it is hard to obtain the high resolution shear-wave velocity structure mainly due to the low signal-to-noise ratio (SNR) of shear-wave data. Receiver function is good at detecting interfaces with a large velocity contrast, but it is less sensitive to the absolute value of wave speed. We combine dispersion curves of short-period surface waves from ambient noise cross-correlation and teleseismic P-wave receiver functions together to obtain the high resolution shear-wave velocity structure in the crust.

4.1. Surface wave empirical Green’s functions

Based on an assumption of randomly distributed ambient noise sources, monthly or longer continuous data can be cross-correlated to
compute the surface wave empirical Green’s functions (EGFs) with sufficient SNR for dispersion analysis and to produce reliable tomographic images (Shapiro et al., 2005; Yao et al., 2006; Bensen et al., 2007).

A series of preprocessing steps should be applied to the waveform data to improve the SNR of the EGFs. Commonly, the ambient seismic data preprocessing consists of band-pass filtering, in our case using the bandwidth 0.2–50 s, cutting the continuous data to one day length, de-trending the zero line slope, removing instrument response, and re-sampling the data to improve the computational efficiency, and in our case from 40 sps to 10 sps, normalization in time domain and spectral whitening in frequency domain (Bensen et al., 2007; Badal et al., 2013). Cross-correlations of ambient seismic noise would be seriously contaminated by the local or teleseismic events without the normalization in the time domain. On the other hand, since the amplitude spectrum of ambient seismic noise is normally not flat in the frequency domain, we use a spectral whitening technique to smooth the amplitude spectrum. This operation can produce comparable amplitudes for the concerned frequency band and mute the influence from the persistent monochromatic seismic sources (Villaseñor et al., 2007; Zheng et al., 2011).

Cross-correlations of vertical-component ambient seismic noise are calculated for each available station pair of the 22 stations using the daily time series, and the cross-correlations consist of positive and negative lags with apparent Rayleigh wave signals. Fig. 3 shows the stacked cross-correlation functions (CFs) in the frequency band 10–20 s extracted from ambient noise cross-correlations with a total of 152-day record between the stations S06 and S17. The difference between Fig. 3(a) and (b) is that the result plotted in (a) is computed without spectral whitening, while the latter (b) is the result with spectral whitening. SNR increases markedly in the latter case.

Cross-correlations between the stations S06 and S17. The difference between Fig. 3(a) and (b) is that the result plotted in (a) is computed without spectral whitening, while the latter (b) is the result with spectral whitening.

4.2. Phase velocity dispersion mapping

We measure the phase velocity dispersion curves from the EGFs based on a far-field approximation and an image transformation technique (Yao et al., 2006). There are two constraints to be satisfied: one is that the distance between station pair has to be at least three wavelengths of the surface wave signal in order to satisfy the far-field approximation; the other is that the SNR has to be greater than 5 to ensure the reliability of phase velocity measurements. Here SNR is defined as the ratio of the maximum amplitude of the signal window and the mean envelope amplitude of the 150 s long noise window right after the signal window around the central period. As an example, the EGF extracted from the data at two stations (Fig. 5a) is shown in Fig. 5b, the SNR value is shown in Fig. 5c. Fig. 5d shows the velocity-period image and the red circles represent the measured phase velocities within the period band 5–20 s.

In order to ensure the reliability of the measurements, we repeat the above process for different monthly data of each station pair. The dispersion curves with significant discrepancies from most of the results are discarded. We measure the Rayleigh–wave phase velocity dispersion curves for all possible station pairs and finally obtain 89 dispersion curves at periods from 5 to 20 s. These curves are plotted together in Fig. 5e. Both the average phase velocity dispersion curve and its standard deviation at different periods are shown in Fig. 5f.

We use the continuous regionalization and the generalized inversion scheme to invert path-averaged phase velocities at each period for 2-D phase velocity distribution (Tarantola and Valette, 1982; Montagner, 1986; Yao et al., 2010). We set a proper 2D inversion region containing our survey line and the grid interval used for inversion is 0.5° × 0.5°. Yao et al. (2010) introduce the details about this scheme that include inversion for both isotropic phase velocities and azimuthal anisotropy. Here, we are only concerned with the isotropic phase velocity variation. For each period, we first obtain period-dependent 2D phase velocity variations. Then we interpolate the phase velocity maps of different periods to the station locations along the survey line. The short-period phase velocities are very sensitive to the shallow geological features.

Fig. 2. Global map showing the selected earthquakes (green circles) for calculating receiver functions, which are located roughly in the same area and have similar epicentral distances (around 35°), back azimuths and ray parameters. The red triangle indicates the approximate location of the seismic array.
Fig. 4. (a) Symmetric EGFs extracted from the daily cross-correlations of the vertical component data recorded at the station S00 with the other stations, stacked within the period band 10–20 s and sorted from top to bottom by the interstation distance. (b) Daily EGFs for a single station pair S02–S11. The stacked trace for all daily EGFs is shown as the heavy solid line on top.

Fig. 5. (a) Passive-source seismic array deployed in northeastern Tibet (triangles) that is used in this study. (b) EGF for the station pair S06–S27 (red triangles). (c) Signal-to-noise ratio (SNR) versus period for the EGF in (b). (d) Velocity-period diagram and the selected phase velocity dispersion curve (red dotted line) within the phase velocity window 2.0–4.5 km/s at periods 5–20 s. (e) All 89 phase velocity dispersion curves measured from EGFs. (f) Averaged phase velocity dispersion curve with vertical bars representing the standard deviations at different periods.
As it can be seen in Fig. 6, the spatial distribution of phase velocity along the survey line is correlated with the surface tectonic features. Phase velocities as low as 2.8 km/s at periods of 5–8 s appear in the vicinity of the North Kunlun fault, a zone crossed by strike-slip and thrust faults. The cause of these low velocities is associated with the Triassic turbidite sediments in the Songpan–Ganzi block. At greater periods, phase velocities close to 3.0 km/s appear under the southernmost transect of the survey line and stretch northward up to the North Kunlun fault. In the southern Kunlun block, there shows a uniform phase velocity of ~3.1 km/s. Across the Gonghe Nan Shan fault, the phase velocity changes from 3.2 km/s between the Gonghe Nan Shan and Riyue Shan faults to 3.3 km/s northward of the Qilian block. However, velocities as low as 2.9–3.0 km/s at periods of 5–8 s appear again in the Guide basin, between the Riyue Shan fault and the South Qilian suture.

4.3. Receiver functions

We compute P-wave receiver functions by iterative deconvolution of the vertical component from the radial component from these selected events in the time domain (Ligorria and Ammon, 1999). As shown in Fig. 1b with green cross symbols, the piercing points at 60 km depth (average Moho depth) of these receiver functions parallel the shear-wave structures of the study area. On the other hand, the piercing points of each station are almost overlapped, which indicates that the receiver functions have similar travel time for the P-to-S converted phases and allow us to stack them by a simple linear scheme. However, if we use all global events to compute the receiver functions, the move-out technique must be applied before stacking. The move-out value could be related to the surface tectonic features. Phase velocities as low as 2.9 km/s appear again in the Guide basin, between the Riyue Shan fault and the South Qilian suture.

5. Joint inversion of surface wave dispersion and receiver function

5.1. Implementation

Receiver functions dominated by P-to-S converted phases can provide good constraints for the shear-wave velocity discontinuities of the Earth, and surface wave dispersion can provide constraints on the average shear-wave velocities in different depth ranges. Joint inversion of these two data sets can provide much better constraints on the crustal shear-wave velocity structure beneath our survey line.

We perform a damped least-squares scheme (Tarantola and Valette, 1982) and follow the joint inversion procedure based on the minimization of an objective function including both surface wave dispersion data and receiver functions (Julia et al., 2000). The computer program package developed by Herrmann and Ammon (2004) is used to obtain the 1-D shear-wave velocity structure for each station. According to the least-squares scheme, we shall provide an initial shear velocity model as accurate as possible to greatly reduce the non-uniqueness of the inversion results. Here we interpolate a 3D Vs model of the crust and uppermost mantle beneath Tibet from ambient noise tomography provided by Yang et al. (2012) as our initial shear wave velocity model. We calculate the P-wave velocity based on the Vp/Vs ratio given by the previous receiver function study (Xu et al., 2014). Then, we determine the density by using the empirical formula \( \rho = 0.77 + 0.32Vp \). The initial model is parameterized by thin layers with 1 km thickness to get an almost continuous velocity model with depth to avoid interfaces from sharply changed velocities (Ammon et al., 1999).

The objective function that we minimize is shown as follows:

\[
S = \frac{1-p}{N_r} \sum_{i=0}^{N_r} \left( \frac{O_{r_i} - P_{r_i}}{\sigma_{r_i}} \right)^2 + \frac{p}{N_s} \sum_{j=0}^{N_s} \left( \frac{O_{s_j} - P_{s_j}}{\sigma_{s_j}} \right)^2
\]

where,

- \( O_{r_i} \): observed receiver function at time \( t_i \)
- \( P_{r_i} \): predicted receiver function at time \( t_i \)

Fig. 6. Rayleigh-wave phase velocities versus periods along the seismic array. On top: surface topography along the survey line with the elevation decreasing very smoothly from south to north, and geographic location of the array stations (red triangles). The faults in Fig. 1b are also indicated here.
The parameter $p$ controls the weight of two data sets in the joint inversion. In practice, considering that the Rayleigh wave phase velocities at periods from 5 to 20 s are mainly sensitive to the upper-middle crustal structure, we first set $p = 0$ thus forcing the solution to be exclusively based on the receiver functions, and use a damping factor 10 to avoid an overshoot of the firstly determined model. Then, we take both data sets by changing the weight $p = 0.2$ to allow both of them to participate in the inversion. The damping factor is also revised to 0.1. In this way, both data sets are fitted simultaneously during the inversion process. Since the periods of phase velocities are too short to constrain the middle-to-lower crustal structure well, we gradually decrease the weight of each layer from 1.0 to 0.1 as the depth increases from 30 km to 50 km, and no weight is applied to layers with depths above 50 km.

5.2. Inversion results

The inversion results for the stations S04 and S16 are shown in Fig. 8. The synthetic receiver functions based on the inversion results (red curves) have really high correlation coefficients of 0.9092 and 0.9522 with the observed receiver functions, respectively. The theoretical dispersion curves (continuous red lines) calculated from the inversion results also fit the observed phase velocities (dotted lines) well. In Fig. 9b, we show the shear-wave velocity image of the crust along the survey line. For comparison, the initial shear-wave velocity model derived from the study of Yang et al. (2012) is also shown in Fig. 9a. In order to check the reliability of the inversion results, we change the layer thickness of the initial model to 2 km and 2.5 km, and the inversion results shown in Fig. 10 are also very similar.

6. Discussion and conclusion

In view of the inversion result, the variation of the shear-wave velocities in the upper crust along the survey line is consistent with the Rayleigh wave phase velocities (Fig. 6). Low phase velocities at 5–17 s periods are obtained at the south of the Kunlun fault, and low shear-wave velocities at depths up to ~20 km are observed correspondingly at the same location. This can be attributed to the Middle to Late Triassic Songpan–Ganzi complex, with an estimated stratigraphic thickness of 10 km and reaching 15 km locally (Weislogel, 2008). At the north part of the survey line, low shear-wave velocities underlie the intramontane Guide basin, probably due to the Cenozoic fill and scattered Neogene/Quaternary deposits, or the tectonic destruction of upper crust from the Riyue Shan fault and South Qilian suture to the surrounding regions. Strike-slip faults mainly develop in the region south of Kunlun fault, such as Mado–Gande fault, and thrust faults develop north of the Kunlun fault, such as Animaqing, Gonghe Nan Shan, Laji Shan, and Qinghai Nan Shan. Considering lateral variations of the upper-crustal shear-wave velocity and the surface tectonic features along the survey line, we suggest that the upper-crustal properties between Songpan–Ganzi block and Kunlun block are probably different.
the upper crust in the Songpan–Ganzi block is more ductile, and in the Kunlun block shows more brittle feature.

There is an obvious phenomenon in the shear-wave velocity structure, that is, a low velocity zone (LVZ) is found in the middle-to-lower crust within the depth range about 40–60 km beneath the Songpan–Ganzi block. The mechanism that causes this LVZ is an important issue. The P-wave velocity model from the active-source seismic study (Zhang et al., 2011) shows that the P-wave velocity gradient agrees well with the average continental crust in the Songpan–Ganzi block, but the absolute value is lower by 0.5 km/s than the continental average.

There are two possible explanations for this observation provided by Zhang et al. (2011). One is that the crust of the Songpan–Ganzi block is more felsic than normal crustal composition; if true, the average crustal Vp/Vs ratio should become lower than normal values. The other reason is related to the crustal temperature beneath the Songpan–Ganzi block, which needs to be 500 °C above the typical continental geothermal curves (Christensen, 1979). If so, the crustal Vp/Vs ratio should be much higher than 1.73 since the Vs values would be significantly smaller. The average crustal Vp/Vs in the Songpan–Ganzi block is ~1.81 provided by the previous receiver function study (Xu et al.,

Fig. 8. 1-D shear-wave velocity inversion results for the stations S04 and S16. For both of them, we show the initial shear-wave velocity model (blue dashed line) and the inversion result (red line) on the left velocity–depth panel. On the right the upper panel shows the observed receiver function (blue curve) and the synthetic result from the inversion model (red curve) with the correlation coefficients 0.9092 and 0.9522, respectively. The lower panel shows the measured Rayleigh-wave phase velocities (black circle) and the synthetic result (red line).

Fig. 9. On top: surface topography along the survey line and geographic locations of the broad-band stations (red triangles). The faults in Fig. 1b are added. (a) The initial shear-wave velocity model with the layer thickness 1.0 km provided by Yang et al. (2012). (b) Shear-wave velocity image of the crust along the study line based on the joint inversion of the Rayleigh-wave phase velocity dispersion and receiver functions.
Several previous studies performed in the northeastern Tibetan plateau have found the existence of the LVZ, but the distribution areas are still obscure. Unsworth et al. (2004) report the existence of widespread fluids at depths 25–50 km in the region 32.5°N–35.5°N and 92°E–94°E. Later, Karplus et al. (2013) extend this region to 32°N–36°N and 92°E–98°E by means of low group velocities found at periods 34–40 s, which are sensitive to the middle-to-lower crust velocity at depths of ~30–60 km. Yang et al. (2012) observe low velocities in a range of 2.9 to 3.3 km/s at depths 20–40 km in the Songpan–Ganzi block and suggest that the reason is partial melting. Two recent studies based on ambient noise surface wave tomography show a prominent LVZ in the middle crust of the Songpan–Ganzi block at depths 20–40 km (H.Y. Li et al., 2014; X.F. Li et al., 2014). These studies show that the LVZ in the NE Tibetan plateau only exists in the region south of the North Kunlun fault, but Jiang et al. (2014) suggest that the LVZ in the NE Tibetan plateau is confined to the region south of the Kunlun fault and west of 98°E, and it penetrates northward into the Qinling orogens to the region north of the Kunlun fault and east of 98°E. In our shear-wave velocity structure, we do not observe LVZ under the Kunlun block north of Kunlun fault. This difference may reflect intrinsic variations of two blocks before their collision, or it is because the Kunlun fault limits the northward extension of the LVZ. According to all the observations, the north limit of the LVZs in the NE Tibetan plateau probably distributes non-uniformly in different regions and more studies need to investigate its distribution in detail.

The LVZ in the northeastern Tibetan plateau is comparable to that in other parts of the plateau, such as the eastern, central, and southern Tibet. Geophysical studies using different methods also find high electrical conductivity, high temperature and low strength zones with low seismic velocities in the middle-to-lower crust beneath the plateau (e.g. Clark and Royden, 2000; Yao et al., 2008; Guo et al., 2009; Li et al., 2009; Zheng et al., 2010; Yang et al., 2012; Zheng et al., 2013; Liu et al., 2014). The effects of this LVZ on the dynamic mechanism and deformation of the NE Tibetan plateau are still controversial topics. Clark and Royden (2000) propose the channel flow model to explain the surface topography of the surrounding regions of the Tibetan plateau. To the east, they suggest that the adjacent crust of the Sichuan Basin is mechanically weak, and the middle-to-lower crust of the plateau can escape to these regions, thus producing the broader plateau margin with a smoother topographic gradient. Yang et al. (2012) point that the LVZ observed in the middle-to-lower crust only based on Rayleigh wave observations is not strong enough to support the argument of the channel flow model. Liu et al. (2014) conclude that the eastward extension of the Tibetan plateau is accommodated by a combination of local crustal flow and strain partitioning across deep faults such as the Longmen Shan fault, and propose that rigid block motion and crustal flow may be reconcilable as mutual modes in the framework of crustal deformation in southeastern Tibet. In order to determine whether the channel flow model could dominantly explain the tectonic process in the NE Tibetan plateau, we should perform more work in the future such as accurate Love wave phase velocity measurements for radial anisotropy, inversion of azimuthal anisotropy, and investigation of lithospheric shear-wave velocities in this region.

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