Lithospheric structure beneath the northeastern Tibetan Plateau and the western Sino-Korea Craton revealed by Rayleigh wave tomography

Yonghua Li,1,2 Jiatie Pan,1,2 Qingju Wu1,2 and Zhifeng Ding1,2

1Key Laboratory of Seismic Observation and Geophysical Imaging, China Earthquake Administration, Beijing 100081, China. E-mail: lyyhgeomag@sina.com
2Institute of Geophysics, China Earthquake Administration, Beijing 100081, China

SUMMARY

A new 3-D shear wave velocity model of the northeastern (NE) Tibetan Plateau and western Sino-Korea Craton is presented. The model is based on Rayleigh waves recorded at 650 portable stations deployed in the region. Interstation phase and group velocity dispersions for more than 18 000 paths were estimated using the two-station method and then inverted to produce phase velocity maps for 10–80 s period and group velocity maps for 10–60 s period. Local 1-D shear wave velocity profiles for each 0.5° × 0.5° grid node were obtained by inverting Rayleigh wave dispersions obtained in this study together with previously published Rayleigh wave group velocities between 60 and 145 s and then assembled into a 3-D shear wave velocity model. The images obtained reveal an obvious mid-crustal low-velocity zone (LVZ) and low-velocity anomaly in the upper mantle beneath the Songpan–Ganzi terrane. The mid-crustal LVZ can be explained by the presence of partial melting, most likely from asthenospheric upwelling. The existence of a relatively weak mid-crustal LVZ beneath the Qilian orogeny is also confirmed. This LVZ is likely caused by the presence of deep crustal fluids. The Yinchuan–Hetao graben is characterized by relatively low velocities that extend to at least 200 km below the Earth’s surface, strongly contrasting with the seismically fast lithosphere of the Ordos and Alxa blocks. The model presented here shows the presence of a relatively thick and high-velocity lithosphere beneath the Ordos and NE Alxa blocks, as well as evidence of thinning of the lithosphere beneath the southwestern Alxa and Qilian blocks, indicating that the Alxa block is not subducting beneath NE Tibet.

Key words: Asia; Seismic tomography; Surface waves and free oscillations; Continental tectonics: compressional; Crustal structure; Dynamics of lithosphere and mantle.

1 INTRODUCTION

The northeastern (NE) Tibetan Plateau is bounded by the Alax and Gobi Deserts to the north and the Ordos basin to the east, and is considered as an ideal site for understanding continental dynamic processes (Yin & Harrison 2000; Tapponnier et al. 2001; Wang et al. 2016). The NE Tibetan Plateau is composed of several NW-trending terranes, including the Qaidam–Qinling, Qilian and Songpan–Ganzi terranes (Fig. 1), which all continue to be deformed in response to the ongoing collision of continental India and Eurasia (Meyer et al. 1998; Yin & Harrison 2000). The presence of Quaternary folds and thrust faults in NE Tibet indicates that this region is undergoing significant shortening and active deformation (e.g. Meyer et al. 1998). However, the mechanism responsible for crustal thickening of the Tibetan Plateau is not fully understood and several models are under debate.

Tapponnier et al. (2001) have suggested that the oblique crustal thrusting accommodated by the extrusion was responsible for the crust shortening and uplift of the Tibetan Plateau, while a crustal channel flow model was proposed by Clark & Royden (2000) to explain the high topography in the eastern plateau. Recent geophysical observations have also identified obvious intra-crustal low-velocity zones (LVZs) beneath the NE Tibetan Plateau (Yang et al. 2012; Bao et al. 2013; Jiang et al. 2014; Li et al. 2014a), which lends support to the crustal flow model. It remains debatable whether these LVZ are continuous and correspond to actual crustal flow.

The Alax and Ordos blocks are regarded as the western part of the North China block, which is part of the Archean Sino-Korean Craton (SKC, Zhao et al. 2005). The graben basins around the Ordos block are major tectonic features of late Cenozoic extension in North China (Zhang et al. 1998; Ren et al. 2002). Two rift systems, the Yinchuan–Hetao graben system and Shanxi graben system
developed in western North China during the Late Tertiary (Fig. 1). The Yinchuan–Hetao graben system consists of two individual grabens: the NNE-trending Yinchuan Graben and the E-W-trending Hetao Graben (Zhang et al. 1998). The widespread development of Mesozoic and Cenozoic extensional basins and extensive intraplate volcanism in eastern China suggest that cratonic mantle lithosphere has been lost beneath the eastern NCC (Deng 1988; Ren et al. 2002), an idea which has been verified by recent seismic studies (Chen et al. 2009; Zhao et al. 2012; Bao et al. 2013; Li et al. 2013b). However, whether the lithosphere beneath the western SKC also suffered similar destruction or not remains in question (Chen et al. 2009; Li et al. 2013b).

The role of the Asian lithosphere in the evolution of the Tibetan Plateau also requires examination. One hypothesis is that southward-subducting Asian lithosphere beneath the northern Tibetan Plateau developed in response to the Indo-Asian collision (Meyer et al. 1998; Tapponnier et al. 2001). Results from some recent P and S receiver function images show prominent south-dipping interfaces beneath northern Tibet (Zhao et al. 2010), which have also been used to support the hypothesis. The recent S-wave receiver function study of Ye et al. (2015) also found a south-dipping interface extending continuously from the Alxa interior to the Qilian Orogen, which has been interpreted as evidence of Alxa block underthrusting beneath NE Tibet. However, the receiver function images of Yue et al. (2012) found no interfaces associated with subduction of the Asian lithosphere. Additionally, a relatively low-velocity structure in the upper mantle was imaged beneath the Qilian Orogen by seismic tomography (Zhang et al. 2011; Ceylan et al. 2012; Nunn et al. 2014), which is also inconsistent with south-directed subduction of Asian lithospheric mantle.

A detailed picture of the structure of the crust and upper mantle is fundamental to improving our understanding of tectonic processes, and the evolution of the NE Tibetan Plateau and its adjacent regions. Such a detailed picture can be provided by surface wave tomography, which has proven to be an efficient tool in determining lateral velocity variations in the crust and upper mantle. Both large-scale regional surface wave results (Priestley et al. 2006; Li et al. 2013b) and several smaller scale regional and local surface wave results (Zhang et al. 2011; Yang et al. 2012; Bao et al. 2013; Li et al. 2013a,b, 2014a; Jiang et al. 2014) have been presented for the NE Tibetan Plateau and the western SKC. However, their resolutions are highly variable because of sparse distribution of seismic stations. Since 2013 September, new broad-band seismic stations have been temporarily installed in this region (Fig. 2), particularly in poorly resolved areas (e.g. Alxa and Ordos blocks). This will allow for new tomographic images to be captured at enhanced resolution and, in this paper, we investigate the crustal and upper-mantle structure beneath the NE Tibetan Plateau and adjacent region using Rayleigh wave tomography. The resulting images provide important insights into the geodynamic processes that have shaped this region.

2 DATA AND DISPERSION MEASUREMENTS

2.1 Data

To measure Rayleigh wave dispersions beneath the NE Tibetan Plateau and its adjacent region, seismograms recorded at 650 broad-band seismic stations from ChinArray (Phase II) were used (Fig. 2). The ChinArray (Phase II) operated from 2013 September to 2016 March and consists of 676 portable seismic stations deployed at the NE margin of the Tibetan Plateau, with station spacing in the 40–70 km range. Each station consists of a Guralp CMG-3ESP with a corner period of 60 s, or a CMG-3T seismometer with a corner
period of 120 s, coupled with a Reftek 130 data acquisition system with a sampling rate of 100 Hz. This array offers a higher station density than has previously been available in this area.

The traditional two-station technique was used for estimation of the phase and group velocities in this study. This method is based on the principle that, for a given earthquake with its epicentre lying along a common great circle joining a pair of stations, the differential travel time of a surface wave between these two stations reflects the underlying structure between these two stations. To ensure that the great circle path (GCP) assumption is valid, we only used station pairs for which the angle between the backazimuths of the far station to the epicentre and to the near station was $\leq 3^\circ$ (Fig. 3a). All events that occurred for the 2013–2015 period with a magnitude $\geq 5.5$ were initially selected from the United States Geological Survey catalogue. To ensure we can obtain well-developed surface waves train, we only chose events with focal depths $\leq 70$ km and epicentral distances between 10$^\circ$ and 120$^\circ$. We also limit the interstation distances to be $\leq 800$.

A total of 147 events were retained that each satisfy the above criteria and were well recorded at two stations (Fig. 4a). Before determining the interstation group and phase velocities, the instrument response, the mean and travel are removed from the vertical components; and then the 100 Hz seismograms are decimated to 1 sample per second. An example of the vertical components of the Rayleigh wave caused by an earthquake located in the Bismarck Sea and used in this study is shown in Fig. S1 in the Supporting Information.

2.2 Rayleigh wave dispersion measurement

The two-station method has been extensively used to measure dispersion curves of surface waves. In this study, the wavelet transformation method of Wu et al. (2009) is adopted to measure interstation phase velocity, which allows for the measurement of dispersion curves with high precision. To determine interstation group velocity, the frequency-domain Wiener filtering technique is first applied (Hwang & Mitchell 1986) to construct the interstation Green’s function. Then, the interstation group velocities are calculated by applying the frequency–time analysis technique (Wu et al. 2009) to the Green’s functions.

For each given pair of stations, all measurements from different earthquakes are combined to compute average Rayleigh wave phase and group velocity dispersion curves and their standard deviations. The measured phase and group velocity dispersions for the station pair of 51 528–63 059 are shown in Fig. 3(b). These average uncertainties lie in the 0.02–0.05 km s$^{-1}$ range for phase velocity measurements and in the 0.02–0.08 km s$^{-1}$ range for group velocity measurements, and is not heavily dependent on the wave period. More than 18 000 phase and group velocity dispersion curves were obtained with different interstation paths (Figs 4b, and 5a and b). The histogram of the number of dispersion measurements at each period shows a maximum in the 20–40 s range (Fig. 4b). The ray path coverage maps (Figs 5a and b) show that most of the study area has good path coverage.

Few reliable group dispersion measurements could be made at periods $> 60$ s and phase dispersion measurements at periods $> 80$ s, which are only confined to the central part of the study area. Therefore, inversion of phase and group dispersion curves has been performed in the period range of 10–80 and 10–60 s, respectively.

3 INVERSION FOR PHASE AND GROUP VELOCITY MAPS

3.1 Tomographic inversion

The retained interstation dispersion measurements are inverted to produce 2-D Rayleigh wave phase and group velocity maps. Using the same seismic data, Pan et al. (2017) generated Rayleigh wave phase velocity maps for periods between 10 and 80 s following a generalized 2-D linear inversion program developed by Ditmar & Yanovsky (1987). Instead of using their phase velocity dispersions directly, we constructed Rayleigh wave group and phase velocity maps on a $0.5^\circ \times 0.5^\circ$ grid using the tomographic method of Barmin et al. (2001).

The tomographic method of Barmin et al. (2001) seeks to minimize a penalty function that depends on damping ($\alpha$), path density ($\beta$) and correlation length ($\sigma$) factors. A large number of inversions were performed with different parameter combinations (Barmin et al. 2001), the final values of $\alpha$ and $\sigma$ are chosen through a standard trade-off analysis (Fig. S2, Supporting Information). In this study, values of $\alpha$, $\beta$ and $\sigma$ were selected as 400, 100 and 100 km, respectively, to obtain the final group and phase velocity maps.

The corresponding resolution of the Rayleigh wave group and phase velocity maps at three selected periods are shown in Fig. 5 and Fig. S3 in the Supporting Information. The resolution lengths for the group velocity map are similar to those for the phase velocity map, although the ray paths for group velocity are slightly sparser. The average resolution is in the 50–100 km range in the central part of the study area, but increases to ca. 150–200 km at the borders of the maps where ray coverage is quite limited. At shorter and...
Figure 4. Map showing (a) the distribution of teleseismic earthquakes used in this study and (b) the number of dispersion measurements as a function of period for Rayleigh wave phase and group velocity measurements.

Figure 5. Two-station path coverage (a, b) and horizontal resolution maps (c, d) for Rayleigh wave phase (left) and group (right) velocity measurements at a period of 40 s.

longer periods, the resolutions are slightly weaker because of the reduced number of dispersion measurements (Fig. 4b and Fig. S3, Supporting Information).

3.2 Rayleigh wave dispersion maps

Rayleigh wave phase and group velocity distribution maps are shown at periods of 10, 24, 32, 40 and 60 s in Fig. 6. Only results within these areas which have a resolution length <300 km are shown. As Rayleigh wave phase velocity maps have been addressed elsewhere (Pan et al. 2017), only the lateral variation in group velocity maps is discussed here.

The sampling depth of Rayleigh waves increases with increasing period, and the depth sensitivity kernels computed for the model of Li et al. (2013b) at 38°N/103°E also show that Rayleigh wave phase velocity is sensitive to deeper features than group velocity at the same period (Fig. 7). The pattern of velocity variation in the group velocity map at a period of 40 s for the study area is similar
to that of the phase velocity map at a period of 30 s, confirming the consistency of the results. Conversely, the two maps at the same period perfectly illustrate the difference in depth sensitivity between group and phase velocities.

Lateral variation of group velocities in this study generally agree at short periods (10–50 s) with previous ambient noise tomography (ANT) results (Li et al. 2014a; Shen et al. 2016) and at long periods (>50 s) with the regional surface wave study of Li et al. (2013b). Owing to the dense ray path coverage and short ray path, the group velocity maps of the study area provide better lateral resolution compared with previous regional-scale group velocity maps (Li et al. 2013b; Shen et al. 2016).

Group velocity maps at short periods (<16 s) show that significant low-velocity anomalies are imaged in sedimentary basins, and...
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Figure 6. (Continued.)

Figure 7. Depth sensitivity kernels of Rayleigh wave group (left) and phase (right) velocities calculated using the model of Li et al. (2013b) at 38°N/103°E.
the rift valley (e.g. Yinchuan graben) around the western edge of Ordos basin, has the lowest velocities (∼2.7 km s⁻¹) in the whole region. In contrast, high-velocity anomalies (∼3.2 km s⁻¹) are correlated with the locations of mountain ranges (e.g. Qinling). These observations have been identified in previous tomographic maps of Rayleigh waves (Yang et al. 2012; Jiang et al. 2014; Li et al. 2014a; Shen et al. 2016).

Intermediate-period (20–50 s) group velocity maps are sensitive to crustal thickness and shear velocity in the lower crust and upper mantle. Low-velocity anomalies observed in the Songpan and Qilian terranes are indicative of thick crust (Li et al. 2014b) or a crust with an elevated temperature and/or fluid content (Priestley et al. 2006; Li et al. 2014a).

Group velocity maps at long periods (>60 s) are sensitive to lower crustal and upper-mantle structure and show that the western Ordos basin, northern Sichuan basin and Alax block are characterized by high-velocity anomalies related to a shield-like lithosphere structure. In contrast, the Songpan and Qilian terranes seem to be encircled by relatively low group velocities, indicating that most of the upper mantle beneath NE Tibet is seismically slow.

3.3 Comparison to previous dispersion maps

To confirm the reliability of the dispersion maps, the phase and group velocity maps presented here are compared with previous dispersion results which overlap with our results at the period range of 10–70 s for phase velocity maps and 10–50 s for group velocity maps.

Comparisons between phase velocity maps from this study and those of Pan et al. (2017) are shown in Fig. 8. The maps of the aforementioned authors applied a different tomographic method to the same data sets. The phase velocities of Pan et al. (2017) show good agreement with this study. For example, at 32 and 40 s, the average difference between the phase velocity maps is ∼0.01 km s⁻¹ and the standard deviation of the difference between the maps is ∼0.04 m s⁻¹, which is within the uncertainties of phase velocities (Pan et al. 2017).

The group velocity maps are also compared with those based on ambient noise data (Shen et al. 2016) and earthquake data (Li et al. 2013b), as shown in Fig. S4 in the Supporting Information. Shen et al. (2016) inverted group velocity maps across China in the 8–50 s period range based on ambient noise data. These two sets of group velocity maps between 10 and 50 s are quite similar. The mean difference between these maps range from 0.01 to 0.05 km s⁻¹ and the standard deviation of the difference range from 0.08 to 0.10 km s⁻¹, which lie within the uncertainties of group velocities (Shen et al. 2016).

There is also an overlapping frequency range (in this case 10–60 s) between results of this study and those measurements from earthquake data based on the single-station method (Li et al. 2013b). The comparison of group velocity maps indicates that, within the seismic array, the two models show good similarity, although some of the small-scale anomalies in the dispersion maps are absent in group velocity maps of Li et al. (2013b). The consistency of results from different data set and methods enables the combination of long-period dispersion measurements from 60 to 145 s (Li et al. 2013b) to be used in constraining the structure of crust and lithospheric mantle.

4 SHEAR WAVE VELOCITY STRUCTURE

4.1 Shear wave velocity inversion

To construct an image of the shear velocity structure of the crust and upper mantle, a 1-D velocity model was inverted for each 0.5° × 0.5° gridpoint using Rayleigh wave dispersion measurements obtained in this study and long-period Rayleigh wave group velocities between 60 and 145 s periods from Li et al. (2013b). These 1-D inversion results were then combined to create a model of the 3-D shear velocity structures beneath the NE Tibetan Plateau and its adjacent regions.

The linearized iterative methodology of Herrmann (2013) was adopted to obtain the shear wave velocity model. Since the longest period of Rayleigh waves used here is 145 s, the model is not well resolved beyond a depth of 200 km (Fig. 7). The starting model is similar to the global 1-D reference model AK135 (Kennett et al. 1995), except that a constant velocity of 4.48 km s⁻¹ in the crust was used. The layer thicknesses of the starting model increase with depth: 1 km at depths of 0–6 km, 2 km at 6–50 km, 5 km at 50–100 km and 10 km at 100–300 km.

A differential inversion scheme was used in this study that damps the differences between velocities in adjacent layers. It was possible to only solve shear wave velocity by coupling P-wave velocities to S-wave velocities using a constant Poisson’s ratio of 1.73 because Rayleigh wave dispersion is primarily sensitive to S-wave velocities. During the inversion, a slightly higher damping value of 10 was used in the initial two iterations and then a lower damping value of 1 for the following 30 iterations.

The inversion of surface waves to obtain seismic velocity structure is a typical non-linear problem and therefore may be influenced by the starting model used in the inversions. To evaluate the uncertainty of the inversion results, numerous inversions were performed for a broader range of starting models and damping factors (Fig. 9). A total of 12 different starting velocity models were used for the velocity inversions (Fig. 9a). The starting models had different velocity gradients in the crust, ranging from 3.0–4.48 km s⁻¹ at the surface to 4.10–4.48 km s⁻¹ at depths >100 km. Thus, 36 inversions were conducted to investigate the sensitivity of the inverted results to the starting model and damping parameters. This included all combinations of the 12 different starting models with three different values of damping factors (1, 5 and 10).

The ranges of shear wave velocities resulting from different starting models and damping parameters are shown in Fig. 9(b). It was found that the general features of the resulting velocity structure remained stable, and statistical evaluation determined uncertainty of the inverted models to be 0.04–0.2 km s⁻¹ depending on depth. It is not surprising that the large uncertainties exist in the upper crust, given that the shortest period is 10 s.

The group and phase velocity maps produced in this study have a lateral resolution of 50–100 km, while the lateral resolution of the regional group velocity maps (10–145 s) produced by Li et al. (2013b) is 200–300 km. To assess whether the final model is controlled by the surface wave dispersion curves in this study or by the group velocity results of Li et al. (2013b), inversions were also performed based only on our data set without using the surface wave group velocities of Li et al. (2013b). A comparison of joint inversion is shown in Fig. S5 in the Supporting Information, and demonstrates that, in the top 120 km at least, the main constraint is higher resolution surface wave dispersion measurements obtained in this study.
4.2 3-D model

The shear wave velocity structure obtained in this study is shown in both horizontal (Fig. 10) and vertical cross-sections (Fig. 11). The structural features shown by these images are well correlated with the primary geological features.

The $S$ waves at shallow depths (5–15 km, Figs 10a and 11) show strong lateral velocity variations, with the lower velocities in Basins because of their thick sedimentary cover, and high velocities for the mountainous regions. These results agree well with those obtained in previous surface wave tomographic studies (Yang et al. 2012; Bao et al. 2013; Shen et al. 2016).
The lateral variation of S-wave velocity is larger at greater depths (e.g. 30 km, Fig. 10b). The 30 km depth represents approximately the middle crust within the NE Tibet and lower crust outside of Tibet. High velocities are evident beneath the Ordos block, eastern Alxa block and eastern part of the western Qingling orogeny. Low velocities are observed for the remaining part of the study area. In agreement with previous ANT studies (Yang et al. 2012; Bao et al. 2013; Jiang et al. 2014; Li et al. 2014a), a prominent mid-crustal LVZ is observed with velocities as low as 3.20 km s\(^{-1}\) at 20–40 km beneath the Songpan–Ganzi terrane and a relatively weak mid-crustal LVZ (as low as 3.35 km s\(^{-1}\)) beneath the northwestern Qilian Orogen (Figs 10 and 11). The mid-crustal LVZ beneath Songpan–Ganzi terrane does not terminate abruptly along the eastern Kunlun fault, but penetrates northward into western Qinling Orogen, which is consistent with the work of Jiang et al. (2014).

At 50 km depth, typical velocities of the lower crust (3.6–3.8 km s\(^{-1}\)) are observed in the NW Qilian Orogen and Songpan–Ganzi terrane, whereas higher velocities of about 4.0–4.3 km s\(^{-1}\) are observed in other regions, indicating the presence of uppermost mantle (Fig. 10c).

In the upper mantle at depths of 75–100 km, a low-velocity anomaly is still observed in the NW Qilian Orogen and Songpan–Ganzi terrane (Figs 10d and e), whereas a dominant high-velocity anomaly is evident beneath the Ordos and Alxa blocks. At depths of 75–200 km, the shear wave velocity model shows low-velocity anomalies along, and to the north, of the eastern Kunlun fault, which is consistent with the Rayleigh wave tomography studies of Zhang et al. (2011) and Li et al. (2013a). It is, however, different from the large-scale surface wave tomographic results of Priestley et al. (2006), who only observed a low-velocity anomaly in the uppermost mantle.

### 4.3 Estimation of crustal thickness

Previous studies (Lebedev et al. 2013, and references therein) show that surface waves are highly sensitive to the depth of the Moho. Surface wave studies also show that the depth with shear wave velocity of 4.0 km s\(^{-1}\) (Bao et al. 2013) or 4.1 km s\(^{-1}\) (Acton et al. 2010) derived from the inversion of dispersion curves correlate well with crustal thickness. In this study, it has been assumed that the depth of the 4.0 km s\(^{-1}\) S-velocity contour is a reasonable approximation for crustal thickness, but it does not mean that 4.0 km s\(^{-1}\) represents the upper mantle Vs in this region.

The depth of the 4.0 km s\(^{-1}\) S-velocity contour and estimations of crustal thickness from receiver functions (Li et al. 2014b; Wang et al. 2017) performed in this region are shown in Fig. 12. Significant lateral variation in the crustal thickness is evident, as is expected from the tectonic complexity of this region. Crustal thickness estimates obtained from Rayleigh waves across the Alxa block, Ordos block and western Qinling orogenic belt are generally in the 40–46 km range. This estimate increases across the Qilian block and western Qinling orogenic belt to 50–60 km west of 104°E, and to 70 km beneath the Songpan–Ganzi terrane. This variation is generally consistent with previous receiver functions (Li et al. 2014b; Wang et al. 2017) and deep seismic sounding (DSS) results (Teng et al. 2013).

An interesting regional feature is the existence of thicker than normal crust of about 46–48 km beneath the Yinchuan–Hetao graben. This thicker crust has been previously detected by receiver function studies (Ge et al. 2011; Wang et al. 2014), surface wave studies (Yang et al. 2012; Bao et al. 2013; Shen et al. 2016) and DSS studies (Tian et al. 2014), but not by the estimates of Sun et al. (1992) from DSS result. Some anomalously thin crust has been found around the western edge of the Ordos basin based on receiver function analysis (Fig. 12b). It is possible that the thick sedimentary structure significantly affects the Ps phase from the Moho at these stations.

Significant discrepancies between the depth of the 4.0 km s\(^{-1}\) contour and the estimation of crustal thickness from receiver function studies exist in the western part of the Songpan–Ganzi terrane. This has been previously observed and interpreted by Acton et al. (2010) to be the result of a low-velocity anomaly in the lower crust and upper mantle.

### 5 DISCUSSION

The 3-D shear wave velocity models presented in this paper show well-resolved structures of the crust and upper mantle below the NE Tibetan Plateau and adjacent areas and provide important insights into the geodynamic processes that have shaped this region. In the following section, the implications of these results are discussed. These include: (1) the origin of the mid-crustal LVZ and low-velocity mantle beneath the NE Tibetan Plateau, (2) the contrasts in lithospheric structure within the SKC and evidence for subduction having occurred beneath NE Tibet and (3) the evolution and formation of the Yinchuan–Hetao graben.
5.1 Low-velocity structure beneath the NE Tibetan Plateau

The tomographic images presented confirm the two mid-crustal LVZs observed by previous ANT studies (Yang et al. 2012; Bao et al. 2013; Jiang et al. 2014; Li et al. 2014a). However, the interpretation of these crustal LVZs is still under debate. For example, Li et al. (2014a) proposed that the LVZs beneath the NW Qilian Orogen may be associated with crustal thickening, but Jiang et al. (2014) interpreted the mid-crustal LVZs to be the result of partial melting.

The LVZs at 20 and 40 km depth beneath the Songpan–Ganzi terrane are the most prominent features in the shear wave velocity model (Figs 10 and 11). Receiver function studies (Wang et al. 2017) show that the average crustal $V_p/V_s$ ratios in this region
Figure 11. Cross-sections of shear wave velocity along six profiles shown in Fig. 10(f) (grey dashed lines). Topography is plotted above each profile. The thin dashed lines represent the estimated Moho from receiver function analysis (Li et al. 2014b; Wang et al. 2017), and the solid line in profile FF’ is the depth of the lithosphere–asthenosphere boundary (LAB) from Ye et al. (2015). The estimated lithospheric thickness from this study (thick dashed line) is also shown in the other five profiles. Note that the different colour scales are used for the crust and mantle and the abbreviations for the tectonic units are the same as in Fig. 1.

Figure 12. Comparison of (a) the depth of the 4.0 km s\(^{-1}\) S-velocity contour and (b) estimations of crustal thickness from receiver functions (Li et al. 2014b; Wang et al. 2017).
are in the 1.76–1.85 range; slightly higher than the global average (Christensen & Mooney 1995). The 2-D resistivity studies in this area imaged high electrical conductivity with resistivities < 1 Ω m starting at a depth of ~20 km (Zhan et al. in preparation), suggesting that the mid-crustal LVZs beneath the Songpan–Ganzi terrane are related to the presence of partial melts (Jiang et al. 2014; Li et al. 2014a). The low-resistive 

\[ V_p/V_s \] ratios revealed by receiver function analysis, however, does not suggest significant partial melting within the NE Tibetan crust (Vergne et al. 2002; Li et al. 2006; Tian & Zhang 2013).

Moreover, a partial melt explanation for the relatively weak mid-crustal LVZs in the NW Qilian Orogen is more difficult. First, the average crustal 

\[ V_p/V_s \] ratios in this area are in the 1.69–1.76 range (Li et al. 2006; Yue et al. 2012; Tian & Zhang 2013; Wang et al. 2017), which is close to the global continental average (Christensen & Mooney 1995). Secondly, relatively low- to moderate-surface heat flow (< 70 mW m\(^{-2}\)) is observed beneath the NW Qilian Orogen (Wang 2001). Thirdly, the NW Qilian Orogen is conductive, with minimum resistivities of ~10–100 Ω m that are higher than typical for pure melts (0.1–1 Ω m) or brine (0.01–10 Ω m; Sass et al. 2014, and references therein). However, the presence of aqueous fluids may explain the observed low shear wave velocity and highly conductive anomaly in the crust.

Xiao et al. (2012, 2013) suggest that these high conductivity layers are caused by the development of imbricated crustal thrusts, which has also been proposed as the primary mode of crustal thickening in NE Tibet (Vergne et al. 2002; Li et al. 2006; Tian & Zhang 2013). Faults and fractures are important channels for fluids which may contribute to the low shear wave velocity observed in this study, but fluid content fails to explain the observed resistivity structure of Zhan et al. (in preparation), where the upper crust beneath the Qilian Orogen is characterized by continuous high resistivity.

The mid-crustal LVZs in the NW Qilian Orogen may represent the ductile middle crust. The catalogue of the China Earthquake Network Center (CENC) shows that most large earthquakes (M > 5.0) in Qilian occur in the upper crust (<20 km), indicative of a brittle upper crust and ductile mid-to-lower crust. The brittle–ductile transition has been proposed as a mechanical trap to deep crustal fluids (Bailey 1990), which may explain the presence of the LVZ found in this study, as well as the resistive upper crust and conductive middle crust observed by Xiao et al. (2012, 2013). These deep crustal fluids may be released during metamorphism induced by subduction–accretion processes beneath the Qilian Orogen (Yin & Harrison 2000).

Previous seismic studies (Priestley et al. 2006; Barron & Priestley 2009; Shi et al. 2013a,b) and ANT tomographic studies (Yang et al. 2012; Shen et al. 2016) show that the mid-crustal LVZ imaged beneath the Songpan–Ganzi terrane continues into the lower crust and uppermost mantle, which may also contain some melt (Barron & Priestley 2009). Two distinct mechanisms and processes have been proposed to explain the partial melting in the crust and upper mantle of NE Tibet Plateau. The first is radioactive heating in the thickened crust (McKenzie & Priestley 2008) and the second is asthenospheric upwelling caused by lithosphere thinning (Chung et al. 2005).

The observation that low-frequency Sn propagate across the entire plateau (Barron & Priestley 2009), together with the intact high-velocity lithospheric lid throughout the plateau (Priestley et al. 2006) has been used to support the idea that the low shear wave velocity anomaly in the crust and sub-Moho mantle derives from radioactive heating of the thickened Tibetan crust. However, if the seismic lithosphere beneath NE Tibet is still intact, the signature of a high-velocity seismic lid should be observed.

The tomographic images presented clearly indicate that the low-velocity anomaly in the Songpan–Ganzi terrane was not confined to the shallow mantle, but continues into the upper mantle with depths of at least 200 km (Fig. 11). This is consistent with previous surface wave and body wave tomographic results (Zhang et al. 2011; Li et al. 2013a,b; Lei & Zhao 2016). The models also show that the low-velocity anomalies with velocities of 4.20–4.35 km s\(^{-1}\) span the eastern Kunlun fault (Figs 10 and 11). Thus, the results and observations presented support the interpretation that partial melting is probably related to asthenospheric upwelling caused by delamination of a thick lithosphere root, which is also consistent with the interpretation of recent surface wave studies (Li et al. 2013a,b).

### 5.2 Lithospheric heterogeneity within the western SKC

A seismically fast lithosphere (lid) has been clearly imaged beneath the western SKC by different scale surface wave tomographic models (Priestley et al. 2006; Bao et al. 2013; Li et al. 2013b; Shen et al. 2016). More complex structures have been found in this region than previous regional- and local-scale surface wave studies because of the installation of new seismic stations in the western SKC. In the models presented, most of the western SKC appears to be underlain by high-velocity anomalies from the top of the upper mantle down to a depth of 130 km (Fig. 10). The maximum shear wave velocities in the lithospheric lid for the western SKC are also found to be lower than in other Precambrian Cratons (Li et al. 2003; Priestley et al. 2006) and heat flow in the western SKC of about 45–65 mW m\(^{-2}\) (Wang 2001) is higher than in other cratonic areas (30–40 mW m\(^{-2}\), Dorofeeva & Lysak 2010).

The new models presented show that the mantle structure of the western SKC is not homogeneous. The maximum lid velocities for the northern part of the Ordos block are slightly higher than beneath the southern segment of the Ordos block (4.45–4.60 km s\(^{-1}\), Fig. 10c). A few small to moderate size (M < 6) earthquakes in the southern Ordos block have been recorded, while the northern Ordos block is seismically quiescent (Fig. 1). It is concluded that the seismic velocity contrast in the lithospheric mantle likely corresponds to differences in lithospheric strength. Inside the Alxa block, several small-scale seismic low-velocity anomalies appear, and similar features are found in the body wave traveltime tomographic model of Zhao et al. (2012). A high-velocity anomaly beneath the western part of the Alxa block is observed down to a depth of about 100 km, indicating a thinned lithosphere but, in contrast, the high-velocity anomaly beneath the Ordos block is observed at greater depths (~120–130 km), indicating a thicker lithosphere in this area.

The velocity models obtained by inverting dispersion of fundamental-mode surface waves do not definitively resolve the lithosphere–asthenosphere boundary (LAB, Li et al. 2003; Eaton et al. 2009). Previous seismic studies show that the subcratonic lithosphere is imaged as a high-velocity lid, which can be regarded as a thick lithospheric root (Priestley et al. 2006; Bao et al. 2013; Li et al. 2013b). In this study, the method of Li et al. (2003) is followed in which the depth to the base of the negative velocity gradient below the high-velocity lid is used as a proxy for the seismic LAB.

The seismological definition of the LAB is only applied to stable terranes (e.g. Alxa and Ordos blocks), where the lithosphere–asthenosphere contrast in terms of seismic velocity is more evident.
The depth of the maximum negative gradient obtained from the 1-D velocity models along vertical depth profiles is shown in Fig. 11. It is found that the lithospheric thickness beneath the western SKC is in the 100–140 km range, which is consistent with the lithospheric thickness reported by Wang & Cheng (2012) and Li et al. (2013b), but is much thinner than the lithospheric thickness of >200 km reported by Chen et al. (2009).

A recent analysis of S-wave receiver functions (Ye et al. 2015) showed that the LAB dips southward from a depth of ~120 km beneath the Alxa block to a depth of ~160 km beneath the Qilian Orogen. This has generally been regarded as evidence of the North China plate underthrusting beneath the NE Tibetan Plateau. If southward-subducting Alxa lithospheric mantle occurred beneath the Qilian Orogen, thick and high-velocity lithospheric mantle would be expected, but a cross-section through the Ordos and Alxa blocks (Figs 11d–f) illustrates a relatively thick region of high velocities beneath the Ordos and NE Alxa blocks, as well as thinning of the lithosphere towards the Qilian Orogen to the west and south. Lower velocities in the upper mantle between the Alxa block and Qilian Orogen have been observed, which is in agreement with recent seismic tomographic studies (Zhang et al. 2011; Ceylan et al. 2012; Nunn et al. 2014) and also argues against subduction of the Qaidam beneath the Tibetan Plateau. The existence of thicker and higher velocity lithosphere beneath the Ordos and eastern Alxa blocks suggests that the lithosphere beneath the western Alxa block must have been modified, possibly during the Indo-Eurasian collision.

5.3 Structures beneath the Yinchuan–Hetao graben

As a result of the newly installed stations, a low-velocity anomaly was observed beneath the Yinchuan–Hetao graben system (Figs 10, and 11c and d). The low shear wave velocities at upper crustal depths (<15 km) throughout the Yinchuan–Hetao graben are associated with thick sedimentary cover that has a maximum depositional thickness of 10 km (Zhang et al. 1998). The DSS study of Sun et al. (1992) and magnetotelluric results of Dong et al. (2014) indicate the presence of a low resistivity and low-velocity layer in the mid-to-lower crust below the Yinchuan–Hetao graben. Recent receiver function H–A analyses (Ge et al. 2011; Wang et al. 2014) show moderate-to-high P/VS (1.76–1.89) in this area. Ge et al. (2011) attributed these high ratios to the presence of partial melt, while Wang et al. (2014) interpreted them to be indicative of mafic lithologies in the crust. In addition, this study finds no evidence of a crustal LVZ in this region which is in agreement with the recent DSS study of Tian et al. (2014). This observation, together with moderate thermal activity of the essentially amagmatic Yinchuan–Hetao graben (heat flow of 56–64 mW m⁻², Dorofeeva & Lysak 2010), may suggest that partially molten rocks at mid-to-lower crustal levels are either absent or too narrow to be resolved.

Normal Pn velocities were observed beneath the Yinchuan–Hetao graben by Hearn et al. (2004), who inferred for it a nascent stage of continental rifting. On the contrary, the tomography results presented here produce clear images of a low-velocity anomaly at uppermost mantle depths (50 and 75 km) beneath the Yinchuan–Hetao graben. This feature is consistent with the recent Pn wave tomography study of Li et al. (2011) that was based on a larger data set than Hearn et al. (2004).

The models presented here show that most of the Hetao–Yinchuan graben is underlain by low-velocity anomalies from the top of the upper mantle down to a depth of 200 km, which is interpreted as being the result of mantle upwelling and thermal weakening of the lithosphere. This may have originated from the complex interaction of the Indo-Asian collision and subduction of the Pacific plate, as has been previously suggested (Ren et al. 2002).

6 CONCLUSIONS

A new shear wave velocity model of the crust and upper mantle beneath the NE Tibetan Plateau and western SKC has been presented. The model is based on surface wave observations from a dense broad-band array deployed in this region. Rayleigh wave phase and group velocities were measured using the two-station technique, and then inverted to produce 2-D dispersion maps, and ultimately a 3-D model of shear wave velocity was constructed by inverting Rayleigh wave dispersion measurements obtained, together with previously published Rayleigh wave group velocity measurements between 60 and 145 s.

The model is demonstrably consistent with previous studies, but also resolves a number of previously unknown features. The presence of two isolated crustal LVZs in the Songpan–Ganzi terrane and NW Qilian Orogen has been confirmed. These LVZs likely have two different origins, that is, the mid-crustal LVZs beneath the Songpan–Ganzi terrane represent a partial melt and the low-velocity structure of the upper mantle is associated with asthenospheric upwelling caused by the delamination of a thick lithospheric root. The relatively weak crustal LVZ beneath the Qilian Orogen is attributed to deep crustal fluids associated with the subduction–accretion processes in this region. The tomographic images highlight a low-velocity anomaly that extends from the uppermost mantle down to 200 km depth beneath the Yinchuan–Hetao graben, which may be related to thermal weakening of the lithosphere. The results display a seismically fast lithosphere beneath the Ordos and NE Alxa blocks to depths of between 100 and 140 km. The southwestern Alxa and Qilian blocks have lower velocities than the Ordos and NE Alax blocks at lithospheric mantle depths, suggesting that the Alxa block is not subducting beneath the NE Tibetan Plateau.

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REFERENCES


**SUPPORTING INFORMATION**

Supplementary data are available at *GJI* online.

**Figure S1.** Vertical-component waveforms recorded at station 51 528 (top) and 63 059 (bottom) used in this study to calculate the Rayleigh wave dispersion curves.

**Figure S2.** Trade-off plot of tested parameter combinations for 2-D inversion of Rayleigh waves group velocity at 32 s. The grey-filled circle marks the chosen set of parameters.

**Figure S3.** Horizontal resolution maps for Rayleigh wave phase (left) and group (right) velocity measurements at periods 10 (top) and 60 s (bottom).

**Figure S4.** Comparison of group velocity maps for periods of 20 (left) and 60 s (right). Top panel shows the 2-D group velocity maps obtained using two-station method for this study. Middle panel shows 2-D group velocity maps obtained using ambient noise tomography (Shen et al. 2016, left) and single-station method (Li et al. 2013b, right). Bottom panel show the difference between these two maps using two different methods and data sets.

**Figure S5.** (Bottom) Comparison between the Vsv model obtained from inversion of the phase and group velocity presented in this study and (top) from the joint inversion of dispersions in this study together with longer period group velocity from Li et al. (2013b). The location of cross-sections are shown in Fig. 10(f) (grey dashed lines).

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