An upper-mantle S-wave velocity model for East Asia from Rayleigh wave tomography

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A B S T R A C T

We present a new shear velocity model of the upper mantle beneath the East Asia region derived by inverting Rayleigh wave group velocity measurements between 10 and 145 s combined with previously published Rayleigh wave phase velocity measurements between 150 and 250 s. Rayleigh wave group velocity dispersion curves along more than 9500 paths were measured and combined to produce 2D dispersion maps for 10−145 s periods. The group velocity maps benefit from the inclusion of new data recorded by the China National Seismic Network and surrounding global stations. The increase in available data has resulted in enhanced resolution compared with previously published group velocity maps; the horizontal resolution across the region is about 3° for the periods used in this study. The new shear-wave velocity models indicate varying velocity structure beneath eastern China, which yields estimates of a lithosphere–asthenosphere boundary depth from around 160 km beneath the Yangtze block to approximately 140 km beneath the western part of the North China Craton (NCC), up to depths of 70−100 km beneath the eastern NCC, Northeast China, and the Cathaysia block. The models reveal the subduction of two opposite-facing continental plates under the southern and northern margin of Tibet. An obvious low-velocity anomaly appears in the top 200 km of the upper mantle beneath northern Tibet, which is inconsistent with the presence of subducted Asian or Indian mantle lithosphere beneath northern Tibet. The Cenozoic volcanism fields in the Mongolian plateau are characterized by an obvious upper mantle negative anomaly, but no signature of deep-seated plume was observed.

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1. Introduction

East Asia is a tectonically complex region with a variety of crustal types and tectonic styles (Fig. 1). The present tectonics of East Asia comprises active continental collision and subduction of oceanic plates, which results in continental orogenic processes coexisting with widespread lithospheric extension (e.g., Molnar et al., 1993; Ren et al., 1999; Yin and Harrison, 2000; Schellart and Lister, 2005; Yin, 2010; Li et al., 2012a). The India−Asia collision is interpreted to have led to crustal shortening and thickening in the Himalaya−Tibet region and possibly in regions as far north as Mongolia and Tienshan (e.g., Molnar et al., 1993; Yin and Harrison, 2000; Tapponnier et al., 2001). The subduction of the Pacific and Philippine Sea plates led to the development of the western Pacific marginal basins and the Mesozoic basins in eastern China (Ren et al., 2002; Li et al., 2012a). Moreover, it induced widespread mantle upwelling under East Asia and possibly as far west as the Baikal rift zone (Schellart and Lister, 2005). It is evident that these tectonic processes have produced strong seismic heterogeneities in the crust and mantle beneath East Asia. A detailed 3D seismic model of the crust and upper mantle is important for understanding the tectonic and geodynamic evolution in this region.

P and S velocity models of the deep earth structure beneath East Asia have been constructed based on the tomographic inversion of traveltime data (e.g., Huang and Zhao, 2006; Li et al., 2006a; Koulakov, 2011; Wei et al., 2012). These studies show that the upper mantle beneath much of Eastern China is characterized by relative low seismic velocities, with high velocities beneath several Precambrian blocks (e.g. the Ordos Basin, Sichuan Basin, and Tarim Basin) extending to depths between 200 and 350 km. However, the vertical resolution of body-wave images is highly variable because most of the incoming rays are nearly vertical. In addition, these body-wave tomographic models typically resolve relative velocities, rather than the absolute velocities, which are more important in understanding how velocities vary with depth.

Surface-wave methods can be used to constrain absolute seismic wave velocity in the lithosphere and asthenosphere with good depth resolution. A large number of surface-wave tomography studies at different scales have been carried out in East Asia and its surrounding regions. For example, the shear-wave
velocity structure beneath East Asia was investigated in larger scale (>500–1000 km) global (e.g. Ekström et al., 1997; Ritzwoller et al., 2002b; Shapiro and Ritzwoller, 2002; Ekström, 2011) or regional (e.g. Villaseñor et al., 2001; Zhu et al., 2002; Huang et al., 2003; Yanovskyaya and Kozhevnikov, 2003; Friederich, 2003; Lebedev and Nolet, 2003; Pasyanos, 2005; Priestley et al., 2006; Feng and An, 2010) surface wave studies. Smaller scale local surface wave studies have also been carried out in different areas of East Asia, such as the Tibetan Plateau (e.g. Yao et al., 2008; Chen et al., 2009b; Fu et al., 2010), Northeastern China (e.g. Li et al., 2012b) and North China (e.g. Pan et al., 2011). These studies based on earthquake data have revealed the lateral variation of mantle structures in different tectonic units, but global/regional-scale tomography suffers from low resolution, and local tomography lacks a broader perspective. Some recent studies based on seismic ambient noise have resolved the structure of the crust and upper mantle beneath mainland China with relatively high resolution (e.g. Zheng et al., 2010b, 2011; Zhou et al., 2012; Yang et al., 2010, 2012). The main limitation of these models is that only shorter period dispersions and shallow velocity structure are obtained.

Furthermore, despite the intensive seismic body- and surface-wave tomography studies (e.g. Villaseñor et al., 2001; Wei et al., 2012, and references above), comparison of these published results shows that they are often inconsistent with each other and even contradictory, which leads to incompatible interpretations regarding many important geological issues in this region, such as the details and fate of mantle lithosphere in the shortened and extended areas and the origin of the intraplate Cenozoic volcanism in NE Asia (e.g. Mongolia and the Baikal region).

Over the last five years, many new permanent seismic stations have been installed in China and adjacent areas (Zheng et al., 2010a). Data from permanent seismic stations located in and around China have accumulated considerably in recent years, allowing for new surface wave tomographic studies with improved path coverage and enhanced resolution. We present here the 3D shear-wave velocity structure of the uppermost mantle beneath East Asia, derived by inverting fundamental mode Rayleigh wave group velocity measurements between 10 and 145 s together with previously published Rayleigh wave phase velocity measurements between 150 and 250 s from Ekström (2011). The Rayleigh wave group velocity dispersions between 10 and 145 s were measured through frequency–time analyses using the Continuous Wavelet Transformation method (CWT) (Wu et al., 2009) and combined to form tomographic group velocity maps for the East Asia region. The models provide improved constraints on the crust and upper mantle structure, yielding new insights on the tectonic and geodynamic evolution in this region.

2. Data selection and analysis

2.1. Dataset and preprocessing

To estimate surface wave group velocities, we used waveform data recorded by approximately 184 permanent broadband seismic stations operating in East Asia (Fig. 2). The most important contribution to this study comes from stations of the China National Seismic Network (CNSN) operated by the China Earthquake Administration. The network was established at the end of 2007 and consists of 145 stations unevenly spaced across mainland China. Most of these stations are equipped with very broadband seis-
mometers (e.g. CTS-1, BBVS-120, KS-2000, STS-2, CMG-3ESP) and ultra broadband seismometers (e.g. STS-1, JCZ-1), which record at periods of 120 s or longer. Only three stations were equipped with broadband seismometers (e.g. CMG-3ESP, CMG-3TB) with a relatively lower recording range (cutoff period 60 s) (Zheng et al., 2010a). Four years of data (2008.01–2012.03) from these 145 broadband stations were made available for this study. In order to increase ray coverage, supplementary data from 39 stations in other networks (e.g. GEOSCOPE, KNET, and KZNET) operating between 1998 and 2012 were obtained from the IRIS (Incorporated Research Institute for Seismology) data management center and incorporated in the analysis. We analyzed seismograms from these stations for 525 intermediate-to-large magnitude shallow events (depth shallower than 70 km and Ms magnitude equal to or greater than 5.5) that occurred within the geographical boundaries of 0°–60°N latitude and 55°–150°W longitude.

The epicenters of the earthquakes as well as the locations of the seismic stations used in this study are shown in Fig. 2. Most of the earthquakes are distributed along the circum-Pacific belt and circum-Indian belt and a few earthquakes are located in western China and southern Russia. In contrast, there are almost no earthquakes within eastern China and the Indian subcontinent. The geographical distribution of earthquakes is similar to that in previous surface wave studies (e.g. Zhu et al., 2002; Huang et al., 2003; Feng and An, 2010), but this study includes data from many new seismic stations that are densely distributed in mainland China.

The waveform data were firstly deconvolved from the instrument response to correct the instrumental delay and filtered in the frequency range of 0.2–0.0067 Hz to protrude the fundamental mode surface waves. These records were then decimated from their original sampling rates to 1 sample per second. Each vertical trace was visually checked to ensure only good-quality recordings with high signal-to-noise ratio were kept for subsequent dispersion measurements.

2.2. Group velocity measurement

For each selected event–station pair, we used the seismogram of the vertical component to analyze the Rayleigh waves. The group velocities for the fundamental mode Rayleigh waves in the period range 10–145 s were obtained by applying the frequency–time analyses using CWT (Wu et al., 2009). Compared with the traditional narrow band pass method (Herrmann, 1973), CWT analysis uses bandpass filters with variable length for each period to obtain spectra with different widths and to ensure good dispersion measurements (Wu et al., 2009).

Each measurement was visually checked to increase the quality of the dataset and around 12,000 Rayleigh wave group velocity dispersions were determined in this study. The initial dataset contains considerable redundancy since many paths were very nearly repeated. The redundancy was used to obtain estimation of data uncertainty by clustering dispersion measurements from similar paths. In this study, measurements with path endpoints lying within 1% of the path length were grouped to produce a cluster with the mean properties of the constituent paths. The uncertainty of the Rayleigh wave group velocity measurements was in the range of 0.02–0.06 km/s and was not frequency-dependent.

After clustering, the maximum number of Rayleigh wave ray paths was more than 9500. Compared with previous studies (e.g. Zhu et al., 2002; Huang et al., 2003), our new data adds over 4500 paths. The almost 100% increase in the total number of paths allows for a considerable improvement in the tomographic resolution of East Asia. The number of ray paths as a function of period for Rayleigh waves is shown in Supplemental Fig. A1. The maximum number of measurements was for Rayleigh waves with periods of about 40 s, with fewer measurements at both shorter and longer periods. There were 8994 ray paths at 40-s period across this region, but only 2921 at 10-s and 2758 at 145-s period. The path lengths used for the surface wave analysis were well distributed between 850 and 7000 km with an average path length of ~2900 km (Supplemental Fig. A1). These ray paths provide good coverage of most of East Asia except the margins (Fig. 3).

3. Group velocity tomography

To invert our group velocity measurements for 2D group velocity maps, we applied a generalized 2D-linear inversion program developed by Ditmar and Yanovskaya (1987) and Yanovskaya and Ditmar (1990). The method is the generalization of the 1D inversion method of Backus and Gilbert (1968) for 2D problems and permits the retrieval of locally smoothed dispersion curves at every grid from a finite set of travel times along the ray paths crossing the study area.

We inverted the Rayleigh wave group velocities for a set of periods between 10 and 145 s on a 1° × 1° grid. The calculations were made by using several smoothing parameters of 0.05, 0.1, 0.2, and 0.3. A smaller model smoothing parameter will enable a rougher model (sharper solution) but a better data fitting (i.e., larger variance reduction), whereas a larger smoothing parameter will result in a smoother model but a worse data fitting (i.e., smaller variance reduction). After testing and thoroughly inspecting inversions for different solution error, model resolution and model smoothness, we selected 0.2 as the smoothing parameter in this study. Another criterion for the quality of the solution is the comparison of the initial mean square travel time residual and the remaining unaccounted residual. If for one path, the travel time residual exceeds three times the unaccounted residual, the corresponding path is eliminated from the data set and the solution is recalculated.

During the tomography processing, the corresponding standard error and spatial resolution at every node was also simultaneously evaluated. The standard errors associated with the regionalized...
group velocities ranged from 0.03 to 0.05 km/s, which is similar to those obtained in previous studies (e.g., Ritzwoller et al., 2002b). Examples of resolution maps and associated path coverage are plotted in Fig. 3 for the 40 s period measurement. The dense ray paths ensure that enhanced resolution is obtained over large parts of East Asia. For most parts of East Asia, the estimated lateral resolution length for the 40-s Rayleigh waves was about 200 km. At shorter and longer periods, the resolution was slightly weaker because of the reduced number of measurements (Supplemental Fig. A2).

Further resolution constraints are also presented in Supplemental Fig. A3, which show the checkerboard resolution test for 10, 40 and 145 s Rayleigh wave data with different synthetic input models. The initial spaced checkerboard models contained alternating velocity values of 3.8 and 4.2 km/s for the low- and high-velocity regions. We performed tests with $2^2$, $3^3$, and $4^4$ input patterns. Group travel times along the same paths as those of the real data (Fig. 3; Supplemental Fig. A2) were calculated for the input model. Then, the synthetic travel times were used to reconstruct the velocity model using the same set of regularization parameters as for the real data inversion. The spatial geometry was well reproduced over mainland China and Mongolia for the $2^2$ checkerboard map, but the magnitude was underestimated. Both the spatial geometry and magnitudes are well resolved in the majority of East Asia for $3^3$ or larger checkerboard maps, but the edge of the model (over the Indian continent and the East China Sea) shows somewhat smeared patterns or lower resolution due to a relatively limited number of crossing paths or poor ray coverage.

4. Rayleigh wave group velocity maps

We obtained fundamental mode Rayleigh wave group velocity maps for 17 periods between 10 and 145 s. Fig. 4 presents the Rayleigh wave group velocity maps at a number of representative periods. These group velocity lateral variations are related to the different tectonic and geological features presented in the complex East Asia region. As Rayleigh-wave energy penetrates deeper as the wave period increases, the group velocity maps demonstrate the lateral variations of the shear-wave velocity structure for increasingly greater depths.

The Rayleigh wave group velocity maps at shorter periods (10–20 s) are primarily sensitive to shear-wave velocity in the upper-middle crust and they show a strong correlation with the sediment thickness map produced by Laske and Masters (1997) (Figs. 4(a) and 4(b); Supplemental Fig. A4). Low group velocities are observed in some major basins of the East Asian continent and large parts of the marginal sea regions where thick sediments are present, as also shown in previous group maps (e.g., Huang et al., 2003; Pasyanos, 2005; Ekström, 2011), however, some significant smaller-scale features and specific velocity variations are identified and also seem to be more reliable in our study. For example, in the 80 and 100 s maps of Pasyanos (2005), the fast Ords and Tarim basin anomaly is missing. In the 100 s group velocity map of Huang et al. (2003), the slow Tianshan anomaly is absent. In addition, our group velocity maps show an increase in geometrical details as well as stronger contrasts as compared to the global group maps of Ekström (2011) (Supplemental Fig. A5).

5. Shear velocity inversion

5.1. Inversion for shear-wave velocity

To obtain an image of the shear velocity structure of the crust and upper mantle, we inverted for a one-dimensional velocity profile for each $1^\circ \times 1^\circ$ grid point in the region shown in Fig. 4 using our Rayleigh-wave group velocity maps from 10 to 145 s periods and Rayleigh-wave phase velocities between 150 and 250 s periods from Ekström (2011), and combined them into a 3D model. Because the longest period of Rayleigh waves used here is 250 s, we invert only for shear-wave speed down to 400 km depth (which will only be reliable to about 300 km). We use only Rayleigh waves, which are predominantly sensitive to vertically polarized shear-wave speeds ($Vsv$), but for simplicity in this study we also refer to it as the shear-wave speed or $Vs$ model.

A linearized iterative algorithm (Herrmann and Ammon, 2004) was performed to obtain our shear-wave velocity model. The reference models in the initial inversions combined the structure in the upper mantle from the global 1D reference model AK135 (Kennett et al., 1995) and the structure in the crust from a new detailed 3D Asian crustal model (APcrust) (Stolk et al., 2013). The S-wave velocity in this crust is linked to the P-wave velocity using a Poisson's ratio of 0.25. The thickness of each layer between the Moho and 100 km depth is 10 km, and 20 km below a depth of 100 km. Since Rayleigh-wave dispersion is primarily sensitive to S-wave velocities, only the shear-wave velocity was inverted. The $Vp/Vs$ and density in each layer was fixed during the inversion and the density was assumed to follow the Birch equation (Birch, 1960). In this study, we used a differential inversion scheme, which minimized both the magnitude of the error vector between the observed and computed velocities and the differences between adjacent layers,
Fig. 4. Fundamental mode Rayleigh-wave group velocity maps for (a) 10 s period, (b) 20 s period, (c) 32 s period, (d) 50 s period, (e) 80 s period, (f) 100 s period, (g) 120 s period, and (h) 145 s period.

to avoid large velocity changes between adjacent layers. When the residual between the observed and predicted dispersions remains constant, the change in models corresponding to various iterations is negligible. Two examples of the inversion results and data fit are shown in Supplemental Fig. A6. The error bars for the Rayleigh-wave group velocity represent uncertainties of two standard errors, as described in Section 3. The formal uncertainties for the phase-velocity model at individual grid points are not determined; thus, the error bars are period-dependent and result entirely from the model strength of the Rayleigh wave phase-slowness models at different frequencies. These predicted dispersions show a good fit with the measured dispersions.

Surface waves are strongly influenced by crustal structure. To evaluate the impact of the initial model, particularly the crustal thickness and velocity structure on the inverted structure, we also performed a test inversion where we use the SEAPS model (Sun et al., 2008) for comparison (Supplemental Figs. A6). A comparison between the mantle models obtained using SEAPS model and APcrust model is given by Supplemental Fig. A7. The maximum difference is observed at a depth of 80 km and is mainly related to the differences of the Moho depth between the SEAPS model and APcrust model. The influence of the initial crustal structure tends to decrease and becomes negligible at greater depth ranges.

The new Asian crustal model APcrust (Stolk et al., 2013) merges information on active seismic experiments and receiver function analyses from the USGS GCS database and recent literature; its crustal thickness for our study region is more reliable and covers a larger area than that of the SEAPS model (Sun et al., 2008). We therefore chose the APcrust-based model as the representation of the mantle structure.

Our group velocity tomography (10–145 s) has a lateral resolution of 200–300 km, while the lateral resolution of the global phase-velocity maps (150–250 s) produced by Ekstrom is much lower (~650 km). Therefore, the joint inversion of both datasets
may result in a significant resolution problem. To assess whether the final model is controlled by the group velocity maps or by the phase velocity maps, we also inverted the group velocity alone and compared the results with the results from the joint inversion of phase and group velocity. The comparison shows that the velocity characteristics of the 100–200-km depth slices are very similar (Supplemental Fig. A8). These very small differences show that our joint inversion results, at least in the top 200 km, are mainly constrained by higher resolution group velocity maps.

5.2. 3D models of East Asia

Depth slices from the preferred shear-wave velocity model are shown in Fig. 5, and cross-sections through the model are shown in Figs. 6–8. Above 250 km depth, our shear-wave velocity model shows predominantly low-velocity anomalies beneath northern Tibet, western Mongolia, Eastern China, and the marginal seas. The major low-velocity structure occurs beneath Eastern China and the marginal seas. The upper-mantle low-velocity anomaly image is broadly consistent with models reported in global and regional-scale tomographic studies (Shapiro and Ritzwoller, 2002; Huang et al., 2003; Friederich, 2003; Huang and Zhao, 2006; Priestley et al., 2006; Li et al., 2008; Feng and An, 2010; Wei et al., 2012). These low-velocity anomalies coincide with the high heat flow associated with widespread rifting and recent Cenozoic volcanism (Wang, 2001; Ren et al., 2002; Wang and Cheng, 2012). Other low-velocity regions are observed clearly beneath northern Tibet and western Mongolia, where they extend down to ~200 km. These low-velocity zones coincide with low Pn velocity and inefficient Sn propagation regions (Ritzwoller et al., 2002a; Pei et al., 2007). These regions are also characterized by magmatic activity that began in the Cenozoic (e.g. Deng et al., 1996; Barry and Kent, 1998; Yin and Harrison, 2000).

One of the most prominent characteristics of the model at 100 km depth is the high-velocity anomalies that occur beneath the stable platform or blocks, such as the Tarim Basin, West Yangtze Craton, Ordos Basin and the Indian continent. Other significant high-velocity anomalies are found beneath the southern edge of the Tibetan Plateau and the Hindu Kush/Pamir region. The models at 150 km depth show a continuation of the high velocities beneath much of the Yangtze Craton and Tarim Basin, whereas beneath the Ordos Basin the velocities are slightly slower than the reference model. At 200 km depth (Fig. 5), no high-velocity anomalies are observed beneath the stable blocks, whereas the Indian–Himalayan collision belt that is well evidenced at 100 km depth is still visible as a continuous high-velocity belt that extends across the Himalaya and Pamir regions. This high-velocity structure was also observed by previous tomographic studies and was associated with the subducting Indian plate (e.g. Friederich, 2003; Huang and Zhao, 2006; Priestley et al., 2006; Li et al., 2008; Hung et al., 2010; Wei et al., 2012).

6. Discussion

Our shear-wave tomographic images show well-resolved structures in the upper mantle beneath the East Asia region and provide strong constraints on the geodynamic processes that shaped this region. In this section, we consider three main topics: (1) contrasts in lithospheric structure beneath Eastern China, (2) the fate of the colliding Indian and Asian plates, and (3) the link between the mantle structure and volcanic activities in the Hangay Dome–Baikal rift regions.

6.1. Contrasts in lithospheric structure beneath Eastern China

Eastern China contains two major Precambrian blocks, the North China Craton (NCC) and the South China Craton (SCC), separated by Phanerozoic orogenic belts (Zhao and Cawood, 2012). The survival of the Archaean and Proterozoic basement as well as the presence of garnet-diamond facies mantle xenoliths from Paleozoic kimberlites in eastern China implies that the NCC and SCC were underlain by a thick cold lithospheric mantle in the Precambrian (Chi et al., 1992; Lu and Zheng, 1996; Griffin et al., 1998). Studies of xenolith-bearing Cenozoic basaltic lavas in eastern China suggest that the cool, depleted mantle root beneath eastern China has been thinning or is being replaced by hot, fertile mantle since the Mesozoic (Deng, 1988; Menzies et al., 1993; Deng et al., 1994; Griffin et al., 1998; Xu et al., 2000). Geophysical evidence of thin and warm lithospheric mantle beneath eastern China was also provided by global and regional seismic tomographic studies (e.g. Shapiro and Ritzwoller, 2002; Huang et al., 2003; Priestley et al., 2006; Wei et al., 2012). A recent body-wave tomography study (Zhao et al., 2012) shows that the boundary between the high-velocity and low-velocity mantle lithosphere within NCC and SCC coincides with the boundary...
of the Yangtze block and western NCC on the surface. Our tomographic images using more data and different tomographic techniques confirm these results (Fig. 5) but do not show the western NCC and Yangtze block as a high-velocity body extending deeper than 350 km (Zhao et al., 2012), which may be an overestimation of the craton root due to vertical smearing.

To estimate the thickness of the seismological lithosphere, we take the depth to the base of the negative velocity gradient below
the high-velocity lid as a proxy for the lithosphere–asthenosphere boundary (LAB) (Weeraratne et al., 2003). The results indicate significant differences in the depth of the LAB beneath eastern China (Fig. 6). The thinnest lithosphere is observed (~70–100 km) beneath most of northeast China, the eastern part of the NCC and the Cathaysia block, in agreement with the lithospheric thickness from surface-wave inversions (Huang et al., 2003, 2009; Priestley et al., 2006; Li et al., 2009, 2012b) and S-wave receiver function studies (Chen et al., 2009a). In the western part of the NCC, the base of the lithosphere is ~140 km deep, increasing in depth to ~160 km under the Yangtze block. This observation is consistent with the lithospheric thickness reported by Huang et al. (2003) and An and Shi (2006), but the estimated lithospheric thickness for western NCC is much thinner than the lithospheric thickness of >200 km reported by Lebedev and Nolet (2003), Priestley et al. (2006), and Chen et al. (2009a). Indeed, the western NCC group velocities for the period range of 100–145 s are similar or slightly lower than the global averages estimated for the AK135 model (Kennett et al., 1995). Therefore, the thinner lithospheric thickness beneath the western NCC (~140 km) cannot be attributed to the inversion process and must be real (Fig. 4; Supplemental Fig. A6). Similar difference of LAB in different studies have been reported in other locations, but the relationship between the LAB depth estimates from tomographic modeling and those from receiver functions is not clear (Fishwick, 2010).

The seismic (surface wave) estimates of lithospheric thickness illustrated in this study are very consistent with the thermal lithospheric thickness proposed for eastern China from surface heat-flow data (Wang and Cheng, 2012). The similarities between two studies that used two such different methods provide further validation for our results. If a very thick (up to 200 km) lithosphere existed beneath the NCC until the mid-Ordovician (Chi et al., 1992; Lu and Zheng, 1996; Griffin et al., 1998), the observed thin lithosphere beneath the entire North China Craton, NE China, and the Cathaysia block could be impacted by the lithospheric delamination or/and thermal upwelling in this region, as suggested by previous studies (e.g. Deng et al., 1994, 1996, 2004; Zheng et al., 1998; Xu, 2001).

6.2. The fate of the colliding Indian and Asian plates and evolution of the Tibetan Plateau

The ongoing collision of India with Asia since the closure of the Tethys Ocean gave rise to the Himalayas and the Tibetan Plateau. To understand the formation and tectonic evolution of the Tibetan Plateau, many geophysical studies have been carried out over the last decades to investigate its deep structure (e.g. McNamara et al., 1994; Huang et al., 2000; Kind et al., 2002; Tilmann et al., 2003; Zhou and Murphy, 2005; Chen et al., 2009b; Fu et al., 2010; Zhao et al., 2010; Hung et al., 2010; Zhao et al., 2011; Zhang et al., 2012). However, many questions regarding the continental lithosphere in this area remain unanswered. For example, did both the Indian and Asian lithospheres subduct like oceanic lithosphere? How far is the Indian plate subducted northward in Tibet? Did lithospheric delamination occur beneath Tibet?

High seismic velocities of the subducting Indian lithosphere have been imaged beneath the southern edge of the Tibetan Plateau in earlier seismic observations of this region (e.g. Tilmann et al., 2003; Huang et al., 2003; Priestley et al., 2006; Huang and Zhao, 2006; Li et al., 2008; Hung et al., 2010; Zhang et al., 2012; Wei et al., 2012). Our work also imaged the Indian slab (Fig. 7) and shows that the sub-horizontal Indian mantle lithosphere underthrusted the plateau as far as the Bangong–Nujiang Suture. From our tomographic results, the Indian plate does not seem to subduct beneath the entire Tibetan Plateau, which does not agree with previous surface-wave tomography studies (e.g. Zhu et al., 2002; Priestley et al., 2006), but is quite similar to recent body-wave tomographic results (Li et al., 2008; Hung et al., 2010; Zhang et al., 2012). In addition, we did not observe a continuous high-velocity anomaly extending to 400 km depth that represents the remnant of the Tethys Oceanic lithosphere (Priestley et al., 2006) or the downwelling Indian mantle (Tilmann et al., 2003; Li et al., 2008). This may imply that the Tethys Oceanic plate had detached and sunk into the lower mantle or been thermally eroded, while the Indian continental lithosphere stalled beneath the Tibetan Plateau because of its buoyant root.

The southward-dipping subducting slabs are also thought to have developed beneath northern Tibet in response to the Indo-Asian collision (e.g. Tappinier et al., 2001). Geophysical evidence that supports this hypothesis comes mainly from P- and S-wave receiver function images showing the prominent south-dipping con-
versions in the uppermost mantle beneath northern and central Tibet (Kind et al., 2002; Zhao et al., 2010; Zhao et al., 2011). In our model (Figs. 5 and 7), the high-velocity anomaly corresponding to what is presumed to be the Asian mantle lithosphere is clearly visible beneath the northern edge of Tibet at 80 km and 200 km depths. The plunging high-velocity anomaly shown on these N–S profiles (Fig. 7) is similar to the ALM/AML interface (Kind et al., 2002; Zhao et al., 2010; Zhao et al., 2011), except that the slab-like velocity anomalies are observed only beneath the Tarim and Qaidam Basins. Indeed, if two opposite-facing slabs overlap beneath central and northern Tibet (Zhao et al., 2010; Zhao et al., 2011), a high-velocity lithospheric mantle to a depth of 200–250 km will be expected beneath most, if not all, the Tibet plateau (Zhu et al., 2002; Priestley et al., 2006). However, our model reveals an obvious low-velocity anomaly in the top 200 km of the upper mantle beneath the northern Tibetan Plateau, which is also inconsistent with the interpretations of wholesale underthrusting of India beneath the entire Tibetan Plateau (Zhou and Murphy, 2005).

The low velocity anomaly beneath the northern Tibet is also a prominent feature in various scale seismic observations (Huang et al., 2003; Friederich, 2003; Huang and Zhao, 2006; Priestley et al., 2006; Li et al., 2008; Chen et al., 2009b; Koulakov, 2011; Zhang et al., 2012; Wei et al., 2012). Our observation (Fig. 7) is in good agreement with some recent tomographic results (Huang and Zhao, 2006; Chen et al., 2009b; Koulakov, 2011; Zhang et al., 2012; Wei et al., 2012), but differs from the large-scale surface-wave tomographic models (Zhu et al., 2002; Priestley et al., 2006) where the low shear-wave velocities below northern Tibet are a relatively shallow feature and limited to a depth of about 100 km. The observations of the low-velocity anomaly along with widespread Quaternary volcanism suggest a weaker and hotter upper mantle beneath the northern Tibetan Plateau. The anomalously hot upper mantle could be related to mantle upwelling induced from subducted Indian lithospheric mantle (Tilmann et al., 2003) or lithospheric delamination (Molnar et al., 1993). At the same time, a weak and relatively hot upper mantle is also needed below northern Tibet to explain stronger deformation associated with the India–Eurasia collision (McNamara et al., 1994; Huang et al., 2000).

6.3. The Hangay Dome–Baikal rift regions

Our images show that there is a significant upper-mantle low-velocity anomaly beneath western Mongolia, south and west of the southern end of Lake Baikal (Figs. 5 and 8). A similar low-velocity anomaly was imaged in other regional and local-scale tomographic studies (e.g., Villaseñor et al., 2001; Friederich, 2003; Yanovskaya and Kozhevnikov, 2003; Priestley et al., 2006; Koulakov and Bushenkova, 2010; Koulakov, 2011). The surface projection of the shear-wave velocity anomaly correlates quite well with diffused Cenozoic volcanism in this region. More details about the igneous activities can be found in Barry and Kent (1998). Distinct geodynamical mechanisms were developed to explain the origin of the magmatism in this region, including but not limited to, (1) sub-horizontal asthenospheric flow driven by a step in the lithosphere thickness (Lebedev et al., 2006) and (2) mantle plumes and hot spots (e.g. Windley and Allen, 1993). In this section we evaluate these possible models according to our results and those from previous studies.

The first type of model (Lebedev et al., 2006) attributed these volcanisms to the sub-horizontal asthenospheric flow, where mantle upwelling crossing from a thicker lithosphere beneath Siberia to a thinner lithosphere beneath the Baikal region and Mongolia leads to rifting and volcanism in this region. On the west–east velocity model cross section (Fig. 8) we observed that the upper mantle low-velocity anomaly expands beyond eastern Mongolia and links up to the west with the low velocities at this depth beneath northeast China. Thus, the lateral extent of the low-velocity anomaly shown in our S-wave tomographic image is too large to be explained by small scale convection driven by changes in lithosphere thickness, i.e. it is difficult for this model to explain volcanic activities about 700 km to the east of the boundary. Therefore, our S-wave tomography model is inconsistent with this type of geodynamic model.

Some authors have suggested a plume or hotspot to explain the Cenozoic volcanism in western Mongolia (e.g. Windley and Allen, 1993). This idea is supported by the observation that a plume-like low-velocity anomaly extends from near the surface to the mantle transition zone (MTZ) (Friederich, 2003; Zhao et al., 2006) or even to the lower mantle (Petit et al., 1998) beneath the southern tip of Lake Baikal. However, our model shows that the low anomalies related with Cenozoic volcanism are mainly limited to shallow mantle depths (e.g. 80–200 km), and become much weaker (approximately 1%) and even absent in the deeper (>250 km depth) structure beneath this region (Figs. 5 and 8). This observation is in agreement with some regional scale surface-wave and body-wave tomographic images (Priestley et al., 2006; Wei et al., 2012), and implies that there is no deep-rooted mantle plume beneath this region or that the plume is too small to be detected by our tomography (Koulakov and Bushenkova, 2010).

7. Conclusions

In this study, we calculated Rayleigh wave group velocity dispersion over the East Asian region using data from the CNSS and surrounding global stations. More than 9500 path-averaged measurements were used to create group velocity maps at periods of 10–145 s. The resulting new group velocity maps have enhanced resolution compared with previous global (Ekström, 1997; Ekström, 2011) and regional group velocity models (e.g. Zhu et al., 2002; Huang et al., 2003; Yanovskaya and Kozhevnikov, 2003; Fasyanos, 2005) and exhibit an excellent correlation with known tectonic features in the region. We also modeled the shear-wave velocity structure of the upper mantle beneath the East Asian region by inverting Rayleigh wave group velocity measurements between 10 and 145 s combined with previously published Rayleigh wave phase velocity measurements between 150 and 250 s from Ekström (2011). The new tomographic model provides important information about the deep structure beneath the East Asian region.

The new shear-wave velocity models reveal significant heterogeneity in the upper mantle beneath eastern China. High-velocity anomalies with varying depths were observed beneath the western NCC and the Yangtze block, whereas low-velocity anomalies were imaged beneathNE China, the eastern part of the NCC and the Cathaysia block. The seismic imaging revealed significant differences in the thickness of the lithosphere beneath Eastern China. The thinnest lithosphere (~70–100 km) was observed beneath most of northeast China, the eastern part of the NCC, and the Cathaysia block. In the western part of the NCC, the base of the lithosphere is ~140 km deep, increasing in depth to ~160 km under the Yangtze block. Our seismic estimates of the lithospheric thickness are similar to estimates from modeling of surface heat flow and thus appear to be robust.

Beneath Tibet, we observed the subduction of two opposing continental plates. The sub-horizontal Indian mantle lithosphere is subducting northward beneath southern Tibet, while the south-dipping Asian mantle lithosphere is imaged beneath the northern margin of Tibet down to 200 km depth. Low velocities compared with the surrounding mantle are observed beneath northern Tibet, indicative of a hotter upper mantle that is inconsistent with the interpretations of wholesale underthrusting of India beneath the
entire Tibetan plateau (Zhou and Murphy, 2005) and of the Asian mantle lithosphere beneath northern Tibet (Zhao et al., 2010; Zhao et al., 2011).

Our tomographic images show a clear correlation between the Cenozoic Basaltic volcanism of the Mongolian plateau and a regional scale upper mantle negative anomaly. This correlation was clear at shallower mantle depths (80–200 km), but absent at greater depth, implying no evidence of a deep-seated plume.

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Appendix A. Supplementary material
Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2013.06.033.

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