Tomographic imaging of the underthrusting Indian slab and mantle upwelling beneath central Tibet

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Abstract: To better understand the pattern of convergence between the Indian and Eurasian plates, we determine a high-resolution P-wave tomography of the crust and upper mantle under southern to central Tibet using a large number of high-quality data collected by the ANTILOPE-II and Hi-CLIMB projects. A significant low-velocity zone is detected above the northward underthrusting Indian slab beneath the Indus-Tsangpo suture, which may reflect fault zones or (incomplete) fragmentation as well as melts and/or fluids associated with the dehydration of the underthrusting Indian slab. The variations in the diving depth and extending distance of the Indian slab, under the ANTILOPE-II and Hi-CLIMB seismic profiles, may be caused by the differences in the viscosity contrast between the Indian slab and the surrounding mantle as well as strong structural heterogeneities in the upper mantle under the study region. Both high and low velocity anomalies are revealed in the mid- to lower crust under southern Tibet, which may reflect the complex pattern of crustal structure. The

rifts in southern Tibet exhibit different velocity structures, suggesting that different structures beneath those rifts.

Keywords: Tomography; rift zones; subduction processes; mantle dynamics; Tibetan Plateau

1. Introduction

The Tibetan plateau (Fig. 1), comprising several terranes (Dewey et al., 1988), is the largest area with thickened crust (Royden et al., 2008) and high elevation (Fielding et al., 1994), and it is an ideal region to study continental dynamics.

Although there are some discrepancies among the models for the tectonic evolution of the Tibetan plateau (e.g., Maheo et al., 2002; Tapponnier et al., 2001; Zhao and Morgan, 1987), these models hold the same viewpoint that the subduction of the Indian plate plays a key role in the evolution of the Tibetan plateau. Unfortunately, the specific pattern of the subduction (or underthrusting) is still not well understood due to the restrictions of data and methods, even in central Tibet where more data have been collected than the other areas. Some important questions are, for example, where is the northern limit of the northward advancing Indian slab? Does it extend to the Bangong-Nujiang Suture, Jinsha–River Suture or just the Main Boundary thrust (e.g., Hung et al., 2011; Li et al., 2008; Priestley et al., 2006; Zhang et al., 2012b)?

An obvious surface feature related to internal deformation of the Tibetan plateau is the north-south trending rift. The formation mechanism of these rifts, however, remains controversial. The rifts may be confined only within the crust, or may extend

to lithospheric depth (Yin and Harrison, 2000). The notion of vertical coherent deformation of the crust and upper mantle is supported by studies of earthquake focal mechanisms (England and Houseman, 1986). Normal faulting is likely to initiate when convective removal of the thickened lower lithosphere takes place (Molnar et al., 1993). Other explanations include a combination of eastward Tibetan extrusion and oblique convergence between the Indian and Eurasian plates (Tapponnier et al., 2001), and mantle upwelling induced by the delaminated lithosphere (Ren and Shen, 2008; Zhang et al., 2011).

Analyses of the seismic data recorded during the Hi-CLIMB project have resulted in several important findings on the structure beneath southern to central Tibet, such as crustal heterogeneity (Nabelek et al., 2009), pattern of the Indian slab underthrusting (Liang et al., 2011), and diversity of S-wave splitting between southern and northern Tibet (Chen et al., 2010). The main purpose of the present study is to image the three-dimensional (3-D) crustal and upper mantle structure under southern to central Tibet, using a large number of high-quality data recorded by the Array Network of Tibetan International Lithospheric Observation and Probe Experiments (ANTILOPE) project and the Hi-CLIMB project. Our present results shed new light on the 3-D seismic structure and dynamic evolution of the Tibetan plateau.

2. Data and method

We used arrival-time data of local, regional and teleseismic events recorded by two seismic arrays in central Tibet: ANTILOPE-II and Hi-CLIMB (Fig. 1). The Hi-CLIMB project has been well documented by several publications (e.g., Hung et

al., 2011; Liang et al., 2011; Nabelek et al., 2009), whereas the ANTILOPE project has resulted in only one receiver-function (RF) study until now (Zhao et al., 2010). ANTILOPE was a project for investigating the seismic structure of the Tibetan Plateau undertaken by the Institute of Tibetan Plateau Research, Chinese Academy of Sciences. Our combined data set includes 231 portable seismic stations: 169 from the Hi-CLIMB and 62 from the ANTILOPE-II (Fig. 1). The ANTILOPE-II data were recorded from September 2005 to October 2006, while the Hi-CLIMB data were recorded from 2002 to July 2005.

The hypocentral parameters of the teleseismic events were determined by the United States Geological Survey (USGS), whereas those of the local and regional events were determined by the China Earthquake Networks Center (CENC). We adopted the following criteria to select the local and regional seismic events for this study: (1) only earthquakes with magnitude (M) > 3.0 are selected; (2) the hypocentral parameters determined by the CENC are adopted, whereas poorly located events with a fixed focal-depth of 33 km are eliminated (the location accuracy and its effects are discussed in the next section); (3) the minimum epicentral distance of Pn wave is set to be 3° (because the Moho discontinuity under the present study region is deeper than that under the other areas) in order to distinguish Pn and Pg phases. Teleseismic events ($M > 5.0$) with epicentral distance from 30° to 90° are used in this study. All the selected events have over 8 P-wave arrivals, while over 90% of the events actually have more than 15 P-wave arrivals. As a result, our data set contains 30,605 P-wave arrivals from 1,263 local and regional events (Fig. 1), and 54,468 first P-wave arrivals

from 1,578 teleseismic events (Fig. 2a). The ray path coverage is very good under the central part of our study region (Fig. 2b,c and d). In Fig. 2 the ray paths are shown schematically by straight lines connecting the hypocenters and seismic stations, but we actually used a 3-D ray tracing scheme (Zhao et al., 1992) to trace the rays precisely in the real computations. Note that the ray paths shown in the N-S and E-W vertical cross-sections in Fig. 2c,d are just part of the rays actually used: only the ray segments located with a 50-km width of profiles A-A' and C-C' (Fig. 1) are plotted.

By considering the influence of frequency-filtering on the accuracy of arrival-time pickings (1 s high-pass filter in this study), the picking error is estimated to be 0.1-0.3 s. Although applying the high or low pass filter could affect the onset points of arrivals, the actual difference in the arrival times is small, especially when the relative travel-time residuals are used. For more details of this issue, see Zhang et al. (2012a).

Theoretical arrival times are taken into account for identifying the P-wave arrivals (red lines in Fig. 3). Moreover, all of the collected Pg, Pn and P phases are the first arrivals at different epicenter distances, so we could easily identify the onset point on the original seismograms. Note that the Pn wave mainly propagates the uppermost mantle directly below the Moho discontinuity, which is quite different from the direct Pg wave. Thus, the Pn wave travel times are strongly affected by the Moho geometry and 3-D velocity variation in the uppermost mantle. The Pg, Pn and P phases are generally distinguishable on the seismograms. The Pn waves exhibit smaller amplitudes (see the upper four traces of Fig. 3a), whereas the Pg waves show up clearly with larger amplitudes (see the remaining 11 traces in Fig. 3a). All of the

P-wave arrivals we collected (including the local event Pg and Pn waves and the teleseismic P phases) have clear onsets, and they were manually picked up from the original seismograms (Fig. 3).

We used the tomographic method of Zhao et al. (1994, 2012) to determine a 3-D P-wave velocity model under central Tibet, which could jointly invert the absolute travel-time residuals of local and regional earthquakes and relative travel-time residuals of the teleseismic events. Based on the ray theory, the travel time of an individual seismic ray is the only matter we are concerned about. That means whether the two arrays deployed at the same time or not is irrelevant to our result. The starting velocity model is the iasp91 Earth model (Kennett and Engdahl, 1991) for the upper mantle and the CRUST1.0 model (Laske et al., 2013) for the crust (Fig. 4a). Both the station elevations and lateral variation of the crustal thickness beneath the study area are considered in the 3-D ray tracing. The Moho geometry is fixed in the inversion because the crustal structures near the two profiles (AA' and BB' in Fig. 1) have been well determined by the previous receiver-function studies (Nabelek et al., 2009; Schulte-Pelkum et al., 2005; Zhao et al., 2010). A 3-D grid is set up in the study region with a grid interval of 20-25 km in depth. The horizontal grid interval is 0.5° beneath the central part of the study region and 1° at the peripheral parts (Fig. 4b and c). Such grid intervals are adopted by considering the density and coverage of the rays in our data set. Because of the better ray path coverage of local and regional events at the shallower depth (Fig. 2b,c,d), the grid interval in the crust is smaller than that in the upper mantle. Similarly, the grid interval is smaller in southern Tibet than that in

northern Tibet because of the denser seismic stations there. The deepest point of the teleseismic rays is located in the lower mantle, but the ray path coverage is not well at depths > 400 km (Fig. 2e, f and g). In addition, our checkerboard resolution tests show poor resolution in areas deeper than 400 km (Fig. S1). Hence the bottom depth of our tomographic model is taken to be 400 km. Only the velocity perturbations at the grid nodes hit by more than ten rays are inverted, and they are shown in the resulting tomographic images. The LSQR algorithm (Paige and Saunders, 1982) with damping and smoothing regularizations (Zhao et al., 1994, 2012) is used to conduct the inversion (Fig. 5a). The variance of all the travel-time residuals is $\sim 0.5658 \text{ s}^2$ before inversion, and it is reduced to $\sim 0.0384 \text{ s}^2$ after inversion. Its reduction is 93%. The final root-mean-square residual is 0.196 s, comparable to the picking errors (0.1-0.3 s).

3. Analyses

3.1 Influence of crustal structures

Errors of the theoretical travel times are mainly caused by the crustal thickness, Moho geometry, station elevations and lateral velocity variations. Therefore, the crustal correction should be included in the ray-tracing and inversion. One layer is added to the 3-D grid model for simulating the high altitude of the Tibet (station elevation), and the ray path in this layer is assumed to be a straight line because of its short propagation distance (usually less than 6 km). Because of the variations of the crustal thickness and lateral velocity heterogeneities, seismic velocities under

different areas are different even at the same depth. To reduce the effect of the complex crustal structures beneath Tibet, we calculated the travel-time residuals using several crustal models: CRUST1.0 model (Laske et al., 2013), CRUST2.0 (Bassin et al., 2000), SEAPS (Sun et al., 2004, 2008), modified iasp91 (simply double the crustal thickness) and the receiver-function results (Nabelek et al., 2009; Zhao et al., 2010). The main features of the obtained tomographic results for the upper mantle remain the same, even if the different crustal models are adopted.

3.2 Effects of every data sets

The tomographic images near the Moho may be affected by vertical smearing, because both the absolute and relative travel-time residuals are included in our data set. Hence, we tried to conduct two tomographic inversions, one is that only the local and regional earthquake data are used, and the other is that only the teleseismic data are used. Because of the subvertical teleseismic rays under the study area and worse ray path coverage at the shallower depths, the shallow structure cannot be constrained well by the teleseismic data alone, and such vertical smearing appears in Figs. S2a and S2b. However, the low-velocity (low-V) zone at $\sim 87.5^\circ\text{E}$ is revealed by both of the inversions, indicating that this feature is not caused by vertical smearing (Fig. S2). Similarly, the high-velocity (high-V) zone in the middle crust is also a reliable feature.

As mentioned above, we did not relocate the local and regional earthquakes in the inversion process simply because our present seismic array (Fig. 1) is unable to locate those events as well as the regional Chinese seismic network does. We made a test to examine the influence of the mislocation errors. We added random errors (± 5 km) to

the epicentral locations (latitude and longitude) of every event before the tomographic inversion. Then, we trace all the rays and conduct the inversion again, and we repeat this process for 100 times to derive many groups of solutions. After that, we calculate the variances of the velocity perturbations at every grid, and most of them are less than 0.06%. In fact, the focal depth is much harder to determine accurately than the epicentral location. When we simultaneously add random noises to the three hypocentral parameters (latitude, longitude and depth), the variances of the velocity perturbations increase to ~0.12%. Although there are only small influence on the variances of velocity perturbations, the variances of all the travel-time residuals are increased to 0.0455 s^2 (with random errors to the epicentral locations) and 0.1592 s^2 (with random errors to the hypocentral parameters), respectively. To better understand the influences of the hypocentral parameters on the tomographic images, we selected randomly two results from the 200 solutions (Fig. S3). Although the two results provide somewhat different information on the details of the geometry of the high-V and low-V zones, their general features are coherent and exhibit four principal characteristics: (1) low-V zones beneath the Pumqu-Xianza rift (PX) at 40-70 km depths; (2) high-V zones around the Lopu Kangri rift (LK) in the middle- to lower crust; (3) low-V zones in the crust being interrupted by some high-V zones. The main features in the upper mantle are almost the same. A noticeable diversity is velocity anomalies around the Tangre Yum Co rift (TYC) at 70 km depth. The small variances suggest the stability of our result, indicating that the small hypocentral mislocations would not affect the main features of the tomographic images. However, the

interpretation of the tomography near the Moho should be careful, especially beneath the TYC.

3.3 Checkerboard resolution tests

A convenient and effective way to assess the resolution and reliability of the obtained tomographic images is the checkerboard test (Humphreys and Clayton, 1988; Zhao et al., 1992). To conduct the checkerboard test, we first assigned negative and positive velocity variations ($\pm 3\%$) alternatively to neighboring grid nodes to build a synthetic input model, and then calculated synthetic travel times for the input checkerboard model. To simulate the picking errors in the observed data, we added random noises (-0.3 s to $+0.3$ s) to the synthetic data, and used the same algorithm for the observed data to invert the synthetic data. The areas where the input checkerboard pattern is recovered are considered to have a higher resolution. Generally speaking, the real velocity structure would not change as strong as the checkerboard input model, especially in the upper mantle. Because of that, our test shows that the anomalies with ~ 1 degree could be identified and it is sufficient for our interpretations. The results of our checkerboard test indicate that the central part of our study area is well resolved, whereas the resolution in the northernmost part is lower due to the sparse distribution of stations there (Figs. S4 and S5). Note that the resolution at 100 km depth is relatively poor, especially to the north of Bangong-Nujiang suture, because Pg or Pn waves from the local events do not penetrate to that depth and the teleseismic ray path coverage is not well there.

4. Results

The resulting 3-D P-wave tomography of central Tibet obtained by this study is shown in Figures 6 and 7. Before depicting the main features and discussing the geological implications of the results, we point out one feature of the tomographic technique used in this study. The joint inversion of local-earthquake arrival times and teleseismic residuals is capable of determining the absolute velocities in the crust and uppermost mantle but only relative velocity, comparing to the starting 1-D velocity model, is shown in our final result. The absolute P-wave velocity in the upper mantle cannot be obtained because relative travel-time residuals from the teleseismic events are used. This is a common feature of teleseismic tomography.

At the shallower depths, the most prominent feature is a relative low-V zone detected under the Himalayan block that extends from 40 to 70 km depth. To the north of this low-V zone there is a relative high-V zone beneath the Lhasa terrane, which likely reflects the differences in the crustal structure between central and southern Tibet. In the southern part of the Qiangtang terrane, the middle to lower crust still shows clearly a low velocity. Comparing the two low-V zones, we find that the low-V zone beneath the Himalayan block is interrupted by a high-V zone, whereas the low-V zone beneath the southern Qiangtang terrane looks more continuous. Moreover, the large early arrivals of teleseismic events (implying higher velocity than that of the reference model) at the stations around the Main Central Thrust also indicate the existence of high-V zones under that region (Fig. 5b).

Another noticeable feature in the crust is the different velocity anomalies along

the three north-south trending rifts. Low-V zones are revealed beneath the PX extending from the surface to ~170 km depth (Fig. 7c), whereas a high-V zone is detected beneath LK. The low-V zone coincides with a cluster of intermediate-depth earthquakes in southern Tibet (Liang et al., 2008). Our test results show that the high-V zone at 40 km depth along TYC is a robust feature. In the uppermost mantle, a significant high-V zone exists in the central part of our study region, but a low-V zone appears in the northern and southern parts (Fig. 6d).

The main feature in the tomographic vertical cross-sections (Figs. 7a and 7b) is a high-V zone beneath the Lhasa terrane at depths of 150-300 km. The high-V zone is 100 to 150 km wide but could be narrower due to the smoothing effect. Along the Hi-CLIMB profile, the high-V zone extends northward to 33°N latitude with a low dipping angle. However, the northern limit of the high-V zone beneath the ANTILOPE-II profile is restricted to ~32.5°N. In addition, a low-V layer (about 50 km wide) is visible beneath the ANTILOPE-II profile. At the southern part of both profiles low-V zones are detected under the Indus-Tsangpo suture. Local earthquakes ($M > 3.0$) that occurred during 1975 to 2011 within a 50-km thick of the profile are also shown in these vertical cross-sections (Fig. 7).

5. Discussions

5.1 The Indus-Tsangpo suture

Although our images of the high-V zone in the upper mantle under central Tibet and the low-V zone under northern Tibet are consistent with some of the previous results (e.g., Pei et al., 2011; Wittlinger et al., 1998; Zhang et al., 2012a; Zhao et al.,

2010; Zhao and Nelson, 1993), a distinct difference appears under the Indus-Tsangpo suture, indicating that complex structures exist beneath southern Tibet (Figs. 7a and 7b). We conducted a synthetic test to check if the low-V zone is caused by vertical smearing. We first make an input model that contains two low-V zones in the crust and upper mantle beneath southern Tibet and a continuous high-V zone in the upper mantle (Fig. 8a), then we inverted the synthetic data for the input model with the same procedure as for the checkerboard test. In this test we focus on the pattern of the underthrusting Indian slab, so other features of the inversion result are removed (Fig. 8b). The result of the synthetic test shows that the surrounding low-V zones may only affect the continuity of the high-V zone (the region marked by a dashed line) but cannot separate or cut through the whole high-V zone. Hence the low-V zone under Indus-Tsangpo suture is a reliable feature. This test also indicates that we could resolve a high-V zone between two strong low-V zones because of the good coverage of rays in our data set, and there is no obvious high-V zone in the south along our profile. We also examined the distribution of relative travel-time residuals of the teleseismic events (Fig. 5b), which clearly shows that delayed arrivals are prevailing in southern Tibet (except for the zone around the Main Central Thrust, as we discussed earlier). Considering the good ray path coverage in this area (Fig. 2), we believe that the low-V zone in the resulting image is a reliable feature.

One possible cause of the low-V zone under the Indus-Tsangpo suture is significant lateral physical or compositional variation within the Indian lithospheric mantle, which is caused by the fragmentation of the Indian slab during its

underthrusting (Liang et al., 2012). It is considered that the fragmentation could result from the curvature of the Himalayan subduction front (McCaffrey and Nabelek, 1998) or the rolling back of the Burma arc (Rao and Kalpna, 2005). Our image beneath the Hi-CLIMB profile supports such an interpretation, but it does not explain the feature beneath the ANTILOPE-II profile. Since low-V zones are revealed by this study whereas the Indian slab was clearly detected by the receiver-function study (Zhao et al., 2010), another interpretation is that there are some faults or an incomplete fragmentation that occurred in the underthrusting Indian slab (the details will be discussed later). Over 100 intermediate-depth earthquakes at the sub-Moho depth were recorded in this area, implying a brittle deformation there (Monsalve et al., 2006). A continuous underthrusting Indian slab, which extends to ~35°N beneath western Tibet, was detected by a tomographic inversion using the same method and the data from ANTILOPE-I (Zhao et al., 2013). Some subcrustal earthquakes may be expected beneath the Himalaya where strain rates are high if the Indian slab is sufficiently cool (Priestley et al., 2008), which further indicate that pervasive fragmentation is not likely to occur.

5.2 On the Indian plate subduction

Due to the collision of the Indian and Eurasian plates, about 1000 to 2800 km of total north-south convergence has taken place (Patzelt et al., 1996; Yin and Harrison, 2000), whereas the reconstruction of the initial geometry of the Indian crust shows that ~500 to 700 km of shortening has been accommodated beneath the Himalayan orogen (DeCelles et al., 2002; Robinson et al., 2007), suggesting that breakoff

probably occurred at an early stage of the Indian plate subduction (Guillot et al., 2003; Maheo et al., 2002). Most of the previous studies suggested that this process occurred at ~30 Ma. If that is the case, the high-V zone in the upper mantle (Figs. 7a and 7b) may not reflect the relict of the Indian slab, because the high-V zone extends vertically down to ~300 km depth and a colder and heavier Indian slab could not stay in the uppermost mantle for such a long time. Some tomographic results support this opinion (e.g., Replumaz et al., 2004; Van der Voo et al., 1999), suggesting that the continental lithospheric mantle delaminated down to deeper mantle (> 1000 km depth). An alternative interpretation is that this high-V zone reflects the detached Lhasa lithospheric mantle. Although our tomographic images more or less support this interpretation, it is hard to explain the hotter upper mantle beneath northern Tibet (e.g. Zhao et al., 2010), seismic anomaly above the 660-km discontinuity (Chen and Tseng, 2007), and the ultrapotassic and adakitic magmas that occurred at ~25 Ma ago (Chung et al., 2005). By considering the other geophysical results (e.g., Li et al., 2008; Zhao et al., 2010), the prominent high-V zone in the uppermost mantle could be interpreted as the northward advancing Indian lithospheric mantle. The depth of high-V zone up to 300 km is somewhat different from the receiver-function results using the same data set. This is not surprising, however, because seismic tomography can image well the lateral velocity variations, but may have a poor vertical resolution when the coverage of ray path is not well in some depth range.

5.3 Lateral structural variations

A significant difference in the tomographic images between the two N-S profiles

(Figs. 7a and 7b) is visible at the sub-Moho beneath the Lhasa terrane, where a low-V thin layer appears in the ANTILOPE profile (Fig. 7b). This feature may indicate different depths of the underthrusting Indian slab. A study of Sn waves suggests the existence of a thin sub-Moho layer with high attenuation there (Barron and Priestley, 2009), and the thin layer may result in a sharp Moho rather than a disrupted Moho because of the strong velocity gradient there. A rapid change north of the Indus-Tsangpo suture, from null to significant birefringence, is found along the Hi-CLIMB profile, whereas this feature is not visible along the INDEPTH-III profile (near the Yadong-Gulu rift, which is the biggest rift in southern Tibet), showing null birefringence through much of the Lhasa terrane (Chen et al., 2010). In addition, the nulls could be attributed to the thin low-V layer beneath the ANTILOPE profile, which also suggest the different underground structures between these two profiles. The Indian slab may not subduct to over 350 km depth beneath central Tibet, and so the thickened mantle transition zone there is not caused by the subducted Indian slab but due to the detached Lhasa lithosphere (Singh and Kumar, 2009).

Combining the geological and geophysical results so far, we propose a possible explanation for the clear low-V zone which makes the underthrusting Indian slab not continuous well. As the Indian slab underthrusts northward, the lower part of the Lhasa lithosphere was removed in the late Oligocene (Chung et al., 2005). Moreover, the Lhasa lithospheric keel should be completely removed due to the great thickness of collision-thickened Lhasa lithosphere (Fielding, 1996). The loss of Lhasa lithosphere allowed the Indian slab to advance further north and encountered the

Qiangtang lithosphere, and downwelling of the thickened Qiangtang terrane occurred at ~15 Ma (Chung et al., 2005). However, only half of the removed lithosphere is detected above the 660-km discontinuity (Chen and Tseng, 2007). Because clear Asian lithospheric mantle was observed beneath the Qiangtang terrane (Zhao et al., 2010), the Qiangtang lithospheric keel may not have been totally removed. The remnant Qiangtang lithosphere blocked the northward advancing Indian slab, leading to the fragmentation on the Indian slab. However, since we detected the Indian lithospheric mantle beneath the ANTILOPE-II profile by using receiver function technique (Zhao et al., 2010), we infer that the fragmentation of Indian slab is not totally completed (at least beneath the ANTILOPE-II profile) and maybe only some faults are generated. As faults or fragmentation are occurred on the northward underthrusting Indian slab, we agree that the nulls in the southern Tibet are related to the sub-vertical mantle flow, which is supported by the normal-faulting earthquakes there (Fu et al., 2008; Langin et al., 2003). Considering all these results, we think that the different diving depths and even the different extents of the Indian lithospheric mantle are more or less attributed to the strong heterogeneities in the Indian slab, including faults or incomplete fragmentation (Figs. 9a and 9b).

5.4 Complex crustal structures

Above the Moho discontinuity, lower velocities are dominant in southern Tibet (Figs. 6a and 6b), where magnetotelluric imaging revealed zones of low electric resistivity at depths of ~20 to 40 km (Bai et al., 2010). Lower electric resistivity is generally considered to indicate the existence of fluids or partial melts in the mid- to

375 lower crust, supporting [the channel flow model](#) (Unsworth et al., 2005). The absolute
376 P-wave velocity in the lower crust under the Indus-Tsangpo suture is ~ 6.99 km/s,
377 while a low-V zone is visible there because our tomographic image shows velocity
378 perturbations relative to [the average 1-D velocity model](#). In contrast, the velocity in
379 the lower crust under Nepal is normal, with a P-wave velocity of 6.4-6.5 km/s
380 (Pandey et al., 1995). The increase of ~ 0.5 km/s in P-wave velocity could not be
381 simply attributed to higher pressure and lower temperature at these depths, hence
382 partial eclogitization of the lower Indian crust is a plausible [interpretation](#)
383 (Schulte-Pelkum et al., 2005). By considering the extremely high heat [flow of up to](#)
384 ~ 150 mW/m² (Francheteau et al., 1984) and low value of the effective elastic
385 thickness in southern Tibet (Hetenyi et al., 2006; Jordan and Watts, 2005), the low-V
386 zone in the middle crust may reflect a weak and ductile crust. [The existence of low-V](#)
387 [and high-V zones in southern Tibet supports the two-layer anisotropy model](#) (Gao and
388 [Liu, 2009](#)). The studies of earthquake focal mechanisms showed that some of the
389 thrusting events occurred in the mid- to lower crust (de la Torre et al., 2007; Priestley
390 et al., 2008). Because no mantle materials seem to intrude into the lower crust at
391 present, we suspect that faults may also exist in the lower crust, in addition to the
392 upper mantle. Why do the fragmentation and/or faults cluster in the crust and upper
393 mantle beneath the Himalayan orogen? A simple explanation is that the faults are
394 caused by the strong interactions between the Indian and Asian plates and the crust is
395 thickened significantly there (Nabelek et al., 2009; Schulte-Pelkum et al., 2005; Zhao
396 et al., 2010). However, we suggest that the mantle materials have not been completely

intruded into the lower crust because of the relatively low velocity and the sharp Moho discontinuity, as revealed by the receiver-function study (Zhao et al., 2010). The temperature could be elevated in the Tibetan crust even without hot mantle materials because of increased radiogenic heating in the thickened Eurasian crust (Nelson et al., 1996) or/and plastic deformation (Hochstein and Regenauer-Lieb, 1998). Alternative interpretation is that partial melting may take place there due to warmer upper mantle directly contact the eclogitized lower crust if the fragmentation occurs on the Indian slab.

The low P-wave velocity of the southern Tibet implies a relatively weak crust (Unsworth et al., 2005) that permits materials flow. Note that the low-V zones in southern Tibet are not interconnected, in accordance with the results of finite-frequency tomography (Hung et al., 2011) and magnetotelluric imaging (Wei et al., 2010). This feature is consistent with the magnetotelluric observation in southeastern Tibet, which is located ~700 km to the east of our study region (Bai et al., 2010), suggesting that wholesale material flow may not exist in the crust under central-southern Tibet at present. There may be interconnected low-V features across most of the middle crust at depths between 20 and 40 km (Yang et al., 2012), but we cannot resolve this feature due to the limited resolution of our tomography model.

In contrast, the low-V zone in northern Tibet looks continuous. The differences in age and chemical components of the volcanic rocks between northern and southern Tibet indicate diverse heating sources (Chung et al., 2005; Ding et al., 2007; Guan et al., 2012; Searle et al., 2011). Analyses of helium isotopes show that the radiogenic

heat production is only restricted in the crust of southern Tibet, whereas a ^3He anomaly is consistent with a mantle contribution in northern Tibet (Hoke et al., 2000). A study of Sn wave propagation further suggests that the velocity variations or attenuation structure extend deeper beneath northern Tibet than beneath southern Tibet (Barron and Priestley, 2009). Although several models have been proposed for explaining the warmer upper mantle beneath northern Tibet, such as mantle diapir (Wittlinger et al., 1996), strain heating (Kincaid and Silver, 1996), and lithospheric delamination (Molnar et al., 1993), our tomographic results suggest that the downwelling of the Indian lithosphere **may have** caused concomitant asthenospheric upwelling (Tilman et al., 2003). **However, the precise range of the upwelling is not very clear because the tomographic resolution is low beneath northern Tibet (north of $\sim 34^\circ\text{N}$).**

Under the Lhasa terrane, **a prominent high-V zone exists** in the middle to lower crust (Figs. 6 and 7), **which may** reflect a brittle crustal structure beneath central Tibet. **A receiver-function study**, using data from the Hi-CLIMB project, showed clear double Moho beneath the Lhasa Terrane, which is caused by partial eclogitization (Nabelek et al., 2009). Similarly, Schulte-Pelkum et al. (2005) suggested that 30% of the lower crust undergoes eclogitization in southern Tibet. Their model shows that the northern boundary of the high-V zone in the lower crust **is limited to $\sim 31^\circ\text{N}$** (Nabelek et al., 2009), which was interpreted as the partially eclogitized lower crust of India underthrusting beneath southern Tibet. Taking into account the results of receiver function, finite-frequency tomography and Birch's law, Tseng et al. (2009) attributed

the [Moho disruption](#) to mechanical deformation. In the same area, Gaussian-beam migration of scattered teleseismic P-waves imaged more than one big impedance contrast in the zone of disrupted Moho, suggesting that mantle materials may have been intruded into the crust and some materials reside in the mid- to lower crust at present (Nowack et al., 2010). [A shear-wave splitting study suggested that the present-day velocity gradient tensor field within the crust and mantle is correlated \(Holt, 2000\).](#) These pieces of geophysical evidence, [together with](#) a geochemistry result that the Cr and Ni concentrations and Mg# are relatively high in this region (Chen et al., 2012), suggest that the high-V zone is probably a mixed product of mantle and crust materials, not only eclogitization. [Considering](#) the breakoff model and the shallower depth of the high-V zone (Hetenyi et al., 2006), [we prefer](#) that the high-V zone in the crust of the Himalaya orogen and the Lhasa terrane may reflect a [mixed](#) product of cooling mantle materials and lower crust partial eclogitization, which may explain the different mechanisms between the interrupted flow under central-southern Tibet and the wholesale extrusion under northern Tibet.

Interestingly, a high-V zone is visible around the Main Central Thrust in southernmost Tibet (Figs. 6a and 6b). This feature may suggest that the present underground structure is related to an earlier activity (e.g., mantle materials intruded into the crust during the slab breakoff at ~30 Ma, [Maheo et al., 2002](#)), but its details are not clear and should be clarified in future studies.

[5.5 The north-south trending rift zones](#)

Rifts in Tibet were initiated broadly during the Late Miocene and earliest

Pliocene, *i.e.*, 15–5 Ma (Taylor and Yin, 2009; Williams et al., 2001), but a recent study showed a younger age (5 Ma) for the origin of these rifts (Maheo et al., 2007). No matter which estimate is correct, these age estimates are quite different from those of the Neogene ultrapotassic and adakitic magmatism (~26 to 13 Ma) in southern Tibet (Chung et al., 2005), implying that they were originated by different tectonic processes. A prominent north–south trending low-V zone at ~88°E is revealed from the surface down to ~180 km depth, right beneath the Pumqu-Xianza rift (Figs. 6 and 7c), whereas high-V zones are visible under the Lopukangri rift and the Tangra Yum Co rift which are not zones of active faulting (Murphy et al., 2010). The former low-V zone is also visible in finite-frequency tomography (Liang et al., 2011), and its shallower part is consistent with Pn wave tomography (Liang and Song, 2006). The low-V belt also coincides well with a similar feature in southeastern Tibet (Ren and Shen, 2008; Zhang et al., 2011), which was interpreted as an upwelling induced by delaminated lithospheric mantle. The different velocity structures under these rifts may reflect different kinds of the rifts or different episodes in the origin of the rifts, though our tomography only shows the current images of the structures (Fig. 9c).

6. Conclusions

Our tomographic images show the existence of pervasive and continuous low-V zones in northern Tibet, whereas the low-V zone in southern Tibet is obviously interrupted by high-V zones. The low-V zones in southern and northern Tibet are caused by different heating sources. We infer that the low-V zone beneath the Indus-Tsangpo suture may reflect faults or an incomplete fragmentation in the Indian

slab. The high-V zones under southern Tibet may reflect a combination of the eclogitized Indian lower crust and cooling mantle materials. The different underthrusting distances and diving depths of the Indian slab may be caused by the differences in the viscosity contrast of the Indian slab and the surrounding upper mantle under different parts of the Tibetan plateau, as well as the strong structural heterogeneities in the crust and upper mantle caused by the strong deformation under the Himalayan orogen. These factors and processes may have also resulted in the different rifts in southern Tibet.

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Figure Captions

Figure 1. Distribution of portable seismic stations of the ANTILOPE (red triangles) and Hi-CLIMB (blue dots) projects on the Tibetan plateau. The black lines denote the tectonic block boundaries in the region. MBT, the Main Boundary Thrust; MCT, the Main Central Thrust; ITS, the Indus-Tsangpo suture; BNS, the Bangong-Nujiang suture; JRS, the Jinsha River suture; TB, Tarim Basin; LK, Lopu Kangri rift; TYC, Tangre Yum Co rift; PX, Pumuqu Xianza rift. The open circles denote the local and regional earthquakes used in this study. The purple dashed lines show the locations of vertical cross-sections in Fig. 7. The green dashed line shows the location of the interrupted low-velocity zones in southern Tibet. The cyan bold line indicates the northern limit of the descending Indian slab revealed by this study.

Figure 2. (a) Distribution of 1,578 teleseismic events (open circles) used in this study with epicentral distances of 30°-90°. The solid triangle denotes the center of the study

region. (b) Distribution of P-wave rays used in this study in (b) map view at 100 km depth, and along (c) the AA' profile and (d) the CC' profile as shown in (b). Distribution of P-wave rays used in this study in map views at (e) 500 km and (f) 600 km depths, and (g) along the BB' profile as shown in (e). Blue triangles show the seismic stations. Black, white and blue triangles show the seismic stations. The purple dashed lines show the locations of the profiles.

Figure 3. (a) An example showing the original vertical-component seismograms recorded by 15 stations from an earthquake that occurred on October 31, 2005. The number on the right denotes the epicentral distance in degree. The red line shows the theoretical arrival times calculated for the 1-D velocity model as shown in Fig. 4; the green dashed line shows the manually picked arrival times. (b) An example showing the original vertical-component seismograms recorded by 22 stations from a teleseismic event that occurred on August 2, 2006. The number on the right denotes the epicentral distance in degree. (c)(d) Map showing the locations of the event (yellow star), seismic stations (red triangles) and the corresponding rays.

Figure 4. (a) The starting 1-D velocity model used in this study. The dashed lines denote different velocities in the crust from the CRUST1.0 model and the receiver-function results (Nabelek et al., 2009; Schulte-Pelkum et al., 2005; Zhao et al., 2010). See text for details. (b)(c) Configuration of the 3-D grid model adopted for the tomographic inversion. The grid spacing is $1^\circ \times 1^\circ$ beneath the surrounding areas (b) and $0.5^\circ \times 0.5^\circ$ beneath the central part (c).

Figure 5. (a) Trade-off curve for the root-mean-square (RMS) travel-time residuals

versus RMS P-wave velocity perturbations (in %) at the grid nodes. The numbers in the open circles denote the damping parameters adopted for the tomographic inversions. The number in the bold circle denotes the optimal damping parameter for the final 3-D velocity model obtained by this study. (b) Distribution of average relative travel-time residuals of teleseismic events at each station, calculated for the 1-D velocity model as shown Fig. 4. The blue crosses and black circles denote delayed and early arrivals, respectively. The scale is shown at the bottom. The travel-time residuals are derived by subtracting the theoretical arrival times from the observed ones, and the relative residuals are obtained by subtracting the mean residual for each event from the raw residuals at every stations.

Figure 6. Map views of the obtained P-wave tomography. The layer depth is shown atop each map. The red and blue colors denote low and high velocities, respectively. LK, Lopu Kangri rift; TYC, Tangre Yum Co rift; PX, Pumuqu Xianza rift. The velocity perturbation scale is shown at the bottom. The black lines denote the tectonic block boundaries in the region.

Figure 7. Three vertical cross-sections of P-wave tomography along the profiles shown in Figure 1. The red and blue colors denote low and high velocities, respectively. The velocity perturbation scale is shown at the bottom. The black lines denote the Moho discontinuity derived by receiver-function studies (Nabelek et al., 2009; Zhao et al., 2010), and the black dashed line shows the trace of Indian lithospheric mantle (ILM) deduced from Zhao et al. (2010). The open circles denote earthquakes ($M \geq 3$) that occurred within a 50-km width of each profile. The yellow

dashed line shows areas with the double Moho (see text for details). MCT, the Main Central Thrust.

Figure 8. The input model (a) and the inversion result (b) of a synthetic test (see text for details). The dashed contour line shows the target area of this test. The velocity perturbation (in %) scale is shown at the bottom.

Figure 9. Schematic diagrams illustrating the dynamic evolution of central Tibet. The dashed lines show the less certain parts of the tomographic image. Hypocentral locations of the local earthquakes (open circles) are determined by the China Earthquake Network Center; the location of the Asian lithospheric mantle (ALM) is derived from the receiver-function results (Kind et al., 2002; Zhao et al., 2010); the low-V zone in north is referred from a study of Sn waves (Barron and Priestley, 2009) and tomographic result (Tilman et al., 2003). (a) The black dashed line in the Indian lithospheric mantle (ILM) in the south depicts the result from Hung et al. (2011); the double Moho is taken from Nabelek et al. (2011); and the disrupted Moho is taken from Nowack et al. (2010). (b) The clear Moho is from Zhao et al. (2010), and the thin sub-Moho layer is from Barron and Priestley (2009). (c) The relatively stable crust is inferred from Murphy et al. (2010), and the partial eclogitized lower crust is from Schulte-Pelkum et al. (2005). LS, Lhasa block; QT, Qiangtang block; MCT, Main Central Thrust; MHT, Main Himalayan Thrust; ITS, Indus-Tsangpo suture; BNS, Bangong-Nujiang Suture.

Graphical Abstract

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9