Crustal and mantle velocity models of southern Tibet from finite frequency tomography

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[1] Using traveltimes of teleseismic body waves recorded by several temporary local seismic arrays, we carried out finite-frequency tomographic inversions to image the three-dimensional velocity structure beneath southern Tibet to examine the roles of the upper mantle in the formation of the Tibetan Plateau. The results reveal a region of relatively high P and S wave velocity anomalies extending from the uppermost mantle to at least 200 km depth beneath the Higher Himalaya. We interpret this high-velocity anomaly as the underthrusting Indian mantle lithosphere. There is a strong low P and S wave velocity anomaly that extends from the lower crust to at least 200 km depth beneath the Yadong-Gulu rift, suggesting that rifting in southern Tibet is probably a process that involves the entire lithosphere. Intermediate-depth earthquakes in southern Tibet are located at the top of an anomalous feature in the mantle with a low Vp, a high Vs, and a low Vp/Vs ratio. One possible explanation for this unusual velocity anomaly is the ongoing granulite-eclogite transformation. Together with the compressional stress from the collision, eclogitization and the associated negative buoyancy force offer a plausible mechanism that causes the subduction of the Indian mantle lithosphere beneath the Higher Himalaya. Our tomographic model and the observation of north-dipping lineations in the upper mantle suggest that the Indian mantle lithosphere has been broken laterally in the direction perpendicular to the convergence beneath the north-south trending rifts and subducted in a progressive, piecewise and subparallel fashion with the current one beneath the Higher Himalaya.

1. Introduction

[2] The unique role of the Himalayan range and Tibetan Plateau in understanding continent-continent collision has led to numerous geological and geophysical studies in the region, including several temporary seismic networks operated from the southern Himalaya in Nepal to the Lhasa Terrane of southern Tibet in China during the past two decades [e.g., Hirn et al., 1995; Nelson et al., 1996; de la Torre and Sheehan, 2005; Nabelek et al., 2005; Sol et al., 2007; Velasco et al., 2007]. Yet questions remain about the crust and upper mantle structures, particularly variations in the direction perpendicular to the India-Eurasia convergence, and their implications for the history and dynamics of the Himalaya and the plateau.

[3] Since the work of Argand [1924], the nature and extent of the underthrusting of the Indian lithosphere under the Tibetan Plateau has been a topic of debate [e.g., Willett and Beaumont, 1994]. Several studies suggest that the Tibetan Plateau south of the Bangong-Nujiang suture (Figure 1a) is underlain by the Indian lithosphere. This interpretation is based on high Pn and Sn velocities [Ni and Barazangi, 1984], gravity anomalies [Jin et al., 1996], seismic converted phases [Owens and Zandt, 1997], SKS splitting [Fu et al., 2008] and tomographic results [Tilmann et al., 2003; Zhou and Murphy, 2005]. Receiver function studies also reveal a doublet phase in the Tibetan crust, which was interpreted as evidence for the underthrusting Indian lower crust beyond the Indus-Yalu suture (IYS) [Kind et al., 2002; Nabelek et al., 2005; Schulte-Pelkum et al., 2005; Jin et al., 2006; Nabelek et al., 2009]. However, receiver functions [Kosarev et al., 1999] and Helium isotope ratios [Hoke et al., 2000] show evidence that the Indian mantle lithosphere starts subduction near the IYS. A recent study of S-to-P converted phases in the western Tibetan Plateau [Kumar et al., 2006] and body wave tomography [Li et al., 2008] suggest that the geometry of the Indian lithosphere varies from west to east: Underthrusting beneath the Himalayas and the entire plateau in the west and subducting at an angle in the east.

[4] There are two other enigmatic geological features in southern Tibet: The north-south trending rifts [e.g., Armijo et al., 1986] and the intermediate-depth earthquakes [Chen et al., 1981; Chen and Molnar, 1983; Zhu and Helmberger, 1996; Chen and Yang, 2004; Monsalve et al., 2006; Liang

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The rifts are commonly attributed to the east-west extension of southern Tibet, which is roughly perpendicular to the India-Eurasia convergence direction [Larson et al., 1999]. Several mechanisms have been proposed for these rifts, including gravitational collapse following the attainment of the maximum elevation [Molnar and Tapponnier, 1978; Tapponnier et al., 1981], lower crust flow under east-west extension, oblique convergence of Indian subduction [McCaffrey and Nabelek, 1998], lithosphere fragmentation accompanied with the regional boundary condition applied throughout East Asia [Yin, 2000], and deformation facilitated by mantle lithosphere delamination [Ren and Shen, 2008]. Whether these rifts are restricted to the upper crust or involve the entire lithosphere has also been under debate [Masek et al., 1994; Yin, 2000]. Ren and Shen [2008] found a significant low Vp and Vs anomaly extending to 350 km depth beneath the Coma rift, one of the north-south trending rifts near the East Syntaxis, and suggested that the rift could be related to a mantle lithosphere delamination process.

Figure 1. (a) Tectonic map of the Tibetan Plateau with the research area of this study outlined by a box. The directions of plate movement are from Zhang et al. [2004]. (b) Map of our study area showing active faults and seismic stations used in this study. The stations from different sources are denoted in different symbols: Hi-CLIMB as red triangles, HIMNT as blue triangles, Bhutan as blue diamond, INDEPHT II (International Deep Profiling of Tibet and the Himalaya) as blue circles, and LSA as red square.
fault zone has been suggested as a transfer zone linking north-south trending Cenozoic extensional structures in the Lhasa and Qiangtang Terranes [Yin and Harrison, 2000].

[8] Here we adopt the definition of Yin [2006] that the IYS divides the Himalayan orogen as the Higher Himalaya, its north slope as the North Himalaya, and its south slope as the South Himalaya, which in turn consists of the Lower Himalaya in the north and the Sub Himalaya in the south. Our study area traverses the boundary between the Himalayan orogen and the Tibet Plateau. For convenience, southern Tibet is used in this paper to describe the general location of our study region, including the southern Tibetan Plateau and the Himalayas.

3. Data

[9] Data used in this study are from several seismic projects in southern Tibet: Hi-CLIMB, HIMNT, BHUTAN and INDEPTH II. The combined data set includes 80 stations: 32 from Hi-CLIMB, 29 from HIMNT, 1 from the Global Seismic Network, 5 from BHUTAN, and 13 from INDEPTH II (Figure 1b). Among these, 16 of the 32 stations in the HI-CLIMB two-dimensional (2-D) network were operated by Peking University during 2004 and 2005 as a part of the Hi-CLIMB project. The remaining stations in the Hi-CLIMB 2-D array were operated by the Institute of Earth Sciences, Academia Sinica, Taiwan and China Academy of Geological Sciences.

[10] Teleseismic P and S phases were used in our tomo- graphic inversion. We selected events with an epicentral distance between 30°–80° to minimize the interference of long-period waves from the mantle transition zone and core-mantle boundary (Figure 2). The raw data were first converted to displacement by removing instrument response. For S wave the waveforms were rotated to the radial and tangential directions. Then, the waveforms were filtered to two frequency bands: 0.5–2 Hz and 0.1–0.5 Hz for P wave, and 0.1–0.5 Hz and 0.05–0.1 Hz for S wave, using the Butterworth filter with four poles and two passes. We first handpicked the arrival times of the phases then obtained the differential traveltimes by the Multiple Channel Cross Correlation method (MCCC) [VanDecar and Crosson, 1990]. The cross-correlation time window is usually a full wave period containing the largest absolute amplitude of the arrival. The traveltimes used in the inversion are 11137 high-frequency and 4121 low-frequency P waves, and 1850 high-frequency and 1233 low-frequency S waves.

4. Method

[11] In this study, a finite-frequency seismic tomographic method was used to invert for the P and S wave crustal and upper mantle velocity structures beneath southern Tibet. The inversion method was described in details by Hung et al. [2004], Yang and Shen [2006] and Ren and Shen [2008]. Here we present a simplified description of the method.

4.1. Finite Frequency Theory

[12] The Born-Fréchet traveltime sensitivity kernels express the influence of velocity perturbations upon a travel
time shift of finite-frequency waves as [Dahlen et al., 2000]:

$$\delta t = \int \int K(x) \frac{\partial c(x)}{c(x)} d^3x,$$

(1)

where $K$ is the 3-D Fréchet sensitivity kernel for a travel time shift $\delta t$, measured by cross correlation of an observed pulse with its synthetics for a reference earth model. Under the assumption of single scattering, the kernel is expressed by the formulation [Dahlen et al., 2000]:

$$K = \frac{1}{2\pi c} \left( \frac{R}{c_T R} \right) \times \int_0^\infty \int_0^\infty \frac{\omega^2}{\omega^2 + \omega^2} \left| s_{xy}(\omega) \right|^2 d\omega,$$

(2)

where $\Delta T$ represent the difference in travel time between the path with a detour through a scatterer between the source and receiver relative to the corresponding direct path; $R$, $R'$ and $R'''$ are geometrical spreading factors for the unperturbed ray, the forward source-to-scatterer ray and the backward receiver-to-scatterer ray, respectively. The presence of the power spectrum of the synthetic pulse $s_{xy}(\omega)$ shows effectively the frequency dependence of the cross-correlated differential travel times.

For regional teleseismic travel time tomography the relative arrival times of $P$ or $S$ phases at a number of stations are measured by MCCC to constrain spatial variations of underlying mantle velocity perturbations. The finite volume sensitivity of a relative delay between two nearby stations 1 and 2, $\delta t_1 - \delta t_2$, is simply the difference of the individual Fréchet kernels for the traveltime shifts, $\delta t_1$ and $\delta t_2$ [Dahlen et al., 2000; Hung et al., 2004],

$$K_{\delta t_1} - K_{\delta t_2} = K_{\delta t_1} - K_{\delta t_2}.$$

(3)

4.2. Model Parameterization and Inversion

The crustal and mantle volume beneath the study area was parameterized with regular 3-D grids of $33 \times 33 \times 33$ centered at $(88.5^\circ E, 28.5^\circ N)$ with total dimensions of $12^\circ$ in longitude, $12^\circ$ in latitude, and $1200$ km in depth (Figure 3). That is, the grid spacing is $\sim 0.37^\circ$ in longitude and latitude and $\sim 37$ km in depth. With this parameterization, the traveltime equations can be written as:

$$d_i = G_{it}m_i,$$

(4)

where $d_i$ is the $i$th travel time data, and $G_{it}$ is the differential value of the integrated volumetric kernels of the $i$th event contributing to the $i$th node and $m_i$, the model parameter at the $i$th node. The inversion problem was resolved by the standard damped least square method [Paige and Saunders, 1982]:

$$\tilde{m} = (G^T G + \theta^2 I)^{-1} G^T d,$$

(5)

where $I$ is the identity matrix. The damping parameter $\theta$ is determined empirically through a space of variance reduction and the model norm represented by a trade-off curve (Figure 4). We chose the damping parameter that yields an optimum variance reduction and a relatively small model norm. The model discussed in this paper is obtained using a damping parameter that yields a variance reduction of $\sim 76\%$ for $P$ wave and $\sim 57\%$ for $S$ wave (shown as a filled square and filled circle, respectively, in Figure 4). The lower variance reduction of the $S$ velocity model reflects relatively noisy $S$ waves and $S$ traveltime measurements.

4.3. Crustal and Elevation Correction

The traveltime anomalies can be caused by crustal thickness variations, station elevations and lateral velocity...
heterogeneities. Since the ray paths of teleseismic body waves are nearly vertical and do not cross at shallow depth, shallow velocity structures are poorly constrained by relative traveltime data in the inversion. Therefore, a time correction for crustal structure is needed in order to reduce the tradeoff between crustal and mantle velocity heterogeneities in seismic tomography. To correct for the crustal effects, the frequency dependent crustal correction for each event station record was calculated from the synthetic crustal response of an incoming plane wave beneath the station [Yang and Shen, 2006]. We used the crustal structure from the CRUST2.0 [Bassin et al., 2000] for stations in Sub Himalaya and SEAPS [Sun et al., 2004; Sun et al., 2008] for stations elsewhere. An additional free term for each station was incorporated into the inversion to absorb travelt ime shifts caused by remaining shallow heterogeneity.

4.4. Vp/Vs Ratio Perturbation

Because the relative travel time of one phase recorded at two stations is used to constrain the underlying mantle velocity perturbation, the inversion solves spatial distribution of relative velocity perturbation but does not provide constraints on the absolute velocity perturbation. Given a 1-D reference model for the study area, the perturbation of Vp/Vs ratio ($\delta V_p$) can be retrieved from Vp and Vs perturbations in percentage $\delta V_p$ and $\delta V_s$, respectively,

$$\delta V_p = \frac{V_{p_{\text{inverted}}} - V_{p_{\text{reference}}}}{V_{p_{\text{reference}}}} \left( \frac{1 + \delta V_p}{1 + \delta V_s} \right) \frac{V_p}{V_s} = \frac{\delta V_p}{1 + \delta V_s}.$$  

$$\delta V_s = \frac{V_{s_{\text{inverted}}} - V_{s_{\text{reference}}}}{V_{s_{\text{reference}}}} \left( \frac{1 + \delta V_p}{1 + \delta V_s} \right) \frac{V_p}{V_s} = \frac{\delta V_s}{1 + \delta V_p}.$$  

Several different approaches have been used to calculate Vp/Vs tomographic models. With good ray coverage for both $P$ and $S$ waves, the Vp/Vs could be calculated directly from separate Vp and Vs models [Nakajima et al., 2001; Nakamura et al., 2003]. When either the $S$ wave data coverage is not good or there is significant difference in the data coverage for $P$ and $S$ waves, mapping Vp/Vs using separate Vp and Vs solutions is inappropriate [Eberhart-Phillips, 1990]. A simultaneous inversion for Vp and Vp/Vs ratio without the assumption of identical $P$ and $S$ coverage should be used [Conder and Wiens, 2006].

In our case with a significant difference in $P$ and $S$ data coverage, artificial features in the Vp/Vs models could be present when structures are resolved only in the $P$ or $S$ inversion [Kennett et al., 1998; Saltzer et al., 2004]. To minimize artificial features in Vp/Vs ratio introduced from independent Vp or Vs inversions, we selected $P$ and $S$ travel times with the objective of finding a subset of the $P$ and $S$ data that have the same data coverage. This results in the use of 1179 low-frequency $P$ phase travel times and the same number of high-frequency $S$ phase relative travel times that share nearly identical source-receiver paths. Consequently, the Vp and Vs perturbations used for the calculation of Vp/Vs were inverted with the same inversion parameters, since the selected $P$ and $S$ travel times have a very similar sensitivity kernel distribution.

The difference in Vp/Vs ratio among different 1-D reference velocity models is small compared to the range of values observed in the study area in the lower crust and upper mantle depths. For example, the maximum difference of Vp/Vs ratio between IASP91 and AK135 is 0.2% from 35 km to 660 km depth.

4.5. Resolution Tests

Horizontal and vertical resolution tests were performed to evaluate the data coverage and the ability of the inversion to recover the mantle structure (Figure 5). For this purpose, the synthetic travel times were computed by multiplying the G matrix with different input velocity models: $\Delta t = G \cdot \Delta v + \text{noise}$, where $\text{noise}$ is random noise added to the synthetic traveltimes with a Gaussian distribution and the standard deviation of $\text{noise}$ is set to 0.04 second and 0.1 second for $P$ and $S$ respectively. We calculated the correlation coefficients and the differences of the input and output models at various depths. With the criteria of the correlation coefficient higher than 85% and the input-output difference less than 25% of the input anomaly, we estimated that the recovered structure meets the criteria.
Figure 5
down to ~350 km depth for P wave models and ~200 km depth for S wave models. The resolution also worsens in shallow depth (~<50 km), particularly for S waves. The horizontal scales of the interpreted features in our tomographic models are about 150 km or larger.

[20] In the second set of tests, for both horizontal and vertical resolution, the input velocity perturbation alternates in sign both horizontally and vertically (Figure 5). Three-dimensional correlation coefficients were calculated for these tests. Within the 33 × 33 × 33 inversion grid, a moving cube with 5 × 5 × 5 grids in both the input and output models was selected and interpolated, then cross-correlation coefficient was calculated between these two cubes as the correlation coefficient for the central grid of the cube. In the center of the models, the 3-D correlation coefficient is greater than 85% to 300 km depth for Vp and to 200 km depth for Vs. Thus in the following discussion we interpret the P wave model in the upper 300 km depth and the S wave model in the upper 200 km depth.

[21] For the Vp/Vs resolution (Figure 5), separate Vp and Vs inversions are carried out using 1179 P and 1179 S traveltimes that have nearly identical ray paths.

[22] We also carried out tests with different levels of \( \sigma_{\text{noise}} \) and found that the results are quite similar for the noise level smaller than 0.3 second.

### 4.6. Null-Space Shuttle Method

[23] We used the null-space shuttle method [Deal and Nolet, 1996; Muñoz and Rath, 2006] to test the robustness of the low Vp/Vs anomaly in our final result. The null-space shuttle method is an approach to determine the permissible changes for a tomographic image while retaining the fit to the data [Deal and Nolet, 1996; Muñoz and Rath, 2006]. For our inversion problem:

\[
d = Gm. \tag{7}
\]

the original inverted model \( m \) can be modified to get a filtered solution \( m_f \). The difference between them is \( \Delta m = m_f - m \). With the requirement that \( \Delta m = \Delta m_{\text{range}} - \Delta m_{\text{null}} \) and the assumption of and \( G\Delta m_{\text{null}} = 0 \), where \( \Delta m_{\text{range}} \) is the model space with good data coverage and \( \Delta m_{\text{null}} \) is the portion of the model space in the null-space, the linear inversion problem could be expressed as,

\[
G\Delta m = G\Delta m_{\text{range}} = \Delta d. \tag{8}
\]

Where the \( \Delta d \) is the difference between the original data \( d \) and the data corresponding to the filtered model \( m_f \), which are the traveltimes residuals only related with the model difference \( \Delta m_{\text{range}} \) with good data coverage. This equation could be solved just like equation (7) by the standard damped least square method, and the final conservatively filtered solution is

\[
m_{\text{consrv}} = m_f - \Delta m_{\text{range}}. \tag{9}
\]

### 5. Results

[24] The resulting three-dimensional P and S wave velocity models from all the usable P and S traveltimes and the Vp/Vs ratio from the selected P and S data that share nearly identical paths are presented in Figures 6a and 6b. Because we used differential traveltimes in this teleseismic tomography, the absolute velocity perturbation could not be constrained. The fast or slow velocity anomalies discussed here are all relative anomalies.

[25] The most prominent feature in these 3-D velocity models is a high Vp and Vs anomaly along the strike of Higher Himalaya that extends from the surface to about 180 km depth. This high-velocity anomaly is attributed to the Indian mantle lithosphere. To the north of this high-velocity anomaly there are strong low-velocity anomalies at about 75 km depth beneath the North Himalaya, which likely reflect the velocity difference between the thickened Tibetan crust and the cold Indian mantle lithosphere.

[26] In the uppermost mantle, there is a strong low Vp and Vs anomaly with a small positive Vp/Vs ratio perturbation extending to ~200 km depth (Figures 6a, 6b, and 7) along the YGR. A less pronounced low-velocity anomaly is present beneath the TYR. These are similar to the low-velocity anomaly along the Coma rift east of our study area close to the Eastern Syntaxis [Ren and Shen, 2008].

[27] At the depth of about 75 km, there is a low Vp/Vs ratio in the region between the MFT and the southern ends of the PXR and TYR (Figures 6a, 6b, and 9). This low Vp/Vs region coincides with a cluster of the hypocenters of the intermediate-depth earthquakes in southern Tibet [Monsalve et al., 2006; Liang et al., 2008]. We used the null-space shuttle method [Deal and Nolet, 1996; Muñoz and Rath, 2006] to test the robustness of this low Vp/Vs anomaly. In the test the Vp/Vs perturbation result is obtained from the selected P and S data set with nearly identical paths. The best-fitting inverted models were filtered to remove the low Vp/Vs anomaly at the south end of the PXR (shown in the second column in Figure 9). Then the null-space shuttle method was applied to the filtered models and the conservatively filtered models were retrieved. The Vp/Vs ratio anomaly near the center of the model was recovered, suggesting that the low Vp/Vs anomaly at the southern end of the PXR is not part of the null-space but is indeed required by the data.

### 6. Discussion

#### 6.1. Influence From Crust Structures

[28] We use teleseismic traveltimes to invert the regional structures in the crust and upper mantle, therefore the steep incident angle at the shallow depth limits the resolution of the upper and middle crust heterogeneity. Differential traveltimes used here are sensitive only to the relative variation among the stations. To correct for the crustal heterogeneities, the crustal correction for each event station record was calculated based on the crust model for the station. We find that the main features of the inversion results are similar using several different crust models: SEAPS [Sun et al., 2004; Sun et al., 2008], CRUST2.0 [Bassin et al., 2000] and CUB2.0 [Shapiro and Ritzwoller, 2002].

[29] Huang et al. [2009] performed a joint inversion of local and teleseismic traveltimes using the data of the HIMNT, INDEPTH II and BHUTAN projects. They found a widespread middle crust low-velocity zone (at 30–50 km depth) beneath the North Himalaya (Tethyan Himalaya), which is not present in all of the crust models we tested for.
crustal correction. This midcrust low-velocity layer as well as a thicker crust could be partially responsible for the low Vp velocity anomaly at 75 km depth, although our relative travel times are not sensitive to horizontal structures.

6.2. Rift and Nonrift Structures

[30] Five 2-D profiles along the direction of the convergence between the Indian and Eurasian plates are presented in Figure 6b to show the differences among these rifts and between rift and nonrift profiles. Three of them are along the rifts in southern Tibet and two (nonrift) profiles in between. Significant differences exist among these three rifts: a pronounced low Vp and Vs anomaly beneath the YGR, a reduced low Vp and Vs anomaly beneath the PXR, and little anomaly beneath the TYR. These differences reflect the lateral variation along the strike of Himalaya. Especially, the strong low Vp and Vs anomaly beneath the YGR, which we discuss in detail later.

[31] All the five profiles show clearly a low Vp anomaly (with a low Vp/Vs ratio, except for the DD' profile) at the
Figure 6b
Figure 6b. Vertical slices along the plate convergence direction. The Indian lower crust (Main Himalaya thrust (MHT)) and Moho [Schulte-Pelkum et al., 2005] are shown as a dotted line and a dashed line on BB’ and CC’, respectively. Geological structures are the same as the ones in horizontal slices. The dot-dash lines enclose the regions with correlation coefficient greater than 85%. The topography is exaggerated by 10 times. The black circles show the seismicity within 70 km of the profiles and projected on the profiles.

Figure 7. Vertical slices are perpendicular to the Yadong-Gulu rift from south to north, showing the low Vp, Vs, and high Vp/Vs ratio perturbations extending to at least 200 km depth.
Moho depth (75 km) beneath the North Himalaya which is also the location of the hypocenters of the intermediate-depth earthquakes (Figure 6b) [Monsalve et al., 2006; Liang et al., 2008]. These earthquakes near the Moho have been suggested to be associated with the eclogitization of the subducting Indian lower crust [Jackson et al., 2004] and if this is true, a positive Vp anomaly should be observed [Monsalve et al., 2008]. This inconsistency is discussed later.

6.3. The Yadong-Gulu Rift

[32] The most notable rift among these roughly north-south trending rifts in southern Tibet is the YGR, which extends southward to 27.5°N. Previous studies have shown that the central segment of the YGR initiated at about 8 Ma ago and has accommodated >20 km of east-west extension [Harrison et al., 1995] while recent analysis of U-Pb, Ar/Ar, (U-Th)/He has suggested a younger age of 5 Ma for the onset of these rifts [Maheo et al., 2007]. Both age estimates are different from the estimate of the Neogene ultrapotassic and adakitic magmatism (~26 to 13 Ma) in southern Tibet [Chung et al., 2005]. The significant age difference between the onset of the rifting and the Neogene magmatism implies that they were associated with different tectonic processes or at least different stages of the tectonic process.

[33] Figure 7 shows three profiles of the resulting Vp, Vs, and Vp/Vs ratio across the rift from south to north. A strong low-velocity anomaly in both Vp and Vs is observed beneath the rift from Moho down to at least 200 km depth. A corresponding positive anomaly in Vp/Vs ratio is centered at 100 km depth on profile BB′ of Figure 7, indicating a likely thermal origin for this feature because of a higher sensitivity of Vs to a temperature increase at these depths. The resolution test with a low Vp and Vs anomaly in the crust beneath the YGR shows that an input velocity anomaly in the crust is not smeared down significantly to the upper mantle in our inversion (Figure 8). Combined with the vertical resolution
test (Figure 5), it is clear that the velocity anomalies in the mantle beneath the YGR are not due to up-down smearing of the crust structure.

[34] Together with the GPS measurements [Larson et al., 1999; Chen et al., 2004] and focal mechanisms of intermediate-depth earthquakes [Chen and Yang, 2004; de la Torre and Sheehan, 2005], these observations indicate that the rift is associated with processes in the mantle. Results from Pn tomography also showed the presence of a low-velocity belt in the uppermost mantle beneath the YGR [Hearn et al., 2004; Liang and Song, 2006; Pei et al., 2007].

[35] A similar low-velocity anomaly in Vp and Vs was also observed at the Coma rift to the east of the YGR [Ren and Shen, 2008]. Beneath the low-velocity zone of the rift, there is a tabular, high-dipping angle, high-velocity, but low Vp/Vs ratio anomaly, which indicates a highly melt-depleted mantle. Ren and Shen [2008] interpreted that high-velocity, low Vp/Vs anomaly as an evidence supporting the delamination of the mantle lithosphere beneath southern Tibet, a process in which the mantle lithosphere peels away from crust and induces a rapid lithosphere thinning. Beneath the low-velocity anomaly of the YGR, we cannot resolve any such anomaly and discriminate the mantle processes that may have caused or facilitated rifting. Nevertheless the observation of low-velocity anomalies beneath the YGR reconfirms that the north-south trending rifts in southern Tibet could be related to a coherent deformation of the crust and upper mantle.

6.4. Eclogitization of the Indian Lower Crust

[36] Le Pichon et al. [1992] pointed out that the volume of the present-day Indian and Eurasian crust cannot balance the total estimated amount of the crust that has entered into the collision zone and the eastern mass extrusion accounts for no more than one third of the deficit. One possible explanation for the missing crust is that a significant amount of the Indian lower crust has been transferred into the mantle by eclogitization, a process that transforms granulite to eclogite. Thermal-kinematic and petrological models suggest that eclogitization could take place at the P-T conditions in the lowermost crust of the Higher Himalaya [Henry et al., 1997]. Incorporating geophysical constraints with thermal-kinematic and petrological modeling, Hetényi et al. [2007] found that the major density jump in the Indian lower crust occurs when it reaches ~70 km depth beneath the Higher Himalaya. Receiver functions reveal a low-velocity contrast across the Moho north of the Higher Himalaya, reflecting a higher velocity in the lower crust, which could be an indication for the presence of partial eclogitization [Schulte-Pelkum et al., 2005].

[37] Evidence for the transfer of eclogitized crust into the mantle has been more elusive. Observations of a high velocity beneath the Moho from a wide-angle reflection line situated between the Higher Himalaya and the IYS were interpreted as evidence for eclogite just beneath the seismic Moho [Spain and Hirn, 1997]. Using P and S wave traveltimes from local earthquakes recorded by the HIMNT network, Monsalve et al. [2008] found a significantly high Vp velocity (over 8.4 km/s) in the uppermost mantle north of 27.5°N. Together with a relatively low Vp/Vs ratio compared to that of an average uppermost mantle, the high Vp velocity was interpreted as evidence for the presence of eclogite under the Moho [Monsalve et al., 2008]. In contrast to this high Vp in the uppermost mantle, our results show a localized, low Vp, normal-to-high Vs, and low Vp/Vs anomaly that extends at the general vicinity of the intermediate-depth earthquakes beneath the North Himalaya (Figures 6a and 6b). The resolution (Figure 5) and null-space shuttle (Figure 9) tests show that this anomaly is a robust feature required by the traveltime data. One possible explanation for the discrepancy between Monsalve et al. [2008] and this study is that the model of Monsalve et al. [2008] reflects a 2-D structure averaged in the direction perpendicular to the plate convergence, while the 3-D teleseismic tomography using relative traveltimes is sensitive to lateral velocity heterogeneities but insensitive to a constant velocity change at the same depth.

[38] A low temperature, a high Magnesium number (Mg number, Mg/(Mg + Fe)), or a low garnet/olivine ratio could cause a low Vp/Vs ratio in the uppermost mantle [Boyd et al., 2004]. Orthopyroxene-rich zones within the peridotitic lherzolitic mantle also show as a low P wave velocity, with only a small decrease in the S wave velocity and a low Vp/Vs ratio [Miller and Lee, 2008]. But none of these mechanisms could explain both the low Vp and high Vs of this anomalous feature around the intermediate-depth earthquakes (Figures 6a, 6b, and 9).

[39] Jackson et al. [2004] attributed the intermediate-depth earthquakes in southern Tibet to the transformation process of the lower crust from granulite to eclogite. Studies from two high-pressure complexes in Norway [Lund et al., 2004] found that deeply subducted dry rocks in the lowermost continental crust may resist metamorphic re-equilibration for geologically significant periods of time. Their observations indicate that the brittle seismogenic state of some lower crustal rocks is largely controlled by their very low volatile contents, which keep the rocks metamorphically metastable despite at mantle conditions of T = 600–800°C and P = 1.5–2.0 GPa. Dry rocks under high stress may fail by high-temperature and high-pressure faulting if crystal plasticity is unable to accommodate the imposed strain [Shelton et al., 1981]. The occurrence of intermediate-depth earthquakes in the Himalayas could represent the presence of metastable dry granulite within the lower continental crust and its ongoing transformation to eclogite.

[40] Eclogite, which is composed predominantly of omphacitic pyroxene and Ca–Fe–Mg-rich garnet, generally has higher Vp and even higher Vs in percentage compared to those of the peridotitic uppermost mantle [Ji et al., 2002]. However, in a rock with coexisting high- and low-pressure phases, such as the granulite-eclogite or garnet granulite-eclogite system, the effective bulk modulus could be significantly lowered if the pressure of seismic waves drives the volume-reducing phase transformation [Anderson, 1989]. Using experimental data, Li and Weidner [2008] demonstrated the softening of the bulk modulus within the two-phase loop of olivine-ringwoodite on time scales of 10 to 1000 s, and suggested that seismic waves with periods between 1 and 1000 s could at least partially drive the phase transition and result in partially relaxed P velocities.

[41] The effects of seismic waves on the effective bulk modulus of a rock with coexisting (garnet) granulite and eclogite are likely controlled by diffusion between the minerals in metastable (garnet) granulite and eclogite, though the exact effects are unknown. Nevertheless, we may obtain a rough estimate of the characteristic time of the transformation
using the iron-magnesium exchange rate for garnet. Following Li and Weidner [2008] the characteristic transformation time ($\tau_1$ in the work of Li and Weidner [2008]) is $\sim 10^{-2}$ s if we assume a grain size of 1 mm, a pressure perturbation caused by seismic waves of $10^{-7}$ GPa, and a width of the pressure of the transformation of $\sim 1$ GPa, and a diffusion coefficient of $10^{-19}$ m$^2$ s$^{-1}$ at 800°C [Freer and Edwards, 1999]. A larger grain size, smaller width of transformation pressure, or lower temperature increases $\tau_1$, while the opposite reduces $\tau_1$. Although $\tau_1$ is quite variable and has large uncertainties, this exercise shows that the softening of the bulk modulus in the (garnet) granulite-eclogite system and thus a reduction in $P$ wave velocity is possible within the frequency band of teleseismic $P$ waves used in this study (0.1–2 Hz). On the other hand, high-frequency $P$ waves used in local earthquake tomography [Monsalve et al., 2008] may have a significantly less softening of the bulk modulus. Using local and teleseismic $P$ waves, Huang et al. [2009] found a high $V_p$ immediately beneath the Moho (76–83 km) in the vicinity of the intermediate-depth earthquakes (28°N, 88°E), but a low $V_p$ anomaly at depths of 99 and 119 km. The crust and uppermost mantle are presumably constrained primarily by local, high-frequency $P$ and $Pn$ waves, while the deeper structure is constrained by relatively lower frequency, teleseismic $P$ waves in Huang et al. [2009]. So the frequency-dependent wave speeds of the granulite-eclogite transformation also offer a possible explanation for the discrepancy between Huang et al. [2009] and this study. The volume-reducing transformation has little effect on the shear modulus and shear velocity. The combined effects of the softening of the bulk modulus in a (garnet) granulite-eclogite system and a relative low temperature may account for the simultaneous occurrence of low $V_p$, high $V_s$ and low $V_p/V_s$ anomalies.

In this scenario, the intermediate-depth earthquakes and the underlying low $V_p/V_s$ anomaly are both indications of the lower crustal material being transferred into the mantle and an ongoing granulite-eclogite transformation. The mechanisms of the transfer can be passive by the coupling of part of the lower crust to the subducting mantle lithosphere or dynamic because of the negative buoyancy of an eclogitized
Geodynamic interpretations superimposed profiles of B02408 Moho geometry. This low Vp/Vs anomaly is possibly controlled by the boundaries of the broken mantle the plateau building and the surface rifts correspond to and are into pieces by the rifting processes started in the late stage of Shen south Tibet imaged in this study and to the east [2009]. The presence of 200 km deep low velocity anomalies beneath the north the Tibetan Plateau. The presence of 200 km deep low velocity anomalies beneath the north the Tibetan Plateau. Figure 6. The Indian lithosphere underthrusts to the Higher Himalaya and the Indian crust feeds into the Tibetan crust. Sinking of the eclogitized lower crust because of negative buoyancy may be partially responsible for the separation of the Indian mantle lithosphere from the seismic Moho farther north. Our tomographic model supports the notion that the Indian lithosphere subducts in a progressive, piecewise and subparallel fashion shown as different blue colors as different episodes of subduction. The model cartoon is modified from Nabelek et al. [2009]. The eclogitized lower crust is shown in green. Moho is shown as a black line. Focal mechanisms mark mantle earthquakes [Chen and Yang, 2004]. The MHT, which becomes a broader midcrustal LVZ (blue wavy pattern) beneath the North Himalaya, accommodates the simple shear of the plate motions and acts as a conduit for the transfer of the Indian upper crust into the Himalayan orogenic prism, as the middle crust low velocity zone imaged by Huang et al. [2009]. The prominent lineations of the upper mantle fabric are shown in gray lines.

Figure 10. Geodynamic interpretations superimposed on the Vp perturbation images along the CC' profiles of Figure 6. The Indian lithosphere underthrusts to the Higher Himalaya and the Indian crust feeds into the Tibetan crust. Sinking of the eclogitized lower crust because of negative buoyancy may be partially responsible for the separation of the Indian mantle lithosphere from the seismic Moho farther north. Our tomographic model supports the notion that the Indian lithosphere subducts in a progressive, piecewise and subparallel fashion shown as different blue colors as different episodes of subduction. The model cartoon is modified from Nabelek et al. [2009]. The eclogitized lower crust is shown in green. Moho is shown as a black line. Focal mechanisms mark mantle earthquakes [Chen and Yang, 2004]. The MHT, which becomes a broader midcrustal LVZ (blue wavy pattern) beneath the North Himalaya, accommodates the simple shear of the plate motions and acts as a conduit for the transfer of the Indian upper crust into the Himalayan orogenic prism, as the middle crust low velocity zone imaged by Huang et al. [2009]. The prominent lineations of the upper mantle fabric are shown in gray lines.

crust depending on the extent of eclogitization and the effective bulk viscosity of the eclogitized crust.

6.5. The Indian Mantle Lithosphere

Whether the Indian mantle lithosphere underthrusts subhorizontally and continuously to the Bangong-Nujiang suture or subducts at an angle [Tilmann et al., 2003; Li et al., 2008] is an important question to understand the evolution of the Tibetan Plateau. The presence of 200 km deep low-velocity anomalies beneath the north-south-trending rifts in south Tibet imaged in this study and to the east [Ren and Shen, 2008] suggests that an underthrusted Indian mantle lithosphere, if present before rifting, would have been broken into pieces by the rifting processes started in the late stage of the plateau building and the surface rifts correspond to and are possibly controlled by the boundaries of the broken mantle

lithosphere [Yin, 2000]. Local asthenospheric upwelling could drive the rifting process at the surface and is observed as the negative velocity anomalies in both Vp and Vs shown in Figure 7. Individual pieces of the Indian mantle lithosphere could have undergone separate evolution since, driven by local forces acting on the individual pieces and resulting in differences in the geometry of the Indian mantle lithosphere from west to east Tibet. Alternatively rifting involves only the overriding Eurasian plate. In this scenario, the northern extend of subhorizontal underthrusting of the Indian lithospheric mantle is limited to the North Himalaya in the study area and the Indian lithospheric mantle that entered the collision zone earlier has either been peeled off from the Moho of the Tibetan crust or has subducted into deeper mantle [Li et al., 2008].

If the low Vp/Vs region indeed reflects eclogitization of crustal material as we suggest, it is inconsistent with a continuous underthrusting of the Indian mantle lithosphere right beneath the seismic Moho, because the low Vp/Vs region extends below the seismic Moho. It requires the separation of the Indian mantle lithosphere from the lower crust beneath the Higher Himalaya. Together with the compressional stress from the collision, eclogitization and the associated negative buoyancy force offer a plausible mechanism that causes the sinking and subduction of the Indian mantle lithosphere beneath the Higher Himalaya. Since our observation is a snapshot of a limited extent, it does not exclude the possibility of the existence of an underthrusted Indian mantle lithosphere right beneath the seismic Moho in north Tibet.

Also inconsistent with a simple, continuous underthrusting Indian mantle lithosphere right beneath the seismic Moho are the north dipping lineations with about a 20° angle in the upper mantle from seismic reflection data [Alsdorf et al., 1996], receiver function images along the main profile of the Hi-CLIMB experiment [Nabelek et al., 2009] and local and teleseismic joint tomography [Huang et al., 2009]. These mantle fabrics cannot be explained with simple underthrusting. Furthermore they suggest that in the past 20 to 25 million years, the mantle has not been subducting along a single well-established interface but rather along distributed, evolving subparallel structures, which imply a degree of decoupling between the crust and the mantle [Nabelek et al., 2009]. Taking these observations into consideration, our preferred scenario is that the Indian mantle lithosphere has been broken laterally in the direction perpendicular to the convergence and has subducted in a progressive, piecewise and subparallel fashion with the current one beneath the Higher Himalaya (Figure 10).

7. Conclusion

Our finite-frequency body wave tomography models show that there is a low Vp/Vs anomaly below a cluster of intermediate–depth earthquakes and the corner of the “ramp-and-flat” Moho geometry. This low Vp/Vs anomaly is unusual as it is associated with a low Vp but normal to high Vs velocities. This anomaly cannot be explained by a simple thermal or compositional (e.g., volatiles) heterogeneity. We attribute it to pressure softening because of the coexistence of eclogite and (garnet) granulate in a partially transformed, former Indian lower crust. If true, the partial removal and
sinking of an eclogitized lower crust provides a possible explanation for the deficit in the mass balance of the short-ened and thickened crust of the Himalayas and Tibetan Plateau.

[47] The low Vp and Vs anomalies extending to more than 200 km depth beneath the YGR and possibly TYYR are evidence that deformation associated with the rifts cuts through the whole crust and mantle lithosphere. This supports the previous observation by Ren and Shen [2008] that the north-south trending rifts in southern Tibet are lithospheric-scale features. Significant variations in velocity structures do exist between the profiles along the rifts and those between the rifts.

[48] There is a high Vp and Vs anomaly that extends from the Moho to about 200 km depth along the Higher Himalaya. This high-velocity anomaly is attributed to the Indian mantle lithosphere. Together with evidence for north dipping lineations in the mantle [Nabelek et al., 2009; Huang et al., 2009], our observations suggest that the Indian mantle lithosphere has been broken laterally in the direction perpendicular to the convergence beneath the North Himalaya and subducted in a progressive, piecewise and subparallel fashion with the current one beneath the Higher Himalaya (Figure 10).

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B02408 LIANG ET AL.: VELOCITY MODELS OF SOUTHERN TIBET B02408
The page contains a list of scientific articles and references related to geophysical studies, particularly focusing on the geological and tectonic activities in southern Tibet. The references span various journals from 1994 to 2008, covering topics such as seismicity, crustal structure, and tectonic deformation. These studies provide insights into the geological processes occurring beneath northeastern Japan and the implications for models of Tibetan Plateau evolution.

Key articles mentioned include:

The references indicate a comprehensive understanding of the tectonic processes in southern Tibet, with a focus on seismic activity and crustal structure.


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