Finite frequency tomography in southeastern Tibet: Evidence for the causal relationship between mantle lithosphere delamination and the north–south trending rifts

Yong Ren and Yang Shen

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[1] While several mechanisms have been suggested to explain the evolution of the Tibetan Plateau, observational constraints on the deep lithospheric processes have been sparse, and previous seismic studies were mostly along profiles perpendicular to the collision front of the Indian and Eurasian plates. In this study, we show tomographic evidence for the delamination of the mantle lithosphere beneath southeastern Tibet, a process in which the entire mantle lithosphere peels away from the crust along the Moho and thus is a mechanism for rapid thinning of the lithosphere. Our P and S wave velocity models show the presence of a low-velocity anomaly in the crust and upper mantle down to ~300 km depth beneath a north–south trending rift zone in southeastern Tibet. This low-velocity anomaly is situated above a tabular, high-dipping-angle, high-velocity anomaly that extends into the upper mantle transition zone. The $V_P/V_S$ ratio of this high-velocity anomaly suggests that temperature variations are not the only cause of the anomaly and a highly melt-depleted mantle is required. These observations suggest a causal relationship between the delamination of mantle lithosphere and the formation of the north–south trending rift in southeastern Tibet.


1. Introduction

[2] Since the collision of India and Eurasia about 50 Ma ago, at least 2000 km of convergence has been accommodated by thickening the crust and elevating the Himalayan-Tibetan Plateau [Yin and Harrison, 2000]. The region has been strongly deformed through a combination of thrust, extension, and strike-slip faulting. Among the various styles of deformation, north–south and northeast–southwest trending rift zones, which indicate generally east–west extension, are enigmatic features. They are distributed mainly in the southern and central Tibetan Plateau [Molnar and Tapponnier, 1978; Armijo et al., 1986, 1989; Rothery and Drury, 1984], where topography reaches high altitudes. Field observations indicate that the magnitude of this extension is small (<1%) [Armijo et al., 1986] and rifting in southern and central Tibet started about 13.5–14 Ma ago [Yin et al., 1994; Coleman and Hodges, 1995; Harrison et al., 1995; Blisniuk et al., 2001].

[3] Whether the north–south trending rifts are shallow crustal features or a surface manifestation of a coherent deformation throughout the entire crust and mantle lithosphere is a key question in the continuing debate about the mechanisms that form the Tibetan Plateau [e.g., Holt, 2000; Yin, 2000]. The late Cenozoic east–west extension of the plateau is commonly attributed to gravitational collapse of the thickened crust [Mercier et al., 1987; Dewey, 1988; England and Houseman, 1989; Liu and Yang, 2003], the consequence of a sharp increase in potential energy after an abrupt rise of the plateau due to convective removal of the lower mantle lithosphere [England and Houseman, 1989]. Numerical modeling indicates that the present high surface elevation and small regional topographic slope within the plateau require the removal of the underlying mantle lithosphere [Jiménez-Munt and Platt, 2006]. However, the time scale involved in the convective removal of the mantle lithosphere is uncertain because the viscosity is not well known. Numerical simulations by different authors show that the duration of the removal can vary drastically from 10 Ma to a few hundreds of million years, depending on the rheological parameters in the simulation experiments [Davaille and Jaupart, 1993; Conrad, 2000; Houseman et al., 2000; Morency et al., 2002]. The roughly concurrent onsets of tectonic processes in the surrounding regions have led to the suggestion of a rapid removal of the mantle lithosphere beneath Tibet at about 8 Ma ago [Harrison et al., 1992; Molnar et al., 1993].

[4] Observations that support the notion of a vertically coherent deformation of the lithosphere [Tapponnier et al.,
England and Houseman, 1986] include the studies of earthquake focal mechanisms which show brittle east–west extension in the mantle directly beneath the north–south trending rifts [Chen and Kao, 1996; Chen and Yang, 2004] and the instability analysis of the spacing between the rifts [Yin, 2000]. On the other hand, the presence of a low-viscosity mid and lower crust may decouple deformation in the shallow crust from that in the mantle lithosphere [Zhao and Morgan, 1987; Bird, 1991; Royden, 1996]. Other explanations for the east–west extension in the Tibetan Plateau include the combination of the eastward extrusion of Tibet and strike-slip faulting along the Himalayan arc caused by oblique convergence between India and Eurasia [Tapponnier et al., 1981; Armijo et al., 1986; McCaffery and Nabelek, 1998; Tapponnier et al., 2001], and the collisional stresses localized along the southern part of the Himalayan arc [Kapp and Guynn, 2004].

[5] Direct observational constraints on the crust and mantle structure beneath the rifts are needed to understand the east–west extension of the Tibetan Plateau and the dynamic processes that generate the rifts. In this paper, we present tomographic images of the lower crust and upper mantle beneath a north–south trending rift in the region of the Eastern Himalayan Syntaxis, which marks the eastern end of the Himalayan orogenic belt (Figure 1a). The results support the notion of the mantle control of the north–south trending rifts and suggest that delamination of the mantle lithosphere plays an important role in the rise of the plateau in the study area.

2. Data

[6] The data used in this study come from three sources in southeastern Tibet: (1) the Namche Barwa seismic experiment, which deployed a 50-station array of broadband seismometers and a 20-element array of short-period seismometers during 2003–2004 [Sol et al., 2007], (2) the Global Seismic Network (GSN) station LSA, and (3) the broadband stations in the Bhutan experiment, which recorded concurrently with the Namche Barwa array in 2003 (Figure 1a). We used 695 events occurred during the operation period of the stations in 2003 and 2004, and with magnitude greater than 5.5 (Figure 1b). With the multichannel waveform cross correlation method [VanDecar and Crosson, 1990], we measured differential travel times of teleseismic P and S waves between different stations in short, intermediate, and long periods: 0.5–2.0, 0.1–0.5, and 0.03–0.1 Hz for P waves and, 0.1–0.5, 0.05–0.1, and 0.02–0.05 Hz for S waves. The P differential traveltimes are measured on vertical component seismograms and those of S on transverse components. In order to avoid the ambiguity of long-period phase arrivals in the triplication range and the effects of the core-mantle boundary, we have considered only data with epicentral distance between 34° and 81°.

[7] Our final data set consists of ~36313 P (19,131, 10,090 and 7092 for short, intermediate, and long periods, respectively) and ~15043 S differential traveltimes (6629, 6040, and 2374 for short, intermediate, and long periods, respectively), which are then utilized to invert for spatial variations in P and S wave velocity perturbations according to the 3-D finite frequency kernel formulation [Dahlen et al., 2000; Hung et al., 2004]. For the inversion, we used either the combination of data at all frequency bands, the combination of only high- and intermediate-frequency data and only the high-frequency data to test the reliability of different data. The different combinations yield similar results, though adding lower frequency data reduces the variance reduction of the final model. This is likely caused by the relatively larger errors in the long-period measurements. The tomographic models discussed in this paper come from the inversion of the high- and intermediate-frequency data. Figure 2a shows the average teleseismic differential traveltime residuals at each station for P and S waves. They are consistent with P and S wave velocity...
anomalies obtained from the inversion, discussed in sections 3–5.

3. Methodology

Most of global and regional tomographic studies are based on ray theory, where seismic waves are assumed to have an infinite frequency and the arrival time of a body wave phase depends only upon the wave speed along the geometrical ray path between the source and receiver. However, because of wave front healing, scattering, and other diffraction effects, the traveltime of a finite frequency seismic wave is sensitive to a three-dimensional structure off the ray path [Dahlen et al., 2000]. Using body wave ray theory in conjunction with the Born single-scattering approximation, Dahlen et al. [2000] derived the formulation of the Born-Fréchet kernels for a seismic phase, which expresses the influence of velocity perturbations upon a finite frequency travel time shift:

$$\delta t = \int \int K(x)\delta v(x)/v(x)d^3x$$

(1)

where $K$ is the 3-D Fréchet sensitivity kernel for a shift $\delta t$, measured by cross correlation of an observed pulse with its spherical Earth synthetics. It is expressed by the formulation [Dahlen et al., 2000]

$$K = \frac{1}{2\pi^2 c} \int_0^{\infty} \omega |s_{\text{syn}}(\omega)|^2 \sin(\omega \Delta T) d\omega$$

(2)

where $\Delta T = T' + T'' - T$ represent the difference in travel time for the path with a detour through a single point scatterer between the source and receiver (time $T' + T''$) relative to the corresponding direct ray path (time $T$); $R, R'$ 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 power spectrum, $|s_{\text{syn}}|^2$, is for the synthetic seismograms used for cross correlation. Cross-correlating with synthetics in different bandwidths takes advantage of the frequency dependence of the traveltimes.

In this study, we used the finite frequency seismic tomography, as described by Hung et al. [2004] and Yang et al. [2006], to retrieve the crustal and mantle structures beneath southeastern Tibet. Our measured data $d_i$ of $P$ and $S$ differential travel times can be inverted to produce

![Figure 2. (a) Mean traveltime residuals at each station for (left) $P$ wave and (right) $S$ wave. Blue indicates an earlier than expected arrival; red indicates a later than expected arrival. Delay magnitude correlates linearly with the size of symbols. (b) Station terms for (left) $P$ wave and (right) $S$ wave, which are solved simultaneously in the inversion in order to absorb travel time fluctuations due to uncorrected shallow structures.](image)
is the identity matrix. The damping parameter 

\[ S \]  

is the position vector in 3-D vector model space, 

\[ \mathbf{x} \]  

is the model parameter vector, and 

\[ \mathbf{P} \]  

is the wave. For 3-D tomography, we have 

\[ d_i = \int_D g_i(x)m(x)d^3x \]  

where 

\[ d_i, i = 1 \ldots N, \]  

represents the \( i \)th differential travel time data; \( \mathbf{x} \) is the position vector in 3-D vector model space, \( D \); and \( g_i \) represents the 3-D Fréchet sensitivity kernel relating to the data \( d_i \) with the model function \( m(\mathbf{x}) \).

The crustal and mantle volume beneath the region of study is parameterized with regular 3D grids of \( 33 \times 33 \times 33 \) centered at \( (93.0^\circ E, 30.0^\circ N) \), and with a dimension of \( 18^\circ \) in longitude, \( 14^\circ \) in latitude, and \( 1200 \) km in depth. This results in a grid spacing of \( \sim 53 \) km in longitude, \( 46 \) km in latitude and \( \sim 40 \) km vertically. With this parameterization, equation (3) can thus be written as

\[ d_i = G_i m_l \]  

where \( d_i \) is the \( i \)th differential traveltime data, \( G_i \) represents the differential value of the integrated volumetric kernels contributing to the \( l \)th node, and \( m_l \) is the model parameter at the \( l \)th node. The inversion problem is resolved by the standard damped least square method [Paige and Saunders, 1982a, 1982b]:

\[ \mathbf{m} = (G^T G + \theta \mathbf{I})^{-1} G^T d \]  

where \( \mathbf{I} \) 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 3). We choose 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 \( \sim 85\% \) for \( P \) wave and \( \sim 80\% \) for \( S \) wave.

\[ \text{Figure 3. Trade-off curve between the model L2 norm and the data variance reduction for different values of the norm damping parameter for } P \text{ and } S \text{ waves. The larger square and dot represent parameters used to obtain the } P \text{ and } S \text{ velocity models discussed in this paper.} \]
(a) Checkerboard resolution tests: (a) Horizontal slices of the input and retrieved \( P \) wave velocity models at depths 112 km and 338 km. The magnitudes of the input anomalies are \( \pm 2.0\% \); the sizes of the anomalies are (left) \( \sim 200 \text{ km} \times 200 \text{ km} \), (middle) \( \sim 150 \text{ km} \times 150 \text{ km} \) and (right) \( \sim 100 \text{ km} \times 100 \text{ km} \). The rectangle with dashed lines on the horizontal slice at 112 km depth marks the area which we consider well resolved and used for Figure 6. (b) (left) Vertical slice along the path A-B through the input and retrieved \( P \) wave velocity model. The horizontal size of the anomalies is \( \sim 200 \text{ km} \times 200 \text{ km} \) and the vertical size is \( \sim 120 \text{ km} \). (right) The same vertical slice for the input anomalies with the horizontal scale of \( \sim 150 \text{ km} \times 150 \text{ km} \) and the vertical scale of \( \sim 120 \text{ km} \).

Figure 4. (a) Checkerboard resolution tests: (a) Horizontal slices of the input and retrieved \( P \) wave velocity models at depths 112 km and 338 km. The magnitudes of the input anomalies are \( \pm 2.0\% \); the sizes of the anomalies are (left) \( \sim 200 \text{ km} \times 200 \text{ km} \), (middle) \( \sim 150 \text{ km} \times 150 \text{ km} \) and (right) \( \sim 100 \text{ km} \times 100 \text{ km} \). The rectangle with dashed lines on the horizontal slice at 112 km depth marks the area which we consider well resolved and used for Figure 6. (b) (left) Vertical slice along the path A-B through the input and retrieved \( P \) wave velocity model. The horizontal size of the anomalies is \( \sim 200 \text{ km} \times 200 \text{ km} \) and the vertical size is \( \sim 120 \text{ km} \). (right) The same vertical slice for the input anomalies with the horizontal scale of \( \sim 150 \text{ km} \times 150 \text{ km} \) and the vertical scale of \( \sim 120 \text{ km} \).

resolved area (rectangle with dashed lines in Figure 4a) as a function of depth (Figure 6a), as well as the standard deviation of the difference between the input and output models as a function of depth (Figure 6b), for the three cases of the input velocity anomalies above. For the \( P \) and \( S \) models with the two larger-sized anomalies, the correlation coefficients between the input and output models are very good in the upper 450 km depth (\( \sim 0.9 \)). The standard
Checkerboard resolution tests: (a) Horizontal slices of the input and retrieved $S$ wave velocity models at depths 112 km and 338 km. The magnitudes of the input anomalies are ±0%; the sizes of the anomalies are (left) ~200 km × 200 km, (middle) ~150 km × 150 km and (right) ~100 km × 100 km. The rectangle with dashed lines on the horizontal slice at 112 km depth marks the area which we consider well resolved and used for Figure 6. (b) (left) Vertical slice along the path A-B through the input and retrieved $S$ wave velocity model. The horizontal size of the anomalies is ~200 km × 200 km and the vertical size is ~120 km. (right) The same vertical slice for the input anomalies with the horizontal scale of ~150 km × 150 km and the vertical scale of ~120 km.

Figure 5.
Figure 6. (a) Correlation coefficient between the input and output (left) $P$ and (right) $S$ models as a function of depth for the cases in which the sizes of the input anomalies are $\sim 200 \text{ km} \times 200 \text{ km}$ (line with squares), $\sim 150 \text{ km} \times 150 \text{ km}$ (line with stars), and $\sim 100 \text{ km} \times 100 \text{ km}$ (line with triangles). The calculations are for the area that we consider well resolved (rectangle with dashed lines in Figure 4a).

(b) Standard deviation of the difference between the input and output models as a function of depth for the three cases of the input anomalies.
deviations of the difference between the input and output models are respectively 0.7% and 1.3% for the P and S models. This means that about 65% of the amplitudes of the input models at these scales can be recovered. For the case in which the size of the input anomalies is \( \sim 100 \text{ km} \times 100 \text{ km} \), the recovery in amplitude is about 55% in the upper 300 km and degrades gradually at greater depths.

5. Results and Discussions

[14] Figure 7a shows P and S wave velocity structures in the crust and upper mantle beneath southeastern Tibet at four different depths. There is a prominent north–south trending, low-velocity structure at \( \sim 92^\circ \) longitude from the lower crust to \( \sim 300 \text{ km} \) depth. This structure is observed on both the P and S velocity models. It extends across the Indus-Yarlung Suture and coincides strikingly well with a rift on the surface. The shallow (75 km and 112 km) velocity structure (Figure 7a) is in agreement with Pn wave tomography [Liang and Song, 2006], which also shows a north–south trending, low-velocity anomaly in the same region, albeit at a coarse resolution and centered somewhat west of the low-velocity feature in this study. At depth above 250 km, there is a high-velocity anomaly east of the north–south trending, low-velocity feature and north of the surface trace of the main boundary thrust fault (MBTF) between the Indian and Eurasian plates. We associate this high-velocity anomaly to the Indian lithosphere underthrusting beneath the plateau. While we agree with the interpretation of Pn tomography that the Indian lithosphere advances further north near the eastern corner than to the west [Liang and Song, 2006], the high-velocity anomaly associated with the Indian lithosphere appears to be bounded by the Jiali fault. There is no evidence for a subducted Indian lithosphere at depths greater than 250 km beneath the study area.

[15] West of the north–south trending, low-velocity anomaly is a high-dipping-angle, high-velocity anomaly
that extends into the upper mantle transition zone, as shown on the cross sections in Figure 7b. Resolution tests using synthetic slabs with a thickness of ~120 km, placed either in the upper and lower parts of the upper mantle, show little vertical smearing in the recovered structures (Figure 8). We conclude that the observed high-velocity anomaly is a robust feature. This anomalous structure is also more or less tabular as opposed to cylindrical. The possible causes for seismic velocity anomalies in the mantle are variations in temperature, chemical composition, and volatile contents. Thermal variations alone cannot fully account for the observed high-velocity anomaly, which coincides with a low $V_p/V_S$ ratio (Figure 7b). A temperature reduction of 400°C for a wide range of upper mantle compositions causes approximately a ~0.5% change in $V_p/V_S$ [Hacker and Abers, 2004; Boyd et al., 2004], much smaller than the observed anomaly (~1.5%). Resolution tests show that for features comparable to the interpreted, delaminated mantle lithosphere and the Indian mantle lithosphere beneath the eastern indentor, the inversion yields well-recovered $V_p/V_S$ ratios, with magnitudes slightly less than those of the input anomalies (Figure 8). So the magnitude of the real $V_p/V_S$ ratio of the high-angle, high-velocity anomaly is likely greater than the observed value in Figure 7b, requiring a compositional change, which could be accounted for by a refractory mantle depleted of volatiles [Hacker and Abers, 2004; Boyd et al., 2004]. The fact that the high-velocity anomaly in the shallow mantle beneath the eastern indentor, which we identify as the Indian mantle lithosphere, also has a low $V_p/V_S$ of a similar magnitude supports the interpretation that the high-dipping-angle, high-velocity anomaly is a sunken mantle lithosphere. The north–south orientation of this feature excludes the possibility that it is a detached Indian plate. We suggest that the low-velocity anomaly immediately east of and above this high-velocity feature represents the mantle asthenosphere that filled the void left by the sunken Eurasian mantle lithosphere (Figure 9). The tabular nature of the high-angle, high-velocity anomaly, the asymmetry of the low-velocity anomaly above it and the rise of this low-velocity anomaly to the base of the crust at ~70 km suggest a process in which the entire mantle lithosphere peeled off as opposed to a partial removal of the mantle from the base of a thickened lithosphere by viscous flow. The shallow asthenosphere and associated heating and weakening of the overlying crust could thus initiate and localize the rifts observed on the surface of the Tibetan Plateau, at least in the region of the Eastern Himalayan Syntaxis.

[16] Roughly coincident with the onset of the east–west extension, potassic volcanism became widespread on the Tibetan Plateau. In southern Tibet, volcanism may have started slightly earlier (~16–20 Ma) and then ceased around 10 Ma [Williams et al., 2004]. The small magnitude of the east–west extension is insufficient to cause decompressional melting of the asthenosphere [McKenzie and Bickle, 1988]. Geochemical modeling shows that the magmas in southern Tibet were derived from a small degree (~2%) of partial melting of metasomatized (phlogopite) peridotite in the spinel stability field (at depths of 65–80 km) [Williams et al., 2004]. Our tomographic results are consistent with the geochemical analysis. The delamination of the mantle lithosphere brought the subcontinental mantle
lithosphere into direct contact with asthenospheric temperatures, so that metasomatized peridotite, which has a lower solidus, underwent partial melting right beneath the crust or released its volatiles into the shallower mantle. The tabular geometry of the north–south trending low-velocity anomaly suggests that the release of volatiles from the sunken mantle lithosphere may contribute significantly to the velocity reduction above it since a buoyant and low-viscosity upwelling tends to form a cylindrical not tabular anomaly. Melting would cease or be greatly reduced once the volatiles in the peridotite are depleted by the melting process. The fact that extension postdates the earliest magmatism by a few million years [Williams et al., 2004] further suggests a causal relationship between the mantle processes and the surface rifts.

[17] The thickening and deformation of the lithosphere caused by the Indo-Eurasia collision is certainly a likely geological setting for the initiation of the lithospheric delamination [Bird, 1979], though the exact mechanism for the initiation of the delamination of the mantle lithosphere is unclear. Shear heating through viscous dissipation in the lithosphere likely plays a role in weakening the lithosphere [Kincaid and Silver, 1996; Schott et al., 2000] and the initiation of the delamination at a localized shear zone. Numerical simulations on the dynamics of the mantle lithosphere show that mantle delamination and detachment likely occur if the lithosphere is substantially thickened and that the process is strongly controlled by the rheology of the lower crust [Schott and Schmeling, 1998; Morency and Doin, 2004]. These experiments show that a low-viscosity and hot lower crust favors especially the delamination phenomenon in a thickened lithosphere and the Tibetan Plateau fulfills well these conditions. Several recent Himalayan-Tibetan tectonic investigations have proposed a weak middle-lower crust, which allows crustal flow, to explain the building of the southeastern margin of the Tibetan Plateau [Clark and Royden, 2000; Beaumont et al., 2001; Schoenbohm et al., 2006]. This idea is supported by results from geophysical studies that show evidence for localized partial melt within the middle crust in Tibet [Kind et al., 1996; Wei et al., 2001] and a radial anisotropy that can be caused by channel flow within the mid-to-lower Tibetan crust [Shapiro et al., 2004]. However, contrary to the notion that the north–south rift zones in the Tibetan Plateau are shallow features, formed by the eastward motion of the shallow crust that are decoupled from the mantle lithosphere by a low-viscosity lower crust, our results suggest an upper mantle origin of the rift zones, at least in southeastern Tibet, where mantle lithosphere delamination plays a key role in the process of the rise of the plateau.

6. Conclusion

[18] Finite frequency tomography of teleseismic P and S travel times shows the presence of a low-velocity anomaly in the crust and upper mantle down to ~300 km depth beneath a north–south trending rift zone in southeastern Tibet. This low-velocity anomaly is situated above a tabular, high-dipping-angle, high-velocity anomaly that extends into the upper mantle transition zone. The $V_P/V_S$ ratio of this high-velocity anomaly suggests that temperature variations are not the only cause and a highly melt-depleted mantle is required. These results provide evidence for the mantle lithosphere delamination and its link to north–south trending rifts in southeastern Tibet. Studies on an extended region in southern Tibet are thus needed to understand whether this phenomenon is limited to the study area or occurs in a broad region where mantle lithosphere delam-
flation plays an important role in the elevation and deformation of the Tibetan Plateau.

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Y. Ren and Y. Shen, Graduate School of Oceanography, University of Rhode Island, South Ferry Road, Narragansett, RI 02882, USA. (ren@gso.uri.edu; yshen@gso.uri.edu)