Seismic signature of the collision between the east Tibetan escape flow and the Sichuan Basin

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1. Introduction

As a result of the continuous convergence between the Indian and Eurasian tectonic plates that has been taking place over approximately the last 45 million years, it is estimated that the crust in the collision zone has been shortened by at least 1500 km (Molnar and Tapponnier, 1975; Armijo et al., 1986; England and Houseman, 1986; England and Molnar, 1997; Yin, 2000). Subsurface compensations may have occurred in a variety of ways, such as by crustal thickening (Allegrè, 1984), denudation (Meng et al., 2006), slip partitioning (Chen et al., 1994; Tapponnier et al., 2001), subduction of the Indian mantle lithosphere (Kosarev et al., 1999; Kumar et al., 2006; Li et al., 2008a), lithospheric detachment (Houseman et al., 1981, Molnar, 1988), subduction of the Asian lithosphere (Willett and Beaumont, 1994; Kind et al., 2002) and eastward escape (Royden et al., 1997, 2008; Clark and Royden, 2000; Klemperer, 2006). GPS displacement vectors (Gan et al., 2007) and SKS anisotropy measurements (Wang et al., 2008) indicate that the Tibetan crust (and possibly also the lithosphere and asthenosphere) is escaping eastwards, and that the main portion of the flow is being redirected towards the south-east after it encounters the Sichuan Basin (Fig. 1). As part of the Yangtze craton, the Sichuan Basin is an old, intact block that remained undeformed during the orogenic events that formed the Tibetan plateau. Anomalously cold temperatures derived from tomographic observations extend down to at least ca 200 km beneath the Sichuan Basin and are also present in the Mantle Transition Zone (MTZ) to the east and south-east of the Sichuan Basin (Li et al., 2008a, Yi et al., 2008). However, the mode of collision that occurs between the Tibetan and Sichuan lithospheres (e.g. thickening or subduction) remains poorly understood. The powerful interaction between eastern Tibet and the Sichuan Basin is characterized by the Longmen Shan (LMS), which has the steepest topography gradient in Tibet (Royden et al., 1997; Clark and Royden, 2000; Klemperer, 2006). The LMS fault system, which marks the border of the Songpan-Ganzi block of
eastern Tibet and the Sichuan block of the Yangtze craton, is dominated by dextral strike-slip and has a significant thrust component (Burchfiel et al., 1995, 2008). The 12 May 2008 Wenchuan earthquake, characterized by a crustal shortening of 8.5 m and an uplift of 7.5 m, confirms the view that the east Tibetan escape flow is responsible for the high topography in this region (Xu et al., 2009).

In order to study the nature of the collision occurring between the Tibetan and the Sichuan lithospheres, we conducted a passive seismic experiment across the LMS fault belt at the eastern rim of the Tibetan plateau between August 2006 and July 2007. This data set has yielded images of crustal variations across the LMS fault belt that supports lower crustal channel flow or tectonic escape (Zhang et al., 2009). The deeper structure of the same profile is analyzed herein. We use the teleseismic P and S receiver functions (Langston, 1977; Vinnik, 1977; Farra and Vinnik, 2000) to detect converted seismic waves from the crust–mantle boundary (Moho), from the Lithosphere–Asthenosphere Boundary (LAB), and from the discontinuities at 410 km and 660 km depth that confine the MTZ. This provides a good opportunity to understand the shallow and deep interactions that occur between the Tibetan escape flow and the resistive Sichuan Basin.

2. Data acquisition and generation of receiver function images

2.1. Moho and LAB from P and S receiver functions

We used a teleseismic data set consisting of one year of data obtained from 29 three-component seismic stations (nine Reftek-130 and twenty Reftek-72A data loggers and Guralp CMG 3ESP sensors) along the profile in the western part of Sichuan Province (Fig. 1). In these observations, stations S01–S07 were located in the Sichuan Basin and stations S13–S29 were located in the Songpan-Ganzi block, with station intervals of ~15 km. Stations S08–S12 were located in the LMS thrust fault zone at intervals of 10 km (Zhang et al., 2009). During the period of observation, we recorded 264 earthquakes of magnitude Ms > 5.0 in the distance range between 30 and 90° (Fig. 2). We obtained 1823 Receiver Functions (RFs). The mean number of retained RFs per station was 64. Fig. 2b shows all individual raw P-receiver functions plotted in trace spacing within a time window of 0–80 s after P arrival for all 29 stations. The traces are moveout-corrected for a fixed ray parameter of 0.0573 s/km (corresponding to a slowness of 6.4 s/° or an epicentral distance of 67°) using the IASP91 reference model (Kennett and Engdahl, 1991). The figure serves to show...
the raw data set. Piercing point locations of the mantle phases are not properly considered. No summation is applied to suppress the noise level. To increase the coherence of the mantle phases a 5-s lowpass filter is applied to delay times later than 35 s. Signals from the Moho, LAB, 410 and 660 km discontinuities are marked. The Moho signal is red, which indicates that it is positive, and has a velocity that increases in the downward direction. A clear change in delay time of the Moho conversion is apparent between eastern Tibet and the Sichuan Basin. These times are about 6–8 s beneath the Songpan-Ganzi block (in eastern Tibet), and about 4–5 s beneath the Sichuan Basin. Details of the internal crustal structure derived from the same data set are given by Zhang et al. (2009) and may be summarized as follows: the Moho shallows from about 55–58 km under the Songpan-Ganzi block in eastern Tibet to 36–40 km under the western Sichuan Basin; a negative phase was observed in the lower crust under Songpan-Ganzi; the average crustal Vp/Vs ratios vary in the range of 1.75–1.88 under Songpan-Ganzi in east Tibet, and in the range of 1.8–2.0 under the LMS fault belt, and decrease to less than 1.70 in the NW part of the Sichuan Basin. The results inferred a low viscosity zone in the lower crust beneath eastern Tibet (Zhang et al., 2009). Fig. 3 shows migrated P-receiver functions down to a depth of 800 km. In Fig. 2b, 410- and 660-km phases can be seen also though they are much weaker compared to Moho conversion phase and the signal/noise ratio is not very high. Then we convert these P-receiver functions in time domain into the migrated P-receiver function image (Kind et al., 2002). The section was smoothed over 5 km in vertical direction and 60 km in horizontal direction.

The IASP91 model was used for migration. Fig. 3 displays the resultant migrated P-receiver function image. A clear blue signal (meaning negative amplitudes which indicate velocity decrease downward) is observed over the entire profile and interpreted as LAB. It is shallower beneath eastern Tibet than beneath the Sichuan Basin. This dip direction is opposite to that of the Moho. The LAB lies at a depth of about 70–80 km under east Tibet, at about 120 km under the western margin of the Sichuan Basin, and then deepens to about 150 km at the eastern end of our profile. The marked LAB in Figs. 2b and 3 might be argued as Moho phase sidelobe. A sidelobe has always more or less the same time difference to the main part of the signal, if the spacing between main part of the phase and the "sidelobe" varies from location to location then it can be considered as an independent phase mostly caused by the geological structure. The blue phase marked LAB in Fig. 3 is indeed at the left side parallel and very close to the Moho which could give rise to the assumption that it is a sidelobe. However, if we look close to the LMS and the Sichuan Basin, LAB and Moho clearly separate which proofs that these two phases are independent. The Moho is shallowing and the LAB is deepening towards the right of the figure. In order to check these results for the Moho and LAB structure, we applied S receiver function analysis, even though this data set is not as large as for the P-receiver functions. The S receiver functions obtained along the profile are shown in Fig. 4. Fig. 4a shows the location of the boxes used for

Fig. 2. (a) Distribution map of the events with Ms ≥ 5.0 and epicentral distances of 30–90° used in this study. (b) Raw P-receiver functions for all 29 stations along the profile, grouped by stations and plotted at equal-spacing. Red and blue colors are used to shade the positive and negative amplitudes, respectively. The later part (>35 s) is lowpass filtered to 5 s. Main discontinuities are marked by Moho, LAB, 410-km and 660-km, respectively.
the summation of the traces. The LAB changes in depth below the LMS fault in an almost step-like fashion (Figs. 3 and 4b) from about 70–80 km below eastern Tibet to 100–150 km below the Sichuan Basin. Similar depth of LAB beneath East Tibet is also reported by surface wave inversion (75–80 km, Lebedev and Agius, 2009). The shallower LAB beneath eastern Tibet than beneath the Sichuan Basin is confirmed by the S receiver function results (Fig. 4b). Lower velocities at 100 km depth below eastern Tibet than below the Sichuan Basin as a consequence of the location of the LAB are supported by ambient noise surface wave tomography studies (Li et al., 2009) and other Rayleigh wave tomography studies (Li et al., 2008a; Yao et al., 2008; Yi et al., 2008).

2.2. 410 km and 660 km discontinuities obtained from the receiver functions

The discontinuities at 410 km and 660 km are also clearly visible in the migrated section shown in Fig. 3. Throughout the entire profile, the 660 is about 10–20 km below its global average value in the IASP91 model with only insignificant deepening below eastern Tibet. The 410, however, is exactly at the IASP91 value below the Sichuan Basin and clearly deepens towards eastern Tibet by about 30 km. In order to confirm this observation, we show in the time domain (Fig. 5) the stacks of the P-receiver functions. The binning stacks were obtained using a 150 km window of piercing points at a depth of 530 km. The window was moved using a step length of 50 km.

Our receiver function signals from the 410 km discontinuity indicate that the clear change in lithospheric structure that occurs between eastern Tibet and the Sichuan Basin is also visible at the MTZ (Fig. 3). The 410 is clearly deeper below eastern Tibet than below the Sichuan Basin. The 660 remains more stable. This results in a thinner MTZ below eastern Tibet than below the Sichuan Basin. No significant variation in MTZ thickness has been observed in central Tibet (Kind et al., 2002). In western Tibet, the MTZ has a uniform thickness between the Tarim Basin and the Kunlun range (Wittlinger et al., 2004). This could imply that any tectonic activity in central and western Tibet is confined to the upper mantle and does not reach down as far as 410 km below the surface. Contrary to that, the MTZ below the eastern margin of Tibet could be influenced by the lithospheric collision above.

The 410 km and 660 km discontinuities are commonly considered as mineral phase transformations of olivine into its high pressure forms, which cause sharp vertical gradients in density and seismic velocity. The 410 km discontinuity marks the transition from olivine to wadsleyite, and the 660 km discontinuity marks the transition from ringwoodite to perovskite + magnesiowustite. Experimental studies have shown that both reactions are sensitive to temperature and have Clapeyron slopes that are opposite in sign (Bina and Helffrich, 1994). In the absence of other effects, a lateral increase in temperature at the level of the transition zone should be reflected in a deepening of the 410 km discontinuity and a shallowing of the 660 km discontinuity (and vice versa). The temperature in the MTZ, which strongly influences the mechanical strength of the mantle materials, can be derived from the thickness of the transition zone (Helffrich, 2000). Compared to the global average MTZ thickness of 250 km in the IASP91 global reference model (Kennett and Engdahl, 1991), the MTZ throughout our profile experiences a significant thinning of about 15 km to a thickness of 235 km beneath eastern Tibet. Beneath the western Yangtze block under the Sichuan Basin, however, it experiences a thickening of about 15 km to 265 km. This effect would lead to temperatures being raised in the MTZ under eastern Tibet by about 300° compared to the Sichuan Basin. Tomographic velocity measurements...
However, show higher velocities (meaning colder temperature and larger MTZ thickness) beneath eastern Tibet than below the Sichuan Basin, which contradicts the hypothesis that our observed thinning of the MTZ below eastern Tibet could be caused by increased temperature. We will return to a discussion of the variation in the thickness of the MTZ in Section 4.

3. Validation assessment of the Moho, LAB, 410 km and 660 km discontinuities

The identification of the Moho and the 410 km and 660 km discontinuities is a routine procedure in the analysis of P-receiver functions. Our observations of these discontinuities are clear. However, we have also made observations of the LAB using the P-receiver functions (Figs. 2 and 3). The use of P-receiver functions to study the LAB is generally characterized by problems caused by crustal multiples, because their presence often masks the LAB. No clear multiples are visible in our data, either in the time domain section (Fig. 2b) or in the migrated section (Fig. 3). In our case, the LAB cannot be mistaken for crustal multiples because the signals associated with these should arrive much later beneath the thick eastern Tibetan crust than beneath the thinner crust of the Sichuan Basin. The phase marked LAB in Fig. 3 shows the opposite trend. The P multiples within the crust below the Sichuan Basin must arrive after about 20 s, because the direct conversion arrives after about 5 s (see e.g. Yuan et al., 2002). This is much later than the arrival time of the LAB beneath the Sichuan Basin of about 12 s. Multiples do not show up in the S receiver functions in Fig. 4b but the LAB is visible there. The LAB in Figs. 3 and 4b can also not be caused by side lobes of the Moho signal because it appears at about the same depth in the longer period S receiver functions in Fig. 4b and in the shorter period P-receiver functions in Fig. 3. Sidelobes would also be expected at a constant delay from the main signal which is clearly not the case in both P and S receiver functions. Observations of the LAB in P-receiver functions are not unusual. Such observations have also been obtained worldwide by Rychert and Shearer (2009). S receiver functions are very useful for the verification of LAB observations, because of the absence of multiples (see e.g. Kumar et al., 2007). They are less commonly applied for this purpose, however, because there are usually fewer high quality S wave data available, especially in short term projects. The Moho and the LAB are visible in the S receiver functions in Fig. 4, and arrive at approximately at the same time as in the P-receiver functions (Fig. 4b), although the piercing points of the P and S receiver functions are not identical. This confirms the observations of the Moho and LAB obtained from the P-receiver functions. In order to confirm the P-receiver function results from the MTZ in Fig. 3, we have split the distribution of the piercing points into two regions (see Fig. 6a) and computed two migrated profiles (Fig. 6b). The essential features of the structure of the 410 km and 660 km discontinuities are preserved especially in the section A–A’ in the northeast part of the region. The depth variation of the 410 km discontinuity is somewhat less pronounced along profile B–B’ which could indicate a lateral variation of the 410 topography. These features are (1) the deepening of the 410 below eastern Tibet, (2) the agreement of the 410 below the Sichuan

![Fig. 3. Migrated P-receiver function image along the profile. P-to-S conversions from the Moho, the Lithosphere–Asthenosphere Boundary (LAB) and the discontinuities around the 410 km and 660 km depth are clearly seen. Reddish colors indicate positive (velocity increasing downwards) and bluish negative (velocity decreasing downwards) signals.](image-url)
Due to a lack of data, we could not obtain S receiver function data from the upper mantle. The determination of the depth using measured differential times requires the use of a velocity model. Velocity models are not easily obtained from receiver function data. Techniques exist for inverting receiver function waveforms into velocity–depth models (e.g. Kind et al., 1995). However, such techniques are non-unique and we prefer not to use them here. Certain information may be obtained from receiver function images even if the velocity model is incomplete, i.e. the differential depths of the two discontinuities. For example, we may obtain the apparent thickness of the MTZ in Fig. 3. The thickness of the MTZ depends only on the velocity within the transition zone and not on the velocity model above 410 km depth. This in turn means that the apparent thinning of the MTZ below eastern Tibet does not depend on the shallower structure. Li et al. (2008a) show high resolution tomographic velocity variations in the region of our profile. Their results indicate a velocity 2–3% higher at 400 km depth below eastern Tibet than below the Sichuan Basin. We observe a thinning of the transition zone of about 30 km, i.e. more than 10%, which means the thinning of the transition zone below eastern Tibet is indeed real and cannot be explained by a lateral velocity increase. It follows from the same argument that the mantle part of the lithosphere (i.e. that is the difference between the LAB and the Moho) is independent of the crustal structure above it. Velocity measurements from around the Sichuan Basin were obtained by Yao et al. (2008), Li et al. (2008a) and Wang et al. (2007) from tomography and controlled source studies. According to Li et al. (2008a) the upper mantle shows almost 4% higher velocities down to at least 200 km depth beneath the Sichuan Basin than beneath eastern Tibet. From this we conclude that the observed time differences between the Moho and the LAB along our profile are caused by changes in the thickness of the mantle lithosphere and not changes in velocity. Otherwise, we would expect to see reversed time differences. Moho depths obtained from surface wave tomography in the south of our profile resulted in about 60 km depth beneath eastern Tibet and in about 40 km below the Sichuan Basin (Yao et al. 2008). Almost identical results for the Moho depths were obtained by Wang et al. (2007) from controlled source experiments. Topographic elevations and sedimentary coverage in the Sichuan Basin were considered in the results by Wang et al. (2007). Xu et al. (2007) also obtained very similar results for the Moho depths beneath the Sichuan Basin and eastern Tibet, using receiver function inversion from data obtained to the south of our profile. We used the IASP91 reference model to achieve the migration of our data and arrived at practically the same Moho depths. This shows that the migration of receiver functions is not very sensitive to the model changes. The reason for this is that in receiver function techniques, the P and S differential times are used and not absolute travel times. Differential times are more sensitive to the Vp/Vs ratio than to the absolute velocities.

4. Discussion of the variation in MTZ thickness across the Longmen Shan fault belt

4.1. MTZ thickening below Sichuan Basin from lower-angle Pacific plate subduction

The reason for the thickened MTZ below the Sichuan Basin relative to the global average is unclear. Changes in the thickness of the
transition zone over such short distances have so far only been observed in subduction zones (Li et al., 2000; Li and Yuan, 2003; Liu et al., 2003; Tonegawa et al., 2006). In the following, we will discuss the possibility of the influence of lower-angle Pacific plate subduction on MTZ thickening beneath the Sichuan Basin. There are indications from tomography studies (e.g. Fukao et al., 2009) that a flat subduction of the Pacific plate in the MTZ in the north of the Yangtze Craton could reach as far as the Sichuan Basin (Lebedev et al., 2003; Li et al., 2008b).

We have looked at P-to-S conversions from the 410 km and 660 km discontinuities at a number of seismic stations in China (see locations in Fig. 7a) and found no indication of a large regional thickening of the MTZ that could be related to a flat Pacific slab inside the MTZ. The waveform data from these stations are shown in Fig. 7b. Fig. 7b indicates the observed differential times of the 410 km and 660 km discontinuities at an epicentral distance of 67°. The IASP91 value is 24 s. No significant contiguous zone of a thickened MTZ exists in

Fig. 6. (a) Piercing points of P-receiver functions at 410 and 660 km depths. They are grouped into two subsets to build two transects (AA′ and BB′); (b) Migrated MTZ images along two transects (AA′ and BB′). (c) Section of all receiver functions. The deepening of the 410 below eastern Tibet is visible especially in the AA′ transect.

Fig. 7. (a) Location map of permanent seismic broadband stations in China. The observed differential times of the 410 km and 660 km discontinuities corrected to a 67° epicentral distance are also shown. The IASP91 value is 24 s. No significant contiguous zone of a thickened MTZ related to the slab stagnation of the western Pacific subduction zone exists. The seismic profile from the Sichuan Basin is marked on the map. (b) Waveforms of the P-to-S conversions from the 410 km and 660 km discontinuities at the stations shown in (a) lined up along the 410 signal and sorted by increasing differential time. The maximum deviation from the 24 s global average value (marked by the blue line) is ±2 s, corresponding to a MTZ thickness variation of less than ±10%.
relation to the slab stagnation of the western Pacific subduction zone. The maximum deviation from the 24 s global average value (marked by the blue line in Fig. 7b) is ±2 s, corresponding to a variation in MTZ thickness of about ±10%.

### 4.2. Deepening of the 410 discontinuity below eastern Tibet

The deepening of the 410 discontinuity under eastern Tibet could possibly be caused by other factors than increased temperature. We hypothesize that the escape flow beneath eastern Tibet may also be turning downwards and subsiding into the MTZ, besides its horizontal component in southeast direction. The thin mantle lithosphere beneath eastern Tibet would appear to support this view. The downwelling flow does not cause any increase in transition zone temperature, although it may be responsible for compositional differences in the MTZ, e.g. water content. Petrological studies have shown that the MTZ contains much more water than the mantle above and below it, because its main components (wadsleyite and ringwoodite) are known to be water-soluble (Meng et al., 1996). The dry and depleted asthenospheric material that sinks into the MTZ may reduce the water content of the MTZ. Laboratory experiments have shown that a decrease in water content in olivine may lower and broaden the 410 km discontinuity but increase the velocity jump across it, and vice versa (Yusa and Inoue, 1997; Chen et al., 2002). It therefore follows that a lateral variation in water content may change the topography of the 410 km discontinuity.

### 5. Sichuan Basin barrier to the eastward escape flow of eastern Tibet

Our receiver function images show from the Sichuan Basin to eastern Tibet a step-like increase of about 20 km in the Moho depth, a decrease of about 50 km in the LAB depth and a decrease of about 30 km in the depth of the 410 discontinuity. Table 1 summarizes the depths of the LAB, and the 410 km and 660 km discontinuities beneath the stations along the profile. The crustal thickness beneath all 29 stations may be seen in Zhang et al. (2009). The previously mentioned systematic variations of crustal thickness, LAB and 410 beneath the LMS fault belt probably indicate that the Sichuan Basin acts as a barrier to the eastward escape flow of East Tibet that may extend to the whole lithosphere, even to the whole upper mantle.

The steepest topography gradient around the Tibetan plateau margin is commonly accepted to result from the strong interaction between the eastern Tibetan escape flow with the resistive Sichuan Basin. From our passive-source seismic experiment, we contend that the deep process (or processes) that have occurred (and/or are occurring) beneath eastern Tibet may also contribute to the steep rise in topography at LMS, in addition to middle/lower crust channel flow and tectonic escape (Royden et al., 1997; Klemperer, 2006; Zhang et al., 2009). The convergence of the eastward escape flow from eastern Tibet with the Sichuan Basin should produce a downward undulation (or detachment according to Houseman et al., 1981) of the lithosphere beneath both the western margin of the Sichuan Basin and the eastern margin of eastern Tibet (Zhang et al., 2009). Our observation of a lithosphere just 70–80 km thick under Songpan-Ganzi (east Tibet) probably results from delamination of a part of the lithosphere, as suggested by Houseman et al. (1981) and Houseman (1996), or from thermal erosion of the bottom of the lithosphere by the hot, eastward flowing, upper mantle (asthenosphere) beneath Tibetan plateau, or from both of these. This obvious lateral transition between eastern Tibet and the Sichuan Basin down to a depth of 400 km leads us to propose a two-phase accommodation of the shortening. The eastward flow of central and eastern Tibet meets the rigid barrier of the Sichuan Basin, which belongs to the western Yangtze block, and escapes partly into southeastern Tibet or north eastern Tibet around the Sichuan Basin, and is partly downwelling deep into the mantle. The lower part of the hot Tibetan lithosphere may be being delaminated due to collision with the colder Sichuan Basin, or alternatively the bottom of the lithosphere may be being removed by the hot asthenospheric escape flow. This idea could also explain the unusually thin mantle lithosphere beneath eastern Tibet. The delaminated material may be depressing the 410 km discontinuity dynamically. The asthenospheric vertical downward flow beneath the LMS fault belt due to the resistance of the Sichuan basin may be considered as an alternative way to compensate for the continuous convergence between the Indian and Eurasian plates. The extreme topographic relief across the LMS fault belt may be caused not only by lower crust channel flow in eastern Tibet (e.g. Royden et al., 1997, 2000; Clark and Royden, 2000; Klemperer, 2006; Zhang et al., 2009), but also by gravitational buoyancy due to lithospheric delamination (Houseman and England, 1986) or lithospheric bottom removal caused by hot asthenospheric escape flow. The residual (beneath the LAB) of the lithospheric delamination cannot confidentially be recognized in our data. Future studies are needed, such as high resolution seismic tomography or gravity modeling.

### 6. Conclusions

At the India–Tibet boundary, GPS displacements indicate a continuous northward flow of the Indian lithosphere of about 10 mm/yr (Gan et al., 2007). The resultant crustal shortening is accommodated by the subduction of the Indian lithosphere below the Tibetan lithosphere, as shown by numerous studies using receiver functions and tomographic methods (Kind et al., 2002, Wittlinger et al., 2004, Kumar et al., 2006, Li et al., 2008a). The Indian crust is probably separating from the mantle by the LMS fault belt due to the resistance of the Sichuan basin.
lithosphere and propagating as far as central Tibet at a shallow angle (Yuan et al., 1997; Kind et al., 2002; Li et al., 2008a; Royden et al., 2008; Nabelek et al., 2009; Zhang and Klemperer, 2010), whereas the mantle lithosphere itself seems to be subducting deeper into the mantle. The depths of the upper mantle discontinuities directly below the collision zone are generally close to their global average values, indicating weak influence of the collision on the deeper part of the upper mantle. Recent receiver function observations in the eastern Himalaya show some modifications to the thickness of the MTZ that could be related directly to the collision (Singh and Kumar, in press). The processes occurring at the boundary between eastern Tibet and the Sichuan Basin seem to differ in a number of ways. GPS observations indicate that the Sichuan Basin is relatively immobile and that Tibet is moving in an easterly direction. GPS vectors obtained prior to the Wenchuan earthquake also show a redirection of the Tibetan flow around the Sichuan Basin and virtually no displacement near the boundary of the Sichuan Basin. Several meters of east–west shortening and uplift to the east of LMS caused by the Wenchuan earthquake have accommodated several thousand years of continuous displacement of Tibet in an easterly direction. The crustal shortening of 8.5 m and an uplift of 7.5 m during the 12 May 2008 Wenchuan earthquake could not be well modeled with crustal deformation mechanism from only middle/lower crustal escape, and suggests that vertical force contribution to crustal deformation beneath Longmen Shan and East Tibet could not be excluded (Zhang et al., 2008).

The Moho under the NW Sichuan Basin is deepening towards Tibet to accommodate collisional shortening (Zhang et al., 2009), similar to the deepening of the Indian crust in a northerly direction. However, the LAB behaves in an opposite sense below the boundaries of Tibet in the south and east. In the south it is also subducting, along with the Moho, while at the border of eastern Tibet it is shallowing towards Tibet. The 410 km discontinuity is deepening directly underneath the deepening Moho in eastern Tibet. These two observations lead us to suggest the occurrence of delamination of the lower lithosphere below eastern Tibet, which reaches the 410 km discontinuity and is contributing to the rise of the eastern Tibetan margin by gravitational buoyancy (in addition to a lower crustal channel flow), which is similar to the suggestion made by Houseman (1996). Alternatively, the asthenospheric vertical downward flow beneath the LMS fault belt could be accelerating due to the rigid resistance of Sichuan Basin, as a means of compensating for the continuous convergence between the Indian and Eurasian plates.

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