Upper Mantle Heterogeneity and Radial Anisotropy Beneath the Western Tibetan Plateau

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Abstract  P-wave radial anisotropy (RAN) tomography of the upper mantle beneath the western Tibetan Plateau is determined using teleseismic arrival time data recorded at 71 temporal seismic stations of the ANTILOPE-I and Y2 arrays, which reveals a complex deformation pattern in the upper mantle. A high-velocity anomaly with prominent positive RAN is visible to the south of 31°N, which may reflect the underthrusting Indian lithospheric mantle (ILM). Further north, another high-velocity zone appears at depths >200 km, which is possibly related to the foundering ILM, or the northward advancing ILM affected by previously attached Tethyan oceanic slab. Considering the continuity of the high-velocity zones and similarity of the RAN anomalies, we suggest that there is no slab tearing or breakoff at present beneath western Tibet. To the north of 31°N, E-W trending variations of velocity and RAN at shallow depths (<150 km) suggest that small-scale convection occurs between 79°E and 81°E, whereas the residue of the Eurasian lithospheric mantle is located in the eastern side. We deem that the formation of the Ya-Re Rift and Puran Graben was caused by variations of the lithospheric thickness instead of slab tearing or asthenospheric upwelling, considering locations of the rift and the graben and our present results.

Plain Language Summary  Our study region, the Tibetan Plateau, is a hot and ideal region to study the mechanisms of the convergence between continental plates. It is generally considered that the Tibetan Plateau is a product of convergence between the Indian and Eurasian plates, but some fundamental issues are still not resolved. For example, what happens in the northward advancing Indian lithospheric mantle (ILM), and what are the responses of the Eurasian lithospheric mantle (ELM)? Because of the extremely hard environment in the western Tibetan Plateau, seismic tomography is an effective way to illuminate the underground structure by processing earthquake waveforms recorded at seismograph stations. Although the interpretations of velocity anomalies are consistent, the anisotropy features result in different conclusions. Our results reveal both horizontal and vertical variations in the deformation patterns, shedding new light on the tectonic history of the study region. First, although different deformation patterns are detected within the ILM, there is no slab tearing or break-off at present. Second, both ELM and mantle wedge are observed in the upper mantle. Third, the formation of the Ya-Re Rift and Puran Graben is related to variations of the lithospheric thickness rather than slab tearing or asthenospheric upwelling.

1. Introduction

It is generally considered that the tectonic evolution of the Tibetan Plateau is influenced by continental collision (e.g., Yin & Harrison, 2000), but many fundamental issues are still not resolved. For example, what happens in the advancing Indian lithospheric mantle (ILM), and what are the responses of the Eurasian lithospheric mantle (ELM)? Zhou and Murphy (2005) suggested that the ILM has subducted beneath the whole Tibet, whereas Kind et al. (2002) detected the ELM in northern Tibet. In recent years, some studies have revealed significant east-west variations of the northward advancing ILM (e.g., J. T. Li & Song, 2018; X. F. Liang et al., 2016). A subvertical high-velocity zone has been revealed at depths of 100–400 km by seismic tomography (Tilmann et al., 2003), and it was interpreted as a downwelling lithosphere in eastern...
In central Tibet, a tomographic study revealed the underthrusting ILM at depths of 100–300 km with a small dipping angle (H. Zhang, D. Zhao et al., 2016). Results of Pn wave tomography generally support a geodynamic model that the ILM has advanced further north in eastern and western Tibet but it has moved a shorter distance in the central part (C. T. Liang & Song, 2006). In contrast, a receiver-function study found that the ILM underthrusts a longer distance beneath western Tibet but subducts with a much larger dip angle in eastern Tibet (J. M. Zhao et al., 2010). Besides the geophysical results, the dischiral subduction of the ILM in southern Tibet was also revealed by a geochemical study (Hou et al., 2006). On the basis of these results, a fragmented ILM has been proposed to explain most of these features in central and eastern Tibet (X. F. Liang et al., 2016). A recent study of high-resolution Sn and Pn wave tomography suggested the slab tearing beneath the whole southern Tibetan Plateau (Li & Song, 2018), but this method can only map the velocity structure directly beneath the Moho discontinuity. If the fragmented ILM or slab tearing indeed occurred in our study area, more evidence should be found at greater depths. In this study, we investigate whether the slab tearing exists beneath the western Tibetan Plateau by analyzing P-wave velocity and radial anisotropy (RAN).

Another controversial issue is whether the formation of the Tibetan rifts (or grabens) is a crustal process or related to the whole lithosphere or the asthenosphere (Yin & Harrison, 2000). Most of the seismic studies prefer the latter case, including slab tearing in the ILM (X. F. Liang et al., 2016), internal deformation of the ILM and ELM (H. Zhang et al., 2018), and upwelling asthenosphere (Ren & Shen, 2008). However, such seismic evidence depends on the limited data recorded by seismic arrays in central and eastern Tibet. Hence, the resolution and reliability of the previous seismic results in western Tibet are relatively low.

Comparing with seismic velocity variations, which are used to roughly identify the lithosphere and/or asthenosphere in this study, the deformation pattern of the underground medium can be better constrained by seismic anisotropy. In western Tibet, shear-wave birefringence (SWB) features are very different from those in central and eastern Tibet. Sudden increases of splitting time have not been observed, which are generally used to delineate the boundary between the ILM and ELM, and the fast polarization directions are more complex (Levin et al., 2008; Wu et al., 2015). The limitation of the SWB method is the lack of depth resolution because this measurement accumulates the anisotropic information along the raypath. Another common way to probe seismic anisotropy is surface wave analysis. Azimuthal anisotropy from Rayleigh-wave tomography shows a depth-dependent pattern in the Tibetan Plateau, where both amplitude and direction of azimuthal anisotropy exhibit east-west trending variations (Pandey et al., 2015). The surface-wave method has a good depth resolution but a relatively lower lateral resolution even if a good data set is used. Although receiver functions can also reveal depth-varying anisotropy, they have a relatively lower lateral resolution and are usually used to detect seismic anisotropy between strong discontinuities, such as the Moho or an interface within the continental lithosphere (Park & Levin, 2002).

There has been no literature on P-wave RAN in the upper mantle beneath western Tibet. Since RAN tomography has a good lateral resolution and can further reduce travel-time residuals (see below), we adopt this method to map three-dimensional (3-D) isotropic P-wave velocity ($V_p$) structure and RAN beneath the study region. Then, we discuss geodynamic implications of the obtained results. From seismograms of the teleseismic events recorded by two temporal seismic arrays in western Tibet, ANTILOPE-I and Y2, we collect a large amount of travel-time data. These two arrays provide a very good station coverage to explore 3-D variations of seismic velocity and RAN, especially near the westernmost N (the Ya-Re rift) and the Puran Graben (Figure 1a). Our findings provide new insights into the relationship between the complex upper mantle structure and formation of the rift and graben in the study region.

**2. Data and Method**

**2.1. Data**

Figure 1a shows locations of 40 stations of the ANTILOPE-I array (2006–2007) and 31 stations of the Y2 array (2007–2011) used in this work. Figure 1b shows 1439 teleseismic events recorded by these two arrays, whose hypocentral parameters were determined by the United States Geological Survey (USGS) and the International Seismological Centre (ISC). These teleseismic events (Mb > 5.0) have epicentral
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distances of 30°–95°. Using a Matlab software package with a waveform cross-correlation technique (Yu et al., 2017), we manually picked at least six $P$ wave arrivals for each event. The waveforms are band-pass filtered between 0.05 and 2 Hz, and the width of the cross-correlation time window is twice the dominant period. As a result, we picked 19,429 $P$ wave arrivals from 536 teleseismic events recorded by the ANTILOPE-I array and 13,567 $P$ wave arrivals from 903 teleseismic events recorded by the Y2 array (Figure 1b).

2.2. Model Setup

Variations of velocity and RAN near the Moho discontinuity are important but hard to constrain because teleseismic rays have nearly vertical incident angles (10°–25°). In this work, the Moho geometry (or Moho depth) beneath the ANTILOPE-I and Y2 profiles are derived from the receiver-function result (Xu et al., 2017), and the CRUST1.0 model (Laske et al., 2013) is adopted for the other parts of our study area. Our initial model for the tomographic inversion has upper mantle velocities below the Moho discontinuity and crustal velocities above the Moho, and the crustal velocities are derived from the CRUST1.0 model. Therefore, in the initial model, the velocity at each grid node may be different at the same depth due to the curved Moho geometry.

We have also tested several upper mantle velocity models, including the iasp91 and ak135. The variance of all the travel-time residuals of our data set is 0.181 and 0.189 s² for the iasp91 and ak135, respectively. The optimal model is found to be the iasp91 model because it leads to the minimum travel-time residual among the three models. Comparing with the results obtained from the ak135 model, the differences of $V_p$ anomaly and RAN change from −0.09% to +0.23% and from −0.23% to +0.4%, respectively. Although some details are different, main features of the tomographic result are very similar.

Considering the main features and the results of resolution tests as shown below, we adopt a lateral grid interval of 1.5° for the 3D anisotropic velocity model, which extends from 29°N to 35°N latitude, 78.5°E to 83°E longitude, and at depths of 0–300 km (Figures 2a–2c).
Applying the RAN tomographic method developed by J. Wang & Zhao, (2013) and D. Zhao et al. (1992), we assume that the upper mantle under the study region has weak RAN with a vertical symmetry axis, and thus P wave slowness is parameterized as (Ishise et al., 2012; D. Zhao, 2015):

\[
S = S_0 (1 + M_1 \cos(2\theta)),
\]

(1)

where \(S\) is slowness in the anisotropic medium, \(S_0\) is the isotropic slowness, \(M_1\) is a RAN parameter, and \(\theta\) is the incident angle. The RAN amplitude is then defined as:

\[
\beta = \frac{V_{ph} - V_{pv}}{2V_0} = \frac{M_1}{1 - (M_1^2)},
\]

(2)

where \(V_{ph}\) and \(V_{pv}\) are the horizontal and vertical \(V_p\) values, respectively, and \(V_0\) is the isotropic \(V_p\).

Using Equation 1, we can get the following relations:

**Figure 2.** (a) Map view and (b) east-west and (c) north-south vertical cross-sections showing the 3-D grid adopted for tomographic inversions. Both the isotropic and anisotropic \(V_p\) parameters at the grid nodes (black dots) are inverted. (d) Trade-off analysis for searching the optimal damping parameter (see text for details). The optimal damping parameter (circled number) is found at the corner of the trade-off curve between the root-mean-square (RMS) travel-time residuals versus RMS \(V_p\) perturbations (in %) after the tomographic inversions.

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Using Equation 1, we can get the following relations:
where \( d \) is the length of the \( k \)th ray segment and \( V_k \) is the isotropic \( V_p \) at its middle point.

For teleseismic tomography, a relative travel time residual \( r_{mn} \) from the \( m \)th teleseismic hypocenter to the \( n \)th station can be derived from a raw travel-time residual \( R_{mn} \) by removing the mean residual \( R_m \) averaged over all stations for each teleseismic event:

\[
\begin{aligned}
\Delta \bar{R} &= R_{mn} - \bar{R}_m = \bar{R}_{\text{obs}} - \bar{R}_{\text{cal}} = \frac{1}{N} \sum_{n=1}^{N} (T_{\text{obs}} - T_{\text{cal}}), \\
&= \sum_p \left( \frac{\partial T}{\partial V_p} \Delta V_p \right) + \sum_q \left( \frac{\partial T}{\partial M_q} \Delta M_q \right) + \frac{1}{N} \sum_{n=1}^{N} \sum_p \left( \frac{\partial T}{\partial V_{pn}} \Delta V_{pn} \right) + \sum_q \left( \frac{\partial T}{\partial M_{qn}} \Delta M_{qn} \right) - \bar{R}_m.
\end{aligned}
\]  

(4)

where \( \Delta \) denotes the perturbation of a parameter, and \( N \) is the total number of stations recording the \( m \)th teleseismic event.

We adopt the damped LSQR algorithm (Paige & Saunders, 1982; Zhao et al., 1992, 2016) to solve the problem in Equation 4. To determine the optimal damping parameter \( \lambda \), we conducted many tomographic inversions with different \( \lambda \) values to construct a trade-off curve between the data fitting and the model update. Considering the balance between the model roughness and the reduction of RMS travel-time residual, we find the optimal damping parameter \( \lambda = 30.0 \) at the corner of the trade-off curve (Figure 2d). The RMS travel-time residual is reduced to 0.425 s, and it is further reduced to 0.348 s by the inversion including RAN (Figure 2d). This RMS residual (0.348 s) is greater than the expected noise level (\( \pm 0.2 \) s), which may be caused by uncertainties of the hypocentral locations of the teleseismic events as well as smaller structural heterogeneities than the resolution scale of our tomographic model. Comparing with the initial data variance, the final 3-D model reduces the data variance by \(~70\%\) after the isotropic inversion and \(~81\%\) after the anisotropic inversion.

4. Results and Resolution Analysis

4.1. Results

Our results of isotropic \( V_p \) and RAN under the western Tibetan Plateau are illustrated in map views (Figure 3) and vertical cross-sections (Figure 4). To the first order, the isotropic \( V_p \) structures are generally consistent with Razi et al. (2016). At depths of 100–250 km, there are two prominent high velocity zones (HVZs) with different RAN features. The first one is located to the south of \( \sim 31^\circ N \), and it shows positive RAN at all depths (labeled as H1 in Figures 3 and 4). The second HVZ with a layered RAN is located between 81°E−84°E and 31°N−34°N, where a positive RAN anomaly appears at shallower depths and a negative RAN anomaly exists at depths >200 km (labeled as H2 and H1’ at shallower and greater depths in Figures 3 and 4, respectively). To the west of the layered HVZ, another notable feature is the low velocity zone (LVZ) with a negative RAN at depths of 100–150 km (labeled as L1 in Figures 3 and 4) above the HVZ with a negative RAN (H1’ in Figures 3 and 4). The L1 anomaly was not revealed at 150 km depth by our previous study because the data from the Y2 array was not included and the resolution was relatively lower in that region (Zhang, J. Zhao, et al., 2016). A common discrepancy between the tomography and receiver function results is the depth of strong discontinuities because vertical smearing is an inevitable disadvantage of tomography, especially for teleseismic tomography. For example, the Indian lithosphere-asthenosphere boundary (LAB) is estimated to be \(~150–200\) km depth along the profile AA’ but our results show a thicker HVZ. Besides, because the Indian LAB exists at \(~200\) km depth, the L1 anomaly may only exist at depths \( \leq 150 \) km (Figure 4a). Hence, in the follow-
ing discussions we focus on the spatial variations of the isotropic $V_p$ and RAN anomalies rather than their exact depths.

4.2. Checkerboard Resolution Tests

Maps of ray crisscrossing suggest that the area near seismic stations can be better resolved (Figure 5). Because we use the data recorded by two seismic arrays (Figure 1a), the ray density is the highest in the central part where the isotropic $V_p$ and RAN anomalies are considered to be reliable. For example, at 100 km depth, an HVZ appears in the northernmost part of our image, and an LVZ exists in the lower right corner (Figure 3a). Only a few crisscrossing rays pass through those areas, hence the results there are not reliable.

To verify our results, we have conducted the checkerboard resolution tests. Velocity and anisotropic anomalies of 3% with alternating signs are assigned on the input model. Synthetic travel time residuals are calculated using the same process as we mentioned above. Random noise ($\pm 0.2$ s) with a standard deviation of 0.1 s is added to simulate the observation error. Then the same inversion process is applied to the synthetic travel times as to the observed data. The consistency between the input models and the inverted models on the synthetic data indicates the reliability of the inversion results on the observed data.

The results show that the velocity and radial anisotropic features in most of the study area could be resolved with a lateral resolution of 1.5° and a vertical resolution of 50 km (Figure 6).
We further evaluated the ability of our inversion to recover some particular features in the upper mantle, such as the amplitude and shape of the $V_p$ and RAN anomalies, because smearing always occurs in seismic tomography even if the ray coverage is good. Therefore, we have conducted three synthetic tests along the vertical cross-sections, which could directly illustrate the effect of smearing and whether we can recover the main features (Figure 7). In the synthetic tests, the obtained images are taken to be the input model. These synthetic tests show that most of the prominent anomalies can be recovered but some insignificant differences appear on their detailed shapes. Comparing with the input model, the amplitudes of the $V_p$ and RAN anomalies are slightly smaller because of the damping regularization applied to the inversion.

4.3. Synthetic and Trade-off Tests

We further evaluated the ability of our inversion to recover some particular features in the upper mantle, such as the amplitude and shape of the $V_p$ and RAN anomalies, because smearing always occurs in seismic tomography even if the ray coverage is good. Therefore, we have conducted three synthetic tests along the vertical cross-sections, which could directly illustrate the effect of smearing and whether we can recover the main features (Figure 7). In the synthetic tests, the obtained images are taken to be the input model. These synthetic tests show that most of the prominent anomalies can be recovered but some insignificant differences appear on their detailed shapes. Comparing with the input model, the amplitudes of the $V_p$ and RAN anomalies are slightly smaller because of the damping regularization applied to the inversion.
The crust itself is likely to be much more heterogeneous (and anisotropic) than the upper mantle, but the crustal structure is hard to constrain by teleseismic tomography. We conducted a synthetic test to assess the effect of isotropic $V_p$ structure and RAN anomalies in the crust on the upper mantle image (Figure S1). In the input model, along the profiles BB' and CC', we assume that isotropic $V_p$ and RAN anomalies exist only in the crust. The synthetic test results show that weak vertical smearing (<0.5%) occurs in the upper mantle (Figure S1). Although the crustal features do not strongly affect the images of the upper mantle, our synthetic tests show that the RAN anomalies in the crust are probably a consequence of vertical smearing, because the synthetic test results show crustal features very similar to those in the real inversion results (Figures 4 and 7).

The trade-off effect between the isotropic $V_p$ and anisotropy must be examined (e.g., D. Zhao et al., 2016). To investigate this issue, additional synthetic tests are performed. In the first two tests, isotropic $V_p$ anomalies without RAN (i.e., anisotropic parameters are all zero) are assigned to the input model and vice versa. The effect of trade-off is considered to be small because there is only weak RAN or isotropic $V_p$ in the output model (Figure S2). In the third test, we have performed an inversion without RAN anomalies, and the pattern of isotropic $V_p$ is quite similar to that obtained by the inversion including RAN (Figure S3). Because teleseismic rays have nearly vertical incident angle, both a high-V anomaly and a fast vertical axis can lead to early P wave arrivals. We conducted a synthetic test with an input model including both a high-V anomaly and negative RAN (opposite to the obtained results) to show whether a strong trade-off occurs in the tomographic model. The test results show that both the high-V anomaly and negative RAN can be well recovered (Figure S4). These synthetic test results show that the trade-off is trivial in our tomography, thanks to the good ray coverage and the inversion method used in this study. As compared with local earthquake tomography, teleseismic tomography cannot determine the absolute velocities in the upper mantle,

Figure 5. (a–d) Map views of P-wave ray paths in a 50-km depth range of each layer. The yellow and purple triangles denote seismic stations in this region.
because relative travel-time residuals are inverted (e.g., D. Zhao, 2015). Although these synthetic tests cannot demonstrate that we can fully recover absolute RAN values, they show clearly that at least the RAN sign (positive or negative, \(\pm 3\%\)) can be revealed.

4.4. Effects of Different Datasets

Because there is no nearby permanent station to be the reference of the relative travel times (the relative residuals have been derived from each array separately before the inversion) and there are only a few overlapping stations between the two arrays, bias may be introduced by simply combining the two datasets, especially in the RAN. Therefore, we conducted two separate inversions for each of the arrays. The obtained results show some differences, but the main features are quite similar (Figure S5 and S6), except for the feature to the north of 33.5°N at 150 km depth. Further comparing this feature with previous studies (e.g., Royden et al., 1997; Zhang, J. Zhao, et al., 2016), the discrepancy still exists, suggesting that such a feature may be an artifact. Hence, the following discussions are mainly focused on the velocity and RAN anomalies to the south of 33.5°N.

4.5. Effects of Model Parameterization

In general, the grid interval of a tomographic model may more or less affect the final result. To confirm the main features of our result, we shift the grid nodes by 75 km in the latitude and longitude directions, and then use the same inversion method and data set. Although the boundary between the HVZ and LVZ is slightly different at 150 km depth, most of the main features obtained with the different 3-D grids are consistent with each other (Figure S7), suggesting that the major results are not affected by the grid setting. We also performed a synthetic test to verify the bottom depth of our modeling space, and the result shows that
vertical smearing occurs at depths >300 km (Figure S8). Considering the significant Vp variations at depths <300 km and the similar features derived from the model with a bottom depth at 400 km (Figure S9), the model bottom is set at 300 km depth in our final Vp model.

5. Discussion

Although it is generally assumed that a negative (positive) RAN anomaly reflects vertical (horizontal) deformation and the lattice preferred orientation (LPO) of olivine (A-type) is the main cause of anisotropy in the upper mantle, the olivine fabric type can be changed by partial melting, pressure, water content and temperature (e.g., S. Karato et al., 2008; Savage, 1999). Especially, the notion of B- and C-type olivine causing different RAN features are incompatible with the A-type olivine. Both vertical and horizontal deformations result in positive RAN anomalies if B-type olivine is assumed, but vertical (horizontal) deformation causes positive (negative) RAN anomalies when C-type olivine is dominant, which is opposite to the A-type olivine. For more details about the olivine type and implications for seismic anisotropy, see S. Karato et al. (2008).

Previous seismic images derived from the ANTILOPE-I array data are limited by the nearly linear distribution of the portable stations (H. Zhang et al., 2016b), whereas the data set used in this study enables us to probe E-W structural variations in western Tibet. The LVZ in the three profiles (Figure 4) is consistent with the previous results, suggesting that the crustal velocity is uniformly slow in this area and there is no evidence for a LVZ in the middle crust (Razi et al., 2014). Because the RAN features in the crust are not very reliable, here we focus on the upper mantle structure. In the following, we first interpret the LVZ and
HVZs with different RAN features in the upper mantle, then we discuss the geodynamic implications of these results.

5.1. Different Deformation Patterns Within the ILM

We present an NW-SE trending vertical cross-section in Figure 3a and 3b, in which HVZs with positive and negative RAN occur in the southern and northern parts, respectively (H1 and H1’ in Figure 4). The large-scale HVZ beneath the Tibetan Plateau has been generally interpreted as the northward advancing ILM, as pointed out by previous tomographic studies (C. Li et al., 2008; H. Zhang et al., 2012; Z. W. Wang et al., 2019). However, different RAN anomalies cannot be reconciled with purely horizontal underthrusting or subvertical subduction, which would predict the HVZs with similar positive RAN anomalies. The southern part of the HVZ shows a positive RAN anomaly (H1 in Figure 3a), which can be interpreted as the northward advancing ILM. A long-term movement of ILM could affect the LPO of olivine, and the fast polarization directions are consistent with those in most of the Indian continent (Saikia et al., 2010; Wu et al., 2015). A few recent studies have also revealed seismic anisotropy in the northward advancing ILM by azimuthal anisotropy tomography (Z. W. Wang et al., 2019; Wei et al., 2016).

In contrast, the HVZ with a negative RAN feature in the northern part of the profile (H1’ in Figure 3a) reflects that the northern part of the ILM was affected by the previously attached Tethyan oceanic slab (H. Zhang et al., 2018) because the negative RAN anomaly is generally related to vertical shear or deformation. Another interpretation of the HVZ is foundering ILM. Since the continental lithosphere is much lighter than the oceanic lithosphere, the continental collision between India and Eurasia involves strong deformation and thickening of the whole continental lithosphere, resulting in a Rayleigh-Taylor instability and subsequent foundering (Pysklywec et al., 2000). Such a foundering continental lithosphere is much less negatively buoyant than the oceanic lithosphere, and so it will not fall into the lower mantle in a short time. As a strong boundary in density and velocity (Kennett et al., 1995), the 660 km discontinuity may effectively prevent the foundering continental lithosphere from sinking into a greater depth (M. Chen et al., 2017). In either case, it suggests a different deformation pattern within the ILM. Although results of deformation experiments and massive data on natural mantle rocks show that the olivine (001) axis can be aligned near vertically below 200 km depth thus resulting in horizontal deformation represented by a negative RAN (Raterron et al., 2009), it is hard to explain that the HVZ with a negative RAN only exists in the north, further suggesting that the H1’ cannot be considered as a purely underthrusting ILM. Comparing with the profile AA’, similar velocity and RAN features are detected along a NE-SW trending vertical cross-section, thus the above explanation is also applicable to the HVZs in the cross-section BB’ (H1 and H1’ in Figure 3c and 3d).

5.2. The ELM and Mantle Wedge at a Shallower Depth

Along the profile BB’, one notable feature is the HVZ with a positive RAN at a shallower depth (H2 in Figure 4c), which may not represent the subducting or foundering ILM. We think that the C-type olivine (vertical deformation causing a positive RAN anomaly) in the uppermost mantle is not a proper candidate here, because it is formed in a water-rich condition (S. Karato et al., 2008) and such a high-V anomaly at this depth is an indicator for the lack of water. Another piece of evidence is the SWB measurements along this profile (green lines in Figure 1), where ENE-WSW fast directions are observed in the north but ~ N-S fast directions appear in the south (Ju et al., 2019). A recent study of surface wave tomography also detected an HVZ with positive RAN near the profile BB’ (<200 km depth), which was attributed to the underthrusting ILM because the deeper HVZ with negative RAN (H1’) is not visible in their results (Li et al., 2020). If the HVZ above 200 km depth represents the ILM, then both similar SWB features from south to north and consistent RAN anomalies from shallow to deep areas should show up. Moreover, we cannot even attribute these different RAN features to a layered ILM (H1’ and H2 in Figure 3c), meaning that there is no mid-lithospheric discontinuity, because a clear velocity drop at different depths is not observed (S. I. Karato et al., 2015).

Therefore, the HVZ with a positive RAN (H2 in Figure 3c) should be interpreted as the ELM rather than the ILM. Based on the receiver-function results, J. M. Zhao et al., (2010) suggested that the Eurasian lithosphere overlies the Indian lithosphere along the profile BB’. Due to the inevitable vertical smearing of
seismic tomography, the LAB derived from the receiver function technique is usually shallower than that from tomographic results. Besides this area, the Eurasian LAB has also been detected in central and eastern Tibet (Kind et al., 2002; W. J. Zhao et al., 2011). Although the Tibetan extension is commonly attributed to the detachment of the ELM beginning at ∼30 Ma, some studies suggested that (partial) residue of ELM may still exist in central Tibet (Chung et al., 2005).

To the west of the ELM, an LVZ with a negative RAN is visible along an E-W trending profile (L1 in Figure 4a and 4e). Since the cold ILM (H1’ in Figure 3a) exists below the LVZ, a plausible explanation for such LVZ is mantle wedge, where low resistivities were also found (Z. J. Zhang et al., 2014). At 80°E, a relatively weak Moho signal in the northern Lhasa Block (LB) further suggests an LVZ beneath the Moho discontinuity (Z. J. Zhang et al., 2014), because the slow uppermost mantle velocity (L1 in Figure 3) is closer to the seismic velocity of the eclogitized lower crust. The distributions of intermediate-depth (70–100 km) earthquakes and large crustal earthquakes in the past ∼60 years follow a similar pattern (blue and white circles in Figures 1 and 3a, respectively): this LVZ seems to be a seismicity gap and the unusual earthquakes occurred at either side. There is evidence that the LVZ is located in the mantle wedge instead of ELM (Ju et al., 2019). Comparing with the SWB observations along the profile BB’, much stronger SWB features with ENE-WSW fast directions were revealed in the LVZ (Figure 1), indicating that the materials in that region are more prone to deform. Although our result cannot resolve the detailed structures in the crust, both low velocity anomaly and resistivity may be an indicator of water in the upper mantle, which affect the progressive process of the eclogitized lower crust (Hetenyi et al., 2007). Using receiver function techniques, the eclogitized lower crust extending to the southern LB has been detected, and a further progressed eclogitized lower crust was observed further north (Z. J. Zhang et al., 2014), suggesting the presence of fluids in the north. It is noted that the water in the lower crust can be generated either from the middle crust and/or from the mantle wedge. Due to the lack of LVZ in the middle crust along the profile AA’ (Razi et al., 2014), the origin of the water should be in the mantle wedge.

5.3. Geodynamic Implications

The E-W extension in the Tibetan Plateau is not only determined by the present-day crustal thickening and consequent gravitational potential energy (England & Houseman, 1986), which means that deeper structures should be involved. The heterogeneities of isotropic $V_p$ and RAN seem to support a geodynamic model that the upper mantle in western Tibet can be divided into three parts: the northward advancing ILM, the (partial) residue of ELM, and the mantle wedge (Figure 8). This complex structure may modify the mantle deformation pattern and further cause complex anisotropy in western Tibet. It is difficult for the ILM to
maintain its uniform deformation pattern during the long-time collision. The different deformation patterns between the northern and southern ILM may result in slab breakoff. In this case, slab tearing occurs in a direction perpendicular to the convergence, and this feature is consistent with the adakitic intrusion in southwestern Tibet during the mid-Miocene (Maheo et al., 2002). However, considering the limited resolution of the tomographic model, continuity of the HVZs and similarity of the RAN anomalies (H1 and H1' in Figures 4a and 4c), we deem that at present a large-scale slab tearing or breakoff (>150 km) does not occur in western Tibet.

On the basis of the above discussions, the LVZ above the ILM can be considered as a “mantle wedge” (L1 in Figure 4a). Results of both numerical simulation and RAN tomography have illustrated the small-scale convection occurring in the mantle wedge (Ishise et al., 2018; Liu & Zhao, 2017), in which the fast polarization directions in the forearc inferred from SWB measurements are perpendicular to the subduction direction (Huang et al., 2011), and negative RAN prevails in the mantle wedge (Liu & Zhao, 2017). The convergence of the ILM and ELM has caused eastward asthenospheric flow (Yin & Taylor, 2011), which may induce the E-W trending SWB features. Since the deformation pattern in the mantle wedge is almost horizontal if there is no small-scale convection (Ishise et al., 2018), a positive RAN should be observed, which is different from our results. Therefore, we suggest that small-scale convection occurs above the ILM (L1 and H1' in Figures 3 and 4).

A broader implication is the relationship between the upper mantle structures and the formation of rift or graben in the western Tibetan Plateau. An obvious LVZ beneath eastern Tibet supports a model that the presence of the Cona rift is likely to reflect the asthenospheric upwelling (Ren & Shen, 2008). However, recent seismic results with more combined data show that the position of LVZs in the upper mantle cannot match the location of surface rifts (J. T. Li & Song, 2018). By taking into account the previous interpretations and the locations of the Ya-Re rift and the Puran Graben, the formation of such rift and graben should not be attributed to the slab tearing or upwelling asthenosphere. Numerical modeling showed that variations of the lithospheric thickness could affect the rift formation and further facilitate the vertical deformation of continental lithosphere (Pascal et al., 2002). Considering the N-S and E-W trending variations of the lithospheric thickness (the LAB and HVZs in Figure 4), the formation of the Ya-Re rift and the Puran Graben can be related to the strong variations of lithospheric thickness.

6. Conclusions

P-wave RAN and velocity anomalies beneath the seismic arrays in the western Tibetan Plateau reflect both N-S and E-W trending structural variations. The main findings of this work are summarized as follows.

(1) Although different deformation patterns are detected within the ILM, there is no slab tearing or breakoff at present considering the continuity of the high-velocity zones and similarity of the RAN anomalies.

(2) Obvious E-W trending variations of velocity and RAN at shallow depths (<150 km) suggest the presence of both the ELM and mantle wedge.

(3) The formation of the Ya-Re Rift and Puran Graben is related to variations of the lithospheric thickness rather than slab tearing or asthenospheric upwelling.

Data Availability Statement

The data for obtaining the images in this study can be obtained from the Third Pole Environment Database (https://data.tpdc.ac.cn/en/data/a54c531f-ed6b-4e2e-a220-b80fe6993a4/). The GMT software is used to draw the figures (Wessel & Smith, 1995). Prof. Laurent Jolivet (the Editor), an Associate Editor, and two anonymous referees provided thoughtful review comments and suggestions, which have improved this paper.

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