EIKONAL TOMOGRAPHY AND AZIMUTHAL ANISOTROPY FOR SE TIBET AND SW CHINA
HUI HUANG*, YU LI†, QIYUAN LIU‡, HUAJIAN YAO§, ROBERT D. VAN DER HILST¶, AND MAARTEN V. DE HOOP∥

Abstract. We perform eikonal tomography on Rayleigh waves extracted from cross-correlations of ambient seismic noise for the SE Tibet Plateau and adjacent areas. Based on eikonal equation, the local wavespeed and propagation direction are obtained from the gradient of the traveltime. The calculation is on the perturbed traveltimes to improve accuracy in interpolation. The isotropic phase velocity and azimuthal anisotropy are studied for arbitrary location in our study region. Then a two-step inversion is performed to reveal the structures of the isotropic VSV and azimuthal anisotropy. Low velocity zones are widespread in the middle and lower crust of the high Plateau and South China block, in sharp contrast to the high wavespeed in deep crust of the Sichuan Basin. Also, strong azimuthal anisotropy (> 2%) is observed in most of the study region throughout the crust except the Sichuan Basin. Although the fast directions at different depths of the crust are all consistent with a clockwise rotation around the Eastern Himalayan Syntaxis, their angular differences are large and thus may imply different deformation patterns at surface and in depth. Furthermore, the synthetic shear-wave splitting from the crustal model in this study exhibits different directions from the observed SKS splitting, which may indicate different deformation directions and patterns in the crust and mantle. Therefore, our results may imply decoupled crust and mantle in the southeastern marginal areas of the Tibetan Plateau.

1. Introduction. The rise mechanism and coupling between the crust and mantle for the southeastern Tibetan Plateau have been debated over the last decades. A lot of modeling works, which propose distinct physical settings in lithosphere, have been conducted to account for topography variation, surface velocity field, stress field, and so on. In general, there are three schools of modeling. The first school proposes that the lithosphere of Tibet is composed of rigid blocks, which move coherently throughout the lithospheric depths [23, 33]. The second school regards the lithosphere as a thin viscous sheet with continuous deformation [9, 10, 11]. The viscous lithosphere obeys a Newtonian or a power law rheology, and the vertical gradients of the horizontal velocity are negligible. Thus, both the rheology and modeled deformation represent the average values over the entire study depth. The third school, the channel flow model, is proposed to explain the either sharp or smooth topography gradient [25]. It argues that the extruded materials from the central collision zone of the Indian and Eurasian plates flow and accumulate in a weak layer in the deep crust and thus cause the uprising the southeastern Tibet. The crust and mantle are decoupled due to the existence of the weak layer in the crust. Besides the above three schools with simple concepts, more complicated numerical modeling has been conducted [7, 6]. However, the rheology is poorly constrained; even if the deformation at the surface is well modeled its patterns in deep lithosphere can differ greatly with different settings of rheology. Progress has been made to address different aspects of the lithosphere of southeastern Tibet and surrounding areas; controversy and dispute, however, still exist. On the one hand, the widespread low velocity zones [41, 44], high Poisson’s ratio [39], low electrical resistivity [1], and strong radial anisotropy [14] all suggest weak middle or lower crust and thus decoupled lithosphere. In addition, azimuthal anisotropy reveals different fast directions in the crust and mantle, indicating possible difference in deformation [43]. On the other hand, based on the direction consistence between the maximum strain-rate (either extension

*Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge MA 02139, USA
†State Key Laboratory of Earthquake Dynamics, Institute of Geology, China Earthquake Administration, Beijing 100871, China
‡State Key Laboratory of Earthquake Dynamics, Institute of Geology, China Earthquake Administration, Beijing 100871, China
§Department of Geophysics and Planetary Sciences, University of Science and Technology of China, Hefei 230026, China
¶Department of Earth, Atmospheric, and Planetary Sciences, University of Science and Technology of China, Hefei 230026, China
∥Department of Mathematics, Purdue University, West Lafayette IN 47907, USA

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or shear) from GPS observations and SKS splitting, Sol et al. [32] and Wang et al. [34] argue for a coupled lithosphere and thus a relative strong crust for the SE Tibet. However, this directional consistence does not necessarily mean coherent lithospheric deformations. The velocity field from GPS only reflects the current deformation patterns at surface. In contrast, SKS splitting measures the integral effect of the anisotropy in the entire mantle and crust and reflect of both current and previous deformation. Hence, the deformation in the lithosphere may vary at different depths even there is some consistence between surface strain-rate and SKS splitting. Therefore, high-resolution wavespeed and anisotropy studies would be helpful to better understand the structures and deformations for the southeastern Tibetan Plateau. In this paper, we use the eikonal equation [38, 29] to extract the local wavespeed and its variation with propagation direction, i.e. azimuthal anisotropy for Rayleigh waves. The Green’s functions for Rayleigh waves are obtained from the cross-correlation functions of the ambient seismic noise. In eikonal tomography, one estimates the local velocity and propagation direction simultaneously from the spatial derivatives of the travelt ime function (or wave fronts) [17, 12]. We focus here on recovering information about azimuthal anisotropy. We examine the variation of azimuthal anisotropy at different depths and compare the results with the surface strain-rate and SKS splitting to improve our understanding of the deformation of the southeastern Tibetan Plateau.

2. Eikonal tomography.

2.1. Ambient noise data and Rayleigh wave Green’s function. We use ambient seismic noise data from a temporary array containing 298 broadband seismometers in western Sichuan Province, China (Figure 1). Only the records of the year 2007 are used because they are sufficient to extract stable surface wave Green’s functions [41, 2]. Following our previous studies [42, 13], Rayleigh wave Green’s functions between two stations are extracted from noise cross-correlation functions (CFs) of the vertical components of the data. In this study a CF is retained for further analysis if (1) its effective stacking length is larger than three months and (2) its signal-to-noise ratio (SNR) [2, 13] is larger than five. Figure 2 is an example of all the extracted CFs connecting a virtual “source” station KCD10 and other stations as “receivers”. Phase velocity dispersion curves in a period band 6-40 s are then measured from the Green’s functions via time-frequency analysis [42]. There are more than 5000 measurements for most of the periods considered (Figure 3), which ensures a sufficiently accurate estimation of the wave front to perform eikonal tomography.

2.2. Eikonal equation. Eikonal tomography is intimately connected to the application of semi-classical analysis to the boundary value problem generating surface waves. The stiffness tensor is assumed to vary slowly in the directions tangential to the surface and more rapidly in the direction normal to it; this is articulated by introducing a small scaling parameter $\epsilon$. We let $k$ denote the surface wave vector, with norm, $|k|$, and direction, $\hat{k}$. Effectively, $\epsilon|k|^{-1}$ becomes the semi-classical parameter. Following an expansion in the semi-classical parameter, the wave solutions can be written in the form of a mode summation; each term of the summation factors into the solution of a surface wave equation and an eigenfunction associated with a one-dimensional spectral problem. The eigenvalue corresponding with the eigenfunction defines an eigenfrequency, which is a function of the surface position and $k$ in general. The eigenfrequency implicitly defines a phase velocity, $c = c(r, \hat{k}, \omega)$. Both the eigenvalue and eigenfunction are symbols of pseudodifferential operators, and are represented by expansions in $\epsilon$ as before. The symbol of the spatial part of the surface wave operator coincides with the mentioned eigenvalue. One typically assumes that $\epsilon$ approaches 0.

Following the parametrix construction of the surface wave equation, we find a traveltime $t(r, r_s, \omega)$ which satisfies

\begin{equation}
|\nabla t(r, r_s, \omega)| = \frac{1}{c(r, \hat{k}, \omega)}
\end{equation}

and

\begin{equation}
\nabla t(r, r_s, \omega) = \frac{1}{c(r, \hat{k}, \omega)} \hat{k}
\end{equation}
Fig. 1. Topography, geological units and seismic stations used in this study. The topography is represented by the background color. The locations of seismic stations are depicted as blue triangles. Our study region can be approximately divided into six geological units according to GPS observations [30]: the Longmenshan block, Yajiang block, Shangrila block, Central Yunnan block, Sichuan Basin, and the South China block. Their boundaries are depicted as thick white dash lines. Major faults are depicted as thin black lines and the abbreviations are: Longriba fault (LRBF), Xianshuihe fault (XSHF), Longmenshan fault (LMSF), Longquan fault (LQF), Litang fault (LTF), Chenzhi fault (CZF), Lijiang fault (LJF), Muli fault (MLF), Anninghe fault (ANHF), Zemuhe fault (ZMHF), Xiaojiang fault (XJF), Shimian fault (SMF), and Luzhijiang fault (LZJF) (after [36, 35, 30]. The inset shows the location of our study region, as the red box, in the context of the whole Tibetan Plateau and its surrounding areas. The white arrows depict the approximate surface motion relative to the Yangtze Craton from GPS observations [45, 30].

Fig. 2. (a) All the raypaths related to the virtual source station KCD10. The station KCD10 is shown as a red triangle, and other stations are shown as blue triangles. (b) The cross-correlation functions (CFs) corresponding to the raypaths in (a). The CFs are filtered between period band 8 - 30 s, and only those with signal-to-noise ratio larger than 10 are plotted.
if \( \hat{k} \) coincides with the phase direction at \( r \) of the bicharacteristic connecting \( r \) with \( r_s \). The amplitude in the parametrix, through an expansion, satisfies frequency-dependent transport equations. Consistent with the assumptions mentioned above, one expects that the amplitude varies slowly and smoothly along the wave front [17, 12]. Tracking the progression of the surface wave in our study, we find that its wave fronts deflects only slightly from circles indeed (Figure 4). We note that the estimation of the amplitude of the Green’s function from ambient noise interferometry has limited reliability [37]. However, its phase can be well recovered. From the phase one straightforwardly obtains the traveltime. Eikonal tomography for the recovery of phase velocity is based on the principle that with a relatively dense set of stations (and hence “sources”) one can determine \( \nabla t(r, r_s, \omega) \) directly, albeit approximately, as long as \( r \) is sufficiently far away from \( r_s \). Here, we follow an alternative, indirect approach.

2.3. Minimum curvature interpolation. We rely on an accurate estimation of traveltimes in a target region. As measurements are only available at station locations, interpolation is necessary to get traveltimes and their gradients on a grid. Subject to our assumptions, the traveltime and its gradient will be smooth. In this study, we use a minimum curvature method, which minimizes the total squared curvature of the interpolation, to approximate the traveltime function [4, 28]:

\[
(2.3) \quad t(r, \ldots) = \sum_{j=1}^{N} \alpha_j(.) \phi_j(r)
\]

The traveltime function is the linear combination of bi-harmonic basis functions. Following [28], the jth basis function \( \phi_j \) is defined on the jth data point \( r_j \):

\[
(2.4) \quad \phi_j(r) = |r - r_j|^2 \log |r - r_j| - 1.
\]

Clearly, there are as many basis functions as data points. The coefficients \( \alpha_j \) are determined from the system of linear equations:

\[
(2.5) \quad t(r_i, \ldots) = \sum_{j=1}^{N} \alpha_j(.) \phi_j(r_i) \quad i = 1, \ldots, N,
\]

where \( r_i \) is the location of the \( i^{th} \) data point. By Eq.(2.5) we require that the interpolated traveltime has values equal to the measured traveltimes at the data points. The gradient of the traveltime,
i.e. the local wave slowness and propagation direction, is calculated from the gradient of the basis functions:

\[
\nabla \phi_j (r) = (r - r_j) [2 \log |r - r_j| - 1].
\]

The direct application of the above interpolation scheme will fail near the virtual source station and the edges of the array. In a smooth medium, the asymptotic expression of the traveltime is in proportion to the travel distance \(|r - r_s|\). Its gradient has a singular point at source location \(r_s\), and thus cannot be correctly estimated using the continuous and smooth basis functions in Eq. (2.6). For this reason, Lin et al. [17] remove the region within two wavelengths of the source station. In this study, this source region elimination cannot be afforded due to the limited size of the region (about 5\(^\circ\) \times 6\(^\circ\)), especially at longer periods. For example, the 30 s Rayleigh wave has a wavelength of about 100 km or 1\(^\circ\). The removal of an area with a radius of two wavelengths would exclude nearly the entire study region. At large distances away from the \(r_j\), the basis functions grow more rapidly than a typical traveltime function. Hence, the interpolation accuracy reduces significantly near the edges of the array.

We follow an alternative method in which we interpolate only the perturbed traveltime. The total traveltime is then decomposed into two parts:

\[
\begin{align*}
t(r, r_s, \omega) &= t_0(r, r_s, \omega) + t_1(r, r_s, \omega), \\
t_0(r, r_s, \omega) &= \frac{|r - r_s|}{c_0(\omega)},
\end{align*}
\]

where \(t_0(r, r_s, \omega)\) is traveltime evaluated in a reference medium, which is, here, homogeneous and isotropic with phase velocity \(c_0(\omega)\), and \(t_1(r, r_s, \omega)\) is the perturbed traveltime which reflects the wavespeed perturbations and anisotropy. For simplicity, we take the reference wavespeed, \(c_0(\omega)\), to be the average of all the raw measurements; hence, it can differ from the average obtained from the final tomography result. Accordingly, the gradient of traveltime, or the slowness in wave propagation
Fig. 5. Perturbed traveltime from the virtual source station KCD10 for 10 s period Rayleigh wave. The background traveltime based on a homogeneous wavespeed model has been removed. This figure shows the procedure of data processing and signal enhancing. The black circles are centered at the source station with 50 km radius. (a) The interpolation is carried on the total traveltime, and then the reference traveltime field is removed. (b) The reference traveltime has been removed before the interpolation is performed. (c) The same as (b) but a damping term is used in the interpolation. (d) A Gaussian filter is applied to smooth the traveltime in (c).

direction, is also decomposed into two terms:

\[
\frac{1}{c(r, k, \omega)} \hat{k} = \nabla t_0(r, r_s, \omega) + \nabla t_1(r, r_s, \omega) = \frac{1}{c_0(\omega)} \frac{r - r_s}{|r - r_s|} + \sum_{j=1}^{N} \alpha_j(\omega)(r - r_j)[2 \log |r - r_j| - 1],
\]

keeping \( r_s \) fixed. As an example, the estimated traveltime surfaces and wavespeed maps associated with the rays emanating from the virtual source KCD10 are illustrated in Figures 4 to 6. The circle-like contours of the total traveltime in Figure 4 reveal the wave front propagating away from the source along the surface of the earth. Their deviations from circles reflect local wavespeed perturbations. We observe a great improvement in the wavespeed calculation near the source station and the edges of the study region if only the perturbed traveltime is involved in the interpolation (Figures 5a, b and 6a, b) while effectively removing the cone near the source from the procedure. Noise in the data and uncertainty will be addressed in the following subsections.

2.4. Damping and smoothing in interpolation. As mentioned above, we have as many basis functions as data measurements. Hence, if the measured traveltime data are not distorted, the above system of linear equations (2.5) has a unique solution for the unknown coefficients, and the obtained traveltime surface passes the data points exactly. However, this exact fitting may cause local oscillations and a large overshoot when there is noise in the data [28, 17, 12]. This effect is
Fig. 6. Phase velocity maps calculated from the traveltime in Figure 5. Locations with poor data coverage have been removed in (d). The improvement of calculation performance is clear from (a) to (d).

not so apparent in the traveltime (Figure 5b) but significant in the phase velocity maps because of amplification of the oscillations upon taking the spatial derivatives (Figure 6b).

Therefore, we construct a surface which only approximately fits the data points instead. Sophisticated algorithms have been and can be designed for this purpose, for example, using a subset of the full set of basis functions or adding extra terms to minimize oscillations in the inversion of the system in Eq.(2.5). In this study, we apply a damping term. The large oscillations, which in nature are coefficients in front of some basis functions, are suppressed (Figure 6c). Even using data from only one source station, structures that are consistent with known geology begin to emerge (Figure 10). Relatively high phase velocities are observed in the interiors of the sub-blocks, and low anomalies are observed on some faults. Furthermore, given that surface waves typically resolve structures comparable to or larger than their wavelengths, we apply a 2-D Gaussian filter to smooth the obtained phase velocities. This procedure plays a role similar to the model regularization in traditional tomography with a corresponding correlation length [43, 13]. When applied alternatingly with the damping, one obtains a strategy which is consistent with the eikonal equation to some degree of accuracy. To some extent, the selection of the standard deviation of the Gaussian function is subjective. Computational experiments indicate that an appropriate choice is 1/4 of the reference wavelength ($c_0 T/4$) with a minimum value of one grid size (here, 0.125°, or about 14 km). The smoothing does not change the pattern or distribution of the anomalies, but it reduces their magnitude and increases their wavelengths (Figures 5d and 6d). We note that one could apply a procedure of the type above to the traveltimes corresponding with all pairs ($r, r_s$) combined.

Finally, we perform a quality control for the resulting phase velocity maps. We only retain
regions with good data coverage [17]. First, a circle centered at the source with radius of two wavelengths or 100 km, whichever is smaller, is removed. Second, the location is only retained if there are data in at least three of the four quadrants of the E-W and N-S axes within 150 km distance (Figure 6d).


3.1. Azimuthal anisotropy at a fixed location. The algorithm in Section 2 is applied to the wave fronts (traveltimes) of all virtual sources. For each location in the study region we have many measurements of phase velocity and corresponding propagation directions (Figure 7). Although the scattering in these measurements is quite large, we observe a cosine variation with a 180° periodicity. To stabilize the analysis of anisotropy we first calculate the mean and standard deviation of the phase velocity in chosen angle bins of 20° width:

$$c(\mathbf{r}, \hat{k}, \omega) = \frac{1}{n} \sum_{\kappa=1}^{n} c_{\kappa}(\mathbf{r}, \omega),$$

$$c_{\kappa}(\mathbf{r}, \omega) = c(\mathbf{r}, \hat{k}_{\kappa}, \omega)$$

for $\mathbf{r}$ fixed, where $\kappa$ labels angle and $n$ signifies the total number of phase velocities in a specific angle bin. To limit the angular range of a bin, and the number of measurements in a bin, we set the minimum value for the mean of the standard deviation to 1%.

In a transversely isotropic medium with a horizontal axis of symmetry (HTI), the phase velocity of the Rayleigh wave attains the form [31]:

$$c(\psi, \omega) = c_{iso}(\omega) + a_1(\omega) \cos(2\psi) + a_2(\omega) \sin(2\psi),$$

$$c_{iso}(\omega) = c_0(\omega) + a_0(\omega)$$

(replacing $\hat{k}$ by the azimuth $\psi$ of the wave vector in our notation) where $c_0$ is the isotropic constant reference phase velocity as before and $a_0$ and $a_1, a_2$ are the isotropic phase velocity perturbation and azimuthal anisotropy coefficients, respectively. This expression was obtained in the planarly layered case, but is here applied locally following the semi-classical analysis assumptions. These coefficients can be determined from weighted least-square fitting using the angular bins introduced above. The fast direction, $\psi_0$, say, and amplitude of anisotropy, $A$, are then calculated from $\psi_0 = \frac{1}{2} \arctan(a_2/a_1)$ and $A = \sqrt{a_1^2 + a_2^2}$ for a given frequency.

3.2. Resolution test. The limited resolution of the isotropic velocity may result in artificial anisotropy (Appendix A). To develop an insight into this effect, we perform separate checkerboard tests for isotropic and anisotropic phase velocities. For the isotropic model, the smallest resolvable anomaly is about 0.75°, at both short and long periods (Figure 8). The anisotropic tests are carried out on a model composed of a realistic isotropic phase velocity model (Figure 19a) from our previous study [13] and a checkerboard anisotropy. These also reveal a resolution of about 0.75° (Figure 9). The resolution of surface wave tomography is primarily determined by two factors: The effective wavelength and the data density or inter-station distance. We note that, in view of the underlying assumptions, the resolution cannot be arbitrarily increased by decreasing the station spacing. The influence of wavelength is reflected in the specification of our Gaussian filter. The standard deviation of the Gaussian function is about 14 and 25 km for the 10 and 30 s Rayleigh waves, respectively, both of which are much smaller than the 0.75° (about 83 km) resolution. Our tests thus suggest that the resolution is primarily determined by data density in this study.

There are about 18,000 rays for both the 10 s and 30 s Rayleigh waves (Figure 3). The average ray count for each source is about 150. For the 5° × 6° study region the average inter-station distance is about 50 km or 0.5°. Our checkerboard experiments show that at least three data points are required in one anomaly cycle (twice the single anomaly) to obtain satisfactory results, exceeding
Fig. 7. The variation of phase velocity against propagation direction at period (a) 10 s, and (b) 30 s for location (102°, 29°). The blue stars represent the results from different sources; the red stars with error bars are the average values and standard deviations of the wavespeed in 20° angle bins. The thick red lines are the 2ψ anisotropy fitting results for the averages values.

Fig. 8. Checkerboard test for isotropic phase velocity: (a) 0.75° × 0.75° checkerboard model; (b) recovered model using 10 s Rayleigh wave data, and (c) recovered model using 30 s Rayleigh wave data.

the Nyquist sampling rate. We find that the resolution in our eikonal tomograph at a 10 s period is about 0.75°, which is lower than the 0.5° resolution in the corresponding traditional tomography [13]. The resolution at a point in traditional tomography is essentially determined by the number of available ray directions at that point. This criterion is naturally different from the one for eikonal tomography which is determined by interpolation. However, eikonal tomography has the ability to resolve anisotropy, which, unless the anisotropy is weak, traditional tomography has not.
Fig. 9. Checkerboard test for azimuthal anisotropy: (a) 0.75° × 0.75° checkerboard model; (b) recovered model using 10 s Rayleigh wave data, and (c) recovered model using 30 s Rayleigh wave data. The black bars describe both the fast direction and amplitude of the anisotropy.

Fig. 10. Eikonal tomography results for the 10 s Rayleigh wave: (a) The isotropic phase velocity; (b) standard error for isotropic phase velocity; (c) azimuthal anisotropy; (d) standard error for the amplitude of azimuthal anisotropy.

3.3. Anisotropy map. The isotropic phase velocity reveals very similar patterns as the ones obtained in our previous study [13]. At short periods (Figure 10a), the Sichuan Basin exhibits a strong low anomaly between the Longmenshan and Longquan faults, representative of the thick sediments in the foreland basin [19, 24]. Low anomalies are also found in the Shangrila block, reflecting felsic and volcanic rock of the Yidun unit [35, 40]. In contrast, relatively high anomalies are found in the interiors of the Longmenshan, Yajiang, and Central Yunnan blocks. At longer
periods (Figure 11a), the study region can be approximately divided into three anomalous sub-regions. The high phase velocity in the Sichuan Basin indicates a cold and rigid basin root. The high Plateau, including the Longmenshan, Yajian, and Shangrila blocks, shows a strong low phase velocity which, we suggest, reflects a relative higher temperature in the middle and lower crust. The Central Yunnan and South China blocks have normal phase velocities.

The azimuthal anisotropy exhibits patterns consistent with known geology and results of a previous study [43]. At short periods (Figure 10c), the fast direction is NW-SE in the Tibetan Plateau and turns to N-S in the Central Yunnan block, consistent with the clockwise rotation of the surface movement from GPS observations [30]. The fast direction is NE-SW near the Longmenshan fault, parallel to the fault strike. At long periods (Figure 11c), although the fast directions of the anisotropy are also consistent with a clockwise rotation in the high Plateau and Central Yunnan block, they exhibit larger angular differences than those at the short periods. For example, the fast directions are more parallel to fault strike near the Xianshuihe fault, and they are NE-SE near the Lijiang-Muli fault line, different from the nearly N-S direction at short periods. The anisotropy is relatively weak in the Sichuan Basin and the South China block at both short and long periods. The isotropic perturbation and amplitude of anisotropy have posterior errors of about or less than 0.5% in the interior of the study region, which are small compared to the magnitude of the perturbations themselves. We conclude, therefore, that the phase velocity anomalies and anisotropy patterns are robust.

4. Recovery of shear wavespeed and anisotropy.

4.1. NA based inversion for an HTI model. Following [21, Eq.(4)], the first-order Rayleigh wave phase velocity (spectral) perturbation in a weakly azimuthally anisotropic medium can be
expressed as

\[
(4.1) \quad \delta c(., \psi, \omega) = \int_0^H \left[ D_{Aciso}(\omega)(\delta A(., z) + Bc(., z) \cos(2\psi) + Bs(., z) \sin(2\psi)) \right.
\]
\[
+ D_{Lciso}(\omega)(\delta L(., z) + Gc(., z) \cos(2\psi) +Gs(., z) \sin(2\psi)) \right] \frac{dz}{\Delta h};
\]

here \( D_{A,Lciso} \) is expressed in terms of the components of the corresponding unperturbed eigenfunction. Furthermore, \( H \) signifies the maximum depth under consideration and \( \Delta h \) is a normalization thickness for the calculation of the sensitivity kernels. The six parameters, \( A, L, B_{s,c}, \) and \( G_{s,c} \) are linear combinations of the elastic stiffness tensor components. They determine the quasi compressional velocity \( v_P \) and quasi vertically polarized shear velocity \( v_{SV} \) [8, 20]:

\[
(4.2) \quad v_P(., z, \psi) \approx \sqrt{\frac{A(., z) + Bc(., z) \cos(2\psi) + Bs(., z) \sin(2\psi)}{\rho(., z)}},
\]

\[
(4.3) \quad v_{SV}(., z, \psi) \approx \sqrt{\frac{L(., z) + Gc(., z) \cos(2\psi) +Gs(., z) \sin(2\psi)}{\rho(., z)}},
\]

where \( \rho \) is the density. Comparing Eq.(3.1) with Eq.(4.1) allows us to separate the isotropic and anisotropic terms:

\[
(4.4) \quad a_0(., \omega) = \int_0^H [D_{Acd}(\omega)\delta A(., z) + D_{Lcd}(\omega)\delta L(., z)] \frac{dz}{\Delta h}
\]

and

\[
(4.5) \quad a_{1,2}(., \omega) = \int_0^H [D_{Acd}(\omega)B_{c,s}(., z) + D_{Lcd}(\omega)G_{c,s}(., z)] \frac{dz}{\Delta h}.
\]

We perform a two-step Neighborhood Algorithm (NA) to estimate these parameters [26, 27]. The first step is to estimate the isotropic \( v_{SV} \) (i.e., \( L \)) using the isotropic part of the Rayleigh wave dispersion data, i.e. \( c_0(\omega) + a_0(., \omega) \). The isotropic \( v_P \) (i.e., \( A \)) is coupled to \( v_{SV} \) using Poisson’s ratios from a receiver function study [39]; the density \( \rho \) is calculated using an empirical relationship [5]. The crust is divided equally into three layers for the entire study region, and the upper crust of the Sichuan Basin is further equally divided into two layers to better account for the velocity changes of the shallow crust due to the presence of thick sediments. We calculate sensitivity kernels which are used in the second step in the NA based inversion for \( G_{c,s} \) and \( B_{c,s} \). As for estimating \( A \) in the first step of the (isotropic) inversion, we couple the \( B_{c,s} \) to \( G_{c,s} \) according to:

\[
(4.6) \quad \frac{B_{c,s}}{A} = \gamma \frac{G_{c,s}}{L}.
\]

If we assume that the relative magnitudes of the anisotropy are the same for \( v_P \) and \( v_{SV} \), the coefficient \( \gamma \) will be equal to 1. After the two-step inversion, we obtain both the isotropic and anisotropic contributions to the shear wave speed:

\[
(4.7) \quad v_{SV}(., z, \psi) \approx \sqrt{\frac{L(., z)}{\rho(., z)} \left( 1 + \frac{Gc(., z) \cos(2\psi)}{2L(., z)} + \frac{Gs(., z) \sin(2\psi)}{2L(., z)} \right)}.
\]

The fast direction and relative magnitude (with respect to the isotropic wavespeed) are \( \psi_0 = \frac{1}{2} \arctan(G_s/G_c) \) and \( A_{SV} = \sqrt{(G_c/2L)^2 + (G_s/2L)^2} \), respectively.
4.2. The 3-D heterogeneity and azimuthal anisotropy. Shear-wave speed $V_{SV}$ and its azimuthal anisotropy are shown in Figure 12. The depths of 5 and 10 km are representative of the upper crust. So $V_{SV}$ and anisotropy are only different in the Sichuan Basin. The depths of 25 and 40 km are representative of the middle and lower crust, respectively. The main features of both the $V_{SV}$ and anisotropy maps were clearly identifiable in the phase velocity maps (Figures 10 and 11) due to the period-depth correspondence of the surface wave sensitivity kernels [22]. Since the structures revealed in the isotropic $V_{SV}$ are very similar to the results in our previous study and have been fully discussed there [13], in this paper, we only introduce some large features and do not go into details of their geological implications. For the Sichuan Basin, the thick sediments and old stable basin root are well represented by the significant low $V_{SV}$ in the upper crust and high $V_{SV}$ in the middle and lower crust. Its $V_{SV}$ gradient in the depth direction is the largest among all the sub-blocks.

For the three blocks in the high Plateau, i.e. the Longmenshan, Yajiang, and Shangrila blocks, in the upper crust, we observe relatively low $V_{SV}$ on major faults, in compared to the interior of the blocks. This may be indication of damage zones on the Xianshuihe, Litang, and Chenzhi faults. In the middle and lower crust, all three blocks exhibit very low $V_{SV}$ and the most dominant low anomalies are within the Yajiang and Shangrila blocks. This feature suggests higher temperature and possible larger deformation to the south of the Xianshuihe fault than to the north, consistent with Liu et al. (2003). In general, throughout the crust the Central Yunnan block always has higher $V_{SV}$ than its surrounding sub-blocks, which may be evidence of the intruded material from the mantle [18]. The South China block also reveals low $V_{SV}$ in the middle and lower crust,
which suggests relative high temperature and weakened rock mechanics. The posterior errors for the isotropic $V_{SV}$ are about 0.5% (about 0.02 km/s) in the upper crust and about 1% (about 0.04 km/s) in the middle and lower crust (Figure 13); these values are small compared to the magnitudes of the revealed wavespeed anomalies (which can reach up to 0.3 km/s). Therefore, the structures revealed in the $V_{SV}$ anomalies are reliable. Azimuthal anisotropy is prominent in our study region. In the upper crust (Figure 12a, b), NW-SE fast directions are observed in the northwestern part of the Longmenshan and Yajiang blocks, and they gradually change to N-S in the Central Yunnan block. This pattern of anisotropy resembles the clockwise rotation of the surface velocity field from GPS observations [30]. Near major faults, e.g. the Longmenshan, Xianshuihe, Lijiang, and Shimian faults, the fast directions are parallel to the strikes of the faults, which may be reflection of cracks or fractures in the damage zones of these faults. Anisotropy is relatively weak in the interior of the Sichuan Basin and South China block. The large anisotropies near the eastern edge of our study region are artifacts due to the combined effects of incomplete angular coverage and recovery of isotropic wavespeed (Figure 19).

In the middle and lower crust (Figure 12c, d), the change in orientation of the fast directions from NW-SW in the Longmenshan and Yajiang blocks to N-S in the Central Yunnan block is also apparent. However, angular differences are obvious when compared with the upper crust. The fast directions are more parallel to the strike of the Xianshuihe faults in the Longmenshan and Yajiang blocks; they are more aligned in the NE-SW direction than N-S near the Lijiang-Muli fault zone; no obvious anisotropy parallel to fault strike is found near the Sinian fault. Anisotropy is relatively small in the Sichuan Basin and South China block as in the upper crust. These results are generally consistent with a previous study of azimuthal anisotropy for a larger region [43]. The fast directions in the upper crust are nearly the same in the areas where our studies overlap. But the fast directions

FIG. 13. Standard errors (in percentage) for $V_{SV}$. 

in the lower crust observed here are closer to their results in the uppermost mantle. In this study we use many more measurements, so we expect that both the isotropic wavespeed and anisotropy are better constrained. The fast directions change more gradually in depth in this study.

The posterior errors of the amplitude of the anisotropy are about 0.3% in the upper crust and 0.5% in the middle and lower crust in the interior of our study region (Figure 14). These values are small compared to the revealed anisotropy (about 3%), so our results about the anisotropy are significant. Near the edges, especially the northern and southern margins, the uncertainties can reach up to 2% due to limited angular coverage and incomplete recovery of the isotropic $V_{SV}$ (Figure 19). Therefore, we do not interpret the changes in magnitude and directions of the anisotropy in the marginal areas. The fast direction of the anisotropy is better determined than its magnitude. The average posterior error is about 5° for our study region (Figure 16). The errors are large and reach 20° only at locations where the anisotropy is small and thus shows no clear preferred fast direction.

5. Discussion.

5.1. Phase velocity maps from eikonal and traditional tomography. Since NA inversion schemes and parameter settings for the isotropic $V_{SV}$ are the same in this research and our previous traditional tomography study [13], the differences in isotropic $V_{SV}$ between the two studies are reflection of their differences in phase velocities. In general, eikonal tomography generates smoother phase velocity maps, especially at short period. We now investigate possible reason. The resolution of surface wave tomography is determined by both data density and wavelength. In our surface wave tomography using a traditional method, the correlation length thus the anomaly size determined from data density is always smaller than one wavelength; therefore, the smallest anomaly that can be resolved is of size of about one wavelength. However, eikonal tomography only
Fig. 15. Angular differences of the azimuthal anisotropy between (a) the upper and middle crust; and (b) the upper and lower crust. The blue bars are for the Sichuan Basin; the green bars are for the Longmenshan, Yajiang, and Shangrila blocks; and the red bars are for the Central Yunnan and South China blocks.

Fig. 16. Comparison of the delay times measured from SKS splitting (lev2006seismic, Sol et al., 2007, wang2008evidence) and predicted from the anisotropic crustal model in this study. The measured and predicted delay times and fast directions are shown as red and black bars, respectively.

uses rays from one virtual source at a time; thus the effective data density is much smaller than that in traditional tomography. As we have shown in section 3.2, the smallest resolvable anomaly size is about 0.75° (83 km) for both short and long periods, which is much larger than the wavelength of short period surface waves (i.e. about 30 km for 10 s Rayleigh wave) and more or less the same as that of long period surface waves (i.e. about 100 km for 30 s Rayleigh wave). We also need to point out that it is due to the consideration of resolution that we use different data sets in these two studies. In the previous research, we retain Green’s functions with SNR larger than 10, and there are more than 5000 measurements for almost all the periods between 4-40 s. These numbers are
sufficient and increasing measurements would not improve the resolution in traditional tomography. In this study, we use a smaller SNR of five in Green’s function selection in order to include more measurements to improve resolution. This reduction of SNR results in about 50% more measurements. If we use the same SNR (i.e., 10) as in the previous study, we would expect smoother phase velocity variations and larger differences between results of eikonal and traditional tomography.

5.2. Azimuthal and radial anisotropy. Azimuthal and radial anisotropy reflect different aspects of the angular variation of wavespeed. Azimuthal anisotropy, or transverse isotropy with a horizontal symmetric axis, measures the variation of wavespeed with polarization direction in the horizontal plane. Radial anisotropy, or transverse isotropy with a vertical symmetric axis, measures the difference between the average wavespeed on the horizontal plane and in the vertical direction. Therefore, a thorough study of the two kinds of anisotropy is necessary to understand the structure and deformation pattern better.

In our previous research, we observed ubiquitous low velocity zones (LVZs) and positive radial anisotropy ($V_{SH} > V_{SV}$) in the middle and lower crust; and we also observe a clear negative correlation between the magnitude of the low speed anomaly and radial anisotropy. These observations lead to three arguments: first, low wave speed in the middle and lower crust indicates relatively higher temperature and thus weakened rock mechanics in our study region; second, positive radial anisotropy reveals the dominant deformation is on the horizontal plane; third, the correlation between LVZs and radial anisotropy confirms that larger deformation occurs in more mechanically weakened regions. In this study, we do not observe a simple correlation between wavespeed anomaly and azimuthal anisotropy (Figure 12). Azimuthal anisotropy in the middle and lower crust is not dominant over that in the upper crust. In fact, azimuthal anisotropy seems to have a little larger amplitude in the lower crust than in the upper crust; but the difference is probably not significant. The correlation between the magnitudes of $V_{SV}$ anomaly and azimuthal anisotropy is not clear. For example, the Yajiang and Shangrila blocks, where the most prominent low $V_{SV}$ anomaly resides in the middle and lower crust, do not exhibit larger anisotropy than the adjacent Longmenshan and Central Yunnan block. This lack of clear correlation between $V_{SV}$ and anisotropy was also found in Yao et al. [43]. The different behaviors of the two anisotropies may have different reasons.

Experiments show that both wavespeed and azimuthal anisotropy cannot be resolved well near the boundaries of anomaly bodies with opposite properties (opposite sign of anomaly for wavespeed and perpendicular fast directions for azimuthal anisotropy) (Figures 8 and 9). The radial anisotropy is calculated from the two speeds $V_{SV}$ and $V_{SH}$, so it has the same patterns of incomplete recovering or smearing as the two wavespeeds. But the calculation of azimuthal anisotropy is independent of wavespeeds and thus can have different local resolutions. For example, the middle and lower crust of the Yajiang and Shangrila blocks is dominated by low wavespeed anomaly and large positive radial anisotropy. In the middle of this low wavespeed zone near the Chenzhi fault, the azimuthal anisotropy changes its fast direction from NW-SE in the north to N-S in the south and only small azimuthal anisotropy is observed. We believe that the small amplitude of azimuthal anisotropy does not necessarily mean little or no anisotropy but rather limited resolution between two areas with perpendicular fast directions. Therefore, it is easier to observe a magnitude correlation between radial anisotropy and $V_{SV}$ anomaly than between azimuthal anisotropy and $V_{SV}$.

Although the amplitude correlation between azimuthal anisotropy and a $V_{SV}$ anomaly is not obvious, we can analyze the deformation variations in the crust through comparing the fast directions at different depths (Figure 15). Between the upper and middle crust, the angular differences of the fast directions are very small in all the sub-blocks except the Sichuan Basin. The angular differences for the Sichuan Basin are distributed nearly homogeneously over the $0° - 90°$ range. This is not surprising considering the small anisotropy and large angular uncertainty in the Basin. Between the upper and lower crust, the angular differences are larger in general. For example, near the Xianshuihe fault the fast directions are more aligned parallel to the fault strike in the middle and lower crust; and in the Central Yunnan block the near N-S directions in the upper crust change to NE-SW in the lower crust (Figure 12). Therefore, the deformation in the upper and middle crust
may be correlated; but the correlation between the upper and lower crust is weaker. Combining the radial and azimuthal anisotropies, we have a more comprehensive picture of the deformation in the SE Tibet and its adjacent regions. Our results show that both the radial and azimuthal anisotropies vary in different layers of the crust, which suggests different deformation patterns in these layers. The largest deformation occurs in the middle and lower crust and mainly on near horizontal planes, revealing positive radial anisotropy in most of our study region. The curvilinear pattern of the fast directions of azimuthal anisotropy may reflect the movement direction of the material extruded from the central Tibet. The deformation or material movement is continuous from the Longmenshan and Yajiang blocks in the north to the Central Yunnan block in the south. But its magnitude may vary over our study region considering the spatial variations of the two kinds of anisotropy. The deformation may go across the Xiaojiang fault into the South China block, because both radial and azimuthal anisotropy are observed there, but with smaller magnitude. These results indicate a weakened middle and lower crust and a decoupled lithosphere, at least locally, in the southeastern Tibetan Plateau.

5.3. Comparison with shear-wave splitting. We make two assumptions in order to compare the anisotropy from this study with SKS splitting. Firstly, we assume the anisotropy is mainly related to the current deformation. However, frozen anisotropy caused by former deformations may cause the comparison not so straightforward. Secondly, we assume A-type lattice preferred orientation (LPO) in mineral following previous studies ([32], [34]). So the fast direction of the azimuthal anisotropy is parallel to the flow direction. We calculate the fast direction and delay time of shear-wave splitting caused by azimuthal anisotropy in the crust. These splitting results for the crustal anisotropy are then compared with the SKS splitting for the whole crust and mantle ([15], [32], [34]) (Figure 16). The synthetic delay times are large to the west of the Longmenshan-Simian-Xiaojiang fault line (black bars), where the values can reach up to 0.5 s. To the east of this fault line, the predicted delay time is very small, less than 0.2 s. The discrepancy between the synthetic and measured splitting is large in both magnitude and direction in most of our study regions. This large difference indicates that the deformations in the crust and in the mantle are not coherent. In some localized areas, i.e. near the northern Xianshuihe fault and the northern part of the Central Yunnan block, we observe some consistency between the synthetic and measured splitting. In these areas, the delay time from crustal anisotropy can contribute up to 50% of the entire delay time. By itself, however, the consistency does not necessarily mean coupled lithosphere. The deformation may still vary in depth direction, but with less difference as in other regions.

6. Conclusions. In this paper, we apply the eikonal equation to Rayleigh wave Green’s functions to study the shear wave speed $V_{SV}$ and its azimuthal anisotropy for the southeastern Tibetan Plateau and adjacent regions. The isotropic wave speed model is in good agreement with surface geology and reveals nearly similar patterns as our previous study from traditional tomography [13]. Significant $V_{SV}$ contrasts are observed between the high Plateau and Sichuan Basin throughout the entire crust, reflecting distinct tectonics on the two sides of the Longmenshan fault. In contrast, $V_{SV}$ varies smoothly from the high Plateau to the Central Yunnan block, reflecting gradual change over the Lijiang-Muli faults zone. Ubiquitous low wavespeed zones in the middle and lower crust of the high Tibetan Plateau are consistent with increased temperature and low mechanical strength, and these zones may play a role in material transport from central Tibet. Strong azimuthal anisotropy is observed mainly in the high Plateau and the Central Yunnan block. Although a curvilinear pattern associated with the clockwise rotation around the Eastern Himalayan Syntaxes is observed in the entire crust, angular differences of anisotropy are large between the upper and lower crust, suggesting different deformation patterns at surface and in depth. This variation of deformation patterns at different depths is further confirmed by the comparison of the obtained anisotropy in the study with the GPS and SKS splitting results. Although we cannot rule out the influence of historic or frozen anisotropy, our results suggest a weakened middle and lower crust and a decoupled lithosphere in the SE Tibetan Plateau. Their inferences are consistent with the channel flow in the middle and lower crust in our study region, but its extent and magnitude may vary spatially and
be influenced by old structures and their different tectonic histories.

**Appendix A. Azimuthal anisotropy and artifacts in inversion.**

In an HTI medium the phase velocities of both Rayleigh and Love waves have the form [31]:

\[ c(., \psi, \omega) = c_0(\omega) + a_0(., \omega) + a_1(., \omega) \cos(2\psi) + a_2(., \omega) \sin(2\psi) + a_3(., \omega) \cos(4\psi) + a_4(., \omega) \sin(4\psi), \]

where \( c_0(., \omega) \) is the reference phase velocity and \( a_0(., \omega) \) and \( a_i(., \omega) \) \( (i = 1, \ldots, 4) \) are the isotropic perturbation and coefficients for azimuthal anisotropy, respectively. In addition, the phase velocity is relatively higher (lower) because the checkerboard is 180° periodicity can rise from an isotropic bias in the inversion [3, 16]. Hence, we first test whether our data contains all or only parts of these anisotropic terms. For models with different anisotropic terms we compare their data fitting, using

\[ \chi^2 = \frac{1}{N} \sum_{i=1}^{N} \frac{(c_{i,\text{pred}} - \bar{c}_i)^2}{\sigma_{c,i}^2} \]

where \( \bar{c}_i \) and \( c_{i,\text{pred}} \) are the average phase velocity and its prediction in the \( i^{th} \) angle bin assuming a subdivision into \( N \) bins. It is not surprising that \( \chi^2 \) reduces when more terms of anisotropy are introduced. Here we use 10 s Rayleigh wave for example (Figure 17). We find that the fitting is greatly improved from the isotropic case to the anisotropic case with only 2ψ terms for most of the study region. Adding \( \psi \) and 4ψ terms does not reduce \( \chi^2 \) by much. This experiment is consistent with the theoretical study that the 2ψ terms have much larger amplitudes than the 4ψ terms [31]. Hence, we only include the 2ψ terms in our analysis of azimuthal anisotropy.

Because of the regularization terms in the model space and various assumptions, the model cannot be fully resolved even with noise-free data. So the tomography results contain some artifacts which may influence our understanding and analysis of the structures. For example, artifacts in the checkerboard test for isotropic \( V_{SV} \) lead to apparent azimuthal anisotropy (Figure 18). The recovered anomalies well exhibit the checkerboard pattern but with a smaller amplitude. This deficit in anomaly amplitude is reflected and compensated for in the angular variations of phase velocity at locations between the positive-negative anomalies. In directions connecting two high (low) anomalies, the phase velocity is relatively higher (lower). Because the checkerboard is 180° symmetric, this type of artifacts appears as 2ψ azimuthal anisotropy. The magnitude of the artificial anisotropy can reach up to 2% with an input isotropic anomaly of 8%. The magnitude of the artificial

**Fig. 17. Test of \( \psi \), 2ψ and 4ψ anisotropy for the 10 s Rayleigh wave: (a) the reduction of \( \chi^2 \) when only 2ψ anisotropy is introduced with respect to the isotropic model; (b) the further reduction of \( \chi^2 \) when both \( \psi \) and 2ψ anisotropies are introduced; and (c) the further deduction of \( \chi^2 \) when both 2ψ and 4ψ anisotropies are introduced.**
Fig. 18. Artificial azimuthal anisotropy obtained from a checkerboard model: (a) $1.0^\circ \times 1.0^\circ$ checkerboard model; (b) recovered model using 10 s Rayleigh wave data. The phase velocity heterogeneities (isotropic part) are shown as background colors. The black bars show the artificial azimuthal anisotropy. We only plot anisotropies with amplitude larger than 1%.

Anisotropy greatly decreases for realistic isotropic models which have no clear periodicity (Figure 19). Except for the edges of the study region, the amplitude of the $2\psi$ anisotropy is less than 0.5%. This asymmetric medium may cause $360^\circ$ periodicity in phase velocity, e.g. $\psi$ anisotropy [16]. Since little or no $\psi$ anisotropy is observed in our data (Figure 17), we do not study this effect here. The experiments here reveal that we need to perform separated checkerboard tests for isotropic phase velocity and anisotropy in order to correctly analyze their resolutions. The isotropic test can use a checkerboard model similar to the one in Figure 18a; the anisotropic test uses a model with a realistic isotropic phase velocity (as in Figure 19a) and checkerboard anisotropy.

Appendix B. Phase velocity and anisotropy of Love waves.

The eikonal tomography scheme can also be applied to the Love wave contributions to Green’s functions. The variations of phase velocity with propagation angle for location $(102^\circ, 29^\circ)$ are shown in Figure 20. In contrast to the results of Rayleigh waves, the anisotropy is not obvious for both 10 and 30 s. This is further confirmed by the data fitting (Figure 21). For most locations the $\chi^2$ reduces less than 20% from the isotropic to the $2\psi$ anisotropic fitting. Introducing $\psi$ and $4\psi$ anisotropy may improve data fitting at some locations. But these locations show little correlation with the known geology. The final results for the models only containing $2\psi$ anisotropy are shown in Figures 22 and 23. The isotropic $V_{SH}$ are very close to our previous results of traditional tomography [13]. Although the depth sensitively kernels are different from those of Rayleigh waves, the $V_{SH}$ perturbations show similar features as the $V_{SV}$ anomalies. The anisotropy shows no clear pattern or connection with known geology or the anisotropy from Rayleigh waves; their amplitudes are small, always less than 2%. Theoretically, in a slightly anisotropic medium the $4\psi$ terms of anisotropy for the Love waves have amplitudes comparable to the $2\psi$ terms [31]. So the $4\psi$ terms are not negligible. Indeed, azimuthal anisotropy for the Love waves is more complicated. In addition, usually the horizontal components of the seismic records have lower signal-to-noise ratio than the vertical components.
Fig. 19. Artificial azimuthal anisotropy from realistic model: (a) a realistic isotropic phase velocity model from traditional tomography at 10 s [13]; (b) recovered model using 10 s Rayleigh wave data. We only plot anisotropies with magnitude larger than 0.5%.

Fig. 20. The variations of Love wave phase velocity against the raypath direction at period (a) 10 s, and (b) 30 s for location (102°, 29°). The blue stars represent the results from different sources; the red stars with error bars are the average values and standard deviations of the wavespeed in the 20° angle bins. The thick red lines are the 2ψ anisotropy fitting results for the averages values.
Fig. 21. Test of $\psi$, $2\psi$ and $4\psi$ anisotropy for the 10 s Love wave, comparable to the test illustrated in Figure 17 for the 10 s Rayleigh wave.

Fig. 22. Eikonal tomography results for 10 s Love wave: (a) the isotropic phase velocity; (b) standard error for isotropic phase velocity; (c) azimuthal anisotropy; (d) standard error for the magnitude of azimuthal anisotropy.

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Fig. 23. The same as Figure 22, but for 30 s Love wave.
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