JGR Solid Earth

RESEARCH ARTICLE

10.1029/2019JB018344

Key Points:

- A new seismic model of the crust and upper mantle beneath the Tibetan Plateau is presented based on dense-array adjoint tomography
- Our model supports the dramatic east-west change of northward Indian plate beneath the Tibetan Plateau
- Obvious thinning and subvertical subduction of the Indian lithosphere are clearly imaged beneath the Bangong-Nujiang Suture

Correspondence to:

N. Fuji and Yuan Gao
nobuaki@ipggp.fr;
qzgyseis@163.com

Citation:

Xiao, Z., Fuji, N., Iidaka, T., Gao, Y., Sun, X., & Liu, Q. (2020). Seismic structure beneath the Tibetan Plateau from iterative finite-frequency tomography based on ChinArray: New insights into the Indo-Asian collision. *Journal of Geophysical Research: Solid Earth*, 125, e2019JB018344. <https://doi.org/10.1029/2019JB018344>

Received 9 JUL 2019

Accepted 17 JAN 2020

Accepted article online 23 JAN 2020

Seismic Structure Beneath the Tibetan Plateau From Iterative Finite-Frequency Tomography Based on ChinArray: New Insights Into the Indo-Asian Collision

Zhuo Xiao^{1,2}, Nobuaki Fuji^{3,4}, Takashi Iidaka⁴, Yuan Gao¹, Xinlei Sun², and Qi Liu⁴

¹Key Laboratory of Earthquake Prediction, Institute of Earthquake Forecasting, China Earthquake Administration, Beijing, China, ²Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, Guangzhou, China, ³Institut de Physique du Globe de Paris, Paris, France, ⁴Earthquake Research Institute, University of Tokyo, Tokyo, Japan

Abstract Indo-Asian continental collision has contributed to the growth of the Tibetan Plateau, which is one of the most prominent uplifts worldwide since Cenozoic. The crustal and upper-mantle structures are key factors in understanding the evolutionary process as well as lateral growth of the plateau. We present a new 3-D seismic model beneath the Tibetan Plateau and its surrounding areas, which uses a large-scale dense array from iterative finite-frequency tomography. The new tomographic images show obvious east-west geometrical change of the northward Indian lithosphere by high-velocity anomalies beneath Tibet. The high-velocity anomaly extends to Pamir in the western side, but it only reaches north Lhasa in central and eastern Tibet. A convective removal of the thickened Tibetan lithosphere is observed as a high-velocity anomaly, followed by a subvertical subduction of the Indian mantle lithosphere beneath the Bangong-Nujiang Suture in central Tibet. We speculate its link to the widespread magmatism and rapid uplift of central Tibet in late Miocene. A high-velocity gap between 100 and 250 km underneath the Longmenshan fault indicates that lithospheric delamination may play an important role in surface evolution between eastern Tibet and surrounding rigid blocks. Our detailed seismological model will give an insight into how the continents collide.

1. Introduction

Tectonic evolution in the Tibetan Plateau (Tibet) has been dominated by crustal compression and shortening as well as mountain building immediately since the initial continental collision between India and Asia about 50–70 Ma (e.g., Molnar & Tapponnier, 1975; Yin & Harrison, 2000). Consequently, we can observe the difference between the western and eastern halves of China, in terms of the elevation (Royden et al., 1997) and thickness of the crust (Kind et al., 2002), horizontal movement rate (Wang et al., 2001; Zhao et al., 2015), and seismicity. Most seismic hazards inside mainland China occurred in the transition zone between Tibet and eastern cratons (the Yangtze Craton in the south and the North China Craton in the north), or “North-South Seismic Zone” (red dashed rectangle in Figure 1). It is thus interesting to study Tibet as a natural laboratory for our understanding of continental dynamics. Various geophysical and geological studies have been carried out in order to understand the tectonic evolution of Tibet, whose continental collision has been explained by many geodynamical modelings over decades (e.g., Chung et al., 2005; England, 1993; Molnar et al., 1993; Royden et al., 2008; Tapponnier et al., 2001). The seismological models beneath Tibet are derived from several international collaborative projects of temporary seismic arrays such as INDEPTH (InterNational DEep Profiling of Tibet and the Himalaya; Kosarev et al., 1999; Li et al., 2008; Tilmann et al., 2003), MIT-CIGMR (a collaborative project among Massachusetts Institute of Technology, Lehigh University, and Chengdu Institute of Geology and Mineral Resources; Yao et al., 2006; Xu et al., 2007), and HiCLIMB (Himalayan-Tibetan Continental Lithosphere During Mountain Building; Nabelek et al., 2009).

However, albeit these efforts, we have not yet constrained the dynamics beneath Tibet, since many controversial discussions are still ongoing over the subsurface structure and evolution beneath Tibet as well as its surrounding areas. For instance, we are still uncertain about the temperature and the lithospheric thickness of the plateau. It remains unclear to us whether the mantle lithosphere beneath Tibet is (i) cold and thus thickened by shortening (Bao et al., 2015; Chen et al., 2017; Huang & Zhao, 2006; Lebedev &

Figure 1. Topography and bathymetry around China and India with tectonic interpretations. Strong earthquakes ($M_w > 7$) and GPS velocity relative to Eurasia (Zhao et al., 2015) are represented by red circles and blue arrows, respectively. QDB and SCB refer to Qaidam Basin and Sichuan Basin, respectively. The North-South Seismic Zone of China is bounded by the red dashed rectangle.

van der Hilst, 2008) or (ii) hot and thus thinned by viscous instability mantle lithosphere beneath Tibet (Curtis & Woodhouse, 1997; Liu et al., 2004; Royden et al., 2008). Another enigma is the geometry of the northward Indian lithosphere (Gao et al., 2016; Kosarev et al., 1999; Tilmann et al., 2003). Is Asian lithosphere subducted or underthrust beneath Tibet (Kind et al., 2002; Nabelek et al., 2009; Zhao et al., 2011)? The lack of understanding of the region is due to several factors: (i) low resolution and uncertainties of seismological studies; (ii) uneven and sparse distribution of seismic stations; and (iii) strong assumptions and regularization imposed during the inversions.

After the Wenchuan earthquake (12 May 2008) that occurred in a transition zone between Tibet and the Sichuan Basin (SCB), intense geophysical surveys have been performed within and around the area, including dense seismic array deployment (ChinArray; e.g., doi: 10.12001/ChinArray.Data). The continuous deployment of seismic stations opened the avenue for an improvement of seismic tomography resolving power (Liu & Gu, 2012). Due to the high performance computer and technical progress in numerical methods (e.g., Komatitsch & Tromp, 2002a, 2002b), full-waveform tomography based on 3-D numerical simulations at regional and global scales has become feasible (e.g., Bozdağ et al., 2016; Chen, Niu, et al., 2015; Konishi et al., 2014; Fichtner et al., 2009; Tao et al., 2018; Tape et al., 2009; Wang et al., 2018;

Figure 2. Spectral-element mesh of isotropic shear wave speed in the initial model and spatial distribution (a), and event and station locations (b). Red beach balls are used in inversion, and black beach balls are used during validations of our inverted models. Permanent stations are denoted by solid orange triangles, and three temporary dense arrays are denoted by open triangles. From south to north, the projects are Himalaya I (open green triangles), ChuanXi Array (open blue triangles), and Himalaya II (open purple triangles).

Zhu & Tromp, 2013; Zhu et al., 2012). In this paper, we present a new seismological model with enhanced clarity on the deep structure of Tibet and its surrounding areas, taking advantage of both the regional dense-array and iterative finite-frequency tomography (Tape et al., 2007).

2. Data

Three-component seismic waveform data were recorded by 1,335 temporary seismic stations (open triangles in Figure 2) from ChinArray and 246 permanent seismic stations (orange-filled triangles in Figure 2) from China Earthquake Network Center and Incorporated Research Institutions of Seismology. The temporary seismic surveys consist of three regional dense arrays, that is, the ChuanXi, Himalaya I, and Himalaya II projects (blue, green, and purple triangles, respectively, in Figure 2). Four hundred seventy-one temporary stations have sensitivity up to a period of 120 s, and the rest up to 60 s.

Considering the trade-off between the computational cost of full-waveform tomography and data coverage in the study area, 51 earthquakes (red beach balls in Figure 2) were selected for iterative finite-frequency tomography. We balanced the numbers of earthquakes per azimuths. Another extra data set containing 91 earthquakes (black beach balls in Figure 2) were later used for validating our inversion results. All the earthquakes involved in this study have a magnitude of M_w 4.9 to 6.7 (Table 1) with a half duration of less than 6 s. We thus can assume a delta function as a source time function for our studies that used band-pass filters of periods larger than 12 s. Source parameters including locations, origin time, and moment tensor solutions are collected from the Global Centroid-Moment-Tensor database (GCMT). 3-D crustal model Crust1.0 (Laske et al., 2013) combined with a global 3-D radially anisotropic mantle model S362ANI (Kustowski et al., 2008) are used as the initial model M00.

The GCMT source depth is of a precision of a few kilometers (Hjörkeufsdóttir & Ekström, 2010), which could result in traveltime differences of at most 0.5 s for P -related phases and 1 s for S -related phases. Should they be systematic, the inversion results could be slightly biased. However, here we do not perform simultaneous GCMT inversions by fixing the CMT, since we are handling traveltimes measured using cross correlation of

Table 1
Events Selected for Seismic Inversions

Date	Longitude	Latitude	Depth (km)	Magnitude
2007-01-08	92.30	8.03	12.00	5.7
2007-04-03	70.62	36.53	223.20	5.7
2007-05-05	81.97	34.33	23.85	5.8
2007-05-16	100.89	20.52	12.57	5.7
2007-06-02	101.13	23.02	12.00	5.7
2008-01-09	85.32	32.30	13.27	6.3
2008-04-02	69.23	−4.98	12.00	5.5
2008-05-25	105.44	32.57	26.95	5.8
2008-06-27	91.82	10.92	17.09	6.4
2008-08-25	83.51	30.61	17.28	6.2
2008-08-27	104.02	51.76	23.46	5.8
2008-10-05	73.64	39.50	12.00	6.4
2008-10-06	90.50	29.66	12.00	6.0
2008-10-28	67.48	30.40	17.17	6.3
2008-11-10	95.75	37.51	27.23	6.4
2011-09-05	97.86	2.88	94.59	6.6
2011-09-18	88.35	27.44	45.99	6.6
2011-10-20	70.44	20.98	12.00	4.9
2011-11-30	118.97	15.54	12.31	6.0
2012-02-04	105.62	42.33	22.54	5.4
2012-02-26	96.00	51.69	20.46	6.4
2012-04-30	89.60	1.73	12.00	5.5
2012-06-29	84.72	43.42	27.60	6.2
2012-07-12	70.76	36.47	192.23	5.6
2012-08-12	82.52	35.88	16.70	6.1
2012-09-14	100.32	−3.58	12.00	6.1
2012-11-11	96.03	22.73	16.81	6.3
2013-01-24	87.75	49.80	19.62	5.3
2013-01-28	79.76	42.57	24.43	6.1
2013-01-30	94.86	32.86	19.60	5.2
2013-02-11	92.41	38.56	19.68	5.3
2013-02-25	85.63	34.30	25.88	5.1
2013-03-03	99.84	25.94	12.00	5.5
2013-04-20	103.12	30.22	21.87	6.5
2013-05-26	67.26	39.94	26.86	6.2
2013-07-21	104.36	34.60	17.15	6.1
2013-10-31	121.39	23.64	19.90	6.3
2014-07-31	95.19	12.30	12.00	5.8
2014-08-03	103.50	27.06	14.59	6.2
2014-09-26	95.18	12.36	12.00	5.5
2014-09-30	67.66	1.51	14.23	5.5
2014-11-20	93.46	23.49	35.15	5.6
2014-11-22	101.78	30.24	24.62	5.9
2014-11-25	101.81	30.16	26.08	5.6
2014-12-29	121.45	8.68	14.96	6.1
2015-02-13	121.39	22.65	29.75	6.2
2015-03-03	98.58	−0.72	23.60	6.1

Table 1 Continued

Date	Longitude	Latitude	Depth (km)	Magnitude
2015-04-26	85.95	27.56	20.61	6.7
2015-05-15	102.14	−2.61	158.37	6.0
2015-06-04	116.65	6.17	12.34	6.0
2015-07-03	78.14	37.58	15.62	6.4

Note. Dates are formatted as YYYY-MM-DD.

waveforms filtered up to 12 s that can smear the potential mislocation of each individual event. Instead, we carefully removed all the events whose observed waveforms are initially correlated lower than 0.5.

3. Iterative Finite-Frequency Traveltime Tomography

3-D numerical simulation of seismic wave propagation in a realistic Earth model has become feasible based on spectral-element method (Komatitsch & Tromp, 2002a, 2002b). Iterative finite-frequency traveltime tomography builds on full 3-D numerical simulations in forward modeling, by calculating sensitivity kernels that can be expressed as cross correlations between forwarding wavefield and back-propagated wavefield (Geller & Hara, 1993; Tromp et al., 2005). These theoretical and numerical developments in global seismology have successfully revolutionized the seismic imaging technique (Fichtner et al., 2009; Tape et al., 2009; Zhu et al., 2012). Here we apply regional iterative finite-frequency traveltime tomography to dense-array observations of mainland China.

3.1. Inversion Strategy

We implement an adjoint-based iterative finite-frequency traveltime inversion that minimizes frequency-dependent phase misfits. A multitaper technique (Zhou et al., 2004) is used to measure frequency-dependent phase differences between three-component observed and synthetic seismograms within the windows selected automatically by FLEXWIN (Maggi et al., 2009). We combine all picked phases including (but not limited to) *P-SV* and Rayleigh waves on vertical and radial components and *SH* waves as well as Love waves on transverse components, in order to simultaneously compute the gradient directions for inversions. We use three frequency bands and multiscale strategy for frequency-dependent phase difference measurement to reduce the nonlinearity of the inverse problem: for the first several iterations, we employed three passbands as 15–40, 30–60, and 50–100 s, and we lowered the periods down to 12–40, 20–50, and 40–100 s in the following iterations in order to refine more detailed structural features. Nevertheless, the misfit of surface waves was not considered in the short periods (12 or 15–40 s) due to the precision limit of initial crust model (Chen, Niu, et al., 2015) and due to the increasing complexity of internal structure and surface topography. The effect of crustal structure and topography on short-period surface-wave waveforms is much more nonlinear than behavior of short-period body wave waveforms due to long-length structure of the upper mantle (the region of our interest). We thus can hardly reconcile the two effects together within our inversions. The overall misfit can be written as follows:

$$\chi = \frac{1}{N} \sum_{i=1}^N \frac{1}{M_i} \sum_{j=1}^{M_i} \left(\frac{\Delta T_{ij}}{\sigma_{ij}} \right)^2 \quad (1)$$

where N is the total number of categories of misfit summed by three components in three period bands, which equals 9 in our study, M_i is the total number of all measurements' windows in the i th category, and ΔT_{ij} is the phase difference measured by cross correlation between observed and synthetic seismograms. σ_{ij} denote associated uncertainties estimated from the j th window at i th record.

3.2. Model Parameterization

Following the model parameterization of previous studies based on adjoint tomography (Chen, Niu, et al., 2015; Zhu et al., 2015), we assume that the bulk modulus remains isotropic, warranting an empirical relationship between relative perturbation in mass density and isotropic shear wave speed. Radial anisotropy can be expressed only in terms of four parameters: the isotropic bulk sound speed c , the vertically polarized horizontal shear wave speed β_v , the horizontally polarized horizontal shear wave speed β_h , and the dimensionless radial anisotropic parameter η . Perturbations in total misfit thus can be expressed as a volume

integral over relative perturbations in these parameters (Zhu et al., 2015):

$$\delta\chi = \int_V (K_c \delta \ln c + K_{\beta_v} \delta \ln \beta_v + K_{\beta_h} \delta \ln \beta_h + K_\eta \delta \ln \eta) dV \quad (2)$$

where V denotes the model volume; K_c , K_{β_v} , K_{β_h} , and K_η are Fréchet derivatives of traveltimes with respect to each parameter.

In order to formally calculate Fréchet derivatives for waveforms, we use the shell (or pixel) perturbation for a seismogram at a receiver $\mathbf{r}_R$ with a source at $\mathbf{r}_S$ in time domain (see general explanations in Fuji et al., 2012; Geller & Hara, 1993; Kawai & Geller, 2010; Tarantola, 1984):

$$\begin{aligned} \delta u_i(\mathbf{r}_R, t; \mathbf{r}_S) = & - \int_{-\infty}^{\infty} d\tau \int_V dV \dot{u}_j(\mathbf{x}, \tau; \mathbf{r}_S) \delta \rho(\mathbf{x}) \dot{v}_{ji}(\mathbf{x}, t + \tau; \mathbf{r}_R) \\ & - \int_{-\infty}^{\infty} d\tau \int_V dV u_{j,q}(\mathbf{x}, \tau; \mathbf{r}_S) \delta C_{jqrs}(\mathbf{x}, \tau) v_{ri,s}(\mathbf{x}, t + \tau; \mathbf{r}_R) \end{aligned} \quad (3)$$

with a forward equation of motion

$$\rho(\mathbf{x}, t) \ddot{u}_i(\mathbf{x}, t) - [C_{ijkl}(\mathbf{x}) u_{k,l}(\mathbf{x}, t)]_{,j} = \delta(\mathbf{x} - \mathbf{r}_S) f_i(t) \quad (4)$$

and a backward-propagated equation of motion

$$\rho(\mathbf{x}, -t) \ddot{v}_{rj}(\mathbf{x}, -t) - [C_{rjkl}(\mathbf{x}) v_{kl}(\mathbf{x}, -t)]_{,j} = \delta(\mathbf{x} - \mathbf{r}_R) \delta_{rj} \delta(t) \quad (5)$$

where f is the external force, $\mathbf{x}$ an arbitrary position, and t time. ρ denotes density, C_{ijkl} denotes elastic moduli, and $\delta\rho$ and δC_{ijkl} are small perturbations to a starting model. u and v respectively denote displacement wavefields for the forward and backward propagated equations of motion. A dot and a comma denote partial derivatives with respect to time and space, respectively. Traveltime partial derivatives can be expressed as (e.g., Dahlen et al., 2000; Zhao & Chevrot, 2011):

$$K_m = \int_{t_1}^{t_2} dt J(\mathbf{r}_R, t; \mathbf{r}_S) \delta u_i(\mathbf{r}_R, t; \mathbf{r}_S) \quad (6)$$

where m is a model parameter such as wave velocity that can be express as a combination of ρ , C_{ijkl} , $\delta\rho$, and δC_{ijkl} and the integral is over the time window $[t_1, t_2]$ that contains the seismic phase, multiplied by the Jacobian

$$J(\mathbf{r}_R, t; \mathbf{r}_S) = - \frac{\dot{u}_i(\mathbf{r}_R, t; \mathbf{r}_S)}{\int_{t_1}^{t_2} dt [\dot{u}_i(\mathbf{r}_R, t; \mathbf{r}_S)]^2}. \quad (7)$$

3.3. Model Updating

Throughout the iterative finite-frequency tomography, we used the gradient-based algorithms to minimize the misfit function and iteratively update model parameters, as the full Hessian matrix (the second derivative of the misfit function) is excessively expensive to compute and store (Chen et al., 2007). We use a conjugate gradient method to update model parameters. All individual event kernels are first summed to form the misfit kernels, which are then multiplied by a preconditioner, that is, the model improvement during the first iteration is as same as the steepest descent gradient direction. Here the diagonal of Hessian is chosen as the preconditioner (Luo et al., 2013) for accelerating the inversion convergence. In addition, we use a Gaussian with a width of 5 km in the vertical plane and 75 km in the horizontal plane to smooth the preconditioned kernels so as to stabilize the inverse problem. For the i th iteration, model updating is performed as follows:

$$\ln \frac{\mathbf{m}_{i+1}}{\mathbf{m}_i} = \alpha_i \mathbf{d}_i \quad (8)$$

where we define the gradient direction by

$$\mathbf{d}_i = -\mathbf{g}_i + \beta_i \mathbf{d}_{i-1} \quad (9)$$

and

$$\beta_i = \frac{\mathbf{g}_i^T \cdot (\mathbf{g}_i - \mathbf{g}_{i-1})}{\mathbf{g}_{i-1}^T \cdot \mathbf{g}_{i-1}} \quad (10)$$

Figure 3. Step length test (a) in the first iteration and variation of misfit value during the whole inversion process (b). For every single iteration, we test the value of step length from 0.01 to 0.07. The total misfit value and frequency-dependent misfit vary with different step lengths. In the first iteration case, we set α as 0.04 for model updating. Blue and red circles represent two stages of multiscale inversion. The misfit value increases at the sixth iteration due to lowered frequency; thus, smaller-scale structures are taken into account.

where $\mathbf{d}_i$ is the i th search direction correlating with gradient and α is a step length along the search direction. $\mathbf{g}_i$ is the gradient direction (the product of Fréchet derivative matrix and waveform residual) for the i th iteration. Here, instead of being obtained through linearized formula (Mora, 1987), the optimal step length α_i is determined based on a line search method (Figure 3) from a subset of events in order to reduce computational cost (Zhu et al., 2015). With the aim of a better representation of the total misfit (Wang et al., 2018), eight events were selected to form the subset considering the surface coverage and measurable FLEXWIN windows of all the data sets. We test α values ranging from 0.01 to 0.07 for line search at each iteration, which necessitates an additional 8×7 forward simulations each time.

The lateral dimension of the model is $60^\circ \times 60^\circ$ (blue rectangles in Figure 2), which almost covers the whole mainland China excepting the northeastern part. We used one “cubed sphere” chunk of a global mesh and five Gauss-Lobatto-Legendre integration points in each direction inside an element (Komatitsch & Tromp, 2002a). In this study, the horizontal element size is approximately 20 km, which makes an average spacing of 5 km between Gauss-Lobatto-Legendre points, resulting in the minimum period being ~ 9 s. We used 400 cores on a SGI AltixUV200 cluster, resulting in a forward wavefield simulation of about 50 and 100 min for an adjoint simulation of an event kernel. As the misfit value converges to a certain range (Figure 3), a new radially anisotropic model of crust and upper mantle beneath the Tibet named TP2019 is constructed after 12 iterations. The total computational time was nearly 836,000 CPU hr.

Figure 4. Improvement of phase difference compared between M00 and TP2019. Red and blue bars represent values for the final model TP2019 and the initial 3-D model M00, respectively. The total number of windows selected by FLEXWIN are given in brackets.

4. Model Validation and Assessment

Seismological structure models should be validated with other data sets in order to assess their robustness. In order not to violate Occam's razor, it is important to propose a model that can explain the data with as simple models as possible. Despite the growing power of seismic tomography as a tool to bring to light the Earth's interior, the model robustness is still suffering from the ill-posedness of seismic inverse problems (Rawlinson et al., 2014). Hence, we need to quantify our obtained models from several aspects.

4.1. Traveltime and Waveform Fitting

Here in this subsection we first discuss the improvement of traveltime and waveform fits. It is first natural to discuss how the travel time differences, the misfit function used in our inversion schemes, have been reduced. We then discuss the waveform fitting although we have not included it in our inversions, in order to access the superior probability of our inverted seismic models. Compared to the initial-model synthetics, both mean value and standard deviation of the misfit of the final-model synthetics decrease significantly. We can also clearly observe the improvement of phase difference between M00 and TP2019 in all three pass-bands (Figure 4). Besides, the total number of measurement windows, which were automatically selected by FLEXWIN, also increase considerably by 25%. The seismic structure of TP2019 reveals more details than M00 (Figure 5) and depicts the eastern cratons as clearly as previous high-resolution tomographic models (Shen et al., 2016; Tao et al., 2018).

Since only phase misfit in selected windows has been considered when we inverted for seismic structure, it is important to validate waveform fitting. Frequency-dependent waveform fitting of one earthquake recorded by stations at different azimuths shows good agreement almost within all the selected windows from three waveform components (Figure 6). It is worth noting that the final-model synthetic waveforms fit better the observed waveforms not only in main phases, used for inversions, but in postcursors whose traveltimes were not included in our seismic inversions.

In order to highlight the improvement of waveform fitting, we checked the waveform difference between the initial model and the final model by a cluster of stations (Figure 7). As a consequence of phase misfit reduction, the short-period phase difference of body waves between synthetics and observations decrease significantly. More remarkably, the obvious reduction of phase difference of surface waves (Rayleigh waves on radial and vertical components), which were not considered in inversions, also prefers the final-model synthetics.

Figure 5. Seismic structure of M00 and TP2019 and several tomographic models. SL2013sv (Schaeffer & Lebedev, 2013), FWEA18 (Tao et al., 2018), and Shen2016 (Shen et al., 2016) are high-resolution seismic models of the globe, East Asia, and mainland China, respectively.

Figure 6. Multiscale waveform fitting between observations and synthetics simulated by TP2019. Both the phase and amplitude of three-component waveform fit well in all the passbands. Gray boxes represent waveform windows selected by FLEXWIN.

Figure 7. Evolution of short-period waveform fitting from M00 to TP2019. The two left panels denote radial components, and the two right panels denote vertical components. Observations and synthetics simulated by M00 and TP2019 are plotted as black, purple and red, respectively. Actually, the short-period surface waves were not considered in inversion due to the precision limit of the initial model, but the misfit also exhibits good improvements.

4.2. Model Validation by Extra Data

A reliable inverted model should not only fit used data but also be valid for those that are not. We thus test an extra data set containing 91 earthquake events (black beach balls in Figure 2), which were deliberately excluded during our inversions, against initial-model and final-model synthetic data. The phase misfit perturbations $(\chi_{TP2019} - \chi_{M00})/\chi_{M00}$ of each earthquake between M00 and TP2019 shown in Figure 8 display the statistical significance of our new seismic model.

Since 3-D global seismic models (i.e., Crust1.0 and S362ANI) are able to well constrain long-wavelength structures, we only show the improvement of waveform fitting in two shorter frequency bands of the extra data set (Figure 9). The decrease of phase difference in FLEXWIN windows and the increase of the number of accepted windows during iterations (Figure 8) are both a confident sign of the statistical significance of

Figure 8. Misfit perturbation of extra data set between M00 and TP2019. (a) Misfit variation of each event; blue circles denote events which undergo a misfit reduction, and yellow circles denote an increase of misfit. (b) Same as Figure 4, but for extra data set.

our models. The reliability of TP2019 was further confirmed by higher similarity between synthetics and observed short-period surface waveforms, which the initial model failed to simulate accurately. Meanwhile, we also compared waveform differences between 1-D and 3-D initial models using the extra data set. It is clear that the similarity between observation and synthetics is higher when a full 3-D seismic model is adopted as the initial model, with more windows available for selection. We propose that the use of 3-D initial models was able to reduce the nonlinearity during inversions.

4.3. Resolution Analysis: The Distribution of Volumetric Kernel and Point Spread Function Test

Due to its heavy computational cost in every single forward modeling, resolution analysis in adjoint tomography in general has limited application in the estimates of composite volumetric sensitivity (Rawlinson et al., 2014; Tape et al., 2009). First, the spatial resolution of finite-frequency tomography is primarily controlled by ray coverage (Luo et al., 2013; Tape et al., 2007). Hence, misfit kernel preconditioned by the approximate Hessian, which can be simply computed by correlating forward and adjoint wavefields, is a reliable proxy for resolution analysis (Baron & Morelli, 2017; Tape et al., 2009). Here we give an approximate estimate

Figure 9. Model validation revealed by extra data set: evolution of frequency-dependent waveform fitting and comparison between 1-D and 3-D initial models. The left column shows short-period (12–40 s) waveforms, and the right column shows intermediate-period (20–50 s) waveforms. Gray rectangles denote waveform windows selected by FLEXWIN under a same standard.

Figure 10. Volumetric coverage of the misfit kernel for β_v at different depths. White areas represent relatively higher resolution, with black to the contrary. The green rectangles represent the area to be discussed in the following sections.

Figure 11. Point spread function test under central Tibet. (a) Horizontal slice. (b) Vertical cross section. The left, middle, and right panels denote the Gaussian spherical perturbation in β_v and corresponding Hessian kernel of β_v and β_h , respectively.

for model uncertainties by spatial distribution of preconditioned misfit kernel (β_v in Figure 10). The results show a correlation between large kernel values and high ray density. Despite the overall decrease of the spatial coverage of misfit kernel with depth, it is worth noting that the misfit kernel at different depths in the crust and upper mantle beneath Tibet shows good spatial coverage.

More recently, point spread function (PSF) test have been proposed and applied to local resolution analysis in adjoint tomography (Chen et al., 2017; Fichtner & Trampert, 2011). The PSF can be used to evaluate resolution in specific regions and trade-off between different model parameters (Zhu, 2018). In this study, a finite-difference approximation (Zhu et al., 2015) is applied to calculate the action of the Hessian on an impulsive model perturbation. Here, we focus on the PSF test targeting the upper mantle of central Tibet. We employ a 3-D Gaussian anomaly centered at 250 km with a 120 km radius and a maximum of 4% perturbation in β_v (Figure 11). The PSF at horizontal and vertical cross sections both recover the main features of the input perturbations, indicating a high quality of tomographical images under central Tibet. The comparison of "Hessian kernel" between β_v and β_h suggests that there is only limited trade-off between the two shear wave speeds of TP2019.

4.4. Comparison With Traditional Tomographical Model

Similarities and differences between new seismic model and existing models will help explain the reliability of the inversion and reveal new features of seismic structure of the Earth's interior. We first compare TP2019 with a shear wave velocity model of mainland China (Bao et al., 2015), constructed by ambient noise and surface-wave tomography. The model exhibits high resolution in shallower lithosphere. Images of three profiles across the entire continental China show a high consistency in the absolute velocity of the shear wave (Figure 12). In particular, both models show distinct low-velocity anomalies beneath sedimentary basins (Tarim, Qaidam, and south Caspian basin), which indicates the ability of TP2019 to describe the crust structure with high qualities.

In contrast, the differences between the two models can be significant in short-wavelength profiles across Tibet, especially in the mantle lithosphere (Figure 13). Given the lower resolution of ambient noise and surface-wave tomography in the upper mantle as well as the limited coverage of seismic stations in Tibet in the previous study (Bao et al., 2015), we consider these discrepancies to be natural due to either

Figure 12. Comparison between TP2019 and seismic model constructed by ambient noise and surface-wave tomography. The elevation and crustal thickness are represented by gray and black lines. All locations of profiles are retrieved from Bao et al. (2015).

Figure 13. Continuation of Figure 11 for profiles beneath Tibet.

Figure 14. Model data of FWEA18 and EARA2014 used in this study. Earthquake events selected for misfit comparison are represented by red beach balls, which are generally located in the overlap region.

regularization terms or different densities of seismic observations. Overall, the comparison suggests that the new seismic model is able to produce reliable results and could be utilized to extract new structural information beneath Tibet, especially in the upper mantle. Our model is smoother without fluctuations and is able to generate better fitting synthetic seismograms than other models with smaller-wavelength fluctuations. It means that we will not need any additional small-wavelength structure to explain our short-period data, according to Occam's razor.

4.5. Waveform Comparison With Previous Full-Waveform Tomographic Models

Recently, two studies (Chen, Niu, et al., 2015; Tao et al., 2018) successively obtained continental scale seismic models of East Asia based on adjoint tomography: EARA2014 (East Asia Radially Anisotropic model) and FWEA18 (Full Waveform Inversion of East Asia in 2018). Despite their focus on the whole East Asia, both models show good resolution beneath western China and Tibet. To further highlight the contribution of the large-scale dense array to waveform tomography, we select events located in the overlap region (Figure 14) of the above two models (Chen, Niu, et al., 2015; Tao et al., 2018) and TP2019 for comparison through waveform simulation. However, as not all data for Tibet are available with EARA2014

Table 2
Phase and Waveform Misfit of Three Models

		Model		
		EARA2014	FWEA18	TP2019
Total window number		41,261	47,941	63,177
Total window length (s)		776,254	981,732	1,308,186
Waveform misfit	In window	0.5651	0.1932	0.2625
	Whole waveform	8.0271	7.7680	4.2651
Phase misfit	12–40 s	1.859	1.649	1.102
	20–50 s	2.181	2.039	1.679
	40–100 s	2.068	1.428	1.134

Note. The windows are automatically selected by FLEXWIN with the same standard.

Figure 15. Waveform comparison between FWEA18 and TP2019: results of long-period (40–100 s) waveforms recorded by vertical component. Black, red, and blue lines represent the observed waveforms, synthetics of TP2019, and synthetics of FWEA18, respectively. All the waveforms are displayed from 0 to 500 s after the source's origin time.

(<https://chenseismolab.org/resources/>), we manually filled in the missing parts using the data of TP2019. The comparison shows that FWEA18 and TP2019 share a relatively small phase misfit of long-period (40–100 s) waveforms (Table 2). Considering that FWEA18 employs EARA2014 as the initial model, we mainly focus on the comparison between FWEA18 and TP2019 in this section.

Given the rather small phase misfit of both models, that is, 1.134 of TP2019 and 1.428 of FWEA (Table 2), the corresponding synthetic long-period waveforms are highly consistent with observations (Figure 15), which indicates the large-scale structures of FWEA18 and TP2019 are almost identical and close to reality. However, a larger difference in phase misfit of shorter-period waveform indicates that the smaller-scale structures vary with each model. For example, for an event that occurred in Tibet, the Vsv of FWEA18 is significantly faster than that of TP2019 at the depths of 80 and 100 km (Figure 16). Despite relatively small phase delay of the SH waves in both models, the Love waves of FWEA18 considerably precede that of both TP2019 and observations in most stations except YULB and YHNB located in Taiwan. The phase difference indicates that TP2019 is closer to the reality than FWEA18 beneath eastern Tibet (Figure 16).

Therefore, we propose that TP2019 could image the structures beneath Tibet in greater detail and with higher accuracy compared with existing full-waveform tomographic models due to the contribution of the dense ChinArray.

5. Results and Discussion

In this paper, we propose a new seismological model beneath the Tibet (green rectangles in Figure 10) in order to geodynamically constrain continental collision between India and Asia. Due to the higher sensitivity of shear wave velocity to physical property changes, we focus on the deep structure beneath the continental collision zone with the isotropic *S* wave model.

Figure 16. Velocity and waveform comparisons between FWEA18 and TP2019. Intermediate-period (20–50 s) of SH and Love phases recorded by transverse component is shown by a window of 200 s. Black, red, and blue lines represent the observed waveforms, synthetics of TP2019, and synthetics of FWEA18, respectively. The phase difference between FWEA18 and TP2019 mainly results from the higher V_{SH} of FWEA18 in uppermost mantle beneath southeastern Tibet.

Figure 17. Isotropic shear wave velocity at each depth beneath Tibet. The reference shear velocity is the 1-D model of TP2019. Green circles represent mantle earthquakes collected from China Earthquake Network Center, GCMT, and previous study (Chen & Yang, 2004). White lines represent vertical profiles, and black lines represent boundaries of tectonic blocks. AKMS = Anymaqen-Kunlun-Muztagh Suture; ATF = Altyn-Tagh Fault; XSH = Xianshuihe Fault; LMS = Longmenshan fault; QL = Qilianshan Orogen; WKL = West Kunlun Orogen; KKF = Karakorum Fault; KLF = Kunlun Fault. WHS and EHS represent western and eastern Himalaya syntaxis, respectively.

Shear velocity structure at different depths shows widespread low-velocity anomalies in central and northern Tibet at a depth around 80 km (Figure 17b). These widespread low-velocity anomalies, combined with the observed low S_n (Ni & Barazangi, 1983) and P_n (Liang & Song, 2006) velocities, suggest that the uppermost mantle beneath central and northern Tibet is hot or fragmented with high attenuation. Low velocity also appears in the uppermost mantle and continues to the depth of 160 km in the east transition zone between Tibet and the Yangtze Craton (Figure 17c), where crustal flow might occur (Royden et al., 1997) and strong earthquakes happen frequently. Seismic tomographic studies (Lei & Zhao, 2016; Liu et al., 2014) have speculated that the eastward extrusion of Tibet. The eastward extrusion might have been blocked by

Figure 18. Vertical cross sections of velocity perturbation relative to STW105 (Kustowski et al., 2008) beneath Tibet. Yellow and green circles represent crustal and mantle earthquakes. Dashed lines denote 410 and 660 km discontinuities. Blue and white contours represent the high- (3%) and ultrahigh- (6%) velocity anomalies, respectively.

the strong SCB and led to weak rock strength in this region, when we looked at relatively low wave velocity in the lower crust and upper mantle. On the other hand, receiver function studies (Zhang et al., 2010) suggested lateral asthenospheric flow or asthenospheric upwelling caused by lithospheric delamination as the potential cause of low velocity.

High velocity can be observed at the depth of 80 km underneath the Tarim Basin, Yangtze Craton, Qaidam Basin, and Ordos Basin (Figure 17b), all of which are relatively stable tectonic blocks. The high-velocity anomaly underneath these geological provinces is only visible in mantle lithosphere. It almost vanishes below 250 km (Figure 17d). We can interpret this as the cold and cratonic lithosphere beneath the stable blocks. However, high velocity body clearly imaged in southern Tibet from the shallow mantle lithosphere to the deep mantle transition zone (Figure 17e) has been previously interpreted as the subduction of Indian lithosphere (IL) (Chen et al., 2017; Li et al., 2008). A depth-dependent pattern of the high-velocity anomalies is also observed, which may correspond to a complex geometry of the subducting IL. More specifically, the high-velocity body stretching from southern Tibet reaches further in the northwest direction at shallower depths (Figures 17b and 17c) and then turns northeastward at deeper depths (Figures 17d and 17e). Such features support the presence of a changing mode of Indo-Asian collision in the east-west direction (Zhao et al., 2010). Vertical cross-section profiles (Figure 17a) from western to the eastern Tibet are thus given in order to constrain the geometry of the above high-velocity anomaly and reveal more details of our new seismic model.

5.1. Western Tibet

Evidence of a large number of mantle earthquakes beneath the Pamir-Hindu Kush region and western Tibet has been reported by studies using focal depth solution (Chen & Yang, 2004; Roecker et al., 1980) and precise earthquake locations (Bai & Zhang, 2015; Pegler & Das, 1998). However, the occurring mechanism of these

Figure 19. 3-D downward (a) and northward (b) visualization of high-velocity and ultrahigh-velocity anomalies in the upper mantle beneath Tibet. The 3% and 6% isosurface of shear velocity perturbation at the depth from 100–400 km are represented by lighter blue and darker blue, respectively.

intermediate-depth earthquakes and their connection with the continental collision are still being debated (Kufner et al., 2016; Li et al., 2018). Hence, more detailed tomographic images are needed to provide new constraints.

A striking feature is that subhorizontal high-velocity (3%) and ultrahigh-velocity anomalies (6%) are imaged beneath western Tibet at depths ranging from 80 to 240 km (Figure 18). The high-velocity structures become thinner beneath the Anymaqen-Kunlun-Muztagh Suture and the crustal low-velocity anomaly correlates well with the crustal thickness (Profile C). Thus, we speculate that the geometry of 3% high-velocity body is basically controlled by the mantle lithosphere considering the lithospheric thickness of western China (Priestley et al., 2006). Another significant feature is the progressively increasing northern extent of ultrahigh-velocity anomaly from east (Profile C) to west (Profile A) in western Tibet, which have also been observed by previous studies (Li et al., 2008; Zhao et al., 2010). Therefore, the ultrahigh-velocity anomalies are probably associated with the cold IL beneath Tibet. Our new seismic model shows that the underthrusting IL may reach as far north as Pamir in the western Tibet (Figure 19).

We divide mantle earthquakes collected in western Tibet into two clusters, one located subhorizontally above underthrusting IL at depths ranging from 80 to 120 km beneath the West Kunlun Orogen (WKL) and the other subvertically at depths ranging from 120 to 200 km beneath Pamir (Profile A). Previous research indicated that mantle earthquakes under the WKL show no indication of either downdip extension or compression. These intermediate-depth earthquakes might not be connected to the remnants of subduction in the Hindu Kush or recent thrust faulting along the Pamir thrust belt (Chen & Yang, 2004). Hence, we suggest the subhorizontal mantle earthquake cluster beneath the WKL is associated to the underthrusting of IL beneath the Asian plate. In contrast, as seismicity and beach balls show a classic pattern of downdip extension (Chen & Yang, 2004) and a southward dipping trend (Bai & Zhang, 2015), deeper and subvertically distributed earthquakes beneath Pamir are more likely the product of the interaction between northward IL

and southward subducted Asian plate. We thus identify a dissimilar deformation model beneath Pamir and WKL from new seismic images and the spatial distribution of mantle earthquakes.

5.2. Central Tibet

East-west extension (Molnar et al., 1993) revealed by north-south trending rifts and normal faults as well as postcollisional magmatism (Figure 17a) featuring potassic, ultrapotassic, and adakitic volcanic rocks (Chung et al., 2005) are two keys in understanding the uplift history of the central Tibet. Recent research suggested a close link between the lithospheric structures and mantle dynamics of central Tibet and the evolution of surface tectonics (Chen et al., 2017; Zhang et al., 2013). However, there have been considerable controversy discussions on the upper-mantle structure of central Tibet, including debate over the morphology of subducting IL (e.g., Chen et al., 2017; Kosarev et al., 1999; Liang et al., 2012; Tilmann et al., 2003; Zhou & Murphy, 2005).

In contrast to the uniform thickness of underthrusting IL (6% contour) and lithosphere-related high-velocity (3% contour) in western Tibet, a thinned IL and a thickened high-velocity body (Figure 18) are observed beneath the Bangong-Nujiang Suture (BNS) in central Tibet. Convective removal (England, 1993) of thickened lithosphere has been proposed to account for both a rapid increase in the mean elevation of Tibet (Molnar et al., 1993) and the magmatism (Chung et al., 2005) in late Cenozoic time. Thus, IL thinning beneath the BNS observed in central Tibet might be crucial evidence for convective removal. Meanwhile, the Lhasa terrane, which lies north of Indus-Yarlung Suture and south of BNS, is characterized by an unusual deep high-velocity (3%) body down to 400 km. Receiver function (Kosarev et al., 1999; Tilmann et al., 2003) and traveltimes tomography (Li et al., 2008; Liang et al., 2016) research also observed a deep high-velocity body beneath Lhasa (DHVL), attributing it to the subvertical subduction of IL. Seismic anisotropy from shear wave splitting measurements indicated that the underthrusting IL would probably convert to subvertical subduction beneath the Lhasa terrane (Chen, Li, et al., 2015; Fu et al., 2008). Despite varying depths of the DHVL from west to east, its northern boundaries remain largely unchanged, which suggests that the DHVL is more likely a result of the northward subduction of the Indian mantle lithosphere beneath Lhasa after the convective removal (Figure 19) rather than the remnants of the convective removal of lithosphere (Chen et al., 2017). Therefore, the subvertical subduction of IL after the convective removal of thickened Tibetan lithosphere may account for the widespread magmatism and rapid uplift of the plateau in late Miocene.

5.3. Eastern Tibet

The complex and strong deformation of eastern Tibet is characterized by a rapid drop of topographic elevation from plateau height to normal height (Clark & Royden, 2000) as well as a prominent clockwise rotation of eastward extrusion of materials observed by fault plane solution (Molnar & Lyon-Caen, 1989) and GPS velocity (Wang et al., 2001). Accordingly, most catastrophic earthquakes in mainland China occurred in eastern Tibet (Figure 1), necessitating the timely investigation of the deformation in eastern Tibet and lateral growth of the plateau. Previous geophysical researches mainly focused on the crust-scale structure of eastern Tibet (e.g., Bai et al., 2010; Liu et al., 2014; Wang et al., 2007; Yao et al., 2008), while recent seismological studies indicated that seismic structure of the upper mantle might also inform the understanding of the kinematics and dynamics of the interaction between eastern Tibet and its surrounding blocks (Huang et al., 2015; Lei & Zhao, 2016). As these tomographical studies, based on ray theory, mainly focused on southeastern Tibet, it is important to consider the upper-mantle structure beneath the entire region of eastern Tibet for further investigation.

The geometry of underthrusting IL shows no significant change in further eastern Tibet without obvious thinning (Profiles G–L in Figure 18). And the horizontal distance of underthrusting IL is being much shorter than which in western Tibet. Furthermore, given that the geometry of the DHVL is profile-dependent, with no clear evidence of the DHVL east of 92° E (Profiles G, H, and I), and that it is consistent with a T-shaped high velocity zone beneath southern Tibet between 83°E and 92°E (Chen et al., 2017), we speculate that the eastern boundary of the vertical subducting IL is around 92°E.

Lateral mantle flow from northeastern Tibet to Alxa has been proposed according to the depth variation of the lithosphere-asthenosphere boundary by receiver function analysis (Shen et al., 2017; Xu et al., 2019). A thicker high-velocity layer beneath the Qaidam Basin contrasts sharply with a thinner high-velocity layer beneath the Qilianshan Orogen (QL) as well as Alxa (Profiles H, I, and J). In other words, the lithospheric mantle of the QL and Alxa is characterized by relatively low velocity, which is supported by observation of a prominent low-velocity anomaly underneath the QL from 125–200 km by surface-wave tomography

(Zhang et al., 2011). Thus, our new images suggest that the QL may act as a possible channel of lateral mantle flow from northeastern Tibet to Alxa should such flow exist.

Our new seismic model shows remarkable low velocity in the uppermost mantle and a high-velocity gap (Profile L) beneath the Longmenshan fault where several strong earthquakes occurred (Pei et al., 2019), including the Wenchuan earthquakes (M_w 7.9). Delamination or removal of lithospheric root driven by asthenospheric escape flow was introduced to explain the sudden rise of lithosphere–asthenosphere boundary from the SCB to eastern Tibet (Zhang et al., 2010). The new images confirm the delamination of lithosphere by a high-velocity gap (Figure 18) and show direct evidence of asthenospheric upwelling by the overlying low-velocity anomaly (Figure 18). The hot upwelling may account for the high heat flow revealed by drilling experiments (Zheng et al., 2016), indicating that the strong surface deformation of Longmenshan fault might be affected also by deep lithospheric dynamics.

6. Conclusion

In this study, we construct a full 3-D seismic model of the crust and upper mantle beneath the Tibetan Plateau based on a large-scale dense array using iterative finite-frequency tomography. The seismic model exhibits new constraints to continental collision dynamics with improved details. The high-velocity anomaly extends to Pamir in the western side, but it only reaches the Bangong–Nujiang Suture in central and eastern Tibet. We also observed an obvious thinning of the Indian lithosphere beneath northern Lhasa in central Tibet, which provides crucial evidence for convective removal. A subvertical high-velocity structure beneath Lhasa penetrating into the mantle transition zone is confirmed by the new images and is likely due to high-angle subduction of the Indian lithosphere after the convective removal. A high-velocity gap overriden by a low-velocity anomaly underneath the Longmenshan fault indicates that lithospheric delamination along with asthenospheric upwelling might occur beneath the transition zone between eastern Tibet and the SCB. Hence, dynamics of deeper lithosphere may also contribute to the strong deformation of surface.

References

- Bai, D., Unsworth, M. J., Meju, M. A., Ma, X., Teng, J., Kong, X., et al. (2010). Crustal deformation of the eastern Tibetan Plateau revealed by magnetotelluric imaging. *Nature Geoscience*, 3, 358–362. <https://doi.org/10.1038/ngeo830>
- Bai, L., & Zhang, T. (2015). Complex deformation pattern of the Pamir–Hindu Kush region inferred from multi-scale double-difference earthquake relocations. *Tectonophysics*, 638, 177–184. <https://doi.org/10.1016/j.tecto.2014.11.006>
- Bao, X., Song, X., & Li, J. (2015). High-resolution lithospheric structure beneath mainland China from ambient noise and earthquake surface-wave tomography. *Earth and Planetary Science Letters*, 417, 132–141. <https://doi.org/10.1016/j.epsl.2015.02.024>
- Baron, J., & Morelli, A. (2017). Full-waveform seismic tomography of the Vrancea, Romania, subduction region. *Physics of the Earth and Planetary Interiors*, 273, 36–49. <https://doi.org/10.1016/j.pepi.2017.10.009>
- Bozdağ, E., Peter, D., Lefebvre, M., Komatitsch, D., Tromp, J., Hill, J., et al. (2016). Global adjoint tomography: First-generation model. *Geophysical Journal International*, 207, 1739–1766. <https://doi.org/10.1093/gji/ggw356>
- Chen, P., Jordan, T. H., & Zhao, L. (2007). Full three-dimensional tomography: A comparison between the scattering-integral and adjoint-wavefield methods. *Geophysical Journal International*, 170, 175–181. <https://doi.org/10.1111/j.1365-246X.2007.03429.x>
- Chen, Y., Li, W., Yuan, X., Badal, J., & Teng, J. (2015). Tearing of the Indian lithospheric slab beneath southern Tibet revealed by SKS-wave splitting measurements. *Earth and Planetary Science Letters*, 413, 13–24. <https://doi.org/10.1016/j.epsl.2014.12.041>
- Chen, M., Niu, F., Liu, Q., Tromp, J., & Zheng, X. (2015). Multiparameter adjoint tomography of the crust and upper mantle beneath East Asia: 1. Model construction and comparisons. *Journal of Geophysical Research: Solid Earth*, 120, 1762–1786. <https://doi.org/10.1002/2014JB011638>
- Chen, M., Niu, F., Tromp, J., Lenardic, A., Lee, C.-T. A., Cao, W., & Ribeiro, J. (2017). Lithospheric foundering and underthrusting imaged beneath Tibet. *Nature Communications*, 8, 15659. <https://doi.org/10.1038/ncomms15659>
- Chen, W. P., & Yang, Z. (2004). Earthquakes beneath the Himalayas and Tibet: Evidence for strong lithospheric mantle. *Science*, 304, 1949–1952. <https://doi.org/10.1126/science.1097324>
- Chung, S.-L., Chu, M.-F., Zhang, Y., Xie, Y., Lo, C.-H., Lee, T.-Y., et al. (2005). Tibetan tectonic evolution inferred from spatial and temporal variations in post-collisional magmatism. *Earth-Science Reviews*, 68, 173–196. <https://doi.org/10.1016/j.earscirev.2004.05.001>
- Clark, M. K., & Royden, L. H. (2000). Topographic ooze: Building the eastern margin of Tibet by lower crustal flow. *Geology*, 28, 703. [https://doi.org/10.1130/0091-7613\(2000\)28<703:TOBTEM>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<703:TOBTEM>2.0.CO;2)
- Curtis, A., & Woodhouse, J. H. (1997). Crust and upper mantle shear velocity structure beneath the Tibetan Plateau and surrounding regions from interevent surface wave phase velocity inversion. *Journal of Geophysical Research*, 102, 11,789–11,813. <https://doi.org/10.1029/96JB03182>
- Dahlen, F. A., Hung, S.-H., & Nolet, G. (2000). Fréchet kernels for finite-frequency traveltimes—I Theory. *Geophysical Journal International*, 141, 157–174.
- England, P. (1993). Convective removal of thermal boundary layer of thickened continental lithosphere: A brief summary of causes and consequences with special reference to the Cenozoic tectonics of the Tibetan Plateau and surrounding regions. *Tectonophysics*, 223, 67–73. [https://doi.org/10.1016/0040-1951\(93\)90158-G](https://doi.org/10.1016/0040-1951(93)90158-G)
- Fichtner, A., Kennett, B. L. N., Igel, H., & Bunge, H.-P. (2009). Full seismic waveform tomography for upper-mantle structure in the Australasian region using adjoint methods. *Geophysical Journal International*, 179, 1703–1725. <https://doi.org/10.1111/j.1365-246X.2009.04368.x>

Acknowledgments

We would like to thank two reviewers and the Editor for their constructive comments, which help improve the manuscript greatly. This research was supported by NSFC Grant 41730212, National Key R&D Project of China (2017YFC1500304 and 2016YFC0600302), and Project supported by the Foundation of Director of Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, China (2019SZJJ-05). The waveform data were provided by China Seismic Array Data Management Center (<http://doi.org/10.17616/R31NJMMK>) at Institute of Geophysics, China Earthquake Administration, and the Data Management Center of IRIS with network code IU, IC, KN, KR, KZ, MY, TW, TM, etc. The open source spectral-element software package SPECFEM3DGlobe and the seismic waveform measurement software package FLEXWIN used for this article are freely available for download via the Computational Infrastructure for Geodynamics (CIG; geodynamics.org). For this study, we have used the computer systems of the Earthquake and Volcano Information Center of the Earthquake Research Institute, the University of Tokyo, and the Super Computation of Tianhe-2. The first author gratefully acknowledges financial support from China Scholarship Council. Figures were made with the GMT (Wessel et al., 2013) and the IRIS EMC 3-D visualization tool. All the data, codes, and 3-D models obtained through this study are available online (https://osf.io/kra7y/?view_only=a74c258f5778450f96d9db7dc6f85887). N. F. was invited to Earthquake Research Institute, University of Tokyo, for 2 months during the summer 2018 and has worked on this project with Z. X. and T. I.

- Fichtner, A., & Trampert, J. (2011). Resolution analysis in full waveform inversion: Resolution in full waveform inversion. *Geophysical Journal International*, 187, 1604–1624. <https://doi.org/10.1111/j.1365-246X.2011.05218.x>
- Fu, Y. V., Chen, Y. J., Li, A., Zhou, S., Liang, X., Ye, G., et al. (2008). Indian mantle corner flow at southern Tibet revealed by shear wave splitting measurements. *Geophysical Research Letters*, 35, L02308. <https://doi.org/10.1029/2007GL031753>
- Fuji, N., Chevrot, S., Zhao, L., Geller, R. J., & Kawai, K. (2012). Finite-frequency structural sensitivities of short-period compressional body waves: 3-D Fréchet kernels for high frequencies. *Geophysical Journal International*, 190, 522–540. <https://doi.org/10.1111/j.1365-246X.2012.05495.x>
- Gao, R., Lu, Z., Klempner, S. L., Wang, H., Dong, S., Li, W., & Li, H. (2016). Crustal-scale duplexing beneath the Yarlung Zangbo suture in the western Himalaya. *Nature Geoscience*, 9, 555–560. <https://doi.org/10.1038/ngeo2730>
- Geller, R. J., & Hara, T. (1993). Two efficient algorithms for iterative linearized inversion of seismic waveform data. *Geophysical Journal International*, 115, 699–710. <https://doi.org/10.1111/j.1365-246X.1993.tb01488.x>
- Hjörkefúsdóttir, V., & Ekström, G. (2010). Effects of three-dimensional Earth structure on CMT earthquake parameters. *Physics of the Earth and Planetary Interiors*, 179, 178–190. <https://doi.org/10.1016/j.pepi.2009.11.003>
- Huang, Z., Wang, P., Xu, M., Wang, L., Ding, Z., Wu, Y., et al. (2015). Mantle structure and dynamics beneath SE Tibet revealed by new seismic images. *Earth and Planetary Science Letters*, 411, 100–111. <https://doi.org/10.1016/j.epsl.2014.11.040>
- Huang, J., & Zhao, D. (2006). High-resolution mantle tomography of China and surrounding regions. *Journal of Geophysical Research*, 111, B09305. <https://doi.org/10.1029/2005JB004066>
- Kawai, K., & Geller, R. J. (2010). Waveform inversion for localized seismic structure and an application to d' structure beneath the Pacific. *Journal of Geophysical Research*, 115, B01305. <https://doi.org/10.1029/2009JB006503>
- Kind, R., Yuan, X., Saul, J., Nelson, D., Soboleva, S. V., Mechie, J., et al. (2002). Seismic images of crust and upper mantle beneath Tibet: Evidence for Eurasian plate subduction. *Science*, 298, 1219–1221. <https://doi.org/10.1126/science.1078115>
- Komatitsch, D., & Tromp, J. (2002a). Spectral-element simulations of global seismic wave propagation-I. Validation. *Geophysical Journal International*, 149, 390–412. <https://doi.org/10.1046/j.1365-246X.2002.01653.x>
- Komatitsch, D., & Tromp, J. (2002b). Spectral-element simulations of global seismic wave propagation-II. Three-dimensional models, oceans, rotation and self-gravitation. *Geophysical Journal International*, 150, 303–318. <https://doi.org/10.1046/j.1365-246X.2002.01716.x>
- Konishi, K., Kawai, K., Geller, R. J., & Fuji, N. (2014). Waveform inversion for localized three-dimensional seismic velocity structure in the lowermost mantle beneath the western Pacific. *Geophysical Journal International*, 199, 1245–1267. <https://doi.org/10.1093/gji/ggu288>
- Kosarev, G., Kind, R., Soboleva, S. V., Yuan, X., Hanka, W., & Oreshin, S. (1999). Seismic evidence for a detached Indian lithospheric mantle beneath Tibet. *Science*, 283, 1306–1309. <https://doi.org/10.1126/science.283.5406.1306>
- Kufner, S.-K., Schurr, B., Sippl, C., Yuan, X., Ratschbacher, L., Akbar, A. S., of Mohammad, et al. (2016). Deep India meets deep Asia: Lithospheric indentation, delamination and break-off under Pamir and Hindu Kush (Central Asia). *Earth and Planetary Science Letters*, 435, 171–184. <https://doi.org/10.1016/j.epsl.2015.11.046>
- Kustowski, B., Ekström, G., & Dziewoski, A. M. (2008). Anisotropic shear-wave velocity structure of the Earth's mantle: A global model. *Journal of Geophysical Research*, 113, B06306. <https://doi.org/10.1029/2007JB005169>
- Laske, G., Masters, G., Ma, Z., & Pasyanos, M. (2013). Update on CRUST1.0—A 1-degree Global Model of Earth's Crust. In *EGU General Assembly*. <https://doi.org/10.6092/1970-9870/128>
- Lebedev, S., & van der Hilst, R. D. (2008). Global upper-mantle tomography with the automated multimode inversion of surface and S-wave forms. *Geophysical Journal International*, 173, 505–518. <https://doi.org/10.1111/j.1365-246X.2008.03721.x>
- Lei, J., & Zhao, D. (2016). Teleseismic P-wave tomography and mantle dynamics beneath eastern Tibet. *Geochemistry, Geophysics, Geosystems*, 17, 1861–1884. <https://doi.org/10.1002/2016GC006262>
- Li, W., Chen, Y., Yuan, X., Schurr, B., Mechie, J., Oimahmadov, I., & Fu, B. (2018). Continental lithospheric subduction and intermediate-depth seismicity: Constraints from S-wave velocity structures in the Pamir and Hindu Kush. *Earth and Planetary Science Letters*, 482, 478–489. <https://doi.org/10.1016/j.epsl.2017.11.031>
- Li, C., van der Hilst, R. D., Meltzer, A. S., & Engdahl, E. R. (2008). Subduction of the Indian lithosphere beneath the Tibetan Plateau and Burma. *Earth and Planetary Science Letters*, 274, 157–168.
- Liang, X., Chen, Y., Tian, X., Chen, Y. J., Ni, J., Gallegos, A., et al. (2016). 3D imaging of subducting and fragmenting Indian continental lithosphere beneath southern and central Tibet using body-wave finite-frequency tomography. *Earth and Planetary Science Letters*, 443, 162–175. <https://doi.org/10.1016/j.epsl.2016.03.029>
- Liang, X., Sandvol, E., Chen, Y. J., Hearn, T., Ni, J., Klempner, S., et al. (2012). A complex Tibetan upper mantle: A fragmented Indian slab and no south-verging subduction of Eurasian lithosphere. *Earth and Planetary Science Letters*, 333–334, 101–111. <https://doi.org/10.1016/j.epsl.2012.03.036>
- Liang, C., & Song, X. (2006). A low velocity belt beneath northern and eastern Tibetan Plateau from Pn tomography. *Geophysical Research Letters*, 33, L22306. <https://doi.org/10.1029/2006GL027926>
- Liu, M., Cui, X., & Liu, F. (2004). Cenozoic rifting and volcanism in eastern China: A mantle dynamic link to the Indo-Asian collision? *Tectonophysics*, 393, 29–42. <https://doi.org/10.1016/j.tecto.2004.07.029>
- Liu, Q. Y., & Gu, Y. J. (2012). Seismic imaging: From classical to adjoint tomography. *Tectonophysics*, 566–567, 31–66. <https://doi.org/10.1016/j.tecto.2012.07.006>
- Liu, Q. Y., van der Hilst, R. D., Li, Y., Yao, H. J., Chen, J. H., Guo, B., et al. (2014). Eastward expansion of the Tibetan Plateau by crustal flow and strain partitioning across faults. *Nature Geoscience*, 7, 361–365. <https://doi.org/10.1038/ngeo2130>
- Luo, Y., Modrak, R., & Tromp, J. (2013). Strategies in adjoint tomography, (2nd ed.). In W. Freedon, Z. Nahed, & T. Sonar (Eds.), *Handbook of geomathematics* (pp. 1943–2001). Berlin Heidelberg: Springer.
- Luo, Y., Tromp, J., Denel, B., & Calandra, H. (2013). 3D coupled acoustic-elastic migration with topography and bathymetry based on spectral-element and adjoint methods. *Geophysics*, 78, S193–S202. <https://doi.org/10.1190/geo2012-0462.1>
- Maggi, A., Tape, C., Chen, M., Chao, D., & Tromp, J. (2009). An automated time-window selection algorithm for seismic tomography. *Geophysical Journal International*, 178, 257–281. <https://doi.org/10.1111/j.1365-246X.2009.04099.x>
- Molnar, P., England, P., & Martinod, J. (1993). Mantle dynamics, uplift of the Tibetan Plateau, and the Indian monsoon. *Reviews of Geophysics*, 31, 357. <https://doi.org/10.1029/93RG02030>
- Molnar, P., & Lyon-Caen, H. (1989). Fault plane solutions of earthquakes and active tectonics of the Tibetan Plateau and its margins. *Geophysical Journal International*, 99, 123–154. <https://doi.org/10.1111/j.1365-246X.1989.tb02020.x>
- Molnar, P., & Tapponnier, P. (1975). Cenozoic tectonics of Asia: Effects of a continental collision: Features of recent continental tectonics in Asia can be interpreted as results of the India-Eurasia collision. *Science*, 189, 419–426. <https://doi.org/10.1126/science.189.4201.419>
- Mora, P. (1987). Nonlinear two-dimensional elastic inversion of multioffset seismic data. *Geophysics*, 52, 1211–1228. <https://doi.org/10.1190/1.1442384>

- Nabelek, J., Hetenyi, G., Vergne, J., Sapkota, S., Kafle, B., Jiang, M., et al. (2009). Underplating in the Himalaya-Tibet collision zone revealed by the hi-CLIMB experiment. *Science*, 325, 1371–1374. <https://doi.org/10.1126/science.1167719>
- Ni, J., & Barazangi, M. (1983). High-frequency seismic wave propagation beneath the Indian shield, Himalayan arc, Tibetan Plateau and surrounding regions: High uppermost mantle velocities and efficient s propagation beneath Tibet. *Geophysical Journal International*, 72, 665–689. <https://doi.org/10.1111/j.1365-246X.1983.tb02826.x>
- Pegler, G., & Das, S. (1998). An enhanced image of the Pamir-Hindu Kush seismic zone from relocated earthquake hypocenters: Pamir-Hindu Kush seismic zone. *Geophysical Journal International*, 134, 573–595. <https://doi.org/10.1046/j.1365-246x.1998.00582.x>
- Pei, S., Niu, F., Ben-Zion, Y., Sun, Q., Liu, Y., Xue, X., et al. (2019). Seismic velocity reduction and accelerated recovery due to earthquakes on the Longmenshan fault. *Nature Geoscience*, 12, 387–392. <https://doi.org/10.1038/s41561-019-0347-1>
- Priestley, K., Debayle, E., McKenzie, D., & Pilidou, S. (2006). Upper mantle structure of eastern Asia from multimode surface waveform tomography. *Journal of Geophysical Research*, 111, B10304. <https://doi.org/10.1029/2005JB004082>
- Rawlinson, N., Fichtner, A., Sambridge, M., & Young, M. K. (2014). Seismic tomography and the assessment of uncertainty. *Advances in Geophysics*, 55, 1–76. <https://doi.org/10.1016/bs.agph.2014.08.001>
- Roecker, S. W., Soboleva, O. V., Nersisov, I. L., Lukk, A. A., Hatzfeld, D., Chatelain, J. L., & Molnar, P. (1980). Seismicity and fault plane solutions of intermediate depth earthquakes in the Pamir-Hindu Kush region. *Journal of Geophysical Research: Solid Earth*, 85, 1358–1364. <https://doi.org/10.1029/JB085iB03p01358>
- Royden, L. H., Burchfiel, B. C., King, R. W., Wang, E., Chen, Z., Shen, F., & Liu, Y. (1997). Surface deformation and lower crustal flow in eastern Tibet. *Science*, 276, 788–790. <https://doi.org/10.1126/science.276.5313.788>
- Royden, L. H., Burchfiel, B. C., & van der Hilst, R. D. (2008). The geological evolution of the Tibetan Plateau. *Science*, 321, 1054–1058. <https://doi.org/10.1126/science.1155371>
- Schaeffer, A. J., & Lebedev, S. (2013). Global shear speed structure of the upper mantle and transition zone. *Geophysical Journal International*, 194, 417–449. <https://doi.org/10.1093/gji/ggt095>
- Shen, X., Liu, M., Gao, Y., Wang, W., Shi, Y., An, M., et al. (2017). Lithospheric structure across the northeastern margin of the Tibetan Plateau: Implications for the plateau's lateral growth. *Earth and Planetary Science Letters*, 459, 80–92. <https://doi.org/10.1016/j.epsl.2016.11.027>
- Shen, W., Ritzwoller, M. H., Kang, D., Kim, Y., Lin, F.-C., Ning, J., et al. (2016). A seismic reference model for the crust and uppermost mantle beneath China from surface wave dispersion. *Geophysical Journal International*, 206, 954–979. <https://doi.org/10.1093/gji/ggw175>
- Tao, K., Grand, S. P., & Niu, F. (2018). Seismic structure of the upper mantle beneath eastern Asia from full waveform seismic tomography. *Geochemistry, Geophysics, Geosystems*, 19, 2732–2763. <https://doi.org/10.1029/2018GC007460>
- Tape, C., Liu, Q., Maggi, A., & Tromp, J. (2009). Adjoint tomography of the Southern California crust. *Science*, 325, 988–992. <https://doi.org/10.1126/science.1175298>
- Tape, C., Liu, Q., & Tromp, J. (2007). Finite-frequency tomography using adjoint methods—Methodology and examples using membrane surface waves. *Geophysical Journal International*, 168, 1105–1129. <https://doi.org/10.1111/j.1365-246X.2006.03191.x>
- Tapponnier, P., Xu, Z., Francoise, R., Bertrand, M., Nicolas, A., Grard, W., & Yang, J. (2001). Oblique stepwise rise and growth of the Tibet Plateau. *Science*, 294, 1671–1677. <https://doi.org/10.1126/science.105978>
- Tarantola, A. (1984). Inversion of seismic reflection data in the acoustic approximation. *Geophysics*, 49, 1259–1266. <https://doi.org/10.1190/1.1441754>
- Tilmann, F., Ni, J., & Seismic Team, INDEPTHIII (2003). Seismic imaging of the downwelling Indian lithosphere beneath central Tibet. *Science*, 300, 1424–1427. <https://doi.org/10.1126/science.1082777>
- Tromp, J., Tape, C., & Liu, Q. (2005). Seismic tomography, adjoint methods, time reversal and banana-doughnut kernels. *Geophysical Journal International*, 160, 195–216. <https://doi.org/10.1111/j.1365-246X.2004.02453.x>
- Wang, C.-Y., Han, W.-B., Wu, J.-P., Lou, H., & Chan, W. W. (2007). Crustal structure beneath the eastern margin of the Tibetan Plateau and its tectonic implications. *Journal of Geophysical Research*, 112, B07307. <https://doi.org/10.1029/2005JB003873>
- Wang, K., Yang, Y., Basini, P., Tong, P., Tape, C., & Liu, Q. (2018). Refined crustal and uppermost mantle structure of Southern California by ambient noise adjoint tomography. *Geophysical Journal International*, 215, 844–863. <https://doi.org/10.1093/gji/ggy312>
- Wang, Q., Zhang, P., Freymueller, J. T., Bilham, R., Larson, K. M., Lai, X., et al. (2001). Present-day crustal deformation in China constrained by global positioning system measurements. *Science*, 294, 574–577. <https://doi.org/10.1126/science.1063647>
- Wessel, P., Smith, W. H. F., Scharroo, R., Luis, J., & Wobbe, F. (2013). Generic mapping tools: Improved version released. *Eos Trans. AGU*, 94, 409–410. <https://doi.org/10.1002/2013EO450001>
- Xu, Q., Pei, S., Yuan, X., Zhao, J., Liu, H., Tu, H., & Chen, S. (2019). Seismic evidence for lateral asthenospheric flow beneath the north-eastern Tibetan Plateau derived from s receiver functions. *Geochemistry, Geophysics, Geosystems*, 883–894. <https://doi.org/10.1029/2018GC007986>
- Xu, L., Rondenay, S., & van der Hilst, R. D. (2007). Structure of the crust beneath the southeastern Tibetan Plateau from teleseismic receiver functions. *Physics of the Earth and Planetary Interiors*, 165, 176–193. <https://doi.org/10.1016/j.pepi.2007.09.002>
- Yao, H., Beghein, C., & van der Hilst, R. D. (2008). Surface wave array tomography in SE Tibet from ambient seismic noise and two-station analysis - II. Crustal and upper-mantle structure. *Geophysical Journal International*, 173, 205–219. <https://doi.org/10.1111/j.1365-246x.2007.03696.x>
- Yao, H., van der Hilst, R. D., & de Hoop, M. V. (2006). Surface-wave array tomography in SE Tibet from ambient seismic noise and two-station analysis - I. Phase velocity maps. *Geophysical Journal International*, 166, 732–744. <https://doi.org/10.1111/j.1365-246X.2006.03028.x>
- Yin, A., & Harrison, T. M. (2000). Geologic evolution of the Himalayan-Tibetan orogen. *Annual Review of Earth and Planetary Sciences*, 28, 211–280. <https://doi.org/10.1146/annurev.earth.28.1.211>
- Zhang, Z., Chen, Y., Yuan, X., Tian, X., Klemperer, S. L., Xu, T., et al. (2013). Normal faulting from simple shear rifting in south Tibet, using evidence from passive seismic profiling across the Yadong-Gulu rift. *Tectonophysics*, 606, 178–186. <https://doi.org/10.1016/j.tecto.2013.03.019>
- Zhang, Q., Sandvol, E., Ni, J., Yang, Y., & Chen, Y. J. (2011). Rayleigh wave tomography of the northeastern margin of the Tibetan Plateau. *Earth and Planetary Science Letters*, 304, 103–112. <https://doi.org/10.1016/j.epsl.2011.01.021>
- Zhang, Z., Yuan, X., Chen, Y., Tian, X., Kind, R., Li, X., & Teng, J. (2010). Seismic signature of the collision between the east Tibetan escape flow and the Sichuan Basin. *Earth and Planetary Science Letters*, 292, 254–264. <https://doi.org/10.1016/j.epsl.2010.01.046>
- Zhao, L., & Chevrot, S. (2011). An efficient and flexible approach to the calculation of three-dimensional full-wave Fréchet kernels for seismic tomography—I. Theory. *Geophysical Journal International*, 185, 922–938. <https://doi.org/10.1111/j.1365-246x.2011.04983.x>
- Zhao, B., Huang, Y., Zhang, C., Wang, W., Tan, K., & Du, R. (2015). Crustal deformation on the Chinese mainland during 1998–2014 based on GPS data. *Geodesy and Geodynamics*, 6, 7–15. <https://doi.org/10.1016/j.geog.2014.12.006>

- Zhao, W., Kumar, P., Mechie, J., Kind, R., Meissner, R., Wu, Z., et al. (2011). Tibetan plate overriding the Asian plate in central and northern Tibet. *Nature Geoscience*, 4, 870–873. <https://doi.org/10.1038/ngeo1309>
- Zhao, J., Yuan, X., Liu, H., Kumar, P., Pei, S., Kind, R., et al. (2010). The boundary between the Indian and Asian tectonic plates below Tibet. *Proceedings of the National Academy of Sciences*, 107, 11,229–11,233. <https://doi.org/10.1073/pnas.1001921107>
- Zheng, Y., Li, H., & Gong, Z. (2016). Geothermal study at the Wenchuan earthquake fault Scientific Drilling Project-Hole 1 (WFSD-1): Borehole temperature, thermal conductivity, and well log data. *Journal of Asian Earth Sciences*, 117, 23–32. <https://doi.org/10.1016/j.jseas.2015.11.025>
- Zhou, Y., Dahlen, F. A., & Nolet, G. (2004). Three-dimensional sensitivity kernels for surface wave observables. *Geophysical Journal International*, 158, 142–168. <https://doi.org/10.1111/j.1365-246X.2004.02324.x>
- Zhou, H., & Murphy, M. A. (2005). Tomographic evidence for wholesale underthrusting of India beneath the entire Tibetan Plateau. *Journal of Asian Earth Sciences*, 25, 445–457. <https://doi.org/10.1016/j.jseas.2004.04.007>
- Zhu, H. (2018). Crustal wave speed structure of north Texas and Oklahoma based on ambient noise cross-correlation functions and adjoint tomography. *Geophysical Journal International*, 214, 716–730. <https://doi.org/10.1093/gji/ggy169>
- Zhu, H., Bozda, E., Peter, D., & Tromp, J. (2012). Structure of the European upper mantle revealed by adjoint tomography. *Nature Geoscience*, 5, 493–498. <https://doi.org/10.1038/ngeo1501>
- Zhu, H., Bozda, E., & Tromp, J. (2015). Seismic structure of the European upper mantle based on adjoint tomography. *Geophysical Journal International*, 201, 18–52. <https://doi.org/10.1093/gji/ggu492>
- Zhu, H., & Tromp, J. (2013). Mapping tectonic deformation in the crust and upper mantle beneath Europe and the North Atlantic Ocean. *Science*, 341, 871–875. <https://doi.org/10.1126/science.1241335>