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RESEARCH ARTICLE

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Key Points:

- New seismic data recorded by 26 broadband seismic stations are used to image the crust and mantle structure beneath northern Sumatra
- Results show that slab morphology beneath northern Sumatra is characterized by slab curving, tearing, deflection and dip angle variations
- A slab tear is revealed at a depth of about 120 km, which is one of the main factors for the strong low anomalies in the uppermost mantle

Supporting Information:

- Supporting Information S1

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Slab Morphology Beneath Northern Sumatra Revealed by Regional and Teleseismic Traveltime Tomography

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Abstract An eikonal equation-based tomography method is used to invert the high-quality regional and teleseismic traveltimes data recorded by 26 broadband seismic stations in northern Sumatra, following which we obtain the *P* and *S* wave velocity structures of the crust and mantle down to a depth of 800 km. The results of both *P* and *S* wave tomography clearly show that the Indo-Australian oceanic plate continuously penetrates downward beneath northern Sumatra and the maximum penetration depth varies approximately from 400 km at the northern tip of Sumatra to about 800 km around the southern boundary of our study area. Significant slab folding or bending reported in the literature as the main feature of the subducted slab beneath northern Sumatra is not found in our results. Instead, our tomographic images demonstrate only a less curved slab that mimics the shape of the Sunda Trench and volcanic arc and generally extends over depths from 120 to 450 km. *P* wave tomography shows broad and pronounced low-velocity anomalies beneath the island of Sumatra in the lower crust and uppermost mantle. Our model also reveals a slab tear approximately at 120-km depth, which has been documented in previous studies and considered to be related to the eruption of the Toba supervolcano.

1. Introduction

The island of Sumatra in western Indonesia is located along the convergent boundary between the Indo-Australian and Eurasian plates (Figure 1). The oceanic Indo-Australian lithosphere subducts obliquely beneath Sumatra with an average convergence rate of 5.7 cm/year (Lange et al., 2010; Sieh & Natawidjaja, 2000). Based on plate tectonic reconstructions, Sumatra was part of continental Asia by the end of the Triassic (Hall, 2012). A Jurassic-Cretaceous intraoceanic arc and accretionary complex that separated from the northern margin of Gondwanaland was sutured onto Sumatra, which was coincident with the closure of the Tethyan Ocean during the early Late Cretaceous (Barber, 2000; Hall, 2012). Australia remained close to Antarctica between 90 and 45 Ma, when subduction beneath Sumatra ceased (Hall, 2009). During this interval, Sundaland was largely surrounded by passive margins characterized by some strike-slip deformation and extension. When Australia began to move northward at approximately 45 Ma, subduction resumed along the Sunda Trench (Hall, 2009). However, the radiometric dates in Rock et al. (1982) show a likely decrease in volcanic activity throughout central Sumatra from 30 to 8.5 Ma, which indicates that slab subduction was probably slow during that period. According to an investigation of Neogene strata exposed on Nias Island (Moore et al., 1980), the present-day fast subduction in northern Sumatra probably began in the late Miocene.

The subducted slab beneath northern Sumatra is characterized by several key features. For example, the Indo-Australian plate, which has a relatively young age varying from the Late Cretaceous to the late Eocene (Carton et al., 2014), subducts obliquely beneath northern Sumatra together with the Investigator Fracture Zone (IFZ) and Wharton Fossil Ridge (Pesicek et al., 2008). The active tectonic processes in northern Sumatra, such as the intense seismicity in the forearc area (Lange et al., 2018; Pesicek, Thurber, Zhang, 2010), the cyclicity of great earthquakes (Lange et al., 2010; Prawirodirdjo et al., 1997), the

Figure 1. Tectonic setting of the study area. The inset map shows the location of the study area (red rectangle) in Southeast Asia. The Indo-Australia oceanic plate converges along the Sunda Trench at an average rate of 57 mm/year. The white arrow in the Indian Ocean shows the plate motion relative to the Eurasia plate. The red triangles denote volcanoes, while the black inverted triangles represent the 26 broadband stations used in this study, including 22 stations deployed in northern Sumatra, 3 in Malaysia, and 1 in Singapore. The Sumatra Fault System (SFS) is marked by dark red lines. The barbed orange curve is the Sunda Trench. The four dashed purple lines represent the four subridges of the Investigator Fracture Zone (IFZ) near the trench. The position of the IFZ is after Koulakov et al. (2016). The numbers in gray squares denote the age of the ocean floor along the trench (Carton et al., 2014). The purple stars to the north of Simeulue Island and near the Banyak Islands mark the locations of the Great Sumatra-Andaman earthquake on 26 December 2004 and the Nias earthquake on 28 March 2004, respectively. The black arrow near Toba Caldera indicates Samosir Island. WFR = Wharton Fossil Ridge.

episodes of volcanic activities (Koulakov et al., 2016; Masturyono et al., 2001; Ninkovich et al., 1978), and the development of the Sumatra Fault System (SFS) (Nishimura et al., 1984; Page et al., 1979; Sieh & Natawidjaja, 2000), are all closely related to the subducted slab.

Over the past four decades, various slab models have been proposed to interpret the subduction of the slab beneath northern Sumatra. Ninkovich (1976) constructed a laterally continuous slab model and inferred a clockwise rotation of Sumatra from ~10 to 4–5 Ma around a rotation pole located either in or near the Sunda Strait. In this model, because of clockwise rotation that reduces the flow in the vertical direction, the penetration depth of the subducted slab decreases from the north to the south. In addition, the maximum penetration depth of the subducted slab was considered to be consistent with the depth of the Wadati-Benioff zone, which decreases from about 600 km to the east of the Sunda Strait to about 200 km beneath the northern tip of Sumatra (Ninkovich, 1976). Widiyantoro and van der Hilst (1996, 1997) developed a similar laterally continuous slab model in the upper mantle by using seismic traveltime tomography. The main difference between the two models is that the maximum

depth of the subducted slab of the latter model is about 200 km deeper than that of the former one. In addition, Widiyantoro and van der Hilst (1996, 1997) showed a detached slab in the lower mantle beneath northern Sumatra. In contrast, based on tomographic images of the mantle beneath Indonesia, Puspito et al. (1993) reported a laterally discontinuous slab model that does not contain any evidence of a detached slab in the lower mantle. Bijwaard et al. (1998) presented a global tomographic model comprising a laterally continuous slab with a maximum subducted depth of ~800 km, which is quite different from the features of the models mentioned above. Almost one decade ago, Pesicek et al. (2010) Pesicek, Thurber, Widiyantoro, et al. (2010) utilized the same tomography technique as Widiyantoro and van der Hilst (1996, 1997) but with a larger seismic dataset enhanced by the 2004–2005 Sumatra-Andaman earthquake sequences. A laterally and vertically continuous slab model was obtained. In contrast to previous models, this slab model exhibits a pronounced SW-NE-oriented fold extending over the depths from 190 to 530 km beneath northern Sumatra. A similar slab model with folding was proposed by Hall and Spakman (2015), but the curvature of the folded slab is relatively small when compared with the model of Pesicek et al. (2008); Pesicek, Thurber, Widiyantoro, et al. (2010). Recently, Liu et al. (2018) developed a slab model with even gentler folding by using teleseismic traveltime tomography.

It is obvious that the debate on the morphology of the slab beneath northern Sumatra is still ongoing. To resolve this dispute, we apply an eikonal equation-based traveltime tomography method (Liu et al., 2018; Tong et al., 2019) to image the crust and mantle structure beneath northern Sumatra by inverting high-quality regional and teleseismic traveltime data from the Agency for Meteorology, Climatology and Geophysics of Indonesia (BMKG). After relocating the regional earthquakes in an initial 3-D model, the regional data are utilized to image the structures of the crust and uppermost mantle. Then, we invert the teleseismic traveltimes in a 3-D model with well-resolved crust and uppermost mantle structures provided by the inversion of regional seismic data. Finally, we obtain the *P* and *S* wave velocity structures beneath northern Sumatra at the depths ranging from 20 to 800 km. A new model is proposed to describe the morphology of the subducted slab, which may provide fundamental information for future tectonic reconstructions of SE Asia.

2. Data and Methodology

2.1. Data Selection

Indonesia has been working on the establishment of a tsunami early warning system since 2005. This system, maintained by the BMKG, consists of more than 160 broadband seismic stations scattered throughout the country (Lauterjung & Letz, 2017). In northern Sumatra (Figure 1), there are 22 BMKG-operated seismic stations, and most of them have been in operation since 2009. To enhance the data coverage, we also include three stations in Malaysia and one station in Singapore (Figure 1).

All the regional earthquakes selected in this study have a magnitude *M* greater than 2.5 and occurred in the period from January 2009 to May 2019. The earthquake list was first obtained from the International Seismological Centre event catalog (<http://www.isc.as.uk>). Then the seismograms of these earthquakes were downloaded from the BMKG seismic network. The waveforms were preprocessed by removing the mean and trend, following which they were bandpass filtered over the frequency range of 1.5–5 Hz. Subsequently, the first *P* and *S* wave arrivals were manually picked by two individuals. To ensure robust *P* wave arrivals, the traveltime difference between the two individuals must be less than 0.2 s, and the average of the two individually picked arrival times was treated as the observed arrival time. In addition, we required that each selected earthquake has at least seven picks. Finally, a total of 9,004 first *P* wave arrivals from 1,005 regional earthquakes were selected (Figure 2a). The same criteria were applied to process regional *S* wave data, but the traveltime difference between the two individuals for each first *S* wave arrival less than 0.3 s was accepted. In the end, a total of 4,498 first *S* wave arrivals from 528 regional earthquakes were selected (Figure 2c).

From the Incorporated Research Institutions for Seismology (IRIS) earthquake catalog (<http://ds.iris.edu/ds/nodes/dmc>), we further selected teleseismic events that occurred during the period from January 2009 to May 2019 with an epicentral distance between 30° and 95° and a moment magnitude M_w greater than 5.5. Again, the seismograms of these teleseismic events were downloaded from the BMKG seismic network. For any selected teleseismic event, the direct *P* wave had to be consistently

Figure 2. Earthquakes selected for traveltime tomography. Colored dots in plots (a) and (c) are the epicenters of the selected regional earthquakes for *P* and *S* wave tomography, respectively. The color represents the depth interval of the events shown in the legend on the right of plot (c). Plots (b) and (d) show the azimuthal distribution of the selected teleseismic events (blue dots). The teleseismic events in plots (b) and (d) are used by teleseismic *P* and *S* wave tomography, respectively. The red circles in both (b) and (d) denote the epicentral distances of 30°, 60°, and 90° from the center of the study area. Earthquakes occurred from January 2009 to May 2019.

recorded by at least five stations in the study area. As a result, a total of 663 teleseismic events (Figure 2b) were selected. After filtering the waveform data within the frequency range from 0.02 to 0.2 Hz, the cross-correlation traveltime differences between all possible station pairs were measured. In teleseismic traveltime tomography, it is always assumed that the heterogeneity outside the study domain is negligible (Liu et al., 2018; Rawlinson et al., 2006). To satisfy this assumption, every station pair must be sufficiently close. Thus, the threshold for the distance of a station pair must be set properly. Allowing a large distance to form a station pair may result in substantial imaging errors, while an excessively short distance may result in poor data coverage. The tradeoff curve between the number of possible station pairs and the allowed maximum distance is presented in Figure S1 in the supporting information. Figure S1 shows that the number of station pairs is generally proportional to the allowed maximum distance. We choose 2.5° as the maximum distance for station pairs so that all the stations in the study area can be included and the influence caused by the heterogeneities outside the study area can be minimized (Zhao et al., 2013). Quantitative evaluation of the relation between the allowed maximum distance and the imaging accuracy is worthy of further investigation. The cross-correlation coefficient, which can be used to evaluate the quality of the data, was required to be greater than 0.9 to allow only high-quality data to be included in our tomographic inversion. Consequently, a total of 9,038 high-quality *P* wave traveltime differences were measured from 664 teleseismic events (Figure 2b). Adopting the same processes, we measured 4,498 teleseismic *S* wave traveltime differences from 623 teleseismic events (Figure 2d).

Figure 3. Schematic diagram showing the relocation of an earthquake. P is an arbitrary source in the Earth with coordinates of the hypocenter and origin time (r, θ, ϕ, t) . The red line is the ray path from the source to one station (black triangle). e_r , e_θ , and e_ϕ are three unit-basis vectors in spherical coordinates. V_R is the tangent unit vector of the ray path at the source. α , β , and γ are the angles between V_R and the three unit-basis vectors e_r , e_θ , and e_ϕ , respectively.

2.2. Method

In this study, the initial model M_0 is composed of a crust structure based on CRUST1.0 (Laske et al., 2013) and a mantle structure following ak135 (Kennett et al., 1995). We relocate all the regional earthquakes in M_0 using Geiger's method (Geiger, 1912). The misfit function for earthquake relocation is defined as follows:

$$f(x_S) = \frac{1}{2} \sum_{i=1}^{N_S} \sum_{j=1}^{N_i} \Delta T_{ij}^2 + \frac{\eta}{2} \|x_S\|_2^2, \quad (1)$$

where $\Delta T_{ij} = (T_{c,ij} - T_{o,ij})$; $T_{c,ij}$ and $T_{o,ij}$ are the j th calculated and observed traveltimes of the i th source, respectively; N_S is the total number of sources; N_i is the number of picks for the i th source; $x_S = (x_{S,1}, x_{S,2}, \dots, x_{S,N_S})$ is the source vector for the N_S sources with each $x_{S,i}$ containing the source location (r, θ, ϕ) (Figure 3) and origin time t ; and η is the damping factor for earthquake relocation. By directly extending the derivatives of $f(x_S)$ with respect to the hypocenter and origin time in a 1-D (radially symmetric) velocity model (Thurber & Rabinowitz, 2000) to a 3-D heterogeneous velocity model, the gradient of $f(x_S)$ with respect to the i th source is written as follows:

$$\left\{ \begin{aligned} \frac{\partial f(x_S)}{\partial x_{S,i}} &= \sum_{j=1}^{N_i} \Delta T_{ij} \frac{\partial \Delta T_{ij}}{\partial x_{S,i}} + \eta x_{S,i}, \\ \frac{\partial \Delta T_{ij}}{\partial x_{S,i}} &= \left(\frac{\partial \Delta T_{ij}}{\partial r_i}, \frac{\partial \Delta T_{ij}}{\partial \theta_i}, \frac{\partial \Delta T_{ij}}{\partial \phi_i}, \frac{\partial \Delta T_{ij}}{\partial t_i} \right), \\ \frac{\partial \Delta T_{ij}}{\partial r_i} &= \frac{\cos \alpha}{v}, \\ \frac{\partial \Delta T_{ij}}{\partial \theta_i} &= \frac{r \cos \beta}{v}, \\ \frac{\partial \Delta T_{ij}}{\partial \phi_i} &= \frac{r \sin \theta \cos \gamma}{v}, \\ \frac{\partial \Delta T_{ij}}{\partial t_i} &= -1, \end{aligned} \right. \quad (2)$$

where v is the velocity at the i th source and α , β , and γ are the angles between the ray direction at the source location and the three spherical unit vectors e_r , e_θ , and e_ϕ , respectively (Figure 3). The fast marching method (FMM; Sethian, 1996) was used to calculate the traveltime field in M_0 ; the ray path of the first arrival (red line in Figure 3) was then obtained by tracing the negative gradient of the traveltime field from the station to the source. The entropy-satisfying FMM algorithm, which combines upwind schemes with a fast sorting technique (Hassouna & Farag, 2007), can calculate the first arrivals in strong heterogeneous media with high accuracy (Rawlinson et al., 2006; Rawlinson & Sambridge, 2004). Considering that the derivatives $\frac{\partial \Delta T_{ij}}{\partial r_i}$, $\frac{\partial \Delta T_{ij}}{\partial \theta_i}$, $\frac{\partial \Delta T_{ij}}{\partial \phi_i}$, and $\frac{\partial \Delta T_{ij}}{\partial t_i}$ have different units, the gradient of the traveltime difference ΔT_{ij} were nondimensionalized by changing $\frac{\cos \alpha}{v}$, $\frac{r \cos \beta}{v}$, $\frac{r \sin \theta \cos \gamma}{v}$, and -1 to $\frac{D \cos \alpha}{vT}$, $\frac{r \cos \beta}{vT}$, $\frac{r \sin \theta \cos \gamma}{vT}$, and $-\frac{D}{vT}$, respectively (Tong et al., 2019). In this study, D and T are set to be 1,000 km and 1 s, respectively. Given the gradient in equation (2), we use the conjugate gradient (CG) method (Hestenes & Stiefel, 1952; Tong, Zhao, Yang, Yang, Chen, & Liu, 2014) to obtain the minimum of equation (1) in an iterative fashion.

After relocating the earthquakes, we inverted the regional traveltime data starting from M_0 to improve the velocity model. The updated 3-D velocity model M_a from the regional tomographic inversion was treated as the initial model for teleseismic tomography. After performing the teleseismic traveltime inversion, we obtained the final model M_b . All the tomographic results discussed in the subsequent sections consist of relative velocity perturbation, that is, the difference between M_b and M_0 divided by M_0 . The least square problem for both regional and teleseismic traveltime tomography is expressed as follows (Liu et al., 2018; Puspito et al., 1993):

$$F(X) = (AX - b)^T C_d^{-1} (AX - b) + \varepsilon X^T C_m^{-1} X, \quad (3)$$

where A is the sensitivity matrix, X is the discrete model perturbation vector, C_d is the a priori data covariance matrix, C_m is the a priori model covariance matrix, and ε is the damping factor for velocity inversion. In regional traveltime tomography, b represents the traveltime difference vector; similarly, b denotes the vector containing relative traveltime differences in teleseismic traveltime tomography. The arrival times of the regional seismic data were picked twice, and the traveltime uncertainties of regional seismic arrivals were estimated by the traveltime differences between the two datasets picked by two individuals. The data uncertainties are estimated to be above 0.1 s and incorporated into the diagonal elements of C_d . Multichannel cross-correlation was utilized to determine the uncertainties of the relative waveform cross-correlation traveltime differences (Allen et al., 2002; Vandecar & Crosson, 1990). Following the error criterion defined in Rawlinson and Kennett (2004), we set the minimal error cutoff to be 0.035 s, which is 75% of the sampling interval of most seismograms. Again, the estimated uncertainties of teleseismic data were incorporated into C_d . Furthermore, in this study, we do not consider the model uncertainty, and let C_m be an identity matrix. The CG method was also used to iteratively solve equation (3). At each iteration of the CG algorithm, a prescribed upper bound for the relative model perturbation was set by choosing an appropriate damping factor. For a series of prescribed upper bounds, an L curve, that is, the tradeoff curve between the model variance and residual variance, can be formed accordingly. The fundamental principle of this strategy is to control the maximum model perturbation at each iteration and steadily update the model using a relatively small updating step. More details of this approach can be found in Liu et al. (2018). Notably, the inversion strategy of setting an upper bound is not sensitive to the number of inversion steps, as the number of inversion steps changes according to the upper bound. Fewer iteration steps correspond to a larger upper bound, while a smaller upper bound requires more iteration steps. In our study, we set the number of iteration steps for the earthquake relocation to be 4 and that for both the regional and the teleseismic traveltime inversion to be 15.

3. Results

3.1. Checkerboard test

We use the grid approach to parameterize the velocity model in this study. A series of checkerboard resolution tests with different grid intervals were conducted to find the optimal model parameterization. For P wave tomography, the optimal intervals of the inversion grid in the longitudinal and latitudinal directions are 1.25° and 1.33° , respectively; in the vertical direction, the grid is located at eight depths, namely, 20, 50, 120, 200, 300, 450, 600, and 800 km. The checkerboard model is constructed by alternatively adding $\pm 4\%$ relative velocity perturbations to the initial model M_0 . In addition, normally distributed random errors with a standard deviation of 0.2 s are added to the synthetic regional and teleseismic traveltime data computed in the checkerboard model. The results of the checkerboard resolution tests are obtained by successively inverting the regional and teleseismic data. Figure 4 shows that the velocity perturbations are generally well recovered, although varying degrees of smearing are observable. At shallow depths (20, 50, and 120 km), the anomalies in the offshore region and beneath mainland Sumatra are well resolved because of the frequent seismicity in these areas. In contrast, the shallow structure in Malaysia cannot be imaged properly due to the limited seismicity therein. At greater depths (>120 km), the resolution in the NE part of the study area (including mainland Sumatra and Malaysia) is improved, but the resolution in the SW part of the study area is relatively poor, mainly due to the lack of stations in the Indian Ocean (Figure 1) and few teleseismic waves arriving from the west of our study area (Figure 2b).

Figure 5 shows the results of the checkerboard tests with two separate datasets, that is, the regional and teleseismic P wave traveltime data. The regional data can effectively resolve the shallow velocity structure from 20 to 120 km (Figures 5a–5c), but the resolution at 200-km depth is greatly reduced (Figure 5d). Because regional earthquakes mainly occurred above at a depth of 200 km (Figure 2a), anomalies at depths of 300–800 km cannot be resolved (Figures 5e–5h). In contrast, the teleseismic data can image the velocity structure at greater depths (200–800 km). It is worth noting that regional and teleseismic data can be inverted simultaneously. We include the results of the joint inversion of regional and teleseismic synthetic data in Figure S2. Because regional data and teleseismic data have different levels of uncertainty (the uncertainty of regional seismic data is mainly in the interval of 0.1–0.15 s, whereas the uncertainty of teleseismic data

Figure 4. Map views of the recovered *P* wave checkerboard model obtained by inverting synthetic *P* wave traveltimes containing Gaussian noise with a standard deviation of 0.2 s. The input relative velocity perturbation is $\pm 4\%$. Regional *P* wave traveltimes tomography is first conducted in the starting model M_0 . Then, we conduct teleseismic traveltimes tomography in the model given by regional *P* wave traveltimes tomography. The inversion process is the same as that for the real data (Figure 7). The depth of each layer is shown in the lower-left corner of each plot. The color scale is shown at the bottom right.

is mainly in the interval of 0.035–0.04 s), the joint inversion seems to be dominated by the teleseismic data. As shown in Figure S2, the alternating pattern is successfully recovered only at depths ranging from 200 to 800 km.

The checkerboard test for *S* wave tomography is presented in Figure S3. The intervals of the inversion grid in the latitudinal and longitudinal directions are 1.6° and 1.43° , respectively, and the inversion nodes in the vertical direction are at depths of 20, 50, 150, 200, 400, 600, and 800 km. Figure S3 indicates that the resolution of *S* wave tomography is slightly lower than the resolution of *P* wave tomography, especially at shallow depths (Figures S3a and S3b).

3.2. Tomographic Results

We use the *L* curve method (Pesicek, Thurber, Widiyantoro, et al., 2010; Rawlinson et al., 2006) to determine the predefined upper bounds of allowed model perturbations at each iteration for earthquake relocation, regional *P* wave data inversion, and teleseismic *P* wave data inversion (Figures 6a, 6c, and 6e). The optimal upper bounds are found to be 1%, 1%, and 0.7%, respectively. Restricted by these upper bounds, Figures 6b, 6d, and 6f present statistical histograms of the traveltime residuals after the earthquake relocation and tomographic inversions. The statistical histogram in Figure 6b shows that the mean and standard deviation of the regional seismic traveltime residuals change from 1.05 ± 1.27 to 0.02 ± 0.71 s after relocating regional earthquakes. The misfit is further reduced to 0.01 ± 0.55 s after regional tomographic inversion (Figure 6d). As

Figure 5. Checkerboard resolution tests from two separate *P* wave data sets. (a–h) Checkerboard test with the regional *P* wave traveltime data; (i–p) checkerboard test with the *P* wave teleseismic traveltime data. For both checkerboard tests, the synthetic *P* wave traveltimes are perturbed with random noise with a standard deviation of 0.2 s. The initial *P* wave velocity model is the ak135 stratified Earth model (Kennett et al., 1995) combined with CRUST 1.0 (Laske et al., 2013) for crustal corrections.

shown in Figure 6f, the mean and standard deviation of the teleseismic relative traveltime misfits decrease from 0.08 ± 0.68 to 0.05 ± 0.37 s. Obviously, the earthquake relocation and tomographic inversion procedures greatly reduce both the mean and standard deviation of the seismic traveltime residuals. For *S* wave tomography, the *L* curves utilized to define the upper bounds and histogram of the residuals are shown in Figure S4. By comparing the residual distributions in Figure S4 with those in Figure 6, we find that the *S* wave arrivals are inferior in quality to the *P* wave arrivals. We present the misfit (*L2* norm of the traveltime residuals) reduction curves in Figure S5. As shown in Figure S5, the misfits for both *P* and *S* wave tomography are greatly reduced after 15 iterations, and the improvements are very minor at the 15th iteration.

3.2.1. Map Views

The final *P* and *S* wave results at different depths are shown in Figure 7. At shallow depths (Figure 7a), the study area is dominated by low *P* wave velocity anomalies beneath mainland Sumatra and the offshore

Figure 6. Tradeoff L curves (left plots) and histograms of the traveltime residuals (right plots). The upper, middle, and lower plots are for the earthquake relocation, regional traveltime inversion, and teleseismic traveltime inversion, respectively. The red dots in the left plots denote the optimal prescribed values. The blue and red histograms in the right plots are the traveltime residuals before and after the inversions, respectively. Dimensionless treatment is applied on the hypocentral coordinates and origin time, therefore the model variance in (a) is dimensionless.

regions. A significant low-velocity anomaly with a negative maximum perturbation over -3.5% is located to the SW of Toba Caldera, which is the site of the world's largest Pleistocene eruption that occurred 74,000 years ago. This caldera is characterized by a topographic depression with a lateral extent of $30 \text{ km} \times 100 \text{ km}$ (Koulakov et al., 2016; Masturyono et al., 2001). Notable low-velocity anomalies with negative perturbation less than -2% are observed beneath Toba Caldera. The regional P wave tomography model in Masturyono et al. (2001) generally shows significant low-velocity anomalies with perturbations exceeding -20% in the middle crust right below and south of Toba Caldera; but some high-velocity anomalies are also observable beneath the northern tip of Samosir Island. The dataset of Masturyono et al. (2001) was reused by Koulakov et al. (2009). Both the P - and S -wave tomographic results in Koulakov et al. (2009) reveal strong low-velocity anomalies below Toba Caldera and along the SFS. However, the low-velocity anomalies in Koulakov et al. (2009) are less significant than those in Masturyono et al. (2001). To address this discrepancy, Koulakov et al. (2009) claimed that the unrealistic negative anomalies in

Figure 7. Map views of the P- and S-wave velocity anomalies at eight depths. The velocity anomalies are plotted with respect to the initial 3-D model M_0 . (a-h) P-wave perturbation; (i-p) S-wave perturbation. The blue curves in (b-d) and (j-l) represent the slab contours from Slab1.0 (Hayes et al., 2012). The depth of each layer is shown in the lower-left corner of each plot. The labels are the same as those in Figure 1.

Masturyono et al. (2001) are probably due to the instability of the underdamped inversion. Surface wave tomography (Stankiewicz et al., 2010) reveals laterally continuous low velocities extending from the shallow crust (2–3 km) to the middle crust (10–15 km) along the long axis of the NW-SE-elongated Toba Caldera. These laterally continuous low velocities in the upper and middle crust have also been revealed by receiver function analysis (Sakaguchi et al., 2006). In addition, the V_P/V_S value (1.87) of the crust beneath Toba Caldera is higher than the global average (1.78) of continental crust (Sakaguchi et al., 2006; Zandt & Ammon, 1995), which is indicative of partial melting and high temperatures within the Toba crust (Masturyono et al., 2001). In the offshore regions, obvious low velocities (–2% to –1.5%) are situated beneath the islands of Simeulue, Nias, and Batu at a depth of 20 km (Figure 7a). The inversion of the Rayleigh wave group velocity dispersion curve from one onshore-offshore station-pair (Harmon et al., 2012) shows a low velocity anomaly with a perturbation of –5% at 15-km depth. Due to the reduction in resolution as depth increases, Harmon et al. (2012) did not present the S wave velocity structure below 15 km; however, this low velocity anomaly probably extends to a depth of 20 km, as inferred by Lange

et al. (2010). These low velocities may be attributed to the accretionary prism and the thick sediments in the forearc basin (Harmon et al., 2012; Lange et al., 2018; Moore et al., 1980). Another possible origin of these low velocities may be the subducted IFZ beneath Nias Island and adjacent areas, considering that subducted seamounts with elevations of 1,000–1,500 m above the surrounding seafloor in conjunction with the thick oceanic crust within the fracture zone may exhibit negative velocity perturbations (Fauzi et al., 1996).

As shown in Figure 7b, the low-velocity anomalies beneath the main island extend to the uppermost mantle with increased magnitudes. No significant velocity perturbations are observed beneath the offshore region, except for the high-velocity anomalies to the SW of the province of Aceh (Figure 7b) and the high-velocity anomalies below Simeulue and Nias Islands. These high-velocity anomalies may reflect the subducted slab. Because of the small temperature contrast between the subducted slab and ambient mantle at a depth of 50 km (Zheng et al., 2016), the tomographic images do not show continuous slab-related high velocities below the forearc basins. At 50-km depth, the maximum low-velocity perturbation is over -4% , which is located below Lake Toba and south of Lake Toba. The regional traveltime tomography study conducted by Koulakov et al. (2016) also reveals a broad low-velocity anomaly crossing their entire study area (1°N – 4°N and 98°E – 99.5°E) at the same depth. In addition, they found that the maximum low-velocity perturbation is located to the south of Toba, but it is narrowly shaped and extends into the offshore area (Koulakov et al., 2016), which is slightly different from our results.

The low velocities in the crust and uppermost mantle revealed by *P* wave tomography are not well resolved by *S* wave tomography (Figures 7i and 7j), which may be due to two factors. First, compared with *P* wave tomography, considerably fewer regional *S* wave arrivals are available (only 4,498 picks). Second, the quality of the *S* wave arrivals is inferior to that of the *P* wave arrivals. The residuals of the regional seismic data evaluated in the final *S* wave model are 0.02 ± 0.79 s (Figure S4d), which is much larger than those (0.01 ± 0.55 s; Figure 6d) in the final *P* wave model. Therefore, our discussion of the low velocities in the crust and uppermost mantle is based on the results of *P* wave tomography.

From the depth of 120 km to the top of the mantle transition zone (MTZ; Figures 7c–7f), the most remarkable feature is the NW–SE-trending slab-related positive velocity anomaly with a maximum perturbation of $\sim 4\%$. The imaged slab is generally parallel to both the volcanic arc and the trench. Additionally, the shape of the slab resembles the contour of the Wadati–Benioff zone (blue curves in Figures 7c, 7d, 7k, and 7l) and exhibits gentle bending (Richards et al., 2007). Fauzi et al. (1996) derived a similar slab geometry from the locations of slab-related earthquakes; however, the slab shape at greater depths (>200 km) was not reported, as the earthquakes occurred mainly above 200-km depth. Nevertheless, based on the tomographic results shown in Figures 7f and 7n, we infer the existence of a curved slab extending to a depth of at least 450 km. Furthermore, at a depth of 250 km, global tomography results presented in Bijwaard et al. (1998) show a much wider slab beneath northern Sumatra, especially beneath Lake Toba. Another global tomography model (Káráson, 2002), which was used as an auxiliary tool to understand the evolution of the mantle beneath SE Asia since 50 Ma (Replumaz et al., 2004), shows fragmented and weak positive velocity perturbations at shallow depths (≤ 200 km). However, the global velocity model UU-P07 inverted from a larger volume of seismic data (Amaru, 2007; Hall & Spakman, 2015) reveals a slab shape similar to ours at shallow depths (≤ 300 km). The slight difference is that the slab in UU-P07 is thicker than our slab beneath Lake Toba and the area north of Lake Toba at depths from 200 to 300 km. Global earthquakes (1964–2006; the data were downloaded from <http://www.isc.ac.uk/isc-ehb>) and earthquakes in the Sumatra region (1918–2006; Engdahl et al., 2007) were used by Pesicek et al. (2008) to image the *P* wave velocity structure of the Burma, Andaman, and Sumatra subduction zones. Beneath northern Sumatra, their model shows slab folding at depths from 190 to 530 km (Pesicek et al., 2008). The model of Pesicek et al. (2008) shows that the curvature of the slab in the upper mantle and MTZ, especially at a depth of 450 km, is considerably larger than that presented in Figures 7c–7f and 7k–7n.

At a depth of 120 km (Figure 7c), the NW portion of the slab seems to penetrate toward the NNE with an additional displacement relative to the SE portion, and thus, a gap seems to appear between them. The slab gap here is most likely the representation of a slab tear beneath northern Sumatra (Hall & Spakman, 2015). Fauzi et al. (1996) found a seismicity gap south of Lake Toba and inferred a similar slab gap around 100-km depth. They considered that the scale of the gap spans tens of kilometers, which is smaller than the gap shown in Figure 7c. A difference in the slab displacements was also found by Hall and Spakman (2015), but at a depth of 300 km, which is much deeper than that found from our analysis. Based on the joint

inversion of local and teleseismic traveltime data, Puspito et al. (1993) showed a slab gap at depths of 250–300 km approximately 200 km to the south of Toba. In contrast, regional seismic tomography (Pesicek et al., 2008; Pesicek, Thurber, Widiyantoro, et al., 2010) based on a nested regional-global approach (Widiyantoro & van der Hilst, 1996) using global seismic data did not show a gap but a laterally continuous slab in the upper mantle. Different from *P* wave tomography, *S* wave tomography does not reveal a slab gap at 120-km depth (Figure 7k) but shows a less curved slab compared with previous results (Pesicek et al., 2008).

Figure 7g shows that the high-velocity perturbations in the northern portion of the study area (2°N–6°N) are relatively weak compared with those in the SE and central portions. The positive anomaly with an amplitude less than 1% (Figure 7g) may be an artefact. At a depth of 600 km, the slab-related positive velocity anomaly is observable from –2°S to 2°N, but it is not observed below the area NW of Toba Caldera (Figure 7g). Obviously, the imaged slab has been greatly distorted and has a spatial shift toward the SW from 0°N to 2.0°N. The SE limb of the folding from the UU-P07 model also suffers from remarkable slab distortion at a depth of 600 km (Figure S6; Amaru et al., 2007; Hall & Spakman, 2015). Additionally, one weak high-velocity anomaly with an amplitude greater than 1% is observed in the south at a depth of 800 km (Figures 7h and 7p).

We present the results of *P* wave tomography by inverting only the regional traveltimes in Figure S7. As indicated by the checkerboard test (Figures 5a–5c), regional seismic traveltime tomography can resolve only the structure in the uppermost 120 km beneath the Earth's surface. At a depth of 20 km, the velocity structures are mostly resolved by regional seismic data, and the results shown in Figures 7a and S7a are nearly the same. At a depth of 50 km, the results in Figures 7b and S7b are generally the same; however, the high-velocity anomaly in Figure 7b beneath the forearc basin is more continuous because teleseismic data can constrain the structures at this depth. At a depth of 120 km, only the slab in the south can be imaged by regional seismic traveltime tomography (Figure S7c).

We conduct two restoration tests (Figure 8) to demonstrate that the main features from *P* wave tomography discussed above are reliable. The slab morphology is constructed mainly based on the slab-related high-velocity anomalies in Figures 7c–7h. The two input slab models in Figure 8 are represented by positive *P*-wave anomalies with a constant amplitude of 4% (Pesicek, Thurber, Widiyantoro, et al., 2010). In the first synthetic model (first column in Figure 8), a gap exists at 120-km depth that separates the slab into two parts, while the slab is continuous in the second synthetic model (third column in Figure 8). The FMM is used to compute the synthetic data, to which random errors are added. The same inversion procedure employed for the real data is applied. The output models show that the recovered slabs are similar to the input models, though image smearing is evident at the some depths (600 and 800 km). Due to regularization, the amplitudes of the inverted velocity anomalies are slightly smaller than those of the input model. Figure 8 demonstrates that the slab tear at 120-km depth can be resolved by our datasets, and the slab gap shown in Figure 7c is not due to a reduced resolution at this depth.

3.2.2. Vertical Views

Cross sections of the inverted *P* wave velocity model are plotted in Figure 9. The imaged slab correlates well with the intraslab seismicity. Earthquakes in northern Sumatra almost all occur above a depth of 300 km (Pesicek, Thurber, Zhang, et al., 2010), much shallower than the maximum depth of seismicity beneath Java (where the deepest earthquakes reach 670 km; Widiyantoro & van der Hilst, 1996; Richards et al., 2007). These relatively shallow earthquakes are attributed to the relatively young age of the subducted slab (~60 Ma; Wortel & Vlaar, 1988). As shown in Figures 9a and 9i, the slab below the northern tip of northern Sumatra penetrates to a shallower depth than the slabs on the other cross sections (BB' through HH'). Clearly, both the *P* wave and the *S* wave tomography results show that the slab on profile AA' just reaches the 410-km discontinuity, and the penetration depth of slab is deeper than that imaged by teleseismic data (Figure S8; Liu et al., 2018). Because the slab on profile AA' abuts the northern boundary of the model, the slab beneath the northern tip of Sumatra may continue beyond the northern boundary of the study area and penetrate to a deeper depth; however, referring to the slab on profile BB' (on which the slab just penetrates across the 410-km discontinuity), we do not consider that the slab on profile AA' penetrates to a much deeper depth beyond the 410-km discontinuity. Nevertheless, some *P* wave tomography models in the literature show that the slab beneath the northern tip of Sumatra penetrates to a depth of at least 500 km (Hall & Spakman, 2015; Pesicek et al., 2008; Richards et al., 2007; Zenonos et al., 2019), although some other studies suggest that the slab penetrates to a much shallower depth (approximately 250 km; Huang et al., 2015).

Figure 8. Slab restoration test. Two synthetic slab models (4% P-wave velocity perturbation), which mimic the geometry of the actual model (the main feature of the slab we discuss) are constructed to test the recovery and degree of smearing of small-scale and slab-like anomalies. The two synthetic models are the same; however, the slab in model 1 is discontinuous at a depth of 120 km, while the slab in model 2 is continuous. The fast marching method (FMM) is used to calculate the synthetic P-wave traveltimes in these two input models. Gaussian noise with a standard deviation of 0.2 s is added to the synthetic data. The amplitudes of the inverted velocity anomalies slightly decrease because of the regularization term added during the inversion.

Figure 9. Cross sections of the velocity perturbations in the study area. The white dots depict seismic events located at distances of less than 30 km from each profile. The hypocentral information is obtained from the IRIS Data Management Center (<http://ds.iris.edu/ds/nodes/dmc>). The cross sections, the positions of which are shown in the map on the bottom left, are oriented roughly parallel to the movement of the Indo-Australian plate, and the distance between two adjacent cross sections is 1°. The two dashed lines on each cross section represent the 410-km discontinuity and 660-km discontinuity. Volcanoes (red triangles) within 50 km are projected onto the cross-sections. Plots (a–h) are the results of *P* wave tomography; plots (i–p) are the results of *S* wave tomography. The color scale is shown on the bottom of the figure.

Figure 10. Slab morphology beneath northern Sumatra. (a) Schematic of the slab morphology beneath northern Sumatra. The gray planes represent the subducting oceanic lithosphere. The red triangle denotes Toba Caldera. (b) Slab contours at several depths defined by 1% velocity perturbation contours.

Toward the SE, the slabs on profiles BB', CC', and DD' (Figures 9b–9d) begin to cross the 410-km discontinuity. The shallow penetration of the slab in the north is probably a real feature, as demonstrated by the restoration test shown in Figure S9. The dips of the slabs on profiles AA'–DD' at depths of 100–400 km are nearly the same (approximately 60°). Further to the SE, the slabs on profiles EE' and FF' penetrate to a depth of 600 km. As shown in Figures 9e and 9m, the slab on profile EE' at depths of 500–700 km is deflected toward the SW, which is similar to the result of Liu et al. (2018; Figure S8e). The dips of the slabs on profiles EE' and FF' are comparable to those on profiles AA' through DD', whereas the dips of the slabs on profiles GG' and HH' (especially HH') are steeper than the others. Furthermore, our tomographic results suggest that the slab does not appear to stagnate in the MTZ. In addition, no slab detachment is found in our results, although this phenomenon has been reported in some previous studies (Liu et al., 2018; Widiyantoro & van der Hilst, 1996).

Despite that about 80% of the teleseismic data used by this study are the same as those of a previous work (Liu et al., 2018), there are still some differences between the new model (Figure 9) and the previous one (Figure S8) at some deep depths. The slab detachment observed on profile CC' in Liu et al. (2018, Figure S8c) is not found in either the *P* wave or *S* wave tomographic results in this study (Figures 9c and 9k). In Figure S8c, the high-velocity anomalies in the MTZ and uppermost lower mantle with a perturbation amplitude of ~1% may not be reliable. Furthermore, the slab deflection on profile EE' (Figure 9e) is not as significant as that in Liu et al. (2018; Figure S8e). However, *S* wave tomography of this study shows significant slab deflection on profiles EE' and FF' (Figures 9e and 9f). Hence, further investigations are necessary to resolve these inconsistent features including slab detachment and slab deflection.

4. Discussion

By examining the slab revealed by both *P* and *S* wave tomography, in Figure 10a, we propose a subducting slab model to interpret the slab morphology beneath northern Sumatra. This model is characterized by low slab curvature, tearing, deflection, and dip angle variations. The slab in the north may extend to an average depth of more than 410 km (410–450 km), while the slab in the south may extend to an average depth of 700 km.

4.1. Slab With Low Curvature

Slab folding, which refers to the drastic variations in both the dip and strike of the slab, was previously suggested to be a primary and continuous feature from the surface to the MTZ beneath Toba (Pesicek et al., 2008; Pesicek, Thurber, Widiyantoro, et al., 2010). However, no obvious slab folding is found in our tomographic results. To clearly visualize the geometry of the slab, in Figure 10b we plot the contour curves along which the relative velocity perturbation is 1%. The results of this study suggest that the subducted slab beneath northern Sumatra is characterized mainly by a slab that is less curved at depths of 120–450 km than those

reported in previous studies (Hall & Spakman, 2015; Pesicek et al., 2008; Pesicek, Thurber, Widiyantoro, et al., 2010). At depths of 120–450 km, the strike of the slab evidently changes from NNW south of Toba to NW north of Toba, which is consistent with the findings reported by Fauzi et al. (1996). The slab contours at depths ranging from 120 to 450 km resemble those inferred from the locations of regional earthquakes (Fauzi et al., 1996; Pesicek, Thurber, Zhang, et al., 2010). However, the relatively shallow seismicity in northern Sumatra makes it impossible to define the shape of the deep slab by slab-related seismicity. Combining the results of Fauzi et al. (1996) with our tomographic images, we infer that the slab bends gently from the surface to the MTZ. Hall and Spakman (2015) also presented gentle slab bending at depths of 100–300 km, which is similar to the slab at shallow depths (120–450 km) shown in Figures 7c–7f. However, Hall and Spakman (2015) reported significant slab folding at depths of 400–500 km, although the curvature of the folded slab is slightly lower than that presented by Pesicek et al. (2008). Pesicek, Thurber, Widiyantoro, et al. (2010). Our recent teleseismic traveltime tomography study revealed gentle slab bending in the upper mantle and MTZ (Liu et al., 2018), and the curvature of the slab in Liu et al. (2018) is much lower than that presented in Pesicek et al. (2008) and Hall and Spakman (2015). One possible reason for the increased slab curvature in Pesicek et al. (2008) is that the high-velocity anomaly beneath the Malay Peninsula (Figures 7d and 7e) may be combined with the slab-related high-velocity anomalies discussed in Pesicek et al. (2008). However, our model shows separated high velocity anomalies and a less curved slab. The high-velocity beneath the Malay Peninsula was also imaged by Huang et al. (2015). It is shown to be separated from the high-velocity anomaly beneath northern Sumatra at shallow depths (15–150 km) and combined with the slab-related high-velocity anomaly at greater depths (200–330 km; Huang et al., 2015).

Fauzi et al. (1996) pointed out that the curvature of the slab may follow the changes in the strike of the trench. In contrast, high-resolution seafloor mapping and multichannel seismic reflection profiles along the Sumatra Trench reveal the deformation of the trench and the strong shortening of the overriding plate in response to slab subduction (Kopp et al., 2008). Here we consider that the curvature of the slab in northern Sumatra may be ascribed to three main factors.

First, the curved slab may be partially caused by trench migration. As suggested by Replumaz et al. (2004), the extrusion of the Sunda plate, which is closely related to the collision between the Indian and Eurasian plates, results in the migration of the Sunda Trench. Based on the migration of the Sunda Trench illustrated by the tectonic reconstruction model of Lee and Lawver (1995), Hafkenscheid et al. (2001) conducted kinematic modeling of the temperature distribution in the mantle beneath Indonesia. In their model the slab represents a cold core with a temperature much lower than the ambient mantle, and they found that the slab is gently bent in the upper mantle and MTZ (Figures 3 and 4 in Hafkenscheid et al., 2001). This kinematic modeling indicates that the migration of the Sunda Trench may be a causing factor of the curved slab beneath northern Sumatra.

Second, the lateral stress within the subducting slab may play an important role in its curvature. As suggested by Liu et al. (1995) and McCaffrey et al. (2000), slip partitioning is correlated with the lateral stress within the subducting slab and has been observed from earthquake slip vectors in many oblique subduction zones such as the Sumatra and Aleutian subduction zones. Lateral movement between an obliquely subducting slab and the overriding plate produces an arc-parallel shear force. According to the force balance equations of a downgoing slab (equation (5) in Liu et al., 1995), slab subduction with a high degree of oblique convergence ($\sim 70^\circ$; Figure 1) may result in a large lateral shear stress within the slab. This large shear stress would further induce slab deformation and result in slab curving in the upper mantle and MTZ (Figure 10b). Because of the reduced viscosity of the subducting slab caused by the increased temperature and the accumulated deformation at great depths, the slab suffers from considerable distortion at a depth of 600 km. We can also observe that the NW portion of the subducting slab is deflected toward the SW (Figures 9e and 9m).

Third, the physical properties of the subducting slab may affect its shape. The subducting slab in northern Sumatra is composed of a warm lithosphere with a relatively young age (Shapiro et al., 2008; Whittaker et al., 2007) and may deform easily as it subducts. In addition, the subducting Wharton Fossil Ridge, which forms the diffuse and deformation boundary between the Indian and Australian plates (Pesicek et al., 2008), and the subducting IFZ, which is characterized by fractures and transform faults (Hall & Spakman, 2015), may serve as a weak zone facilitating the occurrence of deformation.

4.2. Slab Tear

Our results show a possible slab tear beneath Toba Caldera only at a depth of 120 km (Figure 7c). However, as shown in Figure 10a, we speculate that the slab tear extends to a much deeper depth (>400 km). Following Hall and Spakman (2015), we also infer that there exists slab tear at great depth according to the curvature of the slab and its dip angle, which varies from the north to the south. Numerous previous studies have speculated the existence of slab tear beneath northern Sumatra, although no convincing evidence has been presented (Fauzi et al., 1996; Hall & Spakman, 2015; Koulakov et al., 2016; Liu et al., 2018; Page et al., 1979). For instance, Page et al. (1979) inferred the presence of slab tear beneath Toba by analyzing the spatial distribution of active volcanoes. In detail, they observed that the volcanoes south of Toba are located along the SFS; in contrast, the volcanoes north of Toba are shifted to the east of the SFS (Figure 1). The slab tear model of Page et al. (1979) shows that the dip of the slab NW of Toba is shallower than that SE of Toba. Additionally, this model can partially explain the unusual volcanic activity in the Toba area (Koulakov et al., 2016; Nishimura et al., 1984; Page et al., 1979). Fauzi et al. (1996) also suggested that there is a tear or a step in the slab south of Toba by analyzing slab-related seismicity. The tear or step that they inferred is at nearly the same depth as that suggested by our tomographic results (Figure 7c). Moreover, a slab model with a tear beneath northern Sumatra is preferred by Hall and Spakman (2015) as inferred from the observation that the slab to the east of the proposed tear dips more steeply. More recently, both Koulakov et al. (2016) and Liu et al. (2018) speculated the existence of slab tear beneath northern Sumatra based on evidence of slab bending and significant low-velocity zones in the overlying mantle wedge.

Fauzi et al. (1996) considered that the density contrast due to the age differences across the IFZ and the right-lateral shear stress on the slab may constitute two secondary factors, rather than primary factors responsible for the potential tearing of the subducted slab. In addition to these two factors, we consider that the weak physical properties of the IFZ also contribute to the slab tear. The oceanic plate west of the IFZ is significantly younger (Carton et al., 2014) than the slab east of the IFZ by up to ~ 15 Ma (Lange et al., 2010; Müller et al., 1997). This age difference results in a density contrast across the IFZ (i.e., the slab in the north is warmer and more buoyant, while the slab in the south is colder and denser) and further causes an increase in the slab dip from the north to the south. The shallow slab dip north of Toba (Figures 9a–9d) is related to the NNE shift of the NW portion of the subducting slab at 120-km depth (Figure 7c). The volcanic arc generally follows the 120-km depth contour of the slab (Fauzi et al., 1996; Zheng et al., 2016). Therefore, the volcanoes in Toba and those to the NW of Toba are shifted toward the NNE relative to the SFS.

Moreover, in conjunction with the fractures and transform faults in the IFZ, the serpentinization of the uppermost mantle beneath the IFZ makes the slab rheologically weak. An inversion of *S* wave dispersion data from ocean bottom seismometers near the trench showed low *S* wave velocities at a depth of ~ 25 km (Harmon et al., 2012). Harmon et al. (2012) considered 14–24% serpentinization based on the empirical relationship between the *S* wave velocity and percent serpentinization (dunite to lizardite serpentinization; Figure 8A in Christensen, 2004). Furthermore, deep seismic reflection images along two profiles in the Wharton Basin west of the northern tip of Sumatra revealed seismic reflectivities in the uppermost mantle that are considered to represent the serpentinization of mantle rocks along deep fault planes (Carton et al., 2014). These deep faults may serve as pathways for fluid flow between the basement of the oceanic crust and the uppermost mantle. At depths of 25–30 km, hydrated mantle peridotites may be serpentinized at 400–500 °C and ~ 1 GPa (McKenzie et al., 2005). We infer that the same mechanism may occur within the IFZ. Because of the lateral shear stress within the slab, slab tear may have developed along physically weak zones. However, we consider that the orientation of the tear in Figure 10a may be consistent with the bending axis in Figure 10b (dotted line), which is different from the orientations of the nearly north IFZ (Hall, 2012; Lange et al., 2010) and the NNE Wharton Fossil Ridge (Pesicek et al., 2008; Shapiro et al., 2008).

4.3. Low-Velocity Anomalies

P wave tomography reveals significant low-velocity anomalies that are observable in the crust and uppermost mantle (Figures 7a and 7b). The significant low-velocity anomalies in the uppermost mantle may reflect partial melting and are closely related to the Toba supervolcano (Koulakov et al., 2016). The most recent eruption of this volcano occurred 74,000 years ago, during which an enormous volume of material (ignimbrites and tuffs) estimated at $2,500\text{--}3,000\text{ km}^3$ was erupted (Stankiewicz et al., 2010). Based on the results of a regional seismic tomography study, Koulakov et al. (2016) proposed a multilevel magma plumbing

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system beneath Toba to interpret the enormous magma reservoir in the uppermost mantle. Concentrated and strong slab dehydration was found at depths of about 80 and 150 km (Koulakov et al., 2016). Because our model is much larger than that of Koulakov et al. (2016) and the tomographic inversion is performed on a much coarser inversion grid, our results do not reveal this multilevel plumbing system. Nevertheless, we follow the conclusion of Koulakov et al. (2016) and consider that the dehydration of the subducting slab is mainly responsible for the broad and strong low-velocity anomalies in the uppermost mantle. Furthermore, the seamounts within the IFZ are elevated 1,000–1,500 m above the surrounding seafloor. According to the principle of isostasy, the crust thickness in the IFZ is estimated to be 15 km, which is much thicker than the global average thickness of 6–7 km for oceanic crust (Fauzi et al., 1996; Harmon et al., 2012). Moreover, the subduction of the thick crust releases more water into the overlying mantle than that normal subducting oceanic crust. At the same time, the breakdown of serpentine in the subducting slab at a depth of about 100 km may release extra fluids into the mantle wedge (Zheng et al., 2016). In addition, the subducted oceanic lithosphere around the slab tear is relatively more exposed to the ambient mantle, and the fluids there can be easily released from the inside of the subducted slab into the overlying mantle wedge. Consequently, the upwelling of fluids into the mantle wedge reduces the solidus temperature of the rocks and results in partial melting in the mantle wedge (Cheng et al., 2011).

5. Conclusions

We applied an eikonal equation-based seismic tomography method to image the subsurface structures beneath northern Sumatra by inverting both regional and teleseismic data. The tomographic results show that the subducted slab exhibits along-trench variations in the curving, tearing, deflection, and dip angle. Slab folding is considered to be the main feature of the subducted slab in the literature but is not observed in our tomographic results. Instead, we find that the subducted slab displays a gently curved shape. We suggest that three main factors, namely, trench migration, lateral stress on the slab, and the physical properties of the subducted slab, are responsible for the gently curved shape. Additionally, from P wave tomography, a slab tear is imaged at a depth of about 120 km. This slab tear, which results in the additional release of fluids from inside of the slab, acts as one of the main factors for the broad and strong low-velocity anomalies in the uppermost mantle. Furthermore, the large-scale magma reservoir in the uppermost mantle may be closely related to Toba supervolcanism.

Although S wave tomography reveals a slab morphology similar to that revealed by P wave tomography, the slab gap appearing at 120-km depth (Figure 7c) is not distinct from the S wave tomographic results (Figure 7k). Meanwhile, the inverted S wave velocities at shallow depths (20 and 50 km) near the SFS (Figure 7i and 7j) may be not reliable. Thus, it may be necessary to adopt other state-of-the-art imaging methods, such as ambient noise tomography (Yang et al., 2011) and adjoint tomography (Chen et al., 2015; Tao et al., 2018; Tong, Zhao, Yang, Yang, & Liu, 2014; Tromp et al., 2005), to reliably obtain fine-scale S wave structures. In addition, future work including the inversion of seismic data generated by regional earthquakes but recorded by global stations, especially those from 2004 to 2005 Sumatra-Andaman great earthquake sequences (Pesicek et al., 2008; Pesicek, Thurber, Zhang, et al., 2010), may be useful for further improving the resolution of both P and S wave tomography of northern Sumatra.

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