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RESEARCH ARTICLE

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Key Points:

- High-resolution tomographic images of the velocity structure beneath Sumatra are obtained by inverting regional and teleseismic data
- The slab exhibits a sinusoidal shape with a low degree of curvature and hence is free of strong lateral deformation
- Two slab tears are found from our tomographic results. One is beneath Toba caldera; another one is beneath Sunda Strait

Supporting Information:

- Supporting Information S1

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The Geometry of the Subducted Slab Beneath Sumatra Revealed by Regional and Teleseismic Traveltime Tomography

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Abstract The geometry of the subducted Indo-Australian plate beneath Sumatra is still controversial because of the historical lack of a dense seismic network. Since 2005, Indonesia has been establishing a relatively dense seismic network for real time earthquake monitoring and tsunami warning. The seismic data accumulated by this network make it possible to achieve high-resolution tomographic images of the velocity structure beneath Sumatra by using eikonal equation-based seismic tomography method. Our P-wave tomographic images derived from regional seismic and teleseismic traveltime data demonstrate that the slab in the upper mantle generally follows the strike of the trench and the orientation of the volcanic arc. Additionally, the slab exhibits a sinusoidal shape with a low degree of curvature. Our tomographic results also reveal that the maximum penetration depth of the subducted slab increases from the north to south. The subducted slab beneath the northern tip of Sumatra barely arrives at the 410-km discontinuity, while the slab in the south penetrates to a depth of at least 660 km. Our inversion further indicates that the subducted slab is characterized by a tear at a depth of 120 km between 0° N and 2° N, which may be closely related to the supervolcanic eruption of the Toba caldera in northern Sumatra during the Pleistocene. Moreover, we report that the dip of the subducted slab varies significantly (i.e., dramatically decreases) across the Sunda Strait; therefore, we infer that another subvertical slab tear exists beneath the Sunda Strait.

1. Introduction

Sumatra is located on the western margin of Sundaland (Figure 1). According to plate reconstruction, Sumatra has been situated at its current location since the Late Jurassic (Hall, 2012). With the closure of the Neotethys Ocean, the intra-oceanic Woyla Arc accreted onto the western margin of Sundaland, forming present-day Sumatra (Advokaat et al., 2018; Hall, 2012, 2015) during the Late Cretaceous. Subduction beneath Sumatra was minimal from the Late Cretaceous to the Early Eocene, and Australia remained close to Antarctica during this period (Hall, 2012). Based on magnetic and deep-sea drilling data, the subduction of the Indo-Australian plate beneath Sumatra commenced at about 45 Ma when Australia separated from Antarctica (Hall, 2012). At that time, oblique subduction began in the Sumatra subduction zone (Beaudry & Moore, 1985; Sclater & Fisher, 1974). Studies on the deposition of sediment offshore of northern Sumatra and in the vicinity of the Sunda Strait indicate that rapid subduction may have begun in the early Miocene (~20 Ma) (Beaudry & Moore, 1985; Malod et al., 1995). The convergence rate may have peaked in the Pliocene, as was inferred from the intense andesitic volcanism during that period; a high rate of convergence has characterized this subduction zone ever since (Hochstein & Sudarman, 1993). At present, the rate of oblique convergence in the Sumatra subduction zone is approximately 7.2 mm/yr in the south and decreases to 5.2 mm/yr in the north (Figure 1) (Sieh & Natawidjaja, 2000). Moreover, unlike the oblique subduction at the Sumatra trench, the subduction occurring in western Java is normal to the trench axis (Figure 1); consequently, there is an abrupt transition from normal subduction to oblique subduction at the Sunda Strait (Malod et al., 1995). As pointed out by Malod et al. (1995), this abrupt transition is governed by the convergence motion of the Indo-Australian and Southeast Asian plates.

Figure 1. Tectonic setting of Sumatra. The red rectangle includes the study area, the global position of which is shown in the inset map. The red triangles denote active and Quaternary volcanoes (Acocella et al., 2018; Sieh & Natawidjaja, 2000). The inverted triangles denote the 57 broadband seismic stations used in this study. The black, blue and purple triangles indicate the IA, MY and MS networks operated by the Indonesian Agency for Meteorology, Climatology and Geophysics (BMKG), Malaysian National Seismic Network (MNSN) and Singapore National Network (SNN), respectively. The magenta lines represent the Sumatran Fault (Sieh & Natawidjaja, 2000), the dashed coral line describes the Batee Fault (Barber, 2000), and the coral line denotes the Mentawai Fault (Wiseman et al., 2011). These three faults constitute the Sumatran Fault System (SFS) (Sieh & Natawidjaja, 2000). The barbed orange line denotes the Sunda Trench. The cyan line delineates the fused Indo-Australian plate boundary (Bird, 2003; Richards et al., 2007). The red line represents the Wharton Fossil Ridge (WFR) (Shapiro et al., 2008). The dotted red lines indicate the location of the Investigator Fracture Zone (IFZ), which consists of four subridges and is subducting below the Sunda plate along with the Indo-Australian plate (Lange et al., 2010; Koulakov et al., 2016). The orientation of the IFZ is nearly parallel to the convergence vector near the trench. The black arrows indicate the motion of the Indo-Australian plate relative to a fixed Eurasian plate (GPS observations) (Carton et al., 2014; Dokht et al., 2018; Kennett & Cummins, 2005; Richards et al., 2007; Sieh & Natawidjaja, 2000). The convergence rate decreases from south to north within the Sumatra subduction zone. The plate subduction angle is normal in western Java, but becomes oblique across the Sunda Strait in Sumatra (Malod et al., 1995). The angle of obliquity gradually increases from south to north. In the forearc region, the four megathrust earthquakes labeled by purple stars from the NW to the SE are the Sumatra-Andaman earthquake (2004 Mw 9.2), Nias-Simeulue earthquake (2005 Mw 8.7), Mentawai earthquake (2010 Mw 7.8) and Bengkulu earthquake (2007 Mw 8.4). The age of the oceanic lithosphere near the Sunda Trench (Müller et al., 2008) is depicted in different colors, and the scale is shown at the bottom.

In addition to the oblique subduction, the subducted slab in the Sumatra subduction zone is characterized by other features, for example, the slab is relatively young (~ 30 – 90 Ma) (Figure 1) (Carton et al., 2014; Müller et al., 2008) and the Investigator Fracture Zone (IFZ) (Fauzi et al., 1996; Lange et al., 2010) and Wharton Basin (Kopp et al., 2008) are subducting along with the Indo-Australian oceanic lithosphere. These key features are believed to have played critical roles in the geological and geophysical processes throughout Sumatra, including the occurrence of widespread seismicity (Lange et al., 2010; Pesicek et al., 2010b), the eruption of numerous volcanoes (such as the Earth's largest Quaternary volcanic eruption at the site of the Toba caldera) (Acocella et al., 2018; Chesner & Luhr, 2010; Koulakov et al., 2016), the development of fault zones on the mainland (Sumatra Fault) and forearc slivers (Mentawai Fault and Batee Fault) (Sih & Natawidjaja, 2000), the evolution of the forearc basin (Beaudry & Moore, 1985; Malod et al., 1995), the deformation of the trench system (Kopp et al., 2008) and the partitioning of strain within the overriding plate (McCaffrey et al., 2000).

The hypocenters of slab-related earthquakes can effectively delineate the shape of the slab beneath Sumatra (Fauzi et al., 1996; Lange et al., 2010; Pesicek et al., 2010a; Richards et al., 2007). Using the arrival times of P- and S-waves recorded by a seismography network surrounding the Toba caldera for the relocation of slab-related earthquakes, Fauzi et al. (1996) inferred that the slab beneath northern Sumatra is arcuate at depths ranging from 50 to 200 km and generally parallels to the Sumatra trench and volcanic arc. Furthermore, by analyzing the strike of the slab surface near Toba, they speculated a vertical tear beneath the Toba caldera, although more conclusive evidence was required to verify their findings (Fauzi et al., 1997). In a similar study, Lange et al. (2010) deployed more seismic stations to investigate the topography of the subducted slab beneath western Sumatra. Using a dense, amphibious local seismic network to constrain the locations of earthquakes, Lange et al. (2010) observed that the IFZ, which is characterized by rough bathymetry and contains seamounts standing 1,000–1,500 m above the surrounding seafloor (Koulakov et al., 2016), is subducting beneath Toba and the region south of Toba. Moreover, their results showed that the rough topography of the subducted slab influences the occurrence of seismicity at all depths but particularly enhances seismic activity at intermediate depths (80–200 km). Except for the slab topography, little information about the large-scale structure of the subducted slab was revealed. However, the subduction of the IFZ probably has a considerable impact on the morphology of the subducted slab (Koulakov et al., 2016; Liu et al., 2019). In addition to the locations of regional earthquakes, the hypocenters of teleseismic earthquakes throughout the Sumatra subduction zone were determined by Pesicek et al. (2010b) in a 3-D heterogeneous model using the teleseismic double difference method; their results clearly revealed that the dip of the subducted slab south of Toba is steeper than that north of Toba at shallow depths (< 300 km) (Pesicek et al., 2010b). Nevertheless, the focal depths beneath Sumatra occur mostly above 300 km because the subducted oceanic lithosphere is relatively young (only the slab beneath southernmost Sumatra is of Late Cretaceous age) (Widiyantoro & van der Hilst, 1996; Wortel & Vlaar, 1988); thus, it is impossible to infer the slab geometry at greater depths (> 300 km) by slab-related seismicity alone.

In addition to employing slab-related earthquakes, the geometry of the subducted slab beneath Sumatra has also been investigated by seismic tomography (Huang et al., 2015; Li et al., 2018; Pesicek et al., 2008; Puspito et al., 1993; Richards et al., 2007; Widiyantoro & van der Hilst, 1996, 1997; Zenonos et al., 2019). Accordingly, general conclusions regarding the slab shape have been drawn mainly based on global tomography models (Amaru, 2007; Bijwaard et al., 1998; Hall & Spakman, 2015; Lu et al., 2019; Obayashi et al., 2013; Replumaz et al., 2004). Although overlapping datasets, such as the commonly used Engdahl-van der Hilst-Buland (EHB) dataset (Engdahl et al., 1998), have been utilized by different seismic tomographic inversions, the slab geometries beneath Sumatra extracted from global tomography models are not consistent, which may be the consequence of adopting different inversion approaches (Lu et al., 2019). For example, Bijwaard et al. (1998) employed an irregular grid to parameterize the Earth's mantle and inverted the EHB dataset (P, pP and pwP phases); their tomographic inversion resulted in a broadly arcuate high-velocity zone that extended from the surface to at least 810 km depth from Myanmar to Sunda. However, the resolution of their inversion was not high enough to demonstrate the continuity of the subducting slab material. The high-velocity anomalies may also have been smeared out (Plate 1 in Bijwaard et al., 1998). By combining refracted and diffracted core phases (PKP and Pdiff) with the EHB dataset to image the mantle structure, Kárason and van der Hilst (2001) obtained another global mantle model with improved resolution. Because of this improved resolution, the model of Kárason and van der Hilst (2001) was subsequently utilized by Replumaz

et al. (2004) for plate reconstruction over the past 50 Ma by analyzing the position of the downgoing slab at different depths. However, this model revealed little evidence of a slab-related signature at shallow depths (<300 km) (Káráson and van der Hilst, 2001; Replumaz et al. 2004). In contrast, the slab was clearly imaged with enhanced amplitudes at depths exceeding 300 km (Replumaz et al., 2004). Evidently, the slab is parallel to the trench at a depth of 400 km. Furthermore, the tomographic results of Káráson and van der Hilst (2001), clearly demonstrate that the slab beneath Sumatra penetrates to a depth of at least 800 km. However, the slab-related high-velocity anomalies are poorly defined at depths of 100–800 km. Subsequently, by inverting more than twice as many P-wave traveltimes as used by Bijwaard et al. (1998), Amaru (2007) developed a global mantle model, which was recently employed by Hall and Spakman (2015) as an auxiliary tool to interpret the tectonism of an area centered on Indonesia. The high-velocity anomaly in the tomographic images of Amaru (2007) exhibits lateral contortion at 300–500 km depth. Nearly the same images were obtained by Pesicek et al. (2008, 2010a), which were interpreted as folding, while Hall and Spakman (2015) used these same images as evidence to claim that the slab was torn beneath Toba caldera (their p. 27 and Figure 8).

In addition to global tomography models, regional tomographic studies have also been conducted to explore the slab geometry beneath Sumatra (Li et al., 2018; Pesicek et al., 2008, 2010a; Puspito et al., 1993; Widiyantoro & van der Hilst, 1996, 1997; Zenonos et al., 2019). A pioneering study was carried out by Puspito et al. (1993), who presented P-wave seismic tomographic images of the mantle beneath the Indonesian region by performing a joint inversion of regional seismic and teleseismic traveltimes. Their study illustrated that the disconnected slab-related high-velocity anomalies in the upper mantle are separated by low- and normal-velocity blocks. However, because only six seismological stations located in Sumatra were available for the tomographic inversion (Puspito et al. 1993), the revealed slab shape may not be reliable. To increase the resolution of regional models centered on Indonesia, Widiyantoro and van der Hilst (1996, 1997) employed a nested regional-global model-parameterization strategy with a coarse grid outside the study area and a fine grid inside the study area to invert global traveltime data. They found a pronounced feature, namely, detachment beneath Sumatra, and determined that the size of the vertical gap separating the shallow slab from the detached slab increases northward (Widiyantoro & van der Hilst, 1996). Nearly one decade later, the tomographic inversion process utilized by Widiyantoro and van der Hilst (1996) was adopted by Pesicek et al. (2008) to invert an expanded global seismic dataset containing earthquakes that occurred in the Sumatra region during 1918–2006. Because they employed a large number of earthquakes occurring in Sumatra for their tomographic inversion, a high imaging resolution was achieved, and as a result, they were the first to propose a model featuring significant slab folding beneath northern Sumatra (Pesicek et al., 2008). Moreover, their model contains a laterally continuous slab (i.e., featuring no slab detachment) penetrating into the mantle transition zone (MTZ). More recently, Li et al. (2018) performed teleseismic traveltime tomography using data recorded predominantly by temporary seismic stations. They reported a widening slab gap with depth in the MTZ beneath southern Sumatra. A similar result was reported by Huang et al. (2015), who suggested that some portion of the subducted slab beneath southern Sumatra does not penetrate into the MTZ.

As mentioned in the above review of the existing literature, various slab models have been proposed, but because of the insufficient resolution of these models, some key features remain ambiguous. To resolve these discrepancies and explore the slab geometry in greater detail, we conduct regional and teleseismic traveltime tomography to image the crust and mantle structure of Sumatra to a depth of 800 km by applying an eikonal equation-based seismic tomography method (Liu et al., 2018, 2019; Rawlinson et al., 2006; Rawlinson & Sambridge, 2005). Accordingly, we image the subducted slab with a high resolution by inverting a large volume of seismic data recorded by a dense network of broadband seismic stations with good coverage. The geometry of the subducted slab revealed by our study may be essential for interpreting the tectonic evolution of SE Asia.

2. Data and Method

2.1. Data

Indonesia has been working toward establishing a tsunami early warning system since the 2004 Mw 9.2 Sumatra-Andaman megathrust earthquake. The terrestrial component of this early warning system consists of more than 160 broadband seismometers scattering throughout the country (Lauterjung & Letz, 2017).

These seismometers (named the IA network) are maintained by the Indonesian Agency for Meteorology, Climatology and Geophysics (BMKG), and most have been in operation since 2009. Fifty-three seismic stations from the IA network, which are located on mainland Sumatra, western Java, the Sumatran forearc islands, and the islands southeast of Sumatra (Figure 1), are included in our tomographic study. To increase the data coverage, we also include three broadband seismometers from the Malaysian National Seismic Network (MY network) and one from the Singapore National Network (MS network). As shown in Figure 1, all 57 seismic stations have a uniform distribution over the study area.

We obtained a regional earthquake catalog for the period from January 2009 to August 2019 from the International Seismological Centre (ISC) (www.isc.ac.uk). All the earthquakes in this catalog have a magnitude exceeding 2.5. We downloaded the waveform data of the regional earthquakes recorded by the IA network from BMKG, while the waveform data recorded by both the MY and MS networks were downloaded from the Incorporated Research Institutions for Seismology (IRIS) using the Standing Order for Data (SOD) tool (Owens et al., 2004). After the means and trends were removed, the waveform data were further processed with a bandpass filter between 1.5 and 5 Hz. Subsequently, we manually picked the arrival times of the first P-waves. To ensure that the picked arrivals are of high quality, two individuals manually and independently picked the first arrivals. Each first arrival is retained only when the difference between its two manually picked arrival times is less than 0.2 s, and then the average of the two first arrival times is treated as the observed arrival time in the subsequent inversion. According to our earthquake selection criterion, at least 7 measurements must be retained for each event. Consequently, a total of 21,013 first P-wave arrival times from 2,112 regional earthquakes were selected (Figure 2a).

Unlike the regional earthquake catalog, the teleseismic event catalog was downloaded from the IRIS Data Management Center (http://ds.iris.edu/wiber3/find_event). We also downloaded teleseismic waveform data from both BMKG and IRIS. Then, we processed the raw data by deconvolving the instrument response and removing the means and the trends. Subsequently, we filtered the waveform data in the frequency range between 0.03 and 0.3 Hz. Each event was consistently recorded by at least 5 stations in the study area. We computed the relative traveltime difference at every possible pair of seismic stations using the waveform cross-correlation method. To ensure imaging accuracy, only those relative traveltime differences with corresponding cross-correlation coefficients greater than 0.9 were retained for the subsequent inversion (Liu et al., 2018). To reduce the imaging errors caused by heterogeneities outside the study domain, the allowed distance for a station pair must be properly set (Liu et al., 2018). On the one hand, a significantly short distance may result in very few traveltime observations; on the other hand, an excessively large distance may considerably reduce the imaging accuracy. Thus, there is a tradeoff between the imaging accuracy and the allowed maximum distance of a station pair (Liu et al., 2019). We set the maximum allowed distance to 2.5°, which is slightly larger than the distance of any two nearest seismic stations for all 57 stations. Finally, a total of 27,524 cross-correlation traveltime differences from 775 teleseismic events were used as the input data for the tomographic inversion. The distribution of the teleseismic events is shown in Figure 2b.

2.2. Method

Before carrying out the inversion for the P-wave velocity structure, we relocate regional earthquakes by inverting the regional traveltime data in a 3-D initial model M0, which is derived from the global ak135 model (Kennett et al., 1995) and the CRUST1.0 model (Laske et al., 2013) for crustal corrections. Then, we conduct regional seismic traveltime tomography by inverting the updated regional traveltime data, i.e. the traveltime data with updated hypocenters. This process yields model M1, the shallow part (<200 km depth) of which is well imaged by regional seismic data (Liu et al., 2019). Because most of the regional earthquake occur at depths less than 200 km (Figure 2a), the deeper part (>200 km depth) is not well imaged by regional seismic data. M1 is then treated as the initial model for teleseismic traveltime tomography. At last, we obtain the final model, denoted by M2. The tomographic results presented in the subsequent sections are the relative velocity perturbations, i.e., the velocity difference between M2 and M0 divided by the velocity of M0.

Based on the Bayesian optimization criterion (Nolet, 2008), the damped least square problems for earthquake relocation, regional seismic traveltime tomography, and teleseismic traveltime tomography can be uniformly expressed as follows (Puspito et al., 1993):

Figure 2. Earthquakes selected for the tomographic study. (a) Epicentral distribution of the 2,112 regional earthquakes selected for this investigation (colored dots), the first arrivals of which are consistently recorded by at least 7 stations. The colors in (a) denote the focal depth, whose scale is shown at the bottom of (a). (b) Polar projection map showing the distribution of the 775 teleseismic earthquakes (red dots), the first arrivals of which are consistently observed by more than 5 stations. The red triangle in (b) represents the center of the study area. The three concentric circles depict epicentral distances of 30°, 60° and 90° from the center of the study area. The epicentral distances of the selected teleseismic events are between 30° and 90°.

$$E(\mathbf{X}) = (\mathbf{AX} - \mathbf{b})^T \mathbf{C}_d^{-1} (\mathbf{AX} - \mathbf{b}) + \eta \mathbf{X}^T \mathbf{C}_m^{-1} \mathbf{X}. \quad (1)$$

In Equation (1), $E(\mathbf{X})$ is the objective function, $\mathbf{A}$ is the sensitivity matrix, $\mathbf{b}$ is a vector measuring the difference between the observed arrival time and the synthetic arrival time (or the difference in relative traveltime for teleseismic tomography), η is the damping factor, $\mathbf{C}_d$ is the a priori data covariance matrix, and $\mathbf{C}_m$ is the a priori model covariance matrix. In earthquake relocation $\mathbf{X}$ is a vector containing the hypocenter parameters, i.e., the origin time t_s and location coordinates (r_s, θ_s, ϕ_s) of each regional earthquake, where s denotes the source index and r , θ and ϕ represent the depth, latitude and longitude, respectively. For both regional and teleseismic traveltime tomography, $\mathbf{X}$ represents the relative model perturbations on discrete grid nodes (Tong et al., 2019). The sensitivity matrix $\mathbf{A}$ can be computed when the ray paths of the P-waves are available (Rawlinson et al., 2010; Tong et al., 2019; Zhao et al., 1992). To trace the ray paths emitted from one earthquake, we first compute the traveltime field in a 3-D velocity model by solving the eikonal equation (Rawlinson & Sambridge, 2005; Sethian & Popovici, 1999). The fast marching method, which combines an upwind scheme with a fast sorting algorithm (Hassouna & Farag, 2007), can efficiently solve the eikonal equation for the traveltime field in strong heterogeneous media (Sethian & Popovici, 1999). After acquiring the 3-D traveltime field, we can determine every single ray path by tracing the negative gradient direction of the traveltime field starting at the receiver location. We employ M0 as the initial model, but we do not consider any prior model information. Therefore, we set $\mathbf{C}_m$ in Equation (1) to be an identity matrix. For regional seismic traveltime tomography, the data uncertainties are estimated according to the differences between the picked arrivals by the two abovementioned individuals. In contrast, for teleseismic traveltime tomography, the data uncertainties are determined by using the multichannel cross-correlation method (Allen et al., 2002; Vandecar & Crosson, 1990), and are incorporated into the diagonal elements of $\mathbf{C}_d$ to weight the relative importance of each residual in the inversion (Rawlinson & Kennett, 2004; Rawlinson et al., 2006).

The conjugate gradient method (Hestenes & Stiefel, 1952; Tong et al., 2014) is utilized to solve Equation (1) in an iterative fashion. In our real data inversion, the number of iterations for earthquake relocation, regional seismic traveltime inversion and teleseismic traveltime inversion are 5, 15 and 15, respectively. In

each iteration, we adjust the damping factor η to restrict the maximal relative location/velocity perturbation below a prescribed value (such as 1%). The essential rule of this strategy (establishing a prescribed value) is to steadily evolve the model toward the final optimal model (Liu et al., 2018; Tong et al., 2017). To determine the optimal prescribed value, we conduct a series of tomographic inversions with different prescribed values. Thereafter, an L-curve, which describes the tradeoff between model variance and data variance (Eberhart-Phillips et al., 1986; Rawlinson et al., 2006), can be drawn. The optimal prescribed value is found in the vicinity of the point with maximum curvature. Notably, a smaller prescribed value requires more iterations; in other words, the final inversion result is generally independent of the number of iterations, which has been confirmed by a previous study (Liu et al., 2019). We prefer a large number of iterations to achieve a robust inversion. The basic assumption for seismic tomography is that the initial model effectively approximates the target model (Fichtner, 2011), and the maximal relative velocity perturbation should not exceed 10% in a regional-scale model (Koulakov et al., 2009, 2016; Li et al., 2018; Masturyono et al., 2001). Therefore, it is appropriate to set the number of iterations in the interval [5, 20] and choose the prescribed value in the interval (0, 0.2] for the inversion of velocity. We further transform the origin time of the earthquake and its spatial position into relative perturbations for earthquake relocation (Liu et al., 2019). The same intervals for the iteration number and prescribed value are appropriate for the relocation of earthquakes.

3. Checkerboard Resolution Test

In this section, we discuss the results of checkerboard tests to assess the imaging resolution. A grid approach is used to parameterize the model space (Jiang et al., 2009; Tong et al., 2019). The grid spacing in the horizontal direction is approximately 1.2° , while the nodes of the inversion grid in the vertical direction are located at depths of 15, 60, 120, 200, 300, 450, 600 and 800 km. The inversion grid utilized herein is determined after performing tens of checkerboard resolution tests. In these tests, we adjust the horizontal and vertical intervals and visually compare the imaging results. This strategy for setting the inversion grid is the same as the strategies employed by Jiang et al. (2009) and Liu et al. (2018). To construct the checkerboard model, we assign alternative positive and negative relative velocity perturbations of $\pm 4\%$ to the 3-D grid nodes. We calculate synthetic data in the checkerboard model by solving the eikonal equation. Then Gaussian noises with a standard deviation of 0.2 s are added to the synthetic data. Following the same inversion process as that employed by the real data inversion, we obtain the results of the checkerboard resolution test, which are shown in Figure 3.

Figure 3 illustrates that the checkerboard pattern of the target model has been recovered, especially beneath mainland Sumatra. At a relatively shallow depth (15 km, Figure 3a), the velocity perturbations near the trench and beneath the offshore area are well resolved because of the active seismicity in these areas (Figure 2a). In the crust and uppermost mantle (Figure 3a-c), the imaging resolution decreases as the depth increases, which is mainly due to the reduced seismicity in the uppermost mantle (Figure 2a). Because most regional earthquakes occur at depths less than 200 km (Lang et al., 2010; Pesicek et al., 2010b) (Figure 2a), the velocity perturbations at depths from 200 to 800 km are resolved mainly by the teleseismic data (Figures 3d-h). As shown in Figure 2b, most of the teleseismic earthquakes occurred in the Kuril, Japan, Izu-Bonin, Mariana, Solomon and Tonga subduction zones; in contrast, very few earthquakes occurred in the Indian Ocean. Thus, the imaging resolution in the region east of the volcanic arc is higher than that in the west.

4. Results

We present the map and cross-section views of the final P-wave velocity model in Sections 4.1 and 4.2, respectively. The evaluation of the velocity model is discussed in Section 4.3. As mentioned above, the L-curve approach is utilized to define the upper bound of the relative perturbation in each iteration. In the real data inversion, the upper bounds for earthquake relocation, regional traveltime tomography and teleseismic traveltime tomography are set to be 0.7%, 1% and 0.9%, respectively (Figures 4a, c and e). As shown in Figure 4b, the mean and standard deviation of the regional seismic traveltime misfits are greatly reduced after earthquake relocation, changing from 1.16 ± 1.34 to 0.05 ± 0.81 sec. Further reductions in the

misfits are achieved after conducting regional seismic traveltime tomography; the final mean and standard deviation are 0.04 ± 0.58 sec. For teleseismic differential traveltime residuals, the mean and standard deviation after teleseismic traveltime tomography decrease from 0.16 ± 0.78 to 0.06 ± 0.43 sec (Figure 4f). The epicenter shifts relative to the ISC locations after earthquake relocation are demonstrated in Figure S1 in the supporting information.

4.1. Map Views

The tomographic model at nine representative depths are shown in Figure 5. The positive maximum and negative minimum of the relative velocity perturbation reach about $\pm 4\%$, which are larger than those reported in previous studies (Hall & Spakman, 2015; Li et al., 2018; Pesicek et al., 2010a). At a relatively shallow depth (15 km), the areas in the forearc region, including those near the islands of Simelue, Pulau Pini, Pagai and Enggano, are generally characterized by low-velocity anomalies (Figure 5a). These negative velocity anomalies may be attributed mainly to the thick accretionary prism near the trench (Kopp et al., 2008) and the sediments deposited in the forearc region since the Neogene (Beaudry & Moore, 1985). Low-velocity anomalies at this depth in the forearc of central Sumatra (at latitudes ranging from -5 to 3°N) were discovered by a previous tomography study based on the inversion of regional earthquake traveltimes (Lange et al., 2018). Lange et al. (2018) also revealed moderately elevated VP/VS values (up to 1.85) near the Mentawai Fault, and speculated that such high VP/VS ratios are related to the fluids in this area traveling along the pathway created by the Mentawai Fault. If their speculation is correct, saturating fluids traversing the Mentawai Fault could be another factor causing the low-velocity anomalies in the forearc region. As shown in Figure 5a, prominent low-velocity anomalies are observed beneath the volcanoes in central and southern Sumatra. These low-velocity anomalies may be interpreted as the product of heat flow or magma chambers within the crust, both of which are readily relatable to the high-temperature geothermal systems in this area (Hochstein & Sudarman, 1993). From our tomographic results, low-velocity anomalies with amplitudes of about 2% are located only north of the Toba caldera and are absent beneath and south of the Toba caldera. However, a previous regional traveltime tomography model with a horizontal scale of approximately 100 km showed strong low-velocity anomalies beneath Toba (Masturyono et al., 2001). The same data used to create the model reported by Masturyono et al. (2001) were later inverted by Koulakov et al. (2009), who revealed low-velocity anomalies beneath Toba and the Sumatran Fault to the SW of the Toba caldera (Koulakov et al., 2009). Moreover, gravity data reflect remarkable negative Bouguer gravity anomalies in the Toba area and its vicinity (Masturyono et al., 2001), indicating a mass deficit and low-velocity anomalies in these areas. The absence of low-velocity anomalies in this area may be attributed to the relatively coarse grid used for the model parameterization, in which case local-scale features are difficult to be resolved by our tomographic inversion.

Low-velocity anomalies with increased amplitudes extend to a depth of 60 km beneath central and southern Sumatra (Figure 5b). In addition, low-velocity anomalies with amplitudes larger than 1% exist south of Toba, and they are smaller in magnitude than those presented in a previous study of Liu et al. (2019), although the data used in the previous study are all included in the present study. High-velocity anomalies are detected beneath the offshore area, and probably represent the cold oceanic lithosphere that is currently subducting. These high velocities are laterally discontinuous and are separately located west of Aceh Province, beneath the islands of Batu and Siberut, and west of the provinces of Banten and West Java. The amplitudes of the high-velocity anomalies west of western Java are much greater than those beneath offshore Sumatra, with the maximal amplitude exceeding 4%. These pronounced positive velocity anomalies may be due to the relatively old age of the lithosphere (>100 Ma) (Müller et al., 2008; Widiyantoro & van der Hilst, 1996). Because the temperature contrast between the subducted slab and the ambient mantle is small at a depth of 60 km (Zheng et al., 2016), beneath some offshore areas, the slab will not necessarily exhibit anomalously fast VP velocities and thus seems to be discontinuous (Figure 5b).

At depths ranging from 120 to 450 km, the most prominent feature is a northeastward dipping high-velocity body interpreted to be the subducting slab (Figures 5c-f): (1) the strike of the slab is generally parallel to the Sumatra trench and volcanic arc; (2) the slab itself exhibits a sinusoidal shape; and (3) the subducted slab continuously penetrates to a depth of about 410 km. Another striking feature of the slab is detected in

Figure 3. Checkerboard resolution test. A model with alternating seismic velocity anomalies of +4% and -4% is employed as the input model for the synthetic test. The fast marching method is used to generate the synthetic data in the input model. Gaussian noise with a standard deviation of 0.3 s is added to the synthetic data. The inversion process is the same as that employed for the real data that is, the regional synthetic data inversion followed by the teleseismic synthetic data inversion. The depth is shown in each map. The color scale is presented in the lower-right corner of the plot.

the uppermost mantle at a depth of 120 km: additional displacement in the northeast direction occurs in the slab segment north of the Toba caldera compared with that south of Toba (Figure 5c). This additional displacement is probably related to a gap beneath Toba that separates the north and south slab segments (Koulakov et al., 2016; Liu et al., 2019).

As shown in Figures 5g and 5h, prominent positive velocity anomalies are not observed at depths from 600 to 700 km to the north of the equator. However, we do detect high-velocity anomalies to the south of the equator, albeit with amplitudes less than those in the upper mantle (Figure 5c-5f). Therefore, our tomographic results indicate that the subducted slab in the south (latitude $< 0^\circ$) penetrates to a greater depth than that in the north (latitude $> 0^\circ$). In addition, the subducted slab beneath southern Sumatra is laterally continuous in the MTZ.

At a depth of 800 km, high-velocity anomalies are located beneath the Strait of Malacca, the northern portion of southern Sumatra and the western Java Sea (Figure 5i), and the amplitudes of the high-velocity anomalies beneath the Strait of Malacca are larger than those beneath the other two areas. Generally, the amplitudes of the high-velocity anomalies are less than 1% and therefore lower than the amplitudes found at relatively shallow depths (Figures 5c-5h). Because of the extremely faint high-velocity perturbation (Figure 5i), it is difficult to discern the geometry of the subducted slab at 800 km depth. Some high-velocity anomalies may represent the subducted slab beneath southern Sumatra extending to a depth of 800 km, while some anomalies are probably due to smearing effects (Liu et al., 2018). Because the resolution in the southeastern part of our study area at a depth of 800 km is generally good and no high-velocity bodies are found there (Figure 3h), it is likely that the subducted slab beneath southern Sumatra does not consistently penetrate to a depth of 800 km. Again, no visible high-velocity perturbations are observed to the north of the equator at 800 km depth, similar to our findings at depths of 600 and 700 km (Figures 5g and 5h).

4.2. Cross-Sections

In Figure 6, we present tomographic images of eight cross-sections oriented nearly along the convergence direction of the Indo-Australian plate near the trench (Figure 1). At shallow depths (< 300 km), the upper boundary of the imaged high-velocity slab generally correlates well with the slab-related seismicity (white dots in Figure 6). The focal depths of almost all earthquakes in our study area are shallower than 300 km (Widiyantoro & van der Hilst, 1996; Wortel & Vlaar, 1988), while only a few exceptions are at a depth of about 600 km beneath southern Sumatra and the western Java Sea (readers can also refer to Figure 2 in Zenonos et al. (2019)). Hence, it is difficult to investigate the geometry of the subducted slab based only on earthquake hypocenters, especially at deeper depths (> 300 km). Nevertheless, we can observe three important features embedded in our tomographic results (Figure 6).

First, the slab on profiles AA'-CC' (to the northwest of the Toba caldera) with a relatively shallow dip crosses the 410-km discontinuity but seems not to penetrate the 660-km discontinuity. In particular, beneath the northern tip of Sumatra, the slab appears to terminate at the 410-km discontinuity (Figure 6a), which is different from the slab imaged by Richards et al. (2007) that has penetrated into the uppermost lower mantle beneath the Andaman Sea. However, the geometry of the slab on profile AA' is consistent with that of Liu et al. (2019), for that almost the same regional and teleseismic traveltime data are used to image the overlapped study area. On profiles BB' and CC', high-velocity anomalies with amplitudes of less than 1% extend to the 660-km discontinuity (Figures 6b and 6c). Because the focal depths of the regional earthquakes used in northern Sumatra are mainly shallower than 200 km (Figure 2a), velocity perturbations at depths exceeding 200 km are derived mainly from the teleseismic traveltime data. A large portion of the teleseismic data of the present study were used by a previous teleseismic traveltime tomography study of northern Sumatra (Liu et al., 2018), which confirmed the spatial smearing in the MTZ via a discrete spike test. Similarly, we consider that the extremely faint signatures of the anomalies in the MTZ on profiles BB' and CC' (Figures 6b and 6c) are attributed to the smearing effect. In comparisons with the slab derived from teleseismic data alone (Liu et al., 2018), the slab on profiles AA'-CC' has a shallower penetration depth.

Second, beneath central (profile DD', Figure 6d) and southern Sumatra (profiles EE'-GG', Figures 6e-6g), the penetration depth and dip angle of the slab gradually increase from north to south. On profile DD', the

Figure 4. L-curve used to define the prescribed perturbation value in the traveltime inversion (left plots) and histogram of traveltime residuals before (in green) and after (in red) the inversion (right plots). The L-curve (model variance versus data variance) is formed by choosing a spectrum of prescribed perturbation values. (a, b), (c, d) and (e, f) are for the earthquake relocation, regional earthquake traveltime inversion and teleseismic relative traveltime inversion, respectively. The red dots in (a), (c) and (e) mark the determined optimal perturbation values of 0.7%, 1% and 0.9%, respectively. The averages and standard deviations of the data residuals before (green characters) and after (red characters) the inversion are shown in each plot on the right-hand side of the figure.

slab penetrates to a depth larger than 500 km (Figure 6d); farther south, on profiles EE' and FF', the slab reaches a depth of at least 600 km; on profile GG', the slab arrives at the 660-km discontinuity. Moreover, the dip of the slab on profile DD' (Figure 6d) is slightly steeper than that on profile CC' and increases progressively southward and becomes nearly vertical on profile GG' (Figure 6g).

Figure 5. Map views of the tomographic results at nine depths. The velocity perturbations are relative to the global stratified model ak135 with CRUST1.0 for crustal corrections. The color scale is shown at the bottom. The red and blue colors represent low and high relative P-wave velocity perturbations, respectively. According to the checkerboard test, the areas with relatively poor resolution are masked. The depth of each layer is shown on each plot. The circled numbers in plot (a) denote the main geographical locations at the Sumatra subduction zone. (c) The dash line represents the position of the inferred slab tear. The other labels are the same as those in Figure 1.

Third, the dip of the slab beneath western Java is relatively shallow (> 100 km depth), and the slab penetrates into the MTZ. The dip of the slab on profile HH' (Figure 6h) is much shallower than that on profile GG' (Figure 6g), and a dramatic decrease in the dip angle is observed across the Sunda Strait from west to east (Figures 6g and 6h). Our results demonstrate that a segment of the subducted slab is stagnant in the MTZ and that the slab does not penetrate into the lower mantle directly (Figure 6h). Because this segment is close to the edge of our study area, the size of the stagnant slab segment in the MTZ cannot be inferred from our tomographic results. The weak positive anomalies (amplitude less than 1%) in the lower mantle shown in Figure 6h mantle probably stem from vertical smearing.

4.3. Model Evaluation

In this section, we conduct five numerical tests to investigate the reliability of our inverted model. Furthermore, other tomographic models are employed for comparison. Following the model assessment strategy discussed in Tape et al. (2009), independent teleseismic events, which are not included in the tomographic inversion, are used by the numerical tests.

In the first test, we select a total of five teleseismic events with good azimuthal coverage (Figure 7a) for the model evaluation. The same data processing procedure employed in the previous real data inversion is applied in this test. The waveforms corresponding to one respective event in each plot of Figures 7b–7f are aligned at the theoretical first P-wave arrival times. We calculate the traveltime difference between each station pair by using the fast marching method (Sethian & Popovici, 1999). Then, we obtain the relative traveltime residual by subtracting the calculated traveltime difference from the observed traveltime difference. The calculated traveltime differences are computed in four velocity models, namely, the inverted model of the present study, the initial model M0, the global model UU-P07 (Amaru, 2007; Hall & Spakman, 2015) and the global model GAP-P4 (Obayashi et al., 2013). The histograms of the relative traveltime residuals are presented in Figures 7g–7j. In the initial model M0, the mean and standard deviation of the residuals are 0.098 ± 0.73 (Figure 7h), which decrease considerably to 0.024 ± 0.32 in the final model (Figure 7g). The mean and standard deviation in UU-P07 are 0.129 ± 0.56 (Figure 7i), which are close to those in GAP-P4 (0.118 ± 0.54) (Figure 7j). A direct comparison of the mean and standard deviation values in the four models indicates that our final model is more accurate than both UU-P07 and GAP-P4.

In the second test, we conduct a restoration test to verify the main features of the subducted slab as shown in our tomographic images (Figure 6). Following the strategy of Pesicek et al. (2010a), we construct slab models generally according to the tomographic results by setting the amplitudes of the high-velocity anomalies to be consistently 4% (Figure 8). In Figure 8, the first and third columns are the two synthetic models; the difference between these two synthetic models is that the slab in the first model is continuous beneath the Toba caldera (Figure 8a1), whereas the slab in the second model is discontinuous (Figure 8a3). We use the fast marching method to calculate the synthetic traveltime data and implement the same procedure adopted by the real data inversion to invert the synthetic data. Subsequently, we obtain the inverted results shown in the second and fourth columns in Figure 8, which correspond to the first and second synthetic models, respectively. Figure 8 demonstrates that the anomaly patterns are reliably restored and that the geometry of the slab is mostly recovered. In Figure 8a4, the gap beneath the Toba caldera is visible, which indicates that the gap shown in Figure 5c is a reliable feature. Because regularization is applied in the inversion, the amplitudes of the output models are less than those of the input models. We also present cross-sections of the inverted images on to examine the impact of vertical smearing. The input model (Figure S2a–h) is the same as that shown in Figures 8a1–f1. The penetration depth of the slab in the input model is 480 km north of the equator and increases to 660 km south of the equator (Figure S2a–h). The inverted images are displayed in Figure S2i–p, which resemble the input model. Although low-amplitude vertical smearing is observable on all cross-sections, the main shape and dip variation of the slab are well recovered.

The third synthetic test is employed to further investigate the vertical smearing effect of the tomographic inversion. As shown in Figures 9a–h, the synthetic model is similar as the one presented in Figure 8a1–f1, but the velocity perturbations at > 300 km depth are muted. We obtain the inverted model after applying the same inversion process used in the second test. As shown in Figures 9i–p, the synthetic slab is generally well recovered. In addition, vertical smearing is also visible on all profiles (Figures 9i–p). However, the am-

Figure 6. Tomographic results displayed on eight cross-sections, the locations of which are indicated in the lower-right map. The orientations of the cross-sections are generally parallel to the convergence direction of the Indo-Australian plate. The white dots depict seismic events located at an epicentral distance of <30 km from the profile; the event catalog is downloaded from the International Seismological Centre (ISC). The red triangles denote the volcanoes within a distance of 50 km on either side of each profile. The curved blue line in each profile represents the slab contour extracted from Slab1.0 (Hayes et al., 2012), and the two dashed lines denote the 410- and 660-km discontinuities. According to the checkerboard test, the areas with relatively poor resolution are masked. The color scale is shown at the bottom.

plitudes of the extended high-velocity bodies caused by vertical smearing are much smaller than those of the predefined slab-related structures, and the smearing just slightly extends a deeper depth (Figures 9i-p). This synthetic test demonstrates that the penetration depth of the subducted slab revealed by this study is reliable.

The fourth synthetic test is applied to evaluate the reliability of the low-velocity anomaly south of the Toba caldera at a depth of 60 km. Beginning with M0, we construct a P-wave velocity model by adding a -2%

low-velocity anomaly to the region south of the Toba caldera (Figure 10a). Then, synthetic traveltimes are calculated in the newly constructed model (Figure 10a). After that, Gaussian noises with a standard deviation of 0.2 s are added to the synthetic data. Finally, the inverted result is shown in Figure 10b, which resembles the input anomaly. Because damping regularization is applied in the inversion (Equation 1), the amplitude of the inverted anomaly is smaller than that of the input model. Beside the low-velocity anomaly south of the Toba caldera, we can also observe randomly distributed high- and low-velocity anomalies in Figure 10b. The amplitudes of these high- and low-velocity anomalies are generally smaller than 1%, and their sizes are smaller than 100 km (Figure 10b). Based on the current synthetic test, we believe that the low-velocity anomalies with amplitudes larger than 1% (south of Toba in Figure 5b) and a lateral scales greater than 100 km probably be reliable features.

In the fifth test, we choose two teleseismic events that occurred beneath the Indian Ocean to investigate whether the slab penetrates into the MTZ beneath southern Sumatra (Figure 11a). The horizontal slices in Figure 11b are extracted directly from our tomographic results. We mute the high-velocity anomalies in the MTZ beneath southern Sumatra to obtain another model that resembles the one reported by Huang et al. (2015). The ray paths of the two teleseismic events recorded by two stations, IA.PPBI and IA.TPI, sample the velocity structure in the MTZ beneath southern Sumatra. We plot the two rays from the first source (source 1 in Figure 11a) in Figure 11b. The ray paths and traveltimes are computed by using the fast marching method. Once we have the traveltimes, we simultaneously obtain the calculated traveltime differences between the station pair (IA.PPBI and IA.TPI). By subtracting the calculated traveltime differences from the corresponding observed traveltime differences (Figure 11c), we obtain the relative traveltime residuals. Figure 11d illustrates that the relative traveltime residuals of source 1 and source 2 in our model are 0.89 s and 0.81 s (Figure 11a), respectively, whereas the relative traveltime residuals in the model without the slab are 1.46 s and 0.97 s, respectively, which are much larger than those computed in our model. Based on this test, we may conclude that the slab probably penetrates into the MTZ beneath southern Sumatra.

5. Discussion

The subducted slab in Sumatra is characterized by variations in dip angle, two respective subvertical tears beneath the Toba caldera and Sunda Strait, and variations in the penetration depth. These features of the slab geometry are also summarized in Figure 12. The model in Figure 12 exhibits some differences from those in the literature.

5.1. Model Comparison

We conducted teleseismic traveltime tomography of northern Sumatra in 2018 (Liu et al., 2018). Subsequently, we inverted regional and teleseismic traveltime data to update the velocity model of northern Sumatra in 2019 (Liu et al., 2019). More teleseismic data were included in Liu et al. (2019) than in Liu et al. (2018), and the shallow structure was well defined by the inversion of regional seismic data, and therefore the imaging resolution is much higher in Liu et al. (2019) than that in Liu et al. (2018). A detailed comparison of the two previous tomographic studies can be found in Liu et al. (2019). The current study area also includes northern Sumatra, and thus it is necessary to compare the tomographic images of northern Sumatra generated by this study with those presented by our previous studies.

The tomographic images beneath northern Sumatra in Liu et al. (2018), Liu et al. (2019) and this study are presented in Figure S3. Because nearly the same dataset and the same inversion processes are adopted in both Liu et al. (2019) and this study to invert the velocity structures beneath northern Sumatra, the main features of the subducted slab delineated by the high-velocity anomalies in Figures S3a2-h2 and S3a3-h3 correlate well with each other. For example, the dip angle and penetration depth of the slab on the cross-sections in Figures S3a2-h2 and S3a3-h3 are nearly the same. Despite the common features in Figures S3a2-h2 and S3a3-h3, some differences remain between the new model (Figures S3a3-h3) and the previous model (Figures S2a2-h2). For instance, the amplitudes of the high- and low-velocity anomalies in Figures S2a2-h2 are greater than those in Figures S3a3-h3. In addition, the high-velocity anomalies on all the cross-sections in Figures S3a2-h2 are wider than those in Figures S3a3-h3.

Figure 7. Five independent teleseismic events were employed for the model evaluation. (a) Five teleseismic events (red dots) with good azimuthal coverage that occurred in 2019 are selected. The numbers near the red dots indicate the event number. (b)–(f) Waveforms are aligned according to the theoretical arrival time of the first P-wave calculated by the TauP Toolkit (Crotwell et al., 1999). Plots (b)–(f) correspond to teleseismic events 1–5. Plots (g)–(j) show the histograms of the residuals (the difference between the observed relative traveltime difference and the theoretical relative traveltime difference). The observed relative traveltime differences are obtained by applying cross-correlation to the waveforms shown in (b)–(f). The theoretical traveltime differences are calculated for four models using the fast marching method (Sethian & Popovici, 1999). (g) Our model; (h) M0; (i) the global model UU-P07 (Amaru, 2007; Hall & Spakman, 2015); (j) the global model GAP-P4 (Obayashi et al., 2013).

Figure 8. Restoration test. The synthetic slab is constructed based on the images shown in Figure 5. The slab shown in the first column (a1)–(f1) is continuous at a depth of 120 km, while the slab in the third column (a3)–(f3) is discontinuous at the same depth. For the synthetic slab, the relative velocity perturbation is fixed at 4%. The models in the first and third columns are treated as two input models. Then, synthetic regional and teleseismic traveltime datasets are generated by using the fast marching method for the input models. Subsequently, Gaussian noise with a standard derivation of 0.2 s is added to the synthetic data. The same process used for the real data inversion is implemented for the synthetic data inversion; then, the resulting images are obtained ((a2)–(f2) and (a4)–(f4)). The recovered slabs in the second and fourth columns correspond to the synthetic slabs in the first and third columns, respectively. The depths of the horizontal slices from the top to the bottom are 120 km, 200 km, 300 km, 450 km, 600 km and 700 km. The color scale is displayed on the bottom.

Figure 9. Restoration test for a model with shallow slab penetration. The input model (plots (a)–(h)) is constructed by setting the velocity perturbation at depths greater than 300 km to zero, while the velocity perturbations at depths above 300 km are the same as those in the model shown in Figures 8a1–f1. The inverted model is displayed in plots (i)–(p). The locations of the cross-sections are indicated by the blue lines in the map in Figure 6. The other symbols are identical to those in Figure 6.

Figure 10. Synthetic test to evaluate the low-velocity anomalies in the uppermost mantle at a depth of 60 km south of the Toba caldera. (a) Block with a low-velocity anomaly of -2% located south of the Toba caldera; the block serves as an input model to calculate the synthetic traveltime data. Gaussian noise with a standard deviation of 0.2 s is added to the synthetic traveltime data. (b) The inverted model is obtained by the inversion of the synthetic traveltime data. The black triangles are the stations; the red triangles are volcanoes. The color scale is shown at the bottom of the figure.

Overall, there are three noticeable differences between the new model (Figures S3a3-h3) and the model revealed only by teleseismic data (Figures S3a1-h1). First, the dip of the imaged slab is nearly vertical beneath the northern tip of Sumatra in Figures S3a1. However, the dip in Figure S3a3 is much shallower than that in Figure S3a1. Second, slab detachment is found on cross-section CC' in Figure S3c1, but is not revealed in the new model (Figure S3c3). Third, slab deflection at depths of 600–900 km is shown on cross-sections DD' and EE' (Figures S3d1 and S3e1), but this feature is not presented in the new model.

5.2. Slab in the Upper Mantle and MTZ

The geometry of the slab in the upper mantle derived from our tomographic inversion is different from the slab geometries described in the literature. Two major differences can be identified. First, the penetration depth of the subducted slab beneath northern Sumatra is deeper than previously reported. Ninkovich (1976) inferred that the geometry of the slab resembles the Benioff zone; furthermore, in his model, the maximum penetration depth of the subducted slab increases from 200 km in the north to 600 km in the south (Ninkovich, 1976). The shallow penetration depth in the north was attributed to the clockwise rotation of Sumatra, with its axis of rotation located in or near the Sunda Strait between the late Miocene and early Pliocene. Because of this clockwise rotation, subduction beneath northern Sumatra ceased during that period (Ninkovich, 1976). A similar slab model to that of Ninkovich (1976) was proposed by Widiyantoro and van der Hilst (1996) based on their tomographic images, which depicted a subvertical gap between the subducted slab in the upper mantle and a detached slab segment in the lower mantle. The size of this gap increases northward (Widiyantoro and van der Hilst, 1996). Our tomographic results (Figures 5c–f) reveal laterally continuous slab in the upper mantle, which is different from the models proposed by Ninkovich (1976) and Widiyantoro and van der Hilst (1996). The second major difference is that the sinusoidal slab in the upper mantle does not exhibit significant folding (Figures 5c–f). The folding of the slab beneath Sumatra was first proposed by Pesicek et al. (2008) on the basis of regional seismic tomography. They also suggested the occurrence of strong lateral deformation of the slab in the upper mantle and increasing slab curvature with depth (Pesicek et al., 2008). Slab folding was further revealed by their subsequent tomographic inversion with the incorporation of additional traveltime data (Pesicek et al., 2010a). Although the slab imaged by the teleseismic traveltime inversion performed by Liu et al. (2018) is not as significantly folded as that imaged by Pesicek et al. (2008), the curvature of the subducted slab inferred by Liu et al. (2018) is still a little greater than the slab curvature in the present model (Figures 5c–f).

Figure 11. Evaluation of the slab in the mantle transition zone (MTZ) beneath southern Sumatra. Two seismic stations (purple and green triangles, IA.PPBI and IA.TPI, respectively) are employed for this investigation. (b) The two stations are separately located on two islands east of southern Sumatra. (a) Two teleseismic events (red dots) in the Indian Ocean are selected. Source 1 occurred on February 18, 2009 (the latitude, longitude, depth, and magnitude are 53.0031° S, 20.8158° E, 11 km, and Mb 5.5, respectively); Source 2 occurred on March 6, 2015 (the latitude, longitude, depth, magnitude are 41.3165° S, 80.6147° E, 13.1 km, and Mb 5.5, respectively). The numbers near the red dots are for the event numbering. The seismic rays of the two events recorded by the two seismic stations all sample the velocity structure of the MTZ beneath southern Sumatra. We clearly plot the ray paths from source 1 in (b). The velocity model in (b) is from our tomographic results. Another model that does not possess a slab in the MTZ beneath southern Sumatra is constructed based on the tomographic results of Huang et al. (2016), and this model is employed for comparison. The real waveforms recorded by two stations are shown in (c), in which we also present the observed relative traveltime differences between the two stations. The calculated traveltime differences are calculated in our model and the model without a slab in the MTZ beneath southern Sumatra (Huang et al., 2016). The residuals, which are presented in (d), are obtained by subtracting the calculated traveltime difference from the observed traveltime difference. The residuals for source 1 and 2 in our model are 0.89 s and 0.81 s, respectively, while the residuals in the model without a slab in the MTZ are 1.46 s and 0.97 s, respectively.

Figure 12. Sketch depicting the subducted slab beneath Sumatra. The red triangle denotes the Toba caldera. The slab is characterized by mainly subvertical tears, and by variations in the dip angle and penetration depth. One slab tear is beneath the Toba caldera in northern Sumatra, while another is detected beneath the western Java Sea. The dip angle of the subducted slab beneath the area NW of the Toba caldera is shallower than that of the slab SE of the Toba caldera. The dip is nearly vertical beneath the southern tip of Sumatra, whereas the dip of the slab beneath western Java becomes shallow. The deepest part of the subducted slab is stagnant in the MTZ. Beneath Sumatra, the slab NW of the Toba caldera may barely arrive at the 410-km discontinuity, while the slab SE of the Toba caldera penetrates into the MTZ, and the slab arrives at the 660-km discontinuity beneath the southern tip of Sumatra.

The slab beneath southern Sumatra is not separated from that beneath western Java by a lateral gap spanning a few hundred kilometers (Figures 5g and 5h), as was suggested by the tomographic results of Huang et al. (2015), who claimed that the slab beneath the southern tip of Sumatra penetrates only to the upper boundary of the MTZ, and consequently, a slab gap appears in the MTZ. The gap discussed by Huang et al. (2015) may be due to the relatively low imaging resolution in the MTZ beneath southern Sumatra, which is at the southwest corner of a broad study area (Figure 4 in Huang et al., (2015)). More recently, nearly the same study area as that of Huang et al. (2015) was investigated by Zenonos et al. (2019), who suggested that the subducted slab beneath southern Sumatra penetrates to a depth of only about 300 km, which is much shallower than the penetration depth shown in Figure 5h. The differences between the penetration depths revealed in Figures 5g and 5h and the results of Zenonos et al. (2019) are probably caused by the different amounts of data used in the tomographic inversions. Zenonos et al. (2019) used mainly the ISC-EHB dataset, which has limited coverage in southern Sumatra and hence could resolve only the long-wavelength structures (on the scale of a few hundred kilometers) but not the detailed structures of the slab.

The penetration depth of the subducted slab in southern Sumatra is much shallower than that of the slab derived from a simultaneous inversion of earthquake hypocenters and seismic velocities (Richards et al., 2007). The slab described by Richards et al. (2007) subducts to a depth of at least 800 km and then connects to the subducted Neotethys plate beneath southern Sumatra, which contrasts considerably with our results. The global slab model TX2019slab, which was obtained from seismic tomographic inversion with a starting model incorporating a 3-D slab inferred from a theoretical global 3-D thermal model of subducting slabs (Lu et al., 2019), presents the same features as the model of Richards et al. (2007). Based on our results, we propose that the slab beneath southern Sumatra probably terminates at a depth of about 700 km (Figures 5h and 5i).

Slab detachment beneath northern Sumatra was reported in Widiyantoro and van der Hilst (1996) and Liu et al. (2018). However, no slab exists in the uppermost lower mantle or in the MTZ beneath northern Sumatra (Figures 5g–5i). Thus, our tomographic results do not support slab detachment beneath this area. Slab detachment was also not supported by a recent tomographic study of northern Sumatra using both regional and teleseismic traveltime data (Liu et al., 2019).

Based on the results of teleseismic traveltime tomography, Li et al. (2018) claimed that the subducted slab extends into the lower mantle on profiles EE' and FF' (Figure 8 in Li et al., (2018)), which is deeper than the

penetration depth of the slab in Figures 6e–6f. However, from their checkerboard resolution test (Figure 5 in Li et al. (2018)), strong vertical smearing probably resulted in artificial images within the lower mantle. Hence, it is likely that the slab on profiles DD' and EE' (Figures 6d and 6e) reaches a shallower depth than the slab imaged by Li et al. (2018) (Figure 8 in Li et al., (2018)). In addition, the results of Li et al. (2018) do not show a clear southward variation in the dip angle from profile DD' to profile GG'. Compared with Li et al. (2018), we employed more high-quality seismic data, and these data were recorded at evenly distributed seismic stations throughout southern Sumatra; hence our results are able to provide more details of the subducted slab.

Beneath western Java, the dip of the subducted slab in the upper mantle is nearly the same as the dip reported in Luehr et al. (2013) (Figure 3 in Luehr et al., (2013)), which was obtained by the inversion of revised ISC data with a regional tomographic inversion algorithm (Koulakov et al., 2009). However, on profile GG' (Figure 6h), the geometry of our subducted slab in the MTZ is quite different from that of Luehr et al. (2013). Our images indicate that the dip of the slab decreases in the MTZ and that a small segment tends to flatten above the 660-km discontinuity. In contrast, the dip of the imaged slab in Luehr et al. (2013) becomes nearly vertical in the MTZ and then descends into the uppermost lower mantle (660–900 km). The geometry of the downgoing slab on profile GG' is also different from that derived from global seismic tomography (Fukao & Obayashi, 2013). The global seismic tomographic inversion of Fukao and Obayashi (2013) revealed a slab passing nearly vertically through the MTZ and then becoming trapped in the uppermost lower mantle (660–1000 km depth). Our checkerboard test demonstrates that the resolution beneath the western Java Sea at a depth of 800 km is still high (Figure 3h), but our results do not reveal a slab signature in the uppermost lower mantle (Figure 6h).

Receiver function analysis can provide additional constraints on the penetration depth of the subducted slab (Mishra et al., 2020; Saita et al., 2002). Receiver function analysis showed that the 410-km discontinuity beneath the northern tip of Sumatra is depressed and the undulation is about 20 km (Saita et al., 2002). In addition, the thickness of the MTZ is about 40 km thinner than the global average (Figure 3 in Saita et al., 2002). The results from Saita et al. (2002) indicate that the slab beneath the northern tip of Sumatra does not penetrate across the 410-km discontinuity, which may partly support our results presented in Figure 6a. Different from the topographies of the 410-km and 660-km discontinuities beneath the northern tip of Sumatra, the 410-km and 660-km discontinuities beneath both southern Sumatra and western Java are elevated and depressed, respectively (Saita et al., 2002). The results from receiver function analysis (Saita et al., 2002) imply that the subducted slab exists in the MTZ beneath both southern Sumatra and western Java, which may confirm our tomographic results shown in Figures 6g and 6h.

5.3. Variation in Slab Dip

Beneath Sumatra, the dip of the slab north of the Toba caldera is shallow, while the dip in the south is relatively steep and gradually becomes vertical. Furthermore, a dramatic reduction in the dip occurs from the west of the Sunda Strait to the east. Because of these variations in the dip, the slab exhibits a sinusoidal shape in the upper mantle, mimicking the surface expressions of the trench and volcanic arc (Figure 1). Three main factors can account for the variations in dip along the strike.

First, the age of the subducted slab is probably the most important factor controlling the shape of the subducted slab. A compilation of the magnetic anomalies in the eastern Indian Ocean (Bird, 2003; Carton et al., 2014; Singh et al., 2011) suggests that the age of the oceanic plate near the Sumatra trench generally decreases northward from ~90 Ma to ~50 Ma. Accordingly, the subducted slab beneath northern Sumatra is relatively young and therefore more buoyant than that beneath southern Sumatra (Fauzi et al. 1996). Thus, the dip of the subducted slab in the north is less steep than that in the south. This is probably because of the age contrast in the oceanic floor. A distinct feature regarding the age of the oceanic floor is the considerable jump in age that occurs across the IFZ. The age of the oceanic floor east of the IFZ is much older than that west of the IFZ, and the age contrast reaches as high as 15 Ma (Koulakov et al., 2016). Because of this substantial age difference, the dip changes abruptly across the IFZ. Therefore, we consider that the age of the subducted slab is a first-order factor causing the sinusoidal shape of the slab.

The lateral stress within the slab is the second most important factor. The slip partitioning of interplate earthquakes has often been observed in subduction zones (Fitch, 1972), including the western Aleutian, northern New Guinea and Sumatra subduction zones. Such slip partitioning is attributed to lateral stresses within subducting slabs (Fitch, 1972; Liu et al., 1995). Because of the oblique convergence (Figure 1), the subducting slab beneath Sumatra suffers from lateral shear stress oriented NNW-SSE (Liu et al., 1995; McCaffrey et al., 2000). This lateral shear stress is equivalent to an SSE-directed pressure force acting on the downgoing slab. Combined with the age effect mentioned above, this pressure force results in seaward concave deformation in the north and landward concave deformation in the south (Figure 5). Hence, lateral shear stress deforms the subducted slab, which consequently exhibits a sinusoidal shape in map view.

Trench migration is the third controlling factor. If we consider the average convergence rate of the subducted slab to be consistently 60 mm/yr, the slab in the upper mantle was subducted within the past 15 Ma. From the reconstructed positions of the India and SE Asia blocks, the Sumatra trench has experienced seaward migration in the past 15 Ma, although the migration is not significant (Lee & Lawver, 1995; Replumaz & Tapponnier, 2003; Replumaz et al., 2004). Generally, the scale of migration increases southward (Lee & Lawver, 1995). The extrusion of Sundaland that initiated at ~50 Ma due to the collision between India and Eurasia (Replumaz et al., 2004) causes the seaward migration of the Sumatra trench. Trench migration has been included as a constraint for kinematic modeling of slab subduction in SE Asia (Hafkenscheid et al., 2001). The kinematic modeling shows a subducted slab with a sinusoidal shape although the amplitude of the sinusoid is not large (Hafkenscheid et al., 2001). Thus, trench migration is an important (but not dominant) factor that contributes to the sinusoidal geometry of the slab.

Two features regarding the variation in dip deserve further discussion. The first is that the lateral (seaward concave) deformation of the subducted slab beneath northern Sumatra (latitude > 0°) is more significant than that (landward concave) beneath southern Sumatra (latitude < 0°) (Figures 5c-5f). We attribute the stronger deformation in the north to two reasons. On the one hand, the obliquity of plate convergence in the north is larger than that in the south (Figure 1). According to the force balance equations governing the dynamics of the downgoing slab (Equation 5 in Liu et al. (1995)), larger obliquity results in greater lateral shear stresses within the slab, and a greater lateral shear stress results in more significant deformation of the subducted slab. On the other hand, the oceanic lithosphere in the north is rheologically weak. The lithosphere around the extinct Wharton Fossil Ridge (WFR) (Figure 1) is much younger and warmer than the lithosphere in the south (Shapiro et al., 2008; Singh et al., 2011). Young and warm subducting lithosphere deforms more easily than old and cold lithosphere, even under the same lateral shear stress. The second feature is the dramatic change in dip across the Sunda Strait (Figure 6). Although the age of the subducting slab beneath western Java is greater than that of the slab beneath southern Sumatra (Figure 1), the dip of the former is much shallower, which is attributed to the retreat of the Java trench. Based on laboratory models of upper mantle subduction (Schellart, 2010), trench retreat is a causing factor of the reduction in dip. According to the plate reconstruction, the retreat of the Java trench is much more significant than that of the Sumatra trench (Hafkenscheid et al., 2001; Lee & Lawver, 1995). Therefore, the dip dramatically decreases beneath western Java.

Notably, the curvature of the subducted slab beneath northern Sumatra is not as high as previously reported (Hall & Spakman, 2015; Pesicek et al., 2008, 2010a; Richards et al., 2007). As shown in Figure 5, the subducted slab in the upper mantle reveals a seaward concave shape with a low degree of curvature. The slight concave curvature of the slab at depths above 200 km has also been delineated by earthquake hypocenters (Pesicek et al., 2010b). Owing to the lack of seismicity at greater depths (>200 km), it is impossible to infer the geometry of the slab using earthquake hypocenters alone. Based on our tomographic results, we postulate that this concave shape extends from the uppermost mantle to the 410-km discontinuity. This raises the question of why a significantly curved slab is revealed by other tomographic models (Pesicek et al., 2008; Richards et al., 2007). One plausible interpretation is that the slab-related high-velocity anomalies beneath northern Sumatra are lumped together with the high-velocity anomalies beneath Peninsular Malaysia, which are revealed in both Richard et al. (2007) and Pesicek et al. (2008); as a result, the imaged slab exhibits significant curvature. The high-velocity anomalies in the upper mantle beneath Peninsular Malaysia have been repeatedly confirmed by recent regional seismic tomography studies (Huang et al., 2015; Liu et al., 2019), which provided the necessary facts to support this interpretation.

5.4. Two Subvertical Slab Tears

Figure 12 depicts two subvertical slab tears. The first is beneath northern Sumatra; and the second is beneath the Sunda Strait. Two slab tears are inferred from two different clues found in our tomographic images. The first clue is the additional northeastward displacement of the slab north of the Toba caldera at a depth of 120 km (Figure 5c), while the second clue is the remarkable reduction in slab dip beneath western Java (Figure 6h).

Many researchers have hypothesized the presence of a slab tear beneath the Toba caldera and have concluded that a slab tear is responsible for the world's largest Pleistocene eruption of the Toba caldera 74,000 years ago (Acocella et al., 2018; Fauzi et al., 1996; Hall & Spakman, 2015; Kong et al., 2020; Koulakov et al., 2016; Page et al., 1979). The slab tear beneath the Toba caldera was originally inferred by Page et al. (1979) following observations of the distribution of volcanoes in Sumatra. The Plio-Pleistocene stratovolcanoes to the south of Toba are generally situated along the main trace of the Sumatran Fault, whereas the Toba caldera and the volcanoes to the north of Toba are located to the east of the Sumatran Fault (Page et al., 1979). The volcanic arc in Sumatra generally follows the 120-km depth contour of the slab (Fauzi et al., 1996; Syracuse & Abers, 2006). At this depth, strong dehydration of the heated slab generally releases a large volume of fluids (Koulakov et al., 2016; Zheng et al., 2016). Fauzi et al. (1996) also postulated that a slab tear is possible beneath the Toba caldera based on the local hypocenters of slab-related earthquakes, although no conclusive evidence was presented in their study. In particular, at depths of 100–150 km, the seismicity to the NW of the IFZ presents a gentler downgoing trend than that to the SE (Fauzi et al., 1996). On the one hand, this vertical offset in the depth of seismicity across the IFZ may stem from the inferred subvertical tear described above (Fauzi et al., 1996), which may correspond to a deeper eastern slab segment, similar to the result shown in Figure 5c. On the other hand, this vertical offset may also originate from the bending of the slab. The strike of the subducted slab changes from NNE south of the Toba caldera to NNW north of Toba. This sharp change in the strike of the slab near the Toba caldera could also cause deeper seismicity on a vertical plane perpendicular to the IFZ (Figure 3 in Fauzi et al. (1996)). More recently, a slab tear beneath northern Sumatra was suggested by Hall and Spakman (2015), who employed the global P-wave model UU-P07 (Amaru, 2007) to investigate the mantle structure and tectonic history of SE Asia, and found significant slab folding beneath northern Sumatra in the upper mantle and MTZ, especially at depths from 350 to 500 km, resembling the model of Pesicek et al. (2008). However, instead of following Pesicek et al. (2008), who claimed that this folding originates from the deformation of the subducted WFR and the static character of the plate boundary (the diffuse plate boundary is shown in Figure 1), Hall and Spakman (2015) preferred an NNE-trending subvertical tear in the slab to interpret the slab folding revealed by tomographic images (Amaru, 2007). Furthermore, a recent regional tomography study around the Toba caldera illustrated crucial low-velocity anomalies and high VP/VS ratios in the crust and uppermost mantle (Koulakov et al., 2016). By combining the bending of the subducted slab in the upper mantle revealed by a previous regional tomography model (including Sumatra and Java) (Luehr et al., 2013) with their tomographic results, Koulakov et al. (2016) inferred a tear in the slab beneath the Toba caldera to interpret the abovementioned low-velocity anomalies and high VP/VS ratios. A recent study of seismic anisotropy also indicated the presence of a slab tear beneath Toba caldera (Kong et al., 2020). Their shear wave splitting measurement reflect both lateral and vertical distributions of anisotropy beneath the Toba caldera and surrounding areas (Kong et al., 2020). These anisotropy patterns are attributed to mantle flow that enters the mantle wedge through the slab window resulting from the slab tear. According to the study of Kennett and Furumura (2010), the gap beneath the Toba caldera (Figure 5c) may be due to either slab thinning or slab bucking. However, the integration of the evidence mentioned above (tomographic images, seismicity, seismic anisotropy and volcano locations) probably indicates the presence of slab tear.

The existence of a slab tear beneath the Sunda Strait is somewhat tentative compared with that beneath the Toba caldera. This slab tear is partly supported by the gap in subduction-related earthquakes (Malod et al., 1995). In addition, P-wave anisotropy revealed that the direction of fast P-wave velocity has a component perpendicular to the trench at a depth of 450 km (Huang et al., 2015). This anisotropic feature may reflect mantle flow passing through the slab window beneath the Sunda Strait and provides an additional clue for the existence of a slab tear. Moreover, the deepest earthquake shown in Figure 6h (latitude: 4.2640° S; longitude: 106.8422° E; depth: 645 km; magnitude: 4.8 Mb) is situated beneath the fastest velocity anom-

ally and may indicate the presence of a slab tear, similar to the 2015 Bonin earthquake ($M_w \sim 7.9$) (Zhang et al., 2019).

Assuming that both slab tears revealed by our tomographic images are real, we discuss the reasons for these slab tears below. The difference in slab age across the IFZ is probably the primary reason for the first slab tear. As shown in Figure 1, the age of the slab west of the IFZ is much younger than that in the east, and as a result, the dip of the subducted slab is shallower in the west. Therefore, in map view, the slab north of the Toba caldera exhibits additional northeastward displacement (Figure 5c), which produces a subvertical tear in the slab. The large age contrast across the IFZ has also been interpreted as the origin for the slab tear (Koulakov et al., 2016). Two other factors may act as secondary reasons for the first slab tear. On the one hand, the IFZ, which is rich in fractures and faults (Koulakov et al., 2016; Singh et al., 2011), serves as a zone of weakness within which a slab tear could easily form. On the other hand, the serpentinization of the uppermost mantle beneath the IFZ reduces the rigidity of the downgoing plate, which facilitates the generation of a slab tear. Serpentinization of the oceanic crust and uppermost mantle of the downgoing plate has been adopted to explain the low S-wave velocity anomalies beneath the offshore region in northern Sumatra (Harmon et al., 2012). By analyzing the relationship between the S-wave velocity and the percentage of serpentinization (from dunite to lizardite serpentinization at 1 GPa) (Christensen, 2004), Harmon et al. (2012) estimated 14–24% serpentinization of the oceanic crust and uppermost mantle. Such a high degree of serpentinization would weaken the downgoing slab, potentially inducing the occurrence of a slab tear in response to the difference in buoyancy across the IFZ. However, as shown in Figure 12, the orientation of the slab tear is approximately SW–NE, which is not consistent with the SSW–NNE orientation of the IFZ determined based on seafloor bathymetry (Figure 1) (Koulakov et al., 2016). We presume that the lateral stress on the subducted slab (Liu et al., 1995) gradually changes the oblique angle of convergence towards a normal angle as the depth increases. Therefore, the orientation of the IFZ in the subducted slab changes to SW–NE and is no longer the same as the orientation of the IFZ in the Indian Ocean. It is worthy to note that the orientation of the slab tear in this study follows that of Hall and Spakman (2015). Moreover, the lateral shear in the slab is probably not strong enough and may only partly contribute to the orientation of the slab tear. Nonetheless, the orientation and its mechanism require further investigations.

Trench migration is the main factor for the second slab tear. As discussed in Section 5.3, the large amount of trench migration has significantly reduced the slab dip beneath western Java (Lee & Lawver, 1995). The dramatic change in slab dip across the Sunda Strait may have resulted in the second slab tear (Figure 12). Moreover, the variation in the plate convergence angle may be another factor. The Sunda Strait serves as a transition zone between two different convergence systems (Anshori et al., 2017). The convergence angle east of the Sunda Strait is almost perpendicular to the Java trench, whereas that to the west is oblique to the Sumatra trench (Figure 1). This transition in plate convergence is also reflected by the differences in tectonic developments. Typical forearc basins are developed to the east of the Sunda Strait, whereas dextral strike-slip faults oriented NW–SE are developed to the NW of the Sunda Strait (Malod et al., 1995). This transition in plate convergence further generates extensional stresses in the Sunda Strait. The extensional regime in the strait was previously confirmed by a microearthquake survey by analyzing individual and composite focal mechanisms (Harjono et al., 1991). Therefore, we consider that this extensional stress contributed to the second slab tear.

5.5. Variation in Penetration Depth

Figure 10 shows that the slab beneath northern Sumatra penetrates to a shallower depth than the slab in the south; the former penetrates to an average depth of about 410 km, while the latter penetrates to an average depth of about 660 km. The abovementioned changes in both the convergence rate and the obliquity of convergence from north to south may have produced this variation in penetration depth. On the one hand, the convergence rate through the Sumatra trench decreases northward from 7.2 to 5.0 cm/yr (Figure 1) (Sieh & Natawidjaja, 2000), which may be due to the decreased northward motion of the Indian plate. The motion of the Indian plate has slowed since ~ 50 Ma following the collision between the Indian plate and the stable Eurasian plate (Replumaz et al., 2010). This collision also decreased the northward motion of the Indo-Australian plate (Capitanio, 2020). On the other hand, the obliquity of plate convergence increases northward (Acocella et al., 2018), which implies that the convergence component normal to the trench

decreases northward. Furthermore, the slab dip is relatively shallow in the north (Figures 6 and 10), which also partly contributes to the shallower penetration depth compared with the slab in the south. If changes in the convergence rate and obliquity of subduction are both taken into account, the length of the subducted slab north of the equator is about 250 km shorter than that south of the equator (if we consider that the rapid subduction began in the early Miocene as mentioned in the Introduction), which correlates well with our tomographic results.

The shallow penetration depth beneath northern Sumatra is similar to that beneath the Andaman Sea (Mishra et al., 2020). Mishra et al. (2020) performed receiver function analysis and common conversion point (CCP) stacking and discovered that the subducting slab beneath the Andaman Islands penetrates across the 410-km discontinuity but does not extend to the 660-km discontinuity. A similar feature of the subducting slab beneath the Andaman Sea is also revealed by seismic tomography (Pesicek et al., 2008). The common features between the slabs beneath northern Sumatra and the Andaman Sea may indicate that the slab subduction process in northern Sumatra is continued northward to the region of the Andaman Sea.

6. Conclusions and Final Remarks

We conduct regional and teleseismic traveltime tomography to investigate the slab geometry beneath Sumatra. High-resolution images of the subducted slab are obtained based on the high-quality seismic data provided by BMKG and a robust eikonal equation-based seismic tomography technique. Our tomographic images show that the subducted slab in the upper mantle parallels the volcanic arc and exhibits a sinusoidal shape. In addition to that, two slab tears are revealed by our tomographic results, both of which have been previously inferred. One is beneath the Toba caldera, and the other is beneath the Sunda Strait. From our tomographic inversion, the slab in the north penetrates to a shallower depth than that in the south. The slab geometry revealed by our results may add additional constraints on plate reconstructions of SE Asia.

In our previous tomographic study of northern Sumatra (Liu et al., 2019), because of the relatively low quality of the S-wave data, the discussion was based primarily on the P-wave tomographic results (Liu et al., 2019). In this study, we do not conduct S-wave traveltime tomography. Nevertheless, S-wave tomography is important and may provide crucial details regarding the geometry of the subducted slab (Koulakov et al., 2009, 2016). For instance, ambient noise tomography (Yang et al., 2011) may be an efficient technique for revealing the S-wave velocity structure beneath Sumatra; using wave equation-based adjoint tomography to invert high-frequency waveform data (Chen et al., 2015; Tao et al., 2018; Tong et al., 2014) could also be promising in deriving high-resolution images of the S-wave velocity structure beneath Sumatra.

Some issues remain unresolved in this study. We consider that the high-velocity anomalies beneath Peninsular Malaysia are lumped together with the slab-related high-velocity anomalies beneath northern Sumatra. However, the detailed structures of the high-velocity anomalies beneath Peninsular Malaysia are still unknown. Because only four broadband seismic stations are available in Peninsular Malaysia (one in Singapore and three in Malaysia) (Figure 1), it is impossible to resolve the fine details of the high-velocity anomalies. Therefore, future investigations of the velocity structure may require the deployment of broadband seismic stations throughout Peninsular Malaysia. As discussed in Section 5, subduction in northern Sumatra may be closely linked to subduction in the Andaman Sea. However, high-resolution images of the slab beneath the Andaman Sea are absent because few seismic stations are deployed in this region. Because of the highly active seismicity in the Andaman Sea (Pesicek et al., 2010b), in the future, seismic data from regional earthquakes in the Andaman Sea recorded at teleseismic distances can be used for tomographic studies. In this way, the slab morphology beneath Sumatra and the Andaman Sea can be revealed simultaneously.

Data Availability Statement

The picked traveltimes of the regional events, the cross-correlation traveltime differentials of the teleseismic events, the station information, and the source locations can be downloaded from <https://blogs.ntu.edu.sg/geophysics/files/2020/05/Sumatra-data.zip>.

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