

Article

Mantle Transition Zone Structure Beneath the Sunda–Java Trench and Christmas Island: Implications for Deep Dehydration and Petit-Spot Volcanism

Ba Manh Le ^{1,*}, Ting Yang ^{2,3} , Thi-Giang Ha ¹, Van-Duong Nguyen ¹ , Thanh Hai Nguyen ¹, Van Vung Vi ¹, Thuy Linh Nguyen ¹, Van Duan Bui ¹, Minh Tuan Vu ¹, The Truyen Pham ¹ and Anh Duong Nguyen ¹

¹ Institute of Earth Sciences (IES), Vietnam Academy of Science and Technology (VAST), Hanoi 11512, Vietnam; gianght@ies.vast.vn (T.-G.H.); linhnt.91@ies.vast.vn (T.L.N.); tuannvm@ies.vast.vn (M.T.V.)

² OBS Laboratory, Department of Ocean Science and Engineering, Southern University of Science and Technology, Shenzhen 518055, China; tyang@sustech.edu.cn

³ Shanghai Sheshan National Geophysical Observatory, Shanghai 201602, China

* Correspondence: manhle@ies.vast.vn; Tel.: +84-967-577-583

Abstract

Petit-spot volcanism is a type of intraplate volcanism that commonly occurs near the outer rise in subducting plates. Although deformation of the incoming plate is considered important for its development, the deeper mantle processes contributing to this volcanism remain poorly understood. Here, we use more than 1500 broadband seismograms containing long-period SS precursors to investigate the mantle transition zone (MTZ) beneath the Sunda–Java subduction system. Beneath the northern Sunda–Java region, the 660 km discontinuity (d660) reaches ~690 km depth, resulting in an MTZ thicker than 270 km, consistent with the thermal influence of the subducted Indo-Australian slab. The associated weakening of S660S amplitudes may reflect hydration or other factors affecting the lower MTZ. South of the trench, the MTZ beneath most of the Christmas Island Seamount Province is relatively uniform, with a thickness of ~251 km. Beneath Christmas Island, however, the 410 km discontinuity (d410) is depressed to ~420–425 km, whereas d660 remains near 670–675 km, producing a localized thinner MTZ of ~245–250 km. This structure is compatible with localized slab-related mantle processes and may reflect thermal or compositional heterogeneity near the top of the MTZ. Although our observations do not directly resolve volatile transport or melt generation, they provide new constraints on the deep-mantle environment associated with rejuvenated petit-spot volcanism at Christmas Island.

Keywords: mantle transition zone; SS precursors; deep dehydration; petit-spot volcanism; slab-mantle interaction; Christmas Island



Academic Editor: Gianluca Groppelli

Received: 26 May 2026

Revised: 22 July 2026

Accepted: 24 July 2026

Published: 26 July 2026

Copyright: © 2026 by the authors.

Licensee MDPI, Basel, Switzerland.

This article is an open access article distributed under the terms and conditions of the [Creative Commons Attribution \(CC BY\)](https://creativecommons.org/licenses/by/4.0/) license.

1. Introduction

Most global volcanism can be broadly classified into three major types according to their dominant melting mechanisms: plume-related melting that produces age-progressive oceanic islands [1]; decompression melting at mid-ocean ridges [2]; and flux melting at subduction zones [3]. Petit-spot volcanism represents a distinctive form of intraplate volcanism that occurs on the outer rise in subducting plates. Unlike volcanism associated with mantle plumes, mid-ocean ridges, or volcanic arcs, petit-spot volcanoes are characterized by their small size, geologically young eruption ages, and occurrence far from established magma-generation environments [4].

Since their first discovery on the subducting Pacific Plate near the Japan Trench [4], petit-spot volcanoes have been recognized in several subduction systems worldwide, including Tonga, Chile, Mariana, Sunda, and Puerto Rico trenches [5–10]. Their magmas are typically generated by extremely low-degree partial melting and are enriched in volatile components such as H₂O and CO₂ [4,11–16], making them particularly valuable for investigating mantle metasomatism, the deep volatile cycle, and the origin of recycled mantle materials.

Several mechanisms have been proposed to explain the origin of petit-spot volcanism. The most widely accepted model attributes petit-spot volcanism to lithospheric bending prior to subduction. Lithospheric bending generates extensional stress in the outer rise zone. These extensional stresses may induce low-degree decompression melting near the lithosphere–asthenosphere boundary (LAB), and the resulting melts ascend through fractures generated by lithospheric bending [4,12]. Another model proposes that partially molten material is already present beneath the LAB. Lateral pressure variations drive the movement and accumulation of molten material. After accumulation, the molten material undergoes uplift due to buoyancy through the lithosphere, forming petit-spot volcanoes [17]. In contrast, Yang and Faccenda (2020) [18] proposed that hydrous upwelling from the mantle transition zone (MTZ), induced by slab rollback and MTZ uplift, may supply volatile-rich melts to the shallow mantle, providing a deeper geodynamic origin for petit-spot volcanism. Unlike the lithosphere-controlled models, this hypothesis links petit-spot volcanism to deep mantle dynamics and the deep volatile cycle [18]. Assessing the potential role of the MTZ in petit-spot volcanism therefore requires reliable constraints on MTZ structure.

The MTZ, bounded by global seismic discontinuities at average depths of approximately 410 km (d410) and 660 km (d660), is a key interface linking the upper and lower mantle [19–21]. These discontinuities correspond to pressure-induced phase transitions of olivine, from α -olivine to wadsleyite at the d410 [22] and from ringwoodite to bridgmanite plus ferropericlase at the d660 [23]. Because the d410 has a positive Clapeyron slope, whereas the d660 has a negative Clapeyron slope, their topography is highly sensitive to mantle temperature [24,25]. Cold subducting slabs typically uplift the d410 and depress the d660, producing a thicker MTZ, whereas hot mantle upwellings generate the opposite pattern [25,26].

In addition to temperature, water enrichment (which can reach up to ~1.5 wt% in the upper MTZ) also plays an important role in shaping these boundaries [27,28]. Water can affect MTZ discontinuity topography in a similar way to cold thermal anomaly, uplifting the d410 and depressing the d660 [29,30], and can also reduce the sharpness of the d410 [31]. Consequently, the geometry and seismic properties of the d410 and d660 provide valuable constraints on slab dynamics, mantle circulation, and deep volatile storage, making the MTZ an important target for evaluating whether deep mantle processes may be related to the formation of petit-spot volcanism [20,27].

The Sunda–Java subduction system and the rejuvenated volcanism of Christmas Island (CI) provide an exceptional natural laboratory for evaluating whether MTZ processes are linked to the formation of petit-spot volcanism (Figure 1). In this region, the Indo-Australian Plate subducts beneath the Eurasian (or Sunda) Plate along the Sunda–Java Trench [32–34]. The incoming oceanic lithosphere varies in age from ~40 Ma beneath southern Sumatra to >130 Ma beneath eastern Java and converges at rates of ~60–70 mm yr^{−1} [35]. South of the trench lies the Christmas Island Seamount Province (CHRISP), a large volcanic province (~1,000,000 km²) composed of seamounts, guyots, and volcanic islands [36]. Volcanism within CHRISP spans more than 100 Myr, including an Eocene shield-building phase and younger Pliocene volcanism on CI [10,16,36,37]. Geochemical studies indicate that the

younger volcanism contains signatures of recycled and volatile-rich mantle sources. The combination of these geochemical characteristics and the complex tectonic setting of the region make the CHRISP an important natural laboratory for investigating the potential links between deep mantle processes and petit-spot volcanism.

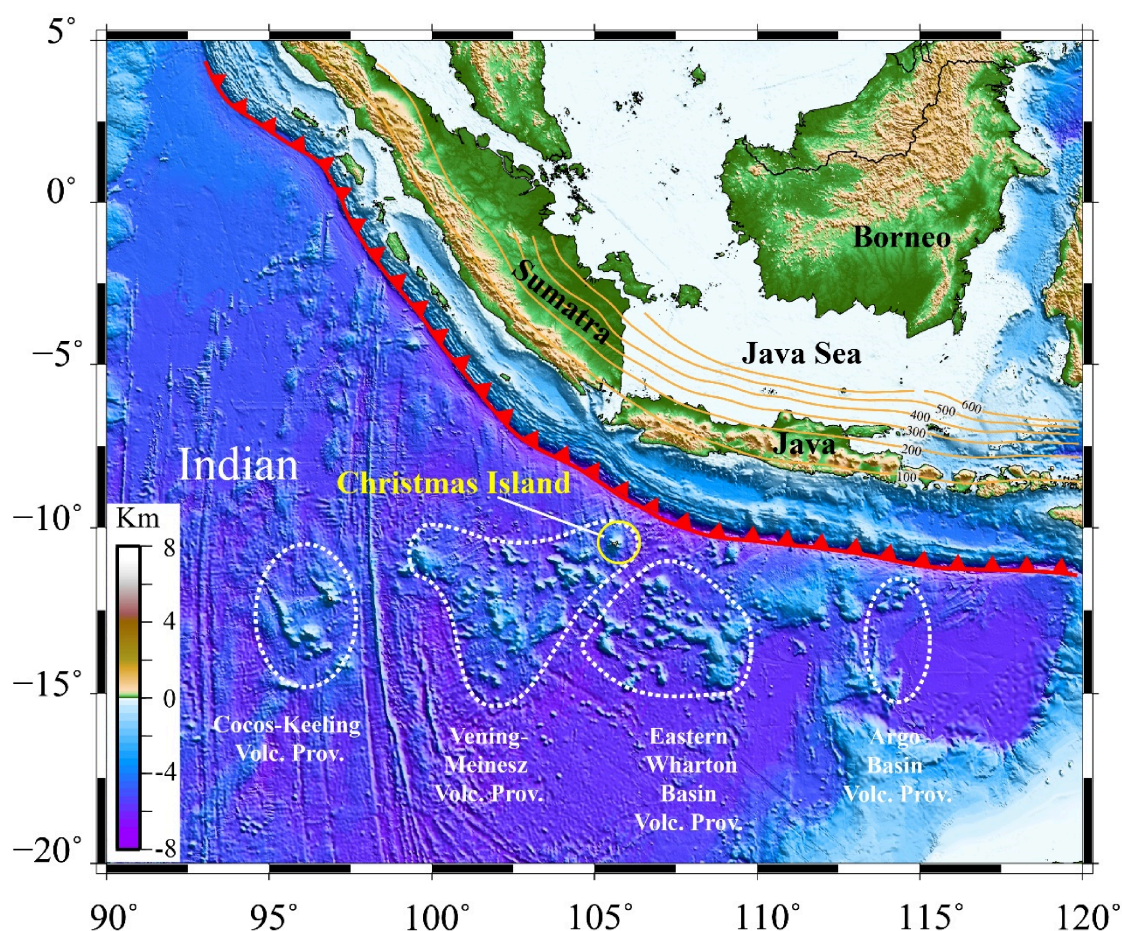


Figure 1. Tectonic setting and regional features of the study area. Background colors denote the regional topography and bathymetry derived from GEBCO 2025 Grid [38], with the color scale provided in kilometers (km). The Sunda–Java Trench, indicated by the red line with triangles, defines the convergent boundary where the Indo–Australian plate subducts beneath the Eurasian plate. Solid orange contours extending across Sumatra, the Java Sea, and Java indicate the depth to the subducting slab in kilometers [39]. Dashed white contours delineate the volcanic sub-provinces within the Christmas Island Seamount Province (CHRISP) and adjacent oceanic regions, including the Cocos–Keeling, Vening–Meinesz, Eastern Wharton Basin, and Argo Basin volcanic provinces [36]. Yellow circle denoted Christmas Island.

Previous tomographic studies have provided important insights into the geometry of the Sunda slab and large-scale mantle heterogeneity beneath Southeast Asia [40–43]. Receiver-function investigations have further constrained MTZ discontinuities, but their observations are largely limited to continental regions and a few island stations because of the requirement for dense seismic station coverage [44,45]. As a result, the MTZ beneath the adjacent oceanic region, including the CHRISP, remains insufficiently characterized.

An alternative approach is provided by long-period SS precursors (S410S and S660S), which are underside reflections from the 410 and 660 km discontinuities and offer an effective means of imaging MTZ structure [20,26,46,47]. Because these phases are primarily sensitive to discontinuity topography near their surface bounce points, located approximately halfway between the earthquake source and the seismic receiver, their spatial

coverage is not constrained by the distribution of local seismic stations [48,49]. This unique ray geometry makes SS precursors particularly well suited for imaging MTZ structure beneath oceanic regions where permanent seismic networks are sparse. Although SS precursor studies have been conducted across Southeast Asia, they have mainly focused on subduction margins and the East Vietnam Sea [50,51]. As a result, the MTZ beneath CHRISP has not yet been investigated using this approach.

In this study, we investigate the MTZ beneath the Sunda–Java subduction system and the adjacent CI region using a high-quality dataset of over 1500 broadband seismograms containing long-period SS precursors that sample the study region. After applying three-dimensional tomographic corrections, we map the topography and reflection amplitudes of the 410 and 660 km discontinuities beneath the study region. We use these observations to characterize MTZ structure beneath the Sunda–Java subduction system and the adjacent Christmas Island Seamount Province and to evaluate the potential implications of the observed MTZ structure for the geodynamic setting of Christmas Island petit-spot volcanism.

2. Materials and Methods

2.1. Data Selection and Processing

In this study, we use seismic records containing SS precursor reflections and only include seismograms with SS bounce points located within the study area. Only seismograms with SS bounce points located within the study area were included in the analysis. We selected earthquakes with moment magnitudes (M_w) greater than 5.5 and focal depths shallower than 75 km [20]. Events within this magnitude range generally produce coherent SS phases with detectable precursor energy, whereas the shallow-depth criterion minimizes interference from depth phases such as sSdiff and sS410S [49]. Source–receiver distances were restricted to 100 – 165° to reduce contamination from topside reflections, including Ss410s and Ss660s, as well as precursor energy associated with ScSScS phases [52,53].

Waveform processing followed standard procedures commonly used in SS precursor studies [54]. Instrument responses were first removed from all seismograms, and the horizontal components were rotated into the great-circle reference frame to isolate the SH component for subsequent analysis. The SH waveforms were then bandpass filtered between 20 and 50 s using a second-order zero-phase Butterworth filter to suppress background noise and enhance weak precursor arrivals. To ensure data quality, seismograms with signal-to-noise ratios (SNRs) lower than 3 were excluded. The SNR was calculated as the ratio between the maximum amplitude within the SS time window and the maximum amplitude within the precursor time window. The SS window was defined as 50 s before and after the predicted SS arrival, whereas the precursor window extended from 30 s before the predicted S660S arrival to 30 s after the predicted S410S arrival based on the ak135 model [51].

Each waveform was subsequently visually inspected to identify high-quality records. For consistent processing, the first major peak of the SS arrival in each trace was aligned to zero time, and waveform amplitudes were normalized by the maximum amplitude of the trace. This normalization reduces variations associated with earthquake source effects and radiation-pattern differences, allowing precursor amplitudes to be compared more consistently across the dataset. After screening, the final dataset consisted of more than 1500 transverse-component (SH) seismograms suitable for SS precursor analysis. The global distribution of earthquakes and seismic stations, together with the corresponding SS precursor ray paths and bounce-point coverage, is shown in Figure 2.

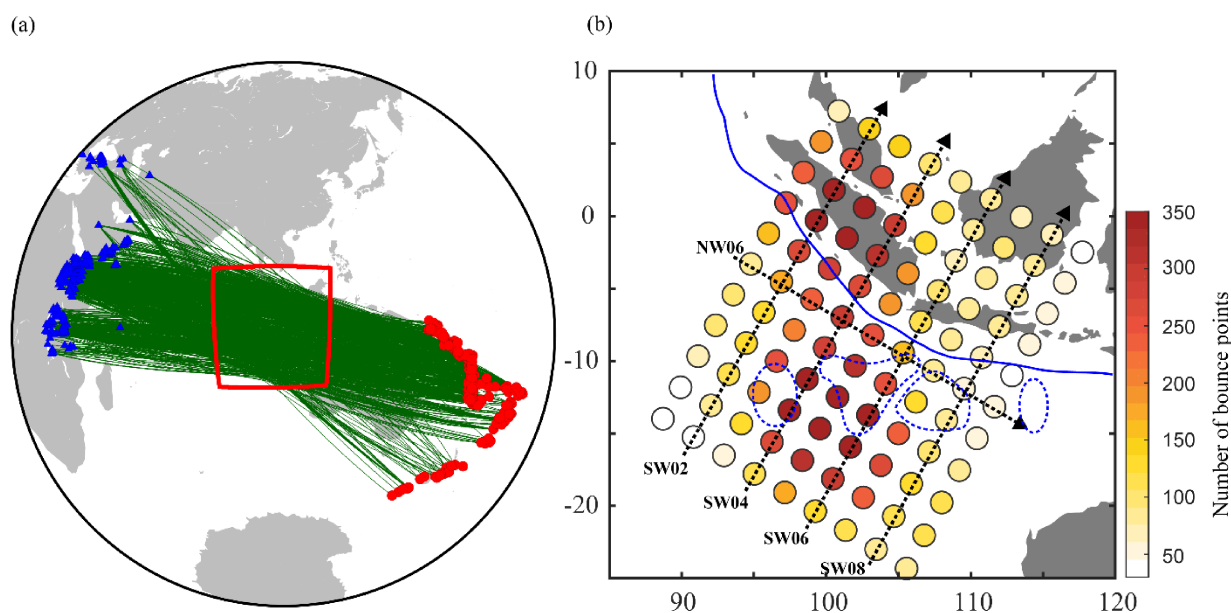


Figure 2. Dataset coverage and spatial binning. (a) Earthquakes (red dots), seismic stations (blue triangles), and SS precursor ray paths (green lines). The study area is boxed in red. (b) Spatial distribution of stacking bins (radius 5° , spacing 2.5°). Bin colors denote the data density (the number of bounce points). Blue solid and dashed lines indicate the trench and the CHRISP region. Black dashed lines show the locations of the cross-sectional profiles analyzed in later figures.

Representative SH-component seismograms are presented in Figure 3. In several records, the S410S and S660S precursor arrivals can be identified prior to stacking. However, because SS precursor phases are typically weak and commonly exhibit amplitudes less than 10% of the main SS phase amplitude, waveform stacking was required to improve signal coherency and enhance precursor detectability [55]. We adopted the linear stacking approach widely used in regional and global SS precursor studies [20]. To implement the stacking procedure, the study area was divided into regularly distributed spatial bins with a radius of 5° , while adjacent bin centers were spaced at intervals of 2.5° (Figure 2) [51,56]. This spatial binning provides sufficient waveform coverage within individual bins to produce stable stacked traces while simultaneously averaging small-scale discontinuity roughness and reducing the influence of scattered seismic energy [20].

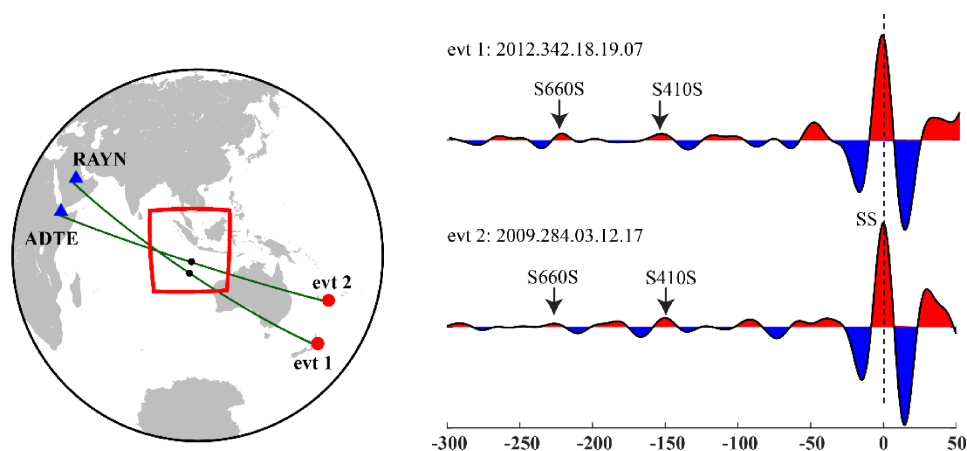


Figure 3. Seismic waveform examples and ray geometry. (Left) Source–receiver distribution showing earthquakes (red dots), stations (blue triangles), and SS ray paths (green lines) with bounce points (black dots) inside the study area (red rectangle). (Right) Two examples of observed SH-component seismograms. Red and blue shading denote positive and negative polarities. Waveforms are aligned

on the main SS arrival (0 s, vertical dashed line). Black arrows indicate the arrivals of S410S and S660S. Horizontal axis shows time relative to the SS phase (in seconds).

2.2. Travel-Time Corrections

The differential travel times between the SS phase and its associated precursors, S410S and S660S, are influenced not only by discontinuity topography but also by three-dimensional (3-D) seismic velocity heterogeneity along the propagation paths [55]. To isolate travel-time variations associated primarily with MTZ structure, several correction procedures were applied prior to stacking. The spatial distribution and magnitude of these corrections are summarized in Figure 4 and Figure S1.

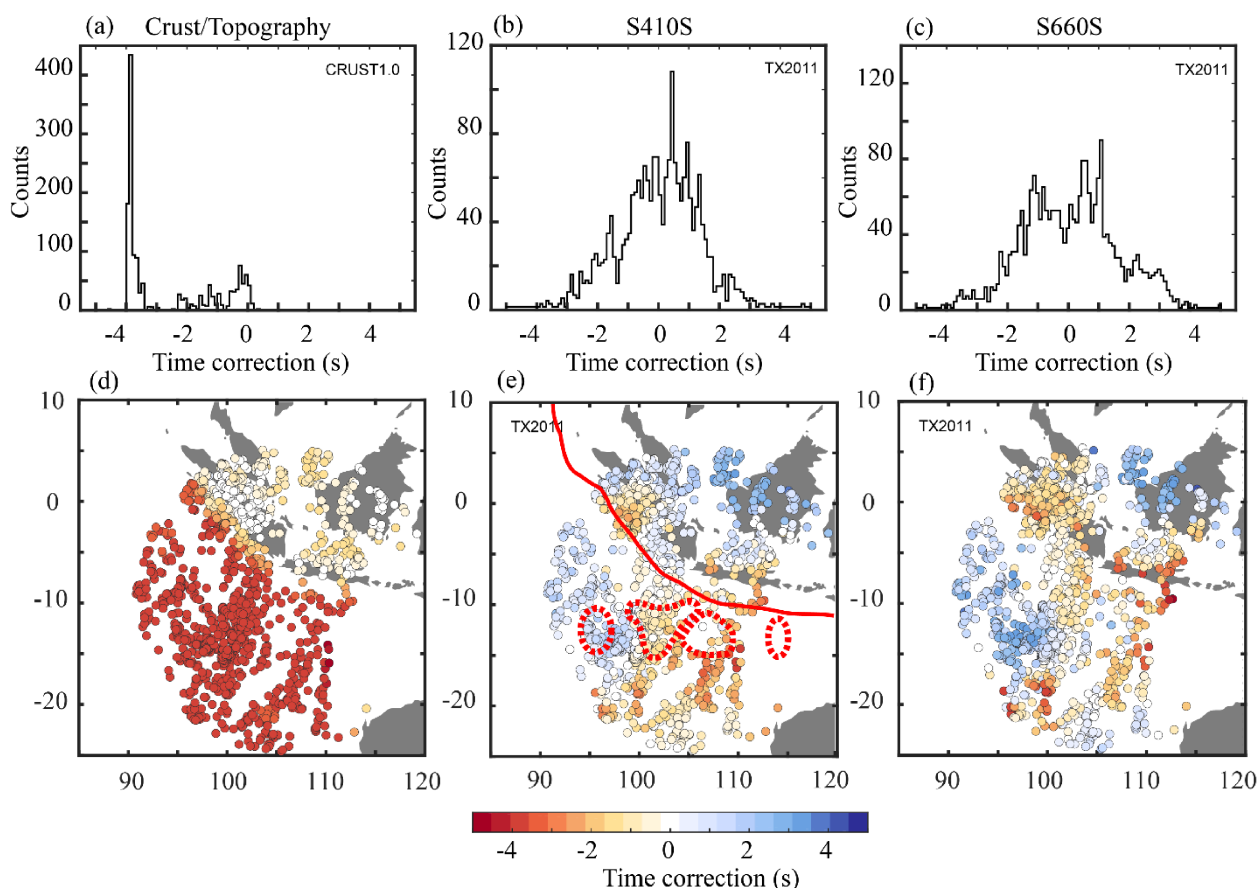


Figure 4. Travel-time corrections. Histograms (**top row**) and spatial distributions (**bottom row**) of the applied time shifts. (**a,d**) Crustal and topographic corrections applied to SS phases using CRUST1.0. (**b,c**) Histograms of the tomographic time corrections for S410S and S660S, respectively, calculated from the TX2011 global velocity model. (**e,f**) Corresponding spatial distributions of the S410S and S660S tomographic corrections, respectively. The color bar represents the time correction in seconds for panels (**d–f**). Red solid and dashed lines in (**e**) indicate the trench and the CHRISP region.

The first correction accounts for near-surface structural effects. Because crustal structure varies substantially between continental and oceanic regions, and because the SS phase reflects at the Earth's surface, SS travel times are particularly sensitive to crustal heterogeneity near the bounce points. Crustal travel-time corrections were calculated using the CRUST1.0 model relative to the 1-D ak135 reference model [19,57]. Surface elevations were additionally corrected to sea level using the ETOPO1 global topography model [58]. The resulting crustal corrections are shown in Figure 4a,b. The largest negative corrections, reaching approximately -4 s, are concentrated in the Indian Ocean region, whereas

corrections north of the subduction zone are generally smaller, mostly ranging between -1 and 1 s.

Because seismic waves propagate through a laterally heterogeneous mantle, velocity anomalies along the ray paths can introduce travel-time perturbations unrelated to discontinuity topography. Tomographic travel-time corrections were applied following previous SS precursor studies [51,55]. Theoretical ray paths for the SS, S410S, and S660S phases were calculated using the ak135 reference Earth model [19], and travel-time perturbations along these paths were estimated using the global tomographic models TX2011 [59].

The tomographic correction was calculated as:

$$\Delta t_{corr} = (T_{SS}^{1D} - T_{SdS}^{1D}) - (T_{SS}^{3D} - T_{SdS}^{3D}) \quad (1)$$

where T_{SS} and T_{SdS} are the travel times of the SS phase and its precursor phase (S410S or S660S), respectively. This correction compensates for mantle-induced travel-time perturbations along the propagation paths. The spatial distribution and magnitude of the tomographic corrections are shown in Figure 4e–f.

Seismograms, which are grouped within each spatial bin, span a broad range of epicentral distances. The relative travel times between the SS phase and its precursors vary systematically with distance. This effect causes phase misalignment during stacking and reduces the coherency of the weak precursor signals. To minimize this effect, move-out corrections were applied following previous SS precursor studies [50–52]. All waveforms were adjusted to a common reference epicentral distance of 130° using theoretical differential travel times predicted by the ak135 model [19]. For each record, the predicted differential travel time at the observed epicentral distance was subtracted from that at the reference distance, and the waveform was shifted accordingly. This correction improves the alignment and coherency of the S410S and S660S precursor phases prior to stacking. After applying all corrections, the seismograms within each spatial bin were combined using linear stacking. Representative stacked waveforms are shown in Figure 5 and demonstrate substantial enhancement of the S410S and S660S precursor phases relative to noise.

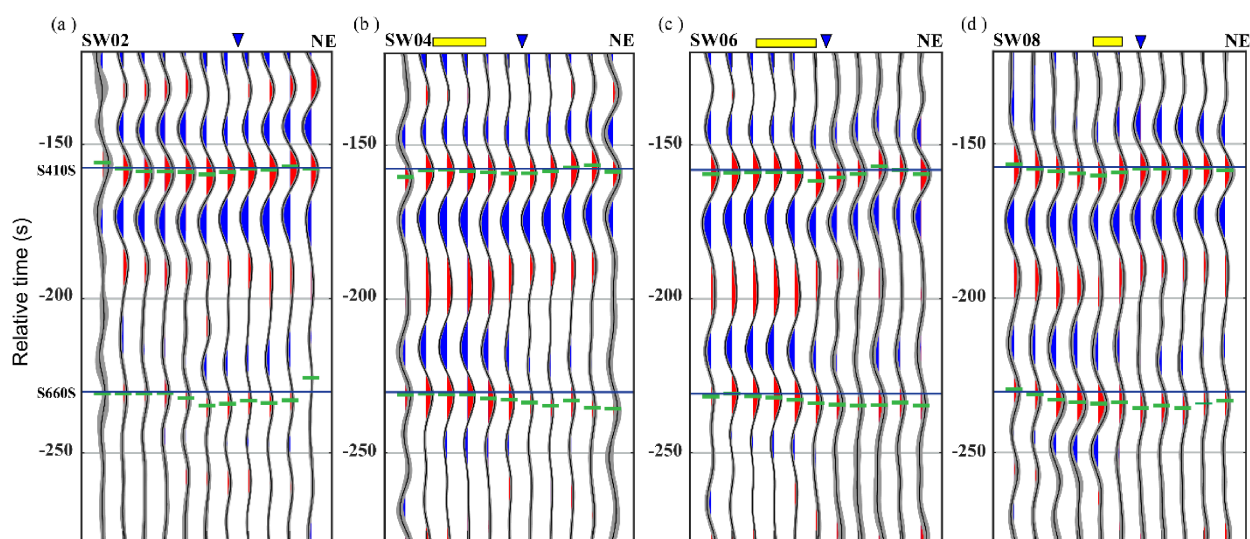


Figure 5. Stacked seismograms for bins along four profiles. (a) Profile SW02. (b) Profile SW04. (c) Profile SW06. (d) Profile SW08. Gray envelopes represent the bootstrap-derived uncertainties. Red and blue shading denote positive and negative polarities. Blue solid lines indicate the differential times ($t_{S410S}-t_{SS}$ and $t_{S660S}-t_{SS}$) predicted by ak135 model, while green markers show the actual peak times picked in this study. Blue triangles and yellow rectangles indicate the trench and volcanic regions, respectively. Profile locations are shown in Figure 2.

3. Results

To improve the robustness of the stacked SS precursor signals and quantify associated uncertainties, we applied a bootstrap resampling procedure following previous SS precursor studies [20,60]. For each spatial bin, the selected seismograms were resampled with replacement and stacked 200 times. The amplitudes and relative travel times of the S410S and S660S phases were measured independently for each bootstrap realization. Final amplitudes and travel times were calculated as the mean values of all realizations, whereas uncertainties were estimated using the corresponding 95% confidence intervals. Furthermore, to ensure that our observations reflect MTZ of study area rather than artifacts of a specific velocity model, we tested several global 3-D tomographic models, including TX2011 [59], HMSL-S06 [61], S40RTS [62], and SEMUCB-WM1 [63]. Although travel time corrections using 3D tomography models differ significantly between different velocity models (Figure S1), the resulting spatial patterns of relative travel times, the depths of the d410 and d660 discontinuities, and the MTZ thickness are highly consistent across the models, even though their absolute values slightly vary (Figures S2 and S3). Our SS precursor imaging reveals lateral variations across the study area. To describe these variations, the study region is divided into three tectonic domains (Figure 6b): the northern region (Sunda plate), the Sunda–Java subduction zone, and the southern incoming oceanic plate.

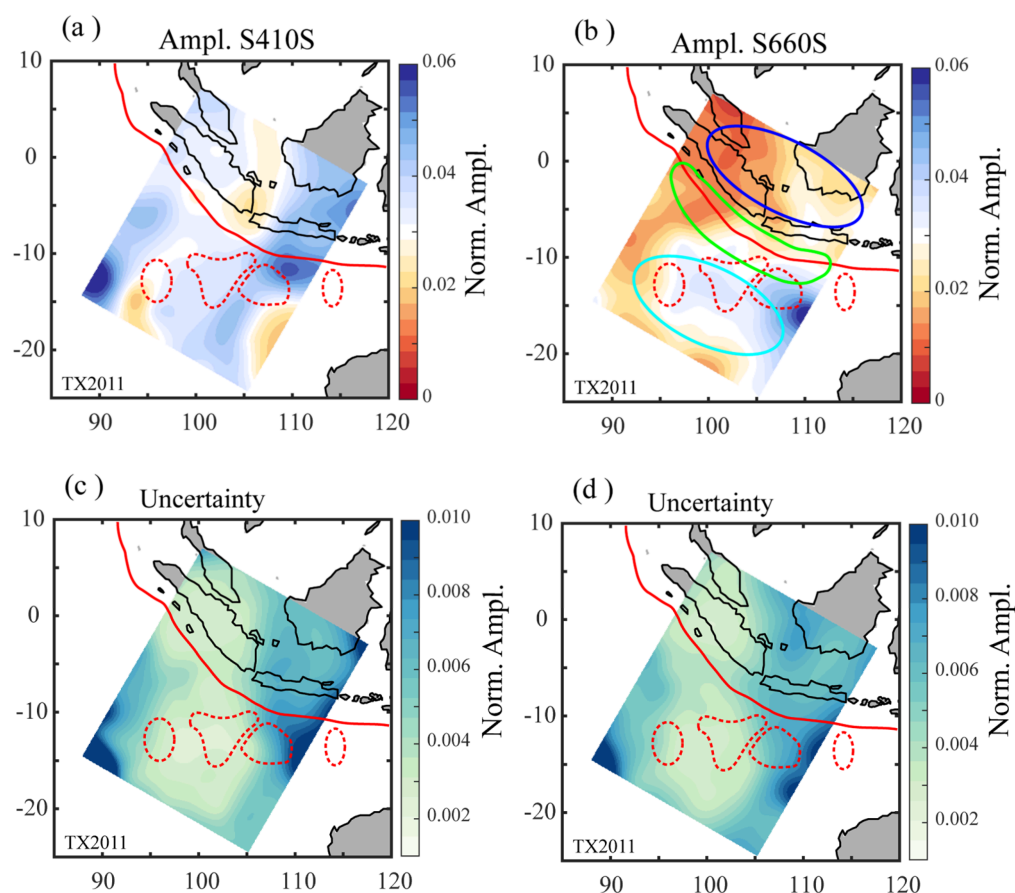


Figure 6. Normalized reflection amplitudes. Spatial distribution of the interpolated, normalized reflection amplitudes for the (a) S410S and (b) S660S phases, and (c,d) their corresponding bootstrap standard deviation, using the TX2011 tomographic model. Red solid and dashed lines represent the Sunda–Java Trench and the CHRISP volcanic region, respectively. Colored ovals in panel (b) outline the three tectonic regions discussed in the text: the northern region (blue), the subduction zone (green), and the southern oceanic plate (cyan).

3.1. Amplitude Characteristics of the SS Precursor Phases

Figure 6 shows the spatial distribution of normalized reflection amplitudes for the S410S and S660S precursor phases together with their bootstrap-derived uncertainty. Uncertainties are generally low throughout the study area for both phases. The S410S amplitudes are relatively homogeneous, whereas the S660S amplitudes display stronger lateral variations. The S410S amplitudes range from 0.023 to 0.055, with an average value of approximately 0.037 (Figure 6a). Spatial variations are limited across most of the study area.

The S660S amplitudes show larger spatial variations (Figure 6b). Lower amplitudes are observed in the northern and northwestern parts of the study area near the subduction boundary, where values decrease to approximately 0.01. Toward the southern oceanic region, amplitudes gradually increase and reach values of approximately 0.035 within the CHRISP region. The north–south amplitude contrast is also visible in the stacked waveform profiles for SW04, SW06 and SW08 in Figure 5.

3.2. Relative Travel Times of the SS Precursor Phases

Figure 7 and Figure S2 show the spatial distribution of differential travel-time residuals for $t_{S410S} - t_{SS}$, $t_{S660S} - t_{SS}$, and $t_{S660S} - t_{S410S}$ using TX2011 and other tomographic corrections. All correction models produce similar large-scale spatial patterns. Uncertainties are generally larger in areas with sparse data coverage, especially near the edges of the study region. However, relatively large uncertainties are also observed in several small areas in the northern region despite the higher data density.

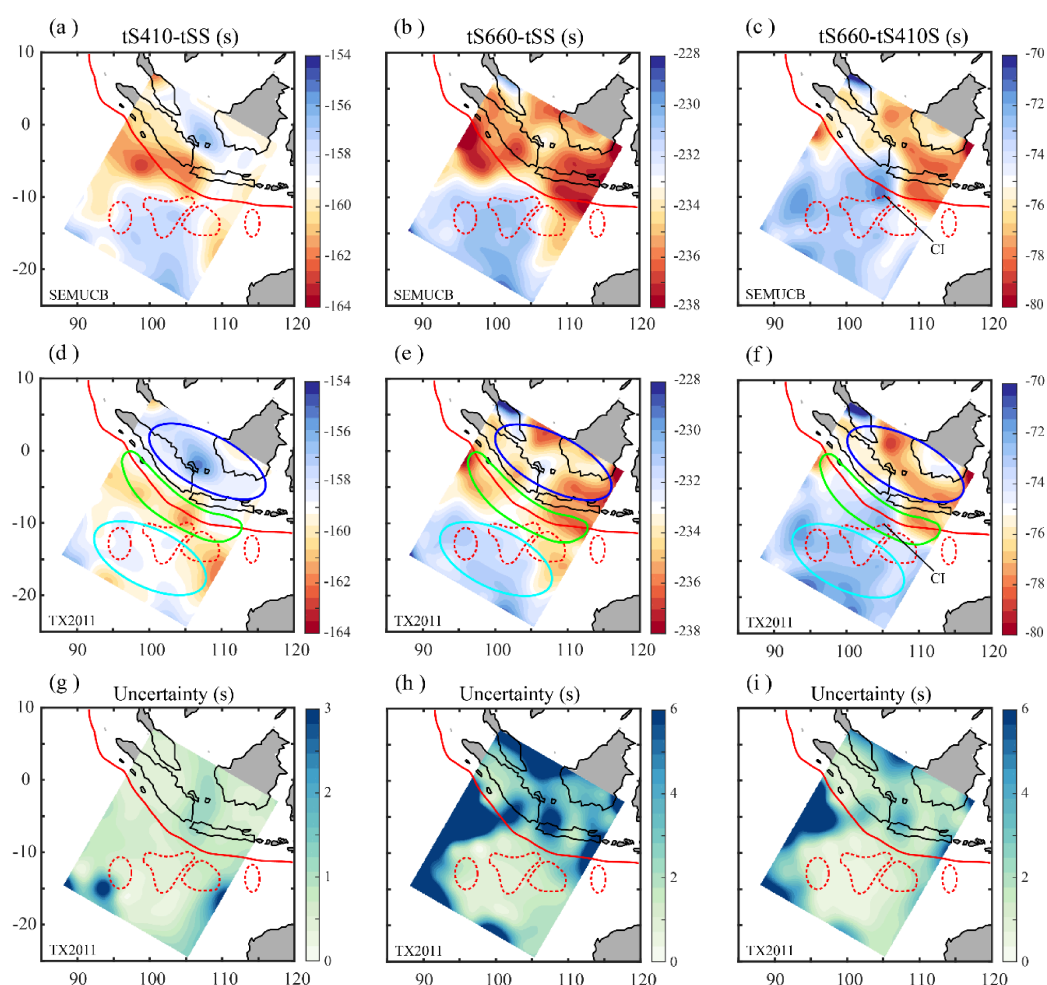


Figure 7. Differential travel times and uncertainties. Maps of differential travel times for (a,d) $t_{S410S} - t_{SS}$, (b,e) $t_{S660S} - t_{SS}$, and (c,f) $t_{S660S} - t_{S410S}$, corrected using the SEMUCB-WM1 [63]

(top row) and TX2011 [59] (middle row) models. Bottom panels (g–i) show the corresponding bootstrap uncertainty based on the TX2011 model. Red solid and dashed lines represent the Sunda–Java Trench and the CHRISP volcanic region, respectively, with “CI” denoting Christmas Island (f). Colored ovals in (d–f) highlight the three tectonic domains discussed in the text: the northern region (blue), the subduction zone (green), and the southern oceanic plate (cyan).

For tS410S – tSS (Figure 7a,d), the most negative residuals occur along the subduction zone, reaching -161 ± 0.8 s relative to the ak135 reference value of -156.7 s. The residuals become less negative toward the northern part, reaching -155 ± 1.5 s, whereas values in the southern oceanic regions approach -157 ± 0.5 s.

The tS660S – tSS residuals (Figure 7b,e) show stronger spatial variations. The most negative values are found along the subduction zone and nearby northern regions, where residuals reach -236 ± 4 s relative to the ak135 reference value of -229.7 s. Toward the incoming oceanic plate, the residuals gradually increase to -232 ± 1 s within the CHRISP region. In certain areas, even with relative dense ray coverage (Figure 2), the associated uncertainties can still reach up to 6 s.

Similar spatial variations are observed for tS660S – tS410S (Figure 7c,f). More negative values occur along the trench and northern regions, reaching -78 s relative to the ak135 reference value of -72.9 s. In the southern region, the differential times become less negative and approach -72 s.

3.3. Discontinuities Depth and MTZ Thickness

The differential travel times tS410S – tSS and tS660S – tSS were converted into depth perturbations relative to the ak135 reference discontinuities using a travel time–depth relationship [21]:

$$\delta h = \frac{\delta t}{\eta} \quad (2)$$

where δt is the differential travel-time residual relative to the reference model and η represents the vertical slowness that governs the sensitivity of the differential travel time to discontinuity depth:

$$\eta = 2\sqrt{V_s^{-2} - \left(\frac{p}{r}\right)^2} \quad (3)$$

Here, V_s is the shear-wave velocity from the ak135 reference model, p is the ray parameter of the SS precursor phase for an epicentral distance of 130° , and r is the radius of the corresponding discontinuity. These calculated perturbations were then utilized to determine the absolute depths of the d410 and d660, with the MTZ thickness defined as d660–d410 (Figure 8 and Figure S3).

The d410 shows relatively small lateral variations across the study area. Depths are around 420 ± 3 km along the Sunda–Java subduction zone. Toward the back-arc region, the discontinuity becomes shallower and reaches 405 ± 4 km west of Borneo. Beneath the incoming Indo-Australian plate, d410 is generally close to 415 ± 2 km, whereas deeper values (~ 425 km) occur beneath the CI region.

In contrast, the d660 exhibits much stronger lateral variations. Depths reach nearly 690 ± 10 km beneath the subduction zone and nearby northern regions, including parts of the back-arc area. In the southern part, the discontinuity becomes shallower and approaches 665 ± 3 km beneath the incoming oceanic plate.

MTZ thickness also varies across the study area. The thickest MTZ occurs along the subduction zone and adjacent northern regions, where values locally exceed 270 km. Between Sumatra and Java, thickness decreases to ~ 255 km. In the southern oceanic and CHRISP regions, MTZ thickness ranges from 245 to 259 km, with an average of ~ 251 km.

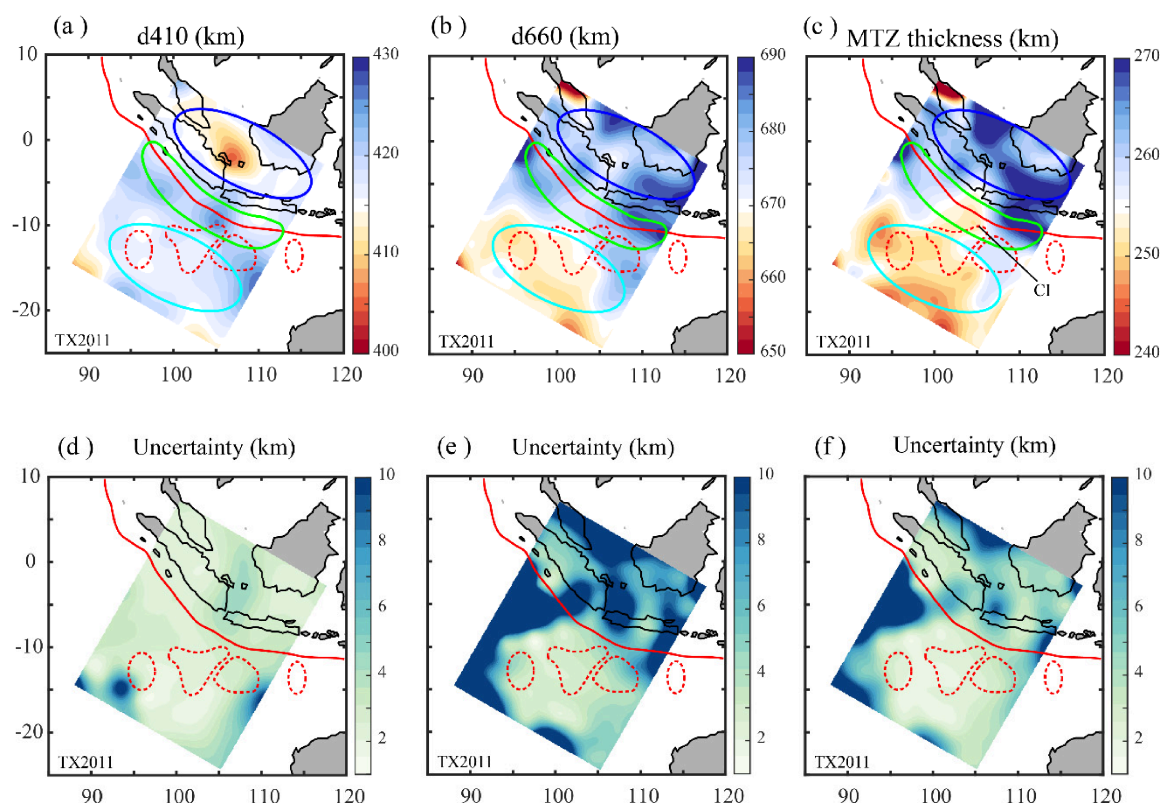


Figure 8. Discontinuity topography and MTZ thickness. Topography of the (a) 410 km and (b) 660 km discontinuities, and (c) MTZ thickness, with their corresponding bootstrap-derived uncertainties (d–f) using the TX2011 tomographic model. The solid red line indicates the Sunda–Java Trench. Colored ovals represent the three tectonic domains discussed in the text: the northern region (blue), the subduction zone (green), and the southern incoming oceanic plate (cyan). Red dashed contours outline the CHRISP, with “CI” denoting Christmas Island (d).

4. Discussion

Previous global SS precursor studies have mainly resolved broad, long-wavelength variations in MTZ structure because of their relatively large stacking caps [20,52]. Although depth migration can improve the imaging of discontinuities with strong lateral topographic variations, previous work has shown that the resulting MTZ structures may not differ substantially from those obtained using linear stacking [50]. In this study, the combination of a regionally focused dataset and three-dimensional tomographic corrections provides improved regional constraints on MTZ topography beneath the Sunda–Java subduction system and the adjacent oceanic region (Figure 8). The similar relative travel-time patterns obtained using different 3-D tomographic models (Figure 7 and Figure S1) further indicate that the main structural features identified here are not strongly dependent on the choice of tomographic model. These characteristics provide a basis for comparing our results with previous observations and for examining regional-scale MTZ variations across the study area.

4.1. Comparison with Previous Studies

Our results broadly agree with previous seismic investigations in identifying a thickened MTZ beneath the Sunda–Java subduction system, while providing more detailed constraints on its lateral variability. The correspondence between the mapped discontinuities and independent shear-wave velocity anomalies (Figure 9) provides additional support for the overall interpretation. However, the comparison with previous studies

also reveals important differences in the spatial expression of MTZ structure across the three tectonic domains identified in Figure 8.

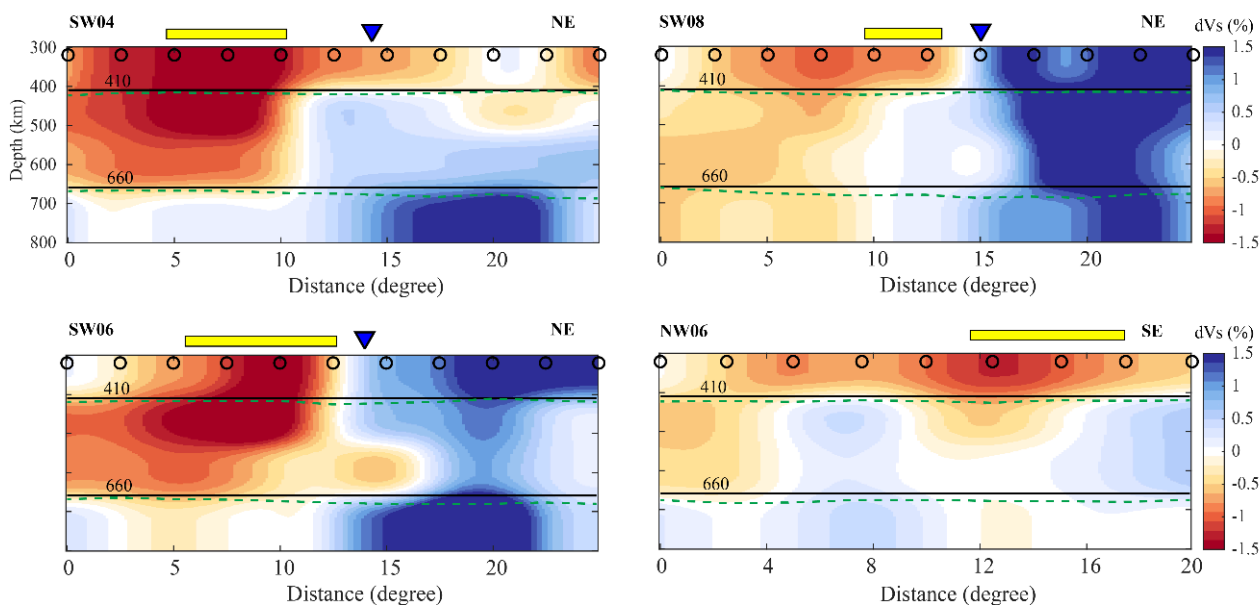


Figure 9. Cross-sectional variations in MTZ topography and seismic velocity. Background contours display shear-wave velocity perturbations (dVs) from HMSL-S06 tomography model [61]. Solid black lines mark global reference depths (410 and 660 km), while dashed green lines represent the absolute discontinuity depths inferred in this study. Top markers denote the Sunda–Java Trench (blue triangles), volcanic provinces (yellow bars), and spatial stacking bin centers (open circles). Profile locations are shown in Figure 2.

Within the northern region, the MTZ is generally thickened, with local thicknesses exceeding 270 km (Figure 8c). This thickening results from the combined effect of an uplifted d410, reaching ~405 km in some areas (Figure 8a), and a depressed d660, locally approaching ~690 km (Figure 8b). The thicknesses observed beneath southern Sumatra and western Java are comparable to those reported by high-resolution receiver-function studies, which measured MTZ thicknesses of up to 295 km and ~269 km, respectively [44,45]. Our observation of an uplifted d410 beneath western Borneo is also consistent with the regional SS precursor study of Dokht et al. (2018) [50]. At a broader scale, the thickened MTZ is consistent with the global SS precursor model of Dai et al. (2026) [53], which identified positive MTZ thickness perturbations beneath many subduction systems. Thus, the pronounced MTZ thickening in the northern region represents a feature consistently observed by different seismic methods and datasets. Our results further show that this broad thickening is spatially heterogeneous rather than uniform across the entire northern region.

South of the trench, our expanded offshore coverage provides constraints on the incoming Indo-Australian plate across the CHRISP. In contrast to the thickened MTZ beneath the northern subduction-related region, most of the CHRISP is characterized by a relatively uniform MTZ thickness of ~251 km (Figure 8c). This value is close to the average thickness predicted by the ak135 reference Earth model [19], suggesting that the incoming oceanic plate does not exhibit a pronounced regional-scale MTZ thickness anomaly before subduction. This result agrees with Dokht et al. (2018) [50], who identified no significant MTZ thickness anomaly south of the trench, and is also consistent with the recent global model of Dai et al. (2026) [53], which reported small average MTZ thickness perturbations across broad oceanic regions. Therefore, our results support a relatively laterally coherent MTZ structure beneath most of the incoming Indo-Australian plate within the spatial scale resolved by our data.

The most distinct difference from this relatively uniform background occurs beneath CI. Along the adjacent Sunda–Java subduction zone, the d660 is depressed to ~690 km and the MTZ is locally thicker than 270 km. Beneath CI, however, the d660 remains relatively stable at ~670–675 km, while the d410 is depressed to ~420–425 km. The resulting MTZ thickness of ~245–250 km represents a localized relative thinning compared with the adjacent subduction-related region. This sharp structural contrast was not clearly resolved in previous regional or global SS precursor models [20,50,53], likely because of differences in spatial coverage, data density, and spatial averaging.

The d410 depression beneath CI is particularly noteworthy because the thermal effect of a cold slab would generally be expected to uplift, rather than depress, this discontinuity [64]. The observed structure therefore cannot be readily attributed to slab cooling alone and suggests that additional processes may affect the upper MTZ beneath CI.

The thermal and compositional implications of the observed discontinuity topography and amplitude variations are examined in Section 4.2, while the possible significance of the CI anomaly for the origin of rejuvenated volcanism is discussed in Section 4.3.

4.2. Thermal Effects and Hydration Processes Within the MTZ

The topography of the 410 and 660 km discontinuities provides constraints on the thermal and compositional structure of the MTZ, whereas SS precursor amplitudes provide complementary information on the strength and sharpness of these seismic boundaries. Hydration can modify the pressure and width of mineral phase transitions and thereby affect the associated seismic impedance contrasts and reflection amplitudes [25,31]. The joint analysis of discontinuity topography and SS precursor amplitudes therefore provides additional constraints on the physical conditions within the MTZ, although amplitude variations may also be affected by discontinuity roughness, lateral heterogeneity, and stacking coherence.

Beneath the Sunda–Java subduction system, the pronounced depression of the d660, which locally reaches ~690 km (Figure 8b), spatially coincides with the high-velocity anomaly interpreted as the subducting Indo-Australian slab (Figure 9). This geometry is consistent with the expected effect of the negative Clapeyron slope of the ringwoodite-to-bridgmanite transition on a cold thermal anomaly [24,65]. In contrast, the d410 shows more modest regional depth variations and does not display a broad uplift comparable to that expected from an extensive cold anomaly (Figure 8a). This subdued topographic expression may partly reflect the steep geometry of the subducting slab beneath Java [32,34,39]. Because long-period SS precursors sample the structure over broad Fresnel zones, a narrow thermal anomaly associated with a steeply dipping slab may be spatially averaged, reducing its expression in the mapped d410 topography [26,50].

The amplitude observations provide complementary evidence for additional heterogeneity within the MTZ (Figure 6). S410S amplitudes remain relatively strong across much of the study region, whereas S660S amplitudes are substantially weaker beneath the subduction zone and adjacent northern region. Strong S410S reflections do not require extensive hydration of the upper MTZ, because water can broaden the olivine–wadsleyite transition and modify the associated seismic velocity contrast [31,66]. In contrast, the combination of a depressed d660 and reduced S660S amplitudes is compatible with hydration near the base of the MTZ. This interpretation is also supported by the stacked seismograms from the northern region, where the S660S arrivals are substantially weaker than the corresponding S410S phases (Figure 5). The subducted Indo-Australian slab may transport water into the MTZ, where dehydration during thermal equilibration or the breakdown of hydrous phases could release water into the surrounding mantle [44]. Experimental and mineral-physics studies have shown that water can shift the relevant phase boundaries, including

depressing the 660 km discontinuity under hydrous conditions [30]. Thus, the observed d660 depression and weak S660S may reflect the combined effects of the cold slab and possible hydration near the base of the MTZ. However, these observations do not uniquely constrain water content, because amplitude reductions may also result from discontinuity roughness, lateral heterogeneity, or reduced stacking coherence.

The S660S amplitudes recovered in this study are generally lower than those reported by the previous regional 3-D migration study of Dokht et al. (2018) [50]. This difference may reflect variations in data coverage, ray-path distribution, and processing procedures. Although our denser dataset improves spatial sampling, it may also increase sensitivity to short-wavelength variations and local roughness along the d660, which can reduce stacking coherence and lower recovered amplitudes. This effect may also contribute to the relatively large travel-time uncertainties (up to 6 s) observed in some regions, including areas with relatively dense data coverage (Figure 7). The dominant raypaths in our dataset are oriented approximately east–west, but the mapped discontinuity topography does not exhibit a systematic elongation in this direction. This lack of an obvious directional imprint suggests that a strong sampling-related artifact is not apparent in the mapped structures, although its contribution to the recovered amplitudes cannot be completely excluded.

Overall, the depressed d660 and reduced S660S amplitudes beneath the Sunda–Java subduction system are consistent with the combined influence of the cold Indo-Australian slab and possible hydration near the base of the MTZ. The amplitude observations therefore provide supporting, rather than definitive, evidence for slab-related hydration.

4.3. Implications for the Origin of Petit-Spot Volcanism at Christmas Island

The MTZ structure imaged in this study provides new geophysical constraints on the origin of the rejuvenated volcanism at CI. Previous petrological and geochemical studies have suggested that the Pliocene volcanic rocks were derived from a volatile-rich and recycled mantle source [10,16,36]. However, the deeper mantle processes that may have contributed to the distribution of these components and their eventual involvement in melting remain uncertain. Our observations of the 410 and 660 km discontinuities therefore provide an opportunity to assess whether the MTZ structure beneath CI is compatible with geodynamic processes proposed for petit-spot volcanism.

Several mechanisms have been proposed for petit-spot volcanism, including lithospheric flexure and faulting [4,12], the ascent of melts stored beneath the lithosphere [9,17], and hydrous upwelling associated with slab-induced small-scale mantle convection [18]. Our observations do not distinguish uniquely between these mechanisms. Instead, they provide constraints on the deeper mantle environment in which such processes may operate.

A key observation beneath CI is the absence of a pronounced d660 uplift. The d660 remains relatively stable at ~670–675 km, in contrast to the marked depression observed beneath the adjacent Sunda–Java subduction zone (Figure 8b). Given the negative Clapeyron slope of the ringwoodite-to-bridgmanite transition, a broad thermal upwelling extending through the MTZ would generally be expected to produce an uplift of the d660. Although a narrow or weak thermal upwelling cannot be excluded, our results provide no evidence for a broad-scale thermal anomaly extending through the MTZ beneath CI and therefore do not require a classical deep-mantle plume as the primary driver of the rejuvenated volcanism.

In contrast, the localized depression of the d410 to ~420–425 km beneath the CI region indicates a localized perturbation near the top of the MTZ (Figure 8a). This anomaly may reflect thermal and/or compositional heterogeneity [24,30], although their relative contributions cannot be distinguished from the available seismic observations. One possible mechanism is that the stagnant Indo-Australian slab modifies the surrounding mantle flow and generates localized return flow or small-scale convection [18,67]. Such circulation

could potentially transport material from the deeper MTZ toward the shallow mantle beneath the incoming plate. If hydrous material is involved, transport across the d410 could promote dehydration, potentially increasing H_2O contents in the upper mantle and lowering the peridotite solidus to facilitate partial melting [27,68]. This interpretation is broadly compatible with geochemical evidence for volatile-rich and recycled mantle components in the source of the CI volcanism [10,24,36,37].

The observed d410 depression should be regarded as a geophysical constraint compatible with, but not demonstrating, upward transport of hydrous material. Our seismic observations do not directly image volatile transport, mantle melting, or pathways connecting the MTZ to individual volcanic centers. Moreover, the broad Fresnel zones of long-period SS precursors and the laterally averaged structures obtained from bootstrap stacking within 5° spatial bins limit the resolution of narrow melt pathways and short-wavelength heterogeneity [20,26].

Within these constraints, we propose the conceptual model shown in Figure 10. The Indo-Australian slab subducts beneath the Sunda–Java margin and stagnates within the MTZ, producing the thickened MTZ and depressed d660 observed beneath the subduction zone. Slab-induced small-scale mantle convection may generate localized flow and upwelling beneath the incoming oceanic plate, potentially contributing to the d410 depression beneath the CI region. If hydrous material is involved, its upward transport across the d410 could release volatiles into the shallow upper mantle and promote partial melting [27,68]. The resulting melts may accumulate near the lithosphere–asthenosphere boundary [69] and subsequently ascend through extensional bending faults generated by outer-rise flexure [4,12]. In this framework, the MTZ provides a possible deep source or pathway for volatile-bearing material, whereas lithospheric flexure and faulting provide a shallow mechanism for melt ascent. Although this model remains conceptual, it provides a geophysically consistent framework linking the observed MTZ structure with existing petrological, geochemical, and geodynamic constraints on rejuvenated volcanism at CI.

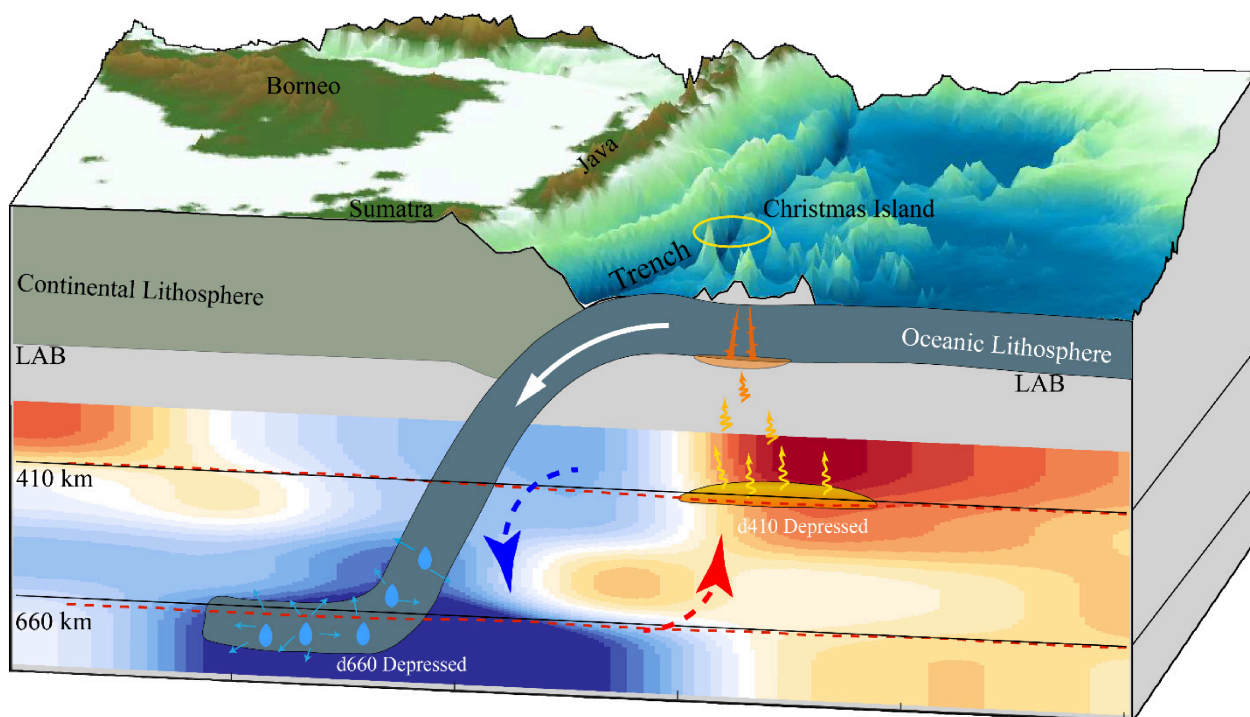


Figure 10. Schematic geodynamic model for the origin of petit-spot volcanism at CI. The Indo-Australian slab stagnates near the bottom of the MTZ, producing a depressed d660 associated with the cold slab anomaly. Proposed slab-induced small-scale convection (dashed blue and red arrows) transports hydrous

MTZ material upward across the 410 km discontinuity (d410), where volatile release and partial melting may occur beneath CI (yellow and orange wavy arrows). The resulting melts accumulate near the lithosphere–asthenosphere boundary (LAB) and ascend through extensional bending faults related to outer-rise flexure near the Sunda–Java trench. The yellow circle marks the location of CI. The tomography model in the subsurface (HMSL-S06) [61] is used to show areas of relatively fast (blue) and slow (red) shear-wave velocity. Dashed red lines indicate the observation of d410 and d660 depths along profile.

5. Conclusions

Our results reveal pronounced lateral variations in MTZ structure beneath the Sunda–Java region, with the strongest contrast occurring between the northern subduction zone and the incoming Indo–Australian plate to the south. In contrast, the MTZ beneath CHRISP is relatively homogeneous, with a thickness close to that predicted by the ak135 reference model. The northern subduction zone is characterized by a thicker MTZ, a deeper d660, and weak S660S amplitudes. This combination is consistent with the combined effects of the cold subducting slab and possible hydration-related compositional heterogeneity within the MTZ.

Beneath CI, the MTZ differs from both the adjacent subduction zone and the surrounding incoming plate. A localized depression of the d410 produces a relatively thin MTZ, whereas the d660 remains close to its regional depth. The absence of a pronounced d660 anomaly beneath CI does not require a large-scale thermal anomaly extending throughout the MTZ, although localized upwelling cannot be excluded. This structure may reflect slab-related mantle circulation that could contribute to the redistribution of volatile-bearing material and shallow-mantle melting. Extensional bending faults associated with outer-rise deformation may provide pathways for subsequent melt ascent. Although our seismic observations do not directly resolve volatile transport or melting processes, they provide a regional framework for evaluating the relationship among MTZ structure, subduction-related mantle dynamics, and petit-spot volcanism.

Supplementary Materials: The following supporting information can be downloaded at: <https://www.mdpi.com/article/10.3390/geosciences16080297/s1>, Figure S1: Tomographic travel-time corrections from alternative global velocity models; Figure S2: Comparison of differential travel times across various tomographic models; Figure S3: Comparison of MTZ discontinuity topography and thickness across different velocity models.

Author Contributions: Conceptualization: B.M.L. and T.Y.; methodology: B.M.L., T.Y. and A.D.N.; software: B.M.L.; validation: T.-G.H., V.-D.N., A.D.N. and T.H.N.; formal analysis: B.M.L. and A.D.N.; investigation: B.M.L., V.V.V. and M.T.V.; resources: V.-D.N., T.T.P. and T.-G.H.; data curation: B.M.L., T.L.N. and V.D.B.; writing—original draft preparation: B.M.L.; writing—review and editing: B.M.L., T.Y., T.T.P., A.D.N., T.H.N., M.T.V., T.L.N., V.D.B., V.V.V., T.-G.H. and V.-D.N.; visualization: B.M.L. and A.D.N.; supervision: B.M.L.; project administration: B.M.L. and T.Y.; funding acquisition: B.M.L. All authors have read and agreed to the published version of the manuscript.

Funding: This research was funded by the Vietnam Academy of Sci. and Tech. (VAST) under Grant THETN.06/24-25; VAST06.03/25-26, and NVKN13.01/26-27; the National-level Independent Project, grant number DTDL.CN-04/24; the National Key Research and Development Program of China, grant number 2024YFF0808201; and the High-Level Special Fund of SUSTech, grant number G03050K001.

Data Availability Statement: The waveform data utilized in this study were obtained from Incorporated Research Institutions for Seismology (IRIS; <https://www.iris.edu>). The global earthquake catalog used was downloaded from the International Seismological Centre (ISC; <http://www.isc.ac.uk>).

Acknowledgments: We sincerely thank our colleagues at the Institute of Earth Sciences, VAST and OBS Laboratory, SUSTech for their constructive discussions and continuous support during this study.

We sincerely thank the two anonymous reviewers for their careful evaluation of this manuscript and for their constructive and insightful comments. The Figures in this paper were created using Generic Mapping Tools (GMT 6).

Conflicts of Interest: The authors declare no conflicts of interest. The funders had no role in the design of the study; in the collection, analyses, or interpretation of data; in the writing of the manuscript; or in the decision to publish the results.

References

1. Morgan, W.J. Convection Plumes in the Lower Mantle. *Nature* **1971**, *230*, 42–43. [\[CrossRef\]](#)
2. Klein, E.M.; Langmuir, C.H. Global Correlations of Ocean Ridge Basalt Chemistry with Axial Depth and Crustal Thickness. *J. Geophys. Res. Solid Earth* **1987**, *92*, 8089–8115. [\[CrossRef\]](#)
3. Stern, R.J. Subduction Zones. *Rev. Geophys.* **2002**, *40*, 3–38. [\[CrossRef\]](#)
4. Hirano, N.; Takahashi, E.; Yamamoto, J.; Abe, N.; Ingle, S.P.; Kaneoka, I.; Ogawa, Y. Volcanism in Response to Plate Flexure. *Science* **2006**, *313*, 1426–1428. [\[CrossRef\]](#) [\[PubMed\]](#)
5. Hirano, N.; Machida, S.; Sumino, H.; Shimizu, K.; Tamura, A.; Morishita, T.; Iwano, H.; Sakata, S.; Ishii, T.; Arai, S.; et al. Petit-Spot Volcanoes on the Oldest Portion of the Pacific Plate. *Deep. Res. Part I Oceanogr. Res. Pap.* **2019**, *154*, 103142. [\[CrossRef\]](#)
6. Hirano, N.; Machida, S.; Abe, N.; Morishita, T.; Tamura, A.; Arai, S. Petit-Spot Lava Fields off the Central Chile Trench Induced by Plate Flexure. *Geochem. J.* **2013**, *47*, 249–257. [\[CrossRef\]](#)
7. Hirano, N.; Koppers, A.A.P.; Takahashi, A.; Fujiwara, T.; Nakanishi, M. Seamounts, Knolls and Petit-Spot Monogenetic Volcanoes on the Subducting Pacific Plate. *Basin Res.* **2008**, *20*, 543–553. [\[CrossRef\]](#)
8. Harmon, N.; Danforth, W.; Ten Brink, U.; Canales, J.P.; Han, S.; Vanacore, E.; Granja-Bruña, J.-L.; Moser, L.; Sokolkova, E. Evidence for Petit-Spot Volcanism in the Puerto Rico Trench. *Geophys. Res. Lett.* **2025**, *52*, e2024GL114362. [\[CrossRef\]](#)
9. Peretyazhko, I.S.; Savina, E.A.; Pulyaeva, I.A. Miocene Petit-Spot Basanitic Volcanoes on Cretaceous Alba Guyot (Magellan Seamount Trail, Pacific Ocean). *Geosciences* **2024**, *14*, 252. [\[CrossRef\]](#)
10. Taneja, R.; O'Neill, C.; Lackie, M.; Rushmer, T.; Schmidt, P.; Jourdan, F. ⁴⁰Ar/³⁹Ar Geochronology and the Paleoposition of Christmas Island (Australia), Northeast Indian Ocean. *Gondwana Res.* **2015**, *28*, 391–406. [\[CrossRef\]](#)
11. Hirano, N. Petit-Spot Volcanism: A New Type of Volcanic Zone Discovered near a Trench. *Geochem. J.* **2011**, *45*, 157–167. [\[CrossRef\]](#)
12. Valentine, G.A.; Hirano, N. Mechanisms of Low-Flux Intraplate Volcanic Fields—Basin and Range (North America) and Northwest Pacific Ocean. *Geology* **2010**, *38*, 55–58. [\[CrossRef\]](#)
13. Okumura, S.; Hirano, N. Carbon Dioxide Emission to Earth's Surface by Deep-Sea Volcanism. *Geology* **2013**, *41*, 1167–1170. [\[CrossRef\]](#)
14. Pilet, S.; Abe, N.; Rochat, L.; Kaczmarek, M.-A.; Hirano, N.; Machida, S.; Buchs, D.M.; Baumgartner, P.O.; Müntener, O. Pre-Subduction Metasomatic Enrichment of the Oceanic Lithosphere Induced by Plate Flexure. *Nat. Geosci.* **2016**, *9*, 898–903. [\[CrossRef\]](#)
15. Machida, S.; Kogiso, T.; Hirano, N. Petit-Spot as Definitive Evidence for Partial Melting in the Asthenosphere Caused by CO₂. *Nat. Commun.* **2017**, *8*, 14302. [\[CrossRef\]](#) [\[PubMed\]](#)
16. Falloon, T.J.; Hoernle, K.; Schaefer, B.F.; Bindeman, I.N.; Hart, S.R.; Garbe-Schönberg, D.; Duncan, R.A. Petrogenesis of Lava from Christmas Island, Northeast Indian Ocean: Implications for the Nature of Recycled Components in Non-Plume Intraplate Settings. *Geosciences* **2022**, *12*, 118. [\[CrossRef\]](#)
17. Yamamoto, J.; Korenaga, J.; Hirano, N.; Kagi, H. Melt-Rich Lithosphere-Asthenosphere Boundary Inferred from Petit-Spot Volcanoes. *Geology* **2014**, *42*, 967–970. [\[CrossRef\]](#)
18. Yang, J.; Faccenda, M. Intraplate Volcanism Originating from Upwelling Hydrous Mantle Transition Zone. *Nature* **2020**, *579*, 88–91. [\[CrossRef\]](#) [\[PubMed\]](#)
19. Kennett, B.L.N.; Engdahl, E.R.; Buland, R. Constraints on Seismic Velocities in the Earth from Traveltimes. *Geophys. J. Int.* **1995**, *122*, 108–124. [\[CrossRef\]](#)
20. Flanagan, M.P.; Shearer, P.M. Global Mapping of Topography on Transition Zone Velocity Discontinuities by Stacking SS Precursors. *J. Geophys. Res. Solid Earth* **1998**, *103*, 2673–2692. [\[CrossRef\]](#)
21. Gu, Y.J.; Dziewonski, A.M. Global Variability of Transition Zone Thickness. *J. Geophys. Res. Solid Earth* **2002**, *107*, ESE 2-1–ESE 2-17. [\[CrossRef\]](#)
22. Ringwood, A.E. Composition and Petrology of the Earth's Mantle. *MacGraw-Hill* **1975**, 618..
23. Ito, E.; Takahashi, E. Postspinel Transformations in the System Mg₂SiO₄-Fe₂SiO₄ and Some Geophysical Implications. *J. Geophys. Res. Solid Earth* **1989**, *94*, 10637–10646. [\[CrossRef\]](#)
24. Bina, C.R.; Helffrich, G. Phase Transition Clapeyron Slopes and Transition Zone Seismic Discontinuity Topography. *J. Geophys. Res. Solid Earth* **1994**, *99*, 15853–15860. [\[CrossRef\]](#)
25. Helffrich, G. Topography of the Transition Zone Seismic Discontinuities. *Rev. Geophys.* **2000**, *38*, 141–158. [\[CrossRef\]](#)
26. Deuss, A. Global Observations of Mantle Discontinuities Using SS and PP Precursors. *Surv. Geophys.* **2009**, *30*, 301–326. [\[CrossRef\]](#)

27. Bercovici, D.; Karato, S. Whole-Mantle Convection and the Transition-Zone Water Filter. *Nature* **2003**, *425*, 39–44. [\[CrossRef\]](#) [\[PubMed\]](#)
28. Karato, S. Water Distribution across the Mantle Transition Zone and Its Implications for Global Material Circulation. *Earth Planet. Sci. Lett.* **2011**, *301*, 413–423. [\[CrossRef\]](#)
29. Wood, B.J. The Effect of H₂O on the 410-Kilometer Seismic Discontinuity. *Science* **1995**, *268*, 74–76. [\[CrossRef\]](#) [\[PubMed\]](#)
30. Litasov, K.D.; Ohtani, E.; Sano, A.; Suzuki, A.; Funakoshi, K. Wet Subduction versus Cold Subduction. *Geophys. Res. Lett.* **2005**, *32*, L13312. [\[CrossRef\]](#)
31. Smyth, J.R.; Frost, D.J. The Effect of Water on the 410-Km Discontinuity: An Experimental Study. *Geophys. Res. Lett.* **2002**, *29*, 123–124. [\[CrossRef\]](#)
32. Hutchings, S.J.; Mooney, W.D. The Seismicity of Indonesia and Tectonic Implications. *Geochem. Geophys. Geosystems* **2021**, *22*, e2021GC009812. [\[CrossRef\]](#)
33. Li, J.; Ding, W.; Lin, J.; Xu, Y.; Kong, F.; Li, S.; Huang, X.; Zhou, Z. Dynamic Processes of the Curved Subduction System in Southeast Asia: A Review and Future Perspective. *Earth-Sci. Rev.* **2021**, *217*, 103647. [\[CrossRef\]](#)
34. Widiyantoro, S.; van der Hilst, R. Structure and Evolution of Lithospheric Slab beneath the Sunda Arc, Indonesia. *Science* **1996**, *271*, 1566–1570. [\[CrossRef\]](#)
35. Seton, M.; Müller, R.D.; Zahirovic, S.; Williams, S.; Wright, N.M.; Cannon, J.; McGirr, R. A Global Data Set of Present-day Oceanic Crustal Age and Seafloor Spreading Parameters. *Geochem. Geophys. Geosystems* **2020**, *21*, e2020GC009214. [\[CrossRef\]](#)
36. Hoernle, K.; Hauff, F.; Werner, R.; van den Bogaard, P.; Gibbons, A.D.; Conrad, S.; Müller, R.D. Origin of Indian Ocean Seamount Province by Shallow Recycling of Continental Lithosphere. *Nat. Geosci.* **2011**, *4*, 883–887. [\[CrossRef\]](#)
37. Taneja, R.; O'Neill, C. Constraining the Age and Origin of the Seamount Province in the Northeast Indian Ocean Using Elastic Thickness and Flexural Modelling. *Mar. Geophys. Res.* **2014**, *35*, 395–417. [\[CrossRef\]](#)
38. The GEBCO_2025 Grid—A Continuous Terrain Model for Oceans and Land at 15 Arc-Second Intervals. 2025. Available online: <https://catalogue.ceda.ac.uk/uuid/05fba4c5b8fe4daea8ff751026daf438/> (accessed on 11 January 2026).
39. Hayes, G.P.; Moore, G.L.; Portner, D.E.; Hearne, M.; Flamme, H.; Furtney, M.; Smoczyk, G.M. Slab2, a Comprehensive Subduction Zone Geometry Model. *Science* **2018**, *362*, 58–61. [\[CrossRef\]](#) [\[PubMed\]](#)
40. Hua, Y.; Zhao, D.; Xu, Y.-G. Azimuthal Anisotropy Tomography of the Southeast Asia Subduction System. *J. Geophys. Res. Solid Earth* **2022**, *127*, e2021JB022854. [\[CrossRef\]](#)
41. Huang, Z.; Zhao, D.; Wang, L. P Wave Tomography and Anisotropy beneath Southeast Asia: Insight into Mantle Dynamics. *J. Geophys. Res. Solid Earth* **2015**, *120*, 5154–5174. [\[CrossRef\]](#)
42. Li, C.; Van Der Hilst, R.D. Structure of the Upper Mantle and Transition Zone beneath Southeast Asia from Traveltime Tomography. *J. Geophys. Res. Solid Earth* **2010**, *115*, B07308. [\[CrossRef\]](#)
43. Liu, F.; Sun, D.; Yang, T.; Dai, Y. Subduction Zones Beneath Indonesia Imaged by Phase Velocity Tomography. *Geosciences* **2026**, *16*, 101. [\[CrossRef\]](#)
44. Kong, F.; Gao, S.S.; Liu, K.H.; Ding, W.; Li, J. Slab Dehydration and Mantle Upwelling in the Vicinity of the Sumatra Subduction Zone. *J. Geophys. Res. Solid Earth* **2020**, *125*, e2020JB019381. [\[CrossRef\]](#)
45. Liu, Z.; Li, J.; Kong, F. Topographic Variations of Mantle Transition Zone Discontinuities in the Java Subduction Zone and Adjacent Regions from Receiver Function Analyses. *Tectonophysics* **2023**, *849*, 229723. [\[CrossRef\]](#)
46. Deuss, A.; Woodhouse, J.H. A Systematic Search for Mantle Discontinuities Using SS-Precursors. *Geophys. Res. Lett.* **2002**, *29*, 90–94. [\[CrossRef\]](#)
47. Gu, Y.J.; Dziewonski, A.M.; Ekström, G. Simultaneous Inversion for Mantle Shear Velocity and Topography of Transition Zone Discontinuities. *Geophys. J. Int.* **2003**, *154*, 559–583. [\[CrossRef\]](#)
48. Guo, Z.; Zhou, Y. Finite-Frequency Imaging of the Global 410- and 660-km Discontinuities Using SS Precursors. *Geophys. J. Int.* **2020**, *220*, 1978–1994. [\[CrossRef\]](#)
49. Sun, S.; Zhou, Y. SS Precursor Imaging Reveals a Global Oceanic Asthenosphere Modulated by Sea-Floor Spreading. *J. Geophys. Res. Solid Earth* **2025**, *130*, e2025JB032011. [\[CrossRef\]](#)
50. Dokht, R.M.H.; Gu, Y.J.; Sacchi, M.D. Migration Imaging of the Java Subduction Zones. *J. Geophys. Res. Solid Earth* **2018**, *123*, 1540–1558. [\[CrossRef\]](#)
51. Shang, Z.; Yang, T.; Jiang, M.; Le, B.M.; Zhou, Y.; Chen, Y.; Morgan, J.P. Two Mechanisms Generating Intraplate Volcanism in Southeast Asia: Insights from Mantle Transition Zone. *J. Geophys. Res. Solid Earth* **2024**, *129*, e2024JB029090. [\[CrossRef\]](#)
52. Gu, Y.; Dziewonski, A.M.; Agee, C. Global De-correlation of the Topography of Transition Zone Discontinuities. *Earth Planet. Sci. Lett.* **1998**, *157*, 57–67. [\[CrossRef\]](#)
53. Dai, Y.; Rychert, C.; Harmon, N. Global Imaging of the Mantle Transition Zone Using SS Precursors Shows Lower Mantle Convection Patterns That Are Broadly Linked to Tectonics. *J. Geophys. Res. Solid Earth* **2026**, *131*, e2025JB032278. [\[CrossRef\]](#)
54. Shearer, P.M. Global Mapping of Upper Mantle Reflectors from Long-Period SS Precursors. *Geophys. J. Int.* **1993**, *115*, 878–904. [\[CrossRef\]](#)
55. Schmerr, N.; Garnero, E.J. Investigation of Upper Mantle Discontinuity Structure beneath the Central Pacific Ocean Using SS Precursors. *J. Geophys. Res. Solid Earth* **2006**, *111*, B08305. [\[CrossRef\]](#)

56. Houser, C.; Masters, G.; Flanagan, M.; Shearer, P. Determination and Analysis of Long-Wavelength Transition Zone Structure Using SS Precursors. *Geophys. J. Int.* **2008**, *174*, 178–194. [[CrossRef](#)]
57. Laske, G.; Masters, G.; Ma, Z.; Pasyanos, M.E.; Livermore, L. Update on CRUST1.0: A 1-Degree Global Model of Earth's Crust. *Geophys. Res. Abstr.* **2013**, *15*, 2658.
58. Amante, C.; Eakins, B.W. *ETOPO1 1 Arc-Minute Global Relief Model: Procedures, Data Sources and Analysis*; NOAA Technical Memorandum NESDIS NGDC-24; NOAA National Centers for Environmental Information: Asheville, NC, USA, 2009. [[CrossRef](#)]
59. Grand, S.P. Mantle Shear-Wave Tomography and the Fate of Subducted Slabs. *Philos. Trans. R. Soc. London. Ser. A Math. Phys. Eng. Sci.* **2002**, *360*, 2475–2491. [[CrossRef](#)] [[PubMed](#)]
60. Efron, B.; Tibshirani, R. Bootstrap Methods for Standard Errors, Confidence Intervals, and Other Measures of Statistical Accuracy. *Stat. Sci.* **1986**, *1*, 54–75. [[CrossRef](#)]
61. Houser, C.; Masters, G.; Shearer, P.; Laske, G. Shear and Compressional Velocity Models of the Mantle from Cluster Analysis of Long-Period Waveforms. *Geophys. J. Int.* **2008**, *174*, 195–212. [[CrossRef](#)]
62. Ritsema, J.; Deuss, A.; van Heijst, H.J.; Woodhouse, J.H. S40RTS: A Degree-40 Shear-Velocity Model for the Mantle from New Rayleigh Wave Dispersion, Teleseismic Traveltime and Normal-Mode Splitting Function Measurements. *Geophys. J. Int.* **2011**, *184*, 1223–1236. [[CrossRef](#)]
63. French, S.W.; Romanowicz, B.A. Whole-Mantle Radially Anisotropic Shear Velocity Structure from Spectral-Element Waveform Tomography. *Geophys. J. Int.* **2014**, *199*, 1303–1327. [[CrossRef](#)]
64. Schmerr, N.; Thomas, C. Subducted Lithosphere beneath the Kuriles from Migration of PP Precursors. *Earth Planet. Sci. Lett.* **2011**, *311*, 101–111. [[CrossRef](#)]
65. Fukao, Y.; Widiyantoro, S.; Obayashi, M. Stagnant Slabs in the Upper and Lower Mantle Transition Region. *Rev. Geophys.* **2001**, *39*, 291–323. [[CrossRef](#)]
66. Frost, D.J.; Dolejš, D. Experimental Determination of the Effect of H₂O on the 410-Km Seismic Discontinuity. *Earth Planet. Sci. Lett.* **2007**, *256*, 182–195. [[CrossRef](#)]
67. Long, X.; Ballmer, M.D.; Córdoba, A.M.-C.; Li, C.-F. Mantle Melting and Intraplate Volcanism Due to Self-Buoyant Hydrous Upwellings from the Stagnant Slab That Are Conveyed by Small-Scale Convection. *Geochem. Geophys. Geosyst.* **2019**, *20*, 4972–4997. [[CrossRef](#)]
68. Hirschmann, M.M. Partial Melt in the Oceanic Low Velocity Zone. *Phys. Earth Planet. Inter.* **2010**, *179*, 60–71. [[CrossRef](#)]
69. Sakamaki, T.; Suzuki, A.; Ohtani, E.; Terasaki, H.; Urakawa, S.; Katayama, Y.; Funakoshi, K.; Wang, Y.; Hernlund, J.W.; Ballmer, M.D. Ponded Melt at the Boundary between the Lithosphere and Asthenosphere. *Nat. Geosci.* **2013**, *6*, 1041–1044. [[CrossRef](#)]

Disclaimer/Publisher's Note: The statements, opinions and data contained in all publications are solely those of the individual author(s) and contributor(s) and not of MDPI and/or the editor(s). MDPI and/or the editor(s) disclaim responsibility for any injury to people or property resulting from any ideas, methods, instructions or products referred to in the content.