Continental Reworking in the Eastern South China Block and Its Adjacent Areas Revealed by F-J Multimodal Ambient Noise Tomography

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Abstract The eastern South China Block (SCB) has experienced complex and drastic reworking processes since the early Neoproterozoic, the mechanisms of which remain unclear. To better understand the continental reworking mechanisms in the eastern SCB, we imaged the shear-wave velocity structure of the crust and upper mantle in the eastern SCB and its adjacent areas with the F-J multimodal ambient noise tomography method. The seismic ambient noise data were acquired by 652 seismic stations in this area. The shear-wave velocity model presents widespread mid-crustal low-velocity zones (MCLVZs) and upper mantle low-velocity zones (UMLVZs), the distribution of which shows a good correlation with regional tectonics. The MCLVZs formed due to compositional and structural layering during the late Neoproterozoic to the early Phanerozoic. The disrupted MCLVZs near the Qinling-Dabie orogen can be linked to metamorphism caused by a continental collision between the SCB and the North China Craton during the Triassic Indosinian event. The thin crustal and lithospheric thicknesses east of the North-South Gravity Lineament may be the result of asthenospheric upwelling due to the subduction of the Pacific plate since the late Mesozoic. In the areas near the Shaoxing-Jiangshan-Pingxiang fault zone and the Tanlu fault zone, upheaval UMLVZs indicate that these ancient fault zones were reactivated by the subduction of the Pacific plate, providing channels for asthenospheric upwelling. Well-preserved MCLVZs north of the Qinling-Dabie orogen can be used to assess compressions and extensions induced by the subduction of the Pacific plate.

Plain Language Summary Continental reworking refers to the destruction of original and older geological systems, due to drastic tectono-thermal events such as continental collisions and ocean plate subduction. Continental reworking changes the composition, structure, and physicochemical properties of the continental lithosphere, which includes the crust and lithospheric mantle, thereby affecting the seismic velocity structure of the continental lithosphere (e.g., the formation of anomalous low- or high-velocity zones). Thus, seismic velocity imaging is an effective tool for studying the process of continental reworking. Here, we used multimodal ambient noise tomography, a new and powerful velocity imaging method, to investigate the shear-wave velocity structure of the crust and upper mantle in the eastern South China Block (SCB) and adjacent areas. Our results include two important features: mid-crustal low-velocity zones and upper mantle low-velocity zones. The detailed shear-wave velocity structure indicates a continental collision between the SCB and the North China Craton during the Triassic Indosinian event and the subduction of the Pacific plate since the late Mesozoic. Yanshanian event, and the results provide crucial knowledge on continental reworking in the eastern South China Block and adjacent areas.

1. Introduction

The South China Block (SCB), which is located in the southeastern part of the Eurasian plate, is generally believed to have developed during the early Neoproterozoic era due to the amalgamation of the Cathaysia Block to the southeast and the Yangtze Craton to the northwest (Figure 1a, Y. Wang et al., 2013). In the Triassic, the SCB collided with the North China Craton (NCC), generating the well-known high and ultrahigh-pressure (HP and UHP) metamorphic belts that constitute the Qinling-Dabie orogen (Mao et al., 2014; J. Zhang et al., 2007). Since the late Mesozoic, the Pacific plate has subducted beneath the SCB, possibly resulting in widespread
magmatism (Z. Li & Li, 2007; L. Liu et al., 2021). With this complex evolutionary history, the SCB, similar to other continents, such as the North American continent, the South American continent, and the NCC, has clearly experienced drastic reworking since its origin and accretion (Zhu et al., 2021).

Over the last two decades, numerous geological, petrological, geochemical, and geophysical studies have been performed to investigate continental reworking in the eastern SCB. Geological and petrological studies (e.g., Chu et al., 2019; Z. Li & Li, 2007; J. Liu et al., 2020; X. Zhou & Li, 2000) have generally suggested that continental reworking in the eastern SCB is associated with the subduction of the Pacific plate based on analyses of the spatiotemporal distribution of granitoids and volcanic rocks. In geochemical studies (e.g., Qiu et al., 2021; Y. Zheng et al., 2007), U-Pb, Hf, and O isotope analyses of detrital zircons from rock samples are usually performed, showing the reworking of ancient crust in the eastern SCB. Geophysical studies, including deep seismic reflection soundings (e.g., Dong et al., 2020; Lü et al., 2015; Z. Zhang et al., 2005; Z. Zhang & Wang, 2007; Z. Zhang et al., 2008; B. Zhao, Zhang et al., 2013), teleseismic tomography (e.g., J. Huang & Zhao, 2006; G. Jiang et al., 2013; C. Li & Van Der Hilst, 2010; L. Zhao et al., 2012), receiver functions (e.g., Ai et al., 2007; He et al., 2013; Q. Li et al., 2013; Wei et al., 2016), ambient noise and teleseismic surface waves (e.g., Lebedev & Nolet, 2003; J. Li, Sun et al., 2020; T. Li, Zhao, et al., 2020; Ouyang et al., 2014; Tsai & Wu, 2000; Y. Zhang et al., 2020), magnetotelluric profiles (e.g., Cheng et al., 2021; S. Xu, Unsworth, Hu, & Mooney, et al., 2019; Y. Xu et al., 2016) and joint inversion between different methods (e.g., Gao et al., 2022; L. Guo et al., 2019; Z. Guo et al., 2018; H. Li et al., 2018), have revealed velocity variations within the continental lithosphere and depth variations in velocity-jump interfaces, including Moho and mantle discontinuities in the eastern SCB. However, despite the amount of evidence provided by these studies, the mechanism for continental reworking in the eastern SCB remains controversial and mainly includes flat-slab subduction (e.g., Z. Li & Li, 2007; Shan et al., 2017; L. Zhou et al., 2012), eclogitic lower continental crust delamination (e.g., Hou et al., 2007; J. Xu et al., 2002), ridge subduction between the Pacific and Izanagi plates (Ling et al., 2009), and the “hot fingers” model related to mantle corner flow (R. Huang et al., 2015). To assess the

Figure 1. (a) Regional tectonics in the study area (after Charvet, 2013; Deng et al., 2014; H. Li et al., 2018; T. Li et al., 2022). The thick solid black lines indicate the boundaries of the main tectonic units, and the thin solid lines indicate the boundaries between provinces. The thick dashed black line represents the North-South Gravity Lineament. The abbreviations are as follows: TLF, Tanlu fault; ZDF, Zhenghe-Dapu fault; SJPFZ, Shaoxing-Jiangshan-Pingxiang fault zone; SBB, Subei Basin; NCB, North China Basin; and JHB, Jianghan Basin. The red straight lines identify the locations of the velocity slices in Figure 8. (b) Distribution of seismic stations used in this study. The red frame indicates the subarea shown in Figure 2. The blue and black frames indicate the subareas next to the red frame. The red, blue, and black solid squares indicate the center points of the stations within the red, blue and black frames, respectively.
proposed models and understand the process of continental reworking in the eastern SCB, detailed knowledge of the crust and upper mantle structures is of vital importance.

The frequency-Bessel transform (F-J) method, which was recently proposed by Wang-Wu-Chen (J. Wang et al., 2019), extracts multimodal surface wave dispersion curves from seismic ambient noise data. When higher-mode dispersion curves are incorporated into shear-wave velocity inversions, the nonuniqueness of the inversion can be significantly reduced, and the shear-wave velocity structure can be better constrained (e.g., Pan et al., 2018; G. Wu et al., 2020; H. Wu et al., 2019). Zhan et al. (2020) successfully applied the F-J method to seismic ambient noise data in Northeast China and acquired an integrated crust and upper mantle shear-wave velocity model, providing new insights into tectonic evolution in Northeast China. Fu et al. (2022) applied the F-J method to ambient noise data recorded by the Long Beach seismic array, providing an improved high-resolution 3-D shear-wave velocity model of Long Beach, CA. Because previous ambient noise surface wave studies (e.g., Shan et al., 2017; Shen et al., 2016; L. Zhou et al., 2012) in the eastern SCB were generally only based on fundamental-mode dispersion curves, the incorporation of higher-mode dispersion curves can effectively improve the knowledge of the crust and upper mantle structures in this area.

In this study, we applied the F-J method to seismic ambient noise data recorded in the eastern SCB and adjacent areas, inverting the multimodal Rayleigh wave dispersion curves to develop a shear-wave velocity model of the crust and upper mantle in this region. Because we incorporated higher-mode dispersion curves, our velocity model has more detailed characteristics than those of previous studies (e.g., Gao et al., 2022; Shen et al., 2016), especially in the crust. Low-velocity zones in both the middle crust and upper mantle present a good correlation with regional tectonics (i.e., the distribution of magmatic rocks and heat flow), providing crucial information on the process of continental reworking in this region. In addition to the continental collision between the SCB and the NCC, the westward subduction of the Pacific plate was crucial in the process of continental reworking in the eastern SCB and adjacent areas.

2. Methods

2.1. F-J Method

Considering that previous studies (e.g., Ma et al., 2022; J. Wang et al., 2019; G. Wu et al., 2020; Zhan et al., 2020) have fully explained the principle and process of the F-J method, we provide only a brief summary of this method here. Under the equipartition assumption (e.g., Campillo & Paul, 2003; Sánchez-Sesma & Campillo, 2006), the cross-correlation function (CCF) of ambient noise recordings between two receivers can be obtained. Then, the F-J method can be utilized to generate an F-J spectrogram in the frequency-velocity (f-c) domain as follows:

\[ I(\omega, c) = \int_0^{+\infty} C(\omega, r) J_0 \left( \frac{\omega}{c} r \right) dr, \]

where \( I(\omega, c) \) is the F-J spectrogram corresponding to a certain angular frequency \( \omega \) and velocity \( c \). \( C(\omega, r) \) indicates the frequency-domain CCF of ambient noise recordings between two receivers spaced at a distance of \( r \). \( J_0 \) is the zero-order Bessel function. In addition, it has been shown that the frequency-domain CCF \( C(\omega, r) \) is approximately equal to the imaginary part of Green's function between two receivers (e.g., Campillo & Paul, 2003; Lobkis & Weaver, 2001; Sánchez-Sesma & Campillo, 2006; Shapiro & Campillo, 2004) using the following formula:

\[ C(\omega, r) \approx A \cdot \text{Im} \{ G(\omega, r) \} , \]

where \( G(\omega, r) \) denotes the frequency-domain Green's function between two receivers, and \( A \) is a constant. The Green's function \( G(\omega, r) \) in a flat-layered and elastic medium (e.g., Bouchon & Aki, 1977; X. Chen, 1993, 1999) is given by:

\[ G(\omega, r) = \int_0^{+\infty} g(\omega, k) J_0(kr) dk, \]
where \( g(\omega, k) \) is a kernel function and \( k \) is the wavenumber, \( k = \omega/c \). Equations 2 and 3 can be substituted into Equation 1, as shown by J. Wang et al. (2019), to obtain:

\[
I(\omega, c) \approx A \cdot \text{Im} \left\{ g \left( \frac{\omega}{c} \right) \right\}.
\] (4)

Relevant surface wave studies (e.g., X. Chen, 1993; Kennett & Clarke, 1983) have shown that the kernel function \( g \left( \frac{\omega}{c} \right) \) approaches infinity at dispersion points. Considering the relationship between the F-J spectrogram and the kernel function, as shown in Equation 4, multimodal dispersion curves can be identified from the F-J spectrogram \( I(\omega, c) \).

### 2.2. Joint Inversion of Multimodal Dispersion Curves

Two prerequisites should be noted before well-constrained inversion is performed on the multimodal dispersion curves. First, we assume that the inversion model is a flat, multilayer velocity model that is determined by basic surface wave theories (e.g., X. Chen, 1993). In addition, most of the study area has an elevation of less than \( \sim 600 \) m, except the area west of the North-South Gravity Lineament (NSGL). The elevation variation (Figure 1) in the study area is relatively small, so the flat-layer assumption is reasonable in practice. Second, the shear velocity is more sensitive to the Rayleigh wave phase velocity than the compression velocity and density (e.g., Pan et al., 2018; Xia et al., 1999, 2003). Thus, only the shear-wave velocity needs to be inverted, and the compression wave velocity and density can be determined with empirical formulas from Brocher (2005) during the inversion. The misfit function for the inversion is given by:

\[
f(V_s) = \frac{1}{m} \sum_j A_j \left\{ \sum_i \left[ c_{ij} (V_s) - c_{ij}^0 \right]^2 \right\} + \alpha \| \Delta V_s \|^2_2,
\] (5)

where \( V_s \) is the shear-velocity model, \( c_{ij}^0 \) and \( c_{ij} \) are the synthetic phase velocity and observation phase velocity at the \( i \)th frequency point of the \( j \)th mode, respectively, \( A_j \) is a weight coefficient corresponding to the \( j \)th mode, and \( m \) is the total number of modes used during the inversion. In this study, we set the weight coefficient of each higher-mode dispersion curve to 1, and that of the fundamental-mode dispersion curve is equal to the number of all higher-mode dispersion curves, that is, \( m - 1 \). This simple strategy ensures the major contribution of the fundamental dispersion curve to the inversion as well as the improvement of the inversion due to the incorporation of higher-mode dispersion curves (Ma et al., 2022). The derivate damping, \( \| \Delta V_s \|_2^2 \), is the square of the L2 norm of the spatial Laplacian of the \( V_s \) map. During the inversion, an \( \alpha \) value of \( 5 \times 10^{-2} \) was selected based on the L-curve criteria.

After obtaining the misfit function, we adopted a quasi-Newton method known as the Broyden-Fletcher-Goldfarb b-Shanno (BFGS) algorithm (Byrd et al., 1995) to invert the shear-wave velocity model. The tolerance value, which is a criterion chosen to terminate a specific iteration of the inversion, was set to \( 1 \times 10^{-10} \) (km/s)\(^2\). To further reduce the nonuniqueness of the inversion, we simultaneously generated 80 randomly selected initial velocity models within a certain velocity range (e.g., G. Wu et al., 2020; Zhan et al., 2020). The weight-averaged value of the 40 best-converged results was used to obtain the final estimated velocity model \( \hat{m} \) as follows:

\[
\hat{m} = \frac{1}{\sum_{i=1}^H w(m_i)} \sum_{i=1}^M w(m_i) m_i.
\] (6)

where \( m_i \) is the final inverted model corresponding to the initial velocity model \( m_i^0 \), and \( w(m_i) \) is a weighted function defined as:

\[
w(m_i) = \exp (-f(m_i)),
\] (7)
Table 1: Broadband Seismic Arrays Used in This Study

| Array         | Number of stations used | Start time | End time | Operated by  |
|---------------|------------------------|------------|----------|--------------|
| H1(ChinArray) | 27                     | 2012       | 2014     | CSADMC       |
| R7(ChinArray) | 17                     | 2014       | 2015     | CSADMC       |
| W0(ChinArray) | 15                     | 2009       | 2010     | CSADMC       |
| TemporaryCo   | 256                    | 2014       | 2018     | Collaborate  |
| Permanent     | 337                    | 2012       | 2018     | CEA          |

where \( f(m) \) is the misfit function of the final model \( m_f \). To assess the accuracy of the final weight-averaged result, the standard deviation was calculated as:

\[
\sigma_m = \sqrt{\frac{1}{M-1} \sum_{i=1}^{M} (m_i - \bar{m})^2}.
\]

3. Data and Processing

The study area is located at 20°–36°N and 108°–123°E and covers the eastern SCB and eastern NCC (Figure 1a). As shown in Figure 1b, the continuous waveform data used in this study have three main parts: the first part (Permanent) was collected from 337 permanent broadband stations (Data Management Centre of China National Seismic Network, 2007; X. Zheng et al., 2010); the second part (ChinArray) was collected from 59 stations (H1, R7, and W0 arrays) operated by the China Seismic Array Data Management Center (CSADMC, ChinArray, 2006); and the third part (TemporaryCo) was collected from 256 stations deployed collaboratively by several institutions including the Chinese Academy of Geological Sciences, the China Earthquake Administration (CEA), Peking University and Nanjing University. Detailed information on the data utilized in this study is summarized in Table 1.

The most important step is to acquire reliable CCFs before implementing the F-J method. First, we selected the vertical components of all continuous waveform data after dividing the data into a series of daily segments. Second, we down-sampled the data to 5 Hz to standardize the sampling rates of different seismic networks, thus improving the computational efficiency. Next, we removed the mean, trend, and instrument response for all data. Finally, we utilized CC-FJpy, a Python package for extracting multiple-mode dispersion curves from seismic ambient noise cross-correlation (Z. Li et al., 2021), to calculate the frequency domain CCFs between all pairs of stations.

After we determined all the CCFs between all pairs of stations, we chose the size of subarrays as 6° × 5° (6° and 5° along the longitudinal and latitudinal directions, respectively, Text S1 and Figure S1 in Supporting Information S1) and divided the study region into 98 overlapping subareas as indicated by the red, blue, and black frames in Figure 1b. In detail, a given subarea (e.g., the red frame in Figure 1b) was successively moved 1° along latitudinal and longitudinal directions until it reached a neighboring subarea (e.g., the blue and black frames in Figure 1b). For each subarea, we first selected CCFs with SNRs greater than half of the mean SNR of all CCFs. Then, we applied the F-J method to the selected frequency-domain CCFs to generate high-quality F-J spectra with clear fundamental- and higher-mode dispersion curves. Finally, with dispersion points picked manually and by using DisperNet (Dong et al., 2021), a dispersion curve extraction tool, we performed the joint inversion of the multimodal dispersion curves based on the BFGS quasi-Newton algorithm.

Figure 2 shows the overall multimodal dispersion curve extraction and inversion processes, which correspond to one subarea (red frame in Figure 1b) in the center of the study region. Figure 2a shows the real part of the stacked CCFs in the frequency domain and Figure 2b shows the stacked CCFs in the time domain. Figure 2c shows the generated F-J spectrum, and the fundamental-mode and six additional higher-order mode dispersion curves can be clearly identified in the frequency range of 0.01–0.80 Hz. Figure 2d shows the final inverted result, which was obtained by weight averaging the 40 best-converged models from 80 randomly selected initial models in the range of ±0.4 km/s relative to the smoothed reference velocity model. We adopted the shear-wave velocity model proposed by Shen et al. (2016) as a reference model. The reference velocity model has a relatively sharp velocity jump near the Moho discontinuity (Figure S2 in Supporting Information S1), which introduces artificial boundaries (An, 2020) during the BFGS quasi-Newton inversion. To prevent the introduction of any artifacts, we smoothed the reference model before performing the inversion (Figure S2 in Supporting Information S1). The corresponding standard deviation of the final weighted model is shown in the right column of Figure 2d.

We calculated the sensitivity kernel map for the multimodal dispersion curves (Figure 3), showing the sensitivity of the dispersion curves to the shear-wave velocity (Pan et al., 2018), at certain depths and frequencies. Figure 3 shows that the fundamental-mode dispersion curve constrains the shear-wave velocity at depths of up to 175 km (Figure 3a), exceeding the objective inverted depth of 150 km. Additionally, higher-order mode dispersion curves are highly sensitive to the shear-wave velocity at depths of 0–50 km (Figures 3b–3f). Therefore, with
the incorporation of the multimodal dispersion curves, the shear-wave velocity structure at 0–150 km should be effectively constrained, as shown in Figure 2d.

4. Results

We used the inverted 1-D shear-wave velocity profiles corresponding to different subareas to obtain an integrated 3-D shear-wave velocity model (Figure 4) of the crust and upper mantle in the study area using the Kriging interpolation algorithm (Virdee & Kotegoda, 1984). Assuming that the shear-wave velocity gradient is the steepest at the Moho discontinuity, the Moho depth (Figure 5, dashed gray lines in Figures 6–8) can be obtained by determining the depth with the largest shear-wave velocity gradient at depths below 20 km. The integrated 3-D shear-wave velocity model is presented as nine vertical slices along the latitudinal direction (Figures 6a and 7a) and nine vertical slices along the longitudinal direction (Figures 6b and 7b). Figure 8 presents six oblique transects across the main tectonic units within the study area. These velocity slices are presented for two depth ranges,
Figure 3. Sensitivity kernels of the multimodal dispersion curves shown in Figure 2. The red shaded areas indicate sensitivity distributions relative to the depth and frequency, while the black points represent the selected dispersion curves of each mode (i.e., phase velocity vs. frequency), the values of which can be read on the right-side vertical axes.
0–60 km (Figures 4a, 6, and 8a) and 0–150 km (Figures 4b, 7, and 8b), to better show the velocity characteristics of the crust and upper mantle, respectively.

4.1. Moho Depth Distribution

Figure 5 shows that the Moho depth within this area ranges from 27 to 44 km, with the depth generally becoming thinner from west to east. The boundary line (thick and green dashed line in Figure 5) across which the Moho depth varies from deep to shallow is outlined. This boundary line generally coincides with the NSGL, which can also be observed in the vertical velocity slices along the latitudinal direction (Figure 6a). The Moho depth across the Tanlu fault presents a different picture; that is, the Moho depth in the Subei Basin to the east is ∼5 km shallower than that in the North China Basin (NCB) to the west. In addition, the south-central part, located between the Shaoxing-Jiangshan-Pingxiang fault zone (SJPFZ) and the Zhenghe-Dapu fault (ZDF), also presents a much shallower Moho depth (∼26–30 km).

4.2. Mid-Crustal Low-Velocity Zones (MCLVZs)

The middle crust (∼10–20 km) includes relatively widespread low-velocity zones (LVZs) with velocity ranges of 3.4–3.5 km/s (Figures 4a, 6, and 8a). For velocity slices along the latitudinal direction (Figure 6a), mid-crustal low-velocity zones (MCLVZs) south of 31°N cannot be continuously traced from west to east, especially in the middle region (29°–31°N) of the study area. In contrast, velocity slices north of 31°N have more noticeable and continuous MCLVZs; however, the MCLVZs in the velocity slices along 32°N and 33°N appear to be curved or slightly disrupted. Several velocity slices (profiles 113°–116°E in Figure 6b, BB’ and FF’ in Figure 8a) have

**Figure 4.** Integrated 3-D shear-wave velocity model, at depths of (a) 0–60 km and (b) 0–150 km, with a corner cut. The topography (same as Figure 1) is covered on the surface.
disrupted MCLVZs near the Qinling-Dabie orogen, which is the boundary of the Yangtze Craton and the NCC. Additionally, the MCLVZs show depth variations across the Tanlu fault (profiles 32°N and 33°N in Figure 6a, 117°E in Figure 6b, and CC’ and FF’ in Figure 8a). However, not all velocity slices (Figures 6 and 8a) present clear morphological variation in the MCLVZs at the boundary of the Yangtze Craton and the Cathaysia Block.

4.3. Upper Mantle Low-Velocity Zones (UMLVZs)

Another prominent feature in the integrated 3-D shear-wave velocity model is the existence of LVZs (4.2–4.3 km/s) at depths of ~60–120 km in the upper mantle (Figures 4b, 7, and 8b). In contrast to the MCLVZs, three clear characteristics of these UMLVZs can be observed. First, the distribution of the UMLVZs is more widespread and continuous. For example, the velocity slices from 29°N to 31°N in Figure 7a show that continuous UMLVZs can easily be traced from west to east, which is not possible for the MCLVZs (profiles 29°−31°N in Figure 6a) at the same latitudes. Second, most velocity slices (Figure 7b, and profile AA’ in Figure 8b) have curved or disrupted UMLVZs near the boundary of the Yangtze Craton and the Cathaysia Block. Third, the UMLVZs in most of the velocity slices do not present the same behavior near the Qinling-Dabie orogen as the MCLVZs in the velocity slices, namely, curved or disrupted shapes (profiles 113°–116°E in Figure 6b, BB’ and FF’ in Figure 8a), with the exception of the velocity slices along 112°E and 113°E in Figure 7b. In addition, the UMLVZs mostly present similarly convex morphology variations (Figure 7, CC’ and FF’ in Figure 8b) near the Tanlu fault zone.

5. Discussion

Our 3-D shear-wave velocity model presents LVZs in both the middle crust and the upper mantle in the eastern SCB and adjacent areas. Compared with the MCLVZs, the UMLVZs are more continuous and widespread. Previous studies (e.g., An & Shi, 2006; Y. Xu, Zheng, Yang, & Xia, 2019) have shown that the lithosphere-asthenosphere boundary (LAB) may coincide with the top boundary of the UMLVZs. Accordingly, we also extracted the depth distribution of the top boundary of the UMLVZs, as shown in Figure 19b. To the east of the NSGL, the average depth (~60 km) of the top boundary for the UMLVZs is consistent with the LAB depth revealed by some previous studies (e.g., Deng et al., 2019; Q. Li et al., 2013; Y. Zhang et al., 2020), although it is ~20 km shallower than that reported by other studies (e.g., Shan et al., 2021; Y. Zhang et al., 2019, 2018). Thus, it is relatively reasonable that the top of the UMLVZs represents the LAB east of the NSGL. The lateral variation in the top depth of the UMLVZs reflects uneven asthenospheric upwelling in the eastern region. In contrast, to the west of the NSGL (area A in Figure 19b), the top depth (~70 km) of the UMLVZs is much shallower than the LAB depth (>150 km) revealed by previous studies (e.g., An & Shi, 2006; T. Li et al., 2022; Shan et al., 2017, 2021). Therefore, defining such a shallow boundary in the upper mantle as the LAB is unreasonable. However, we speculate that this low-velocity layer (LVL) may represent the mid-lithospheric discontinuity (MLD) within the ancient and thick lithosphere in the Yangtze Craton (L. Chen et al., 2014; Fischer et al., 2010; Y. Xu, Zheng, et al., 2019), which calls for more evidence.

In the following sections, we first compare our imaging results with previous tomography results in this area and validate our velocity model with the waveform simulation method (Lu & Ben-Zion, 2022); then, we correlate the spatial distribution of two LVZs with regional tectonics, including the heat flow and magmatic rocks. Subsequently, we discuss the MCLVZs formation mechanism. Finally, we attempt to elucidate the mechanism for continental reworking in the eastern SCB and its adjacent areas.
Figure 6. Vertical slices of the shear-wave velocity structure at depths of 0–60 km. (a) Vertical slices of the shear-wave velocity structure along the latitudinal direction from 26°N to 34°N. (b) Vertical slices of the shear-wave velocity structure along the longitudinal direction from 111°E to 119°E. The dashed gray lines represent the estimated Moho boundaries (depth boundary with the steepest velocity gradient at depths below 20 km). The abbreviations are as follows: NSGL, North-South Gravity Lineament; YC, Yangtze Craton; CB, Cathaysia Block; NCC, North China Craton; TL, Tanlu fault; ZD, Zhenghe-Dapu fault; QD, Qinling-Dabie orogen; and SBB, Subei Basin.
Figure 7. Vertical slices of the shear-wave velocity structure at depths of 0–150 km. (a) Vertical slices of the shear-wave velocity structure along the latitudinal direction from 26°N to 34°N. (b) Vertical slices of the shear-wave velocity structure along the longitudinal direction from 111°E to 119°E. The dashed gray lines represent the estimated Moho boundaries (depth boundary with the steepest velocity gradient at depths below 20 km). The abbreviations are the same as those in Figure 6.
5.1. Model Comparison and Validation

5.1.1. Model Comparison

Here, we compare our imaging results with the results of two previous tomography studies, including Gao et al. (2022) and Shen et al. (2016). Shen et al. (2016) developed a shear-wave velocity model (Shen2016) across the continental crust. Figure 8 shows vertical slices of the shear-wave velocity structure at (a) 0–60 km and (b) 0–150 km along the six profiles identified in Figure 1a. The dashed gray lines represent the estimated Moho boundaries (depth boundary with the steepest velocity gradient at depths below 20 km). The abbreviations are the same as those in Figure 6.

Figure 8. Vertical slices of the shear-wave velocity structure at (a) 0–60 km and (b) 0–150 km along the six profiles identified in Figure 1a. The dashed gray lines represent the estimated Moho boundaries (depth boundary with the steepest velocity gradient at depths below 20 km). The abbreviations are the same as those in Figure 6.
China by collecting ambient noise Rayleigh wave tomography and earthquake tomography with fundamental-mode dispersion curves ranging from 8 to 70 s. According to the surface wave data from Shen et al. (2016), Gao et al. (2022) introduced earthquake body wave data and performed a joint inversion to obtain a shear-wave velocity model (Gao2022) of the SCB. The velocity slice along 33°N is taken as an example (Figure 9) to compare our velocity model with these two models. Compared with Shen2016 (Figure 9f), our result (Figure 9d) and Gao2022 (Figure 9b) present more continuous LVZs in the upper mantle, which is largely due to more independent data (i.e., the earthquake body wave data in Gao et al. (2022) and the higher-mode dispersion curves in this study) incorporated into the shear-wave velocity inversion. Within the depth range of 50–70 km, however, the shear-wave velocity presented in Gao2022 is clearly higher than the shear-wave velocities presented in this study and Shen2016. It is worth mentioning that the UMLVZs in our result (Figure 9d) present clearer morphology variations across the NSGL and the Tanlu fault zone than the other two models. Second, our results (Figure 9c) are significantly different from the results of these two studies (Figures 9a and 9e) in the crust, as we demonstrate the existence of MCLVZs. Gao2022 (Figure 9a) and Shen2016 (Figure 9e) have almost the same characteristics in the crust, although Gao et al. (2022) introduced earthquake body wave data during the shear-wave velocity inversion. The reason for this is that long-period body wave data cannot strongly constrain the shallow velocity

Figure 9. Velocity slices along 33°N from three velocity models, including Gao et al. (2022) (a, b), this study (c, d), and Shen et al. (2016) (e, f). The left and right columns present the velocity structure within the crust and upper mantle, respectively. The abbreviations are the same as those in Figure 6.
structure. Because the same fundamental-mode dispersion data were utilized in both Gao et al. (2022) and Shen et al. (2016), they obtained largely similar velocity structures in the crust.

We chose a northern subarea (Figure 10a) to further explain how higher-mode dispersion curves impact the shear-wave velocity inversion. The center point of the stations in this subarea is (118.06°E, 32.92°N) (the blue diamond in Figure 10a), which is close to the location of the velocity slice (Figures 9c and 9d) along 33°N. Figure 10b presents the corresponding F-J spectrum, from which the fundamental-mode and another three higher-mode dispersion curves can be extracted. First, we obtain the inverted velocity model (Figure 10c, Fundamental) by using only the fundamental-mode dispersion curve. Then, we obtain another inverted velocity model (Figure 10c, Multimodal) by using the fundamental-mode and another three higher-mode dispersion curves. Additionally, we plot two previous velocity models, Shen2016 and Gao2022 (Figure 10c), for comparison. When three higher-order mode dispersion curves are not incorporated, the inverted result (Fundamental) is similar to Shen2016, which was obtained by using only the fundamental-mode dispersion curve (Shen et al., 2016). Moreover, the inverted result (Fundamental) is almost the same as the previous two inverted velocity models, Shen2016 and Gao2022, at depths of less than 40 km, validating that the same fundamental-mode dispersion curve used in Gao et al. (2022) and Shen et al. (2016) determines their similar velocity characteristics (Figures 9a and 9e) in the crust. In contrast, the velocity structure (Figure 10c, Multimodal) within the crust is clearly improved and presents more detailed characteristics (i.e., the existence of the MCLVZ) when the fundamental-mode and another three higher-order mode dispersion curves are incorporated into the inversion. As shown in Figure 11, the
fundamental-mode dispersion curve (Figure 11a) constrains the velocity structure at depths greater than 60 km, while the other three higher-mode dispersion curves (Figures 11b–11d) improve the velocity structure at depths of less than 60 km because of their high sensitivity at this depth range. We also plot the theoretical dispersion curves calculated according to the four velocity models shown in Figure 10c against the selected dispersion points (Figure 10d, Data). The synthetic dispersion curves (Figure 10d, Multimodal) calculated according to the multimodal inverted model are consistent with the real-world data, which includes the fundamental-mode and the other three higher-order mode dispersion curves.

5.1.2. Model Validation

Other than providing valuable information on the tectonic evolution, seismic velocity models can be safely applied in many disciplines, such as dynamic rupture processes and strong ground motion simulations, when they are accompanied by model uncertainty estimates and effective validation. Model uncertainty estimates (Figure S3 in Supporting Information S1) show a generally low uncertainty for our velocity model, although the uncertainty increases slightly at deeper depths. In regard to model validation or evaluation, checkerboard resolution tests are widely used in which a model with a checkerboard pattern is input to generate synthetic traveltime data which subsequently are inverted to recover the input model. However, the checkerboard tests may provide misleading remarks on the true model resolution because the same forward solver is used during the inversion (Rawlinson & Spakman, 2016). In addition, the real data noise may not follow a Gaussian distribution, which is usually assumed in checkerboard tests. More importantly, the F-J multimodal ambient noise tomography method we adopted in

Figure 11. Sensitivity kernels of the multimodal dispersion curves shown in Figure 10. The red shaded areas indicate the sensitivity distributions relative to the depth and frequency, while the black points represent the selected dispersion curves of each mode (i.e., phase velocities vs. frequency), the values of which can be read from the right-side vertical axes.
this study considers a more realistic 3-D wave propagation problem (J. Wang et al., 2019), so the checkerboard test method, which is based on the ray propagation assumption, may not be appropriate for evaluating our results.

Considering the intrinsic pitfalls of checkerboard tests and the specific characteristics of our F-J method, we chose to further validate our velocity model and two other models (Shen2016 and Gao2022), as mentioned in Section 5.1.1, with the waveform simulation method proposed by Lu and Ben-Zion (2022). Compared with the checkerboard test, the waveform simulation method is also a synthetic reconstruction method to evaluate the resolving power of velocity models. However, this method provides some more objective metrics, rather than “self-justification” examinations such as checkerboard tests, to evaluate different tomography models at the same time (Lu & Ben-Zion, 2022). Additionally, validating our model with other independent data (i.e., waveform data) rather than dispersion curves, which were previously used to construct the model, is more reasonable. Due to the large scarcity of earthquakes within the study area, we chose to synthesize only the CCFs using the finite-difference method (W. Zhang & Chen, 2006; W. Zhang et al., 2012). The 60 seismic stations are set as virtual sources (Text S2 Supporting Information S1 and Figure 12), which have relatively uniform spatial distributions. The synthesized CCFs are compared with the observed CCFs (Figures 13 and 14) in four different periods including 3–7, 5–10, 8–15, and 12–20 s. Then, we calculated the mean absolute deviation of traveltime delay (CT_MAD) and mean correlation coefficient \( \overline{CCC} \), which were subsequently regionalized, as shown in Figures 15–18. A detailed description of this waveform simulation method, which is beyond the scope of the current discussion, was comprehensively reported in Lu and Ben-Zion (2022). Here, we focus only on the comparison of regionalized CT_MAD and \( \overline{CCC} \) between these three models.

CT_MAD and \( \overline{CCC} \) evaluate the accuracy of relative velocity variations in the models from traveltime delay and relative amplitude, respectively. Specifically, CT_MAD measures the statistical dispersion of the traveltime delay measurements, while \( \overline{CCC} \) measures the waveform similarity between the observed CCFs and the synthetic CCFs after traveltime delay correction. Generally, the larger CT_MAD is, the smaller \( \overline{CCC} \) is (Lu & Ben-Zion, 2022). Figure 18 shows that these three models have slight differences in the spatial distributions of CT_MAD and \( \overline{CCC} \) for 12–20 s. When periods become shorter (Figures 15–17), our model performs better than the other two models (Shen2016 and Gao2022). In detail, our model shows a relatively higher \( \overline{CCC} \) than the other two models in the southern part (i.e., south of 30°N), especially in the Cathaysia Block. More saliently, our model presents a

![Figure 12. Two examples of setting noise-correlation virtual sources. The 60 stations (red triangles) among 652 seismic stations (gray and red triangles) are selected as virtual sources, which have relatively uniform spatial distributions. The two diamonds in (a) and (b) refer to the current virtual sources (HA.SQ and JS.XIY) used to synthesize the noise-correlation waveforms shown in Figures 13 and 14.](image-url)
relatively lower CT_MAD than Shen2016 and Gao2022 in the north of the Qinling-Dabie orogen (i.e., north of 32°N), which reflects a relatively smaller traveltime delay, as shown in the waveform fitting results (Figures 13 and 14).

5.2. Correlation With Regional Tectonics

In Section 5.1, we have validated our 3-D shear-wave velocity model and thus the reliability of two LVZs. Such widespread MCLVZs and UMLVZs may have played a significant role in shaping the regional tectonics. To better understand the correlation between the spatial distribution of the two LVZs and the regional tectonics, we

Figure 13. Waveform fitting comparison for the virtual source HA.SQ (Figure 12a) between different models (columns) for different periods (rows).
first mapped the depth distribution of the top boundary of the MCLVZs and UMLVZs in Figures 19a and 19b, respectively. Then, we collected the heat flow data (G. Jiang et al., 2019) and the age data of magmatic rocks (Lehnert et al., 2000; J. Liu et al., 2020) in the eastern SCB and mapped their distribution in Figures 19c and 19d, respectively.

As Figure 19 shows, we divided the entire study area into four subareas labeled A, B, C, and D. For area A, the depth distribution of the top boundary of the MCLVZs and UMLVZs (Figures 19a and 19b) has little lateral variation and almost maintains a steady value (~11 km for the MCLVZs and ~70 km for the UMLVZs), indicating a long and quiet tectonic period for area A. Moreover, the ages for most magmatic rocks in area A (Figure 19d) are

Figure 14. Waveform fitting comparison for the virtual source JS.XIY (Figure 12b) between different models (columns) for different periods (rows).
almost over 700 Ma, indicating that these rocks were at least formed before the late Neoproterozoic. In other words, area A may have maintained its original state since the late Neoproterozoic, and it was not modified by the Phanerozoic orogenies, which in turn provides some evidence for our speculation that the UMLVZs in area A may represent the MLD within the ancient and steady lithosphere in the Yangtze Craton. Meanwhile, this result also corroborates that the MCLVZs may have formed before the Phanerozoic orogenies.

Compared with areas A and B, areas C and D show four distinct characteristics: (a) the depths of the top boundary of the UMLVZs (Figure 19b) are much shallower, indicating widespread asthenospheric upwelling in areas C and D; (b) the MCLVZs (Figure 19a) show a less continuous or fragmented distribution than those in areas A and
B, indicating that they were destroyed; (c) the distribution of magmatic rocks (Figure 19d) is more widespread, indicating that more drastic magmatic activity has occurred in areas C and D; and (d) the heat flow (Figure 19c) is clearly higher than that in area B (average heat flow in areas B, C, and D is 57.24, 69.29, and 75.21 mW/m², respectively). In short, well-preserved MCLVZs correspond to the areas with the deeper top boundary of the UMLVZs, the less widespread magmatic rocks and the lower heat flow (Note: the heat flow in area A is overlooked because of its scarcity for data).

Figure 16. Spatial distributions of the CT_MAD (top row) and $\overline{\text{CC}}\overline{\text{C}}$ (bottom row) for Rayleigh waves (5–10 s) for different models (columns).
5.3. Formation Mechanism of the MCLVZs

The MCLVZs, which range in velocity from 3.4 to 3.5 km/s, are relatively widespread in our 3-D shear-wave velocity model (Figures 4a, 6, 8a, and 19a). Previous studies have found many LVZs in the mid-lower crust in the same study area. For example, Z. Zhang et al. (2005) obtained the shear-wave velocity structure below a long (380 km) wide-angle seismic profile from Tunxi, Anhui Province, to Wenzhou, Zhejiang Province, and found an LVZ in the bottom of the mid-crust region (∼20 km). By comparing the \( P \)-wave velocity, \( S \) wave velocity, and Lamé impedances with laboratory data of these physical variables on a series of crustal rock samples, Z. Zhang et al. (2008) inferred that this LVZ (3.4–3.5 km/s) corresponded to gabbro and a rock with a mafic composi-
tion and a granular texture. T. Li, Zhao, et al. (2020) utilized seismic ambient noise imaging to determine the shear-wave velocity structure below a 500 km long dense array spanning from the NCB to the Cathaysia Block. A relatively broad LVZ observed in the mid-lower crust (10–30 km) near the Tanlu fault zone was attributed to partial melting. Near Jiuyi Mountain, J. Li, Sun, et al. (2020) found a linear and continuous low-velocity anomaly in the middle crust (10–20 km), implying the existence of metasedimentary rocks formed by the collision between the Yangtze Craton and the Cathaysia Block. P. Zhou et al. (2020) imaged the mid-crustal LVL below the eastern Cathaysia Block using the receiver function imaging method and suggested that the mid-crustal LVL was produced by fluid accumulation due to dehydration in the Paleo-Pacific subduction plate.

Figure 18. Spatial distributions of the CT_MAD (top row) and $\Delta \Delta C^2$ (bottom row) for Rayleigh waves (12–20 s) for different models (columns).
Figure 19. (a–b) The depth distribution of the top boundary of the MCLVZs (MCLVZs*, (a)) and UMLVZs (UMLVZs*, (b)). (c) The distribution of the measured terrestrial heat flow from the China Heat Flow Database (CHFDB, https://chfdb.xyz, G. Jiang et al., 2019). (d) The spatial distribution of magmatic rocks from J. Liu et al. (2020) and EarthChem Portal (http://portal.earthchem.org/, Lehnert et al., 2000). The study area is divided into four subareas labeled A, B, C, and D. MLD: mid-lithospheric discontinuity. Note: the heat flow in area A is overlooked due to a lack of data.
In conjunction with previous studies (e.g., J. Li, Sun, et al., 2020; T. Li, Zhao, et al., 2020; Z. Zhang et al., 2005, 2008; P. Zhou et al., 2020) and the good correlation (Figure 19) between the spatial distribution of the two LVZs (MCLVZs and UMLVZs) and the regional tectonics (i.e., the spatial distribution of magmatic rocks and heat flow), we believe that the formation of the MCLVZs in our 3-D shear-wave velocity model can be attributed to the development of structural and compositional layering in the continental crust during the reworking process. According to Zhu et al. (2021), although refining and differentiation processes usually lead to a felsic granodioritic upper crust and a mafic gabbroic lower crust, the actual processes are more sophisticated. It is possible that certain low-velocity rocks can be formed at specific temperatures and pressures corresponding to mid-crustal depths (e.g., Z. Zhang et al., 2008), resulting in the formation of widespread MCLVZs. In addition, Deng et al. (2014) combined P wave velocity and Bouguer gravity anomalies to investigate the density structure of the eastern SCB and revealed widespread low-density zones in the middle crust (10–20 km). Moreover, the low-density zones observed by Deng et al. (2014) have densities (2.65–2.75 g/cm$^3$) similar to those of the MCLVZs (2.66 g/cm$^3$), as reported by Z. Zhang et al. (2008); thus, the MCLVZs with low densities may indicate the presence of rocks with low shear-wave velocities and densities.

When were such widespread MCLVZs formed? The SCB underwent rifting activity (Charvet, 2013; J. Wang & Li, 2003; G. Zhang et al., 2013) featuring continental rift basins (e.g., the Nanhua Rift) after it was amalgamated between the Yangtze Craton and the Cathaysia Block. However, the SCB had received steady deposits until at least the Ordovician (J. Wang & Li, 2003; G. Zhang et al., 2013) since the rifting activity ended in the late Neoproterozoic (≥690 Ma). Additionally, as Figure 19 shows, the steady depth distribution (Figure 19a) of the top boundary for the MCLVZs in area A correlates well with the uniformly old age distribution (>700 Ma, Figure 19d) of magmatic rocks, clearly indicating that the SCB may have maintained steady tectonic state since the late Neoproterozoic. In short, this significantly long period of quiescence, that is, from the late Neoproterozoic to the early Phanerozoic, provided a unique opportunity for the continental crust to complete its structural and compositional layering in which widespread MCLVZs were formed. During the Phanerozoic era, however, the eastern SCB experienced three major tectonothermal events (e.g., Y. Wang et al., 2013), including the Kwangsian movement in the middle Paleozoic era, the Indosinian movement in the Triassic era, and the Yanshanian movement in the Jurassic-Cretaceous era (the late Mesozoic era). As a result of these intense tectonothermal events, widespread MCLVZs were unlikely to have been formed and preserved and instead were more likely to have been destroyed during the Phanerozoic orogenies.

5.4. Mechanism for Continental Reworking in the Eastern SCB and Adjacent Areas

As previously discussed, the eastern SCB may have experienced a relatively long and quiet tectonic period from the late Neoproterozoic to the early Phanerozoic (J. Wang & Li, 2003; G. Zhang et al., 2013). The nature of the Kwangsian movement in the middle Paleozoic era is still under debate, with the key disputing point on plate collision versus intracontinental regime (Y. Wang et al., 2013; G. Zhang et al., 2013; Charvet, 2013), for which evaluating its role in continental reworking in the eastern SCB may be difficult. Therefore, most previous studies (e.g., Chu et al., 2019; Deng et al., 2019; Hou et al., 2007; R. Huang et al., 2015; J. Li, Cawood, et al., 2020; Z. Li
Due to the subduction of the Paleo-Pacific plate, the eastern SCB has a large number of NE-trending fault zones, folds, and granitoid-volcanic rocks (e.g., Z. Li & Li, 2007; Ling et al., 2009; L. Liu et al., 2021; Y. Wang et al., 2013; J. Xu et al., 2002; X. Zhou & Li, 2000; X. Zhou et al., 2006) have nearly reached a consensus that tectonic activities since the late Mesozoic played a vital role in the process of continental reworking in the eastern SCB and its adjacent areas. Moreover, the ages of most magmatic rocks in the eastern SCB (Figure 19d) are mainly concentrated in ~100–200 Ma, further indicating widespread magmatism since the late Mesozoic. Generally, previous models can be divided into two categories: those related to slab subduction (e.g., Chu et al., 2019; R. Huang et al., 2015; J. Li, Cawood, et al., 2020; Z. Li & Li, 2007; Ling et al., 2009; L. Liu et al., 2021; Y. Wang et al., 2013; X. Zhou & Li, 2000; X. Zhou et al., 2006) and those related to asthenospheric upwelling or lithospheric delamination (e.g., Deng et al., 2019; Hou et al., 2007; J. Xu et al., 2002). The latter type has difficulties determining what kind of tectonic process or dynamic mechanism drives asthenospheric upwelling and lithospheric delamination (Y. Wang et al., 2013). Furthermore, the momentum of the latter may be derived from the former (e.g., Ouyang et al., 2014; Shan et al., 2017). Therefore, models related to slab subduction are more accepted. Nevertheless, models related to slab subduction have several key differences. Specifically, Z. Li and Li (2007) first proposed a flat subduction model to explain the formation of the 1,300 km wide intracontinental orogen that migrated from the coastal region to the continental interior during the Mesozoic. Chu et al. (2019) and X. Zhou and Li (2000) used a detailed analysis of the spatiotemporal distribution of Mesozoic granitoids and volcanic rocks in the eastern SCB to show that the subduction angle of the Paleo-Pacific plate has varied over time. Y. Jiang et al. (2015) and J. Liu et al. (2020) suggested that there were multiple stages of slab advancement and retreat during the subduction process of the Paleo-Pacific plate. Similarly, J. Li, Cawood, et al. (2020) proposed that the Paleo-Pacific plate repeatedly advanced and retreated by analyzing the geometry, kinematics, and age of the Lianhuashan fault zone. Additionally, Y. Wang et al. (2013) and X. Zhou et al. (2006) stressed the far-field effect of the Paleo-Pacific plate, especially in the initial subducting phase. To explain the distribution characteristics of the Lower Yangtze River metallogenic belt (LYRMB), Ling et al. (2009) proposed a mid-ocean ridge subduction model, in which the mid-ocean ridge between the Pacific plate in the south and the Izanagi plate in the north drifted toward and subducted under the LYRMB. Because the Eurasia plate is subducted sub-vertically beneath the Philippine plate, R. Huang et al. (2015) proposed the “hot fingers” model to explain the underlying mechanism of crustal thinning in the middle-south part of the eastern SCB.

Similarly, we believe that the subduction of the Paleo-Pacific plate since the late Mesozoic has contributed considerably to the reworking process of the eastern SCB. According to the Moho depth distribution (Figure 5), the crustal thickness is clearly thinner east of the NSGL than west of the NSGL, which is consistent with the results of previous receiver function studies (e.g., R. Huang et al., 2015). Moreover, the shallower LAB depth east of the NSGL, which is estimated according to the top of the UMLVZs (Figure 19b), provides solid evidence for lithospheric thinning of the eastern SCB (e.g., An & Shi, 2006) since the late Mesozoic. Because the eastern NCC has also experienced noticeable lithospheric thinning of the eastern SCB (e.g., An & Shi, 2006) since the late Mesozoic. Because the eastern NCC has also experienced noticeable lithospheric thinning since the late Mesozoic and shares a similar seismic anisotropy distribution with the eastern SCB (e.g., An & Shi, 2006) since the late Mesozoic. Because the eastern NCC has also experienced noticeable lithospheric thinning since the late Mesozoic and shares a similar seismic anisotropy distribution with the eastern SCB (e.g., An & Shi, 2006). Thus, the eastern SCB was likely in the Pacific regime (extension) during the late Mesozoic. A large number of previously overlooked Triassic (early Indosinian) NW/WNW-trending faults and folds can also be observed in the eastern SCB (Y. Wang et al., 2013). These faults and folds developed in the Tethyan regime due to compression (e.g., Y. Wang et al., 2013; X. Zhou et al., 2006). Additionally, the Qinling-Dabie orogen, the largest HP and UHP metamorphic zone in the world formed during the Tethyan regime due to the collision between the SCB and the NCC (e.g., G. Zhang et al., 2013). The transition from the Tethyan regime to the Pacific regime may have occurred in the Early Jurassic because this period included a magmatically inactive period of approximately 25 Ma (X. Zhou et al., 2006).

Whether the continental collision occurred during the Tethyan regime or slab subduction occurred during the Pacific regime, identifiable scars are evident in our 3-D shear-wave velocity model. For instance, many velocity slices (profiles 113°–116°E in Figure 6b, BB’ and FF’ in Figure 8a) present disrupted MCLVZs near the Qinling-Dabie orogen, which we believe is due to a continental collision between the SCB and the NCC. As
discussed above, the MCLVZs formed due to compositional and structural layering in the continental crust during the reworking process. Moreover, a long tectonically quiet period with essentially no large tectonothermal events from the late Neoproterozoic to the early Phanerozoic allowed for the formation of MCLVZs. Additionally, because of this long period of tectonic quiescence, the morphology of the MCLVZs was initially smooth and continuous. As a result, a collision between the SCB and the NCC during the Triassic Indosinian movement may have destroyed the initially continuous MCLVZs near the Qinling-Dabie orogen. During this process, low-velocity and low-density rocks in the middle crust were metamorphosed into high-velocity and high-density rocks, disrupting the initially continuous MCLVZs and the mid-crustal low-density zones near the Qinling-Dabie orogen (Deng et al., 2014).

With the exception of the disrupted MCLVZs near the Qinling-Dabie orogen, the impact of the Triassic Indosinian movement on our velocity model was essentially overshadowed by the late Mesozoic Yanshanian movement. The reason for this phenomenon is that the late Mesozoic Yanshanian movement had a considerable influence on the eastern SCB. The Paleo-Pacific plate has subducted beneath the SCB since the late Mesozoic, and its front has extended well into the eastern SCB (≈105°E), as shown by relevant tomography studies (e.g., J. Huang & Zhao, 2006; C. Li & Van Der Hilst, 2010). Due to the extensive subduction of the Pacific plate, some lithosphere-scale suture lines or faults, including the SJPFZ and the Tanlu fault zone, were reactivated, providing natural and convenient channels for asthenospheric upwelling. As a result, the UMLVZs generally have convex structures (Figures 7, 8b, 9d) near these fault zones. More obviously, the top depths of the UMLVZs in areas east of the Tanlu fault and between the SJPFZ and the ZDF (areas C and D in Figure 19d) are much shallower. Furthermore, hot materials due to asthenospheric upwelling were likely transported to the crust along these fault channels, where they reacted with materials from the MCLVZs. Subsequently, these mixed hot materials were driven toward the surface along these large-scale fault zones. On the one hand, this disrupted the continuity of the MCLVZs, making them unclear and untraceable (profiles 26°−31°N in Figures 6a and 19a). On the other hand, mixed hot materials that migrated to the surface likely formed large amounts of late Mesozoic magmatic rocks (Figure 19d).

Notably, MCLVZs north of the Qinling-Dabie orogen (profiles 32°−34°N in Figures 6a and 19a) are more continuous and clearer than those south of the Qinling-Dabie orogen (profiles 26°−31°N in Figure 6a). Previous studies (e.g., J. Huang & Zhao, 2006; C. Li & Van Der Hilst, 2010; L. Zhao et al., 2012) have proven that the subducting front of the Pacific plate not only reached the eastern margin of the NCC (≈118°E) but also extended into the eastern SCB (≈105°E). Due to the lower impact of this subduction, the MCLVZs north of the Qinling-Dabie orogen have been well preserved. In addition, several previous studies (e.g., Chu et al., 2019; J. Li, Cawood, et al., 2020; J. Liu et al., 2020; Y. Jiang et al., 2015) have indicated that the subduction of the Paleo-Pacific plate induced multistage compressions and extensions of the SCB, which can be observed in the morphological variations in the well-preserved MCLVZs. Our results show that the initially smooth and continuous MCLVZs were curved (profile 32°N in Figure 6a) and slightly disrupted (profile 33°N in Figure 6a) because of these repeated compressions and extensions. Interestingly, the NSGL and the Tanlu fault zone appear to be the west and east boundaries of the area (Figure 19a), respectively, where the MCLVZs are curved or disrupted. The crustal thickness between these two boundaries is thicker (Figure 5, profiles 32°−34°N in Figure 6a), most likely due to these alternating extensions and compressions. During the Cenozoic, the SCB underwent two major tectonic events: the westward subduction of the Philippine plate and the continental collision between the Indian and Eurasian plates. These two events may not have a huge influence on continental reworking in the eastern SCB and its adjacent areas because the Cenozoic magmatic rocks (≈50 Ma, area B in Figure 19d) are only distributed in the northern margin of the NCB and near the Tanlu fault zone. The concentrated distribution of Cenozoic magmatic rocks may indicate that the Cenozoic tectonic activities further promote the upwelling of hot asthenosphere materials (Yang et al., 2021) along the Tanlu fault zone. However, the subduction of the Pacific plate since the late Mesozoic, as discussed above, played a crucial role in continental reworking for the eastern SCB and adjacent areas because of its distinctly extensive influence.

By systematically analyzing the characteristics of our detailed shear-wave velocity structure and its correlation with regional tectonics, we provide a generalized description of the process of continental reworking (Figure 20) in the eastern SCB and its adjacent areas. During the long tectonically quiet period, that is, from the late Neoproterozoic to the early Phanerozoic, compositional and structural layering in the crust led to continuous low-velocity zones at mid-crustal depths. During the Triassic Indosinian movement, the SCB collided with the NCC, resulting in
the formation of the Qinling-Dabie UP and UHP metamorphic orogens and disrupting the MCLVZs beneath the Qinling-Dabie orogen. The Pacific plate has subducted beneath the eastern SCB since the late Mesozoic Yanshanian movement. On the one hand, the subduction of the Pacific plate induced asthenospheric upwelling, resulting in thinner lithospheric and crustal thicknesses in the eastern SCB and NCC. On the other hand, the subduction reactivated some ancient and lithosphere-scale suture lines or fault zones including the SJJFZ and the Tanlu fault zone, providing convenient channels for upwelling hot materials from the asthenosphere to be transported into the crust. Furthermore, upwelling hot materials likely reacted with materials from the MCLVZs and were transported toward the surface. The consequences of this are twofold. First, the continuity of the MCLVZs near these fault zones was destroyed. Second, hot and mixed upwelling materials arriving at the surface were likely to form larger amounts of late Mesozoic magmatic rocks. In addition, the alternating compressions and extensions induced by the subduction of the Pacific plate caused the crust to thicken, resulting in morphological variations in the MCLVZs and UMLVZs in the area bounded by the NSGL and the Tanlu fault zone.

6. Conclusions

We determined the shear-wave velocity structure of the crust and upper mantle in the eastern SCB and its adjacent areas according to a nonlinear joint inversion of multimodal dispersion curves. The results show that widespread LVZs exist in both the middle crust and upper mantle, reflecting the process of continental reworking. Near the Qinling-Dabie orogen, the MCLVZs seem to disrupted due to the collision between the SCB and NCC during the Triassic Indosinian event. In areas east of the Tanlu fault zone and between the SJJFZ and ZDF, the MCLVZs are not continuous and seem to be fragmented in pieces. However, the UMLVZs, which are more continuous than the MCLVZs, demonstrate upheaval near these fault zones. Overall, the crustal and lithospheric thicknesses east of the NSGL were considerably thinned. The NSGL and Tanlu fault zone bound the area north of the Qinling-Dabie orogen, where well-preserved MCLVZs show curved or disrupted morphological variations, due to the compression and extension induced by the subduction of the Pacific plate. The distribution of the MCLVZs and UMLVZs has a good correlation with the regional tectonics, that is, the distribution of heat flow and magmatic rocks. In addition to the continental collision between the SCB and the NCC during the Triassic Indosinian movement, continental reworking in the study area can be mainly attributed to the westward subduction of the Pacific plate since the late Mesozoic Yanshanian movement.

Data Availability Statement

The waveform data from the permanent stations were provided by the Data Management Centre of the China National Seismic Network at the Institute of Geophysics, China Earthquake Administration (http://www.seisdmc.ac.cn/; SEISDMC, https://doi.org/10.11998/SeisDmc/SN), China Earthquake Administration. The waveform data from the H1, R7, and W0 arrays were provided by the China Seismic Array Data Management Center at the Institute of Geophysics, China Earthquake Administration (http://www.chinarraydmc.cn/; https://doi.org/10.12001/ChinArray.Data). Other temporary data can be requested from the SinoProbe Center-China Deep Exploration Center (http://124.17.88.221/sinoprobe). The CCF and shear-wave velocity model can be found at https://doi.org/10.1785/0220210042. The reference model by Shen et al. (2016) is available at http://ciei.colorado.edu/Models. The F-J spectrogram calculation is implemented by CC-FJpy (Z. Li et al., 2021), which can be accessed from https://doi.org/10.1785/0220210042.

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