Crustal S-wave velocity structure across the northeastern South China Sea continental margin: implications for lithology and mantle exhumation

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Abstract: The northeastern margin of the South China Sea (SCS), developed from continental rifting and breakup, is usually thought of as a non-volcanic margin. However, post-spreading volcanism is massive and lower crustal high-velocity anomalies are widespread, which complicate the nature of the margin here. To better understand crustal seismic velocities, lithology, and geophysical properties, we present an S-wave velocity ($V_s$) model and a $V_p/V_s$ model for the northeastern margin by using an existing P-wave velocity ($V_p$) model as the starting model for 2-D kinematic S-wave forward ray tracing. The Mesozoic sedimentary sequence has lower $V_p/V_s$ ratios than the Cenozoic sequence; in between is a main interface of P-S conversion. Two isolated high-velocity zones (HVZ) are found in the lower crust of the continental slope, showing S-wave velocities of 4.0–4.2 km/s and $V_p/V_s$ ratios of 1.73–1.78. These values indicate a mafic composition, most likely of amphibolite facies. Also, a $V_p/V_s$ versus $V_s$ plot indicates a magnesium-rich gabro facies from post-spreading mantle melting at temperatures higher than normal. A third high-velocity zone ($V_p$: 7.0–7.8 km/s; $V_p/V_s$: 1.85–1.96), 70-km wide and 4-km thick in the continent-ocean transition zone, is most likely to be a consequence of serpentinization of upwelled upper mantle. Seismic velocity structures and also gravity anomalies indicate that mantle upwelling/serpentinization could be the most severe in the northeasternmost continent-ocean boundary of the SCS. Empirical relationships between seismic velocity and degree of serpentinization suggest that serpentine content decreases with depth, from 43% in the lower crust to 37% into the mantle.

Keywords: South China Sea; continental margin; crustal structure; converted S-wave; $V_p/V_s$ ratio; lithology; serpentinization

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1. Introduction
Passive continental margin provides critical constraints on the processes of rifting and eventual rupturing of the continental lithosphere. The northeastern continental margin of the South China Sea (SCS), which has been well-preserved, has experienced a transition from late Mesozoic subduction to Cenozoic continental rifting and subsequent seafloor spreading (Shi HS and Li CF, 2012; Taylor and Hayes, 1983) (Figure 1).

Multi-channel reflection and deep refraction seismic surveys (Gao JW et al., 2015; Savva et al., 2014; Yan P et al., 2006) show features akin to magma-poor margins in the northeastern SCS. These features include hyperextension of the continental crust, detachment faulting, and an absence of seaward dipping reflectors. Hence, it has been argued that the northeastern margin of the SCS is non-volcanic. However, massive post-spreading volcanism and a lower crustal high-velocity layer (with P-wave velocity of 7.0–7.6 km/s, higher than the normal lower crustal velocity of 6.4–6.9 km/s but lower than the uppermost mantle velocity of 8 km/s; Christensen and Mooney, 1995) are also ubiquitous (Li YQ et al., 2017; Nissen et al., 1995; Wang TK et al., 2006; Wei XD et al., 2011; Zhao MH et al., 2010). The study of high-velocity anomalies is crucial to an understanding of the process of continental rifting and breakup.

The origin of lower crustal high-velocity anomalies in the SCS remains a subject of debate. Based on a two-ship expanding spread profiling, Nissen et al. (1995) suggested that they could be due to the underplating of magmatic material during rifting. Others have ascribed them to post-spreading magmatism (Yan P et al., 2001; Zhao MH et al., 2010). Based on numerical modeling, Song TR (2016) suggested that the high-velocity layer near the continent-ocean boundary (COB) could represent the exhumation of serpentinitized upper mantle. Wan KY et al. (2017) suggested that the high-velocity layer underlying the transitional crust is related to Cenozoic decompression melting, whereas that beneath the Dongsha Rise further north is associated with the Mesozoic sub-
duction of the Paleo-Pacific plate.

These controversies are partly due to lack of lithologic information, which cannot be easily derived. The $V_p/V_s$ ratio is particularly sensitive to the porosity (Castagna et al., 1985) and quartz content of rocks (Christensen, 1996), and can be used to predict the lithology (Neidell, 1985). $V_p/V_s$ ratios can help distinguish between felsic (quartz rich) and mafic (quartz poor) crystalline rocks (Mjelde et al., 2003). Hence, acquisition of additional shear (S-) wave velocities ($V_s$) should be helpful in further constraining the elastic properties of the crust.

Based on both field seismic experiments and laboratory measurements, Holbrook et al. (1992) arrived at $V_p/V_s$ ratios for different crystalline basement compositions; they found that $V_p/V_s$ ratios vary from 1.71 in granite (felsic) and 1.78 in granodiorite, to 1.84 in gabbro (mafic). Using $V_p/V_s$ structures derived from Ocean Bottom Seismometer (OBS) data, Zhao MH et al. (2010) and Wei XD et al. (2011) reported a regional high-velocity layer of mafic composition in the northern continental margin of the SCS. However, no such study has been carried out near the COB, where the Moho depth is much shallower and exhumation of serpentinized mantle peridotite could exist.

In this study, we applied converted S-wave seismic phases to constrain the crustal lithology structure along a transect in the northeastern margin of the SCS, focusing on the composition of high velocity anomalies, possible serpentinization, and their geodynamic origins.

2. Tectonic Setting

The SCS has undergone a complete Wilson cycle in its evolutionary history. The SCS oceanic lithosphere is being subducted beneath the Luzon arc along the Manila Trench (Bautista et al., 2001). The western margin of the SCS is affected by the Ailaoshan-Red River strike-slip shear zone (Leloup et al., 1995). The southern margin of the SCS collided with Borneo (Ding WW et al., 2016; Sun Z et al., 2009; Zhao MH et al., 2010). The north margin has experienced multiple episodes of rifting (Taylor and Hayes, 1980). There are also well-preserved Mesozoic sequences that can provide rich information on the pre-rifting geological process of the SCS (Li C-F et al., 2008; Shi HS and Li C-F, 2012; Fan CY et al., 2017; Wan KY et al., 2017).

The complicated and fascinating features of the SCS have led to substantial discussions of its spreading time and rifting mechanisms. Magnetic data show a general northeast-southeast opening of the SCS during the early Oligocene-Miocene (Taylor and Hayes, 1983; Li C-F et al., 2010). Based on newly acquired deep towed magnetic anomalies and International Ocean Discovery Program Expedition 349 cores, Li C-F et al. (2014) found that the initial seafloor spreading started around 34 Ma in the northeastern SCS. At round 23.6 Ma, there was a southward ridge jump by ~20 km in the East Subbasin; coevally, the Southwest Subbasin started to spread. The seafloor spreading terminated at ~15 Ma in the East Subbasin and at ~16 Ma in the Southwest Subbasin (Li C-F et al., 2014).

Taponnier et al. (1982) and Briais et al. (1993) proposed that the collision between the Indian and the Eurasian plates resulted in a southward escape of the southeastern margin of the Tibetan Plateau, forming the large Ailaoshan-Red Sea strike-slip fault with a lateral movement of more than 500 km, which resulted in extension of the SCS. Another leading hypothesis argued that the slab pull force from the subduction of the proto-SCS could be the main driver of the opening of the SCS (Hall and Spakman, 2002; Pubellier et al., 2004).

The northern continental margin of the SCS developed a wide (100–300 km), highly thinned, and magnetically modified continental crust (Yan P et al., 2001; Song TR, 2016). One of the most dominant features of the crustal structure in the northern margin of the SCS is the Dongsha Rise (Figure 1), which falls in a NE-trending high magnetic anomaly belt (Li C-F et al., 2008). Inversion results indicate that this high magnetic anomaly is produced by 2.5–6 km thick magnetic bodies in a ~150 km wide zone (Hu DK et al., 2008), buried mostly in the upper crust at depths from ~2 to ~20 km (Li C-F et al., 2008). Extensive late Mesozoic magmatic rocks related to the subduction of the Paleo-Pacific plate are found in south China, including those responsible for the high magnetic anomaly of the Dongsha Rise (Li C-F et al., 2008; Zhou D et al., 2006).

3. Seismic Data Acquisition and Preprocessing

A marine coincidental reflection and refraction seismic survey was conducted in 2016 by the RV shiyan-2 of the South China Sea Institute of Oceanology (SCSIO), Chinese Academy of Sciences. The seismic source consisted of four large-volume Bolt air guns with a total volume of 6000 in$^3$ (98.3225 L), fired every 90 s. A total number of 15 OBSs (each with three components and one hydrophone) were deployed in ~18 km intervals from the Chaoshan Depression to the East Subbasin of the SCS (Figure 1); 14 of them returned good data (OBS06 was lost). The OBS data were sampled in 4 ms intervals. Coincidental four-channel reflection seismic data were also collected to define shallow structures (Figure 2a).

Data preprocessing consists of data format conversion, correction of shooting time and shot/OBS positions, and deconvolution (Wan XL, 2018). A band-pass filtering of corner frequencies of 3, 4, 8 and 10 Hz was applied. The converted S-wave data were extracted from the horizontal components of OBS recordings. Because the OBSs were gimballed on the seafloor, the orientation of the horizontal components is arbitrary. Therefore, an energy scanning method (Zhang L et al., 2016) was applied to find the polarization angle, and the horizontal components were then rotated into radial and transverse directions with respect to the shot line. The S-wave energy will be primarily in the radial component.

4. Seismic Velocity Modeling

4.1 Initial P-Wave Velocity Model

Wide-angle compressional (P-) wave refraction and reflection travel time tomography has recently been completed by Wan XL (2018), based on 2-D forward ray tracing (Zelt and Smith, 1992). The final $V_p$ model (Figure 2b) reveals a 3–6 km thick Mesozoic sedimentary layer with velocities of 4.3–5.3 km/s in the Chaoshan Depression; this unit pinches out in the COB below OBS11 and
Two isolated high-velocity zones (with $V_P$ of 7.0–7.5 km/s and 7.0–7.3 km/s, respectively) are found in the lower crust under the continental slope. This $V_P$ model is used as the starting model for $V_S$ modeling.

4.2 S-Wave Velocity Model
The $S$-wave modeling is based on the initial $V_P$ model geometry (Figure 2b). The $V_S$ model can be produced by fixing the depth nodes, but adjusting the Poisson’s ratio for each layer and block of the model.
the $V_P$ model, and determining the location of P-S conversion for each arrival. During this stage, the interface depths and $V_P$ were kept fixed, and only the Poisson’s ratio in each block was changed to improve the agreement between calculated and observed travel times. The $V_P/V_S$ ratio model can then be obtained by dividing $V_P$ by $V_S$.

Two types of S-wave arrivals, PPS and PSS, can be identified in the radial components. The PPS wave travels initially as a P-wave and is converted into S-wave at an interface on the way up, whereas the PSS wave is converted into S-wave on the way down, often at the seafloor or a sediment-crust interface (Au and Clowes, 1984; Tan PC et al., 2016). The PSS phases are the main S-wave phases that give direct estimates of $V_S$ in the crust.

A reduction velocity of 6 km/s was applied to identify PPS and PSS phases. Figure 3 compares the vertical and radial components of OBS13. Identification of S-waves was based on their kinematic and dynamic features such as travel time (Figure 3a–b), apparent velocity, and particle motion (Figure 3e–f). The S-wave in the radial component is slower than the P-wave in the vertical component. Three-dimensional particle motion sections show that the motion of a P arrival is predominantly vertical whereas the motion of an S arrival is mainly horizontal (Figure 3e–f).

To confirm the conversion mode, we calculated the apparent velocities of several converted phases in OBS09 (Table 1). Usually, the S-wave phases of the PPS mode have the same or smaller apparent velocities as the corresponding P-wave phases, while the S-wave phases of the PSS mode have much smaller apparent velocities. The apparent velocity of PPSs4 phases is 5.17 km/s, close to the apparent velocity of Pg phases, but for the PSSucS4 phases, the apparent velocity is 2.88 km/s, much smaller than Pg phase velocity of 5.64 km/s (Table 1). If we further compare vertical with radial components (Figure 9), we can see that the PPS4 phases have the same offset with Pg phases, but shorter travel times, while the PSSucS4 phases have smaller offsets and longer travel times. After confirming the conversion mode, the modeling exercise was then performed from sediment to mantle at each station by repeatedly changing the Poisson’s ratio and converting interfaces until all the calculated arrivals fit well with picked phases. Then, all of the S-wave phases are assigned to the suitable interfaces.

Examples of PPS and PSS arrival picks and their modeled fits show that most of the P-S conversions occurred at top Mesozoic sediment, sedimentary basement, and the Moho, where exist large velocity or impedance contrasts (Figures 4–6). The model is primarily constrained by S waves in the upper (PSSuc phase) and lower (PSSIc phase) crust. Table 2 summarizes numbers of picks, picking error, root-mean-square (RMS) misfit, $\chi^2$ values (normalized misfit between calculated and observed travel time), and conversion interfaces.

The uncertainty in the S-wave modeling largely depends on the uncertainty in the P-wave model (Digranes et al., 1998). High-quality and numerous P-wave arrivals provide good structural controls on our S-wave model. The uncertainties of interpreted travel-

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**Figure 3.** Vertical (a) and radial (b) components of OBS13 reduced by 6 km/s (after 3–10 Hz band-pass filtering). (c) and (d) show the single-trace waveform of Trace 1123 in the vertical and radial component, respectively. (e) and (f) show the particle displacements of Trace 1123 in the vertical (4.44–4.74 s) and the radial (6.10–6.50 s) component sections, respectively.
time curves are generally within 50–160 ms. During the simulation, we control the limit range of the Poisson’s ratios (± 0.01–0.04) by keeping the mismatch between calculated and observed travel time within the picking uncertainty. With fixed $V_P$, variations of ± 0.01–0.04 in the Poisson’s ratio lead to uncertainties of ± 0.01–0.2 km/s and ± 0.03–0.16 in calculated $V_S$ and $V_P/V_S$ ratios, respectively.

A statistical measure of fit in chi-squared ($\chi^2$) close to one is used to evaluate the mismatch between observed and calculated travel times (Zelt and Smith, 1992). After the S-wave modeling, the fitting RMS errors are less than 200 ms and $\chi^2$ are less than 2 for most of the S-wave phases (Table 2). The ray coverage map (Figure 7c) of all observed arrival locations further shows that the crust and most interfaces are well covered by 10–40 times, indicating high reliability and resolution of our S-wave model.

5. S-Wave Velocity Structure and Implications on Lithology

In this section, we infer crustal lithology from P- and S-wave velocities and the $V_P/V_S$ ratio model (Figure 7b).

| Offset (km) | Vertical component | Radial component |
|-------------|--------------------|-----------------|
|             | Reduced time (s)   | Apparent velocity (km/s) | Phases | Reduced time (s) | Apparent velocity (km/s) | Phases |
| –62 to –8   | 2.6–3.5            | 5.64            | Pg      | 4–5              | 5.17            | PPSs4 |
|             | 6.7–9.6            | 2.88            | PSSs4   |
| 11–56       | 3.6–4.1            | 5.11            | Pg      | 4.8–5.3          | 4.67            | PSSs4 |
|             | 8.7–10.2           | 2.58            | PSSsuc4 |

Table 1. Apparent velocities and phases in the record of OBS09

Figure 4. Seismic records, interpretation, and ray tracing of OBS09. (a) Comparison of interpreted (colored) PPS and PSS modes and calculated (solid) travel time curves for OBS09. (b) Ray-paths through the model for the calculated PPS arrivals. (c) Same as (b) for PSS arrivals. Solid lines denote rays of P-waves and dotted lines denote rays of S-waves.
5.1 Sedimentary Layers

Our models show that the $V_S$ increases while the $V_P/V_S$ ratio decreases with depth through the sedimentary layers (Figure 7a–b). There are three Cenozoic sedimentary layers. The uppermost layer is characterized by very high $V_P/V_S$ (4.2–6.6) and low $V_S$ (0.3–0.45 km/s), particularly in the lower slope and oceanic basin. The other two Cenozoic sedimentary layers have $V_S$ of 0.6–1.5 km/s and $V_P/V_S$ ratios of 2.0–3.2. The increased confining pressure with depth reduces porosity and closes cracks, thereby decreasing $V_P/V_S$ ratios (Christensen and Mooney, 1995; Sato and Ito, 2001). We thus attribute the observed vertical decrease in $V_P/V_S$ ratio to increased compaction and consolidation of the sedimentary rocks.

The Mesozoic sedimentary layer at the Chaoshan Depression along our profile has a maximum thickness of 5 km, $V_S$ of 2.4–3.1 km/s, and $V_P/V_S$ ratios of 1.74–1.78; the layer pinches out southwards to the COB. These results of the Mesozoic sedimentary lay-
Table 2. Numbers of picks, picking errors, RMS misfits, $\chi^2$ values, conversion mode and conversion interface (P-S interface) for phases identified along profile OBS2016-2 in the S-wave modeling.

| OBS | Mode | Phase   | No. of rays | Uncertainty | RMS     | $\chi^2$ | P-S interface       |
|-----|------|---------|-------------|-------------|---------|----------|---------------------|
| 1   | PPS  | PPSs4s4 | 141         | 0.05        | 0.048   | 0.915    | Upper Mesozoic      |
|     |      | PPSs4   | 337         | 0.08        | 0.049   | 0.974    | Upper Mesozoic      |
|     |      | PPSb    | 292         | 0.07        | 0.059   | 1.394    | Basement            |
|     |      | PmSm    | 153         | 0.09        | 0.064   | 0.504    | Moho                |
| 2   | PPS  | PPSs4s4 | 181         | 0.16        | 0.078   | 2.421    | Seafloor            |
|     |      | PPSuc   | 181         | 0.16        | 0.078   | 2.421    | Seafloor            |
|     |      | PSSlc   | 157         | 0.12        | 0.056   | 1.269    | Upper Mesozoic      |
|     |      | PmSm    | 107         | 0.12        | 0.092   | 1.049    | Moho                |

Figure 6. Seismic records, interpretation, and ray tracing of OBS14. (a) Vertical component of OBS14. (b) Radial component of OBS14 and interpreted (colored) PPS and PSS modes and calculated (solid) travel time curves are shown. (c) Ray-paths through the model for the calculated PPS and PSS arrivals.
Continued from Table 2

| OBS | Mode | Phase | No. of rays | Uncertainty | RMS | $\chi^2$ | P-S interface |
|-----|------|-------|-------------|-------------|-----|--------|---------------|
| 3   | PPS  | PPSs4s4 | 111         | 0.06        | 0.046 | 0.862  | Upper Mesozoic |
|     |      | PPSs4  | 521         | 0.07        | 0.058 | 1.357  | Upper Mesozoic |
|     |      | PPSb   | 574         | 0.08        | 0.06  | 1.456  | Basement       |
|     |      | PPSnm  | 295         | 0.12        | 0.11  | 1.095  | Moho           |
|     | PSS  | PSSuc  | 111         | 0.17        | 0.055 | 1.218  | Seafloor       |
|     |      | PSSLc  | 80          | 0.16        | 0.027 | 0.295  | Seafloor       |
|     |      | PSSLc2 | 76          | 0.05        | 0.067 | 1.824  | Upper Mesozoic |
|     |      | PPSuc  | 171         | 0.12        | 0.062 | 1.548  | Seafloor       |
|     |      | PSSLc2 | 121         | 0.11        | 0.06  | 1.453  | Sediment 2     |
|     |      | PSSLc  | 191         | 0.12        | 0.068 | 1.844  | Sediment 2     |
|     |      | PSSuc  | 103         | 0.16        | 0.056 | 1.284  | Seafloor       |
|     |      | PPSuc2 | 84          | 0.15        | 0.051 | 1.073  | Upper Mesozoic |
|     |      | PSSLc2 | 475         | 0.14        | 0.052 | 0.849  | Upper Mesozoic |
|     |      | PSSLc  | 185         | 0.16        | 0.075 | 1.143  | Basement       |
|     |      | PnSuc  | 278         | 0.12        | 0.05  | 0.311  | Moho           |
|     |      | PSSLc2 | 155         | 0.12        | 0.051 | 1.038  | Seafloor       |
|     |      | PSSLc  | 49          | 0.15        | 0.051 | 1.048  | Sediment 2     |
|     |      | PSSLc2 | 175         | 0.16        | 0.053 | 1.147  | Sediment 2     |
|     |      | PSSuc  | 100         | 0.12        | 0.097 | 1.172  | Upper crust    |
|     |      | PPSuc2 | 204         | 0.14        | 0.07  | 0.609  | Moho           |
|     |      | PSSLc  | 657         | 0.09        | 0.052 | 1.101  | Basement       |
|     |      | PSSLc2 | 84          | 0.14        | 0.051 | 1.265  | Upper Mesozoic |
|     |      | PSSLc  | 475         | 0.14        | 0.052 | 1.393  | Upper Mesozoic |
|     |      | PSSLc2 | 203         | 0.16        | 0.107 | 1.366  | Upper Mesozoic |
|     |      | PmSm   | 71          | 0.21        | 0.058 | 0.418  | Moho           |
|     | PSS  | PSSLc2 | 95          | 0.12        | 0.057 | 1.496  | Upper Mesozoic |
|     |      | PSSLc  | 373         | 0.15        | 0.043 | 1.453  | Upper Mesozoic |
|     |      | PSSLc2 | 107         | 0.21        | 0.064 | 1.439  | Upper Mesozoic |
|     |      | PSSLc  | 463         | 0.06        | 0.057 | 1.324  | Upper Mesozoic |
|     |      | PSSLc2 | 466         | 0.06        | 0.063 | 1.611  | Basement       |
|     |      | PSSLc  | 40          | 0.09        | 0.06  | 0.459  | Basement       |
|     |      | PSSLc2 | 164         | 0.1         | 0.053 | 1.131  | Moho           |
|     |      | PSSLc  | 95          | 0.12        | 0.057 | 1.496  | Upper Mesozoic |
|     |      | PSSLc2 | 272         | 0.15        | 0.059 | 1.274  | Upper Mesozoic |
|     |      | PSSLc  | 71          | 0.2         | 0.075 | 0.707  | Upper Mesozoic |
er are similar to those along the OBS2006-3 profile (Wei XD et al., 2011), which reported a maximum thickness of 8 km, $V_S$ of 2.6–3.0 km/s and $V_P/V_S$ ratios of 1.74–1.80. The good consolidation of Mesozoic sediments is revealed by the relatively low $V_P/V_S$ ratio and high $V_S$, which also contribute to the P-S conversions.

### 5.2 Upper Continental Slope (Model Distance of 0–130 km)

The $V_S$ in the upper crust of the upper slope is 3.1–3.6 km/s, and the estimated $V_P/V_S$ ratio is 1.75–1.77, characterizing a felsic granite composition typical of continental crust, in good agreement

| OBS | Mode | Phase | No. of rays | Uncertainty | RMS | $\chi^2$ | P-S interface |
|-----|------|-------|-------------|-------------|-----|---------|--------------|
| 10  | PPS  | PPSs4 | 303         | 0.05        | 0.06 | 1.422   | Upper Mesozoic |
|     | PPS  | PnSs4 | 47          | 0.05        | 0.059| 0.439   | Upper Mesozoic |
|     | PPS  | PPSb  | 418         | 0.08        | 0.062| 1.545   | Basement      |
|     | PPS  | PnSb  | 75          | 0.09        | 0.062| 0.482   | Basement      |
|     | PPS  | PmSm  | 169         | 0.1         | 0.041| 0.68   | Moho         |
|     | PSS  | PSSs4 | 88          | 0.1         | 0.053| 1.141   | Upper Mesozoic |
|     | PSS  | PSSlcs4 | 142       | 0.14       | 0.053| 1.148   | Upper Mesozoic |
|     | PSS  | PSSlcb | 37         | 0.15       | 0.05 | 1.017   | Basement      |
| 11  | PPS  | PPSs4 | 244         | 0.05        | 0.039| 0.626   | Upper Mesozoic |
|     | PPS  | PPSb  | 283         | 0.07        | 0.038| 0.582   | Basement      |
|     | PPS  | PmSm  | 115         | 0.09        | 0.054| 1.17   | Moho         |
|     | PSS  | PSSs4 | 60          | 0.1         | 0.058| 1.347   | Upper Mesozoic |
|     | PSS  | PSSlcs4 | 178       | 0.1         | 0.046| 0.858   | Upper Mesozoic |
|     | PSS  | PSSlcb | 70         | 0.15       | 0.042| 0.731   | Basement      |
| 12  | PPS  | PPSs4 | 251         | 0.05        | 0.062| 1.529   | Upper Mesozoic |
|     | PPS  | PPSb  | 248         | 0.07        | 0.065| 1.713   | Basement      |
|     | PPS  | PmSm  | 179         | 0.09        | 0.058| 1.354   | Moho         |
|     | PSS  | PSSs4 | 98          | 0.1         | 0.04  | 0.653   | Upper Mesozoic |
|     | PSS  | PSSn  | 104         | 0.15        | 0.081| 0.815   | Seafloor      |
|     | PSS  | PSSlcb | 76         | 0.1         | 0.05  | 0.94   | Basement      |
|     | PSS  | PSSnb | 80          | 0.1         | 0.094| 1.094   | Basement      |
| 13  | PPS  | PPSb  | 127         | 0.05        | 0.057| 1.322   | Basement      |
|     | PPS  | PnSb  | 123         | 0.09        | 0.064| 0.507   | Basement      |
|     | PPS  | PnSm  | 90          | 0.12        | 0.087| 0.944   | Moho         |
|     | PSS  | PSSlcs4 | 137      | 0.11       | 0.06  | 1.452   | Upper Mesozoic |
|     | PSS  | PSSn  | 55          | 0.15        | 0.048| 0.291   | Seafloor      |
|     | PSS  | PSSlcb | 82         | 0.15       | 0.066| 1.748   | Basement      |
| 14  | PPS  | PPSb  | 141         | 0.05        | 0.06  | 1.47    | Basement      |
|     | PPS  | PnSb  | 131         | 0.09        | 0.04  | 0.197   | Basement      |
|     | PPS  | PnSm  | 154         | 0.12        | 0.046| 0.265   | Moho         |
|     | PSS  | PSSlcb | 135       | 0.1         | 0.063| 1.633   | Basement      |
|     | PSS  | PSSn  | 84          | 0.15        | 0.111| 1.531   | Seafloor      |
| 15  | PPS  | PPSb  | 170         | 0.05        | 0.037| 0.559   | Basement      |
|     | PPS  | PnSm  | 125         | 0.1         | 0.062| 0.481   | Moho         |
|     | PSS  | PSSb  | 98          | 0.15        | 0.068| 1.883   | Basement      |
|     | PSS  | PSSnb | 59          | 0.15        | 0.064| 1.121   | Basement      |
with the average $V_P/V_S$ ratio of 1.768 obtained by Christensen and Mooney (1995). The $V_S$ in the lower crust is 3.6–4.2 km/s and $V_P/V_S$ ratios are of 1.73–1.78, suggesting a wide spectrum of rock composition from felsic to mafic, based on the results summarized in Figure 8.

5.3 Lower Continental Slope (Model Distance of 130–233 km)
This section includes the continent-ocean transition zone, which is highly extended during rifting (Song TR, 2016). The upper crustal $V_S$ is 3.0–3.6 km/s and $V_P/V_S$ ratios are 1.74–1.76, indicating a thinned crust and felsic composition similar to the results of Zhao MH et al. (2010). $V_S$ is 3.6–4.1 km/s and $V_P/V_S$ ratios are 1.75–1.81 in the lower crust, also suggesting a rock composition from felsic to mafic (Figure 8).

5.4 Oceanic Basin (Model Distance of 233–320 km)
The Moho depth near the COB varies from ~11 to ~13 km below sea level, somewhat shallower than in the surrounding areas, probably due to the mantle upwelling confirmed by previous reflection seismic interpretations (Song TR, 2016). The $V_S$ of oceanic crust is 2.7–3.8 km/s, and $V_P/V_S$ ratios are 1.80–1.89. These high $V_P/V_S$ ratios indicate a basic-ultrabasic composition (Wei XD et al., 2011).

6. Discussion
6.1 The Nature of High Velocity Zones
Our S-wave velocity model with quantifiable uncertainty allows the interpretable spatial variations of the $V_P/V_S$ ratio to be determined. Plots of the $V_P/V_S$ ratio versus $V_P$ give good geological discrimination (Eccles et al., 2009). Figure 8 shows such a plot for the average HVZ properties (defined by $V_P > 6.9$ km/s and $V_S > 3.9$ km/s) at 10 km intervals in the continent slope. The average lower crustal temperature and pressure are estimated at each position. Curie-point depths derived from magnetic anomalies (Li C-F and Wang J, 2016) are used to calculate crustal temperature assuming a geothermal gradient of 25 °C/km (Stein and Stein, 1992). Densities $\rho$ (Table 3) of sedimentary layers, upper and lower crust,
Poisson’s ratio
0.24
0.28
0.31
Increased 
Mg
1.90
1.85
1.80
1.75
1.70
1.65 VP/VS
5.5 6.0 6.5 7.0 7.5 8.0
GABBRO
10%Fo 0%Fa
BASALT
DIABASE
GABBRO
GRANULITE
GNEISS
GRANITE
HVZ_2 DUNITE
AMPHIBOLITE
HVZ_1

Table 3. Density calculation from seismic velocity for model layers

|                     | Sediment 1 | Sediment 2 | Sediment 3 | Sediment 4 |
|---------------------|------------|------------|------------|------------|
| Gardner’s Rule (1974) | ρ=1.74×Vp^{0.25} | 2.03        | 2.21        | 2.36        | 2.57        |
| Christensen & Mooney (1995) 10 km | ρ=0.541+0.3601×Vp | Upper crust | 2.67        |
| Christensen & Mooney (1995) 20 km | ρ=0.444+0.3754×Vp | Lower crust | 3.05        |

The magmatic activities were not strong before and during the SCS rifting and seafloor spreading and no high-velocity anomaly was found onland South China (Yin ZX et al., 1999). Post-spreading magmatic activities became more active, and more seamounts developed parallel to the faults in this stage (Ru K and Pigott, 1986; Song XX et al., 2017). Therefore, we infer that the high-velocity anomalies may have been formed by magmatic underplating after the cessation of seafloor spreading.

The effect of magnesium on the Vp/Vs ratios has been explored by calculating the trend of properties for an olivine gabbro from the Oman Ophiolite (Browning, 1984), with systematic substitution of magnesium for iron within the olivine (fayalite to forsterite) using the approach of Hacker and Abers (2004). This trend shows that, with increasing magnesium, the Vp increases but the Vp/Vs ratio decreases, that is, the properties trend towards those of dunite (Figure 8). This indicates that the high velocity anomalies could also be caused by magnesium-rich gabbro produced from mantle melting at temperatures higher than normal. We further argue that the anomalies could be formed after the spreading stopped because significant episodic extensions (≈15.2 Ma and 5.2 Ma) may have thinned the lithosphere, leading to high heat flow (He LJ et al., 1998; Shi XB et al., 2003). Thermal contraction in the oceanic lithosphere may also trigger extension and decompressive melting along weak zones in the continental slope and the oceanic basin (Song XX et al., 2017). As a consequence, high-temperature melting would increase the Mg content at the expense of Fe (White et al., 1992) and reduced the Vp/Vs ratio.

6.2 Serpentinitization in the Lower Crust and Upper Mantle

Serpentinitization is formed by the hydration of olivine-rich ultramafic rocks at relatively low temperatures, which significantly lowers the compressional wave velocity of the primary peridotite (from 8 to 7.2–7.6 km/s), and increases magnetic susceptibility by nearly two orders of magnitude (Evans et al., 2013). Serpentinitization can usually be found in magma-poor continental margins, where the continent-ocean transition zone is highly stretched and seawater can penetrate deep into mantle through cracks and fractures to form serpentinite. In the West Iberian margin, a lower crust layer under a thinned transition zone is found up to 4 km
thick with $V_p$ of 7.3–7.9 km/s (Chian et al., 1999; Dean et al., 2000), which represents mantle peridotite with mean bulk serpentinization of < 25% from the seawater influx during rifting along large numbers of faults (Dean et al., 2000).

In our previous $V_p$ inversion model (Wan XL, 2018), we also observed a high-velocity anomaly ($V_p$ 7.0–7.8 km/s) with a maximum thickness of 4 km in the lower crust near the COB (from 210 to 280 km in offset) (Figure 9b). Analysis of 1-D velocity-depth profiles at this high-velocity zone indicated mantle/lower crust exhumation or serpentinization (Wan XL, 2018). To further examine the nature of this high-velocity anomaly, we analyze the new $V_p/V_s$ ratios (Figure 9c) based on the $V_p$ from tomographic inversion (Wan XL, 2018) and our S-wave velocity.

The HVZ properties near the COB (from 210 to 280 km) are plotted on a $V_p/V_s$ versus $V_p$ diagram (Figure 10). The modelled properties are compared to laboratory results to test different hypotheses, e.g., whether the HVZ comprises gabbros, peridotites, or serpentinites. These measurements showed that the $V_p/V_s$ ratios of HVZ samples range from 1.85 to 1.96, much higher than those of most mafic rocks. The plotted points distribute predominantly around serpentinized peridotites, but fewer around gabbro.

We then further draw a $V_p$ versus $V_p/V_s$ plot of different degrees of serpentinization (Figure 11) to see how the HVZ relates to the serpentinization. Physical properties vary linearly with the degree of serpentinization, with $V_p$ decreasing from 8 km/s in peridotite to 4.8 km/s in serpentine (Christensen, 1966). Our observed HVZ velocities (7.0–7.8 km/s) and $V_p/V_s$ ratios (1.85–1.96) fit well with the trend; most points are scattered around the partially serpentinized peridotite samples with 37–43% serpentinization. These data argue for serpentinization of uppermost mantle near the COB.

In general, marine free-air gravity anomalies vary with the bathymetry. However, we observe a high gravity anomaly (Figure 9a) near the flat COB, correlating spatially to our identified zone with lower crust high velocity anomalies, mantle upwelling, and serpentinization near the COB. On the map view, this high gravity anomaly extends further east in a narrow zone parallel to the COB (Figure 12), indicating that this feature of mantle upwelling is widespread and is potentially more severe in the northeastern-most SCS than elsewhere. From reflection seismic interpretations, Song TR (2016) already identified this COB feature of mantle upwelling ~200 km west to this line, but no appreciable high gravity anomalies were found there to be associated with this mantle upwelling (Figure 12). We suspect that mantle upwelling in other places was either too narrow or too premature to induce noticeable gravity anomalies.
activities in the northeastern SCS continental margin were not strong during the incipient rifting and spreading (66–26 Ma) (e.g., Song XX et al., 2017; Fan CY et al., 2017; Xia SH et al., 2017). Based on magnetic data analysis, Li C-F et al. (2014) found that the full spreading rate of the East Subbasin was higher (~70 km/) at the beginning of seafloor spreading from ~32 to ~29 Ma, and dropped to ~25 km/ on average from ~29 to ~26 Ma. The slow spreading rates support a magma-starved northeastern SCS continental margin prone to serpentinization, since many titled faults have been revealed in the thinned continental-ocean transition zone and seawater can penetrate into the already upwelled mantle.

7. Conclusion

Converted S-wave arrival time has been inverted to produce crustal S-wave velocity model ($V_s$) with good estimates of model uncertainty across the northeastern continental margin of the South China Sea (SCS). The velocities are constrained by high-quality, densely-sampled seismic phases recorded by fourteen four-component ocean bottom seismometers (OBS). The crustal

Figure 10. $V_p$ versus $V_p/V_s$ plot for the third high velocity anomaly (HVZ_3) at the COB corrected to a pressure of 255 MPa and a temperature of 316 °C. Lower crust high velocity samplings at 10 km intervals from 210 to 280 km in the horizontal distance are shown by red circles. The properties of partially serpentinized peridotite (PSP) Samples 3 and 4 from Christensen (1966) have been corrected to the pressure of 255 MPa.

Figure 11. $V_p$ versus $V_p/V_s$ plot for the third high velocity anomaly (HVZ_3) at the COB on the diagram for different grades of partially serpentinized peridotites (PSP) corrected to a pressure of 255 MPa (Christensen, 1966).

Figure 12. Free-air gravity anomaly (V21.1 from Sandwell and Smith, 2009; Sandwell et al., 2013) map of the Northern South China Sea. Black line shows the position of our seismic profile with interpreted zone of mantle upwelling/serpentinization in dashed green and white colors. The high gravity anomaly zone surrounded by the blue dashed line represents the interpreted areal distribution of the mantle upwelling/serpentinization subparallel to the COB of Li C-F and Song TR (2012) and Song TR (2016).
scale $V_p$ and $V_S/V_p$ ratios provide valuable information on mineralogy and lithology, particularly of lower crust high velocity anomalies.

In the continental slope, we find two isolated high-velocity zones in the lower crust. The $V_p/V_S$ versus $V_p$ plot reveals a mafic composition, more likely to be of amphibolite rather than gabbro. Based on the effect of magnesium on the $V_p/V_S$ ratios, we further infer that the HVZs could be caused by magnesium-rich gabbro from post-spreading mantle melting at temperatures higher than normal.

The third high-velocity zone of ~70 km wide located between thinned continental crust and oceanic crust is identified with P-wave velocities of 7.0–7.8 km/s and $V_p/V_S$ ratios of 1.85–1.96. These observed velocities and $V_p/V_S$ values suggest serpentinization to an extent of 37–43% in a layer up to 4 km thick beneath the top oceanic crust. High free-air gravity anomalies over a relatively flat COB further support the imaged high lower crustal velocity anomalies, mantle upwelling, and serpentinization, which appear to be most evident in the northeasternmost SCS. Seawater can penetrate down into the uppermost mantle through fractures in the thinned oceanic crust in the COB of the SCS.

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