Seismic evidence of the Hainan mantle plume by receiver function analysis in southern China

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Abstract The Lei-Qiong region is the largest igneous province in southern China and may be a surface expression of a mantle plume beneath the region (the Hainan mantle plume). To investigate the existence of the Hainan mantle plume, we used P-to-S receiver function to image the major seismic discontinuities beneath this region with a regional dense broadband array. We found that the Moho discontinuity beneath the Leizhou Peninsula, mostly covered by Cenozoic basaltic outcrops, is 10–15 km deeper compared to the adjacent region of Eurasian continental margin, showing a thickened local crust by upwelling mantle materials. Additionally, the imaged 410- and 660-km discontinuities suggest a thinner-than-normal mantle transition zone beneath the region, implying that hot materials penetrate through the transition zone from the lower mantle. Both seismic evidences support the existence of the mantle plume, which might be 170–200°C hotter than the surrounding mantle.

1. Introduction

A mantle plume, bringing up hot materials from the deep mantle, has been successfully used to explain intraplate magmatism in various places with hot spots, such as Hawaii [Li et al., 2000; Wolfe et al., 2009; Cao et al., 2011; Ryichert et al., 2013] and Yellowstone [Duke and Sheehan, 1997; Schutt and Dukeer, 2008; Huang et al., 2015]. The Lei-Qiong region, including the Leizhou Peninsula (abbreviated as Lei) and the northern margin of Hainan Island (abbreviated as Qiong), has the largest igneous province in southern China, also invoking the possibility of a related mantle plume. This region is covered by late Cenozoic basalts (dark red areas in Figure 1a), which are obviously different from the Mesozoic andesitic-felsic igneous basement exposed on the South China Block and the main part of Hainan Island. The volcanic activities in the Lei-Qiong region started in the late Oligocene, then reached their peak in the Pleistocene, and ended in the Holocene [Zhou and Armstrong, 1982].

The geochemical signatures of the Lei-Qiong basalts lay between the categories of ocean island basalt and enriched mid-ocean ridge basalt, indicating an Enriched Mantle II type source, which can be a mixture of upwelling mantle materials with terrestrial sediments [Zhou and Armstrong, 1982; Tu et al., 1991]. Without geophysical evidence of a mantle plume, Tu et al. [1991] explained the magma source as consequences of lithospheric stretching associated with the opening of the South China Sea. Later, global S wave tomography by Lebedev et al. [2000] proposed a mantle plume beneath the Lei-Qiong region (the Hainan mantle plume) as an alternative explanation for the intraplate magmatism. Since then, both global [Lebedev and Nolet, 2003; Montelli et al., 2004] and regional seismic results [Huang and Zhao, 2006; Lei et al., 2009; Wang and Huang, 2012] showed a plume-shaped low-velocity zone (LVZ) from the surface down to at least 1000 km deep with a thinned mantle transition zone (MTZ) east of the Lei-Qiong region. However, the existence of the Hainan plume is still not definite due to the limited data and resolution. Moreover, the dynamic connecting mechanisms of the LVZ with surface volcanism remain unclear, as most of previous local studies were confined to Hainan Island, which is south of the proposed location of the plume [Lei et al., 2009; Wang and Huang, 2012].

We performed the P-to-S receiver function imaging to investigate the detailed structure of the crust and upper mantle beneath the Lei-Qiong region with the PKU Le-Muse Array and 19 permanent stations of the Hainan Seismological Bureau, China Earthquake Administration (Figure 1). From the results, we found that the Moho discontinuity beneath the Leizhou Peninsula is 10–15 km deeper compared to the adjacent region, and the 410- and 660-km discontinuities suggest a thinner-than-normal mantle transition zone beneath the...
The results support the existence of the mantle plume and suggest an abnormal thermal structure with the plume 170–200°C hotter than the surrounding mantle.

2. Data and Methods

2.1. Data Collection

Seismic data used in this study were collected from the temporary PKU Le-Muse Array and 19 permanent stations on Hainan Island (Figures 1a and S1a in the supporting information). The PKU Le-Muse Array was operated from February 2008 to April 2009, consisting of 13 seismic stations spaced 15–20 km apart over the entire Leizhou Peninsula and adjacent areas of the South China Block (red triangles in Figure 1a). Each station contained a Guralp CMG-3ESPC or CMG-3ESP seismometer and a Reftek 130-01 data recorder. In addition, we used data recorded over 6 months (April to August 2009 and January 2010) from 19 stations of the Hainan Seismological Bureau (blue triangles in Figure 1a). Compared to the PKU array, these Hainan stations are located in quieter places so that the well-logged 6 month data with relatively higher signal-to-noise ratios are sufficient to constrain the southern end of the Lei-Qiong region.

From the U.S. Geological Survey Preliminary Determination of Epicenters catalog, we selected 115 mb > 6 earthquakes to perform analysis for the PKU Le-Muse Array data, and 122 mb > 5.5 earthquakes for the Hainan permanent station data (Figures S1b and S1c). The raw seismogram of each event was windowed between –10 s and 160 s relative to the P wave arrival, downsampled to 50 samples per second, and then band-pass filtered between 0.05 and 3 Hz (fourth-order Butterworth, zero-phase shift) to improve the signal-to-noise ratio.

2.2. Receiver Function Analysis

We computed radial receiver functions using time domain-iterative deconvolution technique [Ligorría and Ammon, 1999]. This technique has two major advantages compared to other frequency-domain
deconvolution methods. First, the method can constrain spectral shape at long periods. Second, it does not need an objective water-level parameter to stabilize the deconvolution. A low-pass Gaussian filter \( G(\omega) = \exp(-\alpha^2/4\omega^2) \) was applied to remove high-frequency noise from the receiver functions, where \( \omega \) is the frequency and \( \alpha \) controls the bandwidth of the filter. For the Moho discontinuity, \( \alpha \) is chosen as 2.5, which is equivalent to a \( \sim 1 \) Hz low-pass filter. These receiver function stacks for each station are shown in Figure S2. For the 410- and 660-km discontinuities, we chose \( \alpha \) as 1.5 and 1.0, respectively, to amplify the low-frequency signals converted at greater depths.

After obtaining the receiver functions, we then inverted the \( S \) wave velocity structure beneath each station with a time-domain waveform inversion technique [Ammon et al., 1990] (Text S1). Since this inversion is highly dependent upon the starting model, we started with local seismic refraction results in this region [Jia et al., 2006; Xu et al., 2008] for the crustal structure and combined them with the IASP91 model [Kennett and Engdahl, 1991] from the Moho to 160 km in depth. The velocity model was parameterized as uniform layers with a thickness of 2 km above 40 km depth and 20 km below. Although the velocity models are indicative of the complexities of the Lei-Qiong crust, the nonuniqueness of this inversion [Ammon et al., 1990] prevents the inverted velocity results from being used as more than reference models for migration and stacking in the following step. Therefore, we only inverted the \( S \) wave velocity structure beneath station QIZ as a representative of Hainan Island, because this island is a relatively uniform unit with a roughly consistent velocity structure of the crust [Li et al., 2008].

Finally, we applied a 2-D regularized Kirchhoff prestack migration technique [Wilson and Aster, 2005] to image the Moho discontinuity. This method maps seismic signal to its true subsurface location while collapsing diffraction artifacts from lateral heterogeneity [e.g., Abers, 1998]. The migration was performed as an inverse problem with a second-order Tikhonov regularization to improve resolution and to eliminate migration artifacts [Nemeth et al., 1999]. Compared to other methods, Kirchhoff prestack migration does not assume a horizontal subsurface but treats scattering from 2-D structures [Cheng et al., 2016]. This technique has been successfully implemented to map complex tectonics [e.g., Wilson et al., 2005; Yue et al., 2012] and is suitable for the Lei-Qiong region, where we are expecting large variations of the Moho topography.

The migration was performed with a numerical aperture of 0.4 (incidence angle smaller than 22\(^\circ\)), so that the illuminating size is comparable to the Fresnel zone width at the Moho depth. Traveltimes relative to \( P \) were calculated for each station based on the velocity models from the previous step. Although the regularized migration promotes a more laterally coherent image than standard algorithms, migration artifacts (“smiles”) are still present in places sampled by sparse data. We thus only show the well-illuminated areas and leave the rest masked. The same approach was applied to image the 410- and 660-km discontinuities. The migration aperture was adjusted to be wider according to the Fresnel zone width at greater depths. We did not explore the possible 520-km discontinuity because it is difficult to distinguish the weak \( P520s \) signals from the sidelobes of the 410- or 660-km signals [Lawrence and Shearer, 2006]. Furthermore, the ambiguous mineralogical implications at 500–550 km depths [e.g., Saikia et al., 2008] also limit its capability for indicating upper mantle thermal variations.

### 3. Results

#### 3.1. 1-D S Wave Velocity Models

The final \( S \) wave velocity models (Figures 1b and S3) reveal strong variability of crustal structure throughout the study region. Beneath the Leizhou Peninsula (stations LP01-04 and LP10-LP13), the upper crust \( S \) wave velocities (~4.5 km/s at 10 km depth) are significantly higher than typical continental \( S \) wave velocities (e.g., 3.75 km/s from the IASP91 model [Kennett and Engdahl, 1991]). A clear LVZ is also present in the lower crust through the Leizhou Peninsula. The Moho is not well resolved at these stations possibly due to the lack of appropriate starting models or complex crustal structures. In contrast, the Moho of the South China Block (stations LP06-LP09) and Hainan Island (station QIZ) can be clearly identified at about 25 km depth. In between, results at station LP05 show a mixed velocity structure with a clear Moho and a moderate LVZ in the lower crust.

The nonuniqueness of the \( S \) wave velocity inversion [Ammon et al., 1990] prohibits further detailed interpretations of these velocity models. A better starting model may improve the inversion results for the Leizhou
Peninsula. However, the strong variations suggest that the complex crustal velocities of the Leizhou Peninsula cannot be resolved solely by receiver functions. Therefore, we hesitate to interpret these 1-D velocity models and move on to use them as reference models for migration and stacking.

3.2. Migrated and Stacked Imaging

The complex Moho topography is examined along four cross sections that pass piercing points immediately beneath stations. The Moho can be explicitly identified as positive-polarity phases in the stacked images through all four cross sections (Figure 2), despite the illumination viabilities along the cross sections (Figure S5). The main cross section A-A', striking almost N-S across the study region, unambiguously shows that the Moho depth increases to as deep as 35 km beneath the Leizhou Peninsula. The northern (the South China Block) and southern (Hainan Island) parts of the cross section share similar crustal thickness of about 22 km. The E-W striking cross section B-B' across Hainan Island also suggests that the Moho depth remains nearly unchanged at 20–25 km. The crustal thickness obviously increases eastward (from 20 km beneath LP05 and LP12 to about 25 km beneath LP13) and southward (from 20 km beneath LP05 and LP12 to 30 km beneath LP03 and LP10), despite of the large station spacing along C-C' and D-D'. An indistinct and deeper Moho at 45 km depth is imaged beneath LP11 in the eastern peninsula, corresponding to the complex 1-D velocity structure resolved at the station (Figure S4).

Two parallel cross sections E1-E1' and E2-E2' show the 410- and 660-km discontinuities topography east of the array because of their piercing points (Figure 3). The depth of the 410-km discontinuity varies from 420 km to 450 km with an average value of about 425 km (E1-E1'), whereas the 660-km discontinuity appears to be smoother at about 650 km depth (E2-E2'). Positive-polarity phases at depths of about 470 and 740 km are sidelobes of the 410- and 660-km discontinuities signals. More importantly, the thickness of the MTZ is about 225 km, which is in agreement with receiver function results from a single station QIZ [Yang and Zhou, 2001], but 25 km thinner than the IASP91 model [Kennett and Engdahl, 1991]. The thickness of the
MTZ is only sensitive to the velocity structure between 410 and 660 km and is independent of the velocity models of the crust and upper mantle. Therefore, the resolved migration images strongly suggest a thinner-than-normal mantle transition zone east of the Lei-Qiong region.

4. Discussion

4.1. Crustal Thickening

The stacked images suggest that the crust beneath the Leizhou Peninsula has been thickened by 10–15 km, and probably even more beneath the eastern peninsula (Figure 2b). In contrast, the crustal thickness of the South China Block and the main part of Hainan Island remains 20–25 km, consistent with a typical value for the northern continental margin of the South China Sea [Pin et al., 2001]. The thickened crust collocates with outcropped Cenozoic basalts, implying extensive magma intrusion through the entire Lei-Qiong region. These observations strongly suggest the necessity of the Hainan mantle plume: the plume brings upwelling mantle materials to erupt to the surface as basalts or to intrude into the upper crust as gabbros, and to further thicken the crust. The high-velocity anomaly within the upper crust beneath the peninsula is the seismic signature of the igneous rocks. Similar to that at Yellowstone [Huang et al., 2015], the seismic velocity at greater depths is significantly reduced, suggesting a high temperature and possible partial melting in the lower crust due to the plume.

An alternative explanation for the basaltic magmatism is seafloor spreading associated with the opening of the South China Sea [Tu et al., 1991]. However, recent geochemical studies of isotopic signatures [Zou and Fan, 2010; Wang et al., 2013] explicitly suggest that the magma originates from the lower mantle with an average upwelling rate of < 1 cm/yr. Additionally, the higher than normal mantle temperature [Chen et al., 1991; Yu et al., 1998; Wang et al., 2012] cannot be explained by a passive spreading mechanism. Given the existing seismic evidence [Lebedev et al., 2002; Lebedev and Nolet, 2003; Montelli et al., 2004; Huang and Zhao, 2006; Lei et al., 2009; Wang and Huang, 2012], it is reasonable to attribute the thickened crust to the Hainan mantle plume.

According to cross sections A-A', B-B', and D-D' in Figure 2b, we estimate the mantle plume to be confined between the latitudes of 21.3°N and 19.8°N at Moho depth, without clear eastern and western boundaries.
If the plume has a cylindrical shape near the Moho, a horizontal section at 30 km depth is expected as the shadow area in Figure 2a, which agrees well with the distribution of basaltic outcrops. The diameter of the plume is about 160 km, compatible with the value of 200 km suggested by global tomographic studies [Lebedev et al., 2000; Lebedev and Nolet, 2003; Montelli et al., 2004], but larger than the 80 km value from a regional body wave tomography study [Lei et al., 2009]. This difference is likely because the regional tomographic study by Lei et al. [2009] was confined to Hainan Island, only partially covering the mantle plume, and thus leading to a smaller estimate of the plume size, while our results suggest that the major part of the mantle plume is beneath the Leizhou Peninsula rather than Hainan Island.

4.2. Thinning of the Mantle Transition Zone

The 410- and 660-km discontinuities are usually attributed to isochronal phase transformations of (Mg,Fe)$_2$SiO$_4$ olivine transforms to wadsleyite ($\alpha \rightarrow \beta$) at 410 km depth and ringwoodite transforms to bridgmanite (silicate perovskite) and magnesiowüstite ($\gamma \rightarrow$ bdg + mw) at 660 km depth [e.g., Ringwood, 1975; Liu, 1976; Tschauner et al., 2014]. Since the Clapeyron slope is positive for $\alpha \rightarrow \beta$ but negative for $\gamma \rightarrow$ bdg + mw, the topology of these two discontinuities and the MTZ thickness have been widely used to infer changes in mantle temperature [e.g., Shearer, 1991; Flanagan and Shearer, 1998; Lebedev et al., 2002; Lawrence and Shearer, 2008; Cao et al., 2011]. When temperature increases due to a hot mantle plume, the 410-km discontinuity consequently becomes deeper, whereas the 660-km discontinuity becomes shallower, causing a thinned MTZ. Global seismic studies of SS precursors [Houser et al., 2008; Deuss, 2009] suggest that the MTZ beneath the South China Sea is slightly thinner than the surrounding region.

On the other hand, at temperatures higher than 1800°C and depths around 640–680 km, the majorite-bridgmanite (mj $\rightarrow$ bdg) transition with a positive Clapeyron slope becomes predominant [Hirose, 2002]. Therefore, the 660-km discontinuity can also become deeper within a hot mantle plume, which complicates the thermal interpretations. This garnet phase change (mj $\rightarrow$ bdg) was invoked to explain some observations that image a depressed 660-km discontinuity and an unchanged MTZ thickness at several hot spots [e.g., Tauzin et al., 2008; Cao et al., 2011; Jenkins et al., 2016], while other studies still found a thinned MTZ at these hot spots [e.g., Shen et al., 1998; Lawrence and Shearer, 2008]. The potential temperature of 1541°C measured by Wang et al. [2012] leads to about 1800°C at 660 km depth, leaving it ambiguous to distinguish the major reaction. Our results show a depressed 410- and an elevated 660-km discontinuities, resulting in a thinned MTZ east of the Lei-Qiong region (Figure 3b). This is consistent with the transition $\gamma \rightarrow$ bdg + mw rather than mj $\rightarrow$ bdg, suggesting that the olivine reaction is predominant in this region.

Thermal variations of the Hainan mantle plume can be estimated from the topography of the 410- and 660-km discontinuities. By substituting the Clapeyron slope ($\gamma$) into the relationship between the depth ($y$) and the pressure ($p$) $dp/dy = \rho g/\gamma$, where $\rho$ is the density of the mantle. Based on the values used or measured by Bina and Helffrich [1994], Cao et al. [2011], Tauzin and Ricard [2014], and Jenkins et al. [2016], the Clapeyron slopes and their uncertainties are estimated as $\gamma_{410} = 2.7 \pm 0.2$ MPa °C$^{-1}$ and $\gamma_{660} = -2.6 \pm 0.5$ MPa °C$^{-1}$. With $\rho_{410} = 3633.5$ kg m$^{-3}$, $\rho_{660} = 4186.4$ kg m$^{-3}$, and $g = 10$ m s$^{-2}$ [Turcotte and Schubert, 2002], thermal gradients at the discontinuities are (d$T/dy$)$_{410} = 13.6 \pm 1.0$°C km$^{-1}$ and (d$T/dy$)$_{660} = -16.7 \pm 3.1$°C km$^{-1}$. Therefore, the mantle temperature should be increased by about 200–15°C at the 410-km discontinuity (Figure 3b, depressed by 15 km), and 170 ± 30°C at the 660-km discontinuity (Figure 3b, 10 km shallower than normal). We conclude that the temperature of the MTZ beneath the Lei-Qiong region is about 170–200°C higher than the surrounding mantle. The thermal variation agrees well with the high potential temperature [Wang et al., 2012] and is compatible with the value of 150–210°C suggested for the Iceland mantle plume [Shen et al., 1998; Jenkins et al., 2016].

Unlike for crustal thickening, it is difficult to define the boundary of the plume at MTZ depths. Figure 3b shows that the 410-km discontinuity returns to normal depths north of 22.3°N and south of 17.8°N, whereas the 660-km discontinuity remains elevated throughout the whole cross section. This scale (at least 500 km across) is in agreement with previous studies [Huang and Zhao, 2006; Wang and Huang, 2012] but is much larger than the diameter of the plume suggested by the crustal thickening (160 km in section 4.1). Furthermore, contrary to our observations with a rough 410- but a smooth 660-km discontinuity, several studies showed strong topography correlations between them [Chevrot et al., 1999; Tauzin and Ricard, 2014]. These discrepancies can be explained if the Hainan plume has a more complicated shape than a narrow conduit. The thinned MTZ, with a much larger scale, is probably not only responsible for the narrow plume
conduit in the upper mantle but also reflects the deep and broad root of the plume in the lower mantle. This is supported by a recent global seismic tomography model [Zhao, 2007] that shows a broad LVZ in the lower mantle with a scale of 1000–2000 km, interpreted as a ponding of hot plume materials beneath the MTZ. The rough topography of the 410-km discontinuity also indicates more localized thermal variations near the top of the MTZ due to plume focusing.

5. Conclusion

We used the receiver function method to analyze broadband seismic data from the Leizhou Peninsula and Hainan Island to investigate seismic discontinuities beneath the Lei-Qiong region. Results show that the crustal thickness of the Leizhou Peninsula has been thickened by 10–15 km, and the depth of the 410-km discontinuity varies from 420 to 450 km, while the 660-km discontinuity appears consistently at 650 km depth. All these observations provide direct seismic evidence for the existence of the Hainan mantle plume. Based on the variation of the Moho topography and the distribution of basaltic outcrops, we estimate that the plume is centered at 20.5°N, 110.5°E with a diameter of about 160 km at Moho depth. Because the MTZ beneath this region is thinned by about 25 km, we suggest that the temperature of this plume is about 170–200°C higher than the surrounding mantle.

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