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Key Points:

- Lithospheric V_3 velocity structure beneath the Mount Baekdu volcano is imaged from Bayesian inversion using multiple ambient noise data sets
- Preferred Earth structure characterized by a relatively high velocity middle-to-lower crust and a thick crust beneath the volcano
- Mafic underplating and overlying cooled felsic layers by fractional crystallization is suggested for magmatic structure

Supporting Information:

- Supporting Information S1

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Seismic constraints on magma evolution beneath Mount Baekdu (Changbai) volcano from transdimensional Bayesian inversion of ambient noise data

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Abstract The magmatic process of continental intraplate volcanism (CIV) is difficult to understand due to heterogeneous interactions with the crust and the lithospheric upper mantle. Mount Baekdu (Changbai) volcano (MBV) is one of the prominent CIVs in northeast Asia that has shown a complex history of eruptions and associated magmatic structures. In addition, the relationship between the crustal magmatic structures and upper mantle phenomena are enigmatic due to the lack of consistent seismic constraints for the lithospheric structure. To enhance comprehensive understanding of the MBV magma evolution, we image the lithospheric structure beneath the MBV and surrounding regions using ambient noise data and the following two approaches: (1) multiple measures of ambient noise dispersion are acquired through different methods and (2) a transdimensional Bayesian inversion method is utilized to obtain unbiased results in joint analysis of the multiple data sets. The estimated Earth structure shows a thick crust (~40 km) and a crustal anomaly with relatively high S wave velocity in the depth range 20–40 km. This type of structure extends to ~100 km north from the MBV and is accompanied by the shallow and rapid S wave velocity decrease beneath the mantle lid (~80 km). Through a comparison with previous P wave models, we interpret this structure as a consequence of compositional partitioning by mafic underplating and overlying cooled felsic layers as a result of fractional crystallization.

1. Introduction

Continental intraplate volcanism (CIV) is characterized by diffusive, repeated, and small-volume eruptions without a clear relationship to marginal tectonic activities [e.g., Valentine and Hirano, 2010]. The occurrence of CIV has invoked hypotheses of various mechanisms, such as deep mantle plumes, vertical and horizontal flows in the asthenosphere, and lithospheric extensions and decompressional melting [e.g., Anderson, 2000; Courtillot et al., 2003; Hoernle et al., 2006; Conrad et al., 2011]. However, it has also been suggested that the continental crust and lithosphere play a significant role in the development of the volcanoes [e.g., Fitton et al., 1988; Jurine et al., 2005].

Cenozoic volcanoes are scattered across the border region between China and North Korea (Figure 1). They have been recognized as CIV due to the position where they reside in the margin of the North China craton and distanced from subduction zones by more than 1000 km. The Mount Baekdu (Changbai in Chinese) volcano (hereafter MBV) is a prominent active volcano in this region (Figure 1). The MBV has experienced one of the largest Holocene (~946 Common Era (C.E.)) eruptions and recent unrest from 2002 to 2005 [Horn and Schmincke, 2000; Wei et al., 2007; Xu et al., 2012; Oppenheimer et al., 2017]. Despite the evidence of shallow volcanic activities, the lithospheric magmatic structures and their development are still enigmatic due to conflicting observations. The transition of magma from alkaline basalt to silicic compositions [Wang et al., 2003; Zou et al., 2008; Zhang et al., 2014] and the currently nonextensional feature in the lithosphere [e.g., Wang et al., 2001] are not reconciled with the documented upper mantle upwelling and magma supplies [e.g., Kuritani et al., 2009; Zhao et al., 2009; Tang et al., 2014]. In addition, the estimated crustal structure is mutually inconsistent [e.g., Zhang et al., 2002; Song et al., 2007; Choi et al., 2013], though there is a strong evidence of crustal partial melting [e.g., Kyong-Song et al., 2016]. This emphasizes the necessity for comprehensive imaging studies to capture the entire crustal and lithospheric structures associated with magma activities, which could be potentially affected by magma supplies from the asthenospheric mantle.

In this study, we derive a three-dimensional (3-D), S wave velocity (V_3) model of the lithosphere beneath the MBV and surrounding regions. A procedure that encompasses ambient noise surface wave tomography in

Figure 1. Topography map of the region of study. The area enclosed within red dashed lines is an area sampled sufficiently well by interstation paths to conduct ambient noise tomography. Locations of stations are marked by squares in different colors, volcanoes (MBV: Mount Baekdu volcano), and are shown by triangles, and a transparent irregular shape is the volcanic area (symbols are also explained in the legend). Interface depths of the subducting Pacific slab are shown by blue contour lines (both in the main image and in the inset). Dashed lines are national boundaries. Boundaries of tectonic regions are indicated by solid lines and named as follows: SB – Songliao Basin; TLFZ – Tan-Lu Fault Zone; CMR – Changbai Mountain Range, and KP – Korean Peninsula. Inset: the region of study in a wider geographical and geological context.

the first step, and then inversions for the depth profiles of V_S in the second step, has been widely used to construct regional V_S models in volcanically active regions [e.g., Behr *et al.*, 2011; Kim *et al.*, 2012]. We combine multiple data sets of surface wave dispersion measured in the spectral domain [Ekström *et al.*, 2009] and the time domain [e.g., Bensen *et al.*, 2007] using ambient noise waveform correlations. Between the two methods, the spectral-domain method involves measurement at longer periods [Calkins *et al.*, 2011], which generally sample larger depths. In consequence, a broader range of periods for surface wave dispersion data are exploited from the common continuous waveform data. To invert for the depth profiles of V_S , we use a fully nonlinear inversion method based on the hierarchical and transdimensional Bayesian approach [Bodin *et al.*, 2012], which has been used to infer both regional structure [Young *et al.*, 2013; Kim *et al.*, 2016a] and global compressional velocity field in the lowermost mantle [Tkalcic *et al.*, 2015]. The approach prevents potential overfitting or underfitting to data, which is often the case when parameterization and regularization (damping and smoothing) are applied explicitly in conventional methods (for a review, see, e.g., Sambridge *et al.* [2013]). The hierarchical aspect of the method accounts for varying noise characteristics among different data sets. This is particularly important in the joint inversion of multiple data sets, to weight the contribution of each data set in concordance with the noise associated with each data set.

Our 3-D model depicts V_S structures of the entire lithosphere below the MBV and the NE China. We pay special attention to structures in the crust that are possibly magma related, and their potential link with the proposed upwelling in the asthenospheric mantle [e.g., Guo *et al.*, 2016]. Taking into consideration P wave velocity (V_P) structures from the model of Song *et al.* [2007], the observed features are interpreted in terms of chemical composition and the possible existence of partial melting.

2. Background

The collision of the North China craton and the South China Block formed the east-west Qinling-Dabies belt and the Tancheng-Lujiang (Tan-Lu) fault [e.g., *Faure et al.*, 2001], which extends to northeast China and forms the Tan-Lu Fault Zone (TLFZ) (Figure 1). The Changbai Mountain Range (CMR) and the Songliao Basin are adjacent to the TLFZ (Figure 1). Since Late Mesozoic to Cenozoic, the margin of the old continental bodies has been affected by extensions and back-arc openings in the East Sea (Sea of Japan) due to multiple subductions of different oceanic lithosphere and the interaction between the Indo-Australian and Eurasia plates [*Ren et al.*, 2002]. The marginal thinning of the cratonic lithosphere (thickness is ~60–100 km) is considerable beneath the eastern part of the North China craton and the Korean Peninsula [*Menzies et al.*, 1993; *Feng et al.*, 2010] that could be attributed to the oceanic slab subductions since the Late Cretaceous [e.g., *Menzies et al.*, 1993; *Zhang et al.*, 2003]. Although the bottom of the lithosphere beneath the MBV has not been extensively studied, sublithospheric low-velocity structures and possible upwelling have been proposed [*Guo et al.*, 2016; *Kim et al.*, 2016b].

Seismic models have shown the presence of a stagnant slab at the mantle transition zone beneath NE China and the Korean Peninsula, and low-velocity anomalies in the upper mantle above it [e.g., *Fukao et al.*, 2009; *Zhao et al.*, 2009]. The imaged features motivate the interpretations of hot and wet upper mantle being linked to the occurrence of the CIV above it [e.g., *Zhou and Armstrong*, 1982; *Basu et al.*, 1991; *Hahn et al.*, 2008; *Zhao et al.*, 2009; *Kuritani et al.*, 2011; *Wang et al.*, 2015]. Alternatively, mechanical processes were suggested associated with the subduction-induced plume [*Tang et al.*, 2014] or delamination of the lithosphere [*Chen et al.*, 2007; *Guo et al.*, 2016], apart from the thermal and chemical mechanisms. However, the exact depth of the melt production is under debate [e.g., *Chen et al.*, 2007; *Zhang et al.*, 2014].

The MBV and other neighboring CIVs are distributed across the eastern boundary of the North China block (Figure 1). The volcanic structure is emplaced on the Archean-to-Proterozoic granite basement, similar to other CIVs in northeast China [*Basu et al.*, 1991; *Liu et al.*, 2001]. Calc-alkaline volcanic rocks are distributed associated to extensional tectonic activities since the Late Cretaceous. The MBV consists of the Miocene to Pliocene basalt (20–2.6 Ma) and the Pleistocene shield (2.8–1.5 Ma), the stratocone with trachyte compositions (1.0–0.04 Ma), and the Holocene ignimbrite deposition [*Wang et al.*, 2003; *Wei et al.*, 2007]. The caldera (Chunji in Korean and Tianchi in Chinese) at the top of the MBV was formed by explosive eruptions of trachyte and felsic magmas. In contrast to the basaltic-to-silicic compositional transition, the nearby CIVs (Longgang and the Jingbohu volcanoes in Figure 1) are predominantly basaltic, where the cinder cone-forming eruptions are relatively recent (Holocene) compared to the MBV [*Liu et al.*, 2001, 2009]. These heterogeneous features are difficult to reconcile with a largescale model [e.g., *Zhao et al.*, 2009]. *Zhang et al.* [2014] suggested significant interactions with lithospheric structure during the melt generation process by a magma underplating at the bottom of the crust associated with lithospheric extensions.

The crustal structure around the MBV has been studied by a number of authors [*Zhang et al.*, 2002; *Hetland et al.*, 2004; *Yang et al.*, 2005; *Song et al.*, 2007; *Wu et al.*, 2009; *Kyong-Song et al.*, 2016]. A thick crust (~35–40 km) directly beneath the MBV and a relatively shallow Moho (~30–35 km) in the surrounding region have been suggested from seismological studies (> ~20 km from the summit of the MBV) [*Zhang et al.*, 2002; *Hetland et al.*, 2004; *Song et al.*, 2007; *Tao et al.*, 2014; *Kyong-Song et al.*, 2016]. The crustal V_P/V_S ratio (~1.7–1.8) estimated in the surrounding region is comparable with a stable crust in the North China craton and the Korean Peninsula [*Hetland et al.*, 2004; *Tao et al.*, 2014; *Kyong-Song et al.*, 2016]. However, a significant increase of the average crustal V_P/V_S ratio (~1.9–2.1) has been reported beneath the MBV [*Kyong-Song et al.*, 2016]. They infer multiple values for the Moho depth and V_P/V_S ratio pairs due to a trade-off (V_P/V_S ratios are 2.1 and 1.9 for the Moho depths of 33 km and 39 km). From the MBV to ~100 km to the north, a low V_P structure (the maximum 6.2 km/s and as low as 5.4 km/s at places) was estimated at the midcrustal depths (~15–30 km) from the north-south line of an active-source experiment [*Song et al.*, 2007]. The authors estimated relatively shallow Moho discontinuity (30 km) below the low-velocity structure with 7.6 km/s of the uppermost upper mantle V_P . However, other geophysical studies do not support the existence of the low-velocity structure [*Zhang et al.*, 2002; *Choi et al.*, 2013], while at the same time they point to the presence of structures with low density and low V_P only beneath the MBV. In contrast, it is well documented that a shallow magmatic structure exists near the summit of the MBV (< 5 km), which is thought to have caused earthquake

Figure 2. Stacked (a) empirical Green's functions (EGF) in the time-distance domain and (b) real parts of the complex coherency spectrum (RCS) in the frequency-distance domain for all available interstation paths.

swarms, surface deformation, high seismic attenuation, and low electric resistivity at shallow depths [e.g., Tang *et al.*, 2001; Wu *et al.*, 2006; Xu *et al.*, 2012].

3. Data and Ambient Noise Analysis

3.1. Data and Ambient Noise Correlation

To obtain interstation correlation functions, we collect continuous broadband data on vertical components of a permanent station (MDJ) and 46 stations of two temporary networks (Figure 1). The first network is the 1998–1999 PASSCAL experiment [Hetland *et al.*, 2004], which consists of 19 stations deployed around the MBV and close to the border region between China and North Korea. In addition, we use data from 26 out of 127 stations of the 2009–2011 PASSCAL experiment [Ni, 2009]. We select only the stations from the south-eastern part of the network to obtain an overlap with the northwestern part of the previous PASSCAL experiment (Figure 1).

In this study, interstation correlation functions are calculated using time domain method to obtain empirical Green's functions (EGF) and spectral-domain methods to obtain real parts of the complex coherency spectrum (RCS) [Aki, 1957]. As an initial step, the mean, trend, and instrument responses are removed from the continuous waveform of each station. Subsequently, the data are subdivided into half-hour time windows with 50% overlap. Instead of applying the time domain normalization (e.g., one-bit normalization), we reject windows with large glitches to suppress influences of strong transient signals (e.g., earthquakes) on correlation functions [e.g., Seats *et al.*, 2012]. Such windows are identified when the highest amplitude exceeds the third quartile (75th percentile) to the root-mean-square (RMS) of the entire data. Each subwindow is whitened in the frequency domain [e.g., Bensen *et al.*, 2007]. On average, the obtained EGFs and RCSs show coherent patterns of Rayleigh wave signals (Figure 2). In the frequency-distance plot (Figure 2b), the fringe pattern is clear at frequencies between 0.01 and 0.1 Hz and for distances between 50 and 550 km.

3.2. Measurement of Surface Wave Dispersion Velocities

To measure phase velocity dispersion using RCSs, we use the property of seismic ambient noise based on the spectral formulation known as the spatial autocorrelation (also known as Aki's method) [Aki, 1957; Ekström *et al.*, 2009; Tsai and Moschetti, 2010] where the RCS is expressed with Bessel function (see Appendix A for further details). However, the equality between the RCS and Bessel function is not achieved particularly well in practice due to inconsistent amplitude in time and space of the ambient noise spectrum. To circumvent this problem, a method of comparing zero-crossings of the RCS with Bessel function waveforms was proposed [Ekström *et al.*, 2009; Boschi *et al.*, 2013]. The particular advantage of this approach compared to the time domain methods [e.g., Bensen *et al.*, 2007] is that there is no implicit limitation of the maximum

Figure 3. An example of the phase velocity measurement using the spectral method for a station pair MDJ–CBAI. Diagrams of (a) phase velocity of synthetic Bessel function waveforms and (b) observed RCS as a function of frequency. (c) Estimated amplitude scaling factor based on a grid search using equation (A2). (d) The variation of the variance reduction (equation (A3)) in the frequency–phase velocity domain. The dashed lines indicate the target area of search to avoid phase cycle skipping. Small circles in Figures 3a–3d are measured phase velocities as a function of frequency. (e) A comparison of waveforms for the measured phase velocities. The solid line: observed RCS, red crosses: amplitude-corrected RCS by the estimated amplification factor in Figure 3c, and blue dots: the corresponding synthetic Bessel function waveform.

measurable wavelength (period) for a given interstation distance. It has been suggested that the measurement is reliable for the interstation distance of about half-wavelength [Calkins *et al.*, 2011].

To exploit more information in the RCS, an approach has been suggested using direct fitting of waveforms [e.g., Menke and Jin, 2015]. Here we develop a waveform fitting method based on a grid search technique (see Appendix A for further details). An example of phase velocity measurement is shown (Figure 3) for a station pair between the permanent station MDJ and a station CBAI near the MBV. In frequency–phase velocity domain, the variance reductions in equation (A3) (Figure 3d) are calculated between synthetic Bessel functions (Figure 3a) and observed RCSs (Figure 3b). To fit waveforms directly (Figure 3e), amplification

Figure 4. An example of time domain dispersion measurements for the station pair MDJ and CBAI. The results of (a) Group and (b) Phase velocity measurements. Dashed lines indicate the measurement windows, gray line is the array-mean dispersion curve, and white dots show measured dispersion data. Phase velocity measurements from the spectral method are superimposed with white stars.

factors (Figure 3c) are additionally searched at each grid node. Then phase velocities are estimated by tracking ridges ($>75\%$ of the variance reduction) of the fitness functions (Figure 3d). The phase ambiguity is addressed by using a reference dispersion curve, which is obtained from mean phase velocities from the preliminary measurements. We visually check each measurement not only for zero-crossings but also for amplitudes and shapes of waveforms resulting from the grid search (Figure 3e).

In addition, we measure phase and group velocities using the conventional time domain method [Bensen *et al.*, 2007] (Figure 4). At each wave period, phase and group velocity arrival times are measured by picking the highest amplitude within the predefined length of time windows (within ± 0.75 s of wave periods centered on the predicted arrival times for the mean group velocity across the array and within ± 0.5 s of the predicted arrival times for the mean phase velocity across the array). The measurements are finally selected based on the following criteria: the signal-to-noise ratio ($\text{SNR} > 7$) and the minimum interstation distance (greater than three wavelengths).

In comparison, the spectral approach yields more phase velocity measurements at longer periods (>15 s), while the time domain method is more efficient at short periods (Figure 5a). Therefore, the two methods complement each other to retrieve more measurements in a broader range of periods. Overall, dispersion

Figure 5. (a) Numbers of dispersion measurements for different periods. (b) Average dispersion measurements over all interstation paths using the same symbols as in Figure 5a. The error bar indicates the standard deviation of each data measurement. (c) Differences of phase velocity estimations between the time and spectral-domain methods. (d) A histogram of the phase velocity differences shown in Figure 5c.

curves overlap in the common period range (Figures 4b and 5b). For most station pairs, the residual velocities between the two methods (Figure 5c) are generally small (<0.05 km/s). A good consistency is confirmed statistically by the residual distribution (Figure 5d), which is centered at zero with a small standard deviation (0.03 km/s).

3.3. Surface Wave Tomography

To estimate lateral variations of group and phase velocities, surface wave tomography is performed using interstation dispersion velocity data. Tomographic inversions are carried out for (a) phase velocity measurements (10–36 s) from the spectral-domain method, (b) group, and (c) phase velocity measurements (5–20 s) from the time domain method. We use an iterative nonlinear inversion scheme based on the subspace technique [Kennett *et al.*, 1988]. The fast marching method [Rawlinson and Sambridge, 2004] incorporates a prediction of propagating wave fronts. Based on preliminary inversions, the size of the inversion grid is determined ($0.25^\circ \times 0.25^\circ$). Damping values (0.5 and 2.0 for group and phase velocities, respectively) are chosen by using trade-off curves (Figure S1 in the supporting information) between model roughness and data fitting. The smoothing term is omitted because the inversion produces stable results without applying it. A resolution of tomograms is confirmed by the checkerboard recovery tests. For the input models, $1^\circ \times 1^\circ$ checkerboard patterns are used with ± 0.3 km/s and ± 0.2 km/s velocity perturbations for group and phase velocities, respectively. To produce synthetic data sets, Gaussian random noises are applied with standard deviation of 0.5 s. In addition, we perform additional recovery tests using $0.75^\circ \times 0.75^\circ$ checkerboard patterns (Figure S2), an input model with a single lower velocity anomaly (-0.45 km/s and -0.3 km/s for group and phase velocities to the background velocities) in the MBV region (Figure S3) and $1^\circ \times 1^\circ$ checkerboard patterns with a higher level (1.0 s) of noise (Figure S4).

Figure 6. Tomograms for (a) group and (b) phase velocity using EGF and RCS data for selected periods (5, 12, 20, 28, and 36 s). The period, the mean velocity, the number of used rays, and the type of data are indicated above tomograms. Tectonic boundaries are represented by solid lines, while national boundaries are shown by dashed lines. The selected region is confined within gray lines. Small circles are locations of stations and triangles are locations of volcanoes.

Figures 6 show tomography results for selected periods. Phase velocity maps from RCS data reveal low-velocity anomalies around the MBV for all periods, whereas generally higher velocities are resolved in the TLFZ. Low-velocity anomalies around the MBV become dominant as the period increases, and vice versa beneath the Songliao Basin. Short periods (12 and 20 s) are characterized by a good recovery, while the resolution decreases at longer periods. However, the velocity contrast of input patterns is always preserved between the MBV and surrounding regions. Both group and phase velocity tomography from EGF data (Figures 7b and S2–S4) resolve input patterns in the selected area with relatively high path density (Figure 7a). Estimated anomalies of the phase velocity agree well with those estimated from RCS data (Figure 6).

4. The 3-D V_S Model

4.1. Hierarchical Transdimensional Bayesian Inversions for 1-D Depth Profiles

A 3-D V_S model is constructed by a series of 1-D depth inversions using multiple data sets of group and phase velocity dispersions. Here we use four different surface wave dispersion data (curves) of Rayleigh wave fundamental mode, which are extracted from tomograms of (1) group velocities in the period range

Figure 7. Maps showing (a and b) raypaths (blue lines) and (c and d) results of the checkboard recovery test for selected periods as in Figure 6. The input model for the test is presented at the bottom of Figure 7d.

5–20 s from the time domain method (Figure 6a), (2) phase velocities in the periods range 5–9 s from the time domain method (Figure 6b), (3) phase velocities in the period range 10–36 s from the spectral-domain method (Figure 6b), and (4) phase velocities in the period range 38–50 s from *Kim et al.* [2016b]. To better constrain deeper depths ($> \sim 60$ km), we additionally include this long period data, which are from ambient noise tomography for the broader area in northeast Asia [*Kim et al.*, 2016b]. The hierarchical transdimensional Bayesian method (see Appendix B for further details) provides a systematic way of joint inversions to combine four different data sets. Throughout the hierarchical scheme, the levels of data sensitivities are set differently among data sets by adjusting the levels of input data variances. Hence, the four scaling factors of corresponding data types are simultaneously estimated in the inversion to control the levels of data variances. We assume that the data variance for each data value (equation (B2)) is proportional to the variance of the corresponding data point measured from all data (error bars shown in Figure 5b). Therefore, empirically estimated different data variances between different data points are accounted for in the inversions.

The inversions are performed using flat prior probabilities of parameters, including V_5 (between 2.0 and 5.5 km/s), the interface depths (between 0 and 200 km), the number of layers (between 2 and 30), and the scaling factors (between -4.0 and 4.0). In each Markov chain, posterior models (50,000) are collected after a “burn-in” period, in which 10,000 models are discarded to obtain converged samples. For a single inversion, we use 64 Markov chains in parallel and collect samples from 32 nontempered chains ($\beta = 1$; see equation (B1)). Consequently, the posterior probability densities (PPD) are formed by 1.6 million samples of models.

Figures 8 and 9 show examples of inversion results (PPD) for the selected locations of the MBV region and the TLFZ (locations are indicated in Figure 10). Estimated PPDs and mean V_5 profiles (Figure 8) resolve relatively sharp boundaries at the bottoms of shallow crust (Figures 8b and 8e) and at the Moho (Figure 8e).

Figure 8. Examples of posterior probability distributions (PPD) for the Bayesian inversions of (left) the MBV and (right) the TLFZ regions (for the locations, refer to Figure 10): (a and d) PPD of layer interface depths; (b and e) PPD of model parameters (V_s as a function of depth); (c and f) PPD of the transdimensional parameter (number of layers). The solid and dashed white lines in Figures 8b and 8e present mean models and their ± 1 standard deviation.

This implies that our inversions can resolve features associated to velocity interfaces without any arbitrary parameterization. In comparison to the TLFZ (Figure 8e), the V_s profile beneath the MBV (Figure 8b) clearly reveals features with more smoothly increasing velocities at depths from 20 km to 60 km. In addition, the depth of the low-velocity structure is relatively shallower (~ 80 – 120 km) than beneath the TLFZ (> 100 km).

Figure 9. PPDs of dispersion predicted from the resulting profiles for the MBV region (top; see also Figures 8a–8c) and the TLFZ region (bottom; see also Figures 8d–8f). (a, e) EGF group velocity dispersion and the observed data (gray circles and error bars), (b, f) EGF phase velocity dispersion and data (red circles and error bars), (c, g) RCS phase velocity dispersion and data (green circles and error bars), and (d, h) phase velocity dispersion and data (blue circles and error bars) from an independent study [Kim *et al.*, 2016b]. Insets show PPDs for the scaling parameter of each corresponding data set (with the corresponding color code).

Figure 10. Horizontal slices of the estimated 3-D V_S model at selected depths (left column) and their relative uncertainties (right column). The depths from top to bottom are 10, 22, 36, and 110 km. The depth and the mean value are presented at the top of each panel. Straight dashed lines indicate vertical profiles shown in Figure 11. Nodes of the depth inversion are superimposed with small circles on uncertainty maps. The locations of featured inversions in Figures 8 and 9 are shown by red circles (marked by a and b in the velocity map of 10 km depth for the MBV and the TLFZ locations, respectively). Note that the anomaly described in the main text is indicated by dashed square in left column. Triangles are the locations of the intraplate volcanoes.

Figure 11. Vertical sections along profiles (a) A-A', (b) B-B', and (c) C-C' featured in Figure 10. In each panel, topography is shown on top, V_s in the middle, and V_s perturbation to the mean V_s of each depth (as shown above each panel in Figure 10) at the bottom. Names of major tectonic regions appear in the top panel. The anomalous structure beneath the MBV is indicated by the circle with bold-dashed line. Contour lines (narrow-dashed lines) are presented with the spacing of 0.2 km/s in the middle panels. Contour lines with $V_s = 4.2$ km/s are marked with solid lines in the middle and the bottom panels to roughly indicate Moho (~30–40 km depth) and the bottom of the lithosphere (>60 km depth).

The PPD of interface depths (Figures 8a and 8d) show corresponding peaks around the boundary depths. However, such information on interface depths is not directly interpreted in our analysis due to the nonlinear relationship between the interface depth and sensitivities of surface wave dispersion data (see the further recovery tests in section 5.1). Sampling of the dimensional parameter (Figures 8c and 8f) is not hampered by the used prior range. We define model uncertainties with standard deviations of the PPD of V_s profiles at every depth grid [e.g., Young et al., 2013]. Figure 9 presents distributions of predicted data and scaling factors, which correspond to the PPDs of model parameters (Figure 8). Distributions of predicted data agree with observed data and corresponding empirical error ranges. In general, the PPDs of predicted data are more broadly distributed at data points with higher error (e.g., group velocities) than the points with smaller errors (e.g., phase velocities). Negative values of the estimated scaling factor (insets in Figures 9a, 9c, 9e, and 9g) show how overestimated empirical data errors (Figure 5b) are adjusted by the hierarchical scheme. In addition, different average values and widths of the PPDs of scaling factors show the inherent competition between different data sets in the estimation of structural models.

4.2. The 3-D V_S Model

A 3-D V_S model is composed by interpolation of 1-D mean models from the PPD of the Bayesian inversion (e.g., solid white lines in Figures 8b and 8e). Inversions are performed at each node of a horizontal grid with $0.25^\circ \times 0.25^\circ$ spacing (Figure 10). At upper crust (e.g., top 10 km in Figure 10), a large low-velocity region corresponds to thick sedimentary structures in the Songliao Basin. The MBV volcanic area (Figure 1) is also imaged with low V_S . The TLFZ and the surrounding areas of the MBV volcanic area show high V_S . At middle-to-lower crustal depths (22 km in Figure 10), relatively high V_S anomalies appear in the north of the MBV (indicated by a dashed circle), the southeast of the Longgang volcano (127°E , 42°N), and the north of the Jingbohu volcano (from 127°E , 44°N to 130°E , 45°N). These relatively high V_S regions are correlated to relatively low V_S anomalies (~ 3.8 km/s) around Moho depth (36 km in Figure 10). The relatively low V_S basically indicates thick crust due to the presence of crustal materials compared to high V_S (>4.2 km/s) in the upper mantle in surrounding regions. The pattern of prominent low V_S at the bottom of the model (110 km in Figure 10) shows great similarity with the areas with high middle-to-lower crustal V_S and expected deeper Moho. In contrast, the structure beneath the Songliao basin has consistently low V_S in the crust and generally high V_S in the upper mantle.

Vertical sections (Figure 11) are constructed to aid understanding the structure beneath the MBV (see figure Figure 10). Note that the sections A-A' and B-B' in Figure 11 correspond to the lines of the previous active source experiment [Song *et al.*, 2007]. The structure with a relatively high V_S at the midcrustal depths (~ 10 – 20 km) and a relatively low V_S in the lower crust (~ 20 – 40 km) is evident from beneath the MBV to 100 km north of it (indicated by a dashed circle in Figure 11). This structure is underlain by relatively shallow (~ 80 km), low- V_S (~ 4.0 – 4.2 km/s) structures in the upper mantle (the depths of this interface are indicated by a thick contour line of $V_S = 4.2$ km/s in Figure 11). Though not clear, similar features are observed beneath other volcanoes with slightly greater depth (~ 100 km) of velocity drop (i.e., the Jinbohu volcano in A-A' and the Longgang volcano in B-B' in Figure 11). Beneath the TLFZ and the Songliao Basin, generally higher V_S is estimated, though some complexities are imaged (C-C' in Figure 11).

5. Discussion and Conclusions

The prominent features imaged in our model are (1) a thick crust beneath the MBV (~ 5 – 10 km thicker than the surrounding area, e.g., the TLFZ) extending to approximately 100 km north, (2) a relatively high crustal V_S in the depth range 15–40 km in the region with thickened crust, and (3) the shallow (~ 80 km) and rapid V_S decrease beneath the mantle lid in comparison with the surrounding region. We discuss the reliability and the significance of these features in terms of magma evolution processes of the MBV and nearby intraplate volcanoes.

5.1. Resolution and Uncertainty of the Obtained Model

Resolution analysis is performed to check the reliability of imaged features using three sets of synthetic recovery tests (Figure 12). In the first and the second test (Figures 12a and 12b), we use simplified input models to test sensitivities of our inversions to interfaces of the Moho and the velocity decrease in the upper mantle. The assumed simplified models are designed based on the inverted Earth structure from the observed data (Figure 8b). To account for a realistic data noise, we use Gaussian random errors with amplitudes of data noise estimated through the hierarchical process of the Bayesian inversion (Figure 9). The final test is carried out to show the sensitivity of the relatively high V_S structure at lower crustal depths (Figure 12c).

The effect of Moho depth variation from 25 km to 45 km is tested (Figure 12a). Input models consist of three layers for sedimentary (with the thickness of 3 km and $V_S = 2.6$ km/s), crustal ($V_S = 3.6$ km/s), and upper mantle ($V_S = 4.55$ km/s) layers. Estimated models show that our inversions have sensitivities to the Moho, particularly at shallow depths (<35 km). It is noted that the simple input models with the depth to Moho of 30 km result in the mean models close to inverted structure beneath the TLFZ (Figure 12a). However, the inverted Moho structure beneath the MBV at deep crustal depths is not resolved using these simple models.

To test the decrease of V_S at the bottom of the mantle lid (Figure 12b), we add an additional low V_S (4.2 km/s) layer to the model and keep the Moho fixed at 40 km (Figure 12a). The amount of velocity drop is based on our inverted structure (Figure 8b) and previous studies which estimated V_S lower than 8% [Guo *et al.*, 2016].

Figure 12. Results of synthetic sensitivity tests for (a) the Moho depth, (b) the depth of the lower side of the mantle lid, and (c) a more complex structure with an addition of a middle-to-lower crust high velocity structure. Input values of the parameters tested in the inversion are presented at the upper right corner of each panel. The true values from top to bottom in Figure 12c are V_s , the depth of the upper interface, and the Moho depth. The input is in blue and the resolved models are in red. The inverted structure beneath the MBV (the mean model in Figure 8b) and the TLFZ (the mean model in Figure 8e) are shown with solid and dashed lines in Figures 12a and 12b, respectively. The preferred result of the grid search (see the main text) is indicated with yellow background in Figure 12c.

Despite relatively lower sensitivity compared to the Moho depth test, there is a systematic shift in the depth showing negative V_s as the input depth increases from 60 km to 100 km. In comparison, the model beneath MBV corresponds to 70–80 km depths while the TLFZ model is comparable with V_s reduction at 100 km or deeper. This generally agree with the imaged structure (Figure 11).

Finally, a grid search approach is performed to confirm the existence of the middle-to-lower crustal layer with increased V_s and to find optimum values for thicknesses and V_s of the layer. In these experiments,

Figure 13. Joint marginal probability distributions between the Moho depth (all estimated interfaces within the depth range of 20–50 km) and (a and b) the amount of V_S variations across the interface, and (c and d) V_S of the layer above the interface (indicating V_S at the middle-to-lower crustal depths). The distributions are obtained from the posterior samples in inversions for (Figures 13a and 13c) the MBV and (Figures 13b and 13d) the TLFZ regions (Figure 8). The proposed Moho depths in the resolution analysis (see the main text and Figure 12 for further details) are indicated by vertical lines in each panel. Horizontal lines in Figures 13c and 13d present the proposed V_S of the middle-to-lower crust (3.8 km/s and 3.6 km/s, respectively) in the analysis.

we fix the V_S of the upper crust and allow the depth to Moho to vary between 30 and 40 km. We also vary the depth to the upper interface of the layer (between 15 and 25 km) using $V_S = 3.8, 4.0$, and 4.2 km/s. The results of the grid search show that a relatively thick lower crust (25 km in the range of 15–40 km depth as in the inverted images) with moderately high V_S (~ 3.8 km/s) recovers the gradually increasing velocity model (Figure 12c). Note that the model beneath the TLFZ is explained using a simple Moho model (Figure 12a). Therefore, it is unlikely that the relatively high V_S at the middle-to-lower crust and the thicker crust beneath the MBV are artifacts caused by wide sensitivity of surface wave dispersion or velocity depth trade-off.

The trade-off between the Moho depth and the V_S above it can be further explored to support the estimation of relatively high middle-to-lower crustal V_S with thicker crust in the MBV region. Figure 13 shows joint marginal probability distributions between two parameters in posterior models of the Bayesian inversions, which are between the depth of all estimated interfaces in the range of 20–50 km and V_S variations across corresponding interfaces (Figures 13a and 13b), and V_S in the layer above them (Figures 13c and 13d). In the TLFZ region (Figure 13b), there is an abrupt change at the depth of 30 km from a strong positive correlation to a weak correlation, whereas a positive correlation is consistent to deeper depths in the MBV region (Figure 13a). Compared to the MBV region, the correlation change in the TLFZ region indicates that the increase of the V_S variation starts from shallower depth (< 30 km) and ends at the depth due to the presence of a relatively sharp Moho interface and a mantle lid below it as shown in the PPD (Figure 8e). Second, weak correlations between the V_S and the Moho depth in middle-to-lower crustal depths (Figures 13c and 13d)

show that the estimated V_S (~ 3.8 km/s in Figure 13c and ~ 3.6 in Figure 13d) are less affected by the trade-off with the Moho depth. In addition, the patterns of PPD in the MBV region (Figure 13c) show a possibility that the Moho interface can exist even deeper depths (~ 43 km) than we have presented in the resolution test (Figure 12c).

Estimated model uncertainties are relatively small at crustal depths (<0.17 km/s of mean uncertainty as shown in Figure 10) compared to uncertainties associated with upper mantle structures (0.3–0.5 km/s). The increase of uncertainties with depth shows inherent broader and lower data sensitivities of dispersion data at longer periods. The estimated uncertainties show that crustal structure is better constrained than the upper mantle structure. The strong velocity anomalies imaged in the mantle (110 km in Figures 10 and 11) are subdued by high uncertainties (>0.5 km/s). However, our resolution analysis (Figure 12) indicates that the inverted mean model is interpretable despite the estimated large uncertainties. Thus, we focus only on the features confirmed throughout the resolution tests (Figure 12) for the later discussion in this study. The estimated uncertainties can be explained by a large theoretical error in the inversion process [Dosso and Wilmut, 2006], which includes unaccounted effects (e.g., the assumed forms of data noise, imperfections in forward calculation, and the assumption of Gaussianity for probability distributions). Detailed analysis may be required in future studies about model uncertainties of transdimensional and hierarchical Bayesian inversions.

5.2. Comparison With Previous Studies: Double Layering of Middle-to-Lower Crustal Magmatic Structure

Previously reported Moho depths [e.g., Song *et al.*, 2007; Tao *et al.*, 2014] generally agree with the trend of the V_S transition depths (roughly along the isosurfaces of 4.2 km/s in Figure 11). A relatively deeper Moho interface is estimated beneath the MBV and nearby volcanoes compared to generally uniform crustal thickness in surrounding region including the TLFZ (Figure 1, 36 km map in Figure 10, and Figure 11). Our model shows that a thickened crust is extended to the north by ~ 100 km from the MBV. Previous studies using P wave travel time data [Zhang *et al.*, 2002; Song *et al.*, 2007] reported shallower Moho depth (~ 30 – 35 km) in the area. The recovery tests (Figures 12a and 12c) reveal that our inversions using surface wave dispersion data are sensitive to the Moho depth [e.g., Lebedev *et al.*, 2013], particularly when the crust is thinner (see 30 km in Figure 12a). One plausible explanation for this discrepancy might be different sensitivities of P and S waves to these interfaces. This implies that the interface at shallow level (~ 30 km) has stronger impedance for P waves, while S waves mainly detect the deeper (~ 40 km) interface as the Moho interface. The impedance of P waves is generally greater than S waves at lithological boundaries by mafic intrusion [Goodwin *et al.*, 1989]. In this respect, the consistent observations of deeper Moho beneath the MBV is explained by low contrast of V_P at the shallower interface. The evidences of localized partial melting in crust [e.g., Tang *et al.*, 2001; Kyong-Song *et al.*, 2016] support that the local rheology plays a dominant role in forming the less clear crustal interface for P waves [e.g., Hale and Thompson, 1982; Parsons *et al.*, 1992].

The main feature of our model is that the thick crust (~ 40 km) is accompanied by a relatively high V_S middle-to-lower crustal layer (Figures 11 and 12). Based on the sensitivity analysis (Figure 12c), the MBV structure is the most readily explained by a thick layer (in the range of 15–40 km depth) with a slight increase of V_S (~ 3.8 km/s). This feature has a negative correlation with the previously estimated V_P structure [Song *et al.*, 2007] (see section 2 for more details), where the middle crust is slow and the Moho depth is shallow (30–35 km). In addition, the other wide-angle profile [Zhang *et al.*, 2002] shows that V_P at the deeper crust beneath the MBV can be reduced locally to 6.6 km/s. In an attempt to interpret our results with the V_P models, the middle-to-lower crust can be divided into two different parts at the depth of ~ 30 – 35 km by their different seismic properties (Figures 14 and 15a). The upper part (L1 in Figures 14 and 15) has a low V_P and a relatively high V_S (5.5–6.3 km/s and 3.7–3.9 km/s for V_P and V_S , respectively), while both V_P (6.6–7.6 km/s) and V_S (3.8–4.0 km/s) are high in the lower part (L2 in Figures 14 and 15).

5.3. Compositions of the Magmatic Structures

Together with absolute values of V_P and V_S , V_P/V_S ratios provide a stronger constraint on composition (Figure 14). Although the conservatively broad ranges of V_P and V_S , the seismic properties of L1 clearly indicate the presence of highly felsic compositions at middle crust (Figure 14). Possible candidates to explain the V_P/V_S ratio (1.4–1.7) with compositions are from quartzite to granite or metagranite [Christensen, 1996].

Figure 14. Seismic properties of rocks in terms of V_p and V_s for middle and lower crustal depths (400–800 MPa) after Christensen [1996]. Values for granulite facies [Rudnick and Fountain, 1995] are presented by circles. Particularly, felsic compositions are indicated with red symbols. Corresponding V_p/V_s ratios are indicated by lines with Poisson's ratios in brackets. The shaded areas with labels L1 and L2 show possible ranges of compositions for the inferred double layers in this study (see the main text for further details. Also, see Figure 15).

It was suggested that the low V_p body in L1 is attributed to a high temperature without contribution of partial melting [Song et al., 2007]. However, both of the hypothesis of the enhanced temperature or the presence of partial melting are less likely because V_s would be reduced by these effects and the V_p/V_s ratio should become high [Spencer and Nur, 1976; Mavko, 1980]. Hence, we deduce that the low V_p/V_s ratio is mainly due to the compositional contribution of a cooled felsic body (Figure 15c). In contrast, compositions of L2 structure (Figure 14) are largely mafic and ultramafic [Christensen, 1996]. The mafic granulite facies are also included in the range [Rudnick and Fountain, 1995]. These compositions commonly indicate a mafic underplating at the bottom of the crust and mafic intrusions into the lower crust in a continental environment [Thybo and Artemieva, 2013]. A high V_p/V_s ratio (<2.1) implies the presence of partial melting at the lower crustal depths [e.g., Kyong-Song et al., 2016].

The inverted profile has V_s drop ($\sim 8\%$) at the depth (~ 60 – 80 km) in upper mantle (Figure 11), which is consistent with previous seismic studies [Tang et al., 2014; Guo et al., 2016; Kang et al., 2016]. Based on previous estimations [Menzies et al., 1993; Pasyanos et al., 2014], we interpret this as the interface of

Figure 15. Schematic representation of (a) double layering beneath the MBV, (b) the early stage with shield-forming basaltic magmatism, and (c) the current stage with the mafic underplating overlain by a felsic layer. The background image in Figure 15c is V_s perturbation taken from Figure 11a.

lithosphere-asthenosphere boundary (LAB). The synthetic recovery test (Figure 12b) shows that our inversions are sensitive to the relative variation of LAB. While dispersion only inversions provide less precision in depths, the relative variation of LAB (Figure 11) shows that the lithosphere beneath the intraplate volcanoes is thinner in this region. This corresponds to the geochemical signature of the asthenospheric origin of magmas [e.g., *Hahn et al.*, 2008; *Wang et al.*, 2015].

5.4. Reconstruction of Magmatic Structure and Tectonic Implication

The presented model shows a current structure of lithosphere throughout the magma process associated to the intraplate volcanoes in this region. In continental lithosphere, a felsic structure underlain by a middle-to-lower crustal mafic layer is commonly attributed to crustal intrusions and underplating of Basaltic magmas [*Thybo and Artemieva*, 2013, and references therein] (Figure 15b). The midcrustal felsic layer (Figure 15c) can be formed in a relatively short timescale (<100,000 years) followed by partial melting and fractional crystallization due to the crustal heating of the underplating magma [*Petford et al.*, 2000; *Huang et al.*, 2015]. In this context, it is possible to speculate that the neighboring younger volcanoes (i.e., Jingbhou and Longgang) have experienced mainly basaltic magmatism, which might be more recently emplaced (Figure 15c).

Our observations are generally indicative of low crustal temperature and largely cooled magmatic structures beneath the volcanic area (Figure 1), excluding the region locally beneath the MBV with possibilities of a high thermal state and partial melting. In addition, an evidence for a volume of magma feeding the MBV is not found in this study, which implies the absence of current crustal magma activities or the presence of only small-scale structures with subresolution sizes. Accordingly, it is speculated that the current volcanic activity of the MBV is maintained by a residual small portion of silicic magma at shallow magma chamber or by a localized magmatic structure confined beneath the MBV for which we lack resolution. The lack of the crustal interface and evidences of partial melting in the crust directly beneath the MBV are more likely explained by the hypothesis of localized magmatic structures. The overall feature of low temperature and cooling of the magma bodies indicates that the continuous magma supplies from the asthenosphere [e.g., *Zhao et al.*, 2009; *Tang et al.*, 2014] could play a minor role in the recent evolution of the MBV. It is likely that small-scale mechanisms due to localized flows [e.g., *Hoernle et al.*, 2006; *Guo et al.*, 2016], upper mantle upwelling due to instantaneously ascending blobs [e.g., *Lee and Lim*, 2014], or episodic extensions by slab rollback [e.g., *Zhang et al.*, 2014; *Kim et al.*, 2016b] are dominant processes in the MBV region.

Appendix A: Phase Velocity Measurement From Ambient Noise Spectral Correlations

The RCS, $\Re(\omega)$ (Figure 3b) is expressed by the zeroth-order Bessel function of the first kind, J_0 , which is a function of frequency, ω , interstation distance, r , and phase velocity of the corresponding frequency, $c(\omega)$ (Figure 3a):

$$\Re(\omega) = \text{Re} \left\langle \frac{u_A(\omega) u_B^*(\omega)}{|u_A(\omega)| |u_B(\omega)|} \right\rangle \propto J_0 \left[\frac{\omega r}{c(\omega)} \right] \quad (\text{A1})$$

The coherency is defined by the azimuthal average $\langle \dots \rangle$ of normalized cross spectrum between windowed recordings u_A and u_B (u_B^* indicates the complex conjugate of u_B) at stations A and B , respectively. It is known that the ensemble average of the RCSs of all subwindows for a long time period is a proxy of the azimuthal average of noise signals due to the scattering and directional variations of incident ambient noise wavefield [*Boschi et al.*, 2013; *Tsai and Moschetti*, 2010]. Note that a sign of proportionality is used due to spatially and temporally varying amplitude of the ambient noise spectrum, which hinders direct fitting of the RCS to Bessel functions waveforms in the frequency domain [*Ekström et al.*, 2009].

In the method proposed here, we use a grid search method to directly fit RCS waveforms to Bessel functions. To resolve potential amplitude differences, we introduce an additional amplification factor $A(c, \omega)$ (Figure 3c). Then (A1) can be rewritten as

$$10^{A(c, \omega)} \Re(\omega) = J_0 \left[\frac{\omega r}{c(\omega)} \right] \quad (\text{A2})$$

Then, a grid search is performed in the $c - \omega$ domain (Figure 3). At each grid node, an optimum value of $A(c, \omega)$ is searched simultaneously in the range between -1 and 1 . The waveform similarity is quantified by the variance reduction, v , as,

$$v(c, \omega, A) = 100 \times \left[1 - \frac{\sum_{\omega'=\omega-\omega_c}^{\omega+\omega_c} \{10^A \Re(\omega') - J_0(c, \omega')\}^2}{\sum_{\omega'=\omega-\omega_c}^{\omega+\omega_c} \{10^A \Re(\omega')\}^2} \right] \quad (A3)$$

Note that the grid search is carried out using piecewise waveforms segmented in frequency ranges in $\omega \pm \omega_c$. The width of frequency window ($2\omega_c$) is a controlling parameter. Based on our test, the rule of thumb for this width is a half wavelength of Bessel function waveforms. Finally, phase velocities are obtained by picking the highest value of v at each frequency grid. Our method provides a complete distribution of goodness of fit (variance reduction) in the $c - \omega$ domain, such that spurious measurements can be identified visually (Figure 3d).

Appendix B: Methodology of Hierarchical Transdimensional Bayesian Inversion

We use a similar approach to previously developed hierarchical transdimensional inversion methods [e.g., Bodin *et al.*, 2012; Dettmer *et al.*, 2012; Young *et al.*, 2013; Kim *et al.*, 2016a]. Here we briefly describe important and characteristic features of our method.

The Bayesian theorem states that the posterior probability is proportional to the multiplication of prior and likelihood probabilities. To obtain PPD from the transdimensional Bayesian formulation, we use a Metropolis-Hastings-Green sampling algorithm. The acceptance of a newly proposed state (e.g., model) is decided based on the acceptance probability α :

$$\alpha = \min \left[1, \frac{p(\mathbf{x}')}{p(\mathbf{x})} \left\{ \frac{L(\mathbf{x}')}{L(\mathbf{x})} \right\}^\beta \frac{q(\mathbf{x}|\mathbf{x}')}{q(\mathbf{x}'|\mathbf{x})} |\mathbf{J}| \right], \quad (B1)$$

where current and new state including information for model dimension are expressed by $\mathbf{x}$ and $\mathbf{x}'$, respectively. Prior and proposal (the probability of a new state given the current state) probabilities are shown by $p(\mathbf{x})$ and $q(\mathbf{x}'|\mathbf{x})$, respectively. For the brevity, we write the likelihood probability of the current and the new state as functions $L(\mathbf{x})$ and $L'(\mathbf{x})$. To enhance sampling efficiency, we utilize a parallel tempering algorithm [e.g., Kim *et al.*, 2016a] by applying a tempering parameter, β (≤ 1). To change dimension of the model, we utilize a birth-death scheme [e.g., Dettmer *et al.*, 2012; Kim *et al.*, 2016a]. In our method, the Jacobian term, $|\mathbf{J}|$, is unity.

In the hierarchical framework, the likelihood function is defined by

$$L(\mathbf{x}) = \prod_{i=1}^S \frac{1}{\sqrt{(2\pi)^{N_i} e^{2w_i} |\mathbf{C}_i|}} \exp \left\{ -\frac{1}{2} \mathbf{r}_i^T (e^{2w_i} \mathbf{C}_i)^{-1} \mathbf{r}_i \right\}, \quad (B2)$$

where each type of surface wave dispersion data is indexed with i among S different types. The residual vector between observed and predicted data is indicated by $\mathbf{r}$, and the superscript T is transpose of the vector. The data covariance matrix is represented by $\mathbf{C}_i$ for of i th data type with N_i data points. For the inversion of surface wave dispersion velocities in this study, only diagonal terms of $\mathbf{C}_i$ (data variances) are considered assuming uncorrelated data noise. The data covariance matrix of each data type is scaled by an exponential function of w_i (scaling factors shown in Figure 9), which is a hyperparameter of i th data type for the hierarchical scheme. The vector $\mathbf{r}_i$ contains residuals between observed and predicted data vector.

The 1-D Voronoi-cell approach [Bodin *et al.*, 2012] is adopted to parameterize V_S and V_P/V_S ratio depth profiles. Hence, the inversion method samples for V_S and V_P/V_S ratio of each cell, the position of cells in depth, the number of cells (dimensional parameter, see Figure 8), and the scaling factors. Empirical relationships between V_P and density [Brocher, 2005], isotropy, and elasticity are assumed.

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