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Supporting Information for

Intraplate volcanism controlled by back-arc and continental structures in NE Asia inferred from trans-dimensional ambient noise tomography

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Introduction

The supplementary material contains three sections of text and twelve figures. The first section describes the data processing and the measurement process of the Rayleigh wave fundamental mode of group and phase velocities, the second section presents the procedure of the conventional regularized inversions which are compared with results of the Bayesian inversions presented in the main text, and the third section discusses sensitivities of presenting dispersion maps in different periods to shear wave velocities with depths.

Text S1. Further details on the measurement of dispersion velocities

For continuous broadband data of each station, instrument response, mean, and trend are removed with a frequency-limit between 0.01 and 0.40 Hz. The corrected data are decimated (to 1 sample/s) and windowed to 1-day segments with 50% overlap. To obtain high-quality cross-correlation functions (Figure S2), we apply the frequency-time normalization (FTN) method [Shen et al., 2012]. For the narrow bandpass filtering in the FTN process, a second-order Butterworth filter is applied with 0.01 Hz bandwidth. Finally, about 50,000 inter-station Green’s functions are obtained by stacking cross-correlation functions of individual one-day segments for all available station pairs.
In this study, group and phase velocities of fundamental-mode Rayleigh wave are measured for periods from 8 s to 70 s with 2-s increment. We use a multiple-filter technique (Figure S3), which is now well established for ambient noise data [e.g., Bensen et al., 2007; Lin et al., 2008; Yao et al., 2005]. Phase velocity measurements are obtained based on Lin et al. [2008] and visually checked by comparing with velocity-period diagrams from the method of Yao et al. [2005] (Figure S3c). We measure dispersion curves guided by reference group and phase velocity curves. From starting reference curves obtained from the model LITHO1.0 [Pasyanos et al., 2014], the path-specific curves are iteratively updated by a series of dispersion measurements and 2-D map estimations using a regularized tomographic inversion (described in Text 2). Finally, we apply a spline interpolation technique to discretely picked-velocity data to obtain smoothed measurements (Figure S3).

Recording time-periods of our datasets include volcanic activity (i.e., Aso volcano in Kyushu in Japan), which releases persistent microseism energy [Zeng and Ni, 2010]. To avoid the interference of the signal (Figure S2), we apply a similar approach to Zheng et al. [2011], which exploits only the opposite lag of the cross-correlation function to produce the Green's functions when the signal arrives within the measurement window (2.0-4.5 km/s; Figure S2). Our data clearly show that the signal is dominant only in the period shorter than 18 s (Figure S2). Hence, this procedure is applied within the period ranging between 8 and 18 s. Both the causal and acausal lags are used in periods longer than that (Figure S3). In contrast, our data show no visible signal overlap from the 2011 Tohoku-Oki earthquake and its aftershocks (Figure S2b and c).

To control the quality of dispersion measurements, we select data with the signal-to-noise ratio (SNR) higher than 5. In addition, data are rejected when inter-station distances are shorter than 3 wavelengths. The number of selected data yields the maximum (approximately 45,000) at 18 s, and gradually decreases with increasing period (Figure S4a).

To assess effects of temporal variations in our dataset, we calculate cross-correlation functions and measure dispersion velocities using common station-pairs in different recording time-intervals (2009-2011 and April-October in 2014). Velocity differences between different recording periods show great consistency with zero means and small standard deviations (<0.06 km/s) in all periods (Figure S4b and c).

Text S2. Comparison with regularized inversions

To compare with the results from the Bayesian inversions, we perform regularized tomographic inversions [Rawlinson and Sambridge, 2003] for the same synthetic and observed data. We use the subspace inversion technique [Kennett et al., 1988] with the fast marching ray-tracing algorithm [Rawlinson and Sambridge, 2004]. In the inversion, the objective function consists of a fitness term (i.e. the Mahalanobis distance with a data covariance matrix), a damping term (a model covariance matrix), and a smoothing term [Rawlinson and Sambridge, 2008]. Hence, data and model covariance matrices are prior information in this inversion. However, it is unlikely that those matrices are known a priori in many geophysical inverse problems [e.g., Dosso and Wilmut, 2006]. To form the data covariance matrix (here we assume a diagonal matrix with the same variance values for all data points), we use the pick values of the estimated posterior distributions of the data noise parameter in our Bayesian inversions (Figure S5). This diminishes the effects of the data term in the comparison of the different methods. Due to the damping parameter scales the model covariance matrix, we assign arbitrary standard deviations for the matrix in phase and group velocity inversions (0.1 km/s and 0.2 km/s, respectively). To make the problem simple, we avoid the smoothing term by
using zero for the smoothness parameter. Consequently, the damping parameter is the only factor controlling regularization in the inversions.

To follow a conventional practice of determining regularization parameters, we perform a trade-off test to find an optimum balance between model roughness and data fitness (Figure S6). We use a uniform starting model for each period with the mean of velocity measurements (Figure S4a). Results of the inversions are presented in Figure S7 for synthetic recovery experiments and real data.

**Text S3. Depth sensitivity estimations of dispersion velocities**

In general, a meticulous inversion procedure is necessary to estimate exact depth profiles of physical properties. It is because the relationship between surface wave dispersion velocities and shear wave velocity structure is non-linear (i.e., sensitivity kernels as functions of the velocity structure itself). It is possible to estimate the partial derivative profiles of dispersion velocities with respect to shear velocity perturbations by assuming a specific structural model [Pasyanos et al., 2014]. Figure S12 shows calculated sensitivity profiles for selected locations of IPVs (Mt. Baekdu, Jeju and Ulleung) for periods of 12, 20, 40 and 50 s. In addition, we expect the sensitivities of the phase and group velocity maps to be related to sedimentary basins [Laske and Masters, 1997], thickness of crust and lithosphere [Pasyanos et al., 2014], and the vertically polarized shear wave velocity model at shallow upper mantle depths (<150 km) [Kustowski et al., 2008].

**References**


Figure S1. Density maps of (a) all inter-station paths used in this study and (b) a part of paths excluding the second group of stations (China regional networks during April-October in 2014). Solid lines and red triangles indicate tectonic boundaries and intraplate volcanoes, respectively (see Figure 1).
Figure S2. Examples of calculated cross-correlation functions between (a) TJN station and the F-net stations (from the three-year waveform dataset), and (b) SSE station and three regional stations in the Northeast China (from the seven-month waveform dataset). Top panels show inter-station ray-paths with locations of potential directional sources by volcanic activities (red star) and 2011 Tohoku earthquake (blue star). Bottom panels show cross-correlation functions with respect to inter-station distances for two different frequency ranges as indicated at the top of each panel. The expected arrival times (assuming apparent velocity of 2.5 km/s) of the directional sources are indicated by small circles with the same color-code as in the top panels. The measurement window corresponding to velocities between 2.0 s and 4.5 km/s, is marked by gray lines.
Figure S3. An example of the phase and group velocity measurement between STN in Japan and LYT in NE China. (a) SNR for the period range of 8–70 s (black line). The minimum SNR criterion for the selection of data is indicated by red line. Note that the opposite lag of the cross-correlation function is used in periods shorter than 18 s (dashed line) to avoid the arrival of the microseism signal from the volcano. (b) Frequency (period)-time (velocity) diagram for the group velocity measurement from the multiple filter analysis. The preliminary pickings of group arrival times are shown by white dots, and the black line indicates the spline-smoothed curve. The reference curve is presented by gray-dashed line. (c) The same plot as in (b) but for phase velocity. All of possible phase velocity measurements the method of Lin et al. [2008] are presented with white dots, and the 2-π ambiguity is resolved using the reference curve (gray-dashed). The background image is obtained based on Yao et al. [2005].
Figure S4. (a) Number of dispersion measurements for different periods (solid line). Average group (red dots) and phase (blue dots) velocities are presented by error bars (standard deviations). Histograms representing the difference between two different time-intervals for (b) group and (c) phase velocity data. Green, red, and blue histograms indicate the periods of 12, 30 and 60 s, respectively. Mean values of distributions are presented in the upper left corner with their standard deviations in brackets.
Figure S5. Examples of prior (shaded in gray) and posterior distributions (blue) of parameters for the noise level (left) and the number of cells (right) for group and phase velocities.
Figure S6. Examples of a trade-off test for selected regularized inversions of (a-c) 20 s group velocity and (d-f) 40 s phase velocity. Data used for the inversions are the checkerboard test synthetics (left), the multi-scale test synthetics (middle), and the real data (right). Selected damping values to produce results in Figure S7 are indicated by red diamonds.
Figure S7. Results of regularized inversions for (a) synthetic and (b) real data as shown in Figure 2 and 3, respectively.
Figure S8. Distributions of absolute residuals between the true and inverted maps from our Bayesian inversions as shown in Figure 2. Orange, blue and red (our preferred one) lines indicate boundaries of areas where the associated inversion uncertainties are smaller than 0.2, 0.3 and 0.4 km/s, respectively. Note that the resolution starts to decrease quickly at the defined threshold (0.4 km/s).
Figure S9. Tomographic results for subsets of data. Maps of phase velocity (left panel of each pair) and uncertainties (right panel of each pair) are presented with corresponding periods, mean velocities and numbers of data at the top of each map. Compare with Figure 3b. (a) Results using subsets of data selected within ±2 standard deviation of the data-residual distributions for all data presented in Figure 3b. (b) Same as (a), but now using a subset of data within ±1 standard deviations.
Figure S10. Group velocity maps and their uncertainties for selected period in the range of data measurement (8-70 s), excluding periods shown in Figure 3.
Figure S11. The same as in Figure S10, but for phase velocity.
Figure S12. Calculated normalized depth sensitivities for selected locations of IPVs: (a) Mt. Baekdu (Changbai), (b) Jeju island and (c) Ulleung island (see Figure 1). The left panel of each location shows shear wave velocity profile from Pasyanos et al. [2014]. The right panel presents 12 (red) and 20 s (red dashed) of $\partial U/\partial VS$ and 40 (blue) and 50 s (blue dashed) of $\partial C/\partial VS$. Where U, C and VS indicate group, phase and shear wave velocities, respectively. Shaded areas in yellow, blue and red indicate crust, mantle-lid and sub-lithospheric upper mantle (asthenosphere), respectively.