Temporal change of phase velocity beneath Mt. Asama, Japan, inferred from coda wave interferometry

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[1] We estimated the temporal changes of phase velocity of Rayleigh waves extracted from cross correlations of S-coda waves recorded at 12 stations around Mt. Asama, Japan. First, we extracted a Rayleigh wave by taking cross correlations of S-coda waves for 315 regional earthquakes between October 2005 and February 2009. The dispersion curve of the Rayleigh wave was measured and compared with the one extracted from 18 days of ambient seismic noise. We found that both dispersion curves are consistent with each other, demonstrating the dominance of the fundamental Rayleigh waves. We then divided the entire time period into sub-periods, each of which consists of 80 earthquakes, to measure the temporal changes at frequencies between 0.3 and 0.6 Hz. The result shows the reduction of phase velocity by 1.5 % and the subsequent recovery before the eruption of Mt. Asama in 2008. Citation: Nagaoka, Y., K. Nishida, Y. Aoki, and M. Takeo (2010), Temporal change of phase velocity beneath Mt. Asama, Japan, inferred from coda wave interferometry, Geophys. Res. Lett., 37, L22311, doi:10.1029/2010GL045289.

1. Introduction

[2] Probing temporal changes of the subsurface structure of active volcanoes is a key for understanding the mechanics of volcanic eruptions. One possible solution is to conduct a time-resolved seismic tomography using local earthquakes [Gunasekera et al., 2003; Patanè et al., 2006], but seismic tomography inherently involves uncertainties too large to detect subtle changes. Repeating active source seismic experiments can extract small subsurface changes by taking advantages of known source location and origin time [Nishimura et al., 2000, 2005], but it is generally too costly to be a realistic solution.

[3] Recent studies have revealed that cross correlation of seismic random waveform, such as ambient seismic noise or seismic coda, is capable of estimating seismic structure of the subsurface. Ambient seismic noise is generated by ocean waves, and seismic coda is excited by multiple scattering due to the inhomogeneity in the lithosphere [e.g., Sato and Fehler, 1998]. The basic concept of this method dates back to a pioneering work of Aki [1957], and more recently Weaver and Lobkis [2001] and Campillo and Paul [2003] showed that the cross correlations of random waveform recorded at two receivers can be represented as if the source is at one receiver and the recorder is at the other. This technique, now called seismic interferometry, is suitable for exploring subsurface structure; for example, Shapiro et al. [2005] constructed a tomographic image under California from ambient seismic noise, and Tonegawa et al. [2009] described the oceanic Moho within the Philippine Sea slab by extracting body waves from the S coda.

[4] Seismic interferometry is also suitable for detecting temporal changes of local internal structure because the wavefield extracted from cross correlations is sensitive to the internal structure between two receivers. For example, Wegler et al. [2009] found a sudden drop of S-wave velocity after an earthquake using Rayleigh waves extracted from ambient seismic noise. Brenquiel et al. [2008] monitored the seismic velocity perturbations in the interior of Piton de la Fournaise, La Réunion island, using Rayleigh wave extracted from ambient seismic noise. Seismic velocity changes were also measured beneath Pavlof volcano, Alaska, with coda waves [Haney et al., 2009].

[5] Ambient seismic noise, recorded almost anywhere at any time, takes advantages of the large amount of data, but also has a disadvantage of seasonal variation of the excitation sources as a source of spurious measurements of temporal changes [e.g., Stehly et al., 2006]. In contrast, the amount of coda records is limited since they are generated from earthquakes, leading to the difficulty in gaining enough signal-to-noise ratio by stacking. However, the excitation source of coda is not expected to exhibit seasonal changes so that more accurate measurements of temporal changes could be possible. To avoid the possible biases of the solution by employing ambient noise as described above, here we applied seismic interferometry using coda waves to estimate the temporal change of subsurface structure of Mt. Asama between October 2005 and February 2009. Note that the subsurface of Mt. Asama is highly inhomogeneous with short mean free path [Yamamoto and Sato, 2010], which is advantageous in using coda waves because the waveform becomes equipartitioned quickly after the direct arrival [Hennino et al., 2001; Paul et al., 2005; Campillo, 2006].

2. Mt. Asama


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data during the most recent episode in 2008–2009 suggests that a deeper dike to the west of the summit and a shallower source are responsible for the observed deformation [Aoki et al., 2009c]. Within the framework of these studies the detection of temporal changes in subsurface seismic velocities is expected to provide additional insights into the dynamics of magma transport and resulting eruptions of Mt. Asama.

3. Data Processing

[7] We used vertical seismic records of regional earthquakes at 12 stations (five broadband stations and seven short period stations with a natural frequency of 1 Hz) deployed around the summit crater of Mt. Asama (Figure 1a). We use instrument-corrected data recorded in the frequency band 0.1–1 Hz. We chose 315 regional earthquakes between October 2005 and February 2009 (Figure 1b) with magnitude larger than 4.5 determined by the Japan Meteorological Agency. The focal depth of these earthquakes is deeper than 20 km with epicentral distances less than 10 degrees. Note that the choice of deep earthquakes is to avoid influence of ballistic surface waves in the latter part of the seismogram and the choice of regional earthquakes is to retain the high frequency content in the observed seismogram.

[8] Figure 1c depicts a typical example of an observed seismogram. It is characterized by direct P and S arrivals followed by long-lasting coda waves, which reflect the strong scattering due to the inhomogeneity beneath Mt. Asama [Yamamoto and Sato, 2010]. We used S coda starting from 10 seconds after the direct S wave arrival with a time window of 80 seconds.

[9] We computed the cross spectrum of the S coda for 45 station pairs (Figure 1a) averaged over the entire set of earthquakes as

$$\overline{F_{i}^{obs}(\omega)} = \frac{\overline{F_i}}{|F_i| |F_j|^*},$$

where $F_i$ is the Fourier spectrum of the coda of the $i$th station and $\overline{F_{i}^{obs}(\omega)}$ is the observed cross spectrum at the angular frequency $\omega$ for the $i$th pair of stations (the $i$th and the $j$th stations). Here the overline denotes the average over the entire set of earthquakes and * denotes complex conjugate. Note that we disregarded the amplitudes by normalization. Cross-correlation functions were obtained from the inverse-Fourier transform of the cross spectra. The cross-correlation functions were lined up along their separation distances in the frequency range of 0.3–0.6 Hz. Figure 1d shows the propagation of the seismic wave with an apparent velocity of about 2 km/s, which is consistent with the S-wave velocity of the subsurface structure inferred from an
Typical examples of fitting the Bessel function (solid curve) to the observed cross spectra (solid circles) (a) at 0.3 Hz and (b) at 0.5 Hz. (bottom) Plot of variance reduction \( V_R = 1 - \Delta(A_0(k; \omega), k; \omega)/\sum \Phi_0^{\alpha}(\omega)^2 \) against frequency and phase velocity (c) for the vertical-component S coda and (d) for the vertical-component ambient seismic noise. Dispersion curve can be drawn by picking the phase velocity maximizing \( V_R \) for each frequency.

Figure 2. (top) Typical examples of fitting the Bessel function (solid curve) to the observed cross spectra (solid circles) (a) at 0.3 Hz and (b) at 0.5 Hz. (bottom) Plot of variance reduction \( V_R = 1 - \Delta(A_0(k; \omega), k; \omega)/\sum \Phi_0^{\alpha}(\omega)^2 \) against frequency and phase velocity (c) for the vertical-component S coda and (d) for the vertical-component ambient seismic noise. Dispersion curve can be drawn by picking the phase velocity maximizing \( V_R \) for each frequency.

4. Measurement of Dispersion Curve Inferred From Coda Waves

To estimate the phase velocity of seismic waves extracted by the cross correlations averaged over the entire set of earthquakes, we measured a dispersion curve under an assumption of wave propagation in a horizontally-layered medium. This average dispersion curve will be used as the reference dispersion curve in estimating the temporal change of the phase velocity.

Aki [1957] showed that dispersion curves can be deduced from the cross spectra between pairs of stations of random and stochastic wave fields, which is known as the spatial autocorrelation method. Considering isotropic incidence of the waves and using vertical-component data, the observed cross spectra at angular frequency \( \omega \) against the separation distances can be approximated by the Bessel function as \( A_0(k; \omega) \) (Figures 2a and 2b), where \( A \) is the power spectrum, \( r \) is the separation distance, \( k \) is the wavenumber. Here \( J_0 \) is the Bessel function of zeroth order.

We measured the dispersion by fitting the Bessel function to the observed average cross spectra \( \Phi_0^{\alpha}(\omega) \) [Nishida et al., 2008]. Least square error \( \Delta \) can be written as

\[
\Delta(A, k; \omega) = \sum_{\alpha} \left[ \Phi_0^{\alpha}(\omega) - A J_0(k r_\alpha) \right]^2,
\]

where \( r_\alpha \) is the separation distance of the \( \alpha \)-th pair. In order to find the optimal wavenumber \( k_0(\omega) \) which minimizes \( \Delta(A, k; \omega) \) for each angular frequency \( \omega \), we applied a two-step search. In the first step, for each \( k \) and \( \omega \) we estimated the power spectrum \( A_0(k; \omega) \) by minimizing \( \Delta \) as \( \partial \Delta/\partial A = 0 \). In the second step, we used a grid-search over \( k \) to determine the minimum of \( \Delta(A_0(k; \omega), k; \omega) \) (Figure 2c). We can obtain the phase velocity as \( \tau_0(\omega) = \omega/k_0(\omega) \) by employing the optimal wavenumber \( k_0(\omega) \).

[13] Figure 2c shows that the observed wave has a phase velocity of about 2 km/s at around 0.3 Hz and has lower velocities at higher frequencies. Considering a typical velocity structure of Mt. Asama [Aoki et al., 2009a], this dispersion curve is for the fundamental Rayleigh wave, which is sensitive to the velocity structure from the surface down to the depth of about 1 km.

5. Comparison With Dispersion Curve Inferred From Ambient Seismic Noise

If the measurement of the dispersion curve from a limited amount of coda record is comprehensive enough to agree with that from the data-rich ambient noise, then the coda waves are more suitable for estimation of the temporal change because they are not influenced by the seasonal variation of the excitation source. Here we examined the quality of the dispersion curve extracted from coda wave in comparison to that from ambient noise.

[15] The spatial autocorrelation method can also be applied to the ambient seismic noise assuming its stochastic and stationary excitation. We obtained a dispersion curve from 18-day-long ambient seismic noise records (vertical component) between March 14 and 31, 2006 at 45 pairs of stations (Figure 2d).

[16] Figures 2c and 2d show that both dispersion curves are consistent with each other. They show the dominance of the fundamental Rayleigh waves. The dominance of Rayleigh waves for coda waves is presumed to be due to the excitation of surface waves by inhomogeneous structure beneath Mt. Asama [Yamamoto and Sato, 2010]. The similarity of the two dispersion curves (Figures 2c and 2d) supports our use of coda waves to estimate the temporal change of the phase velocity. A difference is that coda waves retain larger energy than the ambient noise at around 0.1 Hz, which indicates that the coda recorded by short period seismometers can be applied in a lower frequency band.

6. Temporal Change of Phase Velocity

To estimate the temporal change of the phase velocity using coda waves, we divided the entire time period into sub-periods, for each of which we measured the perturbation of the phase velocity with respect to the reference dispersion curve. The length of the sub-period is determined by the number of earthquakes; each sub-period consists of 80 earthquakes with an overlap of 79 earthquakes. The total number of the sub-period examined in this study is 236, and the averaged length of them is approximately eight months.

For the \( \alpha \)-th sub-period, we computed the observed cross spectra \( \Phi_0^{\alpha}(\omega) \), and then measured the dispersion curve by minimizing a least square error \( \Delta' \), where \( \Delta' \) is written as:

\[
\Delta'(A, k; \omega) = \sum_{\alpha} \left[ \Phi_0^{\alpha}(\omega) - A J_0(k r_\alpha) \right]^2.
\]
At first, for each $k$ and $\omega$ we estimated the power spectrum $A_i^0(k;\omega)$ by minimizing $\Delta^i$ as $\partial \Delta^i / \partial A = 0$.

In order to estimate the phase velocity change averaged over the entire volcano, we shifted the dispersion curve of each sub-period against the reference dispersion curve with a constant rate for the frequency band from 0.3 Hz to 0.6 Hz. We summed the normalized least square error over the frequency band from 0.3 Hz to 0.6 Hz, as:

$$S_i^o(C_1) = \sum_{k} \sum_{\omega} \left( \frac{A_i^0(k;\omega) - \hat{A}_i^0(k;\omega)}{\hat{A}_i^0(k;\omega)} \right)^2 / \sum_{k} \sum_{\omega} \left( \hat{A}_i^0(k;\omega) \right)^2$$

where $\delta$ is the rate of shift of the wavenumber against the reference $\hat{A}_0$. The optimal rate of shift for the $i$th sub-period $\delta^i$ was estimated by minimizing $S_i^o(\delta)$. We converted the optimal rate of shift of the wavenumber $\delta^i$ into that of the phase velocity (i.e., the relative change of the phase velocity).

Figure 3a shows the relative change of phase velocity. Standard deviation of the rate of velocity change is estimated to be at most 0.5 % since the perturbations between adjacent sub-periods are about ±0.5 % in 2006 when significant volcanic activities are not expected. We found velocity decrease of about 1.5 % in the latter half of 2007 and velocity recovery toward the eruption in 2008.

7. Discussion and Conclusions

During the 2008 and 2009 eruptions, ground deformation, as well as the phase velocity changes, was also observed (Figure 3). The contraction of the KAHG-AVOG baseline (Figure 3c) is due to volumetric expansion right beneath the summit crater, and the extension of the KVCO-TASH baseline (Figure 3d) reflects the dike intrusion to the west of the summit. The observed time series (Figures 3c and 3d) show the onset of deformation around July–August 2008, slightly before the first eruption in 2008. Our result indicates that the reduction in phase velocity occurs much earlier than the ground deformation. This would indicate that the velocity changes are the results of something other than the mass transport detected by deformation measurements. This fact also implies that the velocity changes could be an independent measure to probe the volcanic processes at shallow subsurface.

This study reveals that phase velocity changes can be inferred using coda waves just like using ambient seismic
noise. It thus offers a possibility that we can trace velocity changes back to the past when continuous seismic data does not exist but only event-triggered data is available.

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References


