Seismic imaging of magma chamber beneath an active volcano

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Abstract

While the location and shape of magma chambers beneath active volcanoes play a key role in understanding magma transport and forecasting volcanic activity, the nature of magma chambers, particularly their shape, is not fully understood. Here we found a low velocity body too small to be detected from conventional techniques by the aid of a modern technique called seismic interferometry. Combining our result with independent observations suggests that the low velocity body is likely to represent a magma chamber. Our findings demonstrate the utility of seismic interferometry in imaging a small scale feature with a size of less than 10 km.

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1. Introduction

When magma rises up from depth, it is usually stored at certain depths where neutral buoyancy acts to the magma (e.g., Schmincke, 2006). Precise knowledge of the location and size of such storage, generally called magma chamber, in active volcanoes is important to understand and forecast the behavior of eruptions in that the style of volcanic eruptions depends partially on the style of differentiation of magma, which depends on the pressure, temperature, thus depth, and the size of the magma chamber at its residence (e.g., Petford, 2003). However, they are poorly understood mainly because the imaging capability with conventional local earthquake tomography is not good enough to image magma chambers (e.g., Lees, 2007); it is rare to obtain optimum earthquake distribution for local earthquake tomography in volcanic regions and it is difficult to obtain reliable first arrival picks of S-waves, which is more sensitive to the existence of magma chamber than P-waves. Here we find a robust low velocity region with a radius of about 5 km at several kilometers to the west of the summit. It is found from Rayleigh wave phase velocity maps at a frequency of 0.1–0.2 Hz derived from seismic ambient noise records, a technique emerging rapidly (Shapiro et al., 2005; Brenguier et al., 2007; Nishida et al., 2008, 2009). The S-wave velocity structure derived from the phase velocity maps (Rawlinson and Sambridge, 2003) indicates that the low velocity zone is limited to depths at 5 km or deeper. Combining our result with seismic exploration results and geodetic data shows that the inferred low velocity region represents a magma chamber, which is detected with such high resolution for the first time.

Mt. Asama is one of active volcanoes in central Japan (Fig. 1) with explosive eruptions with Volcano Explosivity Index (VEI; Newhall and Self, 1982) of 5 or more in 1108 and 1783. Recently, moderate-sized eruption with a VEI of 2 occurred in 2004 and minor ones in 2008 and 2009. With this background, Mt. Asama has been well-instrumented with seismometers and Global Positioning System stations. Shallow magma plumbing system is well defined with this wealth of data; during the 2004 eruption, magma-filled cracks (dikes) intrudes to the several kilometers to the west of the summit preceding the summit eruption (Takeo et al., 2006). An active source seismic experiment imaged the area of the dike intrusion by a high P-wave velocity anomaly (Aoki et al., 2009) which is interpreted to be the remnant of repeating intrusions, indicating the persistence of the inferred magma pathway.

While the active source seismic experiment is capable of resolving shallow seismic structure, it does not resolve any structures deeper than 3–4 km below sea level (Aoki et al., 2009). On the other hand, a regional seismic tomography study images broader-scale structures from the lower crust to upper mantle (Nakajima and Hasegawa, 2007) but is not capable of imaging upper crust in details, where magma chambers likely exist. Here we fill the gap to image the seismic structure of the upper crust beneath Mt. Asama using seismic ambient noise.

An idea of seismic imaging with random signals such as ambient noise dates back several decades (Aki, 1957) but it is...
only around the turn of the 21st century that it is started to be practically applied for acoustics (Lobkis and Weaver, 2001), helioseismology (Duvall et al., 1993), atmospheric infrasound (Haney, 2009), and seismology (Shapiro and Campillo, 2004). Ambient noise has extensively been employed to image the interior of the Earth in the local (Benguier et al., 2007), regional (Shapiro et al., 2005; Lin et al., 2008; Nishida et al., 2008), and global (Nishida et al., 2009) scales. The principle of imaging with ambient noise is that the cross-correlation of ambient noise recorded at two stations represents the wavefield as if a source is at one station and a receiver is at the other. The obtained wavefield thus takes advantages of the sensitivity to the structure only around the station pair.

2. Data analysis

We used vertical seismograms recorded between July 2005 and July 2007 at 89 stations, 18 of which are broadband sensors and 71 short-period with a natural frequency of 1 Hz. We take advantages of better station coverage by choosing this time period when 6 broadband and 11 short-period sensors were temporarily deployed. We first divided the whole region into three subregions (Fig. 1) according to a priori knowledge from a previous study (Nishida et al., 2008) which shows a clear velocity contrast with slower northwest and faster southeast. We then took cross-correlations of all possible pairs within each subregion with appropriate preprocessing including instrument corrections, mean removal, detrend, and spectral whitening, as is done by previous studies (Bensen et al., 2007).

Fig. 2A–C depict cross-correlations as a function of interstation distance for each region. They show the propagation of the Rayleigh wave trains in all regions but with different speed; the

Fig. 1. Spatial distribution of seismic sites used in this study. Three regions for phase velocity measurements are circled. The location of Mt. Asama is also shown. Geographical location of the area of this study is shown by a solid rectangle in the inset.

Fig. 2. An example of paste-ups of cross-correlations and variance reductions ($VR$) in phase velocity measurements as a function of frequency and phase velocity. Top panels are an example of paste-ups with a band pass frequency between 0.1 and 0.4 Hz for (A) region 1, (B) region 2, and (C) region 3, respectively. The bottom panels show variance reductions for phase velocity measurements with various frequency and phase velocities for (D) region 1, (E) region 2, and (F) region 3, respectively.
main wave packet travels by about 3 km/s in region 2, much faster than that in other regions of about 2 km/s, consistent with previous studies (Nishida et al., 2008). Note that wave packets travel shorter distance in region 3 (Fig. 2C) than the region 2 (Fig. 2B) because the region 3 is likely to exhibit more attenuation and scattering by thicker sediment as suggested by low velocity bodies (Nishida et al., 2008). Visual inspection of Fig. 2 also shows that the causal and the acausal parts of the cross-correlations are similar, endorsing the quality of the obtained cross-correlations.

In each region, reference dispersion curves were measured with assumptions (1) the seismic structure within the network can be approximated to vary only with depth, (2) the amplitude of the ambient noise is azimuthally isotropic, and (3) the phase of ambient noise is azimuthally uncorrelated. In that situation, the real part of the cross-spectrum \( \Phi^{re} \) (representation of the cross-correlation, the causal and acausal part stacked, in frequency domain) at given frequency \( \omega \) can be given by a function of the power spectrum \( A(c, \omega) \), interstation distance \( r \), and phase velocity \( c \) as (Aki, 1957; Chávez-Garcia and Luzón, 2005; Sánchez-Sesma and Campillo, 2006)

\[
\Phi^{re}(c, \omega) = A(c, \omega) J_0\left(\frac{c}{r}\right)
\]

where \( A \) and \( c \) are unknowns and \( J_0 \) represents the zeroth order Bessel function of the first kind. The reference dispersion curve for each region is defined as the variance reduction:

\[
V_R(c, \omega) = 1 - \frac{\sum_{ij} \left| \Phi^{obs}_{ij}(c, \omega) - \Phi^{re}_{ij}(c, \omega) \right|^2}{\sum_{ij} \left| \Phi^{obs}_{ij}(c, \omega) \right|^2}
\]

(2)

is maximized (Fig. 2D–F), where the subscript \( ij \) represents the pair of station \( i \) and \( j \), the superscript \( obs \) and \( syn \) represent the observed and calculated cross spectrums, respectively. Note that given \( c \) and \( \omega, A(c, \omega) \) is estimated by the least-squares method to maximize \( V_R \). The obtained reference dispersion curves depict that the phase velocity in region 2 is faster than the others, again consistent with a previous study (Nishida et al., 2008) and visual inspection of cross-correlations. The dense deployment of seismometers in region 1 allows us to measure the phase velocity up to 0.4 Hz, higher than the other regions. The less dense deployment in regions 2 and 3 limits us to measure reliable phase velocities only up to 0.3 and 0.25 Hz, respectively.

3. Phase velocity map

Once the reference phase velocity is obtained, the travel time anomaly is measured as a perturbation from the reference phase velocity for all available station pairs (Fig. 3A–C). Phase velocity anomalies were measured for three frequency bands in region 1 (0.1–0.2, 0.15–0.3, and 0.2–0.4 Hz), three frequency bands in region 2 (0.1–0.2, 0.15–0.3, and 0.2–0.4 Hz) and one frequency band in region 3 (0.1–0.2 Hz), respectively.

We measured the phase shift between the observed waveform and the synthetic waveform from the reference dispersion curve in frequency domain. The optimum phase velocity perturbation \( \gamma \) for each given frequency band minimizes:

\[
S(B, \gamma) = \sum_{ij} \left( \Phi^{obs}(r, \omega, \gamma) - B(r, \omega) \right)^2 \]

(3)

Fig. 3. Perturbations in travel times with respect to the reference dispersions and the obtained phase velocities. (A) and (D) represent the travel time perturbations from those expected from the reference dispersions and the obtained phase velocities at 0.1–0.2 Hz. (B) and (E) are the same as (A) and (C) but for 0.15–0.3 Hz. (C) and (F) are for 0.2–0.4 Hz.
where

\[ \phi_{\text{syn}}(r, \omega, \gamma) = A(\omega) / (1 + \gamma) c_0(r) \]  \hspace{1cm} (4) \]

where we implicitly assume small perturbations, \( c_0 \) represents the reference phase velocity at frequency \( \omega \), and \( B \) represents the amplitude correction factor to be estimated. Introduction of \( B \) indicates that we discard the amplitude information.

For each path, the velocity anomalies \( \gamma \) were measured from the lowest frequency band. At the lowest frequency band, 0.1–0.2 Hz, we searched \( \gamma \) assuming that the perturbation is less than 10%. In subsequent frequency bands, we assumed the perturbation is between \( \gamma' = -0.05 \) and \( \gamma' + 0.05 \) where \( \gamma' \) is the perturbation at the previous frequency band. When \( B \) is estimated to be negative or no minimum is found within the search range of the phase velocity perturbation we judged that the measurement is failed and stopped measuring the phase velocity anomaly in the following frequency bands.

\[ \text{Fig. 3D shows that the obtained phase velocity for a frequency range 0.1–0.2 Hz is slower to the west of the volcano by up to 10%. Furthermore, the measured phase velocities between station pairs were inverted for two-dimensional phase velocity maps.} \]

We applied an iterative nonlinear inversion (Rawlinson and Sambridge, 2003) to estimate phase velocities of a given frequency range at grid points with a spacing of 0.03 degrees or approximately 3 km. Taking the phase velocity map from a previous study (Nishida et al., 2008) as an initial model, we iterated the inversion for ten times to get the final model. The spatial variations of the Rayleigh wave phase velocity at a frequency 0.1–0.2 Hz show a negative velocity anomaly by up to \( \sim 20\% \) to the \( \sim 10 \) km west of the summit (Fig. 3D).

Now a question arises as to whether the localized low velocity seen in the 0.1–0.2 Hz phase velocity map can be resolved from the inversion procedure. To address this question, we conducted a spike resolution test (Aster et al., 2005; Nolet, 2008) by the method of Rawlinson and Sambridge (2003). The result demonstrates that

\[ \text{Fig. 4. Results of a spike resolution test to demonstrate that the low phase velocity region we found is not numerical artifact but a real feature. (A) A spike resolution test with velocity perturbations at where seismic sites are dense. The left panel represents the input velocity perturbations. The baselines in which the cross-correlations are calculated are shown in Fig. 3. The right panel the inverted velocity anomalies. These panels indicate that the velocity perturbations in densely instrumented regions are well recovered. (B) Same as (A), except that the input velocity perturbations are in sparsely instrumented regions, indicating that the velocity perturbations in sparsely instrumented regions cannot be recovered.} \]
our data have enough power to resolve the low velocity of this size (Fig. 4). While the mapped low phase velocity at 0.1–0.2 Hz has a radius of about 5 km (Fig. 3D), the actual size would be smaller because the analysis in this study does not take the effect of finite-frequency into account. Under an assumption of infinite frequency, which is employed here, geometrical rays have a resolution to the

![Fig. 5](image-url)

**Fig. 5.** Sensitivity kernels for the density (green), S-wave velocity (red), and P-wave velocity (blue). (A) Sensitivity kernels at 0.15 Hz, (B) 0.225 Hz, and (C) 0.3 Hz. (For interpretation of the references to color in this figure caption, the reader is referred to the web version of this article.)

![Fig. 6](image-url)

**Fig. 6.** Shear velocity around Mt. Asama. Upper panels represent the plan view of the S-wave velocity at depths of 0 and 5 km, respectively. Lower panels represent the vertical view of the S-wave velocity along the A–B and C–D baselines, respectively, shown in the upper panels.
structure only along the raypath. In reality, however, rays with finite frequency has a resolving power to the structure of approximately one third of the first Fresnel zone width (Yoshizawa and Kennett, 2002; Nishida, 2011), which is approximately 7 km for the 0.1–0.2 Hz case. Because the actual velocity anomaly off the geometrical ray but within one third of the Fresnel zone will be mapped along the geometrical ray, the mapped velocity anomaly will be broader than the real one. We thus conclude that the radius of the velocity anomaly would actually be smaller than 5 km. There will be a trade-off between the size and amplitude of the velocity anomaly where a smaller-sized anomaly gives an anomaly of larger amplitude. However, the size of the velocity anomaly should not be much smaller than 5 km because the amplitude of the anomaly we obtained is already large (Fig. 3).

4. Three-dimensional S-wave velocity structure

The regionalized dispersion curves were inverted for local 1-D velocity models at each point (Tarantola and Valette, 1982; Montagner, 1986). Fig. 5 shows sensitivity kernels of the model of Nishida et al. (2008) or our initial model at 36 24’ N 138 27’ E, where a low phase velocity anomaly was found (Fig. 3) for 0.15, 0.225 and 0.3 Hz calculated by the method of Saito (1988) with an assumption that the P-wave velocity and density are scaled as a function of the S-wave velocity by Brocher (2005), indicating that lower frequency has more sensitivity to structures at depth. Fig. 6 shows the 3-D S-wave tomographic model from a collection of the local 1D S-wave velocity structures. This figure indicates low S-wave velocity at depths of 5–10 km. Such low velocity region to the west of the summit disappears at higher frequencies (Fig. 3E and F). Instead, Fig. 6 displays high S-wave velocity anomalies at shallower depths (<3 km) around the repeating dyking region inferred from an active source seismic exploration (Aoki et al., 2009). The S-wave velocity structure is consistent with the P-wave velocity derived from an active source seismic experiment (Aoki et al., 2009, Fig. 4). In addition, the inferred S-wave velocity at depth of sea level shows the slower phase velocity to the east of the summit than the west, which is also consistent with the result from an active source seismic experiment (Aoki et al., 2009).

Fig. 7 shows the difference among the measured phase velocities, initial phase velocities and corresponding S-wave velocities (Nishida et al., 2008), and the final phase velocities and corresponding S-wave velocities with their uncertainties at three different locations. Uncertainties of the phase velocities are taken to be 0.05 km/s. Choice of this value is rather subjective because it is difficult to make a rigid assessment of the errors in phase velocities when the estimation of cross-spectra as a function of interstation distance are not independent of each other. The absolute value of the uncertainties in S-wave velocities could thus be biased but the relative magnitude of those with depth will be preserved.

The mapped uncertainties of S-wave velocities are shown in the bottom panel of Fig. 7 as error bars which represent a posteriori covariance matrix of model parameters (Tarantola and Valette, 1982); they are on the order of 0.1 km/s at 2 km below sea level or deeper with higher uncertainties at shallower levels. This indicates that the low velocity body to the west of the summit is not a computational artifact but a robust feature. Also the comparison between initial and final models shows that the low velocity body is recovered from a initial model without such anomalies.

What makes the low S-wave velocity from 5 to 10 km? Here we suggest with some evidence that it is magma chamber. We observed tilt motions during the February 2009 eruption (Matsuzawa et al., 2009; Miyamura et al., 2009), much of which are obtained from the horizontal record of broadband seismometers by taking advantage of the capability of broadband seismometers in recording tilt motions (Rodgers, 1968; Graizer,
We excluded tilt records near site the summit with tilting up to 7 μrad toward the summit because they reflect the mass loss at the summit associated with the eruption, which is outside of our scope. Other sites generally tilt toward to the west of the summit by up to 1 μrad (Fig. 8A), implying a source offset to the west. We modeled the observed deformation field by a deflating dike embedded in a homogeneous, elastic, and isotropic halfspace (Okada, 1985) by applying Simulated Annealing (Cervelli et al., 2001), a Monte-Carlo based optimization of a nonlinear problem, to search for the optimum source geometry and amount of dike deflation. The optimum dike is at its top at about 1 km below sea level with northwest-southeast strike (Fig. 8A), the location of which is right at above the low S-wave velocity anomaly.

The existence of a dike participating in an eruption suggests that the low S-wave velocity anomaly represents the magma chamber beneath the dike deflating during the 2009 eruption. The crustal magma chamber has not been imaged at all (Aoki et al., 2009) or only as a broad low velocity zone (Lees, 2007) by conventional seismic imaging studies because of technical difficulties while our method is capable of delineating the size of a crustal magma chamber. Our results also demonstrate that ambient noise cross-correlations are capable of imaging an anomaly of a radius of ~5 km at a depth of 5 km or deeper with an optimal station coverage.

5. Discussion

Taken discussion above together and considering that the location of the deflating dike during the 2009 eruption almost coincides with that intruded during previous eruptions (Takeo et al., 2006; Aoki et al., 2009), Fig. 8B depicts the magma pathway in the upper crust beneath Mt. Asama. When magma rises up to the upper crust, it is stored at the magma chamber. As discussed above, the magma chamber is ~8 km offset to the west from the summit and represented by a low velocity region. Then, the magma further moves up to a depth of ~1 km below sea level as a dike. The shallow seismic and resistivity structures (Aizawa et al., 2008; Aoki et al., 2009) suggest that the magma pathway is subject to structural controls to make a winding path to reach the surface. Such interdisciplinary approach enables us to gain a unified understanding of the magma plumbing system of Mt. Asama from a crustal magma chamber to the surface.

Although many volcanoes have their magma chamber right beneath the summit (e.g., Masterlark et al., 2010), a magma chamber horizontally offset from the summit of a volcano could be a ubiquitous feature. In fact, tiltmeters in Kirishima volcano, southwest Japan, tilted toward a point ~8 km to the west of the summit in response to an eruption on 26 January 2011. This point coincides with the location of a deflation source derived from GPS observations. Although it is impossible to verify whether this deflation source is characterized by a low velocity body because of the lack in dense seismic observations, this tilt observations demonstrate the ubiquity of a horizontally offset magma reservoir and that the magma pathway of an active volcanoes is largely subject to structural controls to make a winding path to make the surface. Dense seismic deployment such as in Mt. Asama will enable the detection of such magma storages in other volcanoes, leading us to enhance the understanding of the mechanics of magma transport beneath active volcanoes.

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