Review

Recent volcanic activity at the Asama volcano and long-period seismic signals

By Minoru TAKEO,*1,*2,† Yosuke AOKI*2 and Takao KOYAMA*2

(Edited by Yoshio FUKAO, M.J.A.)

Abstract: Modern observation systems composed of seismic, geodetic, other geophysical, and geochemical networks developed in and around volcanic areas provide a mass of knowledge about volcanic activities. This paper summarizes the magma pathway and recent volcanic activity of the Asama volcano. The seismic velocity structure beneath the Asama volcano was investigated via seismic ambient noise tomography and active source seismic tomography. The magma pathway in the upper crust beneath the Asama volcano was synthesized by combining the velocity structure with a hypocenter distribution of volcanic earthquakes and ground deformations before and after eruptions. Temporal evolutions of multidiscipline data regarding the volcanic activity from October 2003 to January 2018 revealed that the supplied amount of magma from the magma chamber and the internal condition in the shallow regions of the conduit controlled the recent eruptions.

Keywords: Asama volcano, long-period seismic signal, magma pathway, temporal evolution of volcanic activity

1. Introduction

The Asama volcano is one of the most active andesitic volcanoes in Japan. It is located about 150 km northwest of the Tokyo metropolitan area. On the west, there is a row of ancient quaternary volcanoes, which are collectively called the Eboshi volcanoes. Volcanism appears to have progressed eastward, with the Asama volcano as its eastern end and the youngest member of the row.1) The Asama volcano itself is a complex of three volcanic edifices: Kurofu, Hotokeiwa, and Maekake (Fig. 1). Each one is a different eruptive type. Kurofu is the oldest large andesitic stratovolcano. Hotokeiwa is the second oldest medium-sized volcanic edifice with a rhyolitic to andesitic composition. The youngest volcano, Maekake, is a medium-sized andesitic pyroclastic cone with pyroclastic flows and lava flows.2)3) Maekake began to form just after the cessation of the activity at Hotokeiwa about 10,000 years ago. Its lava is mainly andesite, which is typical of arc volcanoes. Hence, its eruptions are sometimes explosive. The eruptive history of Maekake consists of three repetitions of active-dormant phases. The third active phase has a larger eruptive volume than the previous phases. The third active phase has been continuing about 2,000 years and includes the historical Plinian eruptions of the 4th century, 1108, 1128, and 1783. A Volcanic Explosivity Index (VEI)4) of 4 or more occurred in 1108 and 1783. Recent activity in the first half of the twenty century was high and the eruptions of VEIs were up to 3. In the past few decades, the eruptions have been less explosive. However, moderate-sized eruptions occurred in 1973, 1982, 1983, and 2004, and minor eruptions were recorded in 2008, 2009, and 2015.

Seismic observations in the Asama volcano have been collected for over 100 years. Since 2003 when a modern monitoring network was established around the Asama volcano, high-level observational quality data has been maintained, and several moderate to minor eruptions occurred in 2004, 2008, 2009, and 2015. This paper summarizes the magma pathway and recent volcanic activity of the
Asama volcano. These have been elucidated using a combination of seismic, geodetic, other geophysical, and geochemical data obtained from the modern network.

2. Overview of the observation network at the Asama volcano

The eruptions and seismic swarms in 1909 motivated seismic observations of the Asama volcano in 1910.5)–9) Omori established the Asama-yama seismological observatory in 1911 at Nagamine, which is about 2.3 km west-southwest from the summit crater.7) During the spring to autumn season annually from 1911 to 1916, he operated a portable two-component horizontal tremor recorder with a pendulum period of 4.2 s and a Gray-Ewing vertical component seismometer with a pendulum period of 10 s.8) This observation had finished once, but The Imperial University of Tokyo and the Central Meteorological Observatory took over the seismic observations and conducted constant observations until 1945. The Asama Volcano Observatory was established in 1933 by the Karuizawa town, which was later donated to The Imperial University of Tokyo in 1934. After World War II, the seismic observations were summarized by the University of Tokyo Earthquake Research Institute (ERI) and the Japan Meteorological Agency. To date, these entities have continued observations.

A high volcanic activity was observed at the turn of the 20th century until the 1950s. However, almost all the seismometers installed at the Asama volcano prior to the 1960s were short-period sensors, and the main recording systems were smoked-paper recorders. Thus, a major achievement during the early days of seismic observations was the classification of volcanic earthquakes based on the observed waveforms by Minakami.10) Although seismic observations in the Asama volcano have been collected for more than a century, there were only six seismometers in 1998. The monitoring network was modernized in 2003. Since then, it has rapidly grown. In 2017, it was comprised of 30 seismometers, 19 of which were equipped with broadband sensors (Fig. 1a). Continuous global navigation satellite system (GNSS) observations have been conducted in the Asama volcano region since 1996. As of 2017, there were 22 continuous GNSS sites within 20 km of the summit crater (Fig. 1b). In addition to seismic and geodetic sensors, other kinds of sensors such as microphones, video and thermal cameras, and a detector of chemical components in volcanic gas have been installed in the small caves located on the west and east rims of the summit crater.
3. Overview of the eruption activity from 2004 to 2015

The first moderate-sized eruption since 1983 occurred in 2004. It was preceded by elevated thermal activity at the summit crater, increments of SO2 emission beginning around 2000, and a few minor eruptions between February and April 2003.\textsuperscript{11,12} Figure 2 compares the temporal evolution of the seismic activity with the changes in the GNSS baseline length for sites 950221 and 950268 (see Fig. 1b) between 1996 and 2009. The activity of the volcanic earthquakes was low between September 1997 and March 2000, except for the last several months. Although the activity of volcanic earthquakes remained high after 2001, and the maximum temperature in the summit crater exceeded 200°C in autumn 2002.\textsuperscript{12} Continuous GNSS data showed that the north-south extension of the volcano started in late July 2004 and lasted about 5 weeks until the first eruption in 2004. Volcanic glows were observed in the last 10 days of July 2004, and the maximum temperature at the bottom of the crater exceeded 500°C after the sudden extension of the GNSS baseline length.

A seismic swarm started at 15:10 (JST) on 31 August 2004. More than 600 events occurred before the first eruption. The hypocenters of the swarm, which were determined by Yamamoto \textit{et al.}\textsuperscript{13} using a double-difference algorithm,\textsuperscript{14} were just below the crater and ranged from 300 to 800 m deep, suggesting that the shallowest part of the vent was broken during the swarm activity. A tiltmeter located about 3 km north from the summit started to tilt towards the east about 29 hours before the first eruption, implying an inflation source offset to the west.\textsuperscript{15}

At 20:02 (JST) on 1 September 2004, a Vulcanian eruption with a VEI of 2 occurred for the first time in 21 years. The ash plume reached 3.5–5.5 km above the summit or 6–8 km above sea level. After a short period of quiescence, Strombolian eruptions started on 14 and 16–18 September before Vulcanian eruptions occurred on 23 and 29 September, 10 October, 14 November, and 9 December.

The estimated total amount of ejected tephra was about $1.6 \times 10^6$ kg.\textsuperscript{11} The magma in the 2004 eruption was andesite with a composition similar to magmas erupted over the past 10,000 years. A lava dome emerged at the summit by 16 September when continuous Strombolian eruptions began.\textsuperscript{11} The change in the magma composition and the increasing amounts of juvenile magma in the eruptive products by that time suggests that juvenile magma may have ascended to the summit crater. The lava dome deflated afterward at its center, indicating that part of the ascended magma descended to the depth. Figure 3 exhibits the hypocenter distribution of volcanic earthquakes between 1 January 2004 and 24 May 2016 using the double-difference algorithm.\textsuperscript{14}

After 4 years of repose (see Figs. 2 and 3), the seismic activity started to elevate in July 2008, which was one month before the first eruption. The earthquakes were mainly located approximately 1 km to the west of the summit and about 1 km below sea level. The seismic activity further increased on 8 August, leading to eruptions on 10, 11, and 14 August. These were minor with ash plumes reaching heights of 200–400 m above the summit. The eruptive products did not include juvenile materials. From this, we inferred that the 2008 eruption was the ejection of the remnant magma associated with the 2004 eruption. Since September 2008, an elevated seismic activity has been observed approximately 2 km to the west of the summit and about 1 km below sea level (see Fig. 3).

On 1 February 2009, a day before the largest eruption in 2009, the seismicity just below the summit increased and a tiltmeter about 3 km from the summit inclined towards the east. This signaled inflation beneath the western flank of the volcano, which led to a summit eruption on the next day.
Subsequently, minor eruptions from the summit occurred intermittently on 9–16 February, 15 March, 30 April, and 3 and 27 May, with plume heights less than 500 m. The largest eruption during the 2008–2009 crisis occurred on 2 February 2009 but was smaller than that on 1 September 2004. The ejected mass due to the eruption on 2 February was 2.0–2.4 × 10^7 kg, about 20% of that associated with the 1 September 2004 eruption.16 The ejected materials included only trace amounts of juvenile materials, suggesting that the eruption was merely an ejection of near-surface materials rather than a magma migration from the depth.16

After more than 6 years of repose, the seismic activity gradually started to elevate in late April 2015, and the western side of the volcano turned to a north-south extension in May 2015.17 Tilt motions indicating inflation beneath the western side of the Kurofu mountain ridge started in the beginning of June, and the emission of SO₂ rapidly increased from 200 tons/day on 1 June to 1,700 tons/day on 11 June, leading to eruptions on 16 and 19 June. Both eruptions emitted only trace amounts of ash. However, the SO₂ emissions exceeded 300 tons/day and lasted until February 2016.

4. Magma pathway in the upper crust beneath the Asama volcano

This section reviews the synthesis of the magma pathway in the upper crust beneath the Asama volcano based on seismic and geodetic observations by referencing previous research.12,18–23 Takeo et al.12 and Aoki et al.23 compared the temporal variation in the monthly number of volcano-tectonic earthquakes to that of all volcanic earthquakes with changes in the GNSS baseline length between 1996 and 2011. The extensions in GNSS baseline length for sites 950221 and 950268 (see Fig. 1b) indicated a magma intrusion beneath the western flank of the Asama volcano because the baseline crosses the inferred dyking area, which was deduced from the geodetic data during the 2004 and 2008–2009 eruptions. The hypocenter distribution before 2004 demonstrated that the volcano-tectonic earthquakes occurred beneath the western side of the Asama volcano and other volcanic earthquakes occurred under the summit of the Asama volcano. Dyke locations associated with the baseline contractions before the 2004 eruption19 were also to the west of the summit beneath the eastern end of the Eboshi–Asama volcano group. Thus, it seems reasonable to assume that the overall trend of the hypocenter distribution and the dyke location has not changed in the last decade.

Figure 2 shows a three-fold increase in the number of volcano-tectonic earthquakes in the latter half of 1996, from October 2000 to April 2001, and from May 2002 to August 2002 before the 2004 eruption. These seismic activations coincided with the extensions of the GNSS baseline, suggesting that volcano-tectonic earthquakes are associated with the intrusion of magma beneath the western flank of the Asama volcano. The GNSS baseline contracted three times: from September 1997 to March 2000, from July 2001 to February 2002, and from March 2003 to April 2004. Because the observed contractions were too large to be explained by the tectonic motion of a non-volcanic origin, we speculate that it may indicate migrations of intrusive magma from beneath the western flank of the Asama volcano.

The seismicity between 1 January 2004 and 24 May 2016 (see Fig. 3) revealed a sharp image
composed of two groups. One formed a WNW-ESE trending zone at a depth ranging 0.5–1.5 km below sea level. The eastern end of this seismic zone was beneath the summit crater and extended westward horizontally about 2 km. The other group formed a narrow vertical seismic zone extending from the eastern edge of the former group to just below the summit crater. Aoki et al.\textsuperscript{18} and Takeo et al.\textsuperscript{12} modeled the ground deformation field between June 2004 and March 2005 by inverting a rectangular dyke size, its geometry, and its location in an elastic, homogeneous, and isotropic medium. Neither a spherical source nor a fault dislocation model could explain the ground deformation pattern. While the observed deformation was insufficient to satisfactorily constrain all of the model parameters, the horizontal location and the total volume of the induced dyke were relatively well constrained. The estimated intruded volume was $6.8 \times 10^9\text{m}^3$, which was about three times larger than the volume of emitted magma during the eruption.\textsuperscript{11}

The eastern part of the dyke overlapped with the western end of the hypocenter distribution (Fig. 3). The depth at the top of the dyke coincided with the depth range of volcano-tectonic earthquakes. The distribution of the dike-induced seismicity reflected the distribution of ambient stresses near failure. Thus, the seismicity may have a higher concentration near the dyke tip.\textsuperscript{24} The eastern end of the horizontally elongated seismicity was connected with the narrow vertical seismic zone of volcanic earthquakes extending from 1 km below sea level to the eruptive summit crater.

The western side of the volcano began to exhibit a north–south extension in July 2008, which was a month before the first eruption in 2008. The volcano expanded quasi-linearly until June 2009. The ground deformation field between June and December 2008 was well fitted by an east–west striking dyke with its top at 0.7 km below sea level and a shallow source responsible for the summit inflation.\textsuperscript{23} The east–west striking dyke was located near the one associated with the 2004 unrest, indicating that the dyke intrusion to the west of the summit was ubiquitous during unrest of the Asama volcano. These seismic and geodetic analysis results demonstrated that this distribution of hypocenters is along the magma pathway beneath the Asama volcano.

The precise earthquake locations and deformation fields could be used to delineate a sinuous magma pathway beneath the Asama volcano. Aoki et al.\textsuperscript{23} summarized an active-source seismic tomography and other geophysical observations to reveal the structural control of the magma pathway. An active seismic experiment was conducted in October 2006 with a dense deployment of temporary seismometers.\textsuperscript{20,21} The observed waveforms indicated that seismic waves that did not traverse the dyking area showed the maximum energy at first arrivals, while those traversing the dyking area displayed different waveforms with more energy in later arrivals. Using the dataset provided by the active seismic experiment, the estimated mean free paths for $P$ and $S$ waves were as short as about 1 km for the 8–16 Hz band. These values are much shorter than those for the usual Earth’s crust.\textsuperscript{25}

Using the same dataset, Prudencio et al.\textsuperscript{26} obtained intrinsic and scattering attenuation coefficients of the Asama volcano. The distributions of intrinsic and scattering attenuation anomalies indicated that the western part of the Asama volcano is characterized by high scattering and low intrinsic attenuation. This is consistent with the remnants of dyke intrusions. This feature is probably caused by the scattering of seismic waves in the dyking area where the seismic structure is likely to be inhomogeneous.

The $P$ wave velocity structure obtained by a travel time inversion clearly exhibited a high-velocity zone around the western side of Kurofu (see Fig. 1a) at a depth of 2 km below sea level and shallower, which is the area of the dyke intrusion during the 2004 eruption.\textsuperscript{21} The velocity to the west of the summit was faster than that to the east. Combining this with the velocity along the north–south profile suggested that the high-velocity zone strikes east–west with its top deepening to the east. The north–south extent of the high-velocity zone was as narrow as 5 km. This velocity structure was due to solidified magma, which resulted from repeated dyke intrusions because the location of the high-velocity zone roughly coincided with the dyking area associated with the 2004 eruptions.\textsuperscript{23}

This result is consistent with the insights obtained from other geophysical observations. The spatial distribution of the Bouguer gravity around the Asama volcano had east–west trending local maxima to the west of the summit, a high-velocity zone area, and the eastern end is at the summit. This high saddle-shaped gravity suggested that the high-velocity zone is a high-density zone. This is consistent with the hypothesis that the high-velocity zone is composed of dense rocks due to the slow solidification of repeatedly intruded magma.
The electromagnetic data showed that the high-velocity zone was less conductive than the surrounding area, which is also consistent with the hypothesis because slowly solidified magma has a low permeability, which causes a high resistivity. The hypothesis that the high-velocity zone is due to solidified magma implies that some of the magma should fail to reach the surface during eruptions. The volume of emitted magma was about one third of that of the intruded magma during the 2004 eruptions. Thus, we speculate that a significant portion of the intruded magma did not reach the surface in the Asama volcano. However, it was arrested and solidified in past eruptions as well as in the 2004 eruptions.

Nagaoka et al. revealed the seismic structure of the upper crust beneath the Asama volcano using seismic ambient noise. They first divided the whole region into three subregions according to a priori knowledge from a previous study. Rayleigh wave train propagations in all regions were obtained by taking the cross-correlations of all possible pairs within each subregion. In each subregion, reference dispersion curves were measured assuming the seismic structure varied only with the depth in each subregion. Once the reference phase velocity was obtained, the travel time anomaly was measured as a perturbation from the reference phase velocity for all available station pairs. The obtained phase velocity for a frequency range of 0.1–0.2 Hz was up to 10% slower to the west of the Asama volcano. Furthermore, they applied an iterative nonlinear inversion to estimate the phase velocities of a given frequency range at grid points with a spacing of 0.03° to create 2D phase velocity maps. The spatial variations of the Rayleigh wave phase velocity at a frequency of 0.1–0.2 Hz exhibited a negative velocity anomaly up to about 20% for 10 km to the west of the summit. A spike resolution test demonstrated that the data had enough power to resolve the low velocity of this size. While the mapped low phase velocity at 0.1–0.2 Hz had a radius of about 5 km, the actual size should be smaller because the analysis did not consider the effect of finite frequency.

The regionalized dispersion curves were inverted for local 1D velocity models at each point. Figure 4 shows a 3D S wave tomographic model from a collection of the local 1D S wave velocity structures. A low S wave velocity occurred west of the summit at depths of 5–10 km. At shallower depths (< 3 km), high S wave velocity anomalies were observed around the repeated dyking region, as inferred from the active seismic experiment. The S wave velocity is consistent with the P wave velocity derived from the experiment.

What makes the low S wave velocity change? Nagaoka et al. suggested that the magma chamber was responsible and provided some evidence to support this assertion. Tilt motions during the February 2009 eruption were recorded with tiltmeters and broadband seismometers around the Asama volcano. They excluded tilt records from sites near the summit because these data reflected the mass loss at the summit associated with the eruption, which was beyond their scope. Other sites generally tilted towards the west of the summit, implying a source offset to the west. They modeled the observed deformation field by a deflating dyke, where the optimum dyke was at its top about 1 km below sea level with a northwest–southeast strike, which was located above the low S wave velocity anomaly. The existence of a dyke participating in an eruption suggested that the low S wave velocity anomaly represents the magma chamber beneath the dyke defating during the 2009 eruption.

Figure 5 depicts the synthesis of the magma pathway in the upper crust beneath the Asama volcano based on the observations and the above
discussion. Magma rising to the upper crust was stored in the magma chamber at a depth of 5–10 km. The magma chamber was offset by approximately 8 km to the west from the summit and was represented by a low S wave velocity region. Then the magma moved to a depth of about 1 km below sea level as a dyke. The shallow seismic and resistivity structures suggested that the magma pathway is subject to structural controls, generating a winding path to the surface. Combining these multiple constraints provided a unified understanding of the magma plumbing system of the Asama volcano from a crustal magma chamber up to the surface.

5. Long-period (LP) seismic signals at the Asama volcano

Permanent broadband seismic observations near the summit crater started in October 2003. Two broadband seismic stations were installed on the west and east rims of the summit crater before the first eruption in 2004. This proximal broadband seismic network revealed that several kinds of unique LP seismic signals occurred intermittently, but their amplitudes were too feeble to be recorded outside of the Maekake crater.

Figure 6 shows a half-day monitoring record (12:00–24:00) of the broadband seismometer at KAW on 11 June 2004. Arrows indicate several kinds of unique long-period (LP) seismic signals.

Yamamoto et al. classified these unique waveforms into three categories: a very long-period (VLP) pulse,
an earthquake characterized by gradually attenuating an LP motion with upwards convex, and an LP tremor characterized by a waveform with sharpened tips.

The VLP pulse is an impulsive waveform with a dominant period longer than 5s accompanied by a tilt change. Maeda et al.\textsuperscript{30} and Maeda and Takeo\textsuperscript{31} applied a new waveform inversion method to elucidate the location and the source mechanism of a VLP pulse. Here, we summarize their results.

Section 5.1 focuses on the identity of a VLP pulse. Section 5.2 proposes a mathematical model of the unique LP earthquakes and tremors, and considers the meaning of these LP signals.

5.1. VLP pulse at the Asama volcano.

Figure 7 shows a typical waveform of a VLP pulse observed at the Asama volcano. The initial pulse shapes in the horizontal and vertical components were similar, while a much longer transient signal appeared only in the latter part of the horizontal component. Horizontal broadband seismograms were contaminated by a tilt motion, generating a gravitational force equivalent to an inertial force due to ground acceleration.\textsuperscript{32}–\textsuperscript{34} This gravitational force appeared in a vertical seismogram as the second-order term of the tilt angle, implying that it is negligible for ordinary microradian-order tilt motions.\textsuperscript{35}–\textsuperscript{37} Thus, this longer transient signal is attributed mainly to the tilt motion. VLP pulses still occur but the change of the VLP pulse activity coincides with the volcanic activity. Yamamoto \textit{et al.}\textsuperscript{35} found that similar impulsive pulses have occurred at least since September 2002 based on short-period seismograms recorded on the west rim of the summit crater.

As of August 2008, the summit area had four permanent broadband seismic stations (KAE, KAW, KME, and KMS). Ten temporary broadband seismic stations were installed inside the Maekake crater to reveal the location and the source mechanism of the VLP pulse. The temporal observations were carried out between September 2008 and April 2009. Maeda \textit{et al.}\textsuperscript{30} developed a new moment-tensor inversion method to deal with horizontal seismograms contaminated by tilt motions without any assumptions regarding the tilt motions. Their idea included the effect of tilt motions into Green’s functions instead of decomposing the horizontal seismograms into the translational and tilt responses.

Maeda and Takeo\textsuperscript{31} selected the VLP pulse at 17:18 (JST) on 28 October 2008 as a representative event and conducted inversion analysis. They applied the moment-tensor inversion method to the VLP pulses observed at the Asama volcano so that the time histories of the moment-tensor could be elucidated. The locations of the VLP pulses were surveyed using a grid search algorithm in which candidates of the source location were distributed in and around the summit crater at 50-m grid intervals and elevations ranging from 1850 to 2300 m above sea level. Many previous moment-tensor inversion studies at volcanoes have considered single forces along with the moment-tensor. However, the waveforms of the VLP pulses were well reconstructed without the single forces. Since the source parameters did not take the single forces into account, the resolution and the uniqueness of the solution were improved due to the reduction of the degree of freedom in the inversion problem.

Figure 8 compares the observed (gray lines) and synthetic (black dashed lines) particle motions at each station with the best source location (gray stars). Even after taking the uncertainty of the source location of 50–100 m in each direction into account, the source was located to the north of the crater center at a depth of 100–300 m below the summit. The components of the moment-tensor had similar time histories, which consisted of an initial rapid inflation phase and a subsequent gradual deflation phase. This similarity allowed a unique focal mechanism for the event to be determined.
Taking the peak amplitude of each time history as the values of the corresponding moment-tensor component, the focal mechanism of the VLP pulse was estimated. Maeda and Takeo performed moment-tensor inversions for 23 VLP pulses to examine the stability of the focal mechanism. Nineteen of the VLP pulses displayed moment-tensor time histories consistent among the components. Hence, the time-invariant moment-tensor solutions were obtained for these 19 events. The eigenvectors (principal axes) of these moment-tensor solutions showed only small variations. The eigenvalues (principal values) exhibited some variations, which depended on the amplitude. The ratio of the principal values (max:min) was 5:3:2, where max, min, and min indicate the maximum, intermediate, and minimum principal values, respectively. Assuming a Poisson material, this is interpreted as the principal ratio for a synchronized volumetric change of a tensile crack and a cylinder with symmetry axes oriented orthogonal to each other. The strike and dip angles of the tensile crack were both nearly 70°, and the symmetry axis of the cylinder was inclined about 30° from the vertical. The time histories of the moment tensor were indicative of an initial sudden pressurization followed by a gradual depressurization at the source region. They sought to model the VLP pulse at the Asama volcano by considering the inflow and outflow of gas at the source region, compensated by boiling of water or by exsolution of volatiles from a deeper reservoir.

Kazahaya et al. revealed a relationship between the VLP pulses and volcanic gas emissions through a combination of high temporal resolution SO2 flux measurements, vent process monitoring using a video camera set on the summit crater rim, and broadband seismic observations at the stations. The volcanic gas observations were conducted on 2 June 2009. Seven degassing bursts accompanied the VLP pulses. These bursts originated from the vent located at the crater bottom and were clearly recognized by the video camera. The seismic moments of VLP pulses and volcanic gas emissions were linearly related, strongly supporting the VLP pulse source model proposed by Maeda and Takeo.

The VLP pulses were characterized by a transient signal with a 10–20 s duration and a high-frequency oscillation of 5–10 Hz. Maeda et al. investigated the location and size of the high-frequency oscillation sources (HF sources) using the amplitude source location method. Most HF sources were deeper than the VLP pulse source by about 150 m. Additionally, the signals in two frequency bands of 0.00–0.1 Hz (VLP band) and of 5–10 Hz (HF band) did not show a time lag, suggesting a strong connection between the two sources, in which one immediately responds to the other. The location of the VLP pulse source was considered as the centroid of the inflation source, which was near the roof of the cavity where volcanic gas accumulated, whereas the HF oscillation seemed to be emitted more intensely from the lower and narrower portions of the cavity. This resulted in the source depth differences between the two signal bands.

5.2. LP volcanic tremors and earthquakes. Figure 9 shows three unique LP volcanic tremors, which occurred on 23 June 2004. These tremors showed tips or sawtooth waveforms with periods longer than 5 s and durations longer than 100 s. Before the 2004 eruption, unique tremors occurred in June 2004 intensively. The summit broadband stations were destroyed by the first eruption in 2004 but were restored in October 2007. After the re-start of the summit broadband seismic observations,
these types of tremors were rarely detected until June 2018.

Figure 10 shows a closeup of an LP earthquake characterized by a gradually attenuating motion with an upwards convex followed by two large high-frequency events. The signal was filtered with a low-pass IIR filter in the frequency range below 1 Hz to remove the high-frequency component (bottom panel in Fig. 10) as it clearly showed the features of the LP motion. Hereafter, this is referred to as an LP earthquake. Sometimes, an LP earthquake occurs without large high-frequency events (Fig. 11), which is referred to as a quasi-LP earthquake. LP and quasi-LP earthquakes occurred intensively between late-May 2004 and mid-June 2004, and they were rarely detected until May 2018.

It is plausible that a fluid flowing through channels in a shallow region of volcanic conduit is analogous to a hydraulic circuit. A hydraulic circuit with a control valve is a candidate for the nonlinear system of the LP tremor or the LP earthquake. Hayashi and Ohi\textsuperscript{41} and Hayashi et al.\textsuperscript{42} proposed a
control valve model with an attached valve chamber. Figure 12 schematically depicts the control valve model imaged on a volcanic conduit. The equation of the poppet valve motion is given as an oscillatory system with damping, where the axial fluid force \( F \) is controlled by a pressure change in a valve chamber and a flow change in the pipeline connecting the valve chamber and a fluid reservoir. This system also includes the boundary condition of the poppet valve’s rebound and an empirical discharge coefficient of the valve, \( C_P \). The equation of the poppet valve motion is given as

\[
m \frac{d^2X}{dt^2} + \delta \frac{dX}{dt} + k(X + X_i) = F,
\]

where \( X, m, \delta, k, \) and \( X_i \) are the valve displacement, mass of the poppet, viscous damping coefficient, stiffness of the spring supporting the poppet, and initial spring compression, respectively. \( kX_i \) is the initial spring compression force and transforms into a cracking pressure \( P_{Sr} \) described by \( kX_i = A_P P_{Sr} \), where \( A_P \) is the cross-sectional area of the valve-seat aperture. The axial fluid force \( F \) is formulated as

\[
F = A_P P_C \left[ 1 - 4 C_P \frac{X}{d_p} \sin 2\alpha \right],
\]

where \( P_C, C_P, d_p, \) and \( \alpha \) are the valve chamber pressure, discharge coefficient of the poppet valve, diameter of the valve-seat aperture, and half angle of the poppet, respectively. \( A_P \) can be converted to \((\pi/4) d_p^2\). The pressure change in the valve chamber is given as

\[
\frac{V}{\beta} \frac{dP_C}{dt} = Q_O - Q_C,
\]

where \( V, \beta, Q_O, \) and \( Q_C \) are the volume of the valve chamber, bulk modulus of the fluid, orifice flow rate, and poppet valve flow rate, respectively. Assuming that the length of the supply pipeline \( L \) is short, the dynamics of the pipeline flow can be approximated by the lumped parameter model. Consequently, the flow change in the pipeline is given as

\[
\frac{\rho L}{A_L} \frac{dQ_O}{dt} = P_S - P_O,
\]

where \( A_L = (\pi/4) d_L^2 \) is the cross-sectional area of the pipeline. \( \rho, P_S, \) and \( P_O \) are the density of the fluid, supply pressure, and pressure at the orifice, respectively. The flow rate through the poppet valve is written as

\[
Q_C = C_P \pi d_p \cdot X \sin \alpha \sqrt{\frac{2 \Psi(P_C)}{\rho}},
\]

where

\[
\Psi(P_C) = \begin{cases} P_C & \text{for } P_C > 0, \\ 0 & \text{for } P_C \leq 0. \end{cases}
\]

Here, \( C_P \) is the discharge coefficient of the poppet valve based on the experimental results and is given as

\[
C_P = C_{PO} \left( \frac{\gamma X}{1 + \gamma X} \right),
\]

where \( C_{PO} \) and \( \gamma \) are the maximum discharge coefficient and shape parameter, respectively. The orifice flow rate is controlled by the differential pressure \( P_O - P_C \). Bernoulli’s theorem yields their relation as

\[
P_O = P_C + \frac{\rho}{2 C_O^2 A_0^2} |Q_O| Q_O,
\]

where \( C_O \) and \( A_0 \) are the discharge coefficient and area of the orifice, respectively.

The vibrating poppet sometimes impacts the valve seat. Hence, the numerical simulation applied the following collision condition

\[
\frac{dX(t^+)}{dt} = -e \cdot \frac{dX(t^-)}{dt} \text{ at } X(t^+) = 0,
\]

where \( dX(t^-)/dt \) and \( dX(t^+)/dt \) are the valve velocities just before and after a collision, respectively. \( e \) is the restitution coefficient. Hayashi et al.\(^{43} \)
analyzed the local stability on this model. A self-
excited oscillation could occur even when the sup-
ply pressure $P_S$ was lower than the cracking pressure $P_{Si}$. 

To reformulate these equations into a dimen-
sionless form, the following scales were introduced: 
the characteristic length $L_0$, characteristic mass $M_0$, 
and characteristic time $T_0$, which were set to $L_0 = d_p$, 
$M_0 = m$, and $T_0 = 2\pi(m/k)^{1/2}$. Assuming $\alpha = 45^\circ$, 
the equations can be rearranged into a four-order 
system of ordinary differential equations with model 
variables of $x$, $y$, $p_c$, and $q_o$. The nonlinear differential 
equation system is given as 

$$
\dot{x} = y,
\dot{y} = -4\pi^2 x - a_1 y + \left(\frac{\pi}{4} - a_2 \left(\frac{\gamma_x}{\lambda + \gamma_x}\right)x\right),
\dot{p}_c = \frac{\pi}{4} p_{si},
\dot{q}_o = -c_1 p_c - c_2 |q_o| q_o + c_1 p_s, 
y(t^+) = -e \cdot y(t^-) \text{ at } x(t^+) = 0,
$$

where 

$$
x = \frac{X}{d_p}, \quad p_c = 4\pi^2 \frac{d_p}{k} P_C, \quad p_{si} = 4\pi^2 \frac{d_p}{k} P_{Si},
$$

$$
p_s = 4\pi^2 \frac{d_p}{k} P_S, \quad q_o = 2\pi \sqrt{\frac{m}{k}} \frac{Q_0}{d_p^2}.
$$

The coefficients of the equations are given as 

$$
a_1 = 2\pi \frac{\delta}{\sqrt{km}}, \quad a_2 = \pi C_{PO},
b_1 = 4\pi^2 \frac{\beta}{V} \frac{d_p^4}{k}, \quad b_2 = 4\pi^3 C_{PO} \frac{\beta}{V} \frac{d_p^{5/2}}{k} \sqrt{\frac{m}{\rho}},
c_1 = \frac{\pi}{4} \left(\frac{d_h}{d_p^2}\right)^2 \frac{m}{\rho L}, \quad c_2 = \frac{2d_p^3}{\pi C_{O2}^2 L} \left(\frac{d_h}{d_o^2}\right)^2,
\lambda = \frac{1}{d_p}.
$$

The LP tremor which occurred at 04:23 (JST) 
on 23 June 2004 exhibited a change in the dominant 
period from 5 to 20 s. The characteristic time $T_0$ of 
the dimensionless equations is $2\pi(m/k)^{1/2}$. Therefore, 
a change of the poppet mass $m$ or the stiffness of 
spring $k$ can change the dominant period of the valve 
motion. If the temporal changes of $m$ and $k$ are given as 

$$
m(t) = m_0 f_m(t), 
k(t) = k_0 f_k(t),
$$
and $T_0$ is changed from $2\pi(m/k)^{1/2}$ to $2\pi(m_0/k_0)^{1/2}$, 
the second equation of the nonlinear differential 
equation system can be replaced with 

$$
\dot{f}_m(t) = 4\pi^2 f_k(t) x - a_1 y + \left(\frac{\pi}{4} - a_2 \left(\frac{\gamma_x}{\lambda + \gamma_x}\right)x\right), \quad p_c = \frac{\pi}{4} p_{si}.
$$

This control valve lumps the model parameters 
to deal with a system of ordinary differential 
equations. This simplification affected the results 
quantitatively. Hence, the estimated material 
parameters had some ambiguity in the adoption of the 
control valve system to a volcanic conduit system. 
Therefore, this study focuses on the qualitative 
features of LP tremors and LP earthquakes. The maximum temperature at the bottom of the summit 
ater exceeded 200°C in autumn 2002, and exceeded 
500°C about two weeks before the first eruption in 
2004. Thus, we assumed that the fluid involved in the 
signal excitation was a saturated water vapor with a temperature of 250°C. 

Table 1a lists the conceivable values of the 
physical parameters in a volcanic conduit. Although 
this is a rough estimation and includes a large 
ambiguity, these values are reasonable as a trial 
calculation to examine the qualitative characteristics 
of the signal. Table 1b lists the corresponding 
dimensionless coefficients of the system. Using constant 
dimensionless coefficients of $a_1 = 0.95$, $b_1 = 1 \times 10^3$, $c_1 = 10$, $c_2 = 3$, and $p_{si} = 67$ and changing 
the maximum discharge coefficient $C_{PO} (b_2 = \pi C_{PO})$, the 
dimensionless supply pressure $p_s$, the discharge 
shape parameter $\gamma$, and the restitution coefficient $e$, 
the oscillation capturing the general features of 
LP tremors and LP earthquakes was simulated 
(Fig. 13). 

Figure 13a shows the dimensionless valve motion 
with constant values of $m$ and $k$. These 
conditions displayed a sustained valve oscillation 
with a constant period. Figure 13b shows the 
dimensionless valve motion under a condition in 
which the poppet mass increased with time ($f_m(t) = 1 + t^2/4$). The dominant period of the valve motion 
gradually increased with time, reproducing the 
qualitative characteristics of the latter half of an 
LP tremor. Figure 13c shows another example of the dimensionless valve motion with constant values 
of $m$ and $k$, which resembles an LP earthquake in 
qualitative characteristics.
Despite the discrepancy between the LP tremor’s and the LP earthquake’s waveforms, the same dynamics could excite these oscillations by changing the conduit condition. In the case of an LP tremor, a large value of $C_{PO}$ indicated that the discharge coefficient $C_P$ reached the maximum value $C_{PO}$ even if the valve displacement $x$ was small. The maximum discharge coefficient $C_{PO}$ of the LP tremor was about five times larger than that of the LP earthquake. This parameter setting indicated that the axial fluid force $F$ acting on the valve decreased substantially despite the small valve movement. That is, the valve movement was suppressed. The latter half of the LP tremor could be actualized by increasing the pseudo valve mass. In the case of the LP earthquake, both the maximum discharge coefficient $C_{PO}$ and the discharge shape parameter $\gamma$ were smaller than those of the LP tremor. This parameter setting indicated that a large valve movement was necessary for the discharge coefficient $C_P$ to reach the maximum discharge coefficient $C_{PO}$. However, the maximum value was suppressed, indicating that the fluid flow through the valve was blocked despite the large valve movement. These features qualitatively suggested that a blockage in the conduit progressed during the period when LP tremors and LP earthquakes occurred.

Prior to 2003, the near-field region of the summit crater did not have a seismic station and almost all installed seismometers were short-period sensors. Additionally, the recording systems of seismograms before the 1960s were smoked-paper recorders. The last magmatic moderate-sized eruption in the 1950s occurred on 10 November 1958. The volcano released an intensive outburst, which was regarded as one of the strongest explosive eruptions in recent decades. This eruption was preceded by a high activity of volcanic earthquakes beginning in August 1958, and a repetition of minor ash eruptions began in October 1958. This eruption was followed by other outbursts.
and minor eruptions through the remainder of the year.

Unique short-period seismograms were observed in the seismic activity preceding the 1958 eruption and suspected the activity to be similar to LP tremors and LP earthquakes. Figure 14a shows a large short-period seismogram recorded at San-no-Torii station (SAN) on 10 September 1958. This short-period event exhibited a small preceding signal with a duration longer than 10 s. Figure 14b shows closeup seismograms of the LP earthquake recorded by the broadband seismometer at the summit station (KAW) and by the short-period seismometer at SAN on 12 June 2004. The feature of the lasting small preceding signal resembled the initial portion of the LP earthquake recorded by the short-period seismometer at SAN.

Figure 15a exhibits a short-period tremor composed of intermitted wave packets recorded on 5 October 1958 at the Higashi-Maekake station, which was about 1.5 km east of the summit crater. Figure 15b exhibits the LP tremor at 20:30 (JST) on 23 June 2004 recorded at KAW. The upper trace is the raw signal, while the lower trace is filtered by a high-pass filter with a cutoff frequency of 1 Hz. These unique smoked-paper seismograms suggested that LP tremors and LP earthquakes also occurred during the seismic activity preceding the 1958 eruption.

6. Temporal evolution of seismic activity, LP seismic signal preceding 2004 eruption

Since October 2003, the broadband seismic stations on the west and east rims of the summit crater have been operated by ERI. These stations recorded unique LP seismic signals. In addition, a remarkable earthquake characterized by gradually attenuating long-tail oscillations containing a harmonic spectrum structure was observed. These events are called N-type earthquakes, named after the non-dumping oscillation. Forty-one N-type earthquakes occurred between 20 November 2011 and 1 December 2011. Figure 16 shows their hypocenters (star marks) determined using the double-difference algorithm\[14\] and those of other volcanic earthquakes (open circles). N-type earthquakes occurred in a region limited to several hundred square meters with
a centroid depth around 800 m beneath the crater bottom.

The activities of volcanic earthquakes were compared with the LP seismic signals from October 2003 to the first eruption in 2004. VLP pulse events were detected automatically using a VLP pulse detection algorithm.44) The displacement amplitudes of the detected VLP pulses were calculated by integrating the vertical component of the velocity seismograms, and the activity was evaluated by a daily cumulative amplitude.

Figure 17 shows the daily counts of volcanic earthquakes (orange line) and the daily cumulative amplitude (dark red line). The daily counts of LP tremors, LP earthquakes, and N-type earthquakes are represented by bar graphs with different colors. The scale is indicated by integers in round brackets beside the left vertical axis. Before June 2004, the VLP pulse activity was synchronized with the seismicity. However, it gradually decreased toward the eruption despite the increase in seismicity. VLP pulses were not detected from 24 August 2004 to the first Vulcanian eruption on 1 September 2004. By contrast, an intense seismic swarm started beneath the summit crater on 30 August 2004, and lasted until the first eruption. A reliable interpretation is that the occurrences of LP tremors and LP earthquakes indicated the blockage in the conduit, resulting in the decline of the VLP pulse activity due to shielding gas emission and the increase of seismicity due to stress accumulation in and around the conduit.

N-type earthquakes mainly occurred in July–August 2004, which was before the first eruption for that year. One of the most reliable models of N-type earthquakes is the fluid-filled crack model proposed by Chouet.45),46) He showed that a fluid-filled crack generated a very slow wave propagating along the crack wall, which was termed the “crack wave”. The crack wave leads to more realistic estimates of the resonator size relative to a spherical resonator. Chouet’s model consists of a single isolated fluid-filled crack in an infinite elastic solid body, hence external mass transfer into or out of the crack was not taken into consideration.

The synthetic seismograms calculated by the fluid-filled model exhibited strong similarities to the observed N-type earthquakes in terms of the spectral peaks and long-lasting oscillations.47)–53) If the resonance of fluid-filled cracks caused N-type earthquakes, then the cracks should be isolated. This indicated that the fluid in a certain region of the conduit is confined in the cracks and does not connect with fluid in other regions. The highly frequent occurrence of N-type earthquakes in July and August 2004 suggested that the blockage in the conduit progressed during this period. Thus, the temporal evolution of all seismic data, including the seismicity, the activities of the VLP pulse, LP tremors, LP earthquakes, and N-type earthquakes, indicated that the blockage in the conduit was accelerated about 3 months before the first eruption on 1 September 2004.

7. Temporal evolution of seismic activities, ground deformation, and total magnetic force from October 2007 to January 2018

The summit broadband seismic stations were destroyed by the first Vulcanian eruption on 1 September 2004. They were repaired in October 2007. Figure 18 compiles the daily counts of volcanic earthquakes (orange line), the daily cumulative amplitudes of the VLP pulses (dark red line), and
the relative changes in the GNSS baseline length for sites TASH and KVCO (small pink dots). The small orange circles with red circuits represent the SO$_2$ emissions from the summit crater reported by the Japan Meteorological Agency (JMA) (JMA, personal communication). The small green dots indicate the estimated volcanic signals of the geomagnetic total intensity at KMS. The derivation of the estimation is detailed below. The bar graphs with different colors represent the daily counts of the LP tremors, LP earthquakes, and N-type earthquakes, where the scale is indicated by the blue integers in round brackets beside the left vertical axis.

The Asama volcano and long-period seismic signals

Fig. 17. Temporal evolution of the seismic activity, long-period (LP) seismic signals between 16 October 2003 and 1 September 2004. Orange and dark red graphs indicate daily counts of volcanic earthquakes and daily cumulative amplitude of the VLP pulses, respectively. Integers beside the left vertical axis indicate the scale of the daily counts of volcanic earthquakes. Numbers beside the right vertical axis show the scale of the cumulative amplitudes. Daily counts of the LP tremors, LP earthquakes, and N-type earthquakes are represented by bar graphs with different colors, where the scale is indicated by the blue integers in round brackets beside the left vertical axis.

the relative changes in the GNSS baseline length for sites TASH and KVCO (small pink dots). The small orange circles with red circuits represent the SO$_2$ emissions from the summit crater reported by the Japan Meteorological Agency (JMA) (JMA, personal communication). The small green dots indicate the estimated volcanic signals of the geomagnetic total intensity at KMS. (The derivation of the estimation is detailed below.) The bar graphs with different colors represent the daily counts of the LP tremors, LP earthquakes, and N-type earthquakes, where the scale is indicated by integers in round brackets beside the left vertical axis. From the temporal evolution, the volcanic activity consisted of four periods: October 2007 – mid-2010 (first term), mid-2010 – mid-2013 (second term), mid-2013 – June 2015 (third term), and June 2015 – January 2018 (last term).

A magma budget of the magma plumbing system of the Asama volcano was estimated based on the volcanic gas emission rates and the ground deformation data. During both the extension and contraction stages, the volume of degassed magma was one order of magnitude larger than the volume change of the dyke. This suggested that a significant amount of magma was supplied from the magma chamber to the dyke. Then degassed magma descended back to the magma chamber. During the contraction stages, the shrinkage volumes of the dyke were comparable to or slightly higher than the volume decreases of the magma by the volcanic gas discharge. This fact implied that some amount of volcanic gas is perennially supplied to the shallow region of the conduit, even in contraction stages.

The GNSS baseline switched from a contraction stage to an extension stage in the early phase of the first term. The average levels of the SO$_2$ emission and the VLP pulse activity were high during the extension stage, even though the existence fluctuated. An SO$_2$ emission greater than 2,000 tons/day lasted from November 2008 to February 2009 despite
the low VLP pulse activity. However, this relation flipped from April to August 2009. The VLP pulse activity represented the rapid gas emission through the northern part of the conduit as shown by the focal mechanism and its location. The SO2 emission was an indicator of the total amount of gas emissions from the conduit. The variability of correlation between the SO2 emission and the VLP pulse activity suggested that multi outgassing pathways existed in the shallow region of the conduit and a heterogeneous distribution of gas phases occurred within the conduit.44)

The SO2 emission and the VLP pulse activity were maintained at a low level when the GNSS baseline was in the contracting stage through the second and third terms. However, the contraction of the GNSS baseline indicated that the extra magma intrusion beneath the western flank of the Asama volcano was dormant, resulting in a basic amount of volcanic gas supply to the summit crater. The second and third terms were distinguished by the variation patterns of the geomagnetic total intensity and occurrence frequencies of N-type earthquakes. In the last term, which began with a minor eruption in June 2015, the GNSS baseline exhibited repeated extension and contraction stages with short durations. The higher levels of the SO2 emission and the VLP pulse activity than those in the second and third terms were sustained during the last term.

The minor eruptions in August 2008 and June 2015 were preceded by short-term rapid increases in VLP pulse activity and SO2 emissions, which lasted several days. N-type earthquakes occurred frequently for several months before the eruptions. The LP
earthquakes and LP tremors also occurred one and a half months before the eruptions in August 2008. Large VLP pulses preceded all three minor eruptions in August 2008 by about 2 min. The minor eruption in June 2015 was also preceded by a large VLP pulse, which occurred about 3 min before the eruption. Even if N-type earthquakes which frequently occurred for several months before the eruptions suggested shielding in the conduit, the rapid increment of the SO₂ emission and VLP pulse activity implied that multi outgassing pathways were not blocked completely, and certain portions of the pathways were still alive. The intensive gas emissions eventually caused these four minor eruptions.

The small eruption on 2 February 2009 was preceded by a short-term decrement of the VLP pulse activity for 9 days, and 26 N-type earthquakes occurred in a focused manner 1 day before the eruption. This intense activity of N-type earthquakes and the falling-off of the VLP pulse activity suggested that the shielding in the overall area of the conduit rapidly progressed, resulting in the largest eruption during the 2008–2009 crisis.

In a global sense, the variations of the geomagnetic field are caused by the currents in the ionosphere and magnetosphere along with the dynamo mechanism in the Earth’s outer core. The variations of the geomagnetic field due to volcanic activity are much smaller. Hence, ionospheric, magnetospheric, and core effects must be removed to investigate the thermal conditions in a volcano. A common method used to eliminate non-volcanic variations from the observed total magnetic intensity at a volcano site is to simply subtract the intensity measured at a reference site close to the observation site, assuming that large-scale variations are common to both sites due to their long wavelengths.\(^{55}\)

Total geomagnetic intensity observations started in October 2009. Two total geomagnetic intensity sensors were installed at KMS and KUR by ERI. KMS lies on the southeast side of the summit crater, while KUR is about 3 km northeast. Therefore, it is a reasonable hypothesis that KUR and KMS have common large-scale variations of the geomagnetic field. A simple difference method was employed to estimate the volcanic component of continuous observations due to the volcanic activity in the summit area (Fig. 18, green dots).

The geomagnetic total intensity due to the volcanic activity in the summit area exhibited an increasing secular trend over the entire period with a long dormant period from mid-2013 to mid-2015. Assuming the volcano-magnetic change represented the thermomagnetic effect, the increasing secular trend at KMS suggested an increase in the magnetization of the rocks beneath the summit crater Kamayama in response to cooling. The increment of the geomagnetic total intensity rested for 2 years from mid-2013, but began to increase just after the minor eruption on 16 June 2015. During this dormant period, the geomagnetic total intensity decreased by 10 nT until June 2015 compared with the linear secular trend from January 2010 to August 2013.

The GNSS baseline constructing stage was characterized by the small amount of SO₂ emissions and the restful VLP pulse activity. The geomagnetic total intensity increased linearly along the secular trend during the first half of the constructing stage, while the increment stopped completely in the last half of the constructing stage. This quiet interval of the total intensity variation was well synchronized with a period of the frequent occurrence of N-type earthquakes. The GNSS baseline turned to an extension from May 2015 and was followed by the rapid increment of SO₂ emissions and the activation of VLP. This resulted in an eruption on 16 June 2015. Afterward, the geomagnetic total intensity started to increase linearly along the secular trend. The gradient was similar to that before the quiet interval, and the activity of N-type earthquakes dropped dramatically.

The synchronization between the variation of the geomagnetic total intensity and the activity of N-type earthquakes supports the following hypothesis. A basic amount of volcanic gas was supplied to the conduit even in the contraction stage. Shielding in certain portions of the outgassing pathways began in mid-2013. This trapped high-temperature volcanic gas in the shielding area within the conduit. The trapped gas gradually inhibited cooling of the rocks beneath the summit crater, and N-type earthquakes occurred frequently in the shielding area. The north–south extension of the GNSS baseline at the western side of the volcano from May 2015 indicated a new dyke intrusion to the west of the summit, supplying more volcanic gas to the shallow region of the conduit. The intensive gas emission led to the minor eruption on 16 June 2015. After the eruption, the shielding portions were destroyed, and the volcanic gas passed through the conduit freely. Then the cooling process beneath the summit crater recovered, causing an increase in the geomagnetic total intensity along the secular trend. The dissolution of the shield also depressed the activity of N-type earthquakes.
Thermomagnetic modeling was conducted to verify this hypothesis. The concentric occurrences of N-type earthquakes in the last two months of 2011 provided a precise hypocenter distribution of N-type events (Fig. 16). We assumed that N-type events in the dormant period of the total intensity also occurred in the same area of the conduit and a pause in magnetization arose in this area. The change of the total intensity in the summit area was calculated by uniform demagnetization in a prism source with a width of 200 m and height of 500 m. The prism source was beneath the summit crater with a top elevation of 2,000 m above sea level (Fig. 16, pink shaded red rectangles). The magnetization intensity of the rocks in the source was assumed to be 2.2 A/m. A magnetic anomaly change on the surface induced by demagnetization in the prism source was exhibited by contours with intervals of 1 nT (Fig. 19). The magnetic anomaly of −10 nT at KMS was consistent with the observed change of the geomagnetic total intensity over the entire dormant period. Although an important assumption is that the thermomagnetic effect yields the volcano magnetic change, the geothermal activity is the most intensive phenomenon in an active conduit of a volcano, including the Asama volcano. Therefore, our modeling is plausible to explain the observed anomaly, and the hypothesis regarding the internal process within the conduit seems to be a feasible interpretation.

From the 2004 eruptions to the beginning of 2018, all eruptions occurred in the extension stages of the GNSS baseline change, indicating that the magma intrusion beneath the western flank of the Asama volcano is an essential process for the volcanic activity in the last 20 years. The supplied amount of magma from the magma chamber was located about 8 km west of the summit and spanned to the dykes beneath the western flank of the volcano. The internal conditions in the shallow regions of the conduit seemed to control the eruptive style.

8. Concluding remarks

The seismic observations of the Asama volcano, which began in 1910, are some of the earliest geophysical observations of volcanoes in the world. Minakami archived the classical analysis on the classification of volcanic earthquakes based on the observed waveforms. A modern observation system, which is composed of the seismic and geodetic networks, was developed in and around the Asama volcano in 2003. The seismic velocity structure of the upper crust beneath the Asama volcano was elucidated using data provided by this network and a regional seismic network enclosing the Asama volcano via seismic
ambient noise tomography.\textsuperscript{22} The active source seismic tomography revealed the seismic velocity structure down to several kilometers below sea level.\textsuperscript{20,21,23}

The magma pathway in the upper crust beneath the Asama volcano was synthesized by combining the velocity structures with the hypocenter distributions of volcanic earthquakes\textsuperscript{12} and the ground deformations before and after the eruptions.\textsuperscript{18,19,23} This magma pathway was functional during the recent eruptive activity of the Asama volcano. However, other smaller magma chambers with different pathways could exist because most andesites of the Asama volcano are products of magma mixing.\textsuperscript{2,3,57}

The proximal broadband seismic observations provided several kinds of unique LP seismic signals such as the VLP pulse, LP tremor, and LP earthquake. Using inversion analysis of seismic waveforms and observations of volcanic gas emissions revealed that the VLP pulse was caused by the volcanic gas emission process. The qualitative features of LP tremors and LP earthquakes suggested that the blockage in the conduit advanced in periods where LP tremors and LP earthquakes occurred. The temporal evolutions of multidiscipline data regarding the volcanic activity were investigated using the knowledge on the magma pathway and the unique LP seismic signals. The supplied amount of magma from the magma chamber and the internal condition in the shallow regions of the conduit controlled recent eruptions. Although the Asama volcano has not recently experienced a large-scale eruption like the Tenmei eruption in 1783\textsuperscript{3,58–60} or the Tennin eruption in 1108,\textsuperscript{3,60,61} a different scenario may appear before the next large-scale eruption.

Acknowledgements

We thank the Japan Meteorological Agency for access to their SO\textsubscript{2} emission data. This research was supported by Grants-in-Aid for Scientific Research (19540441, 21740319, 22540431, 22740289, 16K13872, and 22K03774) from the Japan Society for the Promotion of Science.

References

1) Aramaki, S. (1963) Geology of Asama Volcano. J. Fac. Sci. Univ. Tokyo Sect. 2 14, 229–443.

2) Takahashi, M., Yasui, M., Ichikawa, Y., Kamioika, Y., Asaka, N., Sakagami, M. \textit{et al.} (2007) Whole-rock major element chemistry for eruptive products of Asama-Mae-kake Volcano, central Japan: Summary for data of 288 samples. Proc. Inst. Nat. Sci. Nihon Univ. \textbf{42}, 55–70 (in Japanese with English abstract).

3) Yasui, M., Takahashi, M., Tsutsumi, T., Aramaki, S., Takeo, M. and Aoki, Y. (2013) Active volcano in Central Japan: Asama Volcano. Bull. Volcanol. Soc. Jpn. \textbf{58}, B2-1–B2-32.

4) Newhall, C.G. and Self, S. (1982) The Volcanic Explosivity Index (VEI): an estimate of explosive magnitude for historical volcanism. J. Geophys. Res. \textbf{87}, 1231–1238.

5) Omori, F. (1910) The eruptions of Asama-yama. Rep. Imp. Earthq. Invest. Comm. \textbf{67}, 1–26 (in Japanese).

6) Omori, F. (1912) The eruptions and earthquakes of Asama-yama. Bull. Imp. Earthq. Invest. Comm. \textbf{6}, 1–147.

7) Omori, F. (1914) The eruptions and earthquakes of Asama-yama II: List of the volcanic earthquakes instrumentally registered at the Asama-yama Observatory in 1911 and 1912. Bull. Imp. Earthq. Invest. Comm. \textbf{6}, 149–226.

8) Omori, F. (1914) The eruptions and earthquakes of Asama-yama III: Remarks on the seismographical observations at Yuno-taira in 1911 and 1912. Bull. Imp. Earthq. Invest. Comm. \textbf{6}, 227–257.

9) Omori, F. (1914) The eruptions and earthquakes of Asama-yama IV: Strong Asama-yama outbursts, Dec. 1912 to May 1914. Bull. Imp. Earthq. Invest. Comm. \textbf{7}, 1–215.

10) Minakami, T. (1960) Fundamental research for predicting volcanic eruptions (Part 1). Bull. Earthq. Res. Inst. Univ. Tokyo \textbf{38}, 497–544.

11) Nakada, S., Yoshimoto, M., Koyama, E., Tsujii, H. and Urabe, T. (2005) Comparative study of the 2004 eruption with old eruptions at Asama Volcano and the activity evaluation. Bull. Volcanol. Soc. Jpn. \textbf{50}, 303–313 (in Japanese with English abstract).

12) Takeo, M., Aoki, Y., Ohminato, T. and Yamamoto, M. (2006) Magma supply path beneath Mt. Asama volcano, Japan. Geophys. Res. Lett. \textbf{33}, L15310.

13) Yamamoto, M., Takeo, M., Ohminato, T., Okawa, J., Aoki, Y., Ueda, H. \textit{et al.} (2005) A unique earthquake activity preceding the eruption at Asama volcano in 2004. Bull. Volcanol. Soc. Jpn. \textbf{50}, 393–400 (in Japanese with English abstract).

14) Waldhauser, F. and Ellsworth, W.L. (2000) A double-difference earthquake location algorithm: methods and application to the northern Hayward fault, California. Bull. Seismol. Soc. Am. \textbf{90}, 1353–1368.

15) Churei, M. and Katayama, H. (2006) Ground tilt change associated with the 2004 eruption at Asama Volcano, Japan. Bull. Volcanol. Soc. Jpn. \textbf{51}, 91–101 (in Japanese with English abstract).

16) Maeno, F., Suzuki, Y., Nakada, S., Koyama, E., Kaneko, T., Fuji, T. \textit{et al.} (2010) Course and ejecta of the eruption of Asama Volcano on 2 February 2009. Bull. Volcanol. Soc. Jpn. \textbf{55}, 147–154 (in Japanese with English abstract).

17) Volcanic Observations and Information Center, Volcanology Division, Japan Meteorological
Agency (2015) Volcanic activity of Asamayama Volcano (June 2015–October 2015). Rep. Coord. Comm. Predict. Volcanic Erupt. **122**, 108–127 (in Japanese).

Aoki, Y., Watanabe, H., Koyama, E., Oikawa, J. and Morita, Y. (2005) Ground deformation associated with the 2004–2005 unrest of Asama Volcano, Japan. Bull. Volcanol. Soc. Jpn. **50**, 575–584 (in Japanese with English abstract).

Murakami, M. (2005) Magma plumbing system of the Asama Volcano inferred from continuous measurements of GPS. Bull. Volcanol. Soc. Jpn. **50**, 347–361 (in Japanese with English abstract).

Aoki, Y., Takeo, M., Aoyama, H., Fujimatsu, J., Tsutsui, T. and Morita, Y. (2005) Ground deformation associated with the 2004–2005 unrest of Asama Volcano, Japan. Bull. Volcanol. Soc. Jpn. **50**, 575–584 (in Japanese with English abstract).

Aoki, Y., Takeo, M., Aoyama, H., Fujimatsu, J., Matsumoto, S., Miyamachi, H. et al. (2009) P-wave velocity structure beneath Asama Volcano, Japan, inferred from active source seismic experiment. J. Volcanol. Geotherm. Res. **187**, 272–277.

Nagaoka, Y., Nishida, K., Aoki, Y., Takeo, M. and Ohminato, T. (2012) Seismic imaging of magma chamber beneath an active volcano. Earth Planet. Sci. Lett. **333–334**, 1–8.

Aoki, Y., Takeo, M., Ohminato, T., Nagaoka, Y. and Nishida, K. (2013) Magma pathway and its structural controls of Asama Volcano, Japan. Geol. Soc. Lond. Spec. Publ. **380**, 67–84.

Rubin, A.M. and Gillard, D. (1998) Dike-induced earthquakes: theoretical considerations. J. Geophys. Res. **103**, 10003–10015.

Yamamoto, M. and Sato, H. (2010) Multiple scattering and mode conversion revealed by an active seismic experiment at Asama volcano, Japan. J. Geophys. Res. **115**, B07304.

Prudencio, J., Aoki, Y., Takeo, M., Ibanez, J.M., Pezzo, E.D. and Song, W. (2017) Separation of scattering and intrinsic attenuation at Asama volcano (Japan): Evidence of high volcanic structural contrasts. J. Volcanol. Geotherm. Res. **333–334**, 96–103.

Aizawa, K., Ogawa, Y., Hashimoto, T., Koyama, T., Kanda, W., Yamaya, Y. et al. (2008) Shallow resistivity structure of Asama volcano and its implications for magma ascent process in the 2004 eruption. J. Volcanol. Geotherm. Res. **173**, 165–177.

Nishida, K., Kawakatsu, H. and Obara, K. (2008) Three-dimensional crustal S wave velocity structure in Japan using microseismic data recorded by Hi-net tiltmeters. J. Geophys. Res. **113**, B10302.

Rawlinson, N. and Sambrodge, M. (2003) Irregular interface parametrization in 3-D wide-angle seismic travel-time tomography. Geophys. J. Int. **155**, 79–92.

Maeda, Y., Takeo, M. and Ohminato, T. (2011) A waveform inversion including tilt: method and simple tests. Geophys. J. Int. **184**, 907–918.

Maeda, Y. and Takeo, M. (2011) Very-long-period pulsed at Asama volcano, central Japan, inferred from dense seismic observations. Geophys. J. Int. **185**, 265–282.

Rodgers, P.W. (1968) The response of the horizontal pendulum seismometer to Rayleigh and Love waves, tilt, and free oscillations of the Earth. Bull. Seismol. Soc. Am. **58**, 1384–1406.

Aster, R., Mah, S., Kyle, P., McIntosh, W., Dunbar, N., Johnson, J. et al. (2003) Very long period oscillations of Mount Erebus Volcano. J. Geophys. Res. **108**, 2522–2543.

Pillet, R. and Virieux, J. (2007) The effect of seismic rotations on inertial sensors. Geophys. J. Int. **171**, 1314–1323.

Graizer, V. (2006) Tilts in strong ground motion. Bull. Seismol. Soc. Am. **96**, 2090–2102.

Aoyama, H. (2008) Simplified test on tilt response of CMG-40T seismometers. Bull. Volcanol. Soc. Jpn. **53**, 35–46 (in Japanese with English abstract).

Aoyama, H. and Oshima, H. (2008) Tilt change recorded by broad-band seismometer prior to small phreatic explosion of Meakan-dake volcano, Hokkaido, Japan. Geophys. Res. Lett. **35**, L06307.

Kazahaya, R., Mori, T., Takeo, M., Ohminato, T., Urabe, T. and Maeda, Y. (2011) Relation between single very-long-period pulses and volcanic gas emissions at Mt. Asama, Japan. Geophys. Res. Lett. **38**, L13307.

Maeda, Y., Takeo, M. and Kazahaya, R. (2019) Comparison of high- and low-frequency signal sources for very-long-period seismic events at Asama volcano, Japan. Geophys. J. Int. **217**, 389–404.

Battaglia, J. and Aki, K. (2003) Location of seismic events and eruptive fissures on the Piton de la Fournaise volcano using seismic amplitudes. J. Geophys. Res. **108**, 2364.

Hayashi, S. and Ohi, K. (1993) Global stability of a poppet valve circuit. J. Fluids Control **11**, 693–716.

Hayashi, S., Hayase, T. and Kurahashi, T. (1997) Chaos in a hydraulic control valve. J. Fluids Structures **11**, 693–716.

Takenaka, T. and Urata, E. (1968) Section 3: momentum theory in hydraulic valves. In Yurikigaku (Dynamics of Hydraulics). Yoken-do, Tokyo, pp. 48–50 (in Japanese).

Kazahaya, R., Maeda, Y., Mori, T., Shinohara, H. and Takeo, M. (2015) Changes to the volcanic outgassing mechanism and very-long-period seismicity from 2007 to 2011 at Mt. Asama, Japan. Earth Planet. Sci. Lett. **418**, 1–10.

Chouet, B. (1986) Dynamics of a fluid-driven crack in three dimensions by the finite difference method. J. Geophys. Res. **91**, 13967–13992.

Chouet, B. (2003) Volcano seismology. Pure Appl. Geophys. **160**, 739–788.

Chouet, B., Page, R.A., Stephens, C.D., Lahr, J.C. and Power, J.A. (1994) Precursory swarms of long-period events at Redoubt Volcano (1989–1990), Alaska: their origin and use as a forecasting tool. J. Volcanol. Geotherm. Res. **62**, 95–135.

Nakano, M., Kumagai, H., Kumazawa, M., Yamaoka, K. and Chouet, B.A. (1998) The
excitation and characteristic frequency of the long-period volcanic event: an approach based on an inhomogeneous autoregressive model of a linear dynamic system. J. Geophys. Res. 103, 10031–10046.

49) Kumagai, H., Chouet, B.A. and Nakano, M. (2002) Temporal evolution of a hydrothermal system in Kusatsu-Shirane Volcano, Japan, inferred from the complex frequencies of long-period events. J. Geophys. Res. 107, 2236.

50) Kumagai, H., Chouet, B.A. and Nakano, M. (2002) Waveform inversion of oscillatory signatures in long-period events beneath volcanoes. J. Geophys. Res. 107, 2301.

51) Nakano, M. and Kumagai, H. (2005) Response of a hydrothermal system to magmatic heat inferred from temporal variations in the complex frequencies of long-period events at Kusatsu-Shirane Volcano, Japan. J. Volcanol. Geotherm. Res. 147, 233–244.

52) Kumagai, H. (2006) Temporal evolution of a magmatic dike system inferred from the complex frequencies of very long period seismic signals. J. Geophys. Res. 111, B06201.

53) Aoyama, H. and Takeo, M. (2001) Wave properties and focal mechanisms of N-type earthquakes at Asama volcano. J. Geophys. Res. 105, 163–182.

54) Kazahaya, R., Aoki, Y. and Shimohara, H. (2015) Budget of shallow magma plumbing system at Asama Volcano, Japan, revealed by ground deformation and volcanic gas studies. J. Geophys. Res. Solid Earth 120, 2961–2973.

55) Takahashi, K. and Fujii, I. (2014) Long-term thermal activity revealed by magnetic measurements at Kusatsu-Shirane volcano, Japan. J. Volcanol. Geotherm. Res. 285, 180–194.

56) Nakatsuka, T., Utsugi, M., Okuma, S., Tanaka, Y. and Hashimoto, T. (2009) Detection of aeromagnetic anomaly change associated with volcanic activity: An application of the generalized mis-tie control method. Tectonophysics 478, 3–18.

57) Takahashi, M., Yasui, M., Kanamaru, T. and Yamashita, D. (2007) Whole-rock major element chemistry for ash-fall pumice obtained by trenching survey of tephra deposit erupted from the Asama-Maekake Volcano: temporal variation for magmatic chemistry of the Asama-Maekake Volcano since 10 ka. Proc. Inst. Nat. Sci. Nihon Univ. 54, 143–172 (in Japanese with English abstract).

58) Yasui, M., Koyaguchi, T. and Aramaki, S. (1997) Plinian eruptions in the 1783 activity of Asama volcano inferred from the deposits and the old records. Bull. Volcanol. Soc. Jpn. 42, 281–297 (in Japanese with English abstract).

59) Yasui, M. and Koyaguchi, T. (2004) Sequence and eruptive style of the 1783 eruption of Asama volcano, central Japan: a case study of an adesitic explosive eruption generating fountain-fed lava flow, pumice fall, scoria flow and forming a cone. Bull. Volcanol. 66, 243–262.

60) Yasui, M., Takahashi, M., Kanamaru, T. and Nagai, M. (2021) High-resolution reconstruction of the history of large-scale eruptions of Asama-Maekake Volcano based on the stratigraphy of pyroclastic fall deposits. Bull. Volcanol. Soc. Jpn. 66, 293–325 (in Japanese with English abstract).

61) Yasui, M. (2018) Textures of the eruptive products of Asama-Maekake Volcano from the 12th century: indicator of eruptive processes. Proc. Inst. Nat. Sci. Nihon Univ. 53, 37–50.

Profile

Minoru Takeo was born in Aomori Prefecture in 1953 and graduated from Meteorological College, Japan Meteorological Agency, in 1977. He received his M.S and Ph.D. in 1979 and 1982, respectively, from Hokkaido University. He worked as a researcher at Meteorological Research Institute, Japan Meteorological Agency from 1983 to 1987. From 1987, he worked for the Earthquake Research Institute, The University of Tokyo; in 1998, he became a Professor of Geophysics at the same institute. In the course of his career, he was a visiting researcher at the California Institute of Technology (1991–1992), vice director of the Earthquake Research Institute (2013–2015), and head of the Volcano Research Center (2007–2017). His major research fields are analytical and theoretical studies of earthquake and volcanic events based on observational data. He has analyzed rupture processes of tectonic earthquakes based on strong motion seismogram, and pointed out the importance of rotational motion measurement carried by S waves to reveal earthquake source processes. He has also analyzed volcanic earthquakes and tremors to elucidate volcanic activities. Now, he is an Emeritus Professor of The University of Tokyo and a Special Advisor of Japan Meteorological Agency.
Profile

Yosuke Aoki was born in Yokohama in 1973. He received B.S., M.S., and Ph.D. degrees in Geophysics from The University of Tokyo in 1996, 1998, and 2001, respectively. He was a Lamont postdoctoral fellow of the Lamont-Doherty Earth Observatory of Columbia University between 2001 and 2003 before joining the Earthquake Research Institute, The University of Tokyo, in 2003. He is now an Associate Professor there. His main research interest is in space geodesy to measure surface deformation associated with seismic and volcanic activity. For example, he combined geodetic, seismic, and other geophysical observations to propose the magma pathway of the Asama volcano. The work has been published in a series of papers and reviewed in the study by Aoki, Takeo, et al. (2013, Geol. Soc. Lond. Spec. Publ., doi:10.1144/SP380.6). While he has mainly worked on volcanic phenomena, he started to work on seismic deformation as well. For example, he recently delineated the evolution of a slow slip transient in high temporal resolution via kinematic GPS processing (Itoh, Aoki, and Fukuda, 2022, Sci. Rep., doi:10.1038/s41598-022-10957-8). Recently, the technical development of GNSS and SAR data processing attracted his interest; he has been investigating the ionospheric disturbance of GNSS signals induced by explosive volcanic eruptions. Further, he has developed a new method for delineating the along-track component of displacements from InSAR. His work has been published in more than 70 peer-reviewed articles.

Profile

Takao Koyama was born in Tokyo in 1973 and graduated from The University of Tokyo, receiving a Ph.D. in 2002 after completing his dissertation entitled “A study on the electrical conductivity of the mantle by voltage measurements of submarine cables”. He worked as a postdoctoral researcher at the Japan Agency for Marine-Earth Science and Technology (JAMSTEC) from 2002 to 2005 to study the three-dimensional inversion of the semi-global electrical conductivity model and its application to reveal the mantle structure. He moved to the Earthquake Research Institute, The University of Tokyo, in 2005 and engages in volcano research using electromagnetic methods such as magnetotellurics, controlled source EM, and UAV aeromagnetic survey, as well as continuous electromagnetic field measurements.