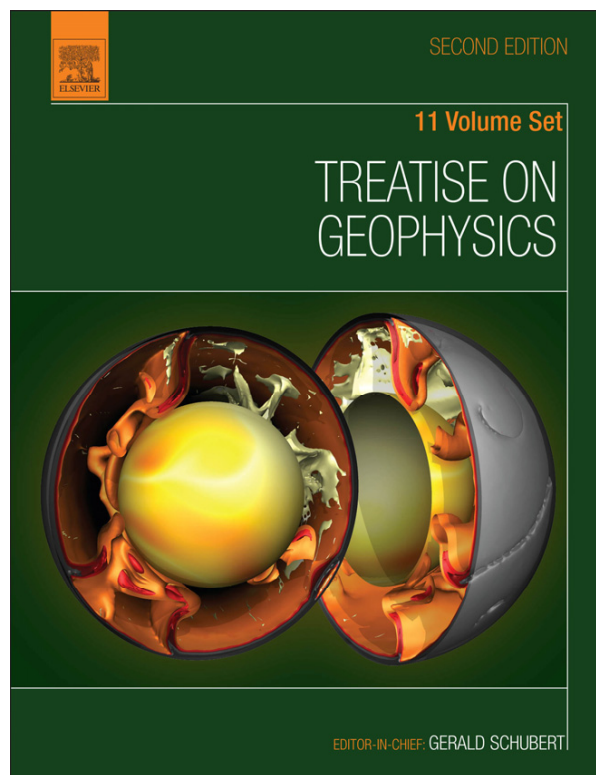


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4.15 Volcano Seismology

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4.15.1 Introduction

Volcano seismology is a field of volcanology in which seismological techniques are employed to help understand the physical conditions and dynamic states of volcanic edifices and volcanic fluid systems to such a level that it eventually contributes to predictions of initiation and cessation of hazardous volcanic activities. Seismology has been a powerful tool for this purpose, especially in the past 20–25 years as new observational and analysis techniques in seismology have become available (e.g., [Chouet, 1996a,b](#); [McNutt, 2005](#)); they include

digital and broadband seismometry and array and moment tensor waveform analyses. While the subjects that should be covered under the title of volcano seismology are quite broad and diverse, this chapter, however, neither intends to nor is capable to cover all of its areas; instead, we try to focus on some of seismological phenomena that appear to be specific to volcanoes. In a sense, volcano seismology here is a field of seismology that deals with problems specific to volcanic areas. Some of the subjects, such as seismic structure of volcanoes (tomography) and eruption monitoring (seismicity), are therefore intentionally omitted. The readers who are interested in

these subjects are advised to refer to the excellent reviews by Chouet (1996a,b, 2003) and McNutt (1996, 2002, 2005).

4.15.2 Volcanic Seismic Signals

For seismologists who have only observed seismograms of ordinary faulting earthquakes, the most intriguing aspect of volcano seismology may be the ample variety of waveforms recorded in volcanic environments. Such volcanic seismic signals range from totally chaotic continuous vibrations lasting for minutes, hours, or sometimes days to purely monotonic vibrations decaying with constant damping rates; they often show several spectral peaks whose frequency may vary during the course of volcanic activities, possibly reflecting the change of physical conditions of volcanic fluid systems; when eruptions occur, a variety of long-period (LP) signals up to as long as a few hundred seconds are generated and sometimes observed at seismic stations around the world. In the following, some of such intriguing volcanic seismic signals will be presented.

Terminology. Based on their appearance on short-period seismometers, volcanic seismic signals have been traditionally classified into four types: high-frequency or A type, low-frequency (LF) or B type, explosion quakes, and volcanic tremors (e.g., Minakami, 1974; for the detail of different terminology, see table 1 of McNutt, 1996). In this chapter, following this tradition, we use high frequency for frequencies higher than 5 Hz and low frequency for frequencies between 5 and 0.5 Hz. On the other hand, the usage of period remains a complicated issue in volcano seismology; the most widely accepted notation to describe volcanic seismic signals is that of Chouet (1996b), in which events of about 1 s are called 'LP'

events and those with longer periods are called 'very-long-period' (VLP) and 'ultra-long-period' events. This definition, however, contradicts with the conventional usage of period in traditional earthquake seismology, where periods longer than the dominant period of the ambient microseismic noise (around 5–10 s) are called 'long-period' (Aki and Lee, 2003; McNutt, 1996, 2005). In this chapter, as we try to cover the subject in a wider context of seismology, we follow the convention of earthquake seismology and use the term LP for periods longer than 5 s and short-period for periods shorter than or around 1 s; for events with a period longer than 50 s, we may also use VLP. To avoid confusion, whenever appropriate, we will try to associate actual number in seconds to describe a period, for example, LP (10 s).

4.15.2.1 Signals Observed by Short-Period Seismometers

The conventional seismometry at active volcanoes, except for a few rare cases, has been conducted using short-period seismometers. Even with such band restricted sensors, wide varieties of waveforms recorded in volcanoes have led scientists to wonder what may be the origins of those signals. Figure 1 shows a classical yet the most fascinating one of those recordings of volcanic tremors observed at Sakurajima Volcano, Japan (Kamo et al., 1977). Volcanic tremors are those sustained ground vibrations commonly observed at volcanoes, typically lasting for longer than 1 min, up to hours or days; they often show periodic spectral features as seen here with the lowest peak frequency around 1 Hz. During the course of activities, relative spectral content and peak frequencies may vary from time to time.

Figure 2 shows examples from Redoubt Volcano, Alaska, where different types of events are displayed: from the top,

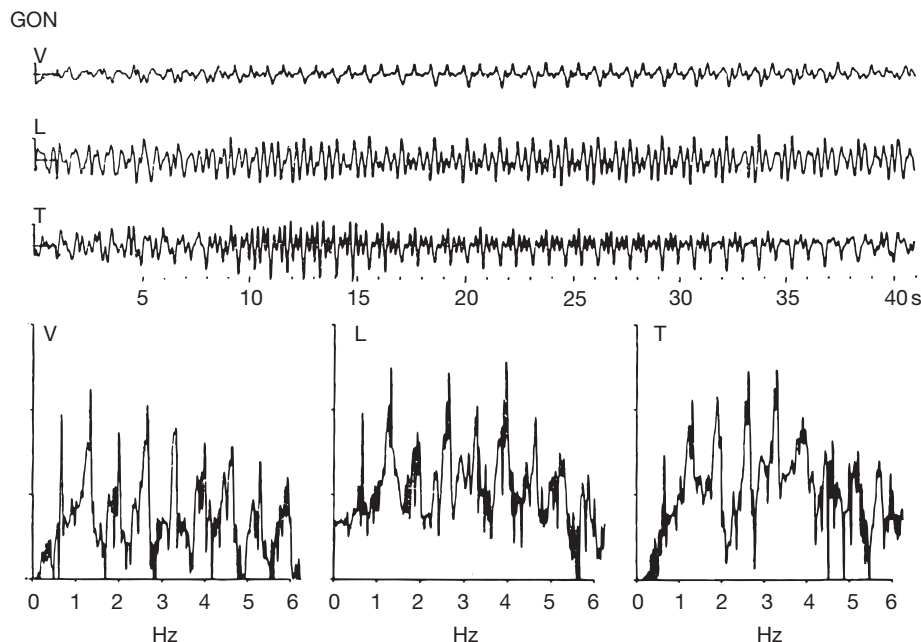


Figure 1 Volcanic tremor observed at Sakurajima. Top three traces show 40 s long seismograms for vertical, longitudinal, and transverse components. Corresponding spectra are shown below (or bottom). Adapted from Kamo K, Furuzawa T, and Akamatsu J (1977) Some natures of volcanic micro-tremors at the Sakurajima volcano. *Bulletin of the Volcanological Society of Japan* 22: 41–58.

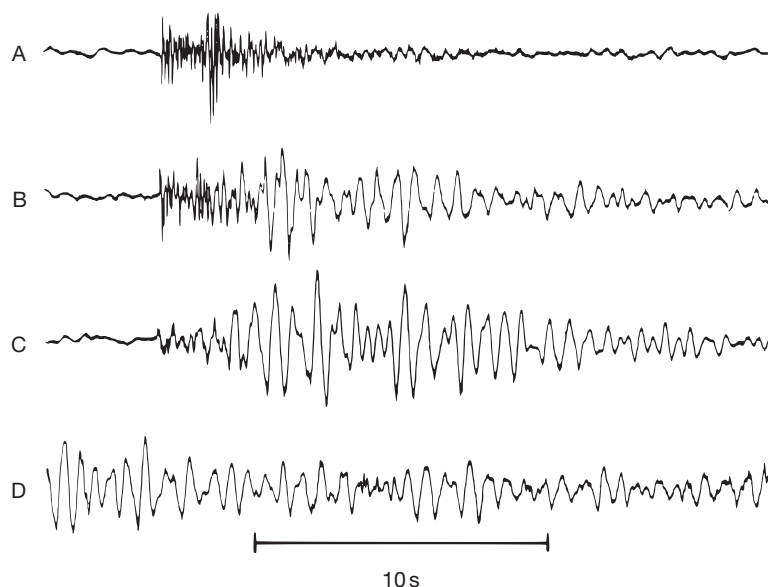


Figure 2 Typical waveforms observed at Redoubt. Adapted from McNutt SR (1996) Seismic monitoring and eruption forecasting of volcanoes: A review of the state-of-the-art and case histories. In: *Monitoring and Mitigation of Volcano Hazards*, pp. 99–146. New York: Springer-Verlag.

(a) high-frequency or volcano-tectonic earthquake, which most likely represents an ordinary brittle failure on a fault; (b) hybrid or mixed-frequency event; (c) LF event with frequency around 1 Hz; and (d) volcanic tremor. Note that the volcanic tremor here appears more random compared to those in Figure 1 and also its frequency content seems similar to that of the LF event (c). It is sometimes considered that tremors and LF events share a common excitation mechanism.

LF events in some volcanoes show characteristic similar spindle-shaped waveforms, suggesting a possible common origin for these events in different volcanoes.

Explosion quakes are observed when explosive eruptions occur and often associated with a high-frequency arrival on seismograms due to a shock wave originated at the vents. Figure 3 shows series of such events recorded in Langila Volcano, Papua New Guinea (Mori et al., 1989), sorted by the ratio of ground-transmitted explosion-wave and air-wave amplitudes. As those without an air-wave are classified as LF events, this figure also indicates that explosion quakes and LF events may share a common origin.

As noted earlier, the spectral content of these LF volcanic seismic signals may vary throughout the course of activities. Figure 4 shows such an example from Montserrat Volcano, West Indies; the temporal change of the system prior to an explosion is clearly manifested.

It appears clear that the LF nature of these seismograms is the most characteristic of volcanic seismic signals. Although they are characterized here just in terms of the frequency content, there exist similarities and diversities among them; their origins may be different from a volcano to another and may change from time to time even in a single volcano. Understanding of their origins and monitoring of their appearances to infer physical conditions and dynamic states of volcanic edifices and volcanic fluid systems are interesting and challenging tasks.

4.15.2.1.1 Deep (volcanic) LF signals

Most of volcanic seismic signals originate within a shallow part of the crust (<5 km). There are, however, certain types of volcanic seismic signals that originate either from the lower crust or from the uppermost mantle (15–50 km) right beneath active volcanoes. As they show similar LF content as volcanic tremors, they are often called deep LF volcanic tremors (or events). They are now reported in many volcanoes around the world (e.g., Aki and Koyanagi, 1981; Hasegawa and Yamamoto, 1994; Nakamichi et al., 2003; Pitt and Hill, 1994; White, 1996). As deep LF volcanic tremors may be related with the deep magma-supplying system of volcanoes, understanding of their origin seems extremely important for the long-term prediction of the activities of volcanoes.

4.15.2.2 Broadband Signals

Availability of broadband seismic records has brought wider perspectives in our view of volcanic activities. Figure 5 shows a very good example observed at Satsuma-Iwojima Volcano, Japan (Ohminato and Ereditato, 1997). The short-period sensor recording at ST3 shows amplitude modulation of high-frequency signal with a periodicity of 46–50 min. When this amplitude becomes small, the low-pass-filtered broadband record at ST1 shows LP (10 s) spike-like signals. This synchronicity of two different band signals continues, while the exact periodicity may fluctuate. Such an interplay of different band signals forces us to view the origin of these signals as a system (i.e., volcanic fluid system).

Figure 6 is another such example of a broadband record of a small phreatic eruption observed at Aso Volcano, Japan (Kaneshima et al., 1996; Kawakatsu et al., 2000). While the integrated displacement record exhibits a VLP (>100 s) signal corresponding to the inflation and deflation stages of a

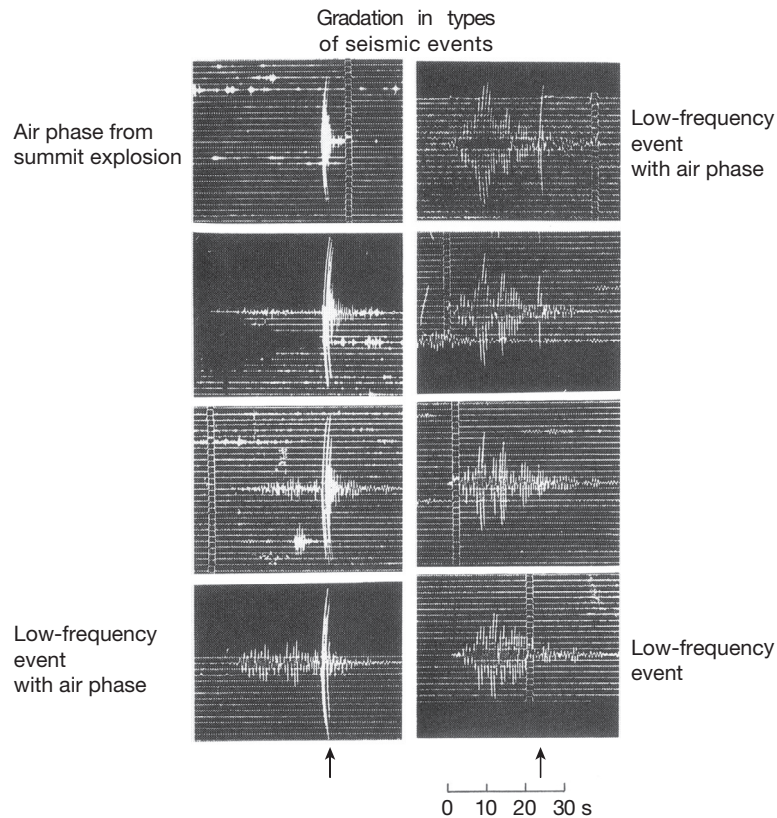


Figure 3 Range of volcanic seismic signals from the high-frequency air-wave phase to LF event observed at Langila. Arrows indicate timings of air-wave phases. Adapted from Mori J, Patia H, McKee C, et al. (1989) Seismicity associated with eruptive activity at Langila volcano, Papua New Guinea. *Journal of Volcanology and Geothermal Research* 38: 243–255.

crack-like conduit (Yamamoto et al., 1999) before and during the eruption, the raw velocity record exhibits short-period signals apparently due to the quick flow of volcanic fluids only seen in the deflation stage (i.e., eruption). Again, signals seen in two different bands allow us to vividly image the process of the eruption of a volcanic fluid system. Different physics apply at different frequencies, and this makes volcanic seismic signals essentially broadband.

4.15.2.2.1 Global observation

The last example (Figure 7) comes from the recordings of VLP (200–400 s) seismic signals generated by the eruption of Pinatubo Volcano, Philippines, observed at broadband stations around the world (Kanamori and Mori, 1992; Widmer and Zürn, 1992). The LP disturbances of the atmosphere generated by the eruption clouds are transmitted back to the solid Earth through the coupling between the two systems. Two spectral peaks observed in the seismic records are well explained by a theory that treats the solid Earth and the atmosphere as a single system (Watada and Kanamori, 2010); such a coupling between the solid and fluid parts of the Earth is also observed in the so-called background free oscillations of the Earth (Nishida et al., 2000). The broadband recordings of the Pinatubo eruption appear to have accelerated the emergence of a new class of seismology that treats the solid Earth, atmosphere, and ocean as a single system (cf. Chapter 4.14).

4.15.3 Description of Volcanic Seismic Sources

The diversity of observed volcanic seismic signals partly reflects the wide variety of seismic source geometries and dynamics present in active volcanoes. A general kinematic description of seismic sources in terms of ‘moment tensor’ and ‘single force’ has become common in volcano seismology, and the waveform inversion solving for them has become a general tool to understand the generation mechanisms of LP signals obtained with broadband instruments.

4.15.3.1 Equivalent Forces of General Seismic Sources

The concept of the seismic moment tensor was first introduced by Gilbert (1970) as a description of seismic sources more general than the conventional double-couple description. Later, Backus and Mulcahy (1976) gave a physical basis for such a general description within the framework of linear elastodynamics by introducing concepts of indigenous source and stress glut. An indigenous source is a seismic source, which originates within the Earth, and its equivalent force system exerts neither total force nor total torque at any instance to the Earth. Seismologists, however, soon realized that there exist such indigenous seismic events, which require single-force equivalent force systems (e.g., Hasegawa and Kanamori, 1987; Kanamori and Given, 1982; Kanamori et al., 1984;

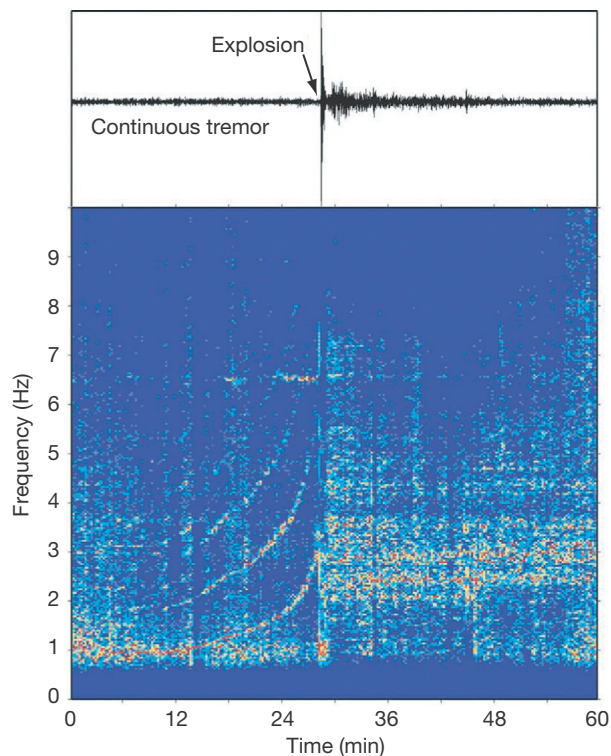


Figure 4 Example of tremor at Montserrat before an explosion. Adapted from Jousset P, Neuberg J, and Sturton S (2003) Modelling the time-dependent frequency content of low-frequency volcanic earthquakes. *Journal of Volcanology and Geothermal Research* 128: 201–223.

Kawakatsu, 1989); most of them happened to occur around active volcanoes. To circumvent this apparent conflict, Takei and Kumazawa (1994) extended Backus and Mulcahy's treatment to include fluid-dynamic description with a mass advection term and introduced a concept of inertial glut; they further gave a clear definition of single-force (and torque) source as a momentum (and angular momentum) exchange between the seismic source volume and the rest of the Earth. In the following, we briefly summarize their treatment.

To make the argument simple, we omit the Earth's self-gravity in the following; although gravity often takes the central role generating single forces in the actual Earth, this simplification does not alter the essence of the argument presented below (or in the following). The true Eulerian equation of motion of the Earth without external force may be written as

$$\rho^t D_t^2 u_j = \partial_i S_{ij} \quad [1]$$

where $\rho^t(x,t)$ denotes the true density, $u_j(x,t)$ is the displacement, $S_{ij}(x,t)$ is the true stress field, and D_t is the particle derivative $D_t = \partial_t + v_k \partial_k$ with $v_k(x,t)$ as the velocity. We compare this true equation of motion with the linear elastodynamic equation of motion

$$\rho^m \partial_t^2 u_j = \partial_i \varphi_{ij} \quad [2]$$

where $\rho^m(x,t)$ and $\varphi_{ij}(x,t)$, respectively, denote the model density and model stress field due to the linear stress–strain constitution equation (Hooke's law), which we use to describe the

wave propagation in the Earth. The equivalent body force $\gamma_j^Y(x,t)$ in the linear elastodynamic system arises as the difference of the two equations [1] and [2] as

$$\rho^m \partial_t^2 u_j = \partial_i \varphi_{ij} + \gamma_j^Y \quad [3]$$

where

$$\gamma_j^Y = \gamma_j^S + \gamma_j^D \quad [4]$$

with

$$\gamma_j^S = -\partial_i \Gamma_{ij} = -\partial_i (\varphi_{ij} - S_{ij}) \quad [5]$$

$$\gamma_j^D = \rho^m \partial_t^2 u_j - \rho^t D_t^2 u_j \quad [6]$$

$\Gamma_{ij}(x,t) = \varphi_{ij}(x,t) - S_{ij}(x,t)$ in eqn [5] is the stress glut introduced by Backus and Mulcahy (1976), which represents failure of Hooke's law due to some nonlinear effect in the source region and is zero outside of the source region. The second term in the right hand of eqn [4] $\gamma_j^D(x,t)$ is named as inertial glut by Takei and Kumazawa (1994) and is the difference of the inertial forces in the actual value and in the model, which is again zero outside of the source region. It can be shown that total force and total torque due to $\gamma_j^S(x,t)$ integrated over the volume are zero at any instance, while those due to the inertial glut $\gamma_j^D(x,t)$ are necessarily not.

The spatial integral of eqn [6] over the whole space

$$\gamma_j^{(0)D} = \int (\rho^m \partial_t^2 u_j - \rho^t D_t^2 u_j) dV \quad [7]$$

can be shown to be

$$\gamma_j^{(0)D} = -\partial_t^2 \int_{V_s} (\rho^t - \rho^m) u_j dV \quad [8]$$

which shows that $\gamma_j^{(0)D}(t)$ originates from a difference between ρ^t and ρ^m in the source region V_s and thus may not vanish. On the other hand, it is shown that the time integral of $\gamma_j^{(0)D}(t)$ during the event vanishes, that is,

$$\int_0^\infty \gamma_j^{(0)D} dt = 0 \quad [9]$$

the requirement that any indigenous source has to satisfy. The authors further defined single-force and torque sources as linear and angular momenta exchange between the source region and the rest of the Earth and showed that, in the LP limit, the single-force source is equal to $\gamma_j^{(0)D}(t)$.

4.15.3.2 The Seismic Moment Tensor

The spatial integral of stress glut is the seismic moment tensor M_{ij} , which describes the overall feature of the seismic source. When the wavelength of observed seismic waves is much longer than the spatial extent of the source, we may approximate the source by a point source moment tensor. Each component of M_{ij} corresponds to one set of opposing forces (dipole or force couple; Figure 8). The seismic moment tensor is a symmetrical second-order tensor and thus has six independent components, while a double-couple equivalent body force for a shear dislocation has only four degrees of freedom. These two extra degrees of freedom in the moment tensor are called non-double-couple

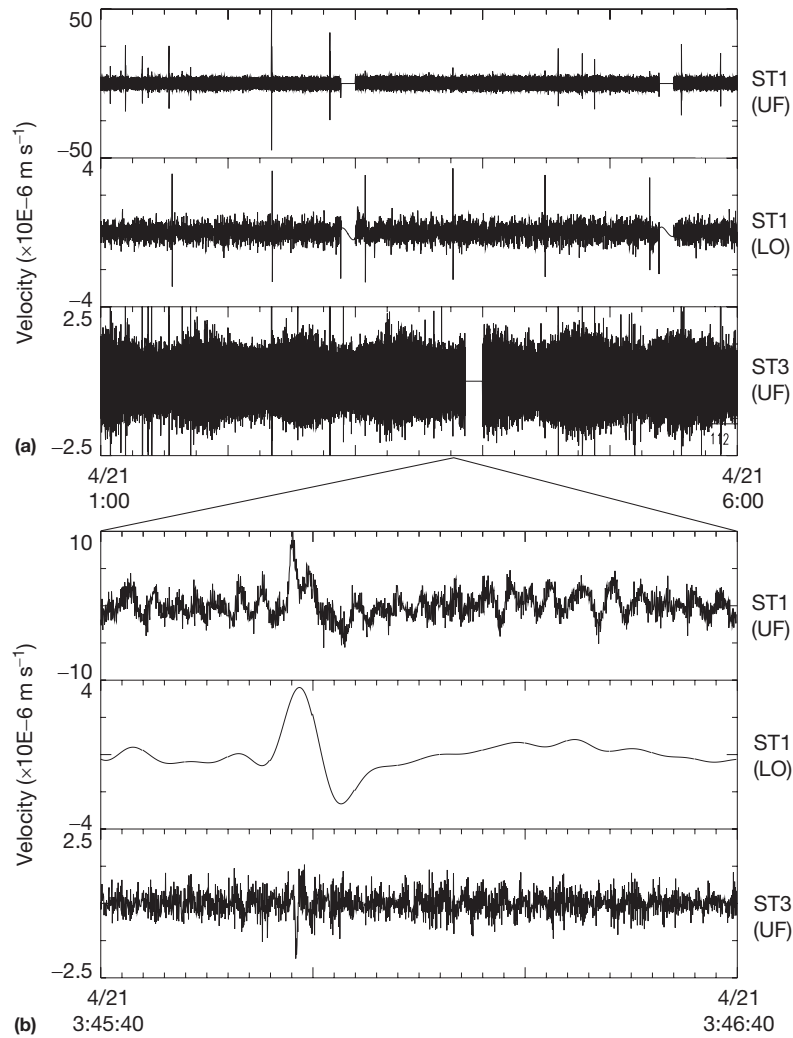


Figure 5 (a) Five-hour records of vertical components of velocity at Satsuma-Iwojima. A trace low-pass-filtered (LO) at 0.2 Hz is shown in between the two unfiltered records. (b) Expansion of the 1 min portion of the top panel. Adapted from Ohminato T and Ereditato D (1997) Broadband seismic observations at Satsuma-Iwojima volcano, Japan. *Geophysical Research Letters* 24: 2845–2848.

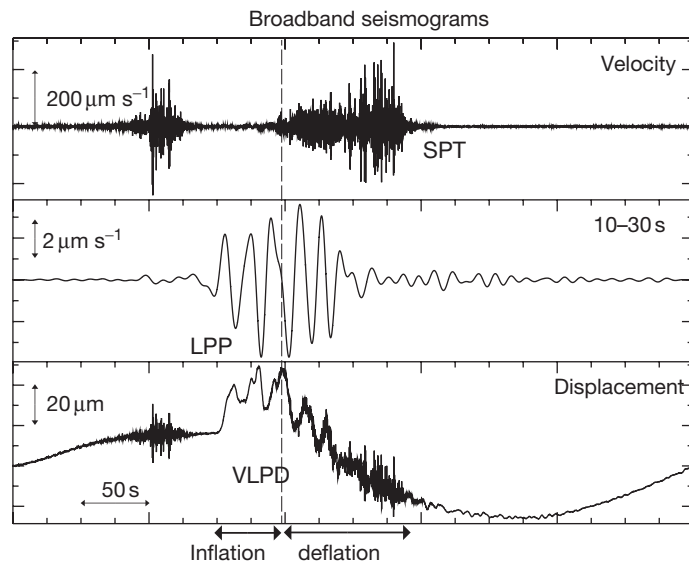


Figure 6 Broadband waveforms of a phreatic eruption observed at Aso. Adapted from Kawakatsu H, Kaneshima S, Matsubayashi H, et al. (2000) Aso94: Aso seismic observation with broadband instruments. *Journal of Volcanology and Geothermal Research* 101: 129–154.

components. For the general discussion on non-double-couple components, readers may refer to the following review articles (Julian et al., 1998; Miller et al., 1998). Seismic sources observed in the volcanic environments often show non-double-couple

components. The end-members of such non-double-couple sources are discussed below (or in the following).

4.15.3.2.1 Spherical source

Let us consider an initial condition that a reservoir contains a certain amount of volcanic fluid (V) under a static pressure P . When a sudden volume increase of ΔV (measured under the same pressure condition) is introduced in the reservoir, either due to an injection of a new volcanic fluid or a thermal expansion, the reservoir should expand and act as a seismic source. This volumetric increase ΔV corresponds to the stress-free volumetric strain introduced by Eshelby (1957) (Aki and Richards, 2002, p. 53) and characterizes the amount of the increase of the volcanic fluid. On the other hand, due to the confining pressure of the surrounding elastic medium (matrix), the actual volumetric increase of the reservoir ΔV_m may be smaller than ΔV , the ratio $\Delta V_m/\Delta V$ depending on the geometry of the reservoir, with a resultant pressure increase ΔP_m .

The corresponding moment tensor for a spherical reservoir is given as

$$M = \Delta V \begin{pmatrix} \lambda + \frac{2}{3}\mu & 0 & 0 \\ 0 & \lambda + \frac{2}{3}\mu & 0 \\ 0 & 0 & \lambda + \frac{2}{3}\mu \end{pmatrix} \quad [10]$$

where the ratio

$$\frac{\Delta V_m}{\Delta V} = \frac{\lambda + \frac{2}{3}\mu}{\lambda + 2\mu} \quad [11]$$

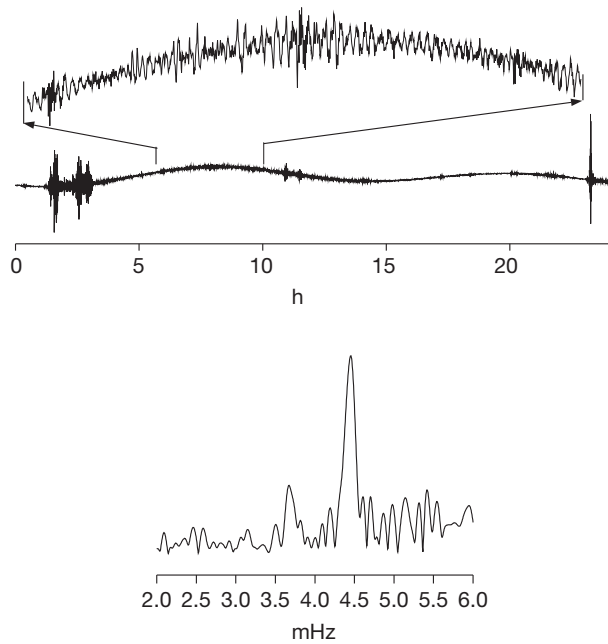


Figure 7 Broadband record of the Pinatubo eruption in 1991 observed in Japan. Adapted from Watada S and Kanamori H (2010) Acoustic resonant oscillations between the atmosphere and the solid earth during the 1991 Mt. Pinatubo eruption. *Journal of Geophysical Research* 115: B12319.

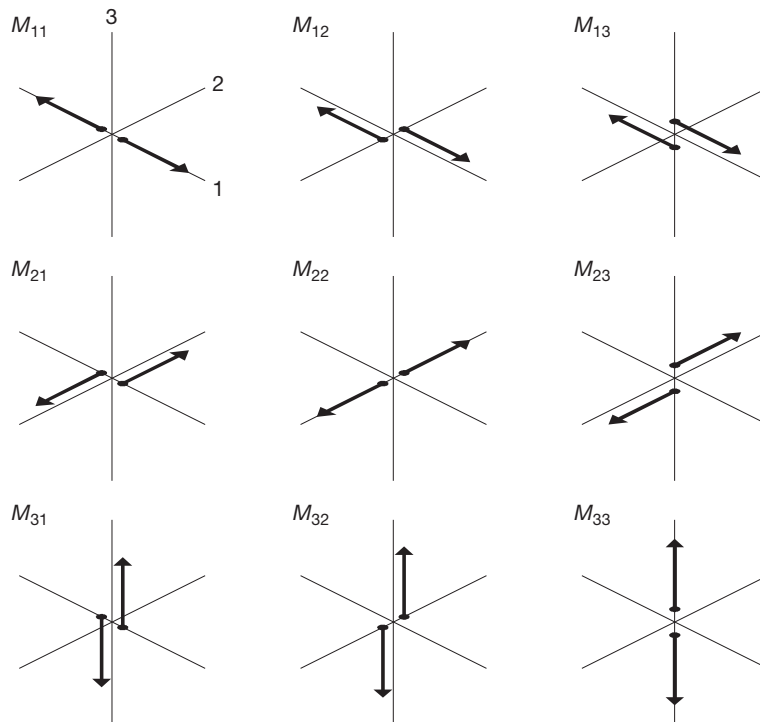


Figure 8 Nine force couples corresponding to moment tensor components.

and the actual volumetric strain and pressure increase may be related as

$$\Delta P_m = \frac{4}{3}\mu \frac{\Delta V_m}{V} \quad [12]$$

(Aki and Richards, 2002, p. 61). As the actual volumetric change ΔV_m is also the volume change often used in geodetic analysis (e.g., Mogi, 1958), we propose to call this 'Mogi volume' to distinguish it from the stress-free volume increase ΔV .

4.15.3.2.2 Cylinder

A moment tensor corresponding to the radial expansion of a cylinder whose symmetry axis is x_1 may be given as

$$M = \Delta V \begin{pmatrix} \lambda & 0 & 0 \\ 0 & \lambda + \mu & 0 \\ 0 & 0 & \lambda + \mu \end{pmatrix} \quad [13]$$

where

$$\frac{\Delta V_m}{\Delta V} = \frac{\lambda + \mu}{\lambda + 2\mu} \quad [14]$$

and

$$\Delta P_m = \mu \frac{\Delta V_m}{V} \quad [15]$$

4.15.3.2.3 Tensile crack

A moment tensor corresponding to an opening of a thin crack in the direction of x_1 -axis may be given as

$$M = \Delta V \begin{pmatrix} \lambda + 2\mu & 0 & 0 \\ 0 & \lambda & 0 \\ 0 & 0 & \lambda \end{pmatrix} \quad [16]$$

In case of a thin crack, two volumetric changes turn out to be the same. The Mogi volume here is given as $\Delta V = \Delta V_m = S \cdot \Delta u$, where S and Δu are the crack area and opening width, respectively, and it is often referred to as 'potency' in seismological literatures (e.g., Ben-Menahem and Singh, 1981).

4.15.3.2.4 Compensated linear vector dipole

The isotropic components of the moment tensors of these three volumetric sources, when expressed in terms of stress-free volumetric change ΔV , are the same:

$$I = \frac{1}{3} \text{trace}(M) = \left(\lambda + \frac{2}{3}\mu \right) \Delta V = \kappa \Delta V \quad [17]$$

where κ is the bulk modulus of the elastic matrix. When a movement of magma from a reservoir to another excites seismic waves, such a seismic source should be observed as a summation of two volumetric sources with opposite signs. Equation [17] indicates that the corresponding moment tensor has zero isotropic component called compensated linear vector dipole by Knopoff and Randall (1970) (Figure 9).

4.15.3.2.5 Determination and interpretation of a moment tensor

Due to structural complexities of volcanic edifices, the conventional first motion polarity analysis method to determine source mechanisms of volcanic events is unlikely to give reliable estimates of moment tensors. Although the amplitude of radiated seismic waves contains more information, often, the absolute amplitude also suffers from structural complexities. Methods that utilize relative amplitudes of waves, which share common ray paths, such as the spectral ratio method of Nishimura et al. (1995), the amplitude ratio method of Julian and Foulger (1996), and the relative moment tensor inversion method of Dahm (1996) appear to be better suited; among them, the linear-programming approach of Julian and Foulger (1996) has been extensively used to show the existence of many non-double-couple earthquakes in different volcanic/geothermal environments (e.g., Foulger et al., 2004; Ross et al., 1996).

An arbitrary moment tensor observed at volcanoes may be decomposed as a linear combination of those four end-member moment tensors (plus a double couple). As such a decomposition is nonunique, there can be different decompositions for a given moment tensor, and thus, there exists always nonuniqueness in interpreting moment tensor solutions in terms of actual phenomenon that is occurring at volcanoes. It is often a custom to choose the simplest possible one or to choose one that is most consistent with known source geometry of corresponding volcano (e.g., the presence of known crack-like conduit). Without such a priori information, on the other hand, one may wish to display a full moment tensor. The source-type plot introduced by Hudson et al. (1989) may be a useful tool to help interpret different moment tensors observed in different volcanic environments (e.g., Foulger et al., 2004; Figure 10). It displays moment tensors without regard to their orientations, and different types of

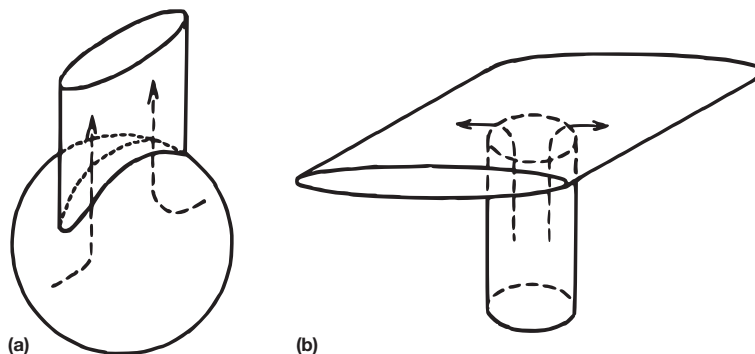


Figure 9 Geometry of selected transport models. Adapted from Chouet B (1996) New methods and future trends in seismological volcano monitoring. In: *Monitoring and Mitigation of Volcano Hazards*, pp. 23–97. New York: Springer-Verlag.

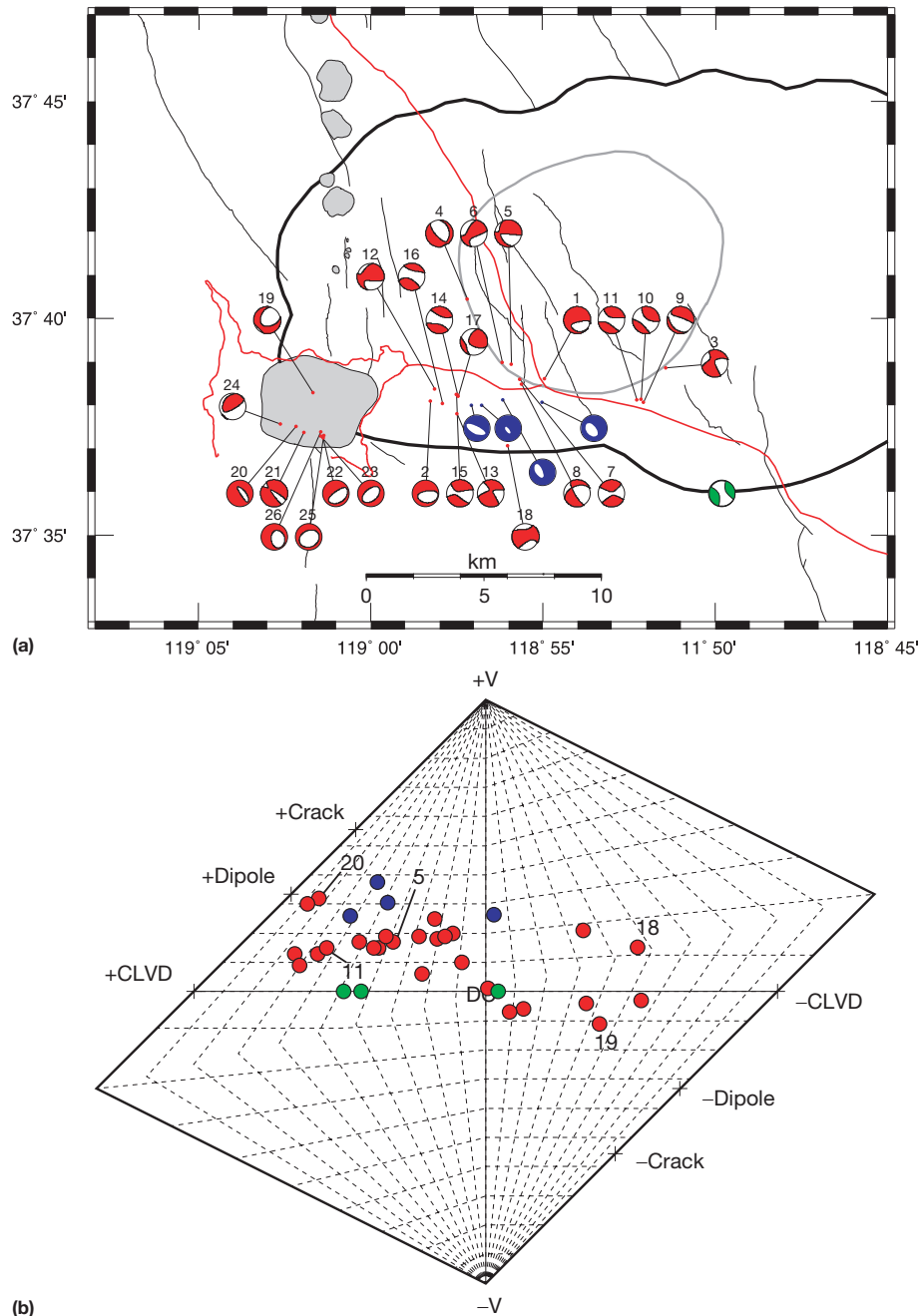


Figure 10 (a) Focal mechanism plot (upper focal hemisphere) of earthquakes in Long Valley. (b) Source-type plot of (Hudson et al., 1989) for the same earthquakes. The vertical axis measures the size of volumetric component, and deviatoric moment tensors are located on the horizontal axis (a pure double couple at the origin). Different colors indicate mechanism solutions determined by different researchers. Adapted from Foulger GR, Julian BR, Hill DP, Pitt AM, Malin PE, and Shalev E (2004) Non-double-couple microearthquakes at Long Valley caldera, California, provide evidence for hydraulic fracturing. *Journal of Volcanology and Geothermal Research* 132: 45–71.

moment tensors (different in relative magnitudes of principal moments) are projected on a rhombic plane in such a way that the area is proportional to the probability of the occurrence of different source types.

4.15.3.3 The Single Force

As described earlier, the single-force equivalent body force is due to a momentum exchange between the source volume and

the rest of the Earth. Whenever a part of the Earth is detached from the rest and gains a momentum, the counterforce of this acceleration is felt by the rest of the Earth. As the detached mass eventually has to stop somehow, there must be a deceleration stage in this process, giving another counterforce with the opposite direction. The total of these two counterforces must cancel out as required from eqn [9].

The single-force equivalence of LP seismic waves excited by a landslide and explosions associated with the 1980 eruption

of Mt. St. Helens was established by Kanamori and Given (1982) and Kanamori et al. (1984). At an LP limit, a landslide may be viewed as a sliding of a box along a slope (Hasegawa and Kanamori, 1987; Kawakatsu, 1989), and its single-force equivalence may be obvious. Kawakatsu (1989) developed a centroid single-force (CSF) inversion method, an extension of the centroid moment tensor inversion method of Dziewonski et al. (1981) to include single-force components, and applied it successfully to several landslide/slump events recorded by global seismic networks. Ekström et al. (2003) applied the CSF inversion to 'glacial earthquakes' observed in southern Alaska and showed that they are represented by stick-slip, downhill sliding of a glacial ice mass, which takes about 30–60 s in duration. Kanamori et al. (1984) also suggested that an equivalent force system for a volcanic eruption can be decomposed into a vertical downward single force and an isotropic moment tensor (Figure 11). A single-force source may be also realized when high-viscous magma flows along a narrow conduit, as suggested by Ukawa and Ohtake (1987) to explain the deep monochromatic LF (1 Hz) event observed beneath Izu-Oshima Volcano, Japan, about 1 year prior to the 1986 eruption. Chouet et al. (2003) also suggested that a piston-like rise of a slug of gas in the conduit produced observed single-force components in the LP signals associated with explosions in Stromboli Volcano, Italy.

4.15.3.4 Waveform Analysis of Volcanic Seismic Sources

The LP part of volcanic seismic signals has been recorded by seismic networks of regional and/or global scales. One of the

advantages of using such LP signals is that waveform analysis to infer excitation mechanisms can be easily performed compared to short-period seismograms. As mentioned, Kanamori and Given (1982) modeled the LP seismic signals accompanied with the eruption of the Mt. St. Helens in 1980 recorded at stations of the global seismic networks. They showed that the single-force mechanism can explain the radiation pattern of LP surface waves and attributed it to the gigantic mass movement of the volcano edifice due to a landslide. Kanamori et al. (1984) also analyzed the LP body waves generated by the same eruption and showed that vertical single forces due to the depressurization of the magma chamber explain the data. With these two single-force mechanisms, they could vividly picture what had occurred during the whole sequence of the eruption. These two studies initiated the applications of 'LP seismology' to study volcanic activities.

For LP seismic signals, it is often appropriate to treat the source originating from a point in space. Then, seismic wave field $u_n(x, t)$ may be written as

$$u_n(x, t) = G_{np,q}(x, x_0, t) * M_{pq}(t) + G_{np}(x, x_0, t) * F_p(t)$$

where $M_{pq}(t)$ and $F_p(t)$ denote a moment tensor and single-force source time function at x_0 , respectively, and $G_{np}(x, x_0, t)$ and $G_{np,q}(x, x_0, t)$, respectively, represent Green's function and its spatial derivative (e.g., Aki and Richards, 2002); the asterisk denotes convolution, and the summation convention for repeated indices is assumed. Knowing the Earth structure (i.e., knowing Green's function) and having enough waveform data, it is then possible to solve for the time history of general seismic sources, $M_{pq}(t)$ and $F_p(t)$.

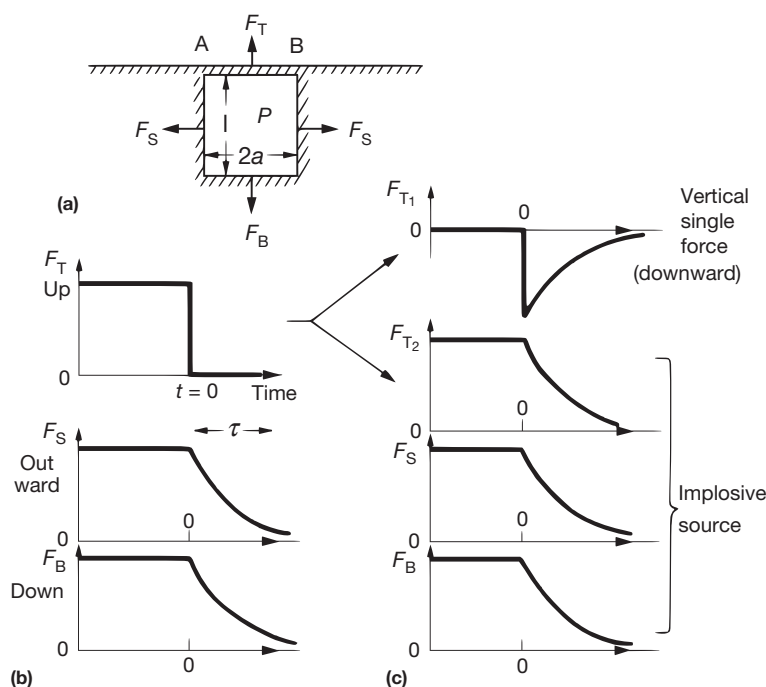


Figure 11 A force system equivalent to a volcanic eruption. (a) A pressure release model for a volcanic eruption. (b) Forces acting on the top, side, and bottom walls. (c) Decomposition of the force to a vertical single force and an impulsive force. Adapted from Kanamori H, Given JW, and Lay T (1984) Analysis of seismic body waves excited by the Mount St. Helens eruption of May 18, 1980. *Journal of Geophysical Research* 89: 1856–1866.

Takeo et al. (1990) extended the waveform moment tensor inversion technique commonly used in the analysis of far-field records in earthquake seismology (e.g., Kikuchi and Kanamori, 1982) to volcano seismology by including the single-force contribution. They applied it to the records observed in a lava drain-back stage of the 1987 eruption of Izu-Oshima Volcano, Japan, and showed that the resolved vertical single forces with opposing directions are due to a rapid collapse of overloading lava into the deeper vent. Ohminato et al. (1998) applied a similar technique to near-field broadband records observed at Kilauea Volcano, USA, and showed that LP signals can be explained by a subhorizontal crack, which periodically injects magma into a larger reservoir. Chouet et al. (2003, 2005) and Ohminato (2006) further incorporated the effect of complex topography of volcanic edifice that may significantly distort seismic signals observed in the near-field by calculating Green's functions using the finite difference scheme developed by Ohminato and Chouet (1997).

The general source analysis for LP signals using moment tensor and single force has become common in seismological studies of volcanic activities. Examples of such studies include Mt. St. Helens (Kanamori and Given, 1982; Kanamori et al., 1984; Kawakatsu, 1989), Long Valley (Aki, 1984; Dreger et al., 2000; Julian, 1983; Wallace, 1985), and Kilauea, USA (Chouet et al., 2010; Eissler and Kanamori, 1987; Ohminato et al., 1998); Asama (Ohminato et al., 2006; Takeo et al., 1984), Izu-Oshima (Takeo et al., 1990), Sakurajima (Uhira and Takeo, 1994), Unzen (Uhira et al., 1994b; Yamasato et al., 1993), Ito-oki (Takeo, 1992), and Tori-shima, Japan (Kanamori et al., 1993); Hengill-Grensdalur volcanic complex (Julian et al., 1997), Bardarbunga (Nettles and Ekstrom, 1998), Iceland; Stromboli, Italy (Chouet et al., 2003, 2008); and Popocatepetl, Mexico (Chouet et al., 2005). Applications of such analysis techniques for shorter-period (~ 1 s) near-field records obtained at volcanoes are also becoming available (e.g., Aoyama and Takeo, 2001; Kumagai et al., 2005; Nakamichi et al., 2003; Nakano et al., 2003; Nishimura et al., 1995; Ohminato, 2006; Uhira et al., 1994a).

4.15.4 Physical Mechanisms for Volcanic Seismic Signals

The presence of ample fluid (gas, vapor, magma, or their mixtures) in volcanic edifice introduces an additional interesting class of vibration/wave phenomena to the well-studied elastic formulations. For example, the so-called crack wave, which is introduced by Chouet (1986) to explain the LF nature of volcanic events, may be understood as a part of a class of waves due to solid/fluid coupling. Julian (1994) showed that the nonlinear effect of such solid/fluid coupling due to fluid flow transient can generate a wide variety of waveforms, which have similar characteristics as volcano seismic signals. This section summarizes some such topics relevant to generation of volcano-specific seismic signals.

4.15.4.1 Slow Waves in Solid/Fluid Composite

There exists a class of waves in the two-phase system of a solid/liquid composite that travel slower than any of the sound velocities of the pure material constituting the

two-phase system (i.e., the wave speed can be slower than that of the liquid phase). These waves include the tube wave (Biot, 1952), the so-called Biot's slow wave in a porous medium (Biot, 1956), the so-called crack wave (Chouet, 1986; Ferrazzini and Aki, 1987), and waves in solid/liquid alternating layers (Schoenberg and Sen, 1983) (Rayleigh wave and Stoneley wave may also be included in the same class, considering that in the high-frequency limit, all the previous slow waves have the same characteristics as an interface wave).

The simplest (and the most relevant to volcano seismology) example of slow waves may be the plane-wave propagation in a medium, which consists of a fluid layer sandwiched by two elastic half-spaces. For this problem, Krauklis (1962) first derived an analytic solution, and later, Ferrazzini and Aki (1987) reformulated the problem aiming to investigate the physical basis of the crack wave found by Chouet (1986). Here, we follow Yamamura (1997) to formulate the problem.

Following Aki and Richards (2002, p. 263), we first express a plane-wave solution propagating parallel to the fluid layer as

$$\begin{pmatrix} u \\ w \end{pmatrix} = \begin{pmatrix} r_1(z) \\ ir_2(z) \end{pmatrix} e^{i(kx - \omega t)} \quad [18]$$

where u and w are, respectively, the displacements for x - and z -directions. The frequency domain equation of motion in x -direction for the fluid layer may then be written as

$$-\omega^2 \rho_f r_1 = -k^2 \kappa_f r_1 - k \kappa_f \frac{\partial r_2}{\partial z} \quad [19]$$

where ρ_f and κ_f denote density and bulk modulus of the fluid, respectively. Integrating [19] over the thickness of the fluid layer (i.e., z -direction) with appropriate boundary conditions (e.g., free-slip condition at the solid/fluid interface), we get the following expression:

$$-\omega^2 \rho_f \int_0^H r_1 dz = -k^2 \kappa_f \int_0^H r_1 dz - k \kappa_f r_2(H) \quad [20]$$

where we assume a symmetrical solution for the z -dependency (Ferrazzini and Aki, 1987) and the thickness of the fluid layer $2H$.

Equation [20] clearly demonstrates the role of the interface: without the last term $r_2(H)$, eqn [20] indicates the averaged displacement in the fluid layer traveling with its own sound velocity; that is, the term represents the way the fluid layer is coupled with the surrounding elastic half-space. If we further rewrite eqn [20] as

$$-\omega^2 \rho_f \int_0^H r_1 dz = -k^2 \kappa_f \left[1 + \frac{1}{k} \frac{r_2(H)}{\int_0^H r_1 dz} \right] \int_0^H r_1 dz \quad [21]$$

it is seen that

$$\kappa_f^e \equiv \kappa_f \left[1 + \frac{1}{k} \frac{r_2(H)}{\int_0^H r_1 dz} \right]$$

represents the effective bulk modulus for the fluid layer; when $\kappa_f^e < \kappa_f$, that is,

$$\frac{r_2(H)}{\int_0^H r_1 dz} < 0 \quad [22]$$

the corresponding traveling wave has a smaller wave speed relative to the sound velocity; thus, we have a slow wave. The condition [22] is satisfied if the denominator and numerator have opposite signs to each other. This makes quite good sense intuitively; when the fluid 'tries' to contract (or expand) as a passage of the wave, due to the 'elastic coupling' along the solid/fluid boundary, there exists a solution in which the solid behaves out of phase to the fluid to expand (contract), deforming the boundary in such a way that fluid 'feels' easy to contract (expand), compared to the single-phase case; that is, for a given amount of displacement in the propagation direction, the actual pressure increment of the fluid phase becomes lower than that of the pure fluid case (Figure 12). The net effect of this elastic coupling is to reduce the 'effective bulk modulus' (i.e., the restoring force) of the fluid, and as a result, the wave speed becomes slower than that of the pure fluid; the stronger the elastic coupling, the slower the wave speed (Figure 13).

This kind of the out-of-phase condition is satisfied by other slow waves such as Biot's slow wave in the porous medium

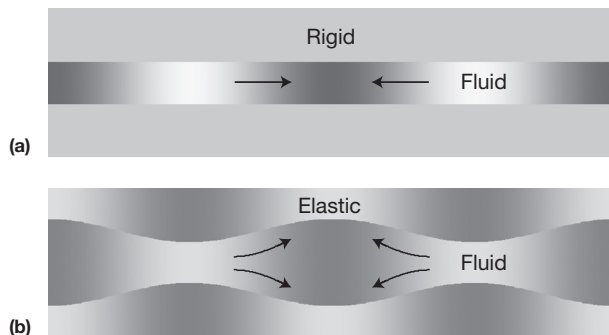


Figure 12 Conceptual image of the slow wave in solid/fluid composites. Out-of-phase motion of the solid (b) lowers the 'effective bulk modulus' of the fluid, and as a result, the wave speed becomes slower than that in the pure fluid (a). (after Yamamoto, 2005)

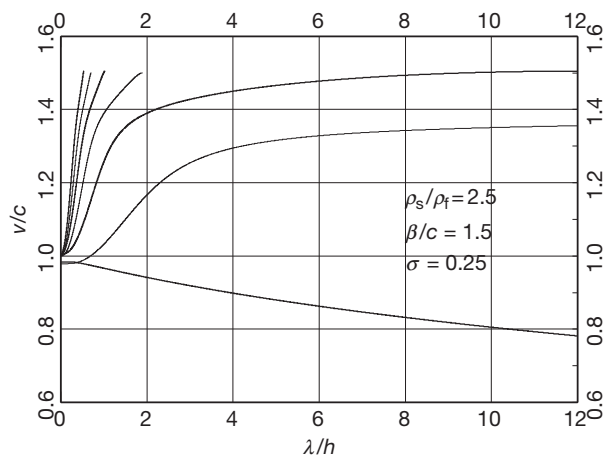


Figure 13 Phase velocity divided by the acoustic velocity of the fluid is plotted as a function of the wavelength divided by the fluid layer thickness. The fundamental branch having a phase velocity below 1 corresponds to the slow wave. Adapted from Ferrazzini V and Aki K (1987) Slow waves trapped in a fluid-filled infinite crack: Implication for volcanic tremor. *Journal of Geophysical Research* 92: 9215–9223.

(Biot, 1956) and waves in solid/fluid alternating layers (Schoenberg and Sen, 1983). Indeed, using the argument presented above, Yamamura (1997) was able to derive a similar set of equations to describe the elastic coupling for the Schoenberg's problem to those Biot's equations of motion and suggested that all these slow waves have a similar characteristic as an interface wave and that, in the short-wavelength limit, they reduce to the Stoneley wave.

4.15.4.2 Resonating Sources: A Crack

The presence of regularly spaced spectral peaks often observed in volcanic seismic signals immediately reminds us of the presence of resonators within the volcanic edifice filled with volcanic fluid. To quantitatively model the volcanic tremor observed at Kilauea, Hawaii, Aki et al. (1977) considered fluid-filled tensile cracks as the most plausible resonating sources. Using the two-dimensional finite difference method, they computed the dynamics of a series of fluid-filled cracks connected by narrow channels, which are excited by jerky extension/opening of channels due to excess fluid pressure. Although their model offers both the driving force and the geometry adequate for the magma transport beneath active volcanoes, the fluid inside the crack acts passively like a cushion so that the stress at the crack wall depends only on local displacements of the crack wall; the fluid also does not support the propagation of pressure perturbation caused by the deformation of the crack wall. To assess active roles of the fluid, Chouet and Julian (1985) extended Aki et al.'s model and investigated dynamic interaction between fluid and elastic solid inside and outside a two-dimensional crack by solving equations for the elastic solid and the fluid simultaneously. The model was further extended to three dimensions by Chouet (1986), and Yamamoto and Kawakatsu (2008) also formulated the two-dimensional problem using the boundary integral method.

Chouet's three-dimensional model consists of a single isolated fluid-filled crack in an infinite elastic solid body, and no external mass transfer into and/or out of the crack is taken into consideration. In the model, the thickness of the crack d is assumed to be much smaller than the wavelength of interest, and thus, the motion of the fluid inside the crack is treated as two-dimensional. Assuming Poisson's ratio to be 0.25 (i.e., $\lambda = \mu$) and taking L and L/α as the characteristic length and time, where L and α are the length of the crack and the P wave velocity in the solid, the response of the crack is calculated by solving the following dimensionless equations simultaneously with boundary conditions at the crack surface and perimeters:

$$\begin{aligned} \text{Solid : } & \frac{\partial \tilde{v}_i}{\partial \tilde{t}} = \frac{1}{3} \frac{\partial \tilde{\tau}_{ik}}{\partial \tilde{x}_k} \\ & \frac{\partial \tilde{\tau}_{ij}}{\partial \tilde{t}} = \frac{\partial \tilde{v}_k}{\partial \tilde{x}_k} \delta_{ij} + \frac{\partial \tilde{v}_i}{\partial \tilde{x}_j} + \frac{\partial \tilde{v}_j}{\partial \tilde{x}_i} \\ \text{Fluid : } & \frac{\partial \tilde{U}_1}{\partial \tilde{t}} = -\frac{1}{3} \frac{\rho}{\rho_f} \frac{\partial \tilde{P}}{\partial \tilde{x}_1} \\ & \frac{\partial \tilde{P}}{\partial \tilde{t}} = -\frac{\kappa_f}{\mu} \frac{\partial \tilde{U}_m}{\partial \tilde{x}_m} - 2 \frac{\kappa_f L}{\mu d} \tilde{v}_z \\ & (\tilde{x} = x/L, \tilde{t} = t\alpha/L) \end{aligned} \quad [23]$$

In these equations, the quantities with tilde represent dimensionless ones, and \tilde{v} and $\tilde{\tau}$ are the velocity and the stress

of the solid; \vec{U} and \vec{P} are the velocity and the pressure of the fluid, ρ and ρ_f are the density of the solid and the fluid, and μ and κ_f are the rigidity of the solid and the bulk modulus of the fluid, respectively. As shown in eqn [23], the behavior of fluid-filled crack critically depends on a nondimensional parameter $c = \kappa_f / \mu \cdot L/d$, which was first introduced by Aki et al. (1977) and named as 'crack stiffness' by Chouet (1986).

Synthetic seismograms calculated by the fluid-filled model show strong similarities to observed volcanic signals in terms of both peaked spectra and long-lasting oscillations and demonstrated the importance of active participation of fluids in source dynamics generating volcanic seismic signals. These studies include the interpretations of volcanic signals observed at Redoubt, Alaska (Chouet et al., 1994); Galeras, Colombia (Gil Cruz and Chouet, 1997; Figure 14), Kusatsu-Shirane, Japan (Kumagai et al., 2002; Nakano et al., 1998), Kilauea, Hawaii (Kumagai et al., 2005), and other volcanoes.

Kumagai and Chouet (1999, 2000) and Morrissey and Chouet (2001) further studied the dependencies of frequencies and attenuations of crack-resonator oscillations on the properties of fluid in more detail using various models of fluid-gas mixtures and fluid-particle mixtures and demonstrated that the fluid properties and compositions can be estimated from frequencies and attenuations of observed signals. These results are used to assess the change of magmatic and hydrothermal system beneath the volcano at, for

example, Kusatsu-Shirane, Japan (Kumagai et al., 2002; Figure 15), Tungurahua, Ecuador (Molina et al., 2004), and Etna, Italy (Lokmer et al., 2008).

4.15.4.3 Other Resonating Sources

Although the presence of regularly spaced spectral peaks may not straightforwardly warrant the presence of a resonator (Julian, 1994), it seems indeed natural to think of such presence like a crack as a dike, a cylinder as a conduit, and a sphere as a magma chamber, as they are all fundamental constituents of a volcano.

Resonators as models of the origins of volcanic seismic signals besides a crack have been suggested for a cylinder (Chouet, 1985; Ferrick et al., 1982), and a sphere (Kubotera, 1974; Sassa, 1935). Crosson and Bame (1985) obtained an impulsive response of a spherical magma chamber with a spherical cavity inside. Fujita et al. (1995) solved eigenoscillation of a fluid sphere embedded in an infinite elastic medium. Fujita and Ida (2003) conducted a systematic survey of the geometric effects on the eigenoscillations and showed that the ratios of higher modes to the fundamental mode frequencies can be indicators of the geometry of the resonator. They also found a class of modes with a low attenuation named 'low-attenuation mode' (LAM), which share the similar out-of-phase characteristics of the slow waves discussed earlier, and suggested that LAM may be the origin of long-lasting LF oscillation of volcanic seismic signals.

The complicated nature of vibrations of cylindrical conduits has been numerically studied (e.g., Jousset et al., 2003; Neuberg et al., 2000; Nishimura and Chouet, 2003). Neuberg et al. (2000) introduced a depth-dependent seismic velocity model to account for the varying gas content in the magma and showed that a pressure change in the conduit was the most likely candidate for the physical process causing the peak-spectral shift often observed in many of volcanoes (Figure 4). Jousset et al. (2003) also showed that the endpoints of the conduit can act as secondary sources of volcanic seismic signals.

The excitation mechanism of these resonators is, on the other hand, not well understood. The proposed models include a jerky extension of the crack tip (Aki et al., 1977), acoustic emissions from collapsing bubbles (Chouet, 1992), rapid discharge of fluids (Ferrick et al., 1982), and a sudden pressure drop caused by unsteady choked flow (Morrissey and Chouet, 1997). Nakano et al. (1998) developed a method to extract the effective excitation function from seismic records based on an inhomogeneous autoregressive model of a linear dynamic system. Nakano et al. (2003) further extended the method to deal with the nonorthogonal nature of eigenfunctions of a resonator embedded in a rock matrix by using the biorthogonal eigenfunction expansion method of Yamamura and Kawakatsu (1998); the resolved effective excitation functions of the LF events in Kusatsu-Shirane Volcano, Japan, are then modeled by the waveform inversion method of Ohminato et al. (1998) to obtain the moment tensor and single-force source time function, which are interpreted as a result of repeated activation of a subhorizontal crack located beneath the summit crater lake.

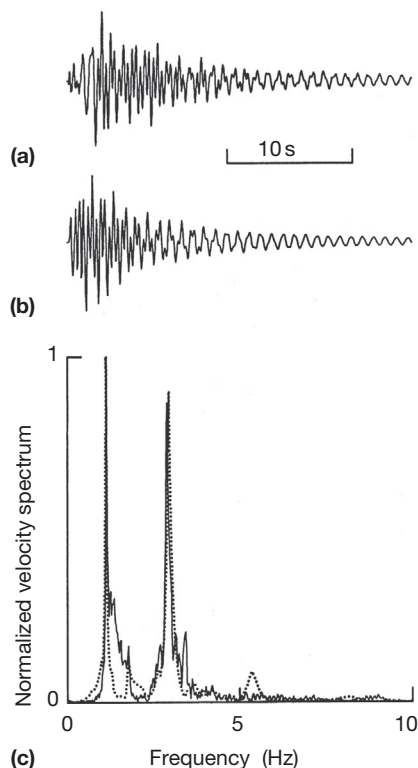


Figure 14 Example of the application of the fluid-filled crack model: (a) Vertical velocity seismogram observed at Galeras Volcano, Colombia. (b) Synthetic waveform obtained from the fluid-filled crack model. (c) Spectrum of observed data (thin line) and synthetic data (dotted line). Adapted from Chouet B (1996) Long-period volcano seismicity: Its source and use in eruption forecasting. *Nature* 380: 309–316.

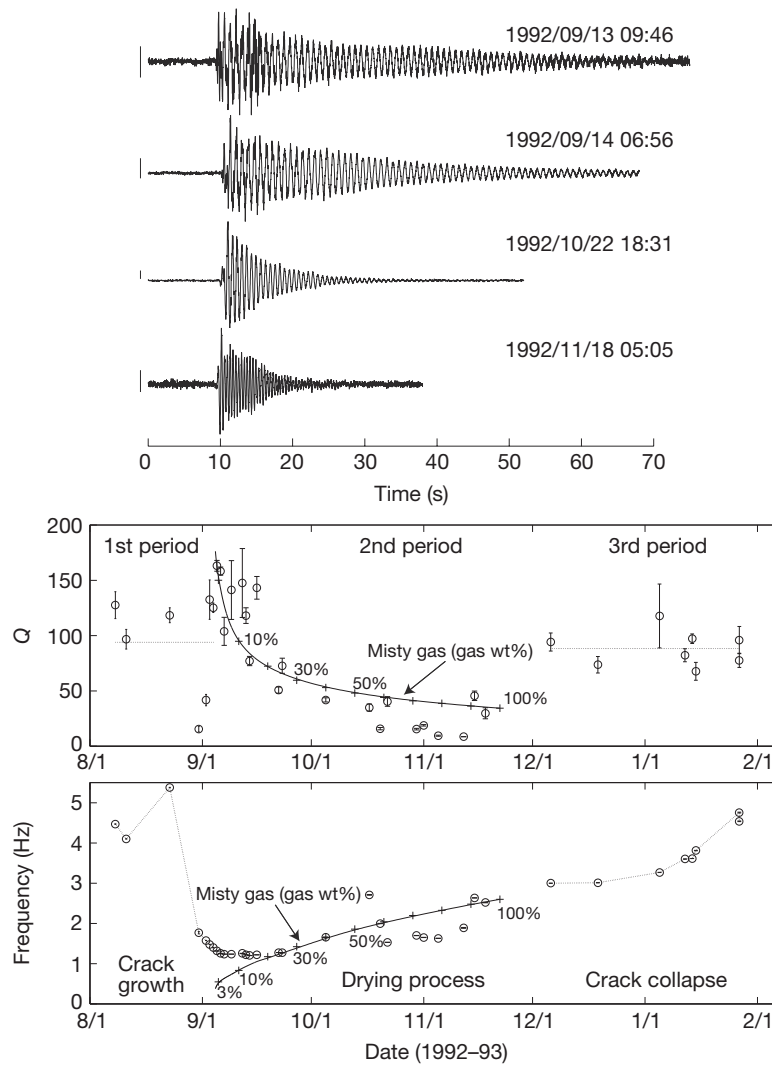


Figure 15 Temporal change of waveforms observed at Kusatsu-Shirane Volcano and the fluid properties estimated from the frequency and the attenuation of signals. Adapted from Kumagai H, Chouet B, and Nakano M (2002) Temporal evolution of a hydrothermal system in Kusatsu-Shirane Volcano, Japan, inferred from the complex frequencies of long-period events. *Journal of Geophysical Research* 107: ESE 1–ESE 9.

4.15.4.4 Flow-Induced Oscillation

Julian (1994) proposed nonlinear flow-induced oscillations in channels transporting volcanic fluid as a possible generating mechanism for volcanic seismic signals, particularly for sustained volcanic tremors. The mechanism may have analogs such as the vibration of vocal chords and musical wind instruments, and the periodic behavior, often characteristic of volcanic seismic signals, occurs without the presence of a resonator.

Figure 16 shows the simplified lumped-parameter model of the flow of a viscous fluid through a constricted channel with elastic walls. The channel walls are modeled as masses connected by a spring and dashpot to represent the effects of the elastic stiffness and radiation loss. The mechanism may be summarized as follows: an increase in a flow speed leads to a decrease in fluid pressure by the Bernoulli effect; as a result, the walls narrow the channel to constrict the flow, causing a pressure increase and forcing the channel open again; this again

decreases the pressure and increases the flow speed. This positive feedback mechanism sustains the oscillation.

Julian (1994) derived a third-order nonlinear system of ordinary differential equations that describe the complete behavior of the system and may be solved numerically. The system without radiation damping exhibits two different types of behavior: oscillations that grow and approach a stable tremor-like limit cycle and transient oscillations that decay, leading to a steady flow. Inclusion of the damping further complicates the behavior, as shown in Figure 17. By increasing the upstream pressure, the system experiences the so-called period doubling, that is, an appearance of a subharmonic with a frequency half that of the fundamental frequency; further increase with successive period doubling eventually leads to chaotic behaviors. Julian suggests that these features resemble behavior of some of the volcanic tremors and the deep LF event observed in Hawaii (Aki and Koyanagi, 1981).

Balmforth et al. (2005) extended Julian's treatment to include the dynamic behavior of the fluid and the elasticity

of the surrounding walls and considered the excitation of propagating waves. Their results confirm that Julian's mechanism is plausible, but the physical condition required for the instability is different. They conclude that magma itself is unlikely to generate flow-induced oscillations but that the rapid flow through fractured rock of low-viscosity fluids exsolved from magma may. Similar flow-induced oscillations are also studied by [Ida \(1996\)](#), [Corona-Romero et al. \(2012\)](#), and [Dunham and Ogden \(2012\)](#).

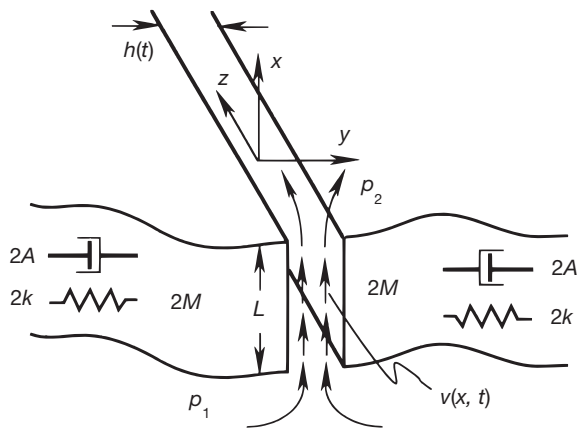


Figure 16 Lumped-parameter model of the generation of volcanic tremor. Viscous, incompressible fluid flows in x -direction from upstream (bottom) to downstream (top) reservoir through a channel with imperfectly elastic walls. Adapted from Julian BR (1994) Volcanic tremor: Nonlinear excitation by fluid flow. *Journal of Geophysical Research* 99: 11859–11878.

4.15.4.4.1 In fluid/gas two-phase system

The flow-induced oscillation described earlier assumes flows of a single-phase fluid. Volcanic fluids such as magma often contain ample gases, especially at a shallow depth, and it may be more reasonable to consider a flow of two-phase fluids consisting of liquid and gas phases rather than single-phase fluids as a source of volcanic seismic signals. Such a flow of liquid/gas two-phase fluid is also seen in, for example, the evaporation pipes in boiler systems, nuclear reactor cooling systems, and chemical industry equipments, and many studies have been conducted extensively in the field of engineering. These studies revealed the existence of flow-induced oscillatory phenomena, which is not observed in a flow of single-phase fluid, caused by complex interactions between two phases.

[Iwamura and Kaneshima \(2005\)](#) considered one of these flow-instability phenomena called 'density wave oscillation,' which was originally studied in the nuclear reactor engineering, as a model of volcanic seismic signals, and evaluated the properties of generated seismic waves through a series of numerical experiments. The density wave oscillation and limit cycle of the system was also studied by [Fujita \(2008\)](#) and applied to the analysis of the banded tremor at Miyakejima Volcano. Another kind of flow-induced oscillation in fluid/gas two-phase flows called 'pressure drop oscillation' was also invoked to interpret the cyclic tilt change observed at Miyakejima Volcano by [Fujita et al. \(2004\)](#).

4.15.4.5 Bubble Dynamics

Gases dissolved in magma are one of the main driving forces to cause various volcanic phenomena. The presence of gas bubbles in magma decreases the density of magma and drives the

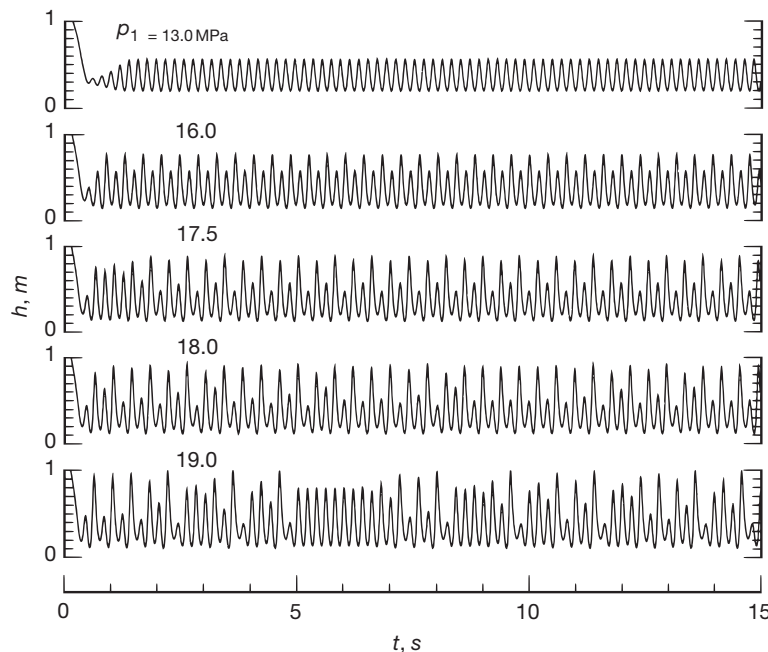


Figure 17 Synthetic tremor time series (channel thickness h vs. time) for different values of the upstream pressure p_1 . Behavior changes from a simple limit cycle (top) to chaos (bottom), with some period doubling in between. Adapted from Julian BR (1994) Volcanic tremor: Nonlinear excitation by fluid flow. *Journal of Geophysical Research* 99: 11859–11878.

magma ascent from the deep part of volcanoes (e.g., [Wilson and Head, 1981](#)). At shallower depths, the relatively high compressibility of bubbles contributes to the generation of oscillatory phenomena specific to volcanoes (e.g., [Chouet, 1996a](#); [Kieffer, 1977](#)), and the transformation and the fragmentation of bubbles affect the behavior of volcanic eruptions (e.g., [Alidibirov and Dingwell, 1996](#); [Ichihara et al., 2002](#)).

In relation to volcano seismology, there are some aspects where bubbles play important roles. As described earlier, volcanic seismic signals are often characterized by long-lasting LF oscillations with sharp spectral peaks ([Figure 1](#)), which are reminiscent of oscillations of some resonator filled with material with a slow acoustic velocity. Since the existence of bubbles in magma drastically changes the compressibility of magma, as pointed out by [Kieffer \(1977\)](#), these characteristics of volcanic seismic signals have been partly attributed to the presence of bubbles in magma. The rapid temporal changes of the observed spectral peaks (e.g., [Benoit and McNutt, 1997](#); [Neuberg et al., 2000](#)) have also been considered to reflect the change in the state of gas bubbles in volcanic conduits.

Bubbles in magma also act as an active source of seismic and acoustic signals: The bursting and the collapsing of bubbles at the lava surface preceding and/or accompanying volcanic eruptions generate acoustic pressure signals, which are sometimes accompanied with seismic waves (e.g., [Ripepe et al., 1996](#); [Vergnolle et al., 1996](#)); the collapse of small bubbles in hydrothermal fluids generates seismic tremors (e.g., [Leet, 1988](#)); the movement of large-scale bubbles (slugs) in magma conduit causes an exchange of their momentum with the surrounding rocks and generates seismic signals that show dominant single-force components (e.g., [Chouet et al., 2003, 2005, 2008, 2010](#); [Ohminato et al., 1998](#); [Ripepe and Gordeev, 1999](#)); the forced oscillations of bubbles by strain waves can rapidly pump dissolved volatiles into bubbles by a mechanism called 'rectified diffusion' and the resultant pressure increase may trigger volcanic activities (e.g., [Brodsky et al., 1998](#); [Ichihara and Brodsky, 2006](#)).

In spite of the difference in their generation mechanisms and resultant phenomena, all of these phenomena are the manifestations of passive and active processes in which the thermochemical energy and/or the gravitational energy of the volcanic fluids is transduced into seismic/acoustic energy. The quantitative understanding of the behavior of bubbles, thus, is a critical key to elucidate the dynamics of volcanic phenomena.

4.15.5 Observation and Analysis Aspects

In addition to the conventional monitoring of active volcanoes using short-period seismometers and geodetic instruments, the digital era has brought new powerful components to seismometry of active volcanoes; broadband and array seismometries may be two of the most notable developments.

4.15.5.1 Broadband Seismometry

The first broadband seismic observation at an active volcano was conducted by [Sassa \(1935\)](#), who installed Wiechert

horizontal-component seismographs (pendulum period $T_0=10.0$ s) and vertical ones ($T_0=4.6$ s), Gallitzin seismographs ($T_0=8.0$ s), and short-period seismographs ($T_0=0.55$ s) at Aso Volcano, Japan. Summarizing the subsequent observations in a period of several years, Sassa classified volcanic seismic signals of Aso into 'eruption earthquakes' and four different kinds of tremors: periods of 0.2, 0.4–0.6, 0.8–1.5, and 3.5–8.0 s. The observation of these tremors in the wide frequency band ranging from 0.2 to 8.0 s clearly demonstrates the efficacy of broadband seismometry at active volcanoes. Since Sassa's pioneering work, until 1990s (when it has become much easier to install portable broadband instruments), there have not been many published reports on attempts to observe LP volcanic seismic signals except for a few cases. For example, [Seidl et al. \(1981\)](#) installed broadband seismometers at Etna, Italy, and observed 4–5 s period signals, which they suggested as one of the fundamental peaks of the Etna tremors.

Examples of deployments of the current-generation broadband seismometers are now plentiful, for example, Sakurajima ([Kawakatsu et al., 1992, 1994](#)), Unzen ([Uhira et al., 1994b](#); [Yamasato et al., 1993](#)), Aso ([Kaneshima et al., 1996](#); [Kawakatsu et al., 2000](#); [Legrand et al., 2000](#); [Yamamoto et al., 1999](#)), Satsuma-Iwojima ([Ohminato, 2006](#); [Ohminato and Ereditato, 1997](#)), Iwate ([Nishimura et al., 2000](#)), Miyake-jima ([Fujita and Ida, 2003](#); [Kobayashi et al., 2009](#); [Kumagai et al., 2001](#)), Usu ([Yamamoto et al., 2002](#)), Bandai ([Nishimura et al., 2003](#)), Hachijo-jima ([Kumagai, 2006](#)), and Asama, Japan ([Ohminato et al., 2006](#)); Stromboli, Italy ([Chouet et al., 2003](#); [Dreier et al., 1994](#); [Falsaperla et al., 1994](#); [Neuberg et al., 1994](#)); Semeru ([Hellweg et al., 1994](#)) and Merapi, Indonesia ([Hidayat et al., 2000, 2002](#)); Arenal, Costa Rica ([Hagerty et al., 2000](#)); Kilauea ([Almendros et al., 2002a](#); [Chouet and Dawson, 2011](#); [Chouet et al., 2010](#); [Dawson et al., 1998](#); [Ohminato et al., 1998](#)), Long Valley ([Hill and Prejean, 2005](#); [Hill et al., 2002](#)), and St. Helens, USA ([Waite et al., 2008](#)); Erebus, Antarctica ([Aster et al., 2003, 2008](#); [Rowe et al., 1998, 2000](#)); Popocatepetl, Mexico ([Arciniega-Ceballos et al., 1999, 2003, 2008](#); [Chouet et al., 2005](#)); and Cotopaxi, Ecuador ([Molina et al., 2008](#)).

Varieties of LP volcanic seismic signals have been observed at different volcanoes. Since the broadband seismometers are also sensitive to tilt motion, which can be analytically described (e.g., [Graizer, 2006](#); [Rodgers, 1968](#)), they have been also used to analyze the deformation due to movements of magma in shallow volcanic edifice (e.g., [Aoyama and Oshima, 2007](#); [Wielandt and Forbriger, 1999](#)).

Besides the advantages already mentioned, the LP seismic wavefield observed with a broadband seismic network may be used to directly monitor the activity of a volcano. [Dawson et al. \(2004\)](#) implemented the radial semblance method of [Kawakatsu et al. \(2000\)](#), originally called 'waveform semblance' for the near real-time monitoring of the activities at the shallow magmatic conduit in Kilauea Volcano, Hawaii ([Figure 18](#)). Similarly, the moment tensor and single-force source real-time monitoring should be possible using a grid-based method as suggested by [Kawakatsu \(1998\)](#) and implemented at the Earthquake Research Institute of the University of Tokyo for the real-time seismicity monitoring ([Tsuruoka et al., 2009](#); http://www.eri.u-tokyo.ac.jp/GRiD_MT/) and as realized at Stromboli Volcano, Italy ([Auger et al., 2006](#)).

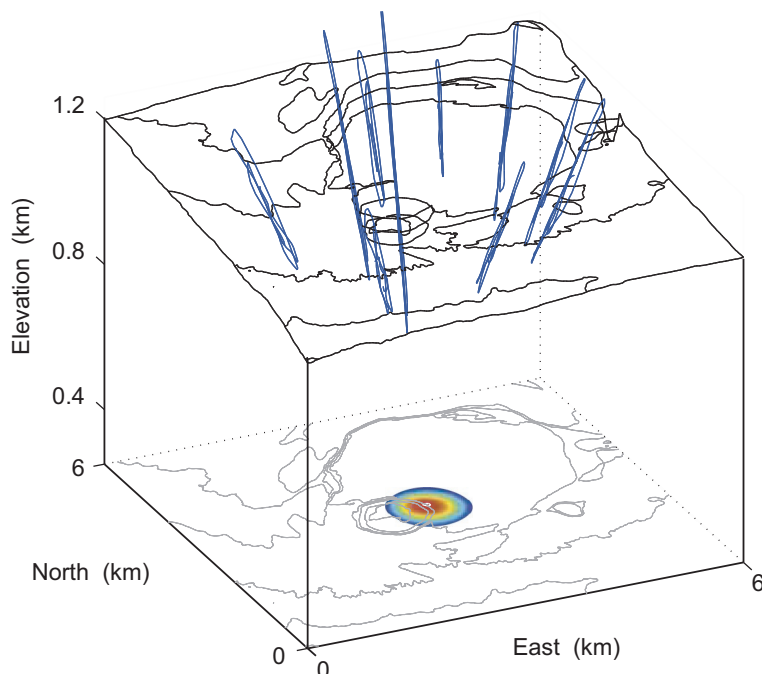


Figure 18 Particle motions recorded by the network during a 1 min long window of long-period signal. The area with bright colors with high semblance values indicates the source region. Adapted from Dawson P, Whilldin D, and Chouet B (2004) Application of near real-time radial semblance to locate the shallow magmatic conduit at Kilauea Volcano, Hawaii. *Geophysical Research Letters* 31: L21606.

Kawakatsu and Montagner (2008) recently clarified the connection between these grid-based source inversion approaches and the adjoint or time-reversal source imaging approach, by showing that the adjoint method is an approximation to the inversion problem.

4.15.5.2 Array Analysis

To locate the source of volcanic signals, various kinds of methods have been applied to the analyses of observed data. Classical hypocenter determination methods using phase arrivals have been applied to signals observed around active volcanoes and used to predict the eruptions (e.g., Scarpa and Tilling, 1996). More advanced methods such as relative hypocenter determination using the cross-spectral and the double difference have also been applied to volcanic earthquakes and have revealed vivid images of volcanic activities like the transportation of magma beneath active volcanoes (e.g., Hayashi and Morita, 2003; Morita et al., 2006).

Nevertheless, there still exist some specific issues to be explored in the analyses of volcanic signals due to complex source processes and complicated wave propagations in inhomogeneous structures. Lack of clear phases and relatively low dominant frequencies of, for example, LF events and volcanic tremors make the direct application of the conventional hypocenter determination means rather difficult. In the past decade, these difficulties have been partly overcome by the use of seismic arrays. In array observations, seismometers are placed at short distance intervals (a few tens to hundreds of meters), and thus, the wave properties of the incident waves can be examined by extracting coherent portion of the signals using the array data. Recent advance in portable instrumentation

enables us to observe volcanic signals close to their sources with dense arrays and sheds new light on the understanding of the processes through the precise determination of location and distribution of volcano seismic sources. For example, the multiple signal classification technique (Goldstein and Archuleta, 1991; Schmidt, 1986) has been applied to the analyses of the array data observed at Kilauea, Hawaii (e.g., Almendros et al., 2002a,b; Goldstein and Chouet, 1994; Figure 19), and at Stromboli, Italy (e.g., Chouet et al., 1997), and revealed magmatic and hydrothermal activities beneath the volcanoes. Other array analysis techniques have been also applied to constrain the source locations and properties of volcanic signals (e.g., Gordeev et al., 1990; Ibanez et al., 2000).

Seismic arrays have also been used to constrain the shallow structure of volcanoes that is indispensable for the quantitative volcano seismology. Here, the wavefield observed by a seismic array is treated as a stationary stochastic one in time and space, and the statistical correlation method (i.e., spatial autocorrelation (SPAC) method) originally proposed by Aki (1957) is applied to the data to determine the wave properties such as wave types and phase velocities (e.g., Chouet et al., 1998; De Luca et al., 1997; Ferrazzini et al., 1991; Metaxian and Lasage, 1997; Saccorotti et al., 2001a).

4.15.6 Models for Volcanic Seismic Signals

Although the basic governing physics, some of which have been discussed earlier, are common, and the methods of data acquisition and analyses employed may be the same, actual phenomena that we observed at volcanoes are different from one volcano to another. To appreciate the diversity, some of

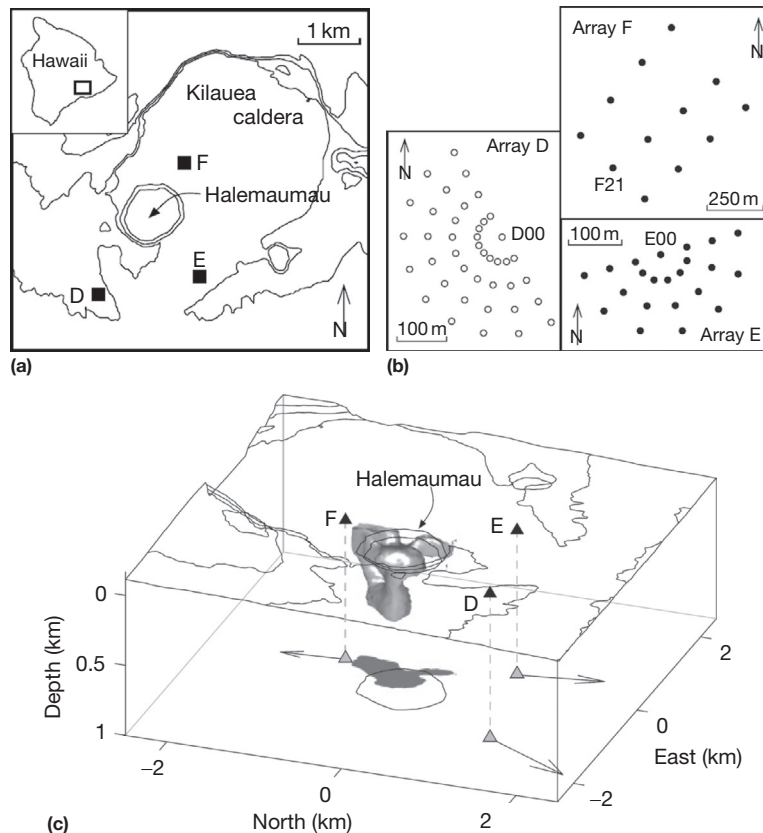


Figure 19 Example of array observation. (a and b) Configuration of the array observation at Kilauea, Hawaii. (c) Source location of the seismic signals obtained by array analyses. The arrows represent the apparent slowness vectors determined by frequency-slowness analyses, and the gray region is the estimated source region. Adapted from Almendros J, Chouet B, Dawson P, and Huber C (2002) Mapping the sources of the seismic wavefield at Kilauea Volcano, Hawaii, using data recorded on multiple seismic antennas. *Bulletin of Seismological Society of America* 92: 2333–2351.

the well-studied volcanoes are reviewed. Since the names of volcanic seismic events sometimes differ from one volcano to another due to the difference in conventions as described in ‘terminology,’ in this section, we keep common terminology locally used at each volcano and not necessarily follow the conventional classification of periods in earthquake seismology.

4.15.6.1 Aso

Aso Volcano is one of the most active volcanoes in Japan. It has erupted in Strombolian style repeatedly with intervals of 5–10 years, and recent activities take place at the youngest central cone, which is composed of seven craters aligned northwest–southeast direction with a length of 1 km. The current eruptive activities have occurred at the first crater located at the northernmost of the chain of the craters.

Among the volcanic tremors observed by Sassa (1935), the most remarkable one is the volcanic microtremor of the second kind for its extraordinary LP (7.5 s) and its highly repetitive occurrence. Observations using modern broadband seismometers revealed that the fundamental period of the long-period tremor (hereafter LPT) is 15 s (Kaneshima et al., 1996; Kawakatsu et al., 2000) and that the second-kind tremor of Sassa is the higher mode of LPTs. The characteristics of LPTs are summarized as follows: (1) they are continually emitted from

the volcano regardless of surface activity; (2) spectra show several common spectral peaks, which align with almost equal spacing (15, 7.5, 5, and 4 s); (3) decay is fairly fast and the duration is only a few cycles; and (4) they are often accompanied with short-period tremors.

Kaneshima et al. (1996) and Kawakatsu et al. (2000) deployed a broadband seismic network in 1994 and analyzed the waveforms of the LPTs and those associated with phreatic eruptions, which ejected composite of mud and water (Figure 6). They concluded that the sources of both phenomena are located a few hundred meters southwest of the first crater and at a depth of 1–1.5 km below ground surface. They further performed a source mechanism inversion of LPTs and obtained a volumetric source mechanism, which can be decomposed into isotropic and vertical tensile crack components (Legrand et al., 2000). The detail of the tensile crack component was further studied by Yamamoto et al. (1999) using the data obtained by a dense broadband observation. Analyzing the spatial distribution of the signal amplitudes of LPTs (Figure 20), they revealed the existence of a crack-like conduit whose strike and width are almost the same as those of the chain of craters; the crack extends nearly vertically from a depth of 300–400 m below the surface to a depth of about 2.5 km, as illustrated in Figure 21. This observation revealed that the chain of craters is simply a surface expression of a buried crack-like conduit.

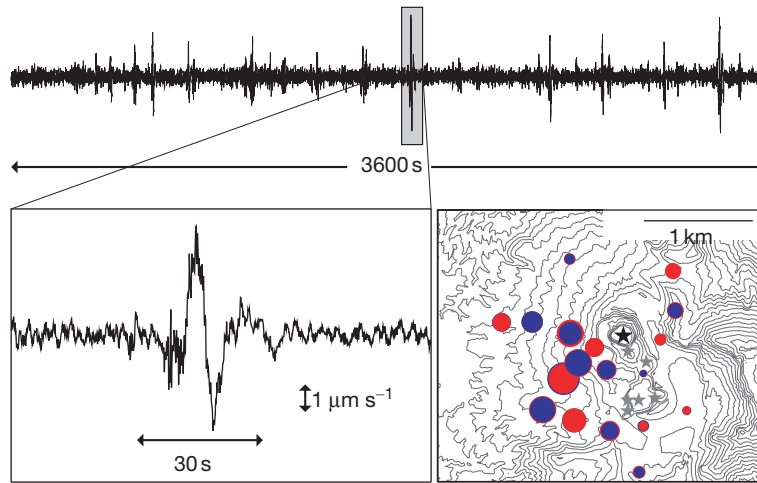


Figure 20 (a) A vertical-component broadband velocity seismogram observed at a distance of 1.5 km. Spike-like signals are LPTs. (b) Close-up view of an LPT. (c) Amplitude variation of LPTs (after Yamamoto, 2005).

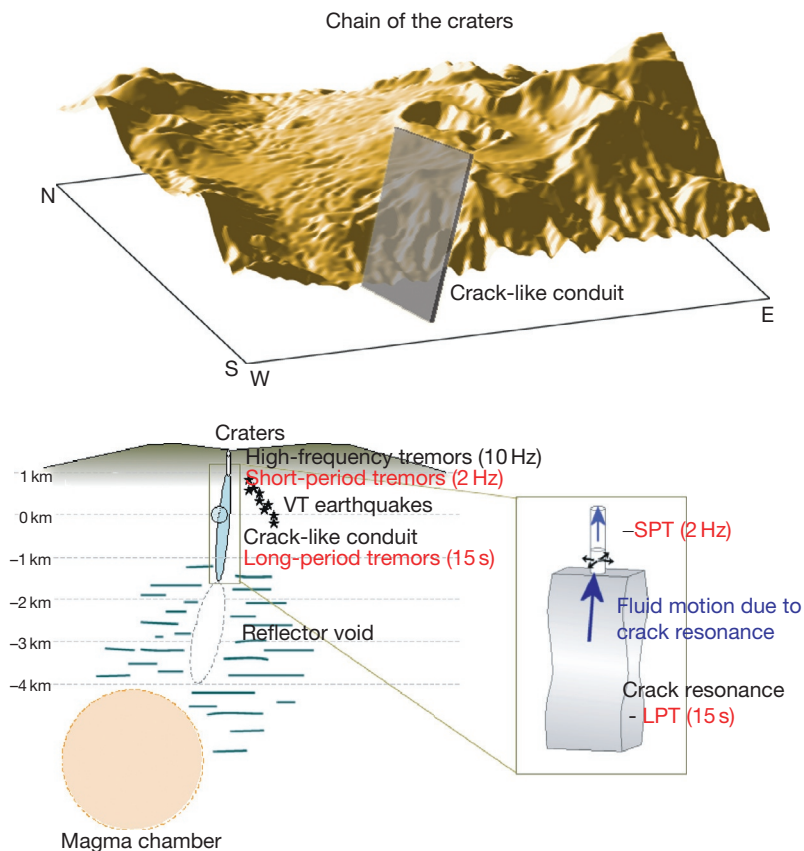


Figure 21 Image of crack-like conduit model. The crack is almost parallel to the chain of craters and the extension of the crack meets the active fumarole at the surface (after Yamamoto, 2005).

The physical properties of the crack-like conduit were studied by Yamamoto (2005), who applied a boundary integral method to the oscillation of a fluid-filled crack with slits at the edges, which allow the fluid to escape, as an extension of Chouet's closed crack model (Chouet, 1986). He demonstrated that the spectral characteristics of LPTs (mode frequencies, spacing, and attenuation) can be explained by the

oscillation of a 25 m-thick crack-like conduit filled with the gas-ash mixture. The existence of a crack-like conduit is also supported by other studies such as a reflection study by Tsutsui and Sudo (2004), and the crack-like conduit is considered as a subsurface path connecting a postulated magma chamber at a depth of around 5 km (Sudo and Kong, 2001) and the surface craters. The nature of other short-period seismic signals

with a period of about 0.5 s and a period of around 0.4–0.1 s has been also studied with modern digital data obtained by dense observations using short-period seismometers (Takagi et al., 2006; Yamamoto, 2005).

Figure 21 schematically summarizes the system beneath Aso Volcano that has been revealed by seismological analyses. Such a line of volcanic conduit systems connecting the magma chamber and the ground surface has not been detected at any other active volcano in the world. Considering that the various volcanic signals are manifestations of dynamic interactions between volcanic fluids and volcanic edifice in the conduit system, Aso Volcano appears to be one of the best fields to study the dynamic behavior of the volcanic fluid system beneath active volcanoes.

4.15.6.2 Kilauea, Hawaii

The Hawaiian volcanoes have been produced by the hot spot, which is presently beneath the Hawaii islands. The volcanoes primarily erupt basaltic magma, and the relatively low viscosity of the magma yields gentle slope of the edifices. Kilauea is the youngest and southeasternmost volcano on the Hawaii islands, and it is located at the splitting point between the Southwest Rift Zone and the East Rift Zone. Owing to the high level of activities and easy access to it, Kilauea may be the best-studied volcano in the world. For instance, a cylindrical cloud of seismicity beneath the volcano, which could plausibly reflect a magma movement in the crust, was detected (e.g., Klein and Koyanagi, 1989; Wolfe et al., 2003; Figure 22), and its relation to the mantle fault zone caused by plate flexure beneath Hawaii has been studied (e.g., Pritchard et al., 2007).

The magma transport system at shallower depths has been also studied using broadband seismic networks and short-period seismic arrays. Performing the moment tensor inversions of broadband data, Ohminato et al. (1998) revealed that the source of the VLP (5–10 s) signals associated with a magmatic activity in 1996 is a crack-like magma pathway at a depth of 1 km, which acts like a buffer for surging magma from beneath the summit of Kilauea to the East Rift Zone. Their analyses suggest that $\sim 1000\text{--}4000\text{ m}^3$ of magma is injected into the crack during a slow (1–3 min) accumulation phase, and it is ejected from the crack toward the East Rift Zone during a rapid (5–10 s) deflation phase, which together produce a sawtooth-like displacement signal shown in Figure 23. Dawson et al. (1998) also analyzed VLP signals observed in 1997 and concluded that the source location is almost the same as that determined by Ohminato et al. (1998), although the source mechanism is different.

The source of LP (1–2 s) signals, on the other hand, is determined using seismic array data (e.g., Almendros et al., 2001; Saccorotti et al., 2001b), and the hypocenter of LP signals is located directly above the source of VLP signals. Kumagai et al. (2005) analyzed the source mechanism of LP signals and imaged an expansion and contraction of a nearly horizontal crack through the moment tensor inversions. Based on the fluid-filled crack model developed by Chouet (1986), they further suggest that LP signals represent a resonance of a hydrothermal crack containing bubbly water and/or steam and that the heat from the underlying magma conduit (VLP source) may cause the pressurization of hydrothermal fluid in the crack

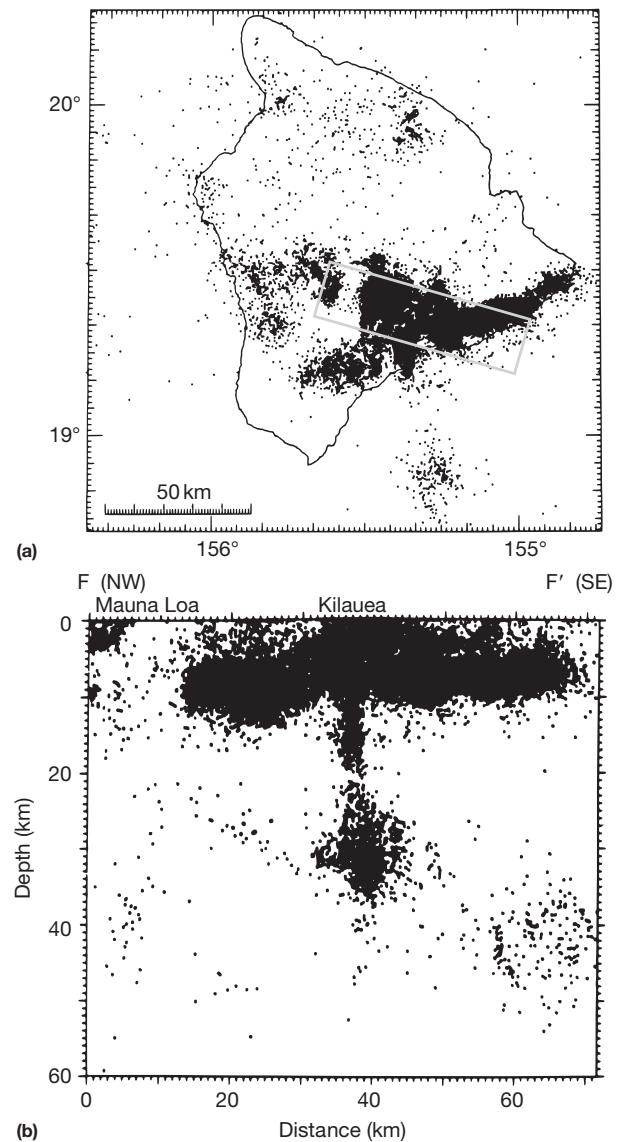


Figure 22 Seismicity of Hawaii. (a) Hypocenters of shallow earthquakes during the period 1970–84. (b) Cross section through Kilauea and Mauna Loa for earthquakes during 1970–83. The area of the cross section is that delimited by the box in the top panel. Adapted from Klein FW and Koyanagi RY (1989) The seismicity and tectonics of Hawaii. In: *The Geology of North America, Vol. N: The Eastern Pacific Ocean and Hawaii*, pp. 238–252. Boulder, CO: Geological Society of America.

and trigger these LP signals. Other geophysical observations, such as a positive correlation between summit SO_2 emissions and the shallow seismicity, which was pointed out by Sutton et al. (2001), also suggest the interaction between magmatic and hydrothermal systems at the shallow part of the volcano.

These results indicate that the detailed quantitative analyses of volcanic seismic signals are critically important for a better understanding of volcanic processes and such analyses may be crucial steps toward eruption prediction and the assessment of volcanic hazards.

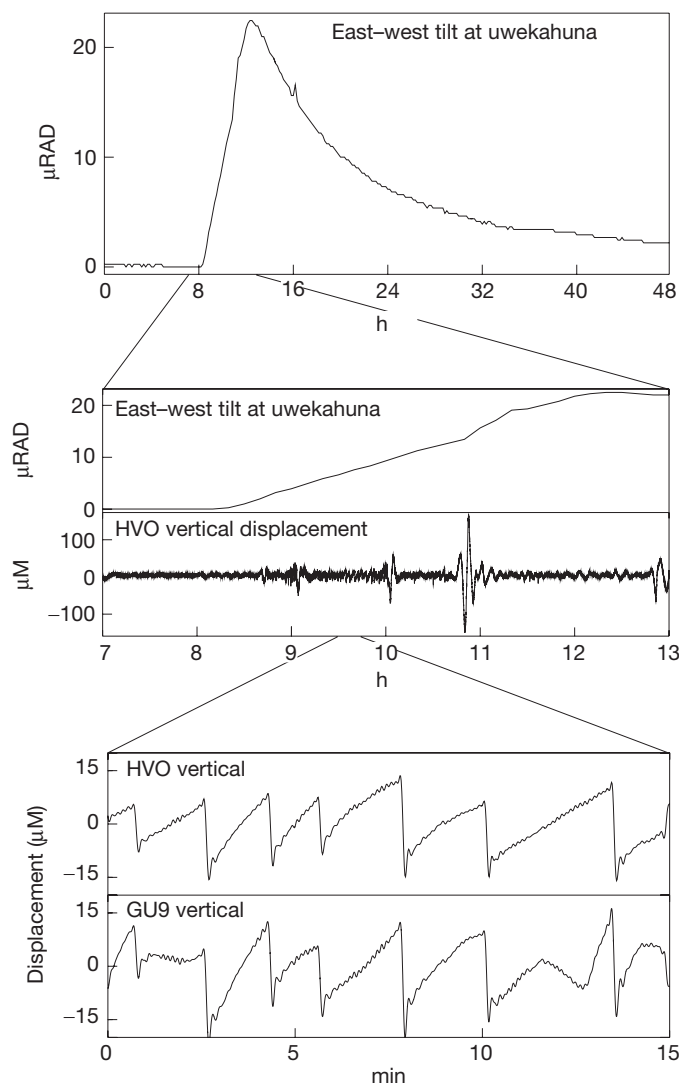


Figure 23 Tilt and broadband record observed around the summit caldera of Kilauea. Adapted from Ohminato T, Chouet B, Dawson P, and Kedar S (1998) Waveform inversion of very long period impulsive signals associated with magmatic injection beneath Kilauea Volcano, Hawaii. *Journal of Geophysical Research* 103: 23839–23862 (after Yamamoto, 2005).

4.15.6.3 Miyake

Miyake-jima is a basaltic stratovolcanic island in the Izu-Bonin arc, Japan, and has erupted quasiperiodically with an approximate interval of 20 years; the recent activities took place in 1940, 1962, and 1983. The eruption in 2000 was expected in this sense; however, what actually happened was totally unexpected: A new caldera was formed for the first time in the past 2500 years (Figure 24). This caldera-forming process was recorded with contemporary geophysical equipments.

Figure 25 shows the seismicity observed during the 2000 Miyake-jima activity (Sakai et al., 2001). The earthquake swarm started right beneath the volcano and migrated to the southwest (Uhira et al. (2005), not well-resolved in Figure 25), which may correspond to the subsurface migration of magma beneath the volcano. The swarm activity next shifted to northwest of the island, indicating a major subsurface magma migration from the volcano to the outside dike system. After a small

summit eruption, the caldera formation took place, and a number of LF earthquakes are observed beneath the summit area. It appears that sucking of magma from the magma storage system beneath the volcano to the northwest dike system and the resulting magma vacancy in the volcano are the direct causes of the caldera formation (Furuya et al., 2003).

During roughly 40 days of the caldera formation, a number of peculiar bell-shaped LP (50 s) seismic (velocity) signals, as well as steps in tiltmeters, are observed once or twice a day (Figure 26). Based on the moment tensor inversion of the near-field and far-field broadband seismic records, Kumagai et al. (2001) associated these LP velocity pulses with inflations of a nearly vertical crack-like magma chamber due to its pressure increase as response to the repetitive stick-slip-type falling of a piston-like mass within the conduit that results in the caldera formation at the summit. Analyzing the signals of the same sequence of events recorded by the tiltmeters and

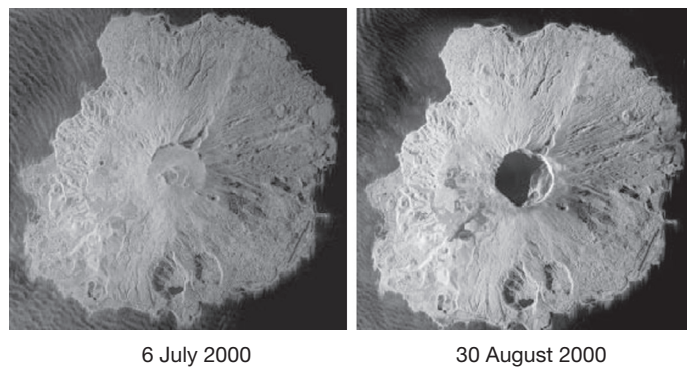


Figure 24 Airborne SAR images of Miyake-jima Volcano before and after the 2000 caldera formation. Adapted from Kumagai H, Ohminato T, Nakano M, et al. (2001) Very-long-period seismic signals and caldera formation at Miyake Island, Japan. *Science* 293: 687–690.

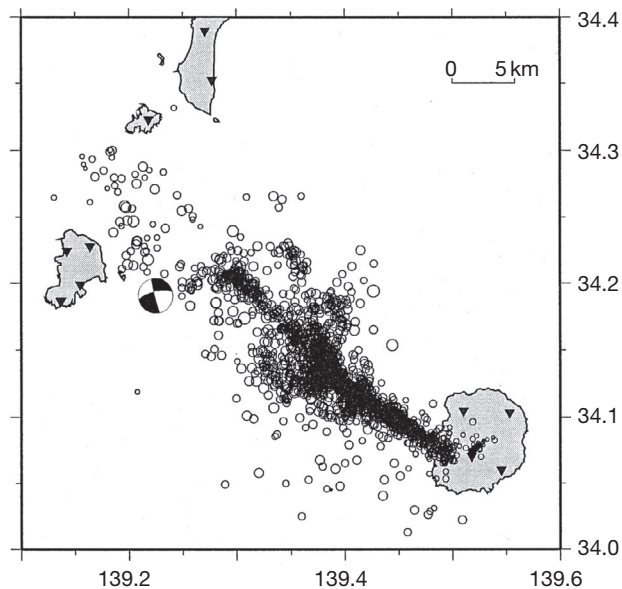


Figure 25 The seismicity of the 2000 Miyake-jima activity. Adapted from Sakai S, Yamada T, Ide S, et al. (2001) Magma migration from the point of view of seismic activity in the volcanism of Miyake-jima island in 2000. *Journal of Geography* 110: 145–155.

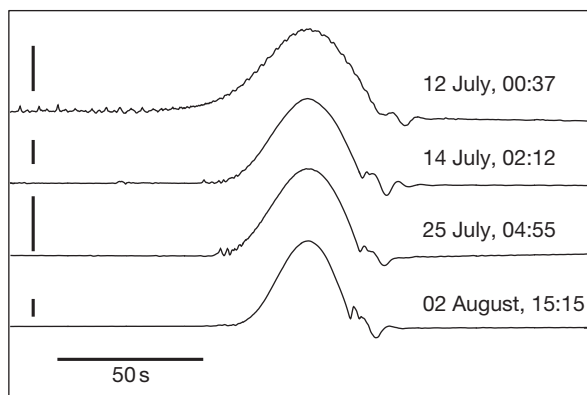


Figure 26 Vertical-component near-field velocity waveforms observed in the Miyake-jima island. Adapted from Kumagai H, Ohminato T, Nakano M, et al. (2001) Very-long-period seismic signals and caldera formation at Miyake Island, Japan. *Science* 293: 687–690.

broadband seismometers deployed in the island, Fujita et al. (2004) arrived at a different conclusion. They suggested that these events are caused by a cyclic expansion of a subsurface sill-like magma plumbing system and introduced a model based on a two-phase flow instability, called a pressure drop oscillation as the mechanism for the repeated activity. Either proposed model seems to explain a part of the observations, but not all. Although the available data may not be able to eventually resolve the difference, nevertheless, the 2000 Miyake-jima activity has provided us a rare opportunity to observe a caldera-forming event with contemporary geophysical instruments.

4.15.6.4 Stromboli

Stromboli is one of the Aeolian islands of Italy and has been almost continuously in eruption for more than at least 2000

years. In contrast to the widely used term 'Strombolian eruption' that is undiscerningly used to describe a variety of volcanic eruptions, the activity of Stromboli is characterized by short-term explosive bursts of lava ejected into the air. Since they eject relatively viscous basaltic lava, building up of the gas pressure is required to fragment the magma and results in intermittent episodic explosions.

From the seismic point of view, such activities generate a considerable number of explosion quakes, persistent volcanic tremors, and LP signals. One of the most characteristic features of seismic activities at Stromboli is the extremely wide frequency contents of these signals (Figure 27). Spindle-shaped signals superimposed on background continuous tremor are generated by summit eruptions, and the nature of these short-period signals has been extensively studied by many researchers (e.g., Del Pezzo, 1992; Del Pezzo et al., 1974; Ripepe and Gordeev, 1999; Saccorotti and Pezzo, 2000). These studies shed new light on the seismic activity of the volcano, which is mainly characterized by a very shallow seismicity. Such very shallow origin of the seismic signals is also confirmed by the analyses of seismic array observations (e.g., Chouet et al., 1997; Saccorotti and Pezzo, 2000), and the sources of explosion quakes and tremors are located at depths shallower than 200 m beneath the summit crater.

Recent observations using broadband seismometers reveal the existence of VLP signals with a period of about 10 s, which are accompanied with explosions, and give additional constraints on the understanding of Strombolian activities (e.g., Chouet et al., 2003, 2008; Neuberg et al., 1994). Analyses of broadband seismograms using the independent component analysis method indicate that further longer-period signals

(30–40 s) are generated at almost the same region as the VLP signals (De Martino et al., 2005).

Based on various field observations and laboratory experiments, Ripepe and Gordeev (1999) and Ripepe et al. (2001) arrived at a model of Strombolian explosions that is reminiscent of the analog experimental model by Jaupart and Vergnolle (1989), in which dissolved gas bubbles are trapped at the roof of a magma reservoir as a foam layer and the periodical collapse of the layer causes ascent of gas slugs through a vertical conduit connecting the reservoir and the surface vent. The growth, flow, and burst of the coalescent gas bubbles observed in their laboratory experiments well capture the characteristics of sequential occurrence of VLP/short-period seismic signals, acoustic signals, and thermal/light emissions observed at Stromboli. Figure 28 shows such a direct link between gas flux, magma volume flux, and seismicity, where the balance between gas/magma flux and gas overpressurization appears to determine the type of activities (i.e., effusive or explosive) (Ripepe et al., 2005). The common trend in the VLP event rate and the SO_2 emission rate during the effusive phase suggest that the rate of gas flux controls the frequency of foam coalescence and the development of gas slugs, which generate VLP signals; the decrease in SO_2 and VLP rates coincides with the increase in the explosion rate at the summit measured by the count of thermal transients and in the amplitude of acoustic pressure; this may be understood in terms of the reduction of gas/magma supplies that results in sealing of eruptive fractures and the upward migration of magma.

Performing detailed analyses of the broadband network data, Chouet et al. (2003) concluded that the source mechanism of VLP events associated with Strombolian explosions

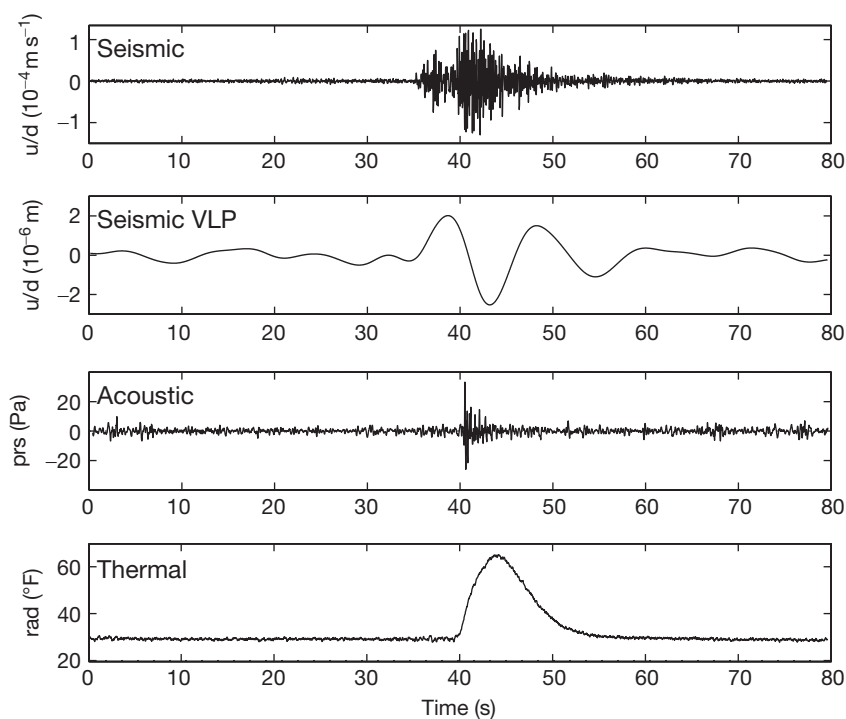


Figure 27 Example of the synchronous occurrence of seismic, acoustic, and thermal signals observed in Stromboli. From Ripepe et al., unpublished data.

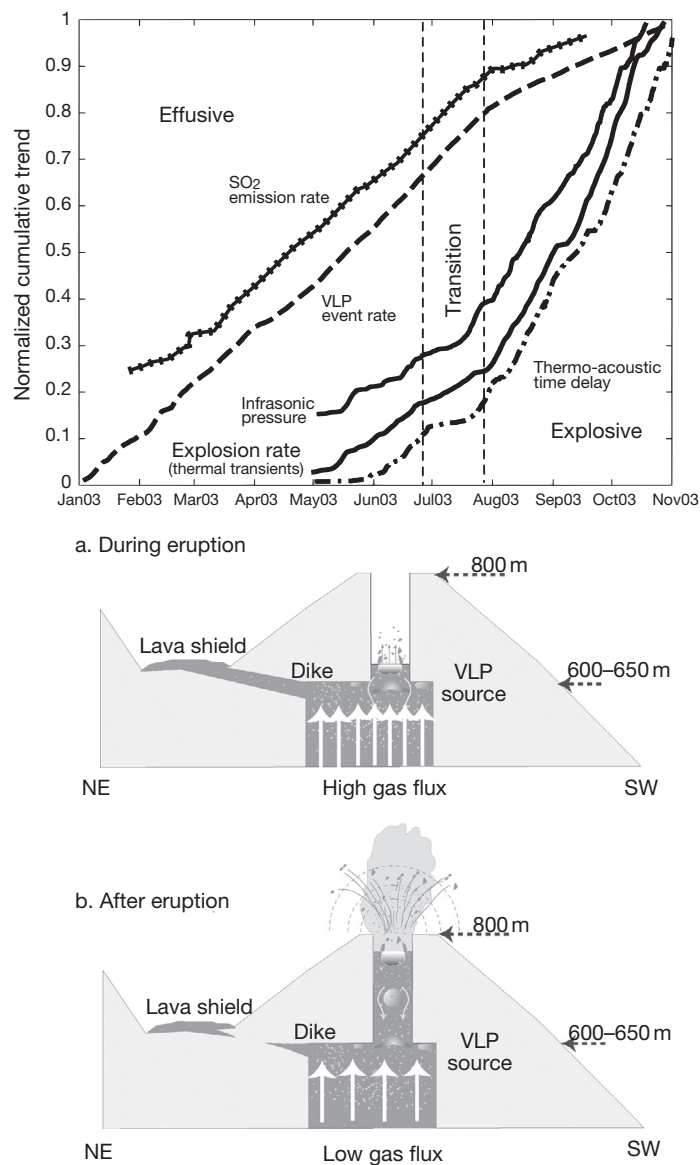


Figure 28 Example of the relation between gas flux rate and seismic/acoustic activities. During the lava effusion phase (model A), few explosions at the summit were observed while the VLP seismicity well correlated with the SO_2 emission rate. On the other hand, in the explosive phase (model B), magma rose up in the conduit due to the sealing of eruptive fractures, resulting in the gas overpressurization and the explosive activity at the summit craters. Adapted from Ripepe M, Marchetti E, Ulivieri G, et al. (2005) Effusive to explosive transition during the 2003 eruption of Stromboli volcano. *Geology* 33: 341–344.

consists of both moment tensor and single-force components, which, respectively, correspond to the opening/closing of an inclined crack just beneath the active vents and the magma flow in the conduit caused by the piston-like rise of a gas slug. To investigate the behavior of coalescence and ascent of gas slugs in such an inclined conduit, James et al. (2004) carried out laboratory experiments of the two-phase flow in vertical and inclined pipes and demonstrated that the inclination of the conduit may play an important role in controlling the flow types and size/velocity distributions of gas slugs in the fluid-filled conduit.

In spite of the difference between models in details, these results suggest that understanding of the bubble dynamics may

be essential to elucidate the source process of the seismic signals and the dynamics of Strombolian activities.

4.15.7 Other Volcano-Specific Issues

4.15.7.1 Explosion Quakes

Explosion quakes are observed when explosive eruptions occur (e.g., Figures 3 and 4). An explosive volcanic eruption (Vulcanian type) can be macroscopically understood as the deflation process of a reservoir system beneath a volcano, which may be observed by geodetic (e.g., Ishihara, 1990) or broadband seismic (e.g., Figure 6) means. The source

mechanism of the accompanied explosion quake may be represented by a combination of a single force and an implosive moment tensor (Figure 11). Since the pioneering work of Kanamori et al. (1984), there are now a number of observations to quantify the force mechanism and size of volcanic explosions (e.g., Chouet et al., 2005; Nishimura et al., 1995; Ohminato et al., 2006).

Explosion quakes are extensively studied at Sakurajima Volcano, Japan; comparing seismic and video records, Ishihara (1985) demonstrated that explosion quakes occur at a depth of 1–2 km beneath the active crater about a few seconds prior to the explosive eruptions at the summit, which generate air shock waves. Uhira and Takeo (1994) showed that an explosive and implosive cylindrical moment tensor component at the source depth is the dominating source mechanism for the explosion quakes. By carefully analyzing borehole seismic network records, Tameguri et al. (2002) were able to decompose explosion quakes into two processes: The first one is an isotropic expansion followed by a contraction of a cylinder at a depth of 2 km, and the second is an isotropic expansion and subsequent horizontal contraction at depths of 0.25–0.5 km beneath the crater bottom, which occur about 1 s after the onset of the first and coincide with the generation of the shock wave. The plausibility of this second process was numerically demonstrated by Nishimura and Chouet (2003), who considered the behavior of a reservoir–conduit system embedded in a homogeneous crust, which is filled with a compressive fluid; it was shown that an abrupt pressure increase occurs just beneath the vent, which can generate pulse-like Rayleigh waves

as observed by Tameguri et al. (2002), and that the strength of the lid played an important role in defining eruption types.

4.15.7.1.1 Scaling law of volcanic explosion

Nishimura (1998) showed that the far-field displacement due to an explosion is proportional to the area (square of radius) of the crater vent and obtained a scaling law of volcanic explosion quakes, as similar to that of earthquakes. Figure 29 compares the vent radii and estimated seismic magnitudes of the largest event for studied volcanoes by various investigators; it shows that the magnitude is essentially proportional to the cross-sectional area of the vent for vent radii ranging from 10 to 600 m, as predicted by the theory, and that the initial excess pressure in the reservoir is of the order of a few megapascals, a similar value as the stress drops of earthquakes. This suggests that the size of explosion quakes may be estimated by knowing the vent radius, although it does not necessarily imply that it is possible to predict the size of future eruptions, as it is not guaranteed that the size of present vent is the same for the future ones.

4.15.7.2 Triggered Seismicity

It has been well known among local observers that volcanoes are 'sensitive,' as seismicity there often shows changes apparently associated with external disturbances (tides or large earthquakes). For example, Sassa (1936) documented diurnal and semidiurnal periodicities in the occurrence of volcanic seismicity of Aso Volcano. McNutt and Beavan (1981) also suggested a tidal correlation of seismicity at Pavlof Volcano (cf., Emter, 1997).

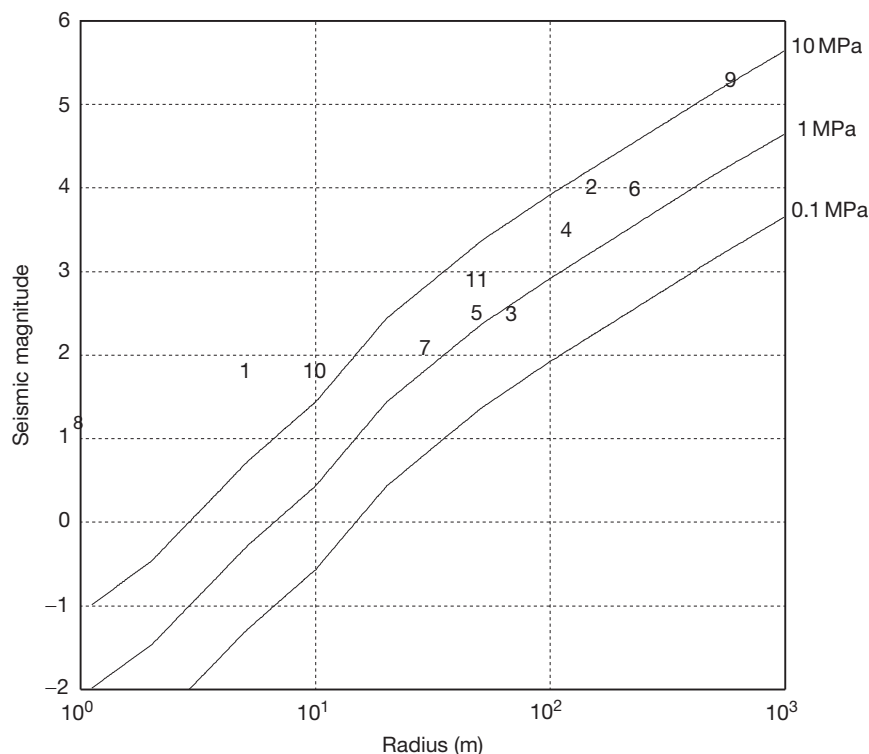


Figure 29 Relation between the seismic magnitude of eruption quakes and the observed vent radius. Numbers from 1 to 11 indicate representative points of different volcanoes. Lines show theoretical relation for assumed initial excess pressures. Adapted from Nishimura T (1998) Source mechanisms of volcanic explosion earthquakes: Single force and implosive sources. *Journal of Volcanology and Geothermal Research* 86: 97–106.

There are also convincing lines of evidence that the passage of large amplitude seismic waves dynamically triggers seismicity, especially at volcanoes. Hill et al. (1993) documented the seismicity increase in Long Valley Caldera, USA, following the $M_w=7.3$ 1992 Landers earthquake, California (also Gomberg et al., 2001; Prejean et al., 2004). Brodsky and Prejean (2005) made a systematic survey of remotely triggered seismicity at Long Valley; showing the evidence for a frequency-dependent triggering threshold, they concluded that LP (>30 s) waves are more effective at generating local seismicity than short-period waves of comparable magnitude; among the several proposed mechanisms for such remote dynamic triggering, they suggested that the mechanism that invokes fluid flow through a porous medium can produce such a frequency dependence of triggering threshold (~ 5 kPa). This mechanism also seems to explain why volcanoes are sensitive to external disturbances, as unclogging fractures to drive fluid flow (Brodsky et al., 2003) may be easily initiated in volcanic/geothermal areas where the constant precipitation from volcanic fluids generate fragile blockages in fractures and faults. West et al. (2005) observed periodically (20–30 s) triggered seismicity at Mt. Wrangell, Alaska, when LP Rayleigh waves generated by the $M_w=9.0$ 2004 Sumatra earthquake swept across Alaska, and concluded that the extensional stresses of ~ 25 kPa due to the Rayleigh waves triggered shear failure on normal faults, which may be also influenced by the aforementioned fluid pumping mechanism. These observations suggest that the presence of ample fluid in volcanic systems makes them ideal experimental field for elucidating the earthquake-triggering mechanism.

4.15.7.3 Seismic Wave Scattering in Volcanoes

Dynamic interactions between volcanic fluids and volcanic edifice along with more large-scale tectonic activities make the volcanic structure very complex, and volcano is considered to be one of the most heterogeneous parts of the Earth's crust. The understanding of inhomogeneities in volcanoes may thus provide us important information on the various volcanic processes as well as the volcanic structure. One way to infer the distribution and movement of volcanic fluids and resultant heterogeneities beneath volcanoes is the use of tomography for seismic velocity and attenuation, which basically relies on the coherent waves (mainly direct waves). The tomographic method, however, has some inherent limitations in its resolution partly due to the wavelength of the waves used in the analyses, and it is rather difficult to image small-scale heterogeneities like intruded dykes and fractures using the method.

An alternative approach to image small-scale heterogeneities is the use of scattered waves. From this point of view, the scattering properties and their spatial variation around active volcanoes have been studied based on various kinds of wave scattering model, and it has been revealed that small-scale heterogeneity beneath volcanoes is one or two orders stronger than that of ordinal crust (e.g., Sato et al., 2012). Such strong heterogeneity causes the multiple scattering of seismic waves and mode conversion between P- and S-waves, and thus, the waveforms of volcanic earthquakes often exhibit spindle-like envelopes having small direct waves and long-lasting coda waves. While these complex waveforms, especially those missing clear coherent phases, seem not

suitable for conventional analyses, recent progresses in theoretical and observational researches on seismic interferometry (e.g., Schuster et al., 2004) reveal that multiply scattered waves instead have a significant advantage that the slight change in the elastic properties of medium, which have little influence on the first arrivals, are amplified and thus readily observed. In addition, the existence of dominant mode conversion due to strong heterogeneity in volcanic environment (e.g., Yamamoto and Sato, 2010) makes the situation more ideal for the application of the seismic interferometry. The method has been effectively applied to monitor the slight medium changes due to magmatic activities at active volcanoes (e.g., Brenguier et al., 2008; Nagaoka et al., 2010), and it may be one of the most promising techniques to monitor the movement of volcanic fluids in volcanic systems and predict volcanic eruptions. Future researches on the elementary physics of seismic wave scattering including the resonant scattering of small fluid inclusions (e.g., Margerin, 2013) may also contribute to better understanding of behavior of volcanic fluids.

4.15.8 Concluding Remarks

One of the ultimate goals of volcano seismology is to extract quantitative information about the volcanic fluid transport systems beneath active volcanoes and to utilize such information for forecasting short- and long-term eruptive behaviors. As described in this chapter, macroscopic behaviors of dynamic interactions between volcanic fluids and volcanic edifices have been extensively studied in volcano seismology, and significant progress has been made; the observational/numerical/experimental studies have established firm foundations for explaining the macroscopic processes responsible for generating observed volcanic seismic signals and other related volcanic activities. Considering the rapid progress in technologies for the field observations and numerical simulations, the future of these lines of studies seems to be promising; the accumulation of high-quality digital data at different volcanoes around the world will further bring significant advances in volcano seismology, as well as volcanology in general; the increase in computer power remarkably extends our capability to simulate complicated dynamic interactions between phases and enables us to identify and separate distinct signatures of volcanic seismic signals; the recent advance in high-sampling, high-precision GPS and InSAR observations may also help our understanding from the observational aspect.

Nevertheless, our understanding of the elementary physics of multiphase systems, such as the dynamics of bubbles in magmatic fluids and their flows, is still immature, although such understanding fused with the established studies is indispensable for the elucidation of the whole volcanic fluid systems. For instance, recently, James et al. (2006) experimentally demonstrated that the geometry of the conduit is the key parameter governing the flow pattern of ascending slugs in viscous magma, which act as a trigger/source of volcanic seismic signals. Such a parameter of the conduit system may be retrieved by acoustic measurements as demonstrated by Garcés et al. (2000), and the integration of these studies will provide a great feedback to volcano seismology. Time-varying

moment-tensor and single-force components estimated from high-quality seismic data may also be an important clue to image nonlinear fluid dynamics in volcanic conduits. Such studies on the modeling of physical processes, in cooperation with multiparameter observations and numerical simulations, may bring a new boost to volcano seismology in the near future. Volcano seismology, thus, remains to be one of the most challenging fields in Earth science as an integrated science of multidisciplinary researches on multiscale/multiphase systems.

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