Bulletin of Volcanology (2022) 84: 86
https://doi.org/10.1007/s00445-022-01586-0

REVIEW ARTICLE

One hundred years of advances in volcano seismology and acoustics

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Received: 30 September 2021 / Accepted: 4 July 2022 / Published online: 20 August 2022

Abstract

Since the 1919 foundation of the International Association of Volcanology and Chemistry of the Earth's Interior (IAVCEI), the fields of volcano seismology and acoustics have seen dramatic advances in instrumentation and techniques, and have undergone paradigm shifts in the understanding of volcanic seismo-acoustic source processes and internal volcanic structure. Some early twentieth-century volcanological studies gave equal emphasis to barograph (infrasound and acoustic-gravity wave) and seismograph observations, but volcano seismology rapidly outpaced volcano acoustics and became the standard geophysical volcano-monitoring tool. Permanent seismic networks were established on volcanoes (for example) in Japan, the Philippines, Russia, and Hawai'i by the 1950s, and in Alaska by the 1970s. Large eruptions with societal consequences generally catalyzed the implementation of new seismic instrumentation and led to operationalization of research methodologies. Seismic data now form the backbone of most local ground-based volcano monitoring networks worldwide and play a critical role in understanding how volcanoes work. The computer revolution enabled increasingly sophisticated data processing and source modeling, and facilitated the transition to continuous digital waveform recording by about the 1990s. In the 1970s and 1980s, quantitative models emerged for long-period (LP) event and tremor sources in fluid-driven cracks and conduits. Beginning in the 1970s, early models for volcano-tectonic (VT) earthquake swarms invoking crack tip stresses expanded to involve stress transfer into the wall rocks of pressurized dikes. The first deployments of broadband seismic instrumentation and infrasound sensors on volcanoes in the 1990s led to discoveries of new signals and phenomena. Rapid advances in infrasound technology; signal processing, analysis, and inversion; and atmospheric propagation modeling have now established the role of regional (15–250 km) and remote (> 250 km) ground-based acoustic systems in volcano monitoring. Long-term records of volcano-seismic unrest through full eruptive cycles are providing insight into magma transport and eruption processes and increasingly sophisticated forecasts. Laboratory and numerical experiments are elucidating seismo-acoustic source processes in volcanic fluid systems, and are observationally constrained by increasingly dense geophysical field deployments taking advantage of low-power, compact broadband, and nodal technologies. In recent years, the fields of volcano geodesy, seismology, and acoustics (both atmospheric infrasound and ocean hydroacoustics) are increasingly merging. Despite vast progress over the past century, major questions remain regarding source processes, patterns of volcano-seismic unrest, internal volcanic structure, and the relationship between seismic unrest and volcanic processes.

Keywords Volcano seismology · Volcano acoustics · Infrasound · History · Geophysical instrumentation · Volcano monitoring · Eruption forecasting

Editorial responsibility: R.Sulpizio

This paper constitutes part of a topical collection:

IAVCEI 1919-2019: One hundred years of international outreach and scientific advances in Volcanology

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State of the art and introduction

Seismic and acoustic (collectively seismo-acoustic) geophysical technologies are complementary in volcano science and monitoring. Volcano seismology involves the analysis, interpretation, and modeling of seismic signals generated inside and around active volcanoes, as well as the application of seismic techniques to image internal volcanic structure (e.g., Aki 1992; Chouet 1996a, 1996b, 2003; McNutt 1992, 1996, 2005; Kumagai 2009; Lees 2007; Neuberg 2011;

Wassermann 2012; Chouet and Matoza 2013; Thompson 2015; Kawakatsu and Yamamoto 2015; McNutt and Roman 2015; Nishimura and Iguchi 2011; Zobin 2016; Saccorotti and Lokmer 2021). Volcanic seismicity occurs from mantle depths to the surface, and elucidates magmatic, hydrothermal, and faulting processes occurring within and around volcanoes (e.g., McNutt 1996; Nishimura and Iguchi 2011; Chouet and Matoza 2013; Kawakatsu and Yamamoto 2015; Matoza 2020). Infrasound (atmospheric acoustic waves with frequencies ~ 0.01–20 Hz) is produced by shallow subsurface and subaerial processes, including explosive eruptions, shallow degassing, surface flow, and mass wasting (Johnson and Ripepe 2011; Fee and Matoza 2013; Allstadt et al. 2018; Matoza et al. 2019a). Infrasound from major explosive eruptions can propagate thousands of kilometers in atmospheric waveguides, enabling regional (15–250 km) and remote (> 250 km) ground-based detection and characterization of explosive eruptions (e.g., Wilson and Forbes 1969; Kamo et al. 1994; Liszka and Garces 2002; Evers and Haak 2005; Le Pichon et al. 2005; Campus and Christie 2010; Fee et al. 2010a; Matoza et al. 2011a, 2018; McKee et al. 2021; Perttu et al. 2020a). Seismo-acoustic wave conversion and coupling commonly occur (e.g., Ichihara et al. 2012; Matoza and Fee 2014; Fee et al. 2016); thus, collocated seismic and infrasonic sensor deployments reduce ambiguity in seismic-acoustic signal type identification and process discrimination (e.g., Iguchi and Ishihara 1990; Garcés et al. 1998; Ripepe et al. 2001; Lees et al. 2004; Johnson et al. 2005; Matoza et al. 2009a, b, 2019b; Ichihara et al. 2021) and in explosive eruption detection and localization (e.g., Matoza et al. 2007, 2017; Sanderson et al. 2020; Le Pichon et al. 2021). At present, seismic and infrasound networks have become indispensable components in tracking the geophysical signatures of unrest and eruption, enabling better monitoring and mitigation of volcanic hazards (e.g., Moran et al. 2008a; National Academies of Sciences, Engineering, and Medicine 2017; Alvarado et al. 2018; Power et al. 2020). In the marine environment, technological advances and increasing availability of hydroacoustic systems and ocean-bottom seismology are expanding volcano seismology and acoustics to partially submerged and submarine oceanic volcanoes (e.g., Talandier and Okal 1987; Yamasato et al. 1993; Caplan-Auerbach and Duennebier 2001; Dziak et al. 2005, 2011; Chadwick et al. 2008, 2012; Green et al. 2013; Metz et al. 2016; Caplan-Auerbach et al. 2017; Metz and Grevemeyer 2018; Tepp et al. 2019, 2020; Fee et al. 2020; Talandier et al. 2020; Tepp and Dziak 2021; Rose and Matoza 2021).

In modern volcano seismology, quantitative source mechanism models based on full-waveform moment-tensor and single-force representations provide detailed source-time histories (e.g., Ohminato et al. 1998a, b; Nakano et al. 2003; Chouet and Matoza 2013; Kawakatsu and Yamamoto 2015). Interpretations of these observations are facilitated by laboratory and numerical experiments investigating a range of seismic source processes in volcanic fluid and solid frictional systems (e.g., Lane and James 2009; James et al. 2004; Lavallée et al. 2008; Arciniega-Ceballos et al. 2015; Spina et al. 2018). Further hypothesis testing is enabled through multi-parametric geophysical and geological field observations (e.g., Tuffen and Dingwell 2005; Pallister et al. 2012; Rasmussen et al. 2018; Unwin et al. 2021). The ability to accurately recover seismic source mechanisms depends on seismic station density and distribution along with known resolution of the internal seismic velocity structure of the volcanic edifice and upper crust (e.g., Bean et al. 2008; De Barros et al. 2011; Dawson et al. 2011; Chouet and Dawson 2016), which are all steadily improving with advances in (for example) portable compact broadband (e.g., Aster et al. 2005; Ibáñez et al. 2016; Lyons et al. 2016; Matoza et al. 2022a) and nodal (e.g., Kiser et al. 2016; Wu et al. 2017; Glasgow et al. 2018) seismic instrumentation, exploited by various tomographic implementations including ambient noise seismology (e.g., Obermann et al. 2016; Wang et al. 2017; Ulberg et al. 2020). Advances in broadband seismic and complementary geodetic instrumentation (e.g., tiltmeters, high-rate Global Navigation Satellite System (GNSS) receivers, and continuous gravity meters) and techniques are expanding the scope of volcano seismology to an increasingly wider bandwidth, including, at longer time-scales, ultra-long-period (ULP, > 100 s period) signals approaching static ground deformation (e.g., Green et al. 2006; Green and Neuberg 2006; Sturkell et al. 2008; Mattia et al. 2008; Maeda et al. 2011, 2017; Chouet and Dawson 2015; van Driel et al. 2015; Wauthier et al. 2013, 2016; Poland and Carbone 2018; Poland et al. 2019; Alvizuri et al. 2021; Soubestre et al. 2021; Bell et al. 2021). The boundary between volcano geodesy and volcano seismology is thus becoming seamless (e.g., Anderson et al. 2010; Segall 2013; Wauthier et al. 2016; Fernández et al. 2017; Segall and Anderson 2021; Neuberg et al. 2022). Rapid advances in computation are enabling more thorough processing and analyses of greater volumes of seismic waveform data and characterization of hundreds of thousands to millions of seismic events recorded during sustained episodes of volcanic unrest and eruption (e.g., Moran et al. 2008b; Rodgers et al. 2015a; Matoza et al. 2015, 2021). Machine learning methods were adopted relatively early in volcano seismology (e.g., Falsaperla et al. 1996; Langer et al. 2003; Scarpetta et al. 2005; Benítez et al. 2007; Ibáñez et al. 2009; Dawson et al. 2010, 2012) but are now in increasing use (e.g., Malfante et al. 2018; Carniel and Guzmán 2021; Dempsey et al. 2020; Shen and Shen 2021) and are poised for massive impact most immediately in event detection and association, classification, and forecasting.

In modern volcano acoustics, quantitative source mechanism models and source inversions have been developed for relatively simple volcano-acoustic sources such as impulsive explosions (Johnson et al. 2008a; Kim et al. 2012, 2015; Iezzi et al. 2019a) and rockfalls (Moran et al. 2008c). The acoustics of more complex sources such as sustained volcanic jet noise signals from sub-Plinian and Plinian eruptions (Matoza et al. 2009a; 2013a; Mckee et al. 2017) are being investigated by laboratory (Swanson et al. 2018; Fernández et al. 2020) and numerical (Cerminara et al. 2016; Brogi et al. 2018) experiments. Non-linearity in source and propagation is being examined in observations and by numerical simulation (Marchetti et al. 2013; Fee et al. 2013a; Maher et al. 2020, 2022; Watson et al. 2021). Acoustic full-waveform inversion methods take into account topographic effects, which are particularly significant at local ranges (< 15 km). Major advances in infrasound propagation theory and numerical implementations incorporating operational atmospheric specifications are enabling increasingly accurate models of regional range (15–250 km) and remote (> 250 km) infrasound propagation through atmospheric waveguides particularly in the troposphere, stratosphere, and thermosphere (e.g., Drob 2019; Waxler and Assink 2019; Schwaiger et al. 2019). In tandem, advances in infrasound technology and signal processing, discrimination, association, and location are improving abilities to detect signals from remote explosive eruptions within the plethora of interfering background ambient infrasound signals, which are sometimes termed clutter (e.g., Garces and Hetzer 2006; Matoza et al. 2013b; Ceranna et al. 2019), and wind noise (e.g., Hedlin and Raspet 2003; Walker and Hedlin 2010; Raspet et al. 2019) and localize these detections to remote volcanoes using sparse ground-based infrasound networks (e.g., Evers and Haak 2005; Arrowsmith et al. 2015; Matoza et al. 2017) or combined seismic and infrasonic networks (e.g., Fee et al. 2016; Matoza et al. 2018; Sanderson et al. 2020; Le Pichon et al. 2021). Infrasound early warning and eruption notification systems are in operation and undergoing testing and refinement (e.g., Garces et al. 2008; Fee et al. 2010b; De Angelis et al. 2012; Ripepe et al. 2018; Matoza et al. 2019a), augmenting spaceborne remote sensing methods for monitoring and quantifying global volcanism (e.g., Wright et al. 2004; Webley and Mastin 2009; Prata 2009; Ramsey and Harris 2013; Patrick and Smellie 2013; Poland 2015; Carn et al. 2017; Poland et al. 2020; Mckee et al. 2021). More broadly, volcano seismology and acoustics have seen progressive integration with a wide array of volcano-monitoring techniques (including, but not limited to) thermal, gas, electromagnetic, volcanic lightning, fumarole and hydrothermal, physical volcanological, and petrological methods utilizing ground-based and spaceborne instrumentation systems (e.g., Martini et al. 1991; Fischer et al. 1994; Harris and Ripepe 2007a; Marchetti et al. 2009; McNutt and Williams 2010; Saunders et al. 2012; Harris et al. 2012; Van Eaton et al. 2016; Neal et al. 2019; Poland et al. 2020).

The occasion of the IAVCEI Centennial (1919–2019) (Cas 2022) is a time to reflect on 100 years of scientific and technological advances in volcano seismology and volcano acoustics; advances which have led to the point at which we are today in 2022. One hundred years is a long time for modern science, and advances in volcano seismology and acoustics have been coupled more broadly to developments, in (including, but not limited to) geophysics, tectonics, volcanology, seismology (broadly), acoustics (broadly), physics, applied mathematics, electrical and mechanical engineering, material science, instrumentation, remote sensing, and computer science. For this necessarily finite review, we limit our scope to a highlight of major trends and changes in instrumentation and technology, new discoveries, and paradigm shifts from 1919 to the time of writing (2021 to 2022). Volcanology is an observational science; over the past 100 years, major technological advances have provided progressively sharper tools to make new observations (e.g., the transition from analog to digital recording, event-triggered to continuous waveform data, short-period to broadband), all of which have led to discoveries of new phenomena as well as major shifts in understanding. Similarly, larger eruptions (VEI > 4; Volcanic Explosivity Index; Newhall and Self 1982) and the associated seismo-acoustic unrest and eruption signatures are only available to observe relatively rarely, and instrumentation must be in place at suitable locations (Moran et al. 2008a). Large eruptions and those with societal consequences have generally provided impetus and catalyzed the implementation of new seismic (and more recently acoustic) instrumentation and led to operationalization of research methodologies (e.g., Alcaraz et al. 1952; Philippine Geodetic & Geophysical Institute 1952; Malone 1990; Tayag and Punongbayan 1994; De la Cruz-Reyna and Siebe 1997; Sparks and Young 2002; Yamasato 2005; Gudmundsson et al. 2010; Neal et al. 2019).

Herein, we use the following definitions to refer to observation period (s) or frequency (Hz) bands of volcano seismic and acoustic signals (Ohminato et al. 1998a, b; Chouet and Matoza 2013):

- Ultra-long-period (ULP) > 100 s or < 0.01 Hz;
- Very-long-period (VLP) 2–100 s or 0.01–0.5 Hz;
- Long-period (LP) 0.2–2 s or 0.5–5 Hz; and
- Short-period (SP) 0.05–0.2 s or 5–20 Hz.

Strictly speaking, this terminology refers just to the band of the signal. In addition to the classification based on frequency content, volcano-seismic signals have also been named according to the inferred physical source process (Lahr et al. 1994; Chouet 1996a). In this latter process-based classification system, the most important distinction is between brittle-failure shear or tensile sources that occur in the elastic solid Earth (including so-called volcano-tectonic or VT seismicity), and volumetric sources that actively involve a fluid (including long-period seismicity, which includes individual LP events and tremor). In general, different physical processes occur on different time and spatial scales, but observed volcanic signals often do not fall neatly into these frequency bands (such as ULP, VLP, LP, and SP). Thus, moment-tensor and single-force source-representations provide a more fundamental basis for signal and process discrimination (e.g., Kumagai 2009; Chouet and Matoza 2013; Kawakatsu and Yamamoto 2015).

Volcano seismology in 1919

Instrumental volcano seismology in 1919

By 1919, quantitative instrumental recording of seismic ground motions was well underway using the seismograph, that is, an instrument for measuring seismic ground motion as a continuous function of time as a waveform (Dewey and Byerly 1969). There was, for example, regular reporting of earthquakes since 1883 in Japan allowing early pioneering observational seismology works by Profs. Sekiya Seikei and Fusakichi Omori (e.g., Omori 1894; Dewey and Byerly 1969; Agnew 2002). A seismoscope is an instrument for recording only the occurrence, time, and in some cases duration of an earthquake, but not a waveform record of ground motion. Seismic monitoring using pendulum seismoscopes began at the Manila Observatory, Philippines in 1868 followed with seismographs during the 1880s (Saderra Masó, 1904; Repetti 1946; Udías and Stauder 1996; Bautista and Bautista 2004; Manila Observatory 2016). Mexico installed its first seismograph in 1904 (Pérez-Campos et al. 2018; Suárez and Pérez-Campos 2020). A first national seismic network was deployed in Chile by 1909 (Brenner 1911; Barrientos and National Seismological Center (CSN) Team 2018).

The first dedicated instrumental volcano-seismological observations (Fig. 1) are typically attributed to Luigi Palmieri, with observations of "continuous tremor" at Vesuvius using his "sismografo elettro-magnetico" (developed by Palmieri around 1856), which is formally considered a collection of electromagnetic seismoscopes (Dewey and Byerly 1969). Osservatorio Vesuviano, the world's first volcano observatory, was founded 1841 (Palmieri 1859; Imbò, 1949; Borgstrom et al. 1999; Giudicepietro et al. 2010). The Palmieri seismoscope ran continuously until 1906, and was replaced in 1914 (Giudicepietro et al. 2010). In Japan, Sakurajima was the first volcano to have a seismometer installed nearby (Fig. 1). A Milne-type seismometer was installed at Kagoshima Weather Station in 1888 and later recorded precursory earthquakes to the 1914 eruption (Omori 1916; Yamasato 2005; Iguchi 2013).

A landmark study by Omori (1912) on eruptions and earthquakes of Mount Asama used Omori's two-component horizontal pendulum seismograph "tromometer" (Fig. 2), which was a modification of the earlier horizontal pendulum seismograph of John Milne (Omori 1899). This formed the basis of the Bosch-Omori seismograph, which was later deployed worldwide (e.g., Dewey and Byerly 1969; Klein and Koyanagi 1980; Moore et al. 2018; Suárez and Pérez-Campos 2020; Ammon et al. 2020). The original Omori seismographs did not include viscous damping, which was added in the Bosch-Omori design (Klein and Koyanagi 1980; Okubo et al. 2014). Omori also conducted pioneering observational seismology studies recognizing the forecasting potential for eruptions of Mount Usu in 1910 (Omori 1911) and Sakurajima in 1914 (Omori 1916; Davison 1924). Omori established the first volcano observatory in Japan at Mount Asama in 1911 (Suwa 1980) (Fig. 1). Even in these

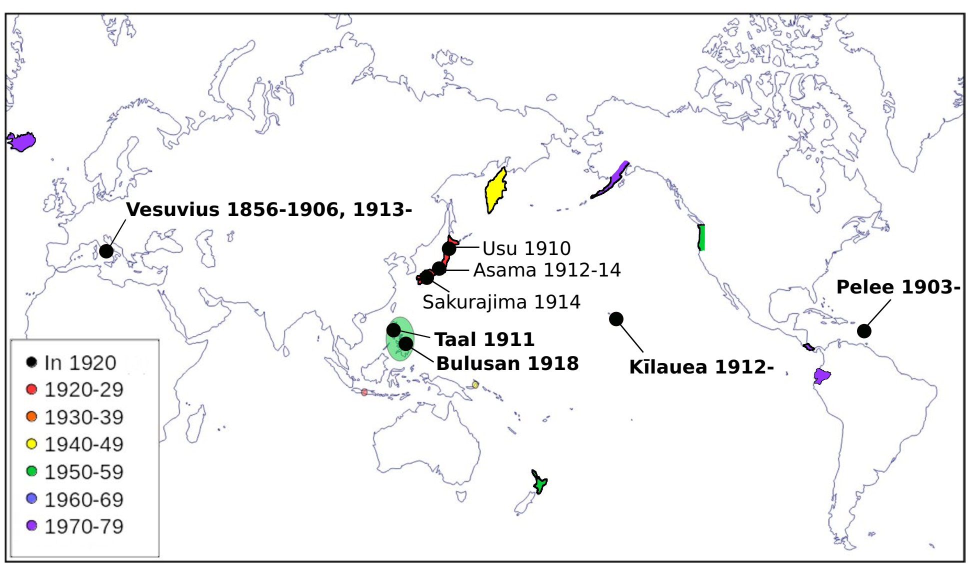


Fig. 1 Expansion of volcano-seismic networks worldwide: 1919 to ~ 1980. Black dots and corresponding labels indicate dates of stations installed on volcanoes by 1919 (bold labels indicate permanent stations). Colored regions indicate timing of initial installation of permanent seismic networks; see Table 1 for details

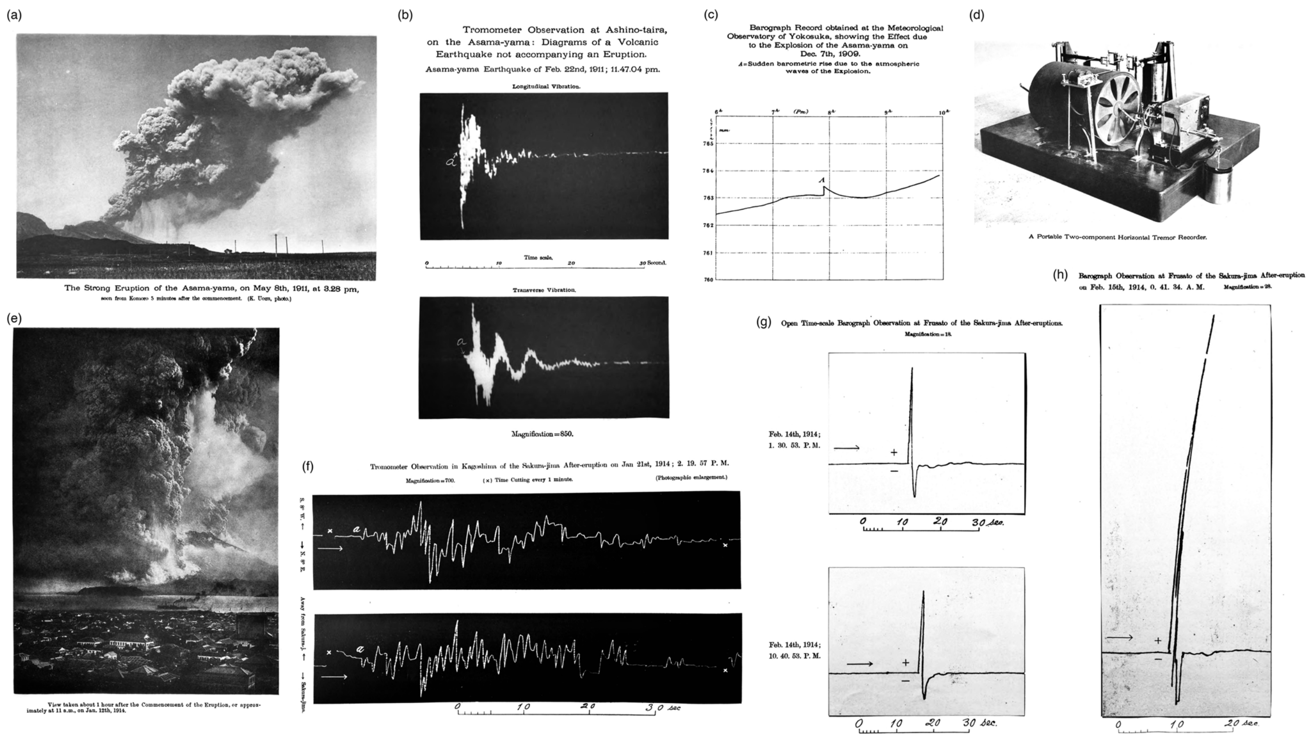


Fig. 2 Seismograph and barograph observations of Mount Asama (**a**–**d**) and Sakurajima (**e**–**h**) by Omori (1912, 1916) [**a**–**d** and **e**–**h** reproduced from Omori (1912) and Omori (1916), respectively; digitally enhanced for image clarity]. (**a**) Mount Asama, 1911 "The strong eruption of the Asama-yama, on May 8th, 1911, at 3:28 pm, seen from Komoro 5 min after the commencement." (**b**) "Tromometer observation at Ashino-taira of the Asama-yama: Diagrams of a volcanic earthquake not accompanying an eruption. Asama-yama earthquake of Feb. 22nd, 1911; 11:47:04 pm." (**c**) "Barograph record obtained at the meteorological observatory of Yokosuka, showing the effect due to the explosion of the Asama-yama on Dec. 7th, 1909." (**d**) "A portable two-component horizontal tremor recorder." (**e**) Sakurajima, 1914 "View taken about 1 h after the commencement of the eruption, or approximately at 11 am, on Jan 12th, 1914." (**f**) "Tromometer observation in Kagoshima of the Sakura-jima after-eruption on Jan 21st, 1914; 2:19:57 pm." (**g**, **h**) "Barograph observation at Frusato of the Sakura-jima after-eruptions" (black-white colors inverted for clarity). In (**g**, **h**), note the asymmetric explosion waveforms which are now commonly captured at volcanoes (including Sakurajima) with modern broadband infrasound instrumentation

earliest instrumental observations it was clear that volcano-seismic signals could be different in character to ordinary crustal earthquakes (Gasparini et al. 1992).

Continuous seismic monitoring at Mount Pelée, Martinique began in 1903 with the installation of a two-component (horizontal) Omori seismograph that operated until 1927 (Fig. 1). However, the station was too far from the volcano (located at a distance of 8.5 km) to detect any weak volcanic seismicity (Lacroix 1904; Hirn et al. 1987). Two of Omori's original seismograph instruments, which were an "ordinary" seismograph and a "heavy" seismograph, were also purchased by Thomas Jaggar and installed at the newly established (founded 1912) Hawaiian Volcano Observatory (HVO) (Klein and Koyanagi 1980; Wright and Takahashi 1989, 1998; Okubo et al. 2014). Jaggar had traveled to Japan in 1909 and met with Omori to learn about the new seismological methods as part of laying the foundation for establishing the HVO (Hawaiian Volcano Observatory 2001; Jaggar 1956). Jaggar later (by July 1913) added two horizontal Bosch-Omori instruments that were operated by HVO until 1963 (Klein and Koyanagi 1980; Apple 1987; Okubo et al. 2014). For further information on the early development of the HVO, the reader is referred to the collections by Wright and Takahashi (1989, 1998) and "The Volcano Letter" collections (see Takahashi 1988).

Simultaneous with Omori's work in Japan, similar pioneering research was conducted by Miguel Saderra Masó in the Philippines at the Weather Bureau (Manila Observatory) (Saderra Masó, 1911a; 1919). The 1911 eruption of Taal was documented in detail by Saderra Masó (1911a) including with observations from Vicentini and Omori seismographs, as well as ten Richard barograph stations installed out to a distance of 242 km (Fig. 3). In a 1911 publication summarizing observations at Taal, Mayon, and Camiguin (Saderra Masó, 1911b), Saderra Masó wrote:

> "The exorbitant toll of human lives levied by the recent eruption of Taal Volcano is a lesson which must not be forgotten, so much the less in view of the fact that, under

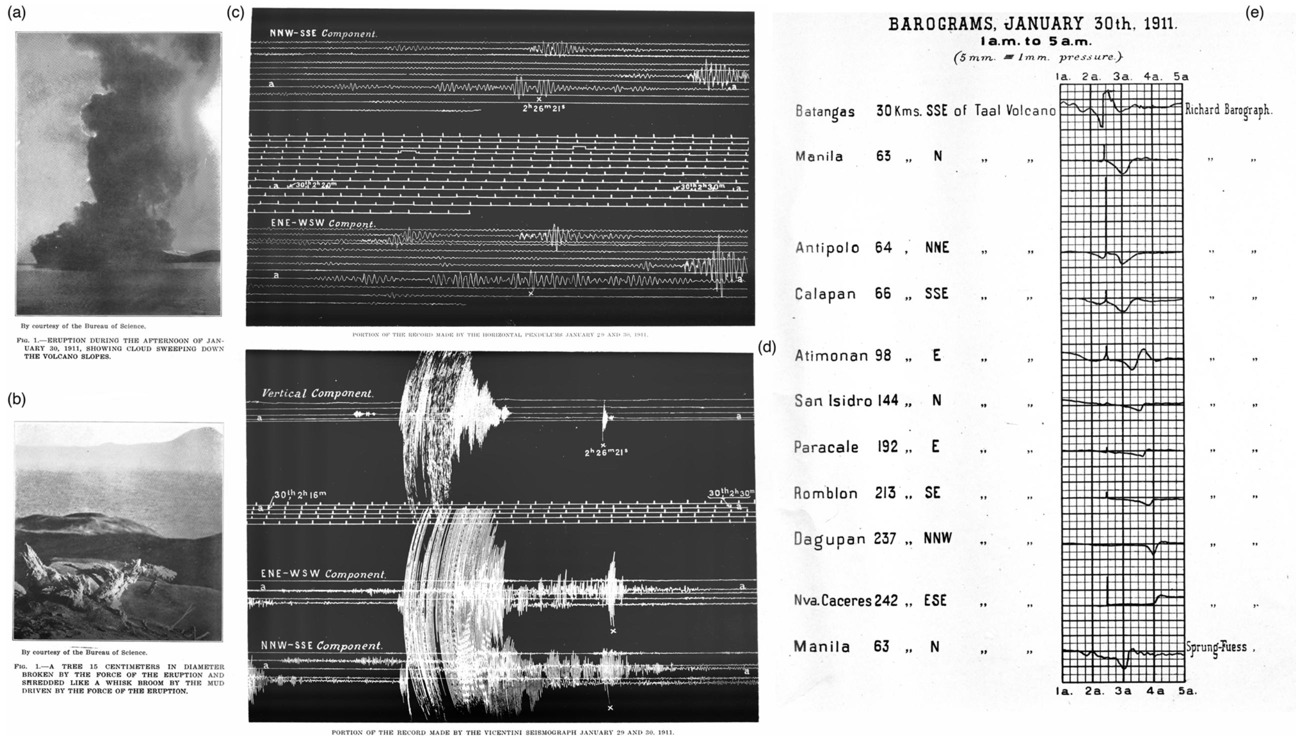


Fig. 3 Seismograph and barograph observations of the 1911 eruption of Taal, Philippines by Saderra Masó (1911a) [Figures reproduced from Saderra Masó (1911a); digitally enhanced for image clarity]. (**a**) "Eruption during the afternoon of January 30, 1911, showing cloud sweeping down the volcano slopes." (**b**) "A tree 15 cm in diameter broken by the force of the eruption and shredded like a whisk broom by the mud driven by the force of the eruption." (**c**) "Portion of the record made by the horizontal pendulums January 29 and 30, 1911." [Omori seismograph] (**d**) "Portion of the record made by the Vicentini seismograph January 29 and 30, 1911." (**e**) "Barograms, January 30, 1911 from 1 to 5 am."; the text on the left gives the location of each barograph station, its range (distance from the source) in kilometers, and the direction from the source (Taal)

> similar circumstances, on a likewise recent occasion (July, 1910) not a single life was lost in Japan [Usu]. These occurrences in Japan and those which we have recently witnessed in connection with the eruption of Taal Volcano, January 30, 1911, prove conclusively that some eruptions can be foreseen; a conclusion likewise stated by the eminent seismologist Prof. F. Omori." (Saderra Masó 1911b)

In a review of Saderra Masó's paper (Saderra Masó 1911b), Harry O. Wood concluded:

> "The moral drawn in the paper is that sundry volcanic eruptions, through the occurrence of earthquakes, or in other ways, can be anticipated in sufficient time to permit the escape of persons whose lives are threatened." (Wood 1912)

Wood was subsequently recruited by Jaggar to establish seismic monitoring at the HVO (Wood 1913), arriving there in summer 1912 (Okubo et al. 2014).

Saderra Masó established a small seismic observatory at Ambulong on the north shore of Lake Taal following the 1911 eruption (Saderra Masó 1911a, 1913; Repetti 1946, 1948). Bulusan volcano had a significant eruption in 1918 which was also documented, including with observations from a seismograph placed at about 8 km distance (Saderra Masó, 1919). In 1920, Saderra Masó represented the, then, world-famous Manila Observatory (Repetti 1948) at the First Pan-Pacific Scientific Conference held in Honolulu, Hawai'i together with Omori and Jaggar who co-organized the seismology and volcanology section (Proceedings of the first Pan-Pacific Scientific Conference 1921).

Volcano seismology scientific framework in 1919

By 1919, it had been well established qualitatively that volcanic eruptions were generally preceded by observable, i.e., felt, seismicity (see, for example, the writings of Pliny the Younger; Sigurdsson et al. 1982), and also that earthquakes at volcanoes do not necessarily lead to eruption (Scrope 1825). Early ideas about the mechanisms by which magmatic processes drove earthquakes were heavily influenced by principles of structural geology. Scrope (1825) posited that earthquakes would occur most strongly at depths where expansive force of magma was strongest. According to Scrope (1825), this led to the, albeit possibly subtle, uplift of shallower strata and consequential, and possibly seismic, dilation/fissuring. This was a prescient connection with the modern continuum between volcano seismology and volcano geodesy. Scrope (1825) further posited that the position of a fissure with respect to its expansive force would control whether magma erupted or remained trapped in the crust, leading to a testable hypothesis about the location and timing of earthquakes with respect to the vent.

In the decades between Scrope's pioneering treatise of 1825 and 1919, the beginnings of instrumental seismology led to debate and refinement of these ideas. Omori (1912) hypothesized that strong volcanic earthquakes resulted from energy released by subterranean explosions that were not simultaneously accompanied by eruption, and that an explosive eruption produced a lower quantity of seismic energy. That is, eruptions were "safety valves" that served to reduce pressure causing large earthquakes (Omori 1912). Omori further hypothesized that the former type of non-eruptive volcanic earthquake would be characterized by a deeper implosive source ("B-type"), and the latter explosion earthquake type by a shallow explosive source ("A-type"). Omori (1912) presented limited evidence for this pattern from a seismograph installed at Mount Asama, which was later refined by Minakami (1960) using data from volcanic and tectonic earthquakes, as shown in Fig. 4. Jaggar (1920), summarizing work by the nascent HVO (Wright and Takahashi 1998), posited that volcanic earthquakes could reflect a wide variety of processes alone or in combination. He hypothesized that volcanic earthquakes occur on existing rift faults stressed past their frictional limit. Jaggar's point that multiple source processes could result in volcanic earthquakes was accompanied by early recognition of a variety of seismic signals such as harmonic and spasmodic tremor (Omori 1914; Jaggar 1920). The early classification scheme based on event depth made by Omori (1912) ultimately evolved into a spectral-based classification scheme, in which spectral differences were hypothesized to correspond to fundamentally different source mechanisms (Minakami 1974; Lahr et al. 1994).

Volcanic waves in the atmosphere in 1919

Atmospheric infrasound (frequency band ~ 0.01 to 20 Hz) is part of a broad spectrum of atmospheric waves produced by volcanic activity that includes gravity waves, acoustic-gravity waves, infrasound, and audible acoustic waves (Gossard and Hooke 1975). By 1919, low-frequency (< 1 Hz) pressure waves from eruptions had been captured instrumentally by meteorological barographs, and research was underway to understand the physics of these atmospheric pressure disturbances and their relation to atmospheric structure. This work based on instrumental observations had begun 36 years earlier with the eruption of Krakatau.

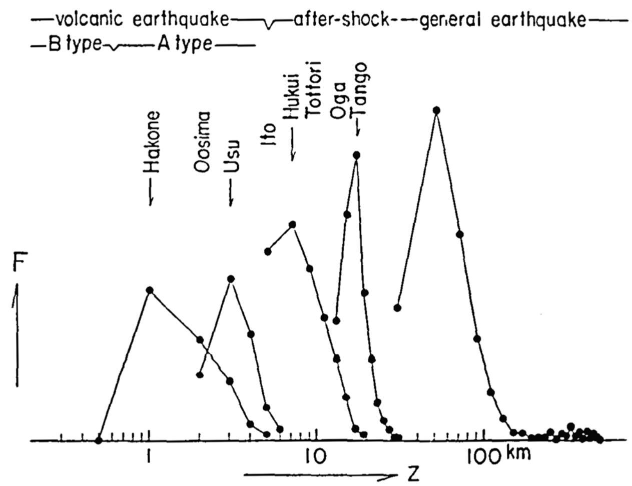


Fig. 4 Comparison of frequency distribution (histogram) of hypocentral depth of "A-type" and "B-type" volcanic earthquakes and tectonic earthquakes from Minakami (1960). Horizontal axis shows approximate depth in km (Z, increasing depth to right; note logarithmic scale), and vertical axis shows occurrence frequency (F, increasing occurrence frequency upwards). Based on these depth distributions, volcanic earthquakes at Oosima (Oshima) and Usu Volcanoes, Japan, were considered to be "A-type" earthquakes, and volcanic earthquakes at Hakone Volcano, Japan were considered to be "B-type" earthquakes. Both types of volcanic earthquakes were shown to have shallower average depths than aftershock sequences following the Ito, Huiki, Tottori, Oga, and Tango mainshock earthquakes, as well as "general" tectonic earthquakes (M > 5) in and near Japan. Figure reproduced from Minakami (1960)

In 1883, over 50 weather barometers around the world recorded (ultra) long-period pressure disturbances from the cataclysmic, VEI 6, August 27 eruption of Krakatau, Indonesia (Scott 1883; Strachey 1884, 1888; Verbeek 1884). A Royal Society of London report compiled the barometric observations and reports of sounds heard (Strachey 1888). Audible cannon-like sounds were reported as far away as ~ 4800 km, similar to historical "earwitness" reports from the earlier 1815 eruption of Tambora (de Jong Boers 1995). The Krakatau atmospheric (ultra) long-period pressure wave propagated around the globe and was recorded as barometric pulses for four minor-arc passages and three major-arc (antipodal) passages (Strachey 1888). It took roughly 1.5 days to make each complete lap, with an average propagation speed of 300–325 m/s; the dominant periods at long range were ~ 100 to 200 min (Gabrielson 2010). These observations stimulated the development of theory to explain what were eventually termed acoustic-gravity waves, and more specifically the surface-guided Lamb wave, and to understand the effects of gravity, buoyancy, and atmospheric structure on their propagation (e.g., LeConte 1884; Lamb 1911; Taylor 1929, 1936; Pekeris 1939; Pierce 1963; Press and Harkrider 1962, 1966; Harkrider

1964; Harkrider and Press 1967; Bretherton 1969; Yeh and Liu 1974; Gabrielson 2010).

The pioneering study of Omori (1912) at Mount Asama, Japan, gave nearly equal emphasis to seismic and atmospheric pressure wavefields, using seismometers and barometers to discriminate between seismic signals associated with airborne explosions ("detonations" and "sound tremors") and non-explosion earthquakes (Fig. 2). Many of the explosion events were audible in settlements at distances of ~ 200 to 300 km, and some were powerful enough to knock out doors and windows. Omori used this information to map the sound propagation and acoustic shadow zones, and began to consider the effects of wind and topography on the acoustic signals; these topics are again active research areas today. Omori continued the analysis of barograph records, for example, at Sakurajima (Omori 1916) (Fig. 2). Saderra Masó (1911a) made similar instrumental (seismograph and barograph) observations for the 1911 eruption of Taal, Philippines (Fig. 3).

The use of weather barometers and infrasonic microphone arrays to study low-frequency (< 1 Hz) atmospheric pressure waves from volcanic explosions at regional to global ranges (tens to thousands of kilometers) continued sporadically throughout the twentieth century, most commonly when large eruptions were recorded on remote barograph or infrasonic microphone arrays, for example, for the eruptions of Mount Pelee, Martinique, 1902 (Anderson and Flett 1903); Bezymianny, Russia, 1956 (Gorshkov 1960); Mount St. Helens, USA, 1980 (Reed 1987; Delclos et al. 1990); El Chichón, Mexico, 1982 (Mauk 1983); Mount Tokachi, Japan, 1988; Sakurajima, Japan, 1989; Pinatubo, Philippines, 1991; Ruapehu, New Zealand, 1995 (Morrissey and Chouet 1997), and Popocatépetl, Mexico (Raga et al. 2002).

Despite early pioneering instrumental studies giving near-equal emphasis to seismic and atmospheric pressure wavefields (e.g., Saderra Masó 1911a; Omori 1912; Perret 1950), advances broadly in seismology and specifically in volcano seismology rapidly outpaced those in atmospheric acoustics until the 1990s (Harris and Ripepe 2007a, b; Fee and Matoza 2013; Chouet and Matoza 2013; Matoza et al. 2019a).

Instrumentation changes 1919–2019

Volcano seismology and acoustics are highly observational fields. The phenomena that can be observed depends upon the available instrumentation. From 1919 to 2019, major advances were made (for example) (1) in instrument sensitivity, i.e., the smallest resolvable amplitude change of ground motion or air pressure that can be measured; (2) in bandwidth, i.e., the frequency range of signals that can be captured; (3) in the portability, compactness, ruggedness, and rapid deployability of instrumentation; (4) in the electronics systems for recording, storing, timing (e.g., GNSS), and telemetering the data; (5) in reducing instrumental power requirements, solar charging, and battery technology; and (6) with the computer revolution, the efficiency with which data could be processed and stored. A comprehensive history of seismometry, microbarograph, and infrasound sensor technology evolution from 1919 to 2019 is beyond our scope. For some of the details, we refer the reader to Dewey and Byerly (1969), Howell (1989), Ben-Menahem (1995), Agnew (2002), Evers and Haak (2010), Ponceau and Bosca (2010), Nief et al. (2019), Marty (2019), and references therein. Major milestones included the transition from analog to digital recording, event-triggered to continuous waveform data, and short-period to broadband, all of which collectively provided a progressively sharper, higher fidelity, wider bandwidth, higher sensitivity, and more temporally continuous capture of the seismic and acoustic signatures of volcanic unrest and eruption. Moreover, a net effect of these technological advances was that the operational seismological monitoring workflow became increasingly efficient, with real-time data transmission and processing enabling the results of seismological analyses to be available more rapidly to inform monitoring decisions (e.g., Klein and Koyanagi 1980; Okubo et al. 2014; Thompson 2015).

Although field logistics at volcanoes will always be demanding, these technological advances have generally also allowed steady expansion in the numbers of seismic and acoustic stations (i.e., increases in network density) at permanently monitored volcanoes (Fig. 5; Table 1) and in campaign research deployments, in turn permitting higher spatiotemporal resolution geophysical inference. Although operating and maintaining permanent seismic monitoring stations at volcanoes is still not straightforward, it is undoubtedly easier now in the days of digital waveform telemetry and low-power ruggedized systems compared to the laborious days of smoked paper or tape recorders. Another promising trend in volcano seismology and acoustics is the increased central archiving and public worldwide sharing of waveform data, which is beginning to allow systematic comparisons and hypothesis testing of seismic and acoustic source processes across varied volcanic systems and tectonic environments.

For operational volcano monitoring, a critical technological advance was the development of radio telemetry (e.g., Eaton 1977; Murray 1992; Lockhart et al. 1992; Thompson 2015). Prior to radio telemetry, data transmission utilized cables including telephone cables. At the HVO, this resulted in miles of overland cables by 1958 and a seismic station distribution limited by cable logistics (Klein and Koyanagi 1980; Klein et al. 1987; Okubo et al. 2014). Radio telemetry thus represented a monumental advance, permitting the expansion of volcano

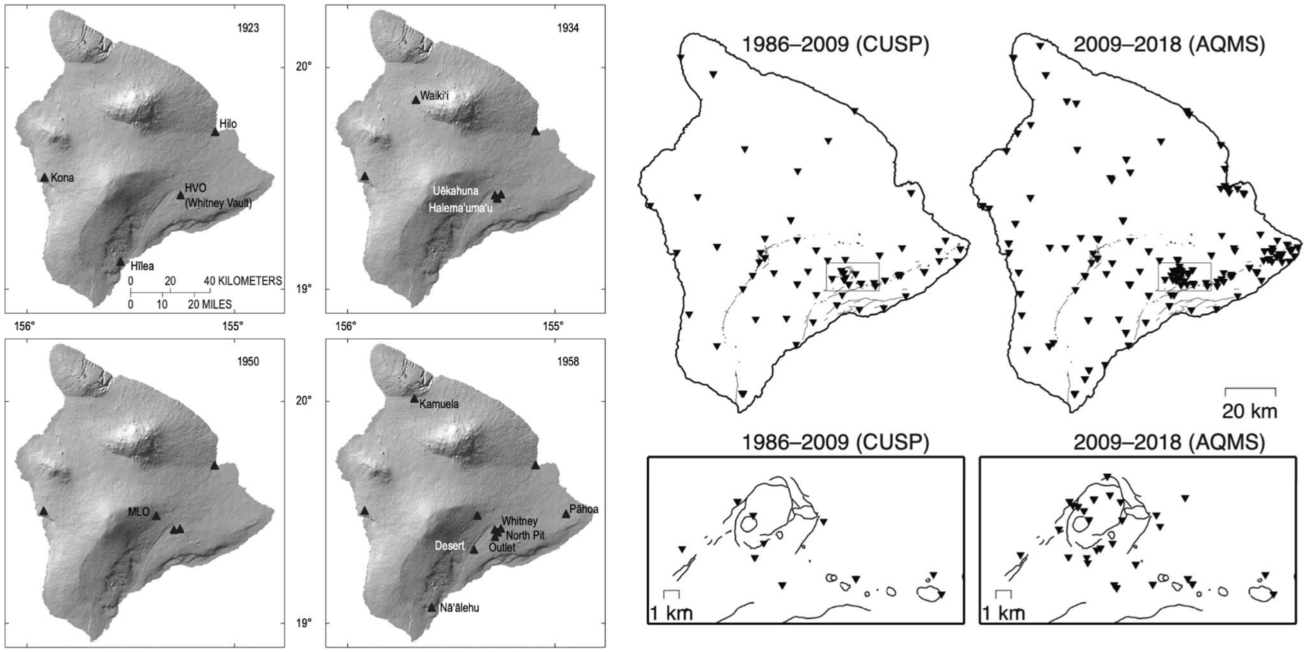


Fig. 5 Expansion of seismic monitoring on the Island of Hawai'i. Figure on left reproduced from Okubo et al. (2014) showing seismic stations (triangles) operating on the Island of Hawai'i in 1923, 1934, 1950, and 1958. Figure on right reproduced from Matoza et al. (2021) showing the HVO seismic network and additional stations on the Island of Hawai'i for which digital event-based waveform data are available from (left) the CUSP system (1986–2009; 144 channels) and (right) the AQMS system (2009–2018; 565 channels). ANSS, Advanced National Seismic System; AQMS, ANSS Quake Management System; CUSP, Caltech-USGS Seismic Processing; HVO, Hawaiian Volcano Observatory. Figures reproduced from Okubo et al. (2014) and Matoza et al. (2021)

Table 1 Expansion of permanent volcano-seismic networks worldwide (illustrative and representative, not complete)

Region/country (volcanoes)	Year(s) of establishment	Reference(s)
Japan (Asama, Aso, Sakurajima)	1910s–1960s	Minakami 1950, Suwa 1980, Wada et al. 1963
Philippines (Taal, Hibok-Hibok)	1910s–, 1950s	Saderra Masó, 1911b; Tayag and Punongbayan 1994
Hawai'i/USA	1910s	Okubo et al. 2014
Indonesia (Merapi, Papandayan, Kelut)	Single station 1924, 1982	Ratdomopurbo and Poupinet 2000, van Padang 1933
Papua New Guinea (Rabaul)	1940s	Fisher 1940
Kamchatka/Russia	1940s	Fedotov et al. 1987, Gorelchik 2001, Gordeev et al. 2006
Pacific Northwest/USA	1950s	Weaver et al. 1990; Norris 1991
New Zealand	1950s	Scott and Travers 2009
Alaska/USA	1960s–1970s	Power et al. 2020
Martinique (Pelee)	1970s	Hirn et al 1987
Iceland	1970s	Einarsson 2018
Ecuador	1980s	Alvarado et al. 2018

seismic monitoring networks worldwide (e.g., Klein and Koyanagi 1980; Ewert and Swanson 1992; Hill 1984; Castellano et al. 2002; Power and Lalla 2010; Giudicepietro et al. 2010; Senyukov et al. 2009; Nishimura and Iguchi 2011). Data from remote and widely distributed instruments could be collected at a central location and analyzed in real time (first on media such as smoked drum paper and later on computerized systems). This advance primarily occurred in the mid-1960s through the early 1970s.

As an illustration of other major technological changes, we consider selected time snapshots containing landmark studies or significant eruptions. We focus the remainder of this section on technological changes from 1970 to 2020, which was a time of major growth in quantitative volcano seismology.

Volcano seismology in the 1970s: limited portability

By the 1970s, field studies at volcanoes using portable seismic instrumentation and computational methods were underway, but the portability was highly limited by today's standards. A 1959 eruption of Kīlauea, Hawai'i, produced a stagnant lava pond at Kīlauea Iki, a pit crater adjacent to Kīlauea summit caldera in the upper east rift zone (Richter et al. 1970), and its slow cooling and solidification provided a landmark opportunity in volcanology (Kauahikaua and Poland 2012; Heiken 2013) and decades of studies including scientific drilling (e.g., Rawson 1960; Wright et al. 1976; Helz 1980, 1993; Helz and Thornber 1987). By the 1970s, the solidified crater floor also enabled seismological investigations, including a refraction experiment performed for a series of geophones deployed along the long axis of the crater floor (Aki et al. 1978) and passive seismic surveys capturing local seismic events originating in the cooling crust of the lake (Chouet 1979). Chouet (1979) developed a quantitative source model for the seismic signals originating within the cooling Kīlauea Iki magma body, parameterized as vertically aligned penny-shaped cracks between columnar basalt joints, with tensile failure (crack opening) due to cooling and solidification of magma. Chouet (1979) presented an analytical expression for the far-field pulse shape of vertical and horizontal ground displacements including attenuation, enabling forward modeling with the crack model to infer cavity volumes, which compared reasonably well with independent estimates based on thermodynamic considerations and a cooling model (also by Chouet 1979).

These were important early studies in quantitative (and computational) volcano seismology, but the limitations of seismic instrumentation technology at the time made installing and maintaining the field equipment highly laborious (B. Chouet, personal communication, 2019 & 2021). For example, a study on coda waves from earthquakes in Hawai'i (Chouet 1976) involved the deployment of 4 mi (6.4 km) of military surplus Spiral-4 cable to connect station OTL (Outlet) on Sand Hill to the HVO. Spiral-4 came in spools weighing 100 lb. (45 kg) each; 40 spools were used. The data were stored on paper (recording at 1 mm/s for months on end), which had to be digitized by hand to be stored on punch cards. The seismic surveys described by Chouet (1979) (conducted in the summer of 1974) used a "portable" Sprengnether MEQ-800 smoke drum recorder with a vertical component (short-period) Mark products L-4C 1 Hz geophone as the sensor (Fig. 6a). All of the data were analog and had to be measured by hand with rulers (B. Chouet, personal communication, 2019 & 2021).

Volcano seismology in the 1980s: digital capture and storage

The 1980s saw the beginnings of digital data capture and storage, based on the recording of event-triggered digital waveforms. The 1980–1986 unrest and eruption sequences of Mount St. Helens, USA, provided opportunities to record signals with new technologies, with access to the crater floor exposed by the 18 May 1980 lateral blast allowing near-field high signal-to-noise ratio recording. A pioneering study by Fehler and Chouet (1982) and Fehler (1983) utilized a prototype 12-bit digital recorder which had been designed and built at the Massachusetts Institute of Technology, USA primarily for ocean bottom deployment (Fig. 6b, c). The deployment consisted of nine short-period seismometers (L4-C 1 Hz geophones with a rapid fall-off in response below 1 Hz) attached to the digital event recorders, which provided event-triggered recording. This involved event-windowed data recording on a magnetic tape that was initiated whenever seismic amplitude rose significantly above the background noise. There was thus no continuous digital recording, but continuous analog recording was made separately on paper chart recorders. The digital recording package for each station consisted of a 4-ft. (1.2 m) tall cylinder containing the signal processing electronics and recorder, but one station was nevertheless deployed in the crater of Mount St. Helens (Fehler and Chouet 1982; B. Chouet, personal communication, 2019 & 2021). Despite these limitations, this deployment provided digital capture of long-period (LP, 0.5–5 Hz) seismicity and tremor at Mount St. Helens (Fig. 6b) (Fehler and Chouet 1982; Fehler 1983), providing new and key observations that initiated a sustained research program to understand the quantitative source mechanism of long-period seismic events and tremor (Chouet and Julian 1985; Chouet 1981; 1985; 1986; 1988; 1992).

Digital recording facilitated digital signal processing, including the application of the Fast Fourier Transform (FFT) (Cooley and Tukey 1965; Cooley et al. 1969) for spectral estimation. Fehler and Chouet (1982) reported LP events with durations ~30 s, spectra peaked in the range 1.7–2.3 Hz, and at depths of between 0 and 5 km. Production of the spectral peaks by a path effect (Malone 1983) was considered inconsistent with the data because the position of the spectral peaks did not change significantly with station location, and a VT earthquake located in the vicinity of the crater observed with the same instruments did not have the same spectral structure as the LPs (Fehler and Chouet 1982). Fehler and Chouet (1982) proposed that the spectral peaks originated from the excitation of a fixed cavity under the active crater. Following Latter (1979), Fehler (1983) also noted the spectral similarity of LP events and tremor, and proposed that tremor consisted of a superposition of randomly occurring LP events. These observations

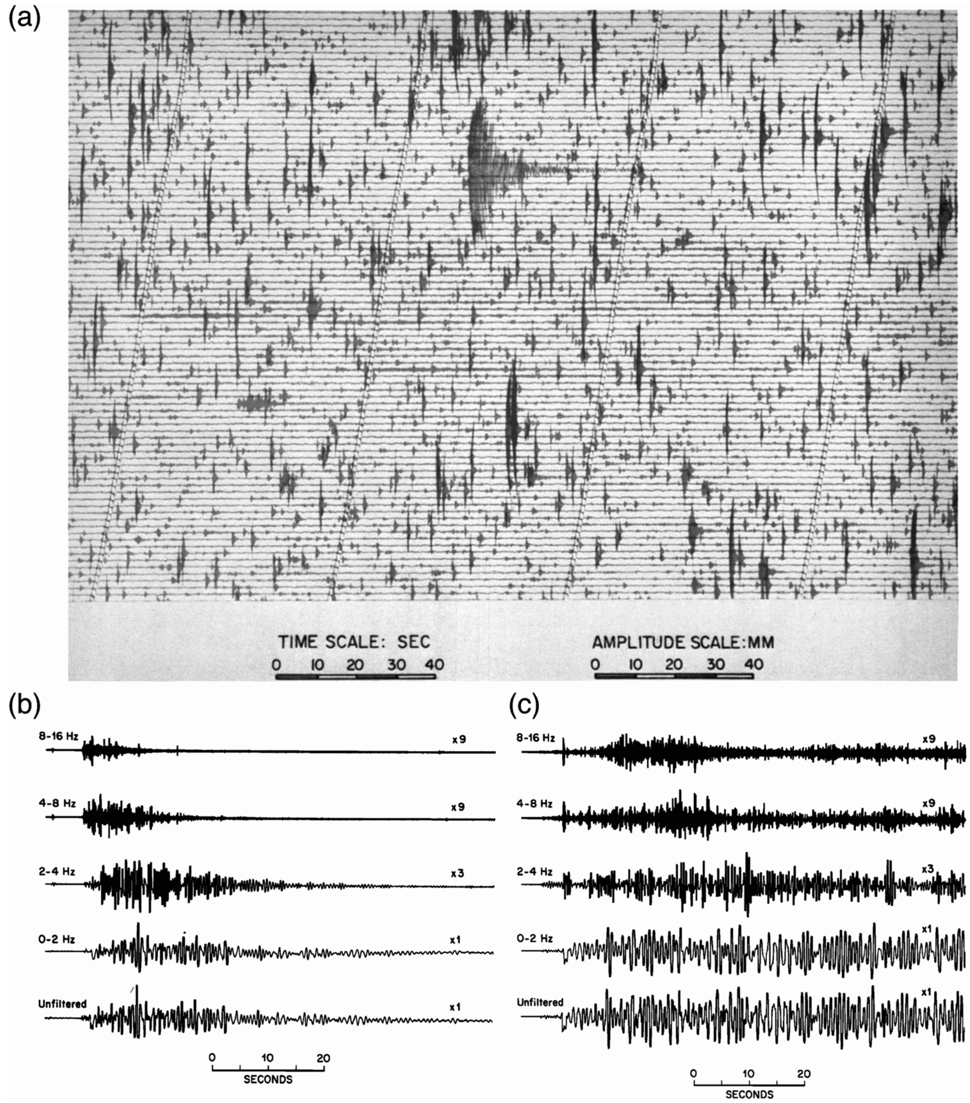


Fig. 6 Instrumentation changes 1970s to 1980s. (**a**) Observation of Kīlauea Iki, 1974 using a Sprengnether MEQ-800 smoke drum recorder with L-4C 1 Hz geophone (Chouet 1979). (**b**, **c**) Prototype 12-bit digital waveform records (also using the L-4C geophone) of (**b**) long-period event in October 1980 and (**c**) tremor at Mount St. Helens (Fehler 1983). The waveforms in (**b**, **c**) are for the same station (vertical component) but filtered in different bands and with different magnification (indicated by text). Figures reproduced from Chouet (1979) and Fehler (1983)

rejuvenated interest in LP event and tremor models in which the fluid plays an active role in generating the signal.

Computer-based earthquake data processing began at HVO in 1979 through digitizing analog tapes. Subsequently, by 1986, automated operational near-real-time seismic network processing with event-triggered digital storage was underway with the Caltech-USGS Seismic Processing (CUSP) system (Fig. 5), which received analog telemetered seismic data and converted it to a digital format (Okubo et al. 2014). Nevertheless, by the time of the 1991 eruption of Pinatubo, Philippines (Tayag and Punongbayan 1994; Punongbayan and Newhall 1999), analog systems remained standard in operational (especially rapid response) monitoring due to their low cost, simplicity of design, and ruggedness (Lockhart et al. 1996). However, digital acquisition, telemetry, and signal processing were becoming increasingly integrated in operations (e.g., Sabit et al. 1996; Ramos et al. 1996, 1999). The 1980s also saw steady expansion of volcano-seismic monitoring capacities worldwide. For example, in 1988, the Instituto Geofísico of the Escuela Politécnica Nacional (IGEPN) of Ecuador began continuous monitoring of Ecuadorian volcanoes with single telemetered seismic stations at Tungurahua, Cotopaxi, Cuicocha, Chimborazo, Antisana, and Cayambe, and seismic and geodetic networks were established at Guagua Pichincha (Alvarado et al. 2018). Several volcanological and seismological observatories were also established in Colombia by the Colombian Geological Survey (formerly Instituto Colombiano de Geología y Minería INGEOMINAS) in the late 1980s (Vargas et al. 2018).

Volcano seismology in the 1990s: portable broadband seismometry, infrasound, continuous digital waveform data

The 1990s saw the advent of portable broadband seismometry at volcanoes capturing waveforms in the VLP and ULP bands, immediately leading to the discovery of new signals and phenomena (Fig. 7) (e.g., Kawakatsu et al. 1992; Neuberg et al. 1994; Kaneshima et al. 1996; Arciniega-Ceballos et al. 1999). The reader is referred to the reviews by Chouet and Matoza (2013) and Kawakatsu and Yamamoto (2015) for an overview of VLP and ULP observations, inversions, and modeling studies during this time. Previously unobservable with standard short-period instrumentation, VLPs represented entirely new signals reflecting slower processes associated with unsteady mass transport, and commonly attributed to fluid–rock interaction or longer-term inertial volume changes in fluid-filled conduits (e.g., Kawakatsu et al. 1992; Ohminato et al. 1998a, b; Nishimura et al. 2000; Kumagai et al. 2003; Kumagai 2006; Chouet and Dawson 2011).

Advances in computer processing and storage by the 1990s also enabled the transition to continuous digital waveform recording and storage for an expanding number of stations, and made tractable full-waveform inversions using synthetic Green's functions taking into account topography (Ohminato et al. 1998a, b; Kumagai et al. 2002a; Chouet et al. 2003, 2005; Nakano and Kumagai 2005). As a result, broadband observations rapidly became quintessential in volcano-seismic monitoring worldwide (e.g., Martini et al. 2007; De Cesare et al. 2009; Neuberg et al. 1998; Kawakatsu et al. 2000; Iguchi 2013). For example, a semi-permanent digitally telemetered (continuous data) 10-station broadband network was established at Kīlauea beginning November 1994, immediately capturing a variety of new signals and processes and augmenting monitoring capacity (Dawson et al. 1998).

Audio range volcanic sound microphone recordings (> 20 Hz)

Frank Perret made probably the first recordings of sounds in the audio range (frequencies > 20 Hz) from volcanoes using moving-coil microphones at Vesuvius in 1906, eventually also recording signals at Etna, Stromboli, Kīlauea, Sakurajima, Mount Pelée, and Soufrière Hills (Perret 1950). The first tape recordings of volcanic sounds were apparently made by the NHK (Nippon Hōsō Kyōkai) Broadcasting Bureau of Japan (Snodgrass and Richards 1956). In 1952, a program of volcanic acoustics was initiated by James Snodgrass at the Scripps Institution of Oceanography, USA, leading to a decade's worth of underwater and airborne acoustic recordings of volcanic sounds with frequencies > 50 Hz (Richards 1963). The paper by Richards (1963) summarizes these observations, relating the various sounds to different idealized styles of volcanic activity.

A pioneering study of acoustic signals (> 20 Hz) by Woulff and McGetchin (1976) represents the first attempt at a quantitative link between acoustic radiation and fluid mechanics at volcanoes using equivalent source theory. This study introduced the idea of using radiated acoustic power and frequency content to infer erupted gas exit velocity for assumed equivalent monopole, dipole, and quadrupole source types. Woulff and McGetchin (1976) only considered

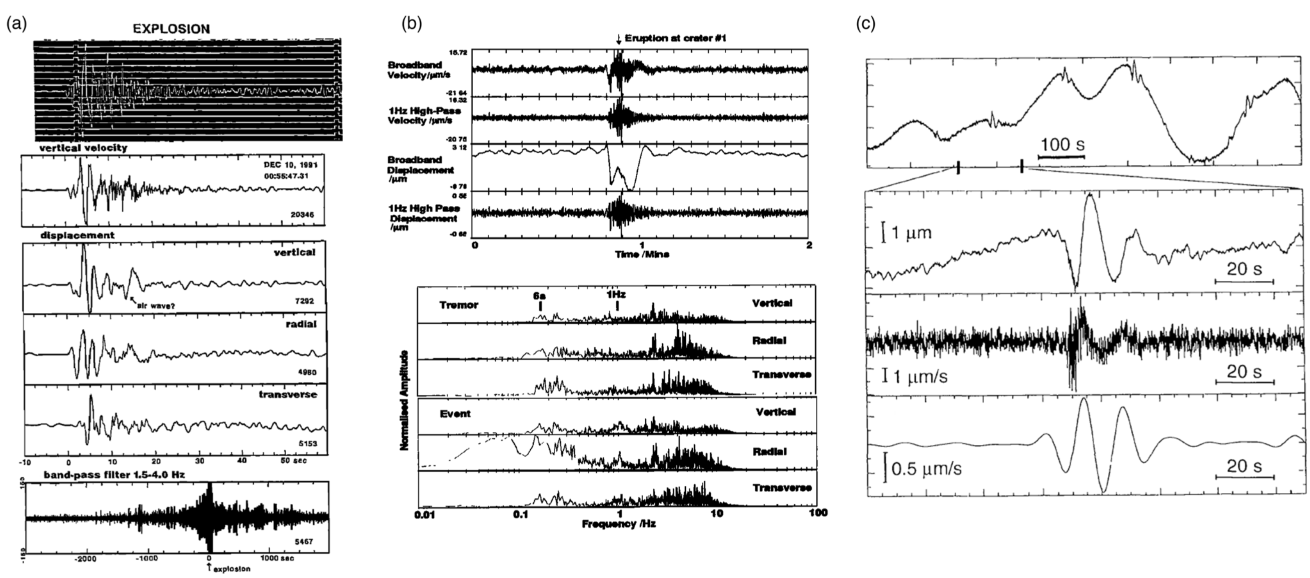


Fig. 7 The advent of broadband volcano seismology (examples). (**a**) Explosion event at Sakurajima reported by Kawakatsu et al. (1992). (**b**) Eruption at Stromboli reported by Neuberg et al. (1994). (**c**) Broadband waveform attributed to a hydrothermal reservoir at Aso reported by Kaneshima et al. (1996). (**a**) Reproduced from Kawakatsu et al. (1992); (**b**) reproduced from Neuberg et al. (1994); (**c**) reproduced from Kaneshima et al. (1996)

audio range acoustic signals > 20 Hz, but later infrasound (< 20 Hz) studies built extensively upon this concept (e.g., Firstov and Kravchenko 1996; Vergniolle et al. 1996; Johnson 2003, Vergniolle and Caplan-Auerbach 2006; Matoza et al. 2009a; Caplan-Auerbach et al. 2010; Kim et al. 2012; Ripepe et al. 2013; Lamb et al. 2015; Delle Donne et al. 2016; Fee et al. 2017; Haney et al. 2018; Iezzi et al. 2019a, 2022; Perttu et al. 2020b). In a reexamination of the original formulation of Woulff and McGetchin (1976) in the context of the current understanding of jet noise, Matoza et al. (2013b) concluded that the formulation of Woulff and McGetchin (1976) can lead to large errors when inferring eruption parameters from acoustic data and thus requires modification.

Remote volcano infrasound observations from the 1960s to the 1990s

In volcano seismology, the frequency band from ~ 0.01 to 20 Hz (which includes VLP, LP, and SP), is particularly important for signals of volcanic unrest and eruption. In atmospheric acoustics, this band is termed infrasound (e.g., Pierce 1981; Bedard and Georges 2000; Hedlin et al. 2002; Evers and Haak 2010). Progress in the field of volcano (atmospheric) acoustics was therefore modest until microphones targeting these frequencies were deployed near active volcanoes. As we reviewed above, barograph records capturing atmospheric pressure wave signals with frequencies < 1 Hz were documented since the 1883 Krakatau eruption. However, the frequency limit of the barograph instrumentation (< 1 Hz), together with their prime usage as weather stations, resulted in an observational bias toward larger eruptions recorded at long ranges.

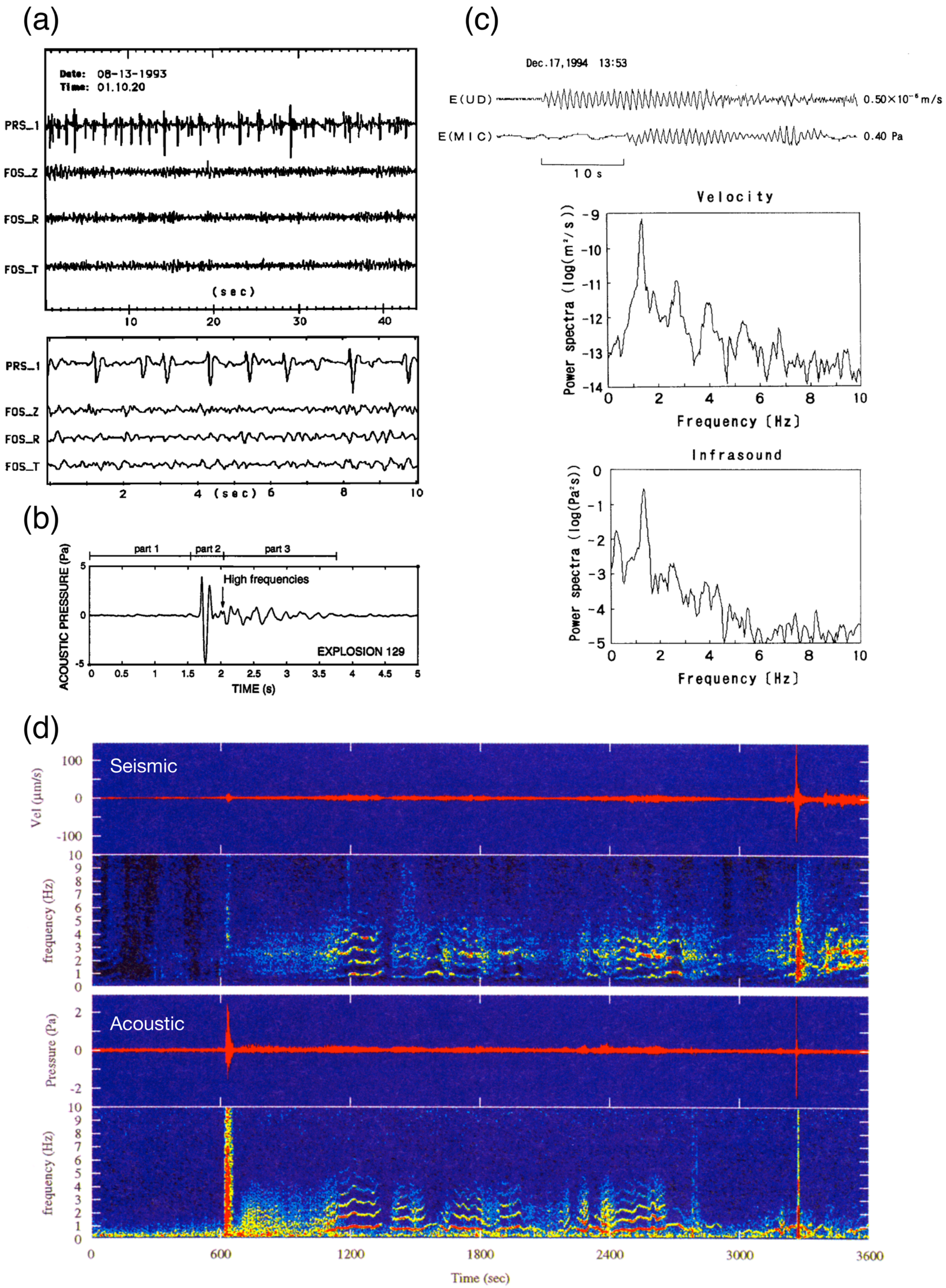
Reviews of some aspects of the history of general infrasound research can be found in Bedard and Georges (2000), Hedlin et al. (2002), and Evers and Haak (2010). The era of atmospheric nuclear testing from 1945 to 1963 (the 1963 Limited Test Ban Treaty then prohibited nuclear weapon tests in the oceans, atmosphere, and space) resulted in active research programs in infrasound, including the development of sensors, spatial wind-noise filtration systems, and array processing methods, particularly between the years 1945 and 1967 (Thomas et al. 1971). Between 1967 and 1985, infrasound research continued with geophysical studies of weather, meteors, aurorae, and volcanoes, and this time period saw the first utilization of low-frequency infrasound microphone arrays to detect remote volcanic eruptions (in the band 0.01–0.1 Hz).

Goerke et al. (1965), Wilson et al. (1966), and Wilson and Forbes (1969) provided some of the first infrasonic microphone array observations of volcanic eruptions in the low infrasound band (0.01–0.1 Hz). The 1963 eruption of Mount Agung, Bali, was recorded 14,700 km away in Boulder, Colorado (Goerke et al. 1965), and the 1967 eruptions of Redoubt and Trident Volcanoes, Alaska, were recorded in Fairbanks, Alaska (Wilson et al. 1966; Wilson and Forbes 1969). The main emphasis of these studies was the atmospheric propagation of the signals. Infrasonic microphone arrays were then installed at Kariya, Japan (Tahira 1982), and Windless Bight, Antarctica, 26 km from Mount Erebus (Dibble et al. 1984). Although limited to the 0.1–1 Hz band, the Kariya array routinely detected explosions from Sakurajima at a range of 710 km and also recorded the 1991 Pinatubo eruption at a range of 2770 km. These data were used to infer eruptive time-histories when visual or instrumental observations close to the volcano were impossible (Tahira et al. 1996).

The first proposal for an acoustic early warning system for explosive eruptions of which we are aware was that of Kamo et al. (1994). Kamo et al. (1994), following work by Tahira (1982), demonstrated that an array at Kariya, 710 km from Sakurajima, was capable of detecting infrasound from volcanoes thousands of kilometers distant and showcased example signals from the 1991 eruption of Pinatubo (see also Tahira et al. 1996). Kamo et al. (1994) concluded that "*this capability forms the basis of a proposal for a worldwide network of air-wave sensors to monitor volcanic explosions,*" proposing the "PEGASAS-VE" ("pressure gage system for air-shocks by volcanic eruptions") early warning system for aviation safety that would consist of a set of infrasonic microphone arrays with a 500–1000 km spacing. Kamo et al. (1994) proposed that PEGASAS-VE "*would be a very effective means of enhancing aviation safety and would be similar to the tsunami warning system, which is in worldwide operation.*" Although PEGASAS-VE was not constructed, the International Monitoring System (IMS) infrasound network was initiated after the Comprehensive Nuclear-Test-Ban Treaty (CTBT) was opened for signature in 1996 (e.g., Christie and Campus 2010; Marty 2019; Le Bras et al. 2021). The proposed 500 km spacing of the PEGASAS-VE design was chosen to provide timely warnings of volcanic eruptions within 30 min based on infrasound propagation time. The average station spacing for the complete IMS infrasound network will be about 2000 km (Christie and Campus 2010), so additional stations will be needed to augment the IMS infrasound network (e.g., Matoza et al. 2007, 2011a,b; 2017, 2018; Garcés et al. 2008; Fee et al. 2010b; Tailpied et al. 2013, 2016; Nishida and Ichihara 2016; Ripepe et al. 2018; Taisne et al. 2019; Perttu et al. 2020a; Le Pichon et al. 2021) and achieve the vision outlined in the original PEGASAS-VE proposal (Kamo et al. 1994).

Volcano infrasound in the 1990s

Volcanic infrasound in the band 1–20 Hz (termed *near-infrasound*) was collected at local recording ranges (defined


(a)
Date: 08-13-1993
Time: 01.10.20
PRS_1
FOS_Z
FOS_R
FOS_T
(sec)
10
20
30
40
PRS_1
FOS_Z
FOS_R
FOS_T
2
4 (sec) 6
8
10
(b)
part 1
part 2
part 3
High frequencies
EXPLOSION 129
ACOUSTIC PRESSURE (Pa)
TIME (s)
(c)
Dec.17,1994 13:53
E(UD)
0.50×10⁻⁶ m/s
E(MIC)
0.40 Pa
10 s
Velocity
Power spectra (log(m²/s))
Frequency (Hz)
Infrasound
Power spectra (log(Pa²s))
Frequency (Hz)
(d)
Seismic
Vel (μm/s)
frequency (Hz)
Acoustic
Pressure (Pa)
frequency (Hz)
Time (sec)

◂**Fig. 8** Volcano infrasound in the 1990s (examples). (**a**) Infrasound and seismic observations at Stromboli by Ripepe et al. (1996). (**b**) Infrasound at Stromboli analyzed by Vergniolle and Brandeis (1994). (**c**) Seismic and infrasonic observation of "C-type" (harmonic) tremor at Sakurajima by Sakai et al. (1996). (**d**) Seismic and infrasonic harmonic tremor at Arenal reported by Hagerty et al. (1997) and Garces et al. (1998). (**a**) Reproduced from Ripepe et al. (1996); (**b**) reproduced from Vergniolle and Brandeis (1994); (**c**) reproduced from Sakai et al. (1996); (**d**) reproduced from Garces et al. (1998)

as < 15 km) during a small number of field studies in Kamchatka and Antarctica in the 1970s and 1980s (e.g., Dibble et al. 1984; Firstov and Kravchenko 1996; Gordeev et al. 1990). Observations of local volcano near-infrasound greatly expanded in the 1990s with field studies particularly in Japan and at Stromboli, Italy (Fig. 8). In Japan, Iguchi and Ishihara (1990) and Yamasato (1997) installed infrasonic microphones at distances of 2–5 km from Sakurajima, Suwanosejima, and Unzen, recording numerous explosions and pyroclastic flows (Yamasato 1997), harmonic infrasonic tremor (Sakai et al. 1996), and impulsive signals associated with LP seismic events (Iguchi and Ishihara 1990; Yamasato 1998). The study by Sakai et al. (1996) at Sakurajima is considered the first observation of infrasonic harmonic tremor, but this publication was followed rapidly by similar observations at Arenal (Hagerty et al. 1997, 2000; Garces et al. 1998) (Fig. 8) and Karymsky (Johnson et al. 1998). Acoustic studies began at Stromboli in the early 1990s (Braun and Ripepe 1993; Vergniolle and Brandeis 1994; Buckingham and Garcés 1996). Since then, it has become increasingly clear that an array of volcanic processes produces a variety of types of infrasound signals across the 0.01–20 Hz frequency range (e.g., Harris and Ripepe 2007a, b; Johnson and Ripepe 2011; Garces et al. 2013; Fee and Matoza 2013; Matoza et al. 2019a; Marchetti et al. 2019).

In 1996, the CTBT was opened for signature, leading to the construction of the IMS. The IMS included a global infrasound network with, as of 2022, 53 certified infrasound stations of a planned total 60 (Marty 2019). Construction of the IMS led to rapid advances in infrasound technology, such as improvements in instrumentation, signal-processing methods, and infrasound propagation modeling. These technologies have all been transferred and adapted to understand and monitor volcanic processes (Garces et al. 2003, 2008; McCormack et al. 2005; Matoza et al. 2007, 2019a; Matoza and Fee 2018).

Volcano seismo-acoustics: 2000 to 2020

During the past 20 years, the fields of volcano seismology and volcano acoustics have been progressively merging, as captured in the term *volcano seismo-acoustics*, and in-line more broadly with the emergence of the discipline of *seismo-acoustics* (Arrowsmith et al. 2010). These complementary geophysical technologies provide more complete capture of the signals of unrest and eruption, from the mantle to the surface, and reduce ambiguity in signal and process identification. From 2000 to 2020, high-quality, broadband, and well-calibrated infrasound sensors have become increasingly portable, lower power, and rapidly deployable, mirroring trends in portable broadband seismology. Digitizers, power systems, and other electronic components of the system are highly similar for infrasound and seismic stations, making these complementary channels easy to record together and leveraging general technological advances in seismology. Telemetry systems have also increasingly moved from analog to digital. The large community EarthScope USArray Transportable Array catalyzed advances in seismo-acoustic systems technology (Busby et al. 2018). For research campaign-style deployments relying on local data storage (no telemetry), high-volume data storage has increasingly permitted the collection of longer multi-year datasets, with more recording channels, at higher sample rates, and with easier field logistics (infrequent data downloads). In addition, cellular modems (where coverage is available) now make possible streaming of remote data without purpose-built telemetry systems (e.g., Busby et al. 2018; Sanderson et al. 2021; Shiro et al. 2021). Numerous other advances in instrumentation technology have driven progress. For example, for reviews of infrasound sensor and wind noise reduction system developments, the reader is referred to Ponceau and Bosca (2010), Nief et al., (2019), and Raspet et al. (2019).

Permanent volcano monitoring networks have progressively established denser seismic networks (Figs. 1 and 5), with continuous digital waveform acquisition, processing, and storage now being the standard. Infrasound technology has been increasingly integrated in volcano-seismic monitoring operations and is also rapidly becoming standard (e.g., Orazi et al. 2013; Ruiz et al. 2013; Iguchi 2016; Coombs et al. 2018; Alvarado et al. 2018; Yokoo et al. 2019; Taisne et al. 2019; Power et al. 2020).

The internet, central data archiving, and legacy data

A general trend in seismology has been toward increased central archiving and public sharing of waveform data facilitated by data management centers worldwide, e.g., IRIS (Incorporated Research Institutions for Seismology), founded 1987 (Smith 1987); GEOFON (GEOFOrschungsNetz), founded 1992 (Quinteros et al. 2021); GEOSCOPE (French Global Network of broad band seismic stations), founded 1982 (Roult et al. 2010); ORFEUS (Observatories and Research Facilities for European Seismology), founded 1988 (van Eck and Dost 1999); MEDNET (Mediterranean Very Broadband Seismographic Network), founded 1987 (Boschi et al. 1991); and POSEIDON (Pacific Orient Seismic Digital Observation Network), founded 1989 (Geller 1974; Shimazaki et al. 1992). The growth of the internet accelerated these trends

(Malone et al. 1993; Malone 1995). These efforts have collectively had profound impacts on seismology (e.g., Malone et al. 1993; Malone 1995; Aster et al. 2004) and, more recently, infrasound (e.g., Hutko et al. 2017; Busby and Aderhold 2020; Stammler et al. 2021) technology and research. Data management centers have become increasingly important with the rise of continuous data streams and growing data volumes. In the past decade, volcano seismo-acoustics research infrastructure has increasingly moved toward the standards of FAIR (Findable, Accessible, Interoperable, and Reusable). The central archiving and public accessibility of seismo-acoustic data allows systematic comparisons across varied volcanic systems and tectonic environments. Preserving data for the future will further ensure that waveform details from landmark eruptions will be available for retrospective analyses with new processing methodologies, or in light of new paradigms and hypotheses.

For the same reasons, the preservation of legacy seismic data is an urgent priority (e.g., Bogiatzis and Ishii 2016; Richards and Hellweg 2020; Hwang et al. 2020; Pérez-Campos et al. 2020). As seismic and acoustic waves propagate over distance, wave amplitude loss occurs from geometrical spreading, attenuation, and scattering, generally resulting in information loss about volcano seismic and acoustic unrest and eruption signals with increasing distance from the volcanic source. Thus, seismic and acoustic instrumentation provide the most information when deployed as dedicated monitoring instruments (< 50 km range). However, seismic signals from eruptions or unrest sequences have still been captured and usable, to some degree, by more distant stations on early instrumental records. For example, some seismicity from the 1943 eruption of Parícutin was recorded by a Wiechert seismograph at the Tacubaya seismic station in Mexico City at ~ 320 km distance from the source (Yokoyama and de la Cruz-Reyna 1990). Similarly, the 1963–1967 eruption of Surtsey, Iceland was recorded by two stations at distances > 100 km (Sayyadi et al. 2021). This underscores the importance of preserving and making available legacy seismic data, especially data from classic eruption case studies in volcanology (e.g., Malone 2020; Thompson et al. 2020; Lee et al. 2020; Sayyadi et al. 2021).

Progression in understanding of volcano seismic source processes 1919–2019

Volcano seismology involves both the analysis of seismic signals generated by volcanic processes and the application of seismic techniques to image internal volcanic structure. We focus our review in this section largely on the former (analysis of volcano-seismic signals). However, these objectives are closely related, since the ability to accurately recover seismic source mechanisms depends upon the resolution of the velocity structure of the volcanic edifice and upper crust (e.g., Bean et al. 2008; De Barros et al. 2011; Dawson et al. 2011). For reviews and perspectives on advances in seismic imaging of internal volcanic structure, we refer the reader to Lees (2007), Chouet and Matoza (2013), Saccorotti and Lokmer (2021), Koulakov and Shapiro (2021), and Thelen et al. (2022).

We also limit our primary focus to the progression in understanding over the past hundred years of volcano-tectonic (VT) and long-period (LP) seismicity (0.5–5 Hz), which includes individual transient LP events and more temporally continuous tremor. This choice is made since VLP seismicity was discovered as recently as the 1990s and has already been adequately reviewed by Chouet and Matoza (2013). Recent advances have also been made at the longer ULP time-scales approaching static. A review of ULP signals is also provided by Chouet and Matoza (2013) and these signals have been increasingly amenable to observation and analysis over the past decade. A primary advance for the ULP band has been the development of waveform inversion methods that account for contributions from both translation and tilt in horizontal seismograms through the use of Green's functions representing the seismometer response to translation and tilt motions (Maeda et al. 2011, 2017; Chouet and Dawson 2015; van Driel et al. 2015; Waite and Lanza 2016; Jolly et al. 2017a). Thus, volcano seismology presently provides quantitative models of the seismic source process related to a variety of volcanic processes over an extremely wide band spanning the LP, VLP, and ULP bands (Maeda et al. 2017; Chouet and Dawson 2015). However, until the advent of broadband seismometry at volcanoes in the 1990s, LP and VT sources were a primary focus of volcano seismology.

Long-period seismicity: LP events and tremor

Long-period (LP, 0.5–5 Hz) seismicity includes individual transient LP events and more continuous tremor (e.g., Kawakatsu et al. 1992; Kaneshima et al. 1996; Narváez et al. 1997; Gil Cruz and Chouet 1997; Neuberg et al. 2000; Aki and Ferrazzini 2000; Saccorotti et al. 2001; Kumagai et al. 2002b; Nakano et al. 2003; Lesage et al. 2006; Waite et al. 2008; Nakamichi et al. 2009; Palo et al. 2009; Alparone et al. 2010; Matoza and Chouet 2010; Buurman and West 2010; D'Auria et al. 2011; Traversa et al. 2011; Arciniega-Ceballos et al. 2012; Rodgers et al. 2013; Matoza et al. 2014a; Unglert et al. 2016; Battaglia et al. 2016a; Lyons et al. 2016; Bell et al. 2017; Frank et al. 2018; Soubestre et al. 2018; Park et al. 2019). The escalation of LP seismicity at shallow depth (< 2 km) in a volcanic edifice is often explained in terms of the pressure-induced disruption of a shallow hydrothermal region, and is one of the most significant indicators of volcanic unrest (e.g., Chouet et al. 1994; Chouet 1996a; Chouet and Matoza 2013). Long-period events are transient signals characterized by a short-lived (~10 s) broadband onset, followed by a coda of decaying harmonic oscillations lasting from tens of seconds to a few minutes in duration (Chouet 1996a). This is commonly interpreted as a broadband, time-localized pressure excitation mechanism (or trigger mechanism), followed by the response of a fluid-filled resonator (Chouet 1996a). Long-period events are typically associated with volumetric source mechanisms when moment-tensor representations are possible to determine (Chouet and Matoza 2013). Volcanic seismic tremor is a more continuous vibration of the ground with observed durations of minutes to hours, or even weeks to years in some cases (McNutt 1992). Observations of volcanic tremor are multifarious and tremor apparently results from a variety of fluid processes (e.g., McNutt 1992; Konstantinou and Schlindwein 2003; Chouet 1996b). Worldwide observations of volcanic tremor show a wide variability in temporal durations, signal amplitudes, and frequency contents. Accordingly, various terms have been introduced over the years to capture the variety in tremor observations and physical interpretations. These include, but are not limited to, harmonic tremor, monotonic/monochromatic tremor, spasmodic tremor, eruption tremor, banded tremor, and tremor storm (e.g., Seidl et al. 1990; McNutt 1992; Konstantinou and Schlindwein 2003). For example, "eruption tremor" is still commonly used to describe broadband tremor directly associated with sustained explosive eruptions (Scandone and Malone 1985; McNutt and Nishimura 2008).

The classification of seismic signals associated with processes operating in complex natural systems is not straightforward, and these descriptive terms have thus consequently been applied in various ways in the literature, and in some cases have evolved over time. One of the earliest distinctions made by Jaggar, following Omori in the early twentieth century, was that between "spasmodic" tremor (i.e., irregular vibrations) and "harmonic" tremor (i.e., more rhythmic vibrations) (Omori 1908, 1911, 1916; Jaggar 1920). However, since the advent of spectral analyzers and later digital signal processing from the 1970s onwards, the term "harmonic tremor" has evolved to generally imply tremor with sharply peaked spectra (Fig. 9), but the tremor spectral peaks do not always follow a simple harmonic progression (Lesage et al. 2006; Matoza et al. 2010). Whether the spectral character of LP and tremor events is related to a source, path, or site effect (Goldstein and Chouet 1994; Chouet et al. 1997) has been discussed extensively (e.g., Malone 1983; Fehler and Chouet 1982; Bean et al. 2014; Chouet and Dawson 2016). Untangling source, path, and site effects has, however, become progressively more robust with more recent data, for example, from denser broadband seismic networks (e.g., Waite et al. 2008; Chouet and Dawson 2016; Lyons et al. 2016; Matoza et al. 2022a). Clear multiparameter evidence for source resonance includes infrasound signals recording the same spectral signature as co-located seismic instrumentation but for a different (atmospheric) path (Garcés et al. 1998) (Fig. 8d), video data capturing breathing mode gas oscillations from a vent coincident with a Helmholtz resonance spectral infrasonic signature (Fee et al. 2010c), and the observation of multiple gas eruption jets related to the production of dual overlapping gliding harmonic seismic spectral evolution (Lesage et al. 2006).

Previously, it was already noted from about the 1980s that LP events and tremor have similar spectral properties and are closely temporally related, with for example swarms of individual LP events merging into tremor and back into LP events (e.g., Latter 1979; Fehler 1983; Neuberg et al. 1998, 2000; Powell and Neuberg 2003; Hotovec et al. 2013). This particular type of tremor thus clearly has a common origin with the individual LP events. These observations led to the interpretation that LP events represent the impulse response of a resonant tremor-generating system, and that some types of tremor consist of the superposition of many individual LP events (Latter 1979; Fehler and Chouet 1982; Fehler 1983; Chouet 1985). This type

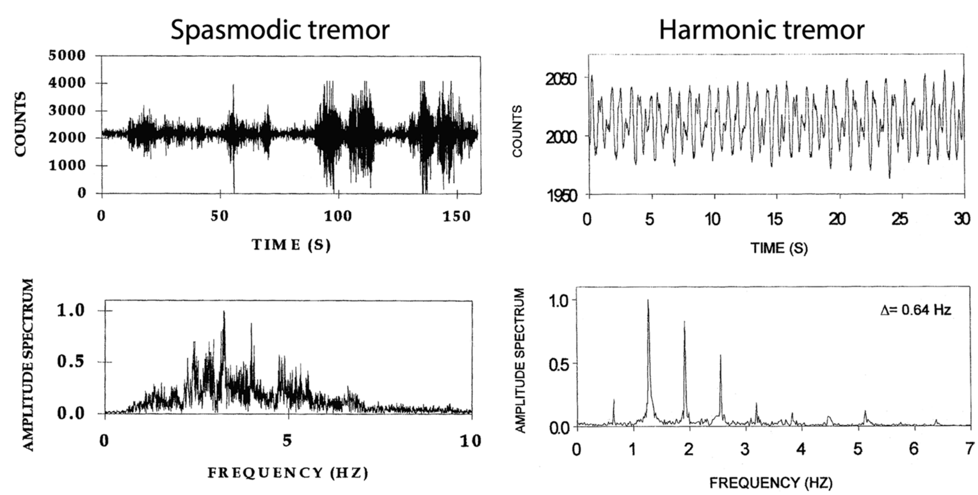


Fig. 9 Example seismograms and their normalized amplitude spectra showing "spasmodic" tremor (left) and "harmonic" tremor (right) at Galeras, Colombia. Reproduced from Gil-Cruz (1999)

of tremor would probably be classified as "spasmodic" tremor in the original terminology of Jaggar (1920).

We next briefly review the development of quantitative models of long-period events and tremor beginning from the 1950s, and again focusing most on the time since the 1970s.

Volcanic tremor: early quantification

Omer (1950) provided one early quantitative model for the source mechanism of volcanic tremor, attributing tremor observations at Kīlauea (Finch 1949) (Fig. 10a, b) to a path effect: the reverberation of near-surface strata excited into motion by magma moving though subsurface feeding conduits. Shima (1958) and Kubotera (1974) instead proposed that a peaked tremor spectrum at Mount Aso (Sassa 1935) (Fig. 10c) was a result of free oscillations of a spherical magma chamber, while Shimozuru (1961) considered the longitudinal resonance of a cylindrical magma column. Steinberg and Steinberg (1975) attributed tremor to pulsating "flow crises" of gas in volcanic vents undergoing the transition from subsonic to supersonic flow. However, these early models did not adequately quantify the driving force of the fluid or predict the elastic radiation from the source region (Chouet 1981). More critically, these models required implausibly large dimensions for the resonating cavities, as pointed out by Ferrazzini and Aki (1987). For example, Kubotera (1974) determined the source of 3.5–7 s period tremor at Aso (Fig. 10c) to be a resonating spherical magma chamber of 2–4 km radius.

Crack propagation source model of Aki

A rigorous quantitative and (early) computational treatment of volcanic tremor was given by Aki et al. (1977), who proposed a mechanism for volcanic tremor at Kīlauea consisting of the sudden extension of dry and fluid-filled tensile cracks (Fig. 11a). Two scenarios were proposed: (1) the jerky extension and propagation of a single crack; (2) the random jerky openings of narrow channels connecting a chain of pre-existing cracks. This simplified two-dimensional model considered both the driving excitation and crack geometry appropriate for magma transport, but the fluid did not support acoustic waves and merely acted as a passive cushion to the motion of the crack wall. Near-field and far-field displacements computed by finite-difference calculations replicated the general properties of the observed tremor. A key parameter in the formulation of Aki et al. (1977) was the *crack stiffness*, defined:

$$C = \frac{bL}{\mu d}, \tag{1}$$

where b is the bulk modulus of the fluid in the crack, L is the crack length, μ is the elastic shear modulus, and d is the aperture of crack opening. The single-crack model (scenario 1, above) was rejected because the growing crack length predicted a significant increase in tremor period, inconsistent with the observations at Kīlauea. Aki et al.'s (1977) scenario 2 was further developed for deep tremor occurring at 30–50 km beneath Kīlauea by Aki and Koyanagi (1981). They defined a measure of tremor amplitude related to the magma flux known as *reduced displacement*:

$$\text{reduced displacement} = \frac{A}{2\sqrt{2}} r = A_{\text{r.m.s.}} r, \tag{2}$$

where A is the peak-to-peak amplitude of ground motion ($\frac{A}{2\sqrt{2}} = A_{\text{r.m.s.}}$, the root-mean-square amplitude), and r is the source-to-receiver distance. Measurement of the reduced

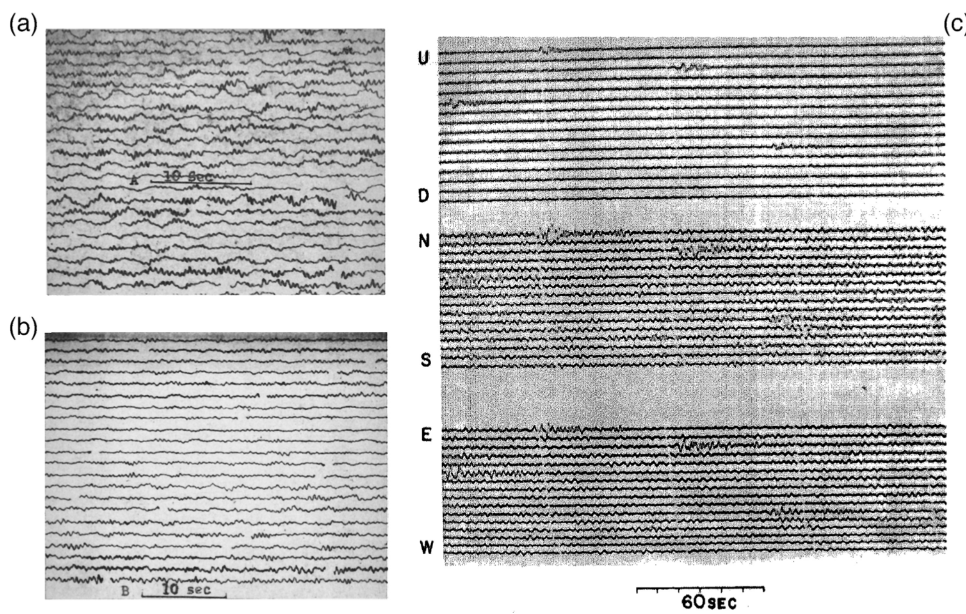


Fig. 10 Continuous observatory waveforms of volcanic tremor that informed early ideas at (**a**, **b**) Kīlauea (Finch 1949) and (**c**) Aso (Kubotera 1974). (**a**, **b**) Recorded on a Bosch-Omori seismograph ~ 3.4 km from Halema'uma'u. (**a**) "March 21, 1921. This is typical of the record when lava is high in Halemaumau. Some microseisms present. Record magnified 3 times" (Finch 1949). (**b**) "May 8, 1924. Record when no lava is visible in Halemaumau but underground movement of lava [sic] in the Puna Rift probable" (Finch 1949). Figures reproduced from Finch (1949) and Kubotera (1974)

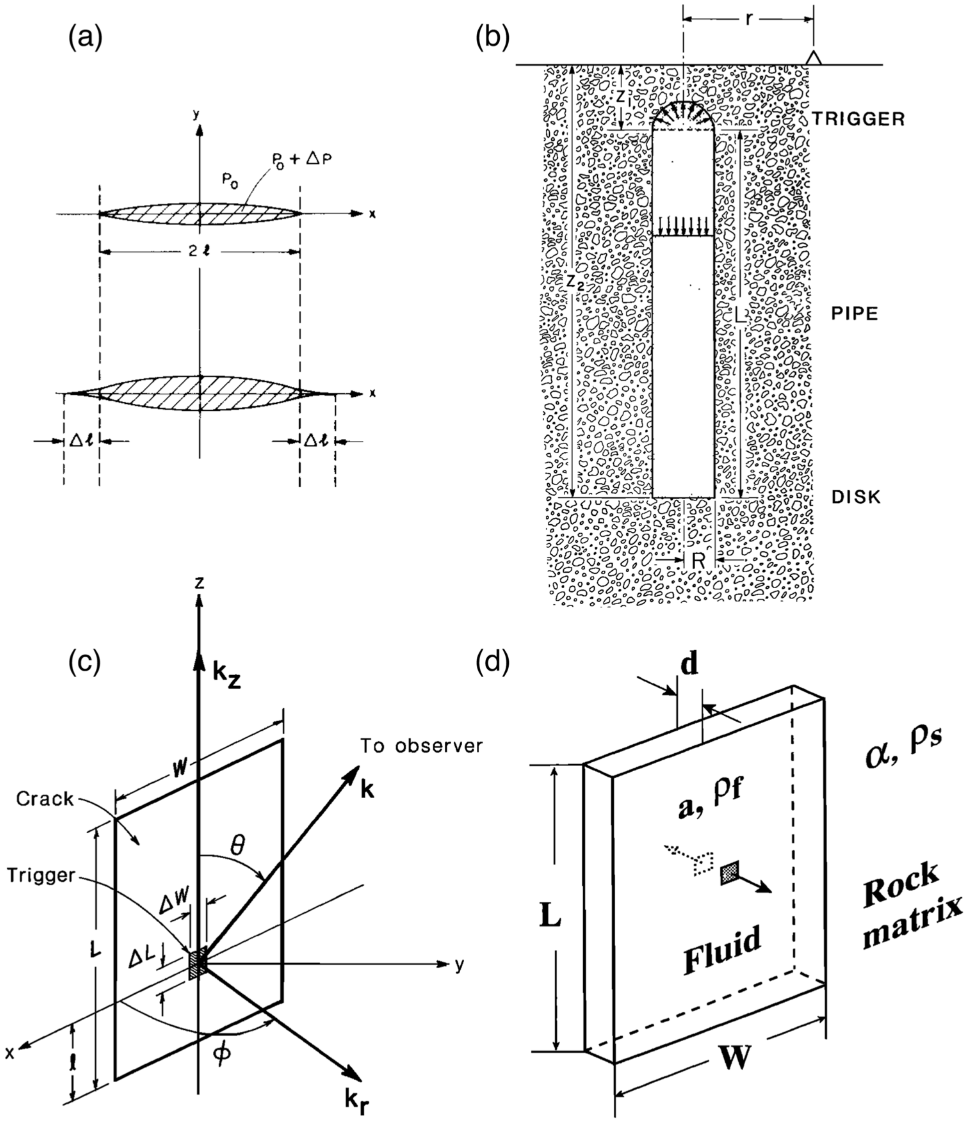


Fig. 11 Evolution of fluid-driven source models. (**a**) Aki et al. (1977) fluid-driven crack models. In this model, the fluid did not support acoustic waves and merely acted as a passive cushion to the motion of the crack wall. (**b**) Chouet (1985) fluid-filled conduit pipe model. The source is composed of a "trigger," a "resonator," and a "radiator"; in this case, the cylindrical conduit resonator produced acoustic resonance organ pipe modes. (**c**) Chouet (1988) resonating fluid-driven crack model. This numerical formulation produced slow solid–fluid interface waves or "crack waves," permitting observed seismicity with long-period (LP, 0.5–5 Hz) frequencies to be explained by a modest-sized compact resonating cavity. (**d**) Kumagai and Chouet (2000) formulation for investigating attenuation in the fluid-filled crack model. For explanation of symbols in each case, the reader is referred to the original references. (**a**) Reproduced from Aki et al. (1977), (**b**) reproduced from Chouet (1985), (**c**) reproduced from Chouet (1988), (**d**) reproduced from Kumagai and Chouet (2000)

displacement as a function of time implied a magma flow rate an order of magnitude lower than that implied by field observations of erupted lava effusion rates reported by Swanson (1972). Thus, Aki and Koyanagi (1981) concluded that most magma transport in the lithosphere takes place aseismically, with only particularly strong barriers to flow acting as seismic sources.

Later, Chouet (1981) further developed the crack model of Aki et al. (1977), calculating near-field and surface displacements for a single crack extension while accounting for interaction with the free-surface and near-surface velocity structure. The effects of varying the structure of the elastic media, source depth, and bulk modulus of the fluid in the crack were explored. This model was expanded into three-dimensions and further described in Chouet (1982, 1983). However, these models still assumed no active participation of the fluid. The fluid could not transmit acoustic waves, and the dynamics of the fluid were not considered in detail. Consequently, the spectral peaks obtained by these models were too weak and too broad, and the long duration of observed LP signals could not be reproduced (Chouet 1988).

Fluid resonance models and "crack waves"

The 1980–1986 unrest and eruptions of Mount St. Helens provided new digital observations of LP events and tremor (Fehler and Chouet 1982; Fehler 1983), rejuvenating interest in LP and tremor models in which the fluid plays an active role. Lawrence and Qamar (1979) and Ferrick et al. (1982) proposed a mechanism involving volcanic fluids analogous to the water-hammer effect in a cavity connecting a magma chamber to the surface. This model consisted of resonance of a conduit in response to unsteady flow conditions (i.e., "fluid transients", resulting from an abrupt disturbance to a fluid system initially at steady state) (Ferrick et al. 1982). These studies were motivated by seismic observations of "icequakes" in glaciers and seismic events originating from a malfunctioning power plant that resembled volcanic LP events.

Chouet (1985) also recognized the importance of the fluid in sustaining resonance, and interpreted individual LP events as the impulse response of a tremor-generating system. Accordingly, he proposed a conceptual system consisting of a "trigger," a "resonator," and a "radiator"; in this case, a hemispherical trigger, overlying a cylindrical conduit resonator (with "organ pipe" modes), terminated at the base by a circular radiator (Fig. 11b). Chouet (1985) proposed that the trigger mechanism was the rapid exsolution of gases from the fluid phase during magma ascent, or flashing of a subsurficial layer of phreatic water to steam due to shallow magma intrusion. An LP event thus corresponded to a single triggering of the system, while continuous tremor would result from continuous triggering. Thus, the goal of understanding the complex source mechanism of volcanic tremor was superseded by the more tractable task of understanding individual LP events.

Chouet and Julian (1985) and Chouet (1986, 1988) further developed the crack models initiated by Aki et al. (1977) and Chouet (1981, 1982, 1983), now allowing the fluid to transmit acoustic energy (Fig. 11c). These models were formulated using the equations of elastodynamics in the elastic solid, as well as conservation of momentum and equations of continuity for the fluid. These fluid-filled crack models were applied to non-double couple earthquakes observed near Long Valley Caldera between 1978 and 1983, hydrofracture events used in hydrocarbon extraction (Bame and Fehler 1986), and volcanic LPs and tremor. The most significant feature of these models was the presence of an interface wave propagating through the fluid and reflecting back and forth at the crack tips. The velocity of this "crack wave" is slower than the acoustic velocity of the fluid at all wavelengths, and is inversely dispersive (i.e., velocity decreases as wavelength increases). The properties of the crack wave are analogous to those of tube waves propagating in a fluid-filled borehole (Biot 1952). However, unlike the tube wave, as the wavelength increases to infinity, the velocity of the crack wave approaches zero in inverse proportion to the square root of wavelength (Ferrazzini and Aki 1987). In the short wavelength limit, the crack wave reduces to the Stoneley wave propagating along a fluid–solid interface (Stoneley 1926; Ferrazzini and Aki 1987).

Ferrazzini and Aki (1987) found analytic expressions of the crack waves by considering normal modes in a fluid layer between two homogeneous half-spaces, producing dispersion relations in harmony with the numerical results of Chouet and Julian (1985) and Chouet (1986). These studies showed that "slow waves" or "crack waves" could produce long-period elastic radiation from only a modest-sized resonating cavity. For instance, Kubotera (1974) had previously determined the source of 3.5–7 s period tremor at Mount Aso (Fig. 10c) to be a resonating spherical magma chamber of 2–4 km radius. By considering crack waves, Ferrazzini and Aki (1987) and Chouet (1988) could model this same tremor signal as resulting from a modest-sized magma body 0.5-m thick and 0.5-km long.

Analysis of the radiation properties from the resonating crack by Chouet (1988) demonstrated the stability of the dominant period in the far-field, while the frequency and width of this spectral peak was a strong function of the crack stiffness and trigger amplitude, area, and location. The crack stiffness (Eq. 1) affects the dispersion characteristics and therefore the resonance frequencies of the crack, while the frequency and duration of the signals are also affected by the impedance contrast between solid and fluid:

$$Z = \frac{\rho_s \alpha}{\rho_f a}, \tag{3}$$

where ρ_s and ρ_f are the density of the elastic solid and fluid, respectively, α is the P-wave velocity of the elastic solid, and a is the sound speed of the fluid in the crack (Chouet 1988). The duration of the LP signal is also related to the viscous damping loss at the fluid–solid boundary:

$$F = \frac{12\eta L}{\rho_f d^2 \alpha}, \tag{4}$$

where η is the viscosity of the fluid and L and d are the crack length and aperture (see Eq. 1) (Chouet 1988). Accordingly, the LP coda contains information on the attenuation properties of fluids in the crack source volume. However, as formulated by Chouet (1988), the crack model accounts for radiation and viscous drag losses only. Intrinsic losses due to dissipation mechanisms within the fluid must be treated separately, and were the focus of follow-up work that examined attenuation in a fluid-filled crack.

Attenuation in volcanic fluid-filled cracks

Attenuation in a fluid-filled crack model (Fig. 11d) was investigated by Kumagai and Chouet (1999, 2000, 2001) and Morrissey and Chouet (2001). The Sompi autoregressive signal analysis method enabled estimates of the quality factor Q of observed LP waveforms (Kumazawa et al. 1990; Nakano et al. 1998). The resultant observed Q is composed of two components:

$$Q^{-1} = Q_r^{-1} + Q_i^{-1}, \tag{5}$$

where Q_r^{-1} and Q_i^{-1} are the radiation and intrinsic losses, respectively. The radiation attenuation Q_r^{-1} is a function of the resonator geometry, as well as the sound speed and density of the fluid, and can be evaluated using the fluid-filled crack model (Kumagai and Chouet 1999, 2000, 2001; Morrissey and Chouet 2001). In contrast, the intrinsic attenuation Q_i^{-1} corresponds to intrinsic losses in the fluid, for example, viscous, thermal, and acoustic damping. Consequently, calculation of Q_i^{-1} requires knowledge of the thermodynamic equations of state for multiphase fluids (e.g., Kieffer 1977; Commander and Prosperetti 1989; Temkin and Dobbins 1966).

Kumagai and Chouet (2000, 2001) evaluated Q_i^{-1} and Q_r^{-1} for various gas–gas mixtures, ash–gas mixtures, and liquid–gas mixtures. They found that Q_i^{-1} was negligible compared to Q_r^{-1} for gas–gas mixtures, but that Q_i^{-1} could be important, for example, in bubbly liquids and in dusty and misty gases under certain bubble-size and particle-size conditions. Kumagai and Chouet (2000, 2001) also noted that the high observed Q (low attenuation) for long-lasting LP codas observed at several volcanoes could be explained by the high Q values of dusty and misty gases with small (~1 μm) particles. This highlighted the importance of these fluids (dusty and misty gases) in generating LP events.

Trigger mechanisms of LP seismicity

The utility of the fluid-driven LP source models reviewed above was restricted to a quantification of the crack resonance and properties of the fluids. These numerical and analytic formulations did not address the excitation (trigger) mechanism of the LP events or tremor. For example, in the computational formulation of Chouet (1986), the spatiotemporal properties of the pressure transient triggering the crack resonance are parameterized as kinematic conditions (e.g., an arbitrary step function in pressure applied to a small patch of the crack wall). Quantifying the physics of the trigger or driving mechanism of LP seismicity, including individual LP events (discrete impulse) and tremor (sustained), remains a work in progress. A wide variety of trigger mechanisms have been proposed (Chouet and Matoza 2013; and references therein), including those ultimately arising from self-sustained oscillations (e.g., Julian 1994; Balmforth et al. 2005; Rust et al. 2008; Dunham and Ogden 2012; De Lauro et al. 2011; Lyons et al. 2013; Takeo 2021), magmatic-hydrothermal interactions (e.g., Latter 1981; Havskov et al. 1983; Chouet 1985, 1996a; Leet 1988; Almendros et al. 2001; Kumagai et al. 2002b; Nakano et al. 2003; Nakano and Kumagai 2005; Lin et al. 2005; Ohminato 2006; Petersen and McNutt 2007; Cusano et al. 2008; Waite et al. 2008; Nakamichi et al. 2009; Matoza and Chouet 2010; Alparone et al. 2010; Arciniega-Ceballos et al. 2012; De Lauro et al. 2012; Cannata et al. 2012; Maeda et al. 2013; Jousset et al. 2013; Syahbana et al. 2014; Matoza et al. 2015; Kato et al. 2015; Caudron et al. 2015; Rodgers et al. 2015b; Padrón et al. 2015; Sgattoni et al. 2016; Jolly et al. 2017a; Park et al. 2019; D'Auria et al. 2011, 2019; Dawson and Chouet 2019; Gresse et al. 2021; Butcher et al. 2021), magmatic degassing (e.g., Chouet and Shaw 1991; Kawakatsu et al. 1992; Neuberg et al. 1994; Benoit and McNutt 1997; Gil Cruz and Chouet 1997; Garcés et al. 1998; Hagerty et al. 2000; Ripepe et al. 2001; Falsaperla et al. 2002; Chouet et al. 2003; Rowe et al. 2004; Molina et al. 2004; Ruiz et al. 2006; Lesage et al. 2006; Saccorotti et al. 2007; Patanè et al. 2008; Arciniega-Ceballos et al. 2008; Johnson et al. 2008b; Palo et al. 2009; Buurman and West 2010; Traversa et al. 2011; Davi et al. 2012; Lyons et al. 2016; Battaglia et al. 2016b), and brittle failure of melt (e.g., Webb and Dingwell 1990; Goto 1999; Neuberg et al. 2006; Tuffen et al. 2003; Tuffen and Dingwell 2005; De Angelis and Henton 2011; Thomas and Neuberg 2012). We refer the reader to the work by Chouet and Matoza (2013) for a review of this literature discussing the various proposed mechanisms and the observational and modeling constraints.

Advances in fluid-driven source models

The fundamental significance of solid–fluid interface waves as possible sources of LP seismicity was demonstrated in the work reviewed above. Work since about 2000 has further explored the parameter space of solid–fluid interface waves, including consideration of other source geometries and investigation of the potential for self-sustained oscillation in volcanic fluid transport systems (e.g., Krauklis and Krauklis 1998; Jousset et al. 2003, 2004; Balmforth et al. 2005; Rust et al. 2008; Dunham and Ogden 2012; Lipovsky and Dunham 2015). Some of these studies have referred to the crack wave as the "Krauklis wave" after Krauklis (1962) (e.g., Korneev 2008, 2011; Frehner 2014; Cao et al. 2021). A series of papers by Maeda and Kumagai (2013, 2017), Taguchi et al. (2018, 2021), and Torres et al. (2021) provide empirical formulations and generalized equations for the frequencies and quality factors of crack resonance, enabling more rapid evaluation of source parameters (forward modeling) compared to previous formulations involving a numerical solution based on finite differences (Chouet 1988).

Volcano-tectonic (VT) seismicity

Since 1919, major advances in the understanding of VT earthquakes and their relationship to magmatic processes accompanied improved observations of event rates, locations, focal mechanisms, and magnitudes; as well as the development of the plate tectonics paradigm and corresponding ideas on magma generation. By the middle of the twentieth century, it was recognized that some volcanic earthquakes, termed “ordinary volcanic earthquakes” by Minakami (1950), and ultimately defined as “volcano-tectonic” or VT earthquakes by Lahr et al. (1994), were nearly indistinguishable from tectonic earthquakes in that their waveforms contained high-frequency *P* and *S* phases. However, following from earlier ideas linking volcanic earthquakes to tensile failure (Jaggar 1920; Reid 1929) and accumulating observations of “great earthquakes” prior to eruptions (MacGregor 1949), a protracted debate about causality, i.e., whether these earthquakes triggered or were triggered by magmatism, arose during the mid-twentieth century. One camp, speaking from an experimental perspective, supported a “stress release hypothesis” that “ordinary volcanic earthquakes” resulted in pressure decreases of sufficient magnitude to generate melting and magma formation, with magma then ascending along the earthquake fault to erupt (e.g., Yoder 1952; Uffen 1959; Uffen and Jessop 1963). Alternatively, others argued, in line with modern understanding, that melting resulted in volume (and thus pressure) changes which strain the crust to trigger VT earthquakes (e.g., Kuno 1958; Minakami 1960; Matsuzawa 1953). For a review of contemporaneous understanding of whether and how purely tectonic earthquakes may trigger magmatic unrest, which is beyond the scope of this article, we refer the reader to Manga and Brodsky (2006).

Deployments of multiple seismic instruments on active volcanoes, and the consequent possibility of locating VT earthquakes, led to both advances in fundamental understanding of VTs and their use in forecasting. An early example involves the recognition that an (ultimately non-eruptive) 1933–37 Montserrat swarm was shallow (e.g., Perret 1939) and of volcanic origin. Powell (1938) showed that this swarm comprised an elongated cluster of epicenters, which MacGregor (1949) linked to the trend of soufrières (recently active vents) on the island, suggesting that the VT earthquakes were connected to a plane of crustal weakness or deep-seated fracture. Due to the expansion of the HVO seismic network, earthquakes preceding the 1942 eruption of Mauna Loa were observed to migrate towards and along the volcano’s southwest and northeast rift zones, leading to an accurate forecast of an eruption along the volcano’s flanks rather than at its summit. Additional observations of propagating VT earthquakes were made in 1960 at Kīlauea, and at Krafla in 1977, among others. Ultimately, the new observations demonstrating linear elongation of VT clusters and temporal propagation lent support to a hypothesized close spatial relationship between VT earthquakes and migrating magma, culminating in the “mesh hypothesis,” i.e., the idea that VT swarms occur on faults connecting magma-filled tension cracks as proposed by Hill (1977). As the station densities of seismic networks increased, it became possible to detect and locate smaller seismic events, and thus expanding quantities of VT seismicity, more precisely. Toda et al. (2002), for example, documented the locations of over 7000 VT earthquakes, accompanying dike intrusion under the Izu islands, Japan, which propagated to ultimately form an elongated cluster. Later, Ágústsdóttir et al., (2016) described the locations of over 30,000 propagating VT earthquakes accompanying dike intrusion and propagation at Bárðarbunga–Holuhraun, Iceland, in 2014–2015.

The expansion of multi-instrument volcano-seismic networks also allowed calculation of VT earthquake focal mechanisms, ultimately shifting early hypotheses that these earthquakes occurred as tensile failure to a double-couple failure model. An early attempt to distinguish “push/pull” mechanisms on the basis of first motions from a single seismometer was made by Sassa (1936). However, later work by Wada and Sudo (1967), using first motions from five stations recording earthquakes during the 1965–1966 eruption of Aso, Japan, documented mixed first motions for “tectonic-type” earthquakes, suggesting a double-couple component of motion. Additional evidence for a double-couple mechanism, dominated by strike slip and/or normal slip, for VT earthquakes emerged in the 1970s (Fig. 12) (Zobin 1971; Minakami 1974; Filson et al. 1973; Francis 1974; Ward and Gregersen 1973). This led to the development of a theoretical framework for dike mechanics (e.g., Pollard 1987; Rubin 1993, 1995 and references therein) in the 1990s that linked VT earthquakes to stresses induced in the host rock by ascending and/or pressurizing magma. This theoretical framework established a new paradigm that VT earthquakes were not necessarily spatially close to their source magma. This paradigm was enforced by observational evidence that induced stresses controlled VT seismicity (Barker and Malone 1991), and included observations of “distal VT” earthquakes preceding eruptions at Pinatubo in 1991 (Harlow et al. 1996), Mount Spurr in 1992 (Power et al. 1995), and Soufrière Hills and Unzen in 1995 (Umakoshi et al. 2008; Roman et al. 2008). Based on a comparative analysis of distal VT seismicity preceding 111 eruptions at 83 volcanoes, in addition to distal VT swarms preceding intrusions at 21 other volcanoes, White and McCausland (2016) made a case that distal VT seismicity was an important precursor for Earth’s most explosive eruptions. In this regard, they argued that distal VT seismicity preceded all VEI $\geq$ 5 explosive eruptions that they considered. They additionally argued that pre-eruptive distal VT seismicity originated on tectonic fault structures up to tens of kilometers laterally from the eruption site, rather than directly beneath the eruption site.

Beginning in the early 1990s, more comprehensive VT focal mechanism catalogs (Barker and Malone 1991; Aspinall et al. 1998) began to show that many VT earthquake focal mechanisms had *P*-axes approximately perpendicular to regional maximum compressive stresses, linking VTs to

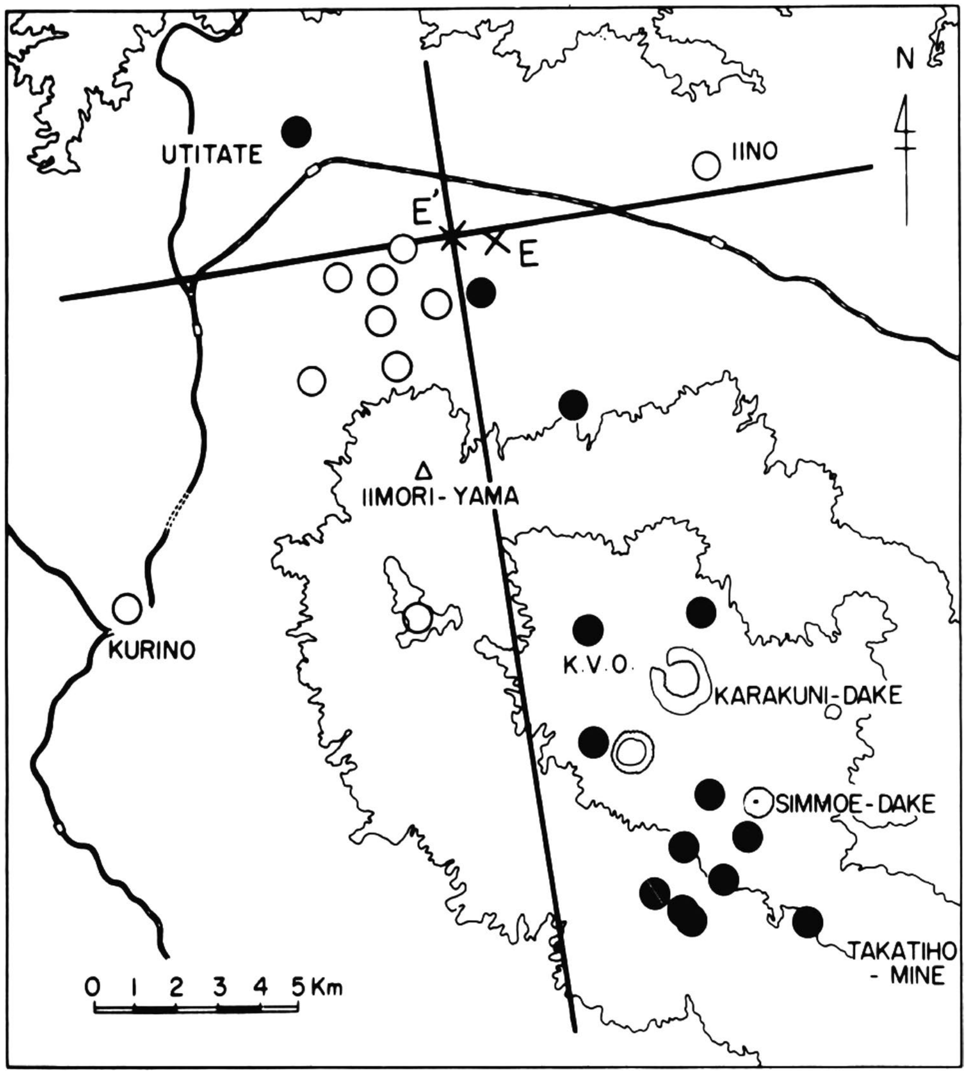


Fig. 12 Evidence from Minakami (1974) of double-couple mechanism for “A-type” volcanic earthquakes: geographical distribution of initial motions of the Kakuto caldera earthquake. E: epicenter, E′: corrected epicenter. Closed circles: upward initial motion. Open circles: downward initial motions. Figure reproduced from Minakami (1974)

stresses resulting from compression of the wall rock around a dike. Later work (Roman and Cashman 2006; Roman et al. 2021) demonstrated that this phenomenon corresponded to magmas with high bulk viscosities, and that VTs resulting from emplacement of low-viscosity magmas most likely represented induced stresses ahead of the dike tip rather than in the dike wall rock.

While VT earthquake magnitudes have not received nearly as much attention as their locations and focal mechanisms, several important observations about VT magnitudes have informed understanding of their mechanism and their utility in eruption forecasting. By the 1970s, it had been recognized that magnitudes of VT earthquakes were generally low $M<4$ (Zobin 1971; McNutt and Roman 2015), although exceptions have since been found (e.g., Yokoyama 2001; Nishimura et al. 2001; Wauthier et al. 2013). Early examinations of temporal patterns of precursory seismic energy release noted that, while some eruptions take place immediately following a decrease in seismic energy (Minakami 1961; Gorshkov 1960), a general increase in earthquake magnitude and cumulative energy release may be useful for forecasting eruption onset (Tokarev 1963; 1966). Furthermore, concepts from earthquake statistics led to recognition that a VT swarm’s *b*-value may be abnormally high compared to “tectonic” earthquake sequences, suggesting that increases in fluid pressure in the seismogenic volume around a magma pocket may play a role in driving VT seismicity (e.g., Mogi 1962; Suzuki 1959; Warren and Latham 1970).

Advances in eruption forecasting using seismicity: 1919 to 2022

In this section, we follow recent terminology for the distinction between an eruption “forecast” and “prediction” using seismicity. This is stated by National Academies of Sciences, Engineering, and Medicine (2017) as follows:

*"An eruption forecast is a probabilistic assessment of the likelihood and timing of volcanic activity. The forecast may also include information about the expected style of activity, the duration of an eruption, and the degree to which populations and infrastructure will be affected (*Sparks 2003*). A prediction, in contrast, is a deterministic statement about where, when, and how an eruption will occur, and a prediction will either be correct or incorrect".*

An ideal forecast of volcanic activity includes the location, timing, character, and magnitude of the potential eruption, and a quantitative estimate of the probability of each of these factors. While the potential of instrumental seismology to inform on eruption forecasting was recognized early-on, true forecasts (as opposed to hindcasts), were, and continue to be, limited by sparse and distant instrumentation (Hirn et al. 1987), lack of real-time telemetry, and inability to distinguish "false alarms" (e.g., MacGregor 1949; Macdonald 1954; Shepherd et al. 1971; Savage and Cockerham 1984; Moran et al. 2011). In addition, forecasting was hindered by the related issue of characterizing background seismicity levels (Wood 1974; Decker 1973). For example, the 1949 eruption of Mauna Loa was preceded by an increase in frequency of earthquakes, but the increase was not sufficiently great, or the pattern sufficiently definite, to make possible a forecast of the eruption (Macdonald 1954). Most early eruption forecasts focused on increases in the number of instrumentally detected discrete events, and indicated only that an eruption was likely, with little or no indication of the timing of the anticipated eruption. An early and rare example is of numerous tremor events that were recorded at Merapi Volcano, Indonesia, in January 1930, with an increase in their occurrence to 25 November. These, together with increased fumarole temperatures, were used to forecast the eruption (BNEIVS 1949; Van Padang 1933; Voight et al. 2000).

Forecasts based on seismic unrest became both more diverse, and more accurate, beginning in the 1960s with the advent of conceptual models, such as those of Minakami (1960, 1974) and the landmark recognition that an increase in B-type (Fig. 4) earthquakes could serve as a short-term precursor. As summarized by Girina (2013), after the first seismic station was installed in 1960 (Tokarev 1981) near Bezymianny Volcano, Russia, earthquake classification led to recognition of reliable patterns of seismic activity leading to phases of lava dome growth and explosive eruptions that could be used to forecast changes in the ongoing eruption (Gorelchik 2001) (Fig. 13). Similarly, the occurrence of volcanic tremor was used to formulate successful short-term (up to 7 days in advance) forecasts of eruptions at Ruapehu, New Zealand (Dibble 1969; Clacy 1972). A notable success occurred at Tolbachik, Russia, in 1976, where a dense network of rapidly deployed seismometers allowed forecasting not only of the time but also the location of the eruption approximately one week in advance (Tokarev 1978). The methods for rapid data analysis developed during these responses provided an early basis for the formalization of the Failure Forecast Method (FFM) later applied at Mount St. Helens in the 1980s (Voight 1988). Additional notable successes in short-term forecasting during these decades also occurred in Iceland at Heimaey in 1973 (Björnsson and Einarsson 1974) and Krafla in 1974 (Einarsson 2018).

Following these efforts between 1960 and 1980, a series of challenges and successes in seismicity-based eruption forecasting continued through the 1980s and 1990s. For example, the start of the 1980–1986 eruption of Mount St. Helens, USA, was successfully forecasted based on earthquakes that began approximately 2 months prior to the major explosive eruption on May 18, 1980 (Endo et al. 1981). Forecasts of subsequent eruptions at Mount St. Helens, based on seismic energy release in combination with observations of tilt and dome expansion, became increasingly

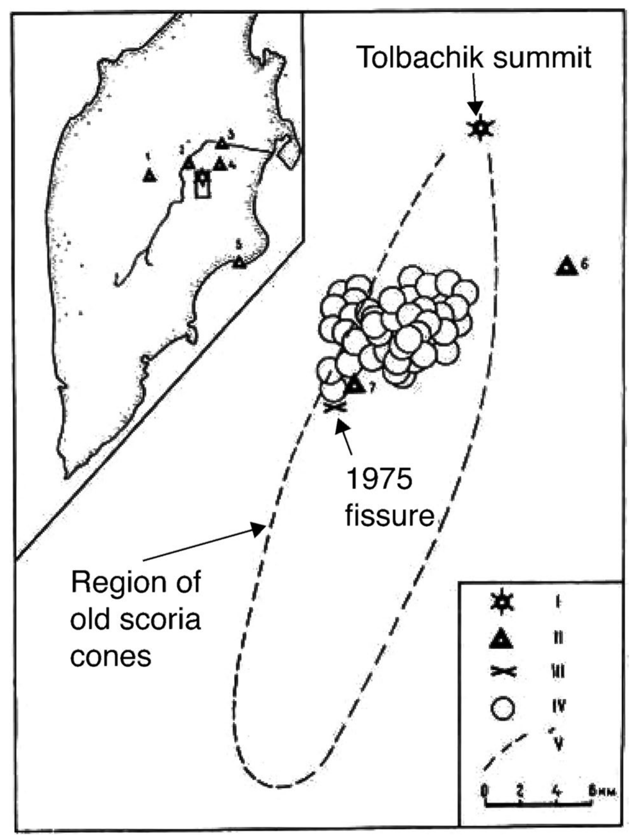


Fig. 13 Location and time of Tolbachik 1975 eruption forecast 3 days beforehand on the basis of epicenter locations. Diagram shows the location of Ploskii Tolbachik volcano (I), seismic stations (II), new crater (III), earthquake epicenters (IV), and boundaries of the area of old, well-preserved scoria cones of fissure eruptions (V). Figure reproduced from Tokarev (1978) (modified with annotation)

precise through the end of the eruptive phase in 1986 (e.g., Swanson et al. 1985; Malone et al. 1981, 1983). In Hawai‘i, an eruption forecast for Mauna Loa, based on 2 years of increased seismicity and deformation, was published in 1983 and was qualitatively correct regarding the timing of the next eruption (Decker et al. 1995). In contrast, although preceded by a detected increase in seismicity almost a year before its cataclysmic eruption in November 1985, Nevado del Ruiz, Colombia ultimately presented a complex situation that resulted in substantial issues with false alarms and communication of warnings (Voight 1996). An eruption at Redoubt Volcano, Alaska, beginning in 1989 led to the formalization of the Real-Time Seismic Amplitude Measurement (RSAM) and Spectral Seismic Amplitude Measurement (SSAM) approaches, which allowed rapid characterization of seismicity levels in multiple frequency bands for forecasting (Endo and Murray 1991; Stephens et al. 1994). RSAM and SSAM soon proved to be critical tools for seismic-based forecasting, as during the 1991 eruption of Pinatubo, Philippines (Pinatubo Volcano Observatory Team 1991; Cornelius and Voight 1994; Power et al. 1994). Regardless of these decades of progress, eruption forecasting based on seismicity continues to be a challenge, particularly for smaller eruptions (Cameron et al. 2018), phreatic eruptions (Roman et al. 2019; Kilgour et al. 2021), and occasional larger eruptions that appear to be preceded only by short and subtle seismic precursors (Johnson et al. 2010).

Volcano infrasound since the 1990s

As reviewed above, volcano acoustics research remained relatively dormant until a significant revival beginning in the 1990s. This revival was facilitated by factors including the availability of new infrasound instrumentation technology and computational capability to perform digital infrasound signal processing and noise discrimination (e.g., Garcés et al. 2003, Matoza et al. 2007; Christie and Campus 2010). Since that time, much progress has been made. The utility of infrasound technology in volcano monitoring is now firmly established, and infrasonic systems are increasingly being implemented as a volcano monitoring tool worldwide. Reviews of various aspects of volcano acoustics are provided in the work by Johnson and Ripepe (2011), Fee and Matoza (2013), Garces et al. (2013), McNutt et al. (2015), Allstadt et al. (2018), Matoza and Fee (2018), Matoza et al. (2019a), Marchetti et al. (2019), Taisne et al. (2019), Ripepe and Marchetti (2019), De Angelis et al. (2019), and Johnson (2019). We here refer the reader to these reviews and do not attempt a comprehensive review of volcano infrasound research since the 1990s, instead limiting the discussion to a highlight of major signals studied and trends in the research progression.

Explosive eruptions are the most obvious volcanic sources producing easily observable high-amplitude infrasound signals. Infrasound signals from explosive eruptions may propagate in the atmosphere over distances of thousands of kilometers under favorable conditions, enabling regional (ranges 15–250 km) and remote (ranges > 250 km) ground-based infrasonic monitoring (e.g., Matoza et al. 2007; Fee et al. 2010a,b; Matoza et al. 2011a; 2018; Ripepe et al. 2018; Lyons et al. 2020). The utility and limitations of infrasound for globally detecting and cataloging Earth's volcanism is presently under investigation (Dabrowa et al. 2011; Matoza et al. 2017; de Negri et al. 2022). We refer the reader to reviews by Matoza et al. (2019a) and Taisne et al. (2019) for discussions on progress and outstanding challenges in developing global eruption notification and acoustic early warning using regional and global infrasound networks for the time covering up to about 2017 (the time of writing of those reviews).

Even considering only explosive eruption sources, a wide variety of infrasound signals have been observed (Fig. 14), capturing the underlying variety of physical explosion mechanisms and mass flux source-time functions (Johnson 2003; Matoza et al. 2014b; Fee et al. 2017). These signals range between (1) discrete explosion waves with relatively simple waveforms lasting from several to tens of seconds (Fig. 14a, b) (e.g., Firstov and Kravchenko 1996; Ripepe and Marchetti 2002; Johnson 2003; Marchetti et al. 2009, 2013), and (2) sustained, broadband, infrasonic tremor signals lasting from minutes to hours (Fig. 14e, f) (e.g., Vergniolle and Caplan-Auerbach 2006; Matoza et al. 2009a; Fee et al. 2010b; Caplan-Auerbach et al. 2010). The latter signals (2) resemble an infrasonic form of jet noise from flight vehicles and have thus been termed volcanic jet noise (Matoza et al. 2009a; 2013a; Fee et al. 2013a, b; McKee et al. 2017). Intermediate signal types (Fig. 14c–e) consisting of a short-duration impulsive explosion waveform followed by sustained jetting (of variable duration) are commonly observed and appear to be a characteristic feature and behavior of intermediate-composition (andesitic) low-level explosive volcanism, as well as strombolian eruptions (e.g., Ishihara 1985; Ripepe et al. 1996, 2007; Johnson 2007, Johnson et al. 2008b; Sahetapy-Engel et al. 2008; Marchetti et al. 2009; Yokoo et al. 2013; Lopez et al. 2013; Firstov et al. 2013; Taddeucci et al. 2014).

The acoustics of all of these complex explosive eruption sources are currently being investigated through dedicated field studies (e.g., Jolly et al. 2017b; Iezzi et al. 2019a; Wallace et al. 2020; Taddeucci et al. 2021; Matoza et al. 2022a) and in the laboratory (e.g., Médici et al. 2014; Médici and Waite 2016; Swanson et al. 2018; Peña Fernández et al. 2020), as well as through numerical modeling (e.g., Taddeucci et al. 2014; Cerminara et al. 2016; Brogi et al. 2018; Watson et al. 2021). The effects of, especially near-source, topography and atmospheric propagation on shaping the

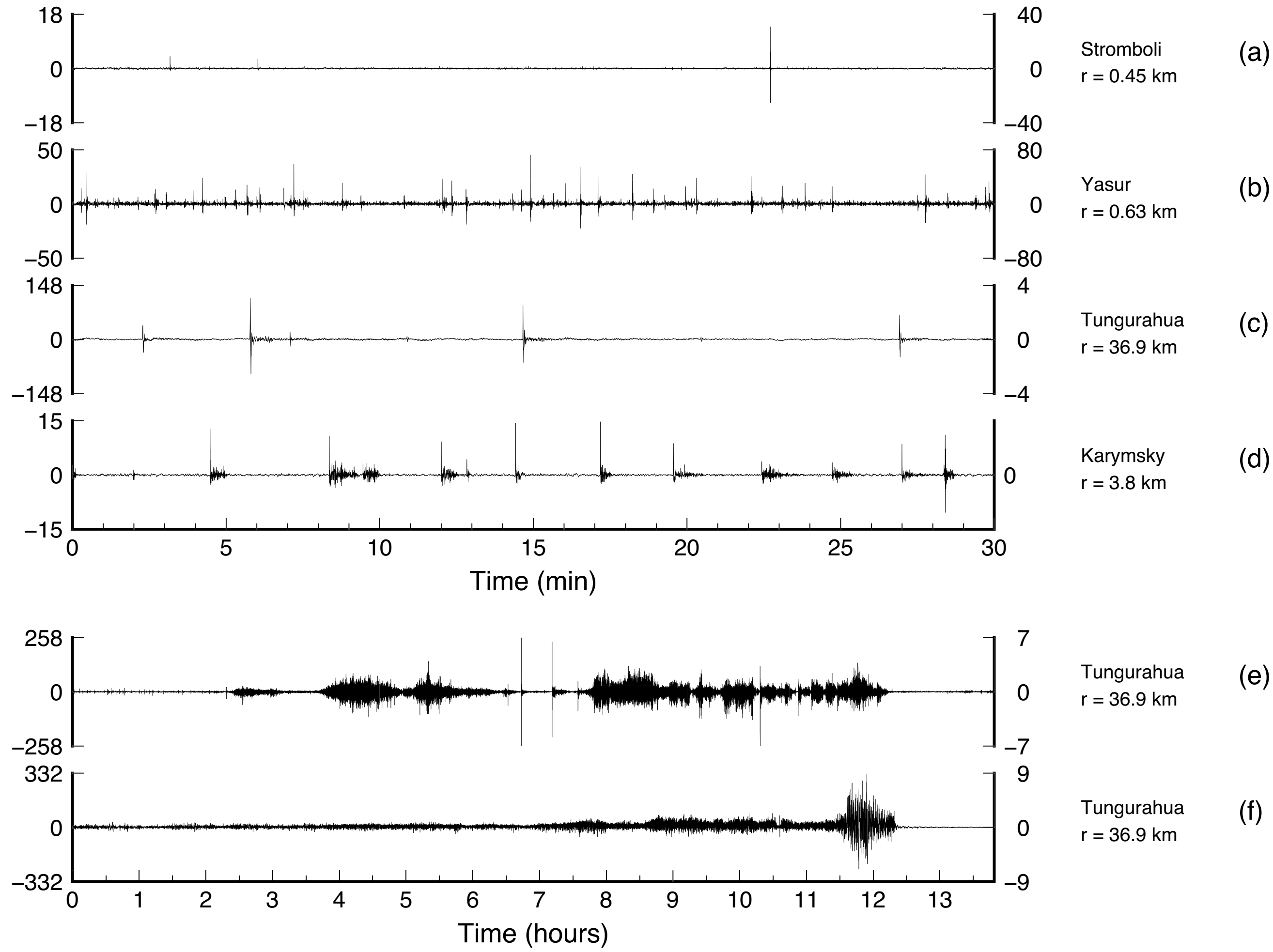


Fig. 14 Example infrasonic pressure waveforms associated with different explosive eruptive styles at selected volcanoes. The top four traces (**a**–**d**) are of 30-min duration, while the lower two traces (**e**–**f**) are of 13.8-h duration. The right-hand labels indicate the volcano and recording distance (range) r [km]. In each case, the right-hand y-axis is the observed acoustic pressure amplitude at that range, while the left-hand y-axis is the amplitude corrected to a reference distance of 1 km from the source by assuming $1/r$ geometrical spreading for approximate comparison. (**a**) Typical strombolian explosions from Stromboli, Italy (Ripepe and Marchetti 2002). (**b**) High-rate repetitive "strombolian" explosions at Yasur, Vanuatu (Matoza et al. 2022a). (**c**) "Strombolian" explosions from Tungurahua, Ecuador, with codas containing harmonic tremor (Fee et al. 2010b). (**d**) Complex explosion waveforms from Karymsky, Kamchatka, with an initial sharp compressional onset followed by short-duration jetting (Lopez et al. 2013; Matoza et al. 2014b) or "blow-off" (Firstov et al. 2013). (**e**) sub-Plinian eruption from Tungurahua, Ecuador, consisting of multiple sustained sequences of volcanic jet noise interspersed with discrete explosions (Matoza et al. 2009a). (**f**) Sub-Plinian to Plinian eruption at Tungurahua: a sustained volcanic jet noise signal with more gradually evolving signal properties

observed signals are being investigated for a variety of scales from local (< 15 km) to remote (> 250 km) (e.g., Fee and Garces 2007; Matoza et al. 2009b; 2011a; Kim and Lees 2011, 2014; Johnson et al. 2012; Assink et al. 2012, 2013; Fee et al. 2013b; Lacanna and Ripepe 2013; Lacanna et al. 2014; Lonzaga et al. 2015; Ortiz et al. 2018, 2021; Sabatini et al. 2019; Iezzi et al. 2019a,b; Waxler and Assink 2019; Ishii et al. 2020; Martire et al. 2022; Maher et al. 2021). Non-linearity in source and propagation is also being examined in observations and by numerical simulation (Marchetti et al. 2013; Fee et al. 2013a; Maher et al. 2020, 2022; Watson et al. 2021). Complex explosive processes and signals from eruptions in partially water-submerged (marine, or crater lake) settings have been documented (e.g., Green et al. 2013; Lyons et al. 2019; Fee et al. 2020; Park et al. 2021; Rose and Matoza 2021).

As reviewed above, infrasonic source resonance signatures in the long-period band, e.g., infrasonic harmonic tremor (e.g., Sakai et al. 1996; Garces et al. 1998; Lyons et al 2013) and seismo-acoustic expressions of LP events (Yamasato 1998; Johnson et al. 2008b; Matoza et al. 2009b) were noted early and observations of these signals have progressively expanded to cover a range of magma compositions and eruption styles. Effusive eruptions, lava flows, lava flowing in tubes, and convecting lava lakes have been observed to produce near-continuous broadband and/or harmonic infrasound (Garces et al. 2003; Cannata et al. 2009; Matoza et al. 2010; Fee et al. 2010c; Ulivieri et al. 2013; Patrick et al. 2016, 2019; Spina et al. 2017; Valade et al. 2018; Barrière et al. 2018; Lyons et al. 2021) as well as seismicity (Harris et al. 2005; Jones et al. 2006). Eruptions sites at Kīlauea, for example, have produced prodigious broadband and harmonic infrasonic tremor associated with effusive and degassing activity and occasional short-duration explosions (e.g., Garces et al. 2003; Matoza et al. 2010; Fee et al. 2010c; Patrick et al. 2016, 2019; Lyons et al. 2021). Overpressurized degassing (gas puffing) also has its own infrasonic signature (Ripepe et al. 2002, 2007; Harris and Ripepe 2007b). Cavities and gas-filled conduits above degassing magma appear to significantly influence the infrasonic signature (Matoza et al. 2010; Fee et al. 2010c; Goto and Johnson 2011; Richardson et al. 2014; Spina et al. 2015; Johnson et al. 2018). In a series of papers, Buckingham and Garcés (1996), Garcés and McNutt (1997), and Garcés (2000) developed a canonical model, deriving an analytic solution for the upgoing sound field (i.e., the airborne Green's function) from a resonant magma or gas-filled conduit. In these conduit resonance models (Buckingham and Garcés 1996), the geometrical idealization of the conduit was similar to that of Chouet (1985), with the exception that the "radiator" was then a diaphragm-like motion of the magma surface radiating sound into the atmosphere. This formulation demonstrated that high-frequency (> 50 Hz) acoustic energy is propagated preferentially in a narrow beam of sound vertically above a conduit, while infrasonic frequencies (< 10 Hz) diffract spherically from the conduit opening (vent), partially explaining why these frequencies are more readily recorded with ground-based sensors. The formulation of Garcés (2000) considered the resonant properties of a tube of fluid connected to the atmosphere with arbitrary variable cross-sectional area that may also be moving at high velocity relative to the sound speed of the flow (i.e., at a high Mach number). More recent work by Watson et al. (2019, 2020) has developed an analytic solution for the shallow crater resonance signature. Similarly to Garces (2000), the Watson et al. (2019, 2020) formulation is axisymmetric and permits a variable cross-sectional area with depth.

A variety of surficial mass movements have now been shown to generate infrasound (Allstadt et al. 2018, and references therein). For example, infrasound signals have been documented from lava dome collapse (Green and Neuberg 2005), pyroclastic flows (Yamasato 1997; Ripepe et al. 2009, 2010; Delle Donne et al. 2014), debris avalanches (Toney et al. 2021), rockfalls (Moran et al. 2008c; Johnson and Ronan 2015), lahars (Johnson and Palma 2015), and explosive blowout of gas-charged blocks impacting the ground (Oshima and Maekawa 2001). As reviewed by Allstadt et al., (2018), this represents significant potential in augmenting monitoring capability for hazardous surficial mass movements. However, much more work is required to quantify the seismo-acoustic source mechanisms from these sources (e.g., Moretti et al. 2012; Allstadt 2013; Farin et al. 2019; Coco et al. 2021; Brosch et al. 2021; Toney et al. 2021) and develop signal processing strategies to identify robustly the sometimes low-amplitude signals of surficial mass movements within realistic persistent and variable background noise (Matoza et al. 2013b) including from (but not limited to) background fluvial infrasound from drainages through which lahars may propagate (Sanderson et al. 2021), distant storms (microbaroms) (Landès et al. 2012), and, in coastal locations, surf infrasound (Garcés et al. 2006).

Future trends

In the first section of this review, we provided a brief snapshot of the current state of volcano seismology and acoustics (also termed seismo-acoustics). One hundred years of advances amounts to a vast amount of progress and changes in instrumentation, analysis and inversion methodologies, as well as in our quantitative understanding of seismic and acoustic sources in volcanic systems. Despite this progress, major questions remain regarding source processes, patterns of volcano-seismic unrest, internal volcanic structure, and the relationship between seismic unrest and volcanic processes. We refer the reader to two recent papers by Thelen et al. (2022) and Watson et al. (2022) published in the Bulletin of Volcanology Special Issue, "Looking Backwards and Forwards in Volcanology: A Collection of Perspectives on the Trajectory of a Science". These two short perspective papers summarize recent progress and look to the future of volcano seismology and acoustics, speculating on advances to come based on current trajectories and trends.

Conclusions

Over the past hundred years, volcano seismology and acoustics have advanced profoundly. By 1919, the basic recognizable components of these fields were already in place, with early instrumental waveform capture of volcanic seismicity and atmospheric pressure waves on seismographs and barographs demonstrating that geophysical monitoring could help track eruption progression and mitigate hazards. Technological advances during the past hundred years have seen the toolkit of volcano seismology advance from smoked paper drum records to robust continuous digital waveform streams, with increasingly sensitive and wider band instrumentation and denser networks capturing new signals and phenomena. The measurement of atmospheric pressure waves (acoustic-gravity waves) from explosive eruptions was a prominent feature of early volcano seismology and volcanology studies through 1919, but advances in volcano seismology rapidly outpaced those in atmospheric acoustics until about the 1990s. Infrasound technology has since become increasingly integrated into volcano-seismic monitoring operations. Regional (15–250 km distance) and remote (at > 250 km) infrasound detection and explosive eruption notification systems are now in operation and undergoing testing and refinement. Computational and quantitative volcano seismological approaches, beginning in the 1970s, have also led to major progressions in understanding of source mechanisms and their relation to volcanic processes. Geophysical monitoring is an essential component of societal resilience to volcanic hazards. Quantitative volcano seismology and acoustics are indispensable for robust volcano monitoring and should continue to provide progressively sharper insights into how volcanoes work.

Postscript: January 2022 eruption of Hunga, Tonga

While this manuscript was in a final editorial stage, the climactic eruption of Hunga volcano, Tonga occurred on 15 January 2022. This event produced atmospheric waves unprecedented in the modern geophysical record and has resulted in exceptional multi-technology observations of rarely captured physical phenomena (e.g., Amores et al. 2022; Astafyeva et al. 2022; Carr et al. 2022; Carvajal et al. 2022; Ern et al. 2022; Harding et al. 2022; Harrison 2022; Kubota et al. 2022; Kulichkov et al. 2022; Lin et al. 2022; Liu et al. 2022; Le et al. 2022; Matoza et al. 2022b; Omira et al. 2022; Otsuka 2022; Poli and Shapiro 2022; Ramírez-Herrera et al. 2022; Saito 2022; Schnepf et al. 2022; Themens et al. 2022; Vergoz et al. 2022; Wright et al. 2022; Yamazaki et al. 2022; Yuen et al. 2022). The prominent Lamb wave was observed propagating around the Earth for the same number of passages (four plus three antipodal) as the historic 1883 Krakatau eruption (Matoza et al. 2022b). As measured by the Lamb wave amplitudes, the Hunga explosion was comparable in size to that of the 1883 Krakatau eruption and over an order of magnitude greater than that of the 1980 Mount St. Helens eruption (Matoza et al. 2022b). As we have reviewed herein, the time from the 1883 eruption of Krakatau (Scott 1883; LeConte 1884; Strachey 1888) to the 2022 eruption of Hunga represents more than a century of remarkable advances in the instrumental recording, technology, analyses, and understanding of seismic and atmospheric waves produced by volcanic eruptions.

Acknowledgements R.S.M. acknowledges NSF grants EAR-1446543 and EAR-1847736. R.S.M. thanks Annie Platoff, UC Santa Barbara Library for assistance with tracking down references. We thank Bernard Chouet, two anonymous reviewers, and editor Andrew Harris for their helpful comments and additional perspective.

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