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INTRODUCTION TO VOLCANIC SEISMOLOGY

SECOND EDITION

VYACHESLAV M. ZOBIN

Introduction to Volcanic Seismology

Introduction to Volcanic Seismology

Second edition

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Elsevier
32 Jamestown Road, London NW1 7BY
225 Wyman Street, Waltham, MA 02451, USA

Second Edition 2012

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British Library Cataloguing-in-Publication Data
A catalogue record for this book is available from the British Library

Library of Congress Cataloging-in-Publication Data
A catalog record for this book is available from the Library of Congress

ISBN: 978-0-444-56375-0

For information on all Elsevier publications visit our website at elsevierdirect.com

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Preface to the Second Edition

The first edition of my book was published 8 years ago. I was glad to recognize that it became a textbook for many universities. Giving my lectures in Volcanic Seismology to graduate students of our Colima University and at other universities of the world, I have begun to realize that many improvements are required within my book, especially regarding the theoretical basis. Many new results have appeared in the study of volcanic earthquakes. The continuous eruption that occurred at our Volcán de Colima during 1998–2011 produced an invaluable collection of thousands of seismic records of volcanic explosions, pyroclastic flows, lahars, together with simultaneous video records. Analysis of this database essentially served to enhance my understanding of the nature of volcanic earthquakes and their relationship with the eruption process.

This book contains data from volcanic earthquakes obtained during a period of about 100 years. The quality and completeness of these data are different depending on the period of observation. However, I preferred to include these characteristics of volcanic earthquakes because they were unique and not repetitive. Once I received the editorial comments to a paper where the editor recommended that I rewrite the manuscript using the seismic records of a larger number of seismic stations around the volcano. It was a good idea but at that time we had only four stations during the eruption, the eruption was over already, and it was impossible to install additional stations going "back to the past." The data included in this book are certainly heterogeneous, but they are unique and describe many important volcanic events. They provide an understanding of the history of our monitoring of volcanic eruptions.

Eight years after the first edition is not such a long interval of time. But the number of publications in *Volcanic Seismology* during the first decade of the XXI century was comparable with the number of publications in this field during the total time of observations of volcanoes prior to this. So I realized that the new, updated, and significantly expanded edition of the book is necessary. I have added new chapters about the theoretical and experimental fundamentals of Volcanic Seismology; about long-period and very-long-period seismic signals, observed at volcanoes; about micro-earthquake swarms associated with explosive and effusive activity at volcanoes; about acoustic signals associated with volcanic processes; about failed eruptions within volcanic regions. I significantly rewrote and expanded the chapters about the seismic signals associated with explosive events and with pyroclastic flows, rockfalls, and lahars.

I hope that this book will serve successfully for a new generation of volcano-seismologists and be a good friend to all interested in the nature of volcanic processes.

Vyacheslav M. Zobin
Colima, México, 12 August 2011

Preface to the First Edition

I just returned from my first field seismological practice in Altai. It was October of 1964, and I was a 20-year-old student of the Faculty of Geology and Geophysics at Novosibirsk State University in Siberia, Russia. Our guys told me that one man from Kamchatka wants to see me. It was Dr. Pavel Tokarev, he had just been appointed chief of a new department for seismic monitoring of volcanic activity in the Institute of Volcanology and recruiting personnel for his department. He invited me to go to Kamchatka the following summer for my second field seismological practice at volcanoes and proposed the following theme for my year's essay: "Volcanic earthquakes and their relationship with volcanic eruptions."

I went to our university library and found a few papers concerning this problem. Among them was a famous paper by Takeshi Minakami (1960) about the prediction of volcanic eruptions, the papers by Georgy Gorshkov and by Pavel Tokarev about seismic study of Klyuchevskoy and Bezymianny volcanoes in Kamchatka, and that was all. No more. Thus, I understood that this field of science was quite open for researchers, and I was lucky to enter to this "terra incognita."

I wrote this essay, presented it at the University student conference, received as a prize a beautiful book "*Elementary Seismology*" by Charles Richter, and next summer went to my Kamchatka practice. I met a wonderful land. I climbed Avachinsky volcano, I had adventures at the flank cones of Klyuchevskoy volcano, and I decided to work there, in Kamchatka. Later, I wrote my M.S. thesis on the seismic wave attenuation beneath Klyuchevskoy volcano and was then accepted in the staff of the Institute of Volcanology in Petropavlovsk-Kamchatky. Two months later, in October 1966, I had my first eruption of the Piip flank cone of Klyuchevskoy volcano; my first publication (together with P. Tokarev and V. Shirokov) about seismic signals associated with this eruption appeared in 1968. I began my life in volcanic seismology.

I spent about 30 years in Kamchatka. In 1971, the General Assembly of the International Union of Geophysics and Geodesy took place in Moscow. It was a great opportunity to meet many outstanding volcanologists of the world. Takeshi Minakami and Izumi Yokoyama became my friends; I was happy to receive Christmas cards from them for many years. In 1975, the great Tolbachik fissure eruption began in Kamchatka. A study of seismic signals related to this eruption strongly improved our knowledge of the nature of volcanic earthquakes. I wrote a book about the source properties of volcanic earthquakes, it was published in Russian, in Moscow, in 1979. It was one of the first books devoted to volcanic seismology.

I have lived in México for the last 9 years. I have prepared and delivered twice the courses in *Volcanic Seismology* for graduate students at the scientific center CICESE in Baja California, México, and recently decided to write a book. You have it now.

Vyacheslav M. Zobin
Colima, México, 5 March 2003

In the photos here I am with my teachers and my students. One photo was taken in Moscow during the 1971 IUGG Assembly. The second in the first row (from the left) is Georgy Gorshkov, he was elected as President of IAVCEI that year; the fourth is Takeshi Minakami. I am the fourth in the third row. The last in the fourth row is Izumi Yokoyama.

In other photos, I am with the graduate students during my visits to Alaska Volcano Observatory, Fairbanks, USA, in September 2010 and to King Saud University, Riyadh, Saudi Arabia, in May 2011. It was a good opportunity to give them the revised version of my lectures on Volcanic Seismology.

1 Introduction

*The subject of **volcanic seismology** is not only the most beautiful and spectacular, but also the most difficult to study of all the subjects seismologists have encountered on Earth. This is because seismic sources in volcanoes involve dynamic motion of gas, fluid and solid, and propagation paths in volcanoes are usually extremely heterogeneous, anisotropic, and absorptive, with irregular topographies and interfaces including cracks of all scales and orientations. Thus, volcanic seismology is the most challenging to seismologists requiring ingenuity in designing experiments and interpreting observations.*

Keiiti Aki, 1982. From the Preface to the Collection of Papers "Volcanic Seismology," IAVCEI Proceedings in Volcanology, Vol. 3, Springer-Verlag, Berlin.

Volcanic seismology is a science about seismic signals originating from volcanoes and associated with volcanic activity. The study of the origin of these signals, their spatial–temporal distributions, their relationships with volcanic processes, and using them as an instrument to investigate the volcano deep structure and to predict a volcanic eruption together create the subject of *volcanic seismology*.

1.1 Terms and Definitions

The subject of Volcanic Seismology lies in the interaction between volcanic and seismo-tectonic processes. Therefore, it is necessary to define the main terms of volcanology, tectonics, and seismology at the very beginning of the book.

1.1.1 Volcanic Terms

A *volcano* is a vent or chimney, which connects a reservoir of molten matter known as *magma*, in the depths of the crust of the Earth, with the surface of the Earth. The material ejected through the vent frequently accumulates around the opening, building up a *volcanic edifice*. The material ejected consists of liquid lava and broken fragments of partially or completely solidified rock (pyroclastic debris), as well as great quantities of gases. The gases are the motivating force and the most important factor in volcanic action (Bullard, 1963).

Figure 1.1 shows the main elements of a volcano. A *volcanic cone* is the result of the accumulation of ejected material around the vent. A *crater* is the surface connection of the *volcanic conduit* through which the ejected material reaches the

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00001-3
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Figure 1.1 The main elements of volcanic edifice.

surface. The crater may be situated on the summit of the cone (*summit crater*) or within the *flank cone*, situated on a slope of the central volcanic cone (*flank crater*). An enormous crater representing a large depression in the truncated summit of volcano is called a *caldera*.

Beneath a volcano, we have the magma feeding system of the volcano that consists of a deep *magma reservoir*, or *chamber* that is a reservoir of magma in the shallow part of lithosphere from which volcanic material is derived; an intermediate *magma storage*, a place between the magma chamber and the Earth's surface, where magma may be collected before an eruption; and the *feeding conduit*, connecting the magma collectors with the surface. The surface appearance of magma is called *lava* (Bullard, 1963). Volcanoes are widely distributed on the surface of the Earth and on the ocean floor (Figure 1.2). Volcanic belts go along *subduction zones* as well as along *oceanic ridges*. Simkin and Siebert (1994) have noted more than 1,300 active volcanoes with more than 5,500 dated eruptions.

A *volcanic eruption* is a process of material transport from the Earth's depths to the surface. Figures 1.3–1.5 show the typical surface manifestation of volcanic features during the activity of *basaltic*, *andesitic*, and *dacitic* volcanoes. According to their origin within the volcanic edifice, the volcanic eruptions may be of *central* (or summit) type with an eruption of the summit crater, *flank* (or lateral) type, with an eruption of the flank crater, or *fissure* type, when the eruption takes place from an elongated fissure, rather than from a central vent.

Figure 1.2 Major lithospheric plates and global distribution of some active and dormant volcanoes. Courtesy of U.S.G.S.

Figure 1.3 Surface manifestation of volcanic features during the activity of basaltic volcanoes. (A) The 1975 fissure eruption of New Tolbachik volcanoes, Kamchatka. Eruption from the new cone and its lava flow. (B) Lava fountain and spatter cone on Kilauea Volcano, Hawaii. (C) The November 1989 lava flow on Etna volcano.
(A) Photograph by V. Zobin.
(B) Photograph by J.D. Griggs; courtesy of U.S.G.S.
(C) Photograph by A. Amantia; courtesy of INGV Catania.

Figure 1.4 Surface manifestation of volcanic features during the activity of andesitic volcanoes. (A) Block lava flow moving down the flank. Bezymianny volcano, Kamchatka. (B) Lava dome in the crater of Volcán de Colima. (C) Explosion at Kizimen volcano, Kamchatka.
(A) Photograph by P. Izbekov; courtesy of Alaska Volcano Observatory.
(B) Photograph by M. Bretón; courtesy of Colima Volcano Observatory.
(C) Photograph by E. Vlasov; courtesy of the Institute of Volcanology and Seismology.

Figure 1.5 Surface manifestation of volcanic features during the activity of dacitic volcanoes. (A) Lateral lava dome Showa Shinzan (Usu volcano, Hokkaido). (B) Surface deformation caused by the subsurface moving of dacitic lava. The 2000 lateral eruption of Usu volcano, Hokkaido. (C) Explosive activity at Karymsky volcano, Kamchatka. Photographs by V. Zobin.

Magnitude of eruption is described by *Volcanic Explosivity Index* (VEI) that combines the total volume of explosive products, the eruptive cloud height, descriptive terms, eruption type, duration, and so forth, and have the grades from 0 to 8 (Newhall and Self, 1982). Newhall and Self (1982) proposed the range of VEI for five main types of volcanic eruptions: Hawaiian (VEI 0−1); Strombolian (VEI 1−2); Vulcanian (VEI 2−4); Plinian (VEI 4−8); and Ultra-Plinian (VEI 5−8). The first two types of eruptions, Hawaiian and Strombolian, are common for *basaltic* volcanoes. Vulcanian eruptions may be observed mainly at *andesitic* and *dacitic* volcanoes. Plinian and Ultra-Plinian eruptions occur at *dacitic* and *rhyolitic* volcanoes. During an eruptive process, the type and style of activity and the nature and type of products can change.

1.1.2 Tectonic Terms

The movement of magma in the interior below the volcanic structure leads to the change in hydrostatic equilibrium and to the fracturing of the Earth material. Tectonic *faults* and *fractures* represent the breaks in rock material of different scales. We define *fractures* as the ruptures of length less than 100 m; greater ruptures would be *faults*. As a rule, the system of fractures forms around general faults and reflects the local rupturing of different parts of the fault. A type of movement along the fault plane characterizes tectonic faults. There are three main types of tectonic movements: normal faulting, thrust, and strike-slip. The *normal fault* and the *thrust* are characterized by a low angle of inclination with reference to a horizontal plane. With normal faulting, the hanging wall of the fault has been depressed, relative to the footwall; the thrust is a reverse fault. The *strike-slip* represents the horizontal movement along the vertical plane. Volcanoes usually lie along a tectonic fault or intersection of faults, usually within a *rift* structure where normal faulting is dominant (Table 1.1).

Table 1.1 General Characteristics of the Main Types of Volcanic Eruptions (Newhall and Self, 1982; Vergniolle and Mangan, 2000; Cioni et al., 2000)

Type of Eruption	Volume of Ejecta (m^3)	Column Height (km)	Type of Magma	Eruptive Style	Volcanic Features
Hawaiian	10^4−10^6	0.1−1	Basaltic	Fissure, central, flank	Lava flow, fire fountain
Strombolian	10^4−10^7	0.1−5	Basaltic	Central	Lava flow
Vulcanian	10^6−10^9	1−25	Andesitic, andesito-basaltic, dacitic	Central, flank	Lava flow, pyroclastic flow, lava dome
Plinian	10^8 to $>10^{12}$	10 to >25	Rhyolitic, trachytic, phonolitic	Central	Lava dome, pyroclastic flow
Ultra-Plinian	10^9 to $>10^{12}$	>25	Rhyolitic, trachytic	Central	

1.1.3 Seismological Terms

The rupturing along the fractures and faults as well as the eruptive activity produces the Earth's surface as *earthquakes*. The site of initial radiation of seismic waves in the interior is *earthquake focus*, or *hypocenter*; its projection to the surface is the *epicenter*.

The seismic waves are divided into *body waves* and *surface waves*. The body waves *P* (compressional waves with volumetric changes) and *S* (shear waves without volume change) radiate from the earthquake focus; the surface *Rayleigh* and *Love* waves are formed as a result of the interaction of body waves with the surface of the Earth. The record of seismic waves is a *seismogram* (Figure 1.6). The earthquake size is described by *magnitude M* (or *m*), the conventional value proportional to the earthquake energy, and by *seismic moment Mo*. The *focal mechanism* characterizes the stress system acting in the earthquake source zone, while the *stress drop* gives the value of stress release during the earthquake.

1.2 Subject of the Book

Volcanic seismology has not yet any plain theory; it is more a science of observation and cumulative experience. This book is an introduction to *Volcanic seismology*. It is devoted to the study of the relationships between volcanic earthquakes and eruptive process and of the nature of volcanic earthquakes.

The book consists of 19 chapters. Chapters 2 and 3 introduce the history and the topic of *volcanic seismology* and discuss the theoretical and the experimental models that were developed for the study of the origin of volcanic earthquakes.

The next four chapters give the theoretical basis for the occurrence of volcano-tectonic earthquakes and their swarms, and describe the case stories of volcano-tectonic activity associated with the eruptions at basaltic, andesitic, and dacitic volcanoes. For this book, the author selected the 40 cases of volcanic eruptions (or episodes) at

Figure 1.6 Seismogram of shallow local earthquake. Vertical component of short-period instrument.
Courtesy of Colima Volcano Observatory.

20 volcanoes that occurred all over the world from 1910 to 2005. This selection was argued by different reasons. The most important of them was the possibility of having a catalog of seismic events for special analysis; among others, the quality of seismic observations, and the historical significance of the eruption. Chapters 8–10 describe the general regularities of volcano-tectonic earthquake swarms, their participation in the eruptive process, their source properties, and the hazard of strong volcano-tectonic earthquakes.

Seven chapters are devoted to the description of eruption earthquakes. The author gives the theoretical basis for the occurrence of eruption earthquakes and describes the volcanic tremor, seismic signals associated with pyroclastic flows, rockfalls and lahars, explosion earthquakes, long-period and very-long-period seismic signals at volcanoes, micro-earthquake swarms, and acoustic events. The author discusses their relationship to the eruptive process.

The next two chapters discuss the mitigation of volcanic hazard. There is a brief discussion about the methodology of seismic monitoring of volcanic activity, and about the experience in forecasting of volcanic eruptions by seismic methods. The author also discusses the seismic activity in the regions of dormant volcanoes.

Acknowledgments

I thank my colleagues from Kamchatka, Russia, and Colima, México for collaboration and useful discussions. My co-authors in the investigations in volcanic seismology Valentina Gorelchik, Pavel Tokarev, Valeria Levina, Tonatiuh Domínguez, and Pavel Firstov helped me to understand better many aspects being discussed now in this book. Seismic data obtained by the Kamchatkan regional seismic network and the Colima seismic network RESCO gave me an experience in the study of seismic signals recorded at volcanoes. The catalogs of volcanic earthquakes provided by Abdullah Al-Amri (Harrat Lunayyir, Saudi Arabia); Stefano Gresta and Domenico Patané (Etna volcano, Italy); Zenon Jiménez (El Chichon volcano, México); Jun'ichi Miyamura (Usu volcano, Japan); Jennifer Nakata (Kilauea volcano, Hawaii); Kodo Umakoshi (Unzen volcano, Japan); Javier Lermo-Samaniego (Popocatépetl volcano, México); and Willy Aspinall (Soufriére Hills volcano, West Indies); as well as the Catalog of earthquakes of Pinatubo volcano (the Philippines) prepared by the U.S.G.S. team, the 1980 and 2004 catalogs of earthquakes of Mount St. Helens volcano (Cascades, U.S.A.) prepared by the Pacific Northwest Seismic Network, the Kamchatka Regional Earthquake Catalog, the Catalog of earthquakes of Icelandic Meteorological Office (Grimsvøtn volcano), and the catalog of Harvard CMT solutions were very important in the preparation of this book. I use in my book the seismic records and video images obtained at Colima Volcano Observatory by the teams headed by Gabriel Reyes and Mauricio Bretón, respectively. Anastasia Chebrova, Luca D'Auria, Susanna Falsaperla, Diego Gómez, Jun'ichi Miyamura, Enrique Guevara, Orlando Macedo, Stephan Malone, William Melson, Yasuaki Sudo, and Mae Yamamoto provided me with the seismic records of volcanic events. Raúl Arámbula, Miguel González, and Justo Orozco helped me with the data acquisition. The processing of the digital signals was performed using the program DEGTRA provided by Mario Ordaz and Seismo-Volcanalysis provided by Philippe Lesage. Izumi Yokoyama, Daisuke Shimozuru, and Viktor Ivanov sent the biographical information and photos of the founders of volcanic seismology.

My work was benefited by numerous comments made during the preparation of the papers that formed the basis for this book. Stefano Gresta, Kazuhiro Ishihara, Fred Klein,

Jim Luhr, Steve McNutt, Seth Moran, Alejandro Nava, among others, helped me to better understand many aspects of volcanic seismology. Nick Varley read the text, prepared for the first edition, and helped greatly in polishing my English. The comments of Andrea Cannata, Hiroyuki Kumagai, Robin Matoza, Silvio Di Angelo, Aldo Martuzano, Natalia Stalboun, Nick Varley, and Gregory Waite helped me improve the text prepared for the second edition.

I also thank all copyright holders, who allowed me to reproduce figures in my book: American Geophysical Union, Birkhäuser Publishers Ltd., Earthquake Research Institute of Tokyo University, Elsevier Science (Academic Press), Geological Society of London Publishing House, Institute of Volcanology and Seismology, Instituto de Geofísica, Istituto Nacionale di Geofisica, Seismological Society of America, Nature Publishing Group, Societá Geologica Italiana, Springer-Verlag and Tohoku University, John Wiley and Sons, University College Dublin, Volcanological Society of Japan and U.S. Geological Survey, Department of the Interior/USGS. I thank S. De Angelo, M. Iguchi, V. Ivanov, M. Kasahara, O. Macedo, S. Matsumura, H. Okada, P. Firstov, K. Tanaka, H. Watanabe, H. Yamasato, and I. Yokoyama for the kind permission to reproduce their published figures.

Finally, I want to say that I have prepared this book considering that the basic volcanological and seismological ideas and theories may be taken from special textbooks. So, I recommend to my readers some textbooks in seismology and volcanology:

Francis, P. (1998). Volcanoes. A Planetary Perspective. Clarendon Press, Oxford. 443 p.
Lay, T., Wallace, T.C. (1995). Modern Global Seismology. Academic Press, San Diego. 521 p.
Schmincke, H.-U. (2004). Volcanism. Springer-Verlag, Berlin. 324 p.
Stein, S., Wysession, M. (2003). An Introduction to Seismology, Earthquakes, and Earth Structure. Blackwell Publishing, Oxford. 498 p.

I recommend also the following review papers:

Chouet, B. (2003). Volcano seismology. Pure Appl. Geophys. 160, 739−788.
Kawakatsu, H., Yamamoto, M. (2007).Volcano seismology. Treatise Geophys. 4, 389−420.
Kumagai, H. (2009). Volcano seismic signals, source quantification of. In: Meyers, R.A. (Ed.), Encycl. Complexity and System Sci. Springer-Verlag, Berlin. pp. 9899−9932.
McNutt, S.R. (2000). Volcanic seismicity. In: Sigurdsson, H. (Ed.) Encyclopedia of Volcanoes. Academic Press, San Diego. pp. 1015−1034.
Newhall, C.G. (2007). Volcanology 101 for seismologists. Treatise Geophys. 4, 351−388.

Japanese readers can also use the book on Volcanic Seismology published in Japanese:

Nishimura, T., Iguchi, M. (2006). Volcanic Earthquakes and Tremor in Japan, Academic Press, Kyoto University, Kyoto. 242 p.

2 Seismicity at Volcanoes

2.1 History of Seismic Monitoring of Volcanic Activity

Earthquakes were associated with volcanic eruptions from ancient times. Pliny the Younger gave us the first scientific description of the eruption of Vesuvius in 79 A.D. He wrote about numerous earthquakes related to this eruption. For many years, people considered volcanoes as the main origin of earthquakes in general. In the classification of earthquakes that was prepared in the beginning of the twentieth century, earthquakes of volcanic origin were in the first place.

The volcano Vesuvius occupies a special place in the history of the study of seismic signals associated with volcanic activity. It was the first volcano whose earthquakes were mentioned in scientific literature, it was the first volcano to have a volcano observatory in 1847, and it was the first to be monitored using seismological equipment. The Palmieri electromagnetic seismograph, built in 1862, was the first seismic instrument to record volcanic seismicity. The seismograms were written on a telegraph tape. Later, in 1903, the Bosch-Omori seismograph was installed at Mont Pelée, Martinique, 1 year after the great eruption of 1902 (Ferrucci, 1995). Seismic observations at the Kilauea, Hawaii, volcano began in 1912 by H. Wood just after the foundation of Hawaiian Volcano Observatory (Wood, 1913, 1915; Macdonald, 1972).

As a science, *volcanic seismology* was born when the Japanese seismologist Fusakichi Omori (Box 2.1) began his investigations of seismic signals related to the 1910 eruptions of Usu and Asama volcanoes (Omori, 1912; 1911–1913) and the 1914 eruption of Sakurajima volcano (Omori, 1914, 1922). Omori defined the volcanic earthquake as "seismic disturbance, which is due to the direct action of the volcanic force, or one whose origin lies under, or in the immediate vicinity of, a volcano, whether active, dormant, or extinct" (Omori, 1912).

The three-component seismic station, with a magnification of 100 and the period of the pendulum about 4 s, was installed by Omori in July–August of 1910 at a distance of about 1–2 km from the craterlets of Usu. This station allowed him to record the earthquakes originating from the volcano and the volcanic micro-tremors whose period was about 0.5 s (Figure 2.1). The micro-tremors were observed only during volcanic activity. Omori demonstrated a good relation between the appearance of micro-tremors and explosions of the volcano.

In the same paper, Omori (1911–1913) noted also that many eruptions of Japanese volcanoes were preceded by numerous earthquakes. "In such a case," he wrote, "seismograph observations made at the vicinity of the center of volcanic

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00002-5
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Box 2.1

Prof. **Fusakichi Omori** (1868–1923) was a founder of Volcanic Seismology. He began his work in seismology with John Milne and became in 1896 the professor of Imperial (Tokyo) University and secretary of the Imperial Earthquake Investigation Committee. For decades, he was the natural leader in all Japanese investigations on volcanoes and earthquakes. Omori's instrumental observations of seismic events during the 1910 Usu, the 1910–1914 Asama and the 1914 Sakurajima eruptions showed the great significance of seismic monitoring of volcanic eruptions for the understanding of volcanic processes and forecasting eruptions. Prof. Omori left us the detailed descriptions of these seismic events; researchers still use his papers in volcanic seismology. His death was dramatic. D. Shimozuru wrote to me in 1999: "It is well known the dispute between Prof. Omori and Prof. Imamura about the possibility of an occurrence of a great earthquake in the Tokyo area in 1923. During his participation in the Congress at Australia, the great Kanto earthquake occurred on September 1, 1923. Prof. Omori was defeated. He was heartbroken during his return voyage from Australia, and soon after he passed away."

activity would give people a warning of the approaching outburst." Analysis of volcanic earthquakes recorded in the course of volcanic activity by Omori had shown the great potential of seismic observations. They became the main instrument for volcanic activity monitoring. Among the most important stages in the development of *volcanic seismology*, we can note the foundation of seismic observations at Asama Volcano Observatory on Honshu Island, Japan in 1934, and at Kamchatka Volcano Observatory in Klyuchi, Kamchatka, Soviet Union in 1946.

Asama Volcano Observatory was established by Takeshi Minakami (Box 2.2) in 1934, and from the very beginning it was equipped by continuously operated seismographs. The observations of volcanic and seismic activity at Asama volcano allowed Minakami to propose a classification of volcanic earthquakes into four

Figure 2.1 Seismogram of volcanic earthquakes (a, b, d) and volcanic tremor (continuous signal) recorded on August 7–8, 1910 at Sobets seismic station near Usu volcano, Hokkaido, Japan, with a magnification of 100; component WE.
From Omori, 1911–1913.

types, which is the basis for the study of volcanic earthquakes up to the present day. He obtained a good correlation between the long-term variation in the number of volcanic earthquakes and volcanic eruptions (Figure 2.2). The increase of volcanic B-type earthquakes before the volcanic outburst became the criteria for the development of the method of eruption forecasting (Minakami, 1960).

The continuous seismological observations at Kamchatka (Klyuchi) Volcano Observatory began in 1946. They were headed by Georgy Gorshkov and later by Pavel Tokarev. Gorshkov (Box 2.3) was the first to discover that S-waves of distant earthquakes greatly attenuate the volcano; it allowed him to estimate the depth of upper mantle magma chamber of Klyuchevskoy volcano (Gorshkov, 1956). The great eruption of Bezymianny volcano in 1955–1956 gave numerous data for the detailed analysis of the interrelation of seismic and volcanic events. This was the first complete history of a great eruption described in terms of seismic information (Gorshkov and Bogoyavlenskaya, 1965). The following analysis of the seismic activity related to the annual small eruptions of Bezymianny in 1957–1961 allowed Tokarev (Box 2.4) to develop a method of forecasting volcanic eruptions based on a change in energy release of the preceding seismic events (Figure 2.3).

A network of high-gain seismometers at Hawaii Volcano Observatory has operated on Kilauea volcano since the 1950s and until 1985 it consisted of 50 seismic stations (Klein et al., 1987). The seismic studies allowed the construction of a fine model of the earth's crust near the Hawaiian volcanoes. The basic model of Kilauea was

Box 2.2

Prof. **Takeshi Minakami** (1909–1985) graduated from Tokyo University and at the age of 25 became the first director of the first Japanese Volcano Observatory at Asama volcano. He initiated the seismological observations at Asama, and his researches of Asama volcanic earthquakes allowed him to elaborate their classification. He developed a methodology for the statistical forecastings of volcanic eruptions at Asama by seismic data. His 1960-year paper "Fundamental research for predicting volcanic eruptions" helped to turn a volcanic seismology into a science. Later T. Minakami studied seismic events associated with the eruptive activity of Usu, Miyake-sima, and Arenal volcanoes. As Professor of Earthquake Research Institute of Tokyo University he prepared a new generation of Japanese volcano-seismologists. D. Shimozuru wrote in 1971 that Prof. Minakami "obtained pioneering results, which led to great progress towards the better understanding of volcanic phenomena."

proposed (Eaton and Murata, 1960) with a magma reservoir at the depth of about 3–4 km beneath the caldera. The generalized histories of Kilauea earthquakes represent the detailed history of volcanic activity of Kilauea from 1970 (Klein et al., 1987). The regularities of inflation–deflation cycle of magma reservoir and the appearance of characteristic related seismic events combined with ground deformation monitoring allowed the development of the short-term forecasts of Hawaiian eruptions (Klein, 1984).

These basic investigations had become the foundation for the construction of volcanic seismology as a science. During the last 30 years of the twentieth century and the first decade of the current century, the detailed and complex studies of some eruptions (among them, New Tolbachik volcanoes, Kamchatka, 1975–1976; Mount St. Helens, Cascades, 1980, 2004–2005; Redoubt, Alaska, 1989–1990; Pinatubo, Philippines, 1991; Soufriére Hills, Montserrat, 1995–1999; Volcán de Colima, México, 1997–2010) strongly increased our understanding of the seismo-volcanic

Figure 2.2 Relation between eruptions and shallow earthquakes (B-type) of Asama, Honshu, Japan. A, B, and C indicate the calm, preeruptive, and explosive eruption stages of the volcano. *S* is the daily frequency of explosions; *E* is the kinetic energy of strong explosions; and *F* is the daily frequency of shallow earthquakes.
From Minakami, 1960.

process. It was shown that the source of the majority of A-type volcanic earthquakes is similar to the source of tectonic earthquakes (Zobin, 1970, 1972); it was shown that the records of explosive volcanic earthquakes can be interpreted as Lamb pulses excited by a nearly vertical single force that represents the counter force of the eruption (Kanamori and Given, 1983; Nishimura, 1998); it was shown that shallow earthquakes in volcanic areas often have mechanisms with isotropic components, indicating volume changes of either explosive or implosive polarity (Miller et al., 1998).

Many models have been proposed for explaining the nature of volcanic tremor (Steinberg and Steinberg, 1975; Aki et al., 1977; Schick, 1981; Ripepe and Gordeev, 1999) and long-period volcanic earthquakes (Chouet et al., 1994). And, also, explaining the development of methodology for volcanic eruption forecasting (Tokarev, 1978; Malone et al., 1981; Harlow et al., 1996; Reyes-Dávila and De la Cruz-Reyna,

Box 2.3

Dr. **Georgy Stepanovich Gorshkov** (1921–1975) entered volcanic seismology in the very beginning of his volcanological carrier. In 1948 he was appointed Director of Kamchatka Volcano Observatory in Klyuchi village. A seismic station was installed here in the end of 1946, and Gorshkov became the first who studied the seismic records of the Kamchatka volcanoes. Two of his seismological results were fundamental. He discovered the screening effect of magma on S waves of distant earthquakes and determined the depth of the magma chamber of Klyuchevskoy volcano in the upper mantle. He studied in detail the relationships between seismic energy release and volcanic activity associated with the great 1955–1956 eruption of Bezymianny volcano. Dr. Gorshkov worked in volcanic seismology only for about ten years. He was a petrologist, and later he worked mainly in this area of volcanology. His monographs "Bezymianny Volcano and the Peculiarities of its Last Eruption, 1955–1963" (1965; with G.E. Bogoyavlenskaya) and "Volcanism and Upper Mantle" (1969) became the important events in a volcanological science. At the end of his life, Georgy Gorshkov was elected as the President of the International Association of Volcanology and Chemistry of the Earth's Interior.

2002; etc.). The installing of broadband digital equipment at volcanoes gives a good perspective for better understanding the nature of seismic signals originating with volcanic activity (Chouet, 1996a).

2.2 Classification of Volcanic Earthquakes

The first observations of volcanic earthquakes showed that the seismic signals originating from volcanic activity had different waveforms. Studying the seismic

Box 2.4

Dr. **Pavel Ivanovich Tokarev** (1923–1993) graduated from Moscow University, Russia and began his scientific activity in the Laboratory of Volcanology in Moscow in the middle of the 50s. His first papers were published in 1958–1959 and were devoted to the study of the structure of the Kurile-Kamchatka subduction zone and the relationship between seismicity and volcanism in this zone. In 1958, he moved to Kamchatka and began to work at Kamchatka Volcano Observatory in the village of Klyuchi. Tokarev installed the first seismic network for the monitoring of Klyuchevskaya Group of volcanoes and began to develop the method of seismic forecasting of volcanic eruptions for Bezymianny volcano. His first publications in this field appeared in 1963–1971 and became classic. Later Tokarev headed the Department of Geophysical Monitoring of Volcanic Activity in the Institute of Volcanology in Petropavlovsk-Kamchatsky, and studied the seismic activity of Klyuchevskoy, Bezymianny, Karymsky, and New Tolbachik volcanoes. His forecasting of the 1975 fissure eruption of New Tolbachik volcanoes was published in regional newspaper a few days before the beginning of the eruption and allowed a group of volcanologists to travel to the site of forthcoming eruption one week before the event begun. Pavel Ivanovich founded a system of seismic monitoring for volcanic activity in Kamchatka and prepared a team of seismo-volcanologists who continue now his efforts in seismic studies of volcanoes.

signals during the 1910 Usu eruption, Omori (1911–1913) wrote about "occurrence besides the proper volcanic earthquakes, of well-defined small quick unfelt vibrations, which may be termed micro-tremors" (see Figure 2.1). The first and still very popular classification of volcanic earthquakes was proposed by Minakami (1960, 1974a). He divided them into four types (Figure 2.4) according to the location of their foci, their relationship to the eruptions, and the nature of the earthquake motion.

Figure 2.3 Variations of the cumulative strain release [in $erg^{1/2}$] recorded at Bezymianny volcano, Kamchatka in 1958–1961. Arrows show the eruptions. The curves indicate an exponential increase in seismic energy release before an eruption.
From Tokarev, 1963.

Figure 2.4 An illustration of Minakami's classification of volcanic earthquakes. From Minakami, 1974a.

A-type. These are earthquakes originated from the bases of volcanoes or from the depths of about 1−20 km. These earthquakes take place previous to and during the first stage of eruptive activity and occur in swarms. They are generally less than 6 in magnitude. The nature of earthquake motions cannot be distinguished from those of general shallow earthquakes (tectonic earthquakes). The *P*-phase and *S*-phase of seismic waves also are clearly defined.

B-type. The hypocenters are limited to an area of about 1 km in radius around the active crater. The hypocenters of these earthquakes are shallower than those of the A-type volcanic earthquakes. These earthquakes take place in swarms in the shallow part ranging from the earth's surface to depths of several hundred meters. Their magnitudes are generally extremely small. The surface waves are predominating and the *S*-phase is not clear.

Explosion earthquakes. These are earthquakes, which accompany individual explosive eruptions. The amplitude of these earthquakes is related to the magnitude of the explosive eruptions. The earthquake motions show an exceeding predominance of longer wavelength as comparable with those of the A-type volcanic and tectonic

earthquakes; the initial motion is "push" in every direction. On the seismograms of explosion earthquakes, we often find disturbances caused by air shocks. The hypocenters are situated beneath the active crater floor.

Volcanic pulsation or continuous volcanic micro-tremors. The main part of these vibrations consists of the surface waves. Volcanic tremor has a form of an irregular sinusoid of rather long duration compared with earthquakes of the same amplitude.

Volcanic earthquakes of all four types can participate in eruptive activity during the same eruption. Figure 2.5 illustrates the seismic history of the 1966 flank eruption of Klyuchevskoy volcano, Kamchatka (Tokarev et al., 1968). Prior to the eruption, volcanic earthquakes of A- and B-types were observed (bars 5). An eruptive fissure opened after the cessation of earthquake sequence. With the beginning of the eruption, volcanic tremor (curve 3) started simultaneously with explosion earthquakes (curve 2). Volcanic tremor stopped just before the second swarm of B-type earthquakes (curve 4) but explosion earthquakes continued up to the end of the eruption.

Minakami's classification stays basic and popular up to today. Of course, for every volcano, seismologists use their own classification with more detailed description of every subtype of earthquakes (Tokarev, 1966; Latter, 1979; Malone, 1983; Lahr et al., 1994; Harlow et al., 1996; Ratdomopurbo and Poupinet, 2000; McNutt, 1996, 2000a).

Majority of the seismic signals defined by these authors were observed during the 1998–2011 unrest at andesitic Volcán de Colima, México. Figure 2.6 shows broadband seismic records of these volcanic earthquakes. Video monitoring allowed identification of the nature of majority of them. The seismic records of volcano-tectonic

Figure 2.5 Seismic history of the 1966 flank eruption of Klyuchevskoy volcano, Kamchatka. The curves 1–5 show the variations: (1) in the cumulative strain release measured in $[10^{-8}]$ erg$^{1/2}$ for A- and B-type earthquakes; (2) in the energy of air waves (A.W.) produced by explosions and measured in $[10^{-15}]$ erg/day; (3) in the energy of volcanic tremor (V.T.) measured in $[10^{-14}]$ erg/day; (4) in the number of B-type earthquakes of the second swarm; (5) in the number of A- and B-type earthquakes of magnitude $M > 0$. N is the number of earthquakes.
From Tokarev et al., 1968.

Figure 2.6 Types of seismic signals (velocity, vertical component) recorded during the 1998−2011 unrest at andesitic Volcán de Colima (left side) and corresponding Fourier spectra (right side). VT, volcano-tectonic earthquakes; EXP, seismic signals associated with magmatic explosive events; HYB, hybrid signals associated mainly with white-cloud degassing; LP, long-period signal recorded at different stages of unrest; ME, micro-earthquake signals, forming together with short hybrid signals the micro-earthquake swarms, associated with extrusive and explosive activity; TR, volcanic tremor recorded at different stages of unrest; PF, signals associated with pyroclastic flows or rockfalls of incandescent rocks; and LAH, signals associated with lahars. All seismic signals were recorded by broadband seismic station at a distance of 4 km from the crater and corrected for instrument response.
Courtesy of Colima Volcano Observatory.

earthquakes (VT) were characterized by clear arrivals of *P* and *S* waves. The next two types of seismic records (EXP and HYB) were associated with Vulcanian explosive activity. EXP events with low-frequency initial pulse were associated with dark magmatic explosions while HYB events with high-frequency initial pulse were mainly associated with white-cloud degassing. Rare LP signals were associated with small explosive events or were observed without any direct relation to eruptive activity. The swarms of short-duration and low-amplitude ME events were associated with the magma movement before lava extrusion and during different stages of strong explosive activity. The volcanic tremor (TR) signals were observed at different stages of eruptive activity without any regularity. The signals marked as PF and LAH were associated with the movement of pyroclastic flows and rockfalls of incandescent rocks, and lahars, respectively.

In this book, the author considers two groups of volcanic earthquakes that include all these types. The first group includes earthquakes generated by stresses, which arise during magma migration in the Earth's crust. They may be caused by shear or tensile failure and occur during magma movement from the depth to the Earth's surface through conduits, dikes, and so on, producing rupture at the tip of magma body and within adjacent fault systems. The author calls them *volcano-tectonic earthquakes*; they correspond to A- and B-types of Figure 2.4. Their study is a study of physics of magma movement to the surface. These earthquakes are very important for forecasting of volcanic eruptions.

The second group includes earthquakes originated due to magmatic activity within the volcanic conduit; during volcanic explosions; associated with the lava dome growth, the movement of lava flows, pyroclastic flows, and lahars. The author calls them *eruption earthquakes*. Their study is a study of physics of volcanic eruptions.

2.3 Sequences of Volcanic Earthquakes

Majority of volcanic earthquakes occur in sequences. Mogi (1963) described three main types of earthquake sequences (Figure 2.7) as main shock—aftershocks sequence (Type 1); foreshocks—main shock—aftershocks sequence (Type 2); and swarm (Type 3).

Earthquake swarms are generally defined as a sequence of events closely clustered in time and space without a single outstanding shock. In practice, the difference in magnitude between the largest and second largest event of a swarm is 0.5 magnitude units or less (McNutt, 2000a). If this difference is greater, we have the main shock—aftershock sequence.

2.3.1 Sequences of Volcano-Tectonic Earthquakes

Volcano-tectonic earthquakes may occur in all three types of sequences. The two first types of sequences are rare, and they occur prior to flank or fissure eruptions. The swarm sequences are the most common for volcanic earthquakes.

Figure 2.7 Three main types of successive occurrences of elastic shocks accompanying fractures that were obtained from experimental destruction of material and their relations to their structures and the applied stresses.
From Mogi, 1963.

The type 1 sequence was observed prior to the 30 September 1996 flank eruption of subglacial volcano Grimsvøtn in Iceland (Zobin, 1999; Konstantinou et al., 2000). The main shock with magnitude *mb* 5.2 was the first in this earthquake sequence (Figure 2.8A); the second largest earthquake occurred after about 3 h with magnitude *mb* 4.3 (ISC, 1996).

Seismic activity developed during three stages (Zobin, 1999). The first stage (first 19 h) included the main shock with magnitude *mb* 5.2 (*Ms* 5.4) and its aftershocks along the northern slope of the Bárdarbunga volcano, situated about 20 km to NW from Grimsvøtn (Figure 2.8B). Then during the next 17 h, the seismic foci marked a line connecting the two volcanoes, Bárdarbunga and Grimsvøtn. This stage culminated in the opening of a fissure and the beginning of the eruption. The third stage was observed during the eruption (from midnight of 30 September until 7 October), and included continuous seismic activity along the northern slope of the Bárdarbunga volcano as well as more western distributed epicenters, which together had clustered along the ring-fault epicentral zone.

The type 2 sequence was observed before the January 1996 subaqueous phreatomagmatic eruption in Akademia Nauk caldera (Karymsky Lake), Kamchatka. This eruption occurred after 4800 years of dormancy and resulted from the injection of juvenile basaltic magmas through a fissure (Fedotov, 1998). The preliminary seismic activity (Zobin and Levina, 1998) was represented by a foreshock–main shock–aftershock sequence with a main shock with magnitude *Ms* 6.7 (*Mw* 7.1) (Figure 2.9).

Figure 2.8 Variations in 6-h numbers (A) and epicenter distribution (B) of the earthquakes associated with the September 30, 1996 subglacial flank eruption of Grimsvøtn volcano, Iceland. The calderas of volcanoes Bárdarbunga and Grimsvøtn are shown. The diamonds are epicenters; the star shows the epicenter of the main shock. From Zobin, 1999.

The seismic foci of foreshocks migrated gradually from the forthcoming site of eruption to the south-west up to the *Mw* 7.1 earthquake epicenter. The aftershocks spread to the south-west and north-east from the main shock, reaching the site of eruption; the total area occupied by aftershocks was of about 50 km × 20 km. The seismic foci were distributed within the depth ranging from 0 to 60 km. The eruption in Karymsky Lake began during a sharp decrease in the number of aftershocks.

The third type of earthquake sequence, and the most frequent type, is a swarm sequence. It occurs roughly with numerous earthquakes (Figure 2.10). Figure 2.11 shows the spatial-temporal characteristics of earthquake swarm associated with the September 1977 deflation of Krafla volcano, Iceland (Brandsdottir and Einarsson, 1979). A small basaltic eruption broke out on a 0.9-km-long fissure the northern rim of caldera and continued for about 5 h. The earthquake activity

Figure 2.9 Variation in hourly numbers (A) and epicenter distribution (B) of the earthquakes associated with the January 2, 1996 subaqueous phreatomagmatic eruption in Akademia Nauk caldera (Karymsky Lake), Kamchatka. The star shows the site of eruption in Akademia Nauk (A.N.) caldera; dots are the epicenters of foreshocks and crosses are the epicenters of aftershocks; large white cross shows the epicenter of the main *Ms* 6.7 shock. The proposed time interval of the beginning of the eruption is shown.
Data from Kamchatka Regional Earthquake Catalog.

and volcanic tremor began about 2 h before the eruption and accompanied the eruption.

The first very small earthquakes were located in the central part of the caldera; later the earthquake epicenters migrated southward with a rate of about 0.5 m/s. The seismic activity culminated just after ceasing of lava eruption with eight earthquakes with magnitude 3 and greater (Figure 2.11). It is interesting to note that the 1138-m-deep well, situated along the way of earthquake migration, throw out the glowing pumice at the time of earthquake migration. It indicated that the earthquake foci migration was directly associated with the magma lateral migration.

Swarms may occur before the eruption, during the eruption, and after the eruption of volcano. Very frequently the appearance of a swarm near a volcano may indicate a forthcoming eruption of the volcano. Two main groups of volcano-tectonic earthquakes swarms can be distinguished according to their character of occurrence through time, their spatial position, and their relationship with volcanic activity (Gorelchik and Zobin, 1971).

Group 1. These swarms precede generally the summit eruptions of andesitic and dacitic volcanoes. The swarms consist of shallow volcanic earthquakes, the epicenters of which lie at a distance of 3–5 km from the crater and the foci of which are

Figure 2.10 Earthquake swarm at Volcán de Colima, México, November 1997. The seismogram was recorded at a distance of 4 km from the crater. Minute marks are shown. Courtesy of Colima Volcano Observatory.

distributed within the volcanic edifice at the depth range between 0 and 5 km above sea level (a.s.l.). Depending on the deep structure of the feeding zones at andesitic volcanoes, earthquake having focal depths of 10−15 km are feasible. The maximum magnitudes of earthquakes may reach 5−6. The swarm duration can be from 7−10 days to 3−5 weeks. During these swarms, the total energy, frequency, and

Figure 2.11 The temporal variations of earthquake magnitudes in the swarm that occurred during the September 1977 deflation of Krafla volcano, Iceland (A) and the earthquake swarm epicenters (B). Symbol I marks the initial part of eruption; F marks the fin of eruption. It is clearly seen a group of equal-magnitude earthquakes in (A). Dots are epicenters located with standard error less than 0.5 km; circles mark the epicenters of worse location.
From Brandsdottir and Einarsson, 1979.

maximum energy of earthquakes increase continuously and evenly, reaching their maximum values at the moment of eruption.

Group 2. These earthquake swarms precede generally flank eruptions at basaltic and andesitic volcanoes. Very shallow earthquakes predominate in the swarms. The epicenters of the earthquakes lie at a distance of 3–10 km from the place of formation of a new crater, and the foci are located at the depths of 0–5 km. The swarms preceding the flank eruptions at strato-volcanoes last for 1–10 days, their maximum earthquake magnitudes reach 5–7. In these swarms the number of events and their maximum energy sharply increase during the initial stage of the swarm,

but reaching the peak frequency, began to decrease gradually. The earthquakes cease a few hours before the flank fissure opening.

The characteristic feature of volcano-tectonic earthquake swarms is the appearance of an earthquake family (a group of earthquakes having similar waveform at fixed seismic station). Kaufman and Burdick (1980) have noted that during the swarm of volcano-tectonic earthquakes accompanied the collapse of Fernandina caldera, Galapagos Islands in June 1968, the teleseismic records of large volcanic earthquake with magnitude *Ms* 5.0−5.5 were spectacularly consistent from event to event (Figure 2.12). The major seismic phases like *P*, *PP*, *SV*, and the Rayleigh waves correlate closely. Even the fine details of the coda between the major phases match.

The families of small earthquakes, recorded at a distance of about 2 km, were founded during the dacitic dome growing at Usu volcano, Hokkaido (Okada et al., 1981; Okada, 1983). The waveforms in each family (Figure 2.12) resemble each other in details from the onset to the coda, even in high frequencies (≥10 Hz).

Figure 2.12 Characteristic waveforms of volcano-tectonic earthquake families recorded at teleseismic distance during the caldera collapse of Fernandina volcano, Galapagos in 1968 (long-period records), and at near distance (about 2 km) during the dome growth at Usu volcano, Hokkaido in 1978 (short-period records). For Fernandina records, the minute marks are shown.
After Kaufman and Burdick, 1980 and Okada et al., 1981.

Hypocenters of one family cluster in a very small region. Figure 2.12 shows that the families OU1 and OU2 have completely different waveforms, although they are located less than 200 m apart.

Okada (1983) noted that about 80% of 2300 located volcanic earthquakes ($M > 3.5$) occurred only in four families. Earthquakes of any particular family tend to exist over a narrow range of their size distribution. The waveforms for family earthquakes vary only in the amplitude and not in the spectral shape.

2.3.2 Sequences of Eruption Earthquakes

Eruption earthquakes also may occur in sequences. Their clustering depends on the style of volcanic activity. The destruction of lava dome may cause a long-term sequence of seismic signals produced by the collapse of the lava body pieces and their movement along the volcanic flanks (Figure 2.13). The permanent process of degassing can produce a sequence of explosion earthquakes (Figure 2.14). The magmatic processes beneath the crater floor generate continuous volcanic tremor (Figure 2.15). The nature of these sequences of volcanic earthquakes differs from the nature of volcano-tectonic earthquake swarms and is related to the dynamics of deep and surface manifestations of magmatic activity.

Figure 2.13 Sequence of the seismic signals produced by rockfalls during the 2002 block-lava eruption at Volcán de Colima, México. One-day short-period records at a distance of 1.7 km from the crater are shown.
Courtesy of Colima Volcano Observatory.

Figure 2.14 Sequence of explosion earthquakes recorded at Karymsky volcano, Kamchatka during the 1997 eruption. The 8-h broadband records at a distance of 1.5 km from the crater are shown.
From Johnson and Lees, 2000.

Figure 2.15 Sequence of continuous volcanic tremors recorded by short-period instrument during sustained lava eruption of Kilauea volcano, Hawaii.
From Koyanagi et al., 1987.

3 Fundamentals of Volcanic Seismology

The Earth's volcanism is the manifestation at the planet's surface of internal thermal processes through the emission at the surface of solid, liquid, or gaseous products (Francis, 1998). The variety of eruptive processes, related to magma movement at depth, and occurring during the emission at the surface of the solid, liquid, or gaseous products, may generate the seismic signals. Therefore, the fundamentals of *volcanic seismology* represent not only the mathematical approximation of the sources generating the seismic waves but also the description of magmatic processes, able to produce volcanic earthquakes, and of the physical properties of medium where these earthquakes occur. In this chapter, the author discusses the physical aspects of magma flow at depth, the experimental evidences of generation of the seismic signals at the different stages of magma flow, and the tensor representation of seismo-volcanic sources.

3.1 Magma Flow within the Volcanic Conduit

The ascent of magma, its movement to the surface from the magma reservoir, may be considered as steady or unsteady multi-phase flow. Figure 3.1 shows a general scheme of magma ascent from the deep magma reservoir through the volcanic conduit. The magma reservoir contains melt, crystals, and dissolved gas. Magma is saturated in volatiles at the top of the reservoir. Bubbles nucleate at slightly higher level (the exsolution surface), as magma rises and experiences a decrease in pressure. Bubbles are critical because they provide the driving force for eruptions, either by increasing buoyancy or by increasing magma reservoir overpressure. Between the exsolution and fragmentation surfaces, the magma is a mixture of silicate melt and gas bubbles. Fragmentation occurs when gas occupies 70–80% of the available volume. At this point, the magma transforms from a liquid with suspended bubbles to a gas with suspended clots of liquid. The abrupt density decrease (expansion) that accompanies this transformation results in acceleration of the gas mixture out of the vent, in the form of a volcanic plume (Cashman et al., 2000).

Seismic signals may be generated at any point in this system, but the type of seismicity depends on the physical processes and the state of the magmatic fluids at each stage of magma flow. The fluid-filled cavities and cracks within the volcanic conduit may be considered as two-phase solid–liquid systems and serve as the

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00003-7
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Figure 3.1 Schematic illustration of processes that occur in volcanic conduits.
From Cashman et al., 2000.

oscillators generating seismic waves. The magma ascent within the conduit before an explosion also generates the seismic signals.

3.1.1 Magma Flow Regimes

Magma flow in the conduit is characterized by three regimes (Melnik et al., 2005; Gonnermann and Manga, 2007).

(1) *Homogeneous magma flow* occurs in the lowest zone, between the magma reservoir and the exsolution (nucleation) surface of the conduit. The ambient pressure, p, within this zone is higher than the nucleation pressure, p_{nuc}, and the magma enters the conduit. Here, for the given initial concentration of dissolved gas c_0 and solubility coefficient k_c ($p > p_{\mathrm{nuc}} = c_0^2/k_c^2 - \Delta p_{\mathrm{nuc}}$), a standard model of a viscous liquid can be applied.

(2) *Bubbly liquid magma flow* takes place in the intermediate zone, between the exsolution (nucleation) and fragmentation surfaces where $p < p_{\mathrm{nuc}}$. Within the nucleation region the bubbles are formed with a number density N_{d}. Bubble nucleation requires a supersaturation pressure to overcome the energy barrier provided by surface tension. Supersaturation may be achieved if diffusion of volatiles from the melt into the bubbles cannot keep pace with the decrease in volatile solubility caused by ascent-driven decompression. Once bubbles nucleate, ascent-driven decompression continues to decrease volatile solubility, causing volatile diffusion from the melt to the melt–vapor interface. Due to viscous resistance, the pressure in the growing bubble, p_{g}, decreases more slowly than the pressure in the surrounding melt, p_{m}. This can result in a large overpressure in the growing bubble, $\Delta p = p_{\mathrm{g}} - p_{\mathrm{m}}$, providing the magma ascent rate and magma viscosity are suitably high.

The bubble growth rate depends on the rate-limiting processes such as the diffusion of volatiles to the melt–vapor interface, the viscous flow of the melt, and the rate of change in solubility caused by decompression. The merging of smaller bubbles to make larger ones promotes the separation of gas from the melt. Coalescence of the bubbles depends on melt viscosity and is probably only important in low-viscosity magmas. Volcanic rocks typically contain a power-law or exponential distribution of bubble sizes. The exsolution of volatiles from the melt phase during open-system degassing may occur as the gas bubble migration from deeper to shallow parts of magma, or gas may flow from bubble to bubble through small apertures in the surrounding melt shells, or bubble walls may be ruptured during magma fragmentation, allowing gas to escape.

(3) *A gas-particle dispersion magma flow regime* occurs above a narrow region of fragmentation that separates a zone of high-density, high-viscous magma from a zone of

low-density gas-particle dispersion, the resistance of which is determined by turbulent viscosity of the gas phase and is negligibly small. When Δp exceeds a critical value, fragmentation of the bubbly media occurs. A competing process is the coalescence of the bubbles with development of a permeable porous structure and outflow of gas from the magma through a system of interconnected pores. This process reduces gas pressure and can also lead to a collapse of the porosity to form dense magma. The total resistance of the conduit and average weight of the mixture are determined by the position of the fragmentation region.

The actual process of fragmentation is probably different in magmas of different rheologies. The likely fragmentation mechanism in low-viscosity magmas, such as Hawaiian fire fountains and Strombolian bubble bursts, is the *fluid instability*. The fragmentation of an already-vesiculated magma occurs by *sudden decompression and bubble rupture*. This mechanism may be applied to the failure of volcanic domes, and may also describe explosive disruption of viscous melts during Vulcanian eruptions. Relative rate of bubble growth, magma transport, and gas loss will control the style of eruptive behavior (explosive versus effusive) (Cashman et al., 2000).

The interaction of magma (or lava) with groundwater or surface water can lead to *phreato-magmatic fragmentation*. The depth of fragmentation is likely to be determined by the depth of the aquifer. Phreato-magmatic fragmentation results in a spectrum of eruptive activity ranging from passive quenching of the magma/lava to explosive ejection of ash (Morrissey et al., 2000).

3.1.2 Modeling of Magma Flow Regimes

Gonnermann and Manga (2007) performed the model calculations for conduit flow regimes (conduit radius is 25 m) for two end-member cases of bubble number densities $N_d = 10^9$ m^{-3} and $N_d = 10^{15}$ m^{-3}. The eruptive style (explosive or effusive) is largely controlled by the magma flow rate Q_m, which is modulated by processes within the volcanic conduit below the fragmentation surface. Bubble number density, N_d, controls time scale of bubble growth and the ability of volatiles to exsolve efficiently. For $Q_m < 10^6$ kg s^{-1} degassing is solubility limited. The rate of permeable gas flow through the vesicular magma eventually exceeds the rate of volatile exsolution so that significant outgassing occurs. Eventually, magma viscosity becomes sufficiently large for brittle shear deformation near the conduit walls. At $Q_m > 10^6$ kg s^{-1} supersaturation leads to secondary bubble nucleation (labeled *N*) and shifts to larger N_d. Upon further ascent, viscous heating near the conduit walls results in shear localization and prevents the occurrence of shear brecciation. Finally, overpressure results in fragmentation (*labeled F)* at $p_m \cong 10$ MPa, which is equivalent to several kilometers below the surface, because of low wall friction of the gaseous flow above the fragmentation depth. With increase in magma flow rate Q_m, the style of eruption may change from effusive to explosive.

These characteristics of magma flow constrain the type of volcanic seismicity during an eruption. Modeling of different stages of the magma flow and the experimental study of rock rupturing at high pressure and temperature allow the formulation of the conditions for seismic source generation.

3.2 Experimental Studies of the Volcanic Processes and Their Applications for the Seismic Sources

The experimental studies discussed in this section may be divided into two groups: the brittle fracturing in flowing magma and the modeling of ascending magma within volcanic conduits. The first group of experiments explains the nature of volcano-tectonic, hybrid, and long-period earthquakes; the second group of experiments explains the nature of volcanic tremor and the variety of seismic signals associated with explosive eruptions.

3.2.1 Experimental Grounds of the Brittle Fracturing in the Rocks at High Temperatures and High Pressure

High-temperature fracturing of volcanic rocks. It has long been assumed that seismogenic faulting is confined to cool, brittle rocks, with a temperature upper limit of 600 °C. Tuffen et al. (2003; 2008) deformed high-temperature silica-rich magmas under simulated volcanic conditions in order to test the hypothesis that high-temperature magma fracture is seismogenic. Uniaxial and triaxial deformation experiments have been done on samples of both glassy and crystalline lavas at temperatures up to 900 °C. The acoustic emissions recorded during experiments show that seismogenic rupture may occur in both crystal-rich and crystal-free silicic magmas at eruptive temperatures, extending the range of known conditions for seismogenic faulting. According to Tuffen et al. (2003), fracture occurs in magma if

$$\varepsilon' \eta > \tau^s \tag{3.1}$$

where ε', η, and τ^s are the shear strain rate, magma viscosity, and shear strength of magma, respectively. Magma viscosity must be in the range 10^9-10^{14} Pa s for fracture to occur, assuming typical magma shear strength of 10^6-10^7 Pa and a plausible range of strain rates between 10^{-2} and 10^{-6} s^{-1} (Tuffen et al., 2003).

Figure 3.2 demonstrates a representative acoustic-emission waveform generated by high-temperature fracturing of rhyolitic obsidian and its power spectrum. Energy is predominantly in the 100−300 kHz range. The onset is abrupt, and the waveform is similar to the waveforms that were recorded during brittle failure of obsidian at room temperature. Post-experiment sample analysis showed that numerous predominantly axial cracks had formed, with curved, conchoidal surfaces and local zones of cataclasis, where slip had occurred between major subparallel cracks (Figure 3.2C, upper image). Scanning electron microscopy (SEM) images of surfaces (Figure 3.2C, lower image) show micrometer-scale hackle markings, typical of brittle glass fracture. This experiment shows the possibility of generation of volcano-tectonic earthquakes within high-temperature rocks.

High-pressure deformation of dry and water-saturated basalt samples. Benson et al. (2010) carried out deformation experiments in which dry and water-saturated basalt from Mount Etna volcano was deformed and fractured at an effective confining

Figure 3.2 Experimental results from high-temperature fracture of rhyolitic obsidian. (A) Axial stress, normalized cumulative acoustic-emission (AE) energy, and acoustic-emission *b*-values against time for uniaxial deformation of rhyolitic obsidian at 645 °C and $10^{-4.3}$ s^{-1}. Jumps in cumulative energy correspond with stress drops (arrows) and drops in compliance, indicating that cracking of the sample is associated with release of acoustic energy. Error bars show 95% confidence limits. (B) Waveform (top) and power spectrum (bottom) of a typical acoustic-emission event, showing the sharp onset and high-frequency content (predominantly 100–300 kHz) that are characteristic of brittle failure. (C) Photomicrograph (top) of post-experiment obsidian sample, sectioned normal to applied load, showing formation of gouge on curved brittle fracture surfaces; SEM image (bottom) showing detail of typical fracture surface. From Tuffen et al., 2008.

pressure representative of conditions under a volcanic edifice. In the case of a water-saturated experiment, a confining pressure of 60 MPa was used with a pore pressure of 20 MPa; for the dry experiment, a confining pressure of 40 MPa was used. An array of 18 piezoelectric sensors to detect acoustic emission (AE) was installed around the sample (12) and embedded in the upper and lower steel platens (3 per platen), of which 16 were used for AE recording. Waveforms recorded during deformation of dry samples consisted of high-frequency events (Figure 3.3A) compared to waveforms

Figure 3.3 Acoustic-emission signals and their spectrograms obtained during deformation of a dry sample (high-frequency events, panel A), of a saturated-water sample (hybrid events, panel B), and of a saturated-water sample which was induced by venting the water pore fluid via the top part of the apparatus (low-frequency events, panel C). The top row in each panel illustrates representative waveforms, with power spectrograms plotted underneath (color denotes power). From Benson et al., 2010.

received during deformation of the water-saturated samples which frequently exhibited hybrid-type waveforms, characterized by a high-frequency onset and a low-frequency component in the tail (Figure 3.3B). Benson et al. (2010) suggest that hybrid seismic events are likely to be produced by the dual process of crack nucleation and deformation producing the high-frequency onset and, once these fluid pathways are created, fluid moving through the damage/crack network producing low-frequency resonance seen in the coda of the waveforms.

A notable trend was observed of reducing instances of hybrid events leading up to the failure stage in the experiments, suggesting that the pore fluid present in the rock moves sufficiently quickly to provide a resonance, seen as a low-frequency coda. Figure 3.3C shows a waveform of a purely low-frequency event from the same experiment with a saturated-water sample. This was induced by venting the water pore fluid via the top part of the apparatus. It has the effect of isolating the high-frequency generation mechanism (micro-cracking) showing a clear low-frequency signal with significant power at low frequencies (50–100 kHz). Benson et al. (2010) concluded that hybrid seismic events are, in fact, a common type of volcanic seismic event with either high-frequency or low-frequency events representing end members, and whose proportion depends on pore fluid being present in the rock-type being deformed, as well as how close the rock is to failure.

3.2.2 Experimental Grounds of the Origin of Seismic Signals During the Magma Ascending Within the Volcanic Conduit

Our knowledge about the parameters of magma ascending within volcanic conduits is rather approximate. Sahagian (2005) reviewed a set of models of conduit processes presented during the 2002 *Volcanic Conduit and Eruption Modeling* International workshop and summarized the parameters of ascending magma taken by researchers as the output parameters of their models. The summary of the quantities characterizing the processes of magma ascent are presented in Table 3.1.

Table 3.1 Generalized Model of Ascending Magma Parameters Within Conduit (by Sahagian, 2005)

Parameter	Quantity range
Fragmentation depth	1,080–5,650 m
Fragmentation pressure	12–29 MPa
Magma gas volume fraction at fragmentation	0.60–0.83
Exit gas volume fraction at the vent	About 1
Exit dissolved water concentration	0.5–2.0 wt.% H_2O
Exit pressure	0.1–6.1 MPa
Discharge rate	0.7×10^7–7.7×10^7 kg/s
Exit velocity	118–182 m/s

The author discusses here two types of models: the models of explosive processes with rupturing of the surface of bubbly magma (Alidibirov and Dingwell, 1996; Koyaguchi and Mitani, 2005; Lane et al., 2001) and the models that do not require the rupture of diaphragm to generate a gas-driven explosion (Jaupart and Vergniolle, 1988; Ripepe et al., 1996; Chouet et al., 1997).

The shock-tube model. This model explains the explosive eruption by rapid decompression of magma in a shock tube (Alidibirov and Dingwell, 1996; Koyaguchi and Mitani, 2005). According to this model, a bubbly magma at a high pressure is separated from air at the atmospheric pressure by a diaphragm. As the diaphragm is ruptured, a shock wave propagates into the air, and a rarefaction wave propagates into the bubbly magma. As a result, the bubbly magma is decompressed and expands. The magma fragments and the flow are assumed to be changed from bubbly flow to gas-pyroclast dispersion when the hoop stress or the gas volume fraction reaches a given threshold. Two types of fragmentation mechanisms are recognized: (1) high-viscosity magma fragments as the hoop stress reaches the tensile strength of the melt (stress fragmentation) and (2) the hoop stress does not grow in low-viscosity magma so that fragmentation occurs after bubble expansion when the gas volume fraction reaches a threshold (expansion fragmentation). This model places the explosion trigger at the magma surface.

Arciniega-Ceballos et al. (2011) modeled the micro-seismicity produced by volcanic explosions in a shock-tube apparatus where the rapid depressurization of volcanic rocks was achieved. Several experiments with samples with different porosity were performed under controlled pressure conditions (ranging from 4 to 20 MPa), at room temperature. Arciniega-Ceballos et al. (2011) wrote that the inflation−deflation states of the pipe-like conduit and the fragmentation process after the rapid removal of the diaphragm were clearly recognized in the micro-seismic records.

Modeling of expanding gas-fluid flows. Expanding gas-liquid flows, designed to be analogous to those in volcanic conduits, were generated in the laboratory by Lane et al. (2001) using organic gas-gum rosin (natural pine resin) mixtures expanding in a narrow vertical tube of borosilicate glass connected to a vacuum chamber. The tube was isolated from the vacuum chamber by a thin plastic diaphragm and an eruption was triggered by the rupture of the diaphragm. Depending on the level of volatile supersaturation, different flow regimes were observed, ranging from gentle unfragmented to violent fragmenting flows.

Figure 3.4A shows the pressure traces obtained at various heights in the tube during a fragmenting diethyl ether-driven flow. Colors represent different flow regimes identified in video and still-camera images, and in optical and pressure sensor data. Figure 3.3B gives schematic representation of the flow regimes and pressure oscillations in the fragmenting foam flow. Pressure oscillations are not detectable in *R4* (source liquid) despite oscillations being present in the flow regions above this region. The wave packets are dominantly present when the flow is fully detached from the tube wall but still interacting with the wall on occasion. Some wave packets have a broadband signature and represent impacts of gum rosin on the tube walls, while others display a series of narrow-band peaks attributed to resonant oscillation of the gas between foam slugs.

Figure 3.4 Modeling of expanding gas-fluid flows and pressure oscillations in the fragmenting foam flow. (A) Representation of the fragmenting diethyl ether-driven flow in the terms of height in tube versus time plot of optical and pressure data. The positions of the pressure (P) and optical (O) transducers are indicated at the left and right, respectively. Open circles mark the times at which the first 270 foam fragments are detected, and filled circles indicate the onset of unfragmented flow. The circle at zero height represents the time at which P14 detected pressure falling below 99 kPa on decompression. The pressure data have been high-pass filtered at 5 Hz to remove transducer drift. The noise-free section of pressure data between 0.1 and 0.3 s is synthetic data included to reduce high-pass filter end effects. The pressure range (2 kPa) is indicated by the scale bar. Shading identifies regions in which different flow regimes are observed. (B) Schematic representation of flow regimes and pressure oscillations observed in a diethyl ether-driven fragmenting foam flow. The positions of the pressure transducers are indicated at the left and the flow regions are representative of the conditions 0.3 s into the flow (see part A). The "?" in (A) indicate the uncertainties in the border estimations between the regions R with different flow regime.
From Lane et al., 2001.

Pressure oscillations were detected over the whole range of flow conditions explored by Lane et al. (2001), and spectral analyses of these oscillations indicated that the dominant oscillation frequencies could be attributed to resonant oscillations in both foam and open gas spaces within the experimental tube. Flows in which foam fragmentation occurred displayed broader, higher frequency, and more energetic spectrum than unfragmented flows.

Chouet (2003) noted that some of the pressure data obtained by Lane et al. (2001) show remarkable similarities to the seismic signals recorded for volcanoes. Figure 3.5 compares the vertical ground displacement record of very-long-period event obtained at Kilauea volcano with pressure pulses recorded during the diethyl ether-driven flow near the point of optically detected fragmentation. The Kilauea events occurred during transition from slow deflation to rapid inflation of the volcano. Chouet (2003) noted that the reasons for these similarities are not yet clear, but they suggest that similar fluid mechanical processes may be operated in both laboratory experiments and in the fluids moving beneath Kilauea.

The foam-collapse model. Foam is a mass of small bubbles formed in or on a liquid. Juapart and Vergniolle (1989) proposed the process of the generation and collapse of a foam layer at the root of basaltic magma chamber to explain the different regimes of eruptions at basaltic volcanoes, such as fire fountaining at Kilauea or Strombolian explosions. In an attempt to provide a unifying framework for these regimes, the phenomenon, induced by degassing in a reservoir which empties into a small conduit, was investigated. Laboratory experiments were done in a cylindrical tank topped by a thin vertical tube. Gas bubbles were generated at the tank bottom with the known bubble diameter and total gas flux. The bubbles rose through the tank and accumulate in a foam layer at the roof.

Depending on the behavior of this foam layer, three different regimes can be distinguished: (1) steady horizontal flow of the foam leading to bubbly flow in the conduit; (2) alternating regimes of foam build-up and collapse leading to the eruption of a single, large gas pocket; and (3) flow of the foam partially coalesced into larger gas pockets leading to intermittent slug flow in the conduit. These regimes have natural counterparts in basaltic volcanoes.

Juapart and Vergniolle (1989) showed that the necessary and sufficient condition for foam layer collapse is that the foam thickness h exceeds a critical value

$$h_c = 2\sigma / \varepsilon \rho_1 g R \tag{3.2}$$

where σ is the coefficient of surface tension, ρ_1 is the liquid density, g is the acceleration due to gravity, R is the bubble radius, and ε is the gas volume fraction in the foam. In a liquid of given physical properties, this occurs when the gas flux exceeds a critical value which depends on viscosity, surface tension, and bubble size.

This model places the explosion trigger at the initial deep source. The seismological application of the foam model was shown by Ripepe et al. (2001). They carried out a laboratory experiment, where bubbles in water were formed by air pumped at a constant flow rate from the bottom of a cylindrical tank of Plexiglas

Figure 3.5 Comparison between (A) VLP displacement waveforms recorded at Kilauea Volcano on 12 February 1997, during a transition from slow deflation to rapid inflation, and (B) pressure waveforms obtained at the boundary between regions R1 and R2 (see Figure 3.4) in the diethyl ether-driven fragmenting foam flow. The record in (A) has been band-pass filtered in the 0.01−0.1 Hz band, and the record in (B) has been similarly filtered in the 0.5−10 kHz band. Date and time at the start of the Kilauea record are indicated at the upper left. Notice the different time scales.
From Chouet, 2003.

filled with water. They merged into a pipe imitating magma conduit. The gas foam accumulated at the roof of the tank and then collapsed into a large bubble inside the pipe. The bubble began to flow within the pipe and then finally broke at the liquid surface. During the movement of the bubble, an acoustic sensor in the tube outside the water recorded a low-frequency signal; when the bubble breaks at the liquid surface, a high-frequency signal was recorded (Figure 3.6A). The laboratory signals presented strong similarities to the seismic signal produced by a Stromboli explosion (Figure 3.6B).

Figure 3.6 Signals recorded during the laboratory experiments ((A) and (B)) and the seismic (C), infrasonic (D), and light sensitive (E) recordings of the Stromboli explosion. (A) Low-frequency signal that was recorded as soon as the gas slug starts to rise (t_l) followed by a high-frequency part (t_h) as soon as the bubble film breaks at the water surface; (B) Strong decompressive signal that was recorded by the sensor inside the water. Both signals are comparable to the signals recorded at Stromboli ((C) and (D)). The time delay between seismic (C) and infrasonic (D) indicates that the source dynamics needs to trigger the explosion at the magma free surface. The second time delay, Δt_e, between infrasonic (D) and light sensitive (E) sensors depends on the gas jet velocity and the sound speed inside the conduit.
From Ripepe et al., 2001.

3.3 General Description of the Source of Seismic Signals at Volcanoes

3.3.1 Equivalent Force System Acting in the Earthquake Source

According to Backus and Mulcahy (1976), the seismic source of a tectonic earthquake is an indigenous source that originates within the Earth; its equivalent force system exerts neither total force nor total torque at any instance to the Earth and may be described completely by the seismic moment tensor. At the same time, the existence of seismic events at active volcanoes that can be described by a single force equivalent system was demonstrated by Kanamori and Given (1982) and other seismologists. The displacement field generated by a seismic source at volcanoes is described by the representation theorem and may be written for a point source as (Chouet, 1996):

$$U_i(t)=F_j(t) * G_{ij}(t)+M_{jq}(t) * G_{ij,q}(t) \tag{3.3}$$

where $U_i(t)$ is the i-component of seismic displacement at the receiver at time t, $F_j(t)$ is the time history of the force applied in the j-direction, $M_{jq}(t)$ is the time history of the jq-component ($j, q = x, y, z$) of the moment tensor, and $G_{ij}(t)$ is the Green's tensor function which relates the i-component of displacement at the receiver position with the j-component of impulsive force at the source position. The notation q indicates spatial differentiation with respect to the q-coordinate and the symbol * denotes convolution. Summation over repeated indices is implied. Therefore, the general source of seismic signals at volcanoes may be represented by a combination of its elements such as single force F, seismic moment M, and Green's delta function $G_{ij}(t)$. A short description of these three elements is given below.

3.3.2 Green's Functions

Green's function $G_{ij}\ (x,\ t;\ \eta,\ \tau)$ is the ith component of displacement at $(x,\ t)$ excited by the unit impulse applied at $x = \eta$ and $t = \tau$ in the j-direction. Green's functions combine the elastic and inelastic effects of propagation from the source to receiver and describe the signal that would arrive at the seismometer if the source-time function were a delta function (Stein and Wysession, 2003).

3.3.3 Single Force

The force F_j applied at a particular point ξ_0 is given as

$$F_j(\xi_0, \tau) = \iiint_V f_j(\eta, \tau) dV(\eta). \tag{3.4}$$

Takei and Kumazawa (1994) defined a single force F_j as a momentum exchange between the seismic source volume and the rest of the Earth. They stated that the single force originates from the difference between the density structure of the prescribed model and that of the actual value in the source region before the event and from the temporal change of the density structure in the source region caused by the finite displacement of mass during the event. The latter is a nonlinear effect caused by mass advection, which may occur by fluid flow, especially in volcanic regions. According to Chouet (2003), during the ascent of a slug of gas in a column of liquid toward the surface, the liquid moves downward to fill the void left behind by the ascending gas. The sinking of dense liquid changes the density structure of the fluid column and releases gravitational energy. Another example of a vertical single force is the recoil force generated by volcanic jet during an eruption (Kanamori et al., 1984; Chouet et al., 1997). The single force source may be used for the study of volcanic explosions and landslides (Kanamori and Given, 1982; Nishimura, 1995).

3.3.4 Seismic Moment Tensor

The symmetric second-order seismic moment tensor $\boldsymbol{M_{ij}}$ describes the overall feature of the seismic source and consists of six independent components:

$$M_{ij} = \begin{pmatrix} M_{11} & M_{12} & M_{13} \\ M_{21} & M_{22} & M_{23} \\ M_{31} & M_{32} & M_{33} \end{pmatrix} \tag{3.5}$$

Each component of $\boldsymbol{M_{ij}}$ corresponds to one set of opposing forces (Figure 3.7). The seismic moment tensor, as a symmetric second-order tensor, has six independent components while the double-couple equivalent body force for a shear dislocation has only four degrees of freedom. These two extra degrees of freedom are the components of non−double-couple source which may be considered as spherical, cylindrical sources or tensile crack taken as its end members (Julian et al., 1998; Kawakatsu and Yamamoto, 2007).

A double-couple source is applied for the description of the source process of volcano-tectonic earthquakes. Non−double-couple sources may be used for the study of volcanic explosions, volcanic tremor, and low-frequency earthquakes (Aki et al., 1977; Crosson and Bame, 1985; Chouet, 1986; Jousset et al., 2003), as well as other unusual volcano-tectonic earthquakes (Ekström, 1994; Panza and Saraó, 2000). For non−double-couple sources that represent the reservoirs of different forms filled by volcanic fluid, the amplitude ratios of three eigenvectors of the seismic moment tensor $\boldsymbol{M_{ij}}$ can serve as criteria for discrimination between spherical, cylindrical, or tensile crack sources. Table 3.2 presents the ratios $M_{11} : M_{22} : M_{33}$ calculated for different relationships between the elastic modulus (Lamè coefficients λ and μ). Normally, it is considered for rocks that $\lambda = \mu$; however, for rocks

Figure 3.7 Nine force couples corresponding to moment tensor components.
From Kawakatsu and Yamamoto, 2007.

Table 3.2 Ratios of Three Diagonal Terms of the Seismic Moment Tensors M_{11}, M_{22}, and M_{33} for Some Non-Double-Couple Sources of Volcanic Earthquakes (Kumagai, 2009; Kawakatsu and Yamamoto, 2007) for Different Relations Between Elastic Modulus (Lamè coefficients λ and μ)

Type of Source	M_{11}: M_{22}: M_{33} ($\lambda = \mu$)	M_{11}: M_{22}: M_{33} ($\lambda = 2\mu$)
Spherical	1: 1: 1	1: 1: 1
Vertical cylinder	2: 2: 1	3: 3: 2
Vertical tensile crack	3: 1: 1	4: 1: 1
Horizontal tensile crack	1: 1: 3	1: 1: 2

Note. If crack is inclined, the ratios M_{11}: M_{22}: M_{33} are slightly different from the values proposed in the table.

at or near liquidus temperatures, the Lamè coefficients would be as $\lambda=2\mu$ (Murase and McBirney, 1973).

Therefore, a general description of a volcanic seismic source comprises a combination of six seismic moment tensor components and three single force components for a point source (Kumagai, 2009). Mathematical aspects of seismic sources at volcanoes are given in the review papers by Kawakatsu and Yamamoto (2007) and Kumagai (2009).

3.3.5 Waveform Inversion

The seismic source parameters of volcanic earthquakes may be determined from waveform inversion by finding the best fit between observed and synthetic seismic

Figure 3.8 Detailed map of summit area of Stromboli volcano, Italy, showing broadband seismic stations (solid dots) in relation to crater (ticked line) and eruptive vents (small dots). Stations prefixed by "T" denote those of the "T ring" of sensors, and by "M" those of the "M ring," located at the top and of the volcano, respectively. The arrows point to the locations of the two active vents at the time of the experiment in September 1997. Contour lines represent 100 m contour intervals.
From Chouet et al., 2003.

Figure 3.9 Seismic record (A) of the east component of velocity at station T6 (see Figure 3.8 for location) and the same record obtained by band-pass filtering the data in (A) in the 2–30 s band (B). The event under study is shown in the gray box.
From Chouet et al., 2003.

waveforms. Waveform inversion of seismic signals of volcanic earthquakes is a more complicated task than the inversion of a tectonic earthquake. While the source of a tectonic earthquake is represented by the moment tensor corresponding to double-couple, seismic signals at volcanoes, as was noted above, may be generated by complex source processes, requiring a general source representation consisting of the single force and the moment tensor. Furthermore, instead of a step-like function for the source-time function of a tectonic earthquake, the complex time history at the source of volcano-seismic signals, represented by Green's functions, involves more unknowns.

Figure 3.10 Waveform match obtained for the 22 September 1997 event, in which six moment-tensor components and three single-force components are assumed for the source mechanism. Thin lines indicate synthetics, and bold lines represent observed velocity waveforms. The station code and component of motion are indicated at the upper right of each seismogram.
From Chouet et al., 2003.

The use of appropriate Green's functions is critically important to obtain the correct solution in either the time or the frequency domains. If the source is located in a medium that can be approximated by a homogeneous or layered structure with a flat surface, Green's functions can be calculated by using propagator matrices and the discrete wave number method. However, this situation is not common in volcanic regions, and it is necessary to take into account steep topographies and the heterogeneous structure of volcanoes. The boundary integral equation method and boundary element method can be used to quantify the effect of three-dimensional topography in a homogeneous or layered structure. To treat both topography and structural heterogeneity simultaneously, the finite-difference method and the discrete lattice method are suitable (Kumagai, 2009).

The time domain waveform inversion of the seismic signals associated with the 22 September 1997 Strombolian explosion observed at Stromboli volcano, Italy, was performed by Chouet et al. (2003). Broadband velocity seismic records of 13 digital seismic stations were used in the inversion; the stations nearest to the crater are shown in Figure 3.8. The station configuration gives a good coverage of the volcanic crater. The seismic record of the event under study, identified in the gray box, is shown in Figure 3.9A. Before inversion, the seismic record was band-pass (2−30 s) filtered (Figure 3.9B). According to analysis of particle motions, the source was positioned within the small region northwest of the crater. In the inversion, six moment tensor and three single force components for the point source were assumed. Green's functions were calculated by the finite-difference method

Figure 3.11 Source-time functions obtained for the 22 September 1997 event, in which six moment-tensor components and three single-force components are assumed for the source mechanism. Shading marks the interval during which the initial volumetric expansion of the source occurs.
From Chouet et al., 2003.

Figure 3.12 Source location of the 22 September 1997 explosive event, occurring at Stromboli volcano. A south-east looking view of Stromboli is shown at the top to help identify the location of the sources within the volcanic edifice. The bold black line on the north-west flank of the edifice bounds the surficial extent of the domain considered in the bottom view. Contours represent 6%, 8%, and 10% errors.
From Chouet et al., 2003.

using the topography of the volcano. Figure 3.10 demonstrates a close match between the calculated and observed waveforms.

The result of this inversion is shown in Figure 3.11. The volumetric components of the moment tensor clearly dominate. The amplitude ratio of the diagonal axes of the moment tensor is (1:1:2) that corresponds to a horizontal tensile crack source, considering that $\lambda=2\mu$ for rocks at or near liquidus temperatures (Table 3.2).

Volumetric source components are accompanied by a dominantly vertical force. The force amplitude, shown in Figure 3.11, contributes roughly 13% of the signal amplitude relative to the contribution from the moment tensor.

The source centroid was located at an elevation of 480 m, 190 m north-northwest of vent 2 (Figure 3.12).

These seismological, volcanological, and experimental fundamentals of *volcanic seismology* will be used in the following chapters in order to develop models for the generation of seismic signals at different stages of volcanic activity.

4 Origin of Volcano-Tectonic Earthquakes

In this chapter, the author considers a group of volcano-tectonic earthquakes that includes the earthquakes generated by stresses, which arise during magma migration in the Earth's crust. They correspond to A and B-types events of Minakami's classification. These earthquakes may be caused by shear or tensile fracturing during magma movement from depth to the Earth's surface through conduits and dykes. Volcano-tectonic earthquakes occur within and around the volcanic edifice and reflect the interaction of two general geological processes: the magma migration to the Earth's surface and the crustal tectonic activity. In Chapter 3, we discussed seismo-generating processes occurring mainly within the flowing magma; this chapter is devoted to the nature of seismic events occurring during fracturing of host rocks where magma migrates to the surface.

4.1 Migration of Magma and Its Seismic Potential

Figure 4.1 gives a general scheme of magma melting in mantle and its rising through the ruptured crust to the surface at zones of ocean spreading and subduction. Magma is formed in mantle in two principal ways: the first way involves simple decompression of hot, solid mantle material, whereas the second involves lowering the melting temperature of the mantle material by the addition of volatiles (Perfit and Davidson, 2000). Plate tectonics show (Figure 4.1) that magma is generated mainly below divergent plate boundaries (oceanic spreading ridges) and below convergent plate boundaries (subduction zones). Intraplate volcanism is supported by magma that is generated within so-called *hot spots* producing *mantle plumes*. The location of the magma generation constrains its physical properties and constitution.

Partial melting of mantle generates basaltic magmas; the silicic magmas are generated by partial melting of crustal rocks (Rogers and Hawkesworth, 2000). Common magmas vary nearly continuously in composition from rhyolitic (75% SiO_2) through to basaltic (50% SiO_2). Generally, three types of magma are distinguished: acid (dacitic−andesitic) containing more than 66% of SiO_2, intermediate (andesito−basaltic) containing between 52% and 66% of SiO_2, and basic basaltic, containing less than 52% of SiO_2.

After magma generation, it commences its journey to the surface. At mantle depths, magma rises very slowly within convection currents without fracturing the

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00004-9
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Figure 4.1 Illustration of the process of magma melting in the mantle and its rising at the mid-ocean spreading centers and subduction zones to the surface. Volcanoes overlie the zones where slowly moving oceanic lithosphere is pushed beneath the thicker lithosphere of the continent. The engines that drive this movement are rising heat currents in the mantle. The molten magma rises buoyantly toward the Earth's surface.
After Fisher et al., 1998.

Figure 4.2 Cross-section of *P*-wave velocity perturbations from the average velocity at a given depth beneath Pinatubo volcano, Philippines. The dots show the hypocenters of post-eruption seismicity from 29 June to 16 August 1991.
From Mori et al., 1996a.

surrounding material. Within the shallow crust there is storage of magma, and differentiation begins to occur.

Seismic studies allow the location of the zones of magma storage because these represent a zone of low velocity and high attenuation of seismic waves. Figures 4.2 and 4.3 show the locations of low-velocity zones beneath the volcanoes Pinatubo,

Figure 4.3 Cross-section of *P*-wave velocity perturbations beneath Mount St. Helens volcano, Cascades. Contours of absolute *P*-wave velocities (background plus perturbation) are provided for reference.
From Lees, 1992.

the Philippines and Mount St. Helens, Cascades. A large zone, representing a relative decrease of *P*-wave velocity (5−10%) beneath Pinatubo Volcano (Figure 4.2), was located between depths of 6 and 11 km and is estimated to have a volume of 40−90 km^3 (Mori et al., 1996a). For Mount St. Helens (Figure 4.3), a *P*-wave low-velocity region began at 1.5 km depth, directly below the crater. This localized low-velocity zone extends to the depth of 3.5 km where it then appears to broaden (4−5 km diameter) and continues to the depth of 7.5 km (Lees, 1992). Therefore, the regions of magma storage are situated at very shallow depths.

Two types of igneous rocks are distinguished: intrusive igneous rocks and extrusive igneous rocks. Intrusive igneous rocks crystallize when molten rock intrudes into unmelted rock masses deep in the Earth's crust while extrusive igneous rocks are formed when magma erupts at the surface. The intrusive and extrusive rocks can have the same content but different textures. The main types of intrusions are *sills* and *dykes*. Sills are concordant, tabular bodies that are emplaced essentially parallel to the bedding of the country rock and occur in relatively unfolded regions at shallow crustal level. Low viscosity of magma is required to produce this sheet-like form. Dykes are thin, tabular, discordant intrusive bodies that cut across the foliation or bedding of the country rock. They range in thickness from less than a meter to several hundred meters, and a few have been traced along strike for tens of kilometers. Typically emplaced into already existing fracture systems, they may occur singly or in swarms (Figure 4.4). In some areas, dykes occur as radiating

Figure 4.4 Swarm of dykes in the crater area of Tokachi volcano, Hokkaido. Photograph by V. Zobin.

swarms centered on an intrusion or on the flanks of a volcano. In rare cases, vertical- or outward-dipping ring dykes or inward-dipping cone sheets are distributed in oval or circular patterns around the intrusion (Blatt and Tracy, 1996).

Extrusive rocks form different types of lavas. The term *lava* is used both for magma that has been erupted onto the Earth's surface and for the rocks that have solidified from it (Blatt and Tracy, 1996).

When magma rises from its storage zone, it begins to move along the system of preexisting or newly formed tectonic fractures and the volcanic conduit to the surface as a dyke or sill. The magma ascent rate is a function of the pressure in the magma storage region, the density and viscosity of the magma, the conduit diameter, and the resistance to flow in the conduit that connects the magma storage zone to the surface (Rutherford and Gardner, 2000). All these factors affect the generation of the volcano-tectonic seismic activity and define the style of eruptive process.

Magma is not pure liquid but has rheological properties, and its ability to flow depends strongly on its viscosity. Figure 4.5 shows that viscosity of different melted rocks at high temperature (about 1000 °C) can change by 10 orders from komatite (less than 1 Pa s) to rhyolite (10^{10} Pa s) (Spera, 2000). This difference in viscosity determines the ability of different magma types to migrate through the cold rocks of the interior of the Earth. Low-viscosity basaltic magma prefers to go

Figure 4.5 Viscosity as a function of temperature at a pressure of 1 bar for natural melts spanning the compositional range rhyolite to komatite. All compositions are volatile free. From Spera, 2000.

along the stratified structures (or faults) in accordance with their strike and does not cut them. The intermediate-viscosity andesite–basaltic magma may choose both ways. High-viscosity dacitic magma slowly cuts stratified structures and forms obelisks. The average ascent rate varies from 0.001 to 0.015 m/s for extrusive magmas. During explosive eruptions, the assessment of magma ascent rate is characterized by much greater variations and leaves the top of conduit at essentially sonic velocities for truly explosive eruptions (Rutherford and Gardner, 2000). This intrusive–extrusive activity of magma indicates the seismic potential of magmatism within the upper crust.

4.2 Volcanism and Tectonics

Volcanoes normally occur within rift (or graben) type structures along long-lived faults. A graben is often formed above intruding dykes in volcanic rift zones because a dyke creates two zones of maximum tensional strain at the surface, parallel to the dyke axis (Perfit and Davidson, 2000). The regional tectonic faults where the volcano is situated are accompanied commonly by numerous local faults. These structures could be mid-oceanic ridges or rift valleys (depressions) along the convergent boundary (subduction zone) or above a hot spot (see Figure 4.1). They could be the basis for a long volcanic chain or a small volcanic group. Basaltic volcanoes have a well-defined framework of local faults where the flank cones are formed.

Figure 4.6 shows the structural map of Mount Etna volcano, Sicily (Azarro, 1999). This volcano has grown at the intersection of two regional faults: the Malta Escarpment and the Messina-Fiumefreddo, striking, respectively, north-northwest and northeast, and at the front of the Apennine-Maghrebian thrust belt, the crustal

Figure 4.6 Simplified structural map of Mount Etna. (1) Faults with downthrown side; (2) eruptive fissures; (3) cinder cones; (4) coseismic surface faulting zones; (5) strike-slip components; (6) caldera rims; (7) limit of Etna volcanic edifice; (C.C.) central craters. Box indicates area of the Timple fault system. Inset map (a) shows the regional geological setting: AMF, front of the Apennine-Maghrebian thrust belt; IF, Iblean Foreland; ME, Malta Escarpment; MFL, Messina-Fiumefreddo line.
From Azarro, 1999.

scale fault. The tectonic setting of Etna's flanks results from the interaction of regional and local tectonics. The Timple system at the volcano's eastern flank represents the northernmost prolongation of the Malta Escarpment and forms a north-northwest–south-southeast (NNW–SSE)-trending system of parallel east-facing step-faults of considerable length (8–10 km). To the north, this system is interrupted by the east–west Pernicana fault that cuts a large part of the volcanic edifice and by the northwest–southwest Ripe della Naca system. Two isolated structures, the Trecastagni and Tremestrieri faults, are situated at the southern flank of the volcano.

The ground breakage that was observed in the volcanic region generally follows preexisting active fault traces. Azarro (1999) also shows evidence of movement along the hidden and semi-hidden local faults. The flank volcanic eruptions occur normally along the local tectonic faults, mainly on the eastern flank of the volcano (Gresta et al., 1990). Among them was the 1991–1993 eruption that occurred along the NNE–SSE Timple system of faults (Ferrucci and Patane, 1993). The existence of numerous cinder cones illustrates the process of magma intrusion along the local faults.

Another basaltic volcano, Kilauea, Hawaii (Figure 4.7), is situated within a complex tectonic framework formed by the Ka'Oki, Koa'E, and Hilina fault systems and

Figure 4.7 Kilauea volcano, Hawaii, and its rifts and fault systems.
From Parfitt and Peacock, 2001.

the southwest and east rift zones. The normal fault systems at the south flank of Kilauea do not participate directly in the magmatic activity but reflect longer-term deformation of the volcanic edifice. The fault systems consist of *en echelon* fault segments that have significant intersection and linkage (Parfitt and Peacock, 2001).

Rift zones, which intersect the flanks of the volcano, are filled by dykes that supply magma and form cinder cones. The dykes propagate downrift at narrow blade-like intrusions from shallow magma reservoirs beneath the summit caldera. Rift eruptions and many summit eruptions begin when the magma issues from long extension cracks that fracture the surface immediately prior to the eruptions. These eruptive fissures are oriented parallel to the regional trend of the rift zones and mark the intersection of propagating dykes with the Earth's surface (Dieterich, 1988).

The precise location of small (magnitude 1−2) volcano-tectonic earthquakes that were observed during the 1983 dyke intrusion fissure eruption (Rubin et al., 1998) shows a pronounced tightening of seismicity along the dyke trend (Figure 4.8A). The foci of earthquakes are located about 0.5 km south of the eruptive fissure, but a dyke that dipped only 7° from vertical would pass through them. The hypocenters of the earthquakes lie between 3 and 4 km depth (Figure 4.8B) that indicate the origin of seismicity from a very narrow depth interval, despite the fact that the dyke breached the surface.

The andesitic stratovolcano Volcán de Colima is located within the north-south −trending Colima Rift Zone, in the western part of the Mexican Volcanic Belt, and together with the Pleistocene volcano Nevado de Colima forms the Colima Volcanic Complex (Figure 4.9). The Colima Volcanic Complex is situated above the subduction zone, where the Rivera plate goes beneath the North American plate. At the same time, the Colima Rift represents the eastern boundary of the

Figure 4.8 Clustering of volcano-tectonic earthquakes (crosses) along the eruptive fissure (thick line) during the 1983 dyke intrusion in East rift zone of Kilauea. (A) Epicenters of earthquakes; (B) cross-section along I−II line of Figure 4.7A. The cinder cones are shown by ovals. From Rubin et al., 1998.

Figure 4.9 Structural position of Volcán de Colima in southwestern México. Three major rift zones are shown: CRZ, Colima Rift Zone; ChRZ, Chapala Rift Zone; and TZRZ, Tepic-Zacoalco Rift Zone. The dotted line running through Volcán de Colima is the proposed Tamazula Fault zone (Garduño-Monroy et al., 1998). Other major volcanic centers from the western Mexican Volcanic Belt are shown as triangles: (1) Nevado de Colima; (2) Volcán Cántaro; (3) Sierra la Primavera; (4) Volcán Tequila; (5) Volcán Ceboruco; (6) Volcán San Pedro; (7) Volcán Tepetiltic; (8) Volcán Sangangüey; (9) Volcán las Navajas; (10) Volcán San Juan; (11) San Sebastian Volcanic Field; (12) Mascota Volcanic Field; (13) Los Volcanes Volcanic Field. From Zobin et al., 2002c.

Jalisco block that is moving toward the ocean from the continent. The Colima volcano also marks the intersection of two large local tectonic structures, the north-south–trending Colima Rift and the northeast-southwest–trending Tamazula fault (Garduño-Monroy et al., 1998).

The November 1998 block-lava extrusion at Volcán de Colima was preceded by a 12-month period of seismic activity that included five earthquake swarms (Zobin et al., 2002b). The epicenters of the earthquakes that were recorded during June–July 1998 swarm clustered across the north-south–trending Colima Rift, while the epicenters of the October–November 1998 earthquake sequence were located along the rift structure (Figure 4.10). This earthquake clustering may reflect the magma movement along the two local faults that were active at different stages of magma uplift to the surface.

The dacitic volcano Usu, Hokkaido, is situated on the southern rim of the nearly circular Toya caldera that is of about 10 km in diameter, surrounded by a steep wall and partially filled now by Lake Toya. Usu volcano has a characteristic distribution of its central and flank cones (Figure 4.11) that outlines a system of circular faults in the volcanic edifice. The craterlets of the 1977 eruptions were formed within the summit caldera, whereas a chain of dacitic cones was formed in

Figure 4.10 Epicenters of earthquakes recorded at Volcán de Colima, México prior to the November 1998 block-lava extrusion in June−July (diamonds) and October−November, 1998 (triangles). VC and NC are Volcán de Colima and Nevado de Colima, respectively. The contours of 3000 and 3500 m are shown.
From Zobin et al., 2002b.

Figure 4.11 Flank cones that were formed at Usu volcano, Hokkaido, during the 1910, 1943−1944, and 1977 eruptions. Contours with dots show approximate outlines of cryptodomes. MS is the 1910 upheaval; K and E have no dates. Thick contours show lava domes. SS indicates the 1943 lava dome; GN and NM are the 1977 Gin-numa craterlet and one of the 1977 peaks, respectively. The stars indicate the wells. Double broken lines show the lateral extent of volcanic activity.
From Yokoyama and Seino, 2000.

Figure 4.12 Deep slices of earthquake foci located during the 1977–1978 eruption of Usu volcano for depths from 0 km (sea level) to 2.0 km. Numbers in the left corner show the number of located events for each depth slice.
From Okada et al., 1981.

1910 and 1943–1944 along the 200-m contour line of volcanic edifice at a distance of about 2–3 km from the summit (Yokoyama and Seino, 2000). These surface manifestations of volcanic activity reflect the movement of dacitic magma along circular faults.

The earthquakes located during the 1977–1978 summit eruption of Usu (Figure 4.12) show circular distribution of earthquake clusters around its caldera (Okada et al., 1981). The earthquakes surround an event-free zone beneath the central part of the summit and indicate deep circular fault distribution. Examining a slice at the 0.5 km depth, the epicenters for an arc in the southwest part of volcano are seen, while with the 1.0 and 1.5 km depth slices, the northeast arc is outlined.

4.3 Source Nature of Volcano-Tectonic Earthquakes

4.3.1 Waveform and Spectra

Figures 4.13 and 4.14 show the typical seismograms of volcano-tectonic earthquakes. The high-frequency waveform of volcano-tectonic earthquake, shown in Figure 4.13, is characterized by clear arrivals of *P*- and *S*-waves. Both spectra have the dominant frequencies between 5 and 20 Hz, but the spectrum of *P*-wave is of higher frequency comparing with the spectrum of *S*-wave. Low-frequency waveform of volcano-tectonic earthquake (Figure 4.14) shows not-so-clear arrivals of *P*- and *S*-waves. The dominant spectra of both types of body waves are between 1 and 4 Hz. As usual, low-frequency volcano-tectonic earthquakes are shallower than high-frequency events. Their waveform duration is longer.

Figure 4.13 Waveform and Fourier spectra of *P*- and *S*-waves of high-frequency volcano-tectonic earthquake. 20 August 2007, 21:11. Nevado de Ruiz volcano, Colombia. Courtesy of INGEOMINAS.

Figure 4.14 Waveform and Fourier spectra of *P*- and *S*-waves of low-frequency volcano-tectonic earthquake. 15 November 1998, 10:32. Volcán de Colima, México. Courtesy of Colima Volcano Observatory.

4.3.2 Tensor Representation of the Source of Volcano-Tectonic Earthquake

The complete seismic moment tensor of an earthquake source may be described as a sum of its isotropic and deviatoric parts. The source of volcano-tectonic earthquakes, produced by movement along tectonic fractures, may be represented by its deviatoric part, which is responsible for the production of fracture. The deviatoric part may be decomposed into two source types: a double-couple (DC) and a compensated linear vector dipole (CLVD) (Figure 4.15). The dominant type may be discriminated by the relationship between the smallest and the largest eigenvalues of the seismic moment tensor ε; it is equal to 0 for a DC source and 0.5 for a CLVD source.

Seismological practice and theoretical studies showed the double-couple source (Figure 4.16) as the most adequate for shear faulting that produces volcano-tectonic earthquakes. This model predicts the quadrant distribution of *P*-wave first motions using stereographic projection (Figure 4.17) and allows the discrimination of the three main types of faulting in earthquake foci: strike-slip, thrust, or normal fault. The values of ε that were obtained for large volcano-tectonic earthquakes related to the eruptions of Mount St. Helens (1980), Myakejima (1983), Oshima (1986),

Figure 4.15 Model for decomposing the deviatoric part of a general moment tensor. DC, the double-couple source; CLVD, the compensated linear vector dipole. Numbers to the left of the diagrams give the principal moments for each mechanism, in units of 10^{17} N m.
From Julian et al., 1998.

Figure 4.16 Radiation pattern proposed for double-couple earthquake source mechanism. (A) Plane view of a vertical strike-slip shear fault in an isotropic medium, showing the direction of slip (open arrows), and the equivalent distribution of double-couple force systems (solid arrow, with forces applied at white dots). (B) Radiation patterns of compressional *P*-waves, showing fault plane (solid lines), and auxiliary plane (dashed lines). Adjacent lobes have opposite polarity.
From Julian et al., 1998.

Figure 4.17 Focal mechanism solutions of volcano-tectonic earthquakes from the earthquake swarm preceding the 2000 eruption of Miyake-Jima volcano. Equal area projection of *P*-wave first motions are projected on the upper hemisphere. Gray and open areas show extensional and compressional quadrant, respectively. Mechanisms of strike-slip-type with northeast–southwest stress axis are predominant, although various focal mechanism types are recognized.
From Uhira et al., 2005.

and Teishi Knoll (1989) from the Harvard CMT solutions (Zobin, 1992) varied from 0.10 to 0.18. The focal mechanism solution may be obtained from the direct plots of the *P*-wave first motions on the stereographic projection or by modeling with synthetic seismograms.

The non-double-couple sources, or the sources with values $0.2 \leq \varepsilon \leq 0.5$, were determined for a few volcano-tectonic earthquakes located at ring fault structures in Iceland (Ekström, 1994; Nettles and Ekström, 1998) and at Etna and Vesuvius, Italy (Panza and Saraó, 2000; Saraó et al., 2001). The focal mechanism solution for the non-double-couple sources may be obtained by modeling with synthetic seismograms; it has a form shown for CLVD in Figure 4.16.

4.4 Models of Volcano-Tectonic Earthquake Sequences

It has been shown that volcano-tectonic earthquakes occur due to magmatic activity and tectonic fracturing. The movement of magma to the surface occurs within a dyke-type body. Rubin and Gillard (1998) proposed a model of dyke-induced seismicity based on dyke intrusion observations at Kilauea Volcano, Hawaii (Rubin and Gillard, 1998; Rubin et al., 1998). They studied the inelastic outcrop-scale deformation produced within several meters of the dyke tip when the internal pressure variations due to viscous flow of magma must be considered.

The schematic diagram of blade-like dyke is shown in Figure 4.18. In this model, the dyke moves within the host rocks that are subject to the action of ambient stress. The region near the tip of dyke (tip cavity) where magma cannot penetrate may be occupied by magmatic volatiles or host rock pore fluids. The region influenced by the increase in stress is comparable in size to the length of the tip cavity. The possibility of the magnitude 1−2 earthquakes with the source dimension of about 100 m in the vicinity of the tip cavity is discussed for three cases (Figure 4.18):

1. Fault slip away from the tip cavity.
2. Fault slip near the tip cavity.
3. Shear failure of intact rock.

The examination of elastic stress fields surrounding propagating fluid-filled cracks showed that the stress state is favorable to failure occurring near the tip cavity when the cavity pressure is maintained by the influx of host rock pore fluids, rather than by exsolution of magmatic volatiles. Even in this case, however, shear fracture of previously intact rock seems unlikely. According to these results, most

Figure 4.18 Schematic diagram of blade-like dyke (vertical section) and the three types of deformations that could be related to the moving dyke: (1) slip on existing faults away from the tip cavity; (2) slip on existing faults adjacent to the tip cavity; and (3) shear failure of intact rock adjacent to the tip cavity. The least principal stress σ_3 is perpendicular to the dyke. The tip cavity of dyke is shown by empty space.
From Rubin and Gillard, 1998.

dyke-induced seismicity should result from slip suitably aligned along existing fractures (Rubin and Gillard, 1998).

Therefore, Rubin and Gillard suggest that the distribution of earthquake foci of dyke seismicity reflects the distribution of ambient stresses that are near to failure, and does not necessarily reflect the extent of the dyke. They reject fracturing as the trajectory of the dyke movement. At the same time, the dyke can trigger the fracturing in the vicinity of its tip along existing fractures. We could consider these fractures as magmatic fractures. Rubin and Gillard (1998) consider that slips could occur at distances over 70−100 m from the dyke if a favorably oriented fault connected two *en echelon* dyke segments.

The conceptual model of the dyke-induced volcano-tectonic earthquake swarms occurring at basaltic volcanoes was proposed by Hill (1977). It is based on interaction between dykes and faults. Hill suggested that the clusters of magma-filled dykes exist within brittle volumes of the crust in regions of active magmatic intrusions and the dykes within a cluster are systematically oriented in the regional deviatoric stress field. Their long dimension is parallel to the greatest principal stress σ_1. Hill (1977) considered shear failures occurring along oblique fault planes connecting adjacent tips of offset dykes when critical combination of the fluid pressure P in the dykes and the regional stress difference $\sigma_1 - \sigma_3$ is attained.

Figure 4.19 shows the schematic representation of the model. The dykes are forced open by a gradual increase in the tectonic stress difference; magma is passively drawn into the dykes. A swarm develops when the stress field within the volume reaches the critical state, and a series of shear failures occur along cracks between the dykes. Hill (1977) suggests that the distribution of earthquake magnitudes will be related to the size and spacing of dykes in the swarm volume in addition to such factors as "roughness" in the variation of frictional properties along individual fractures.

The numerical models of volcano-tectonic earthquake swarms induced by the magma-filled dykes were calculated by Roman (2005). She calculated the inflation-induced Coulomb stress changes on preexisting faults around the volcanic conduit. In her study, dykes trend parallel to regional σ_1 and inflation is in the direction of the regional σ_3. The modeling of Coulomb stress changes on preexisting faults indicates that a relatively small degree of dyke inflation induces slips on faults in dyke walls with a wide range of orientations, including strike-slip motion with reversed sense of slip relative to regional faulting.

Figure 4.19 Schematic representation of dykes and conjugate fault planes with respect to greatest and least principal stresses σ_1 and σ_3. A and B illustrate typical patterns of crack interactions near adjacent dyke tips in homogeneous media.
From Hill, 1977.

Figure 4.20 Variations in the azimuth of compressional *P*-axis obtained from the focal mechanism solutions of volcano-tectonic earthquakes preceding the 1975 fissure Tolbachik, Kamchatka eruption. Dashed lines separate the date before and after the stress field rotation. After Zobin, 1979b.

The stress field rotations obtained from focal mechanism solutions and predicted by this model were observed during the swarms of volcano-tectonic earthquakes associated with the 1975 Tolbachik fussure eruption in Kamchatka (Zobin, 1979b) (Figure 4.20) and the 1992 Crater Peak eruption in Alaska (Roman et al., 2004). During the Tolbachik swarm, this rotation was observed just after two $M = 4.9$ earthquakes occurred.

These three models allow a description of the relationship between the magma inflation and tectonic fracturing. For laterally migrated magma, Rubin−Gillard's and Hill's models work. For vertically migrated magma, the application of Roman's model could be more realistic.

5 Volcano-Tectonic Earthquakes at Basaltic Volcanoes: Case Studies

Basaltic volcanoes have characteristic lava of low viscosity ($10-10^3$ P), high temperature (1000−1300°C), with a small amount of volatiles (0.2−2 wt.%), and high velocity of movement. They are characterized by Hawaiian and Strombolian types of eruption and produce extensive lava flows (Vergniolle and Mangan, 2000; Spera, 2000).

The eruptions of basaltic volcanoes may occur from the summit crater of the volcano, but flank or fissure eruptions are very frequent too. The low-viscosity and high-temperature magmas are very mobile and able to form the passageway to the surface along or out of the central conduit of a volcano. These properties of magma constrain the possible seismic reaction to the magma movement. The fracturing of rock during the forming of new volcanic vents on the flanks of a volcano or along a new fissure produces volcano-tectonic earthquakes. These earthquakes can be observed also during the movement of magma along the volcanic conduit before the summit eruptions. At the same time, the summit eruptions of basaltic volcanoes may frequently occur without any preliminary volcano-tectonic activity. Lava effusion from the summit and flank (fissure) vents is accompanied by volcanic tremor and explosions.

5.1 Volcano-Tectonic Earthquakes Associated with Shield Volcanoes

Shield volcanoes are broad and have gentle slopes. A few shield volcanoes have persistent lava lakes in their summit calderas. Among them are Mount Erebus (Antarctica) and Nyiragongo (Congo). The most active shield volcano is Kilauea in Hawaii.

5.1.1 Kilauea Volcano, Hawaii

Kilauea is a tholeitic shield volcano, and the eruptions of a very liquid magma occur mainly in the summit caldera or from the southwest and east rift (ERZ) zones (see Figure 4.7). Formation of vertical dykes appears to be the dominant process in rift-zone intrusion. Kilauea's volcanic earthquakes occur beneath the rift zones and summit caldera, at the depth of less than 5 km (Klein and Koyanagi, 1989).

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00005-0
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Volcanic earthquakes at Kilauea typically occur in swarms. The history of Kilauea's earthquake swarms that occurred in 1963–1983 (Klein et al., 1987) shows that the swarms can be divided into fast, intense swarms lasting for a few hours to a couple of days (they precede nearly all eruptions of summit caldera and the rift-zone intrusions) and slow swarms lasting several days to a few weeks between the eruptive events.

The author selected three episodes of volcanic activity: the 1969 rift-zone intrusion; the 1974 summit caldera eruption with a rift-zone intrusion; and the 1982 summit caldera eruption. These three episodes were preceded by seismic activity. The detailed descriptions of the events are presented in Klein et al. (1987). The author used the catalog of Hawaiian Volcano Observatory for the analysis of the seismic process.

The 1969 rift-zone intrusion occurred between 22 and 28 February within the ERZ. The eruptive vents were formed at a distance of about 8 km from the summit caldera (Figure 5.1). Seismic activity was represented by two seismic sequences: an inflation swarm and an intrusive swarm (Figure 5.2). The first swarm began on 17 February and was associated with summit inflation recorded by deformation network from 17 to 21 February, but was not followed by summit eruption.

The next, more intensive swarm began on 22 February at 5:23 h and lasted for 10 h. The maximum number of earthquakes per hour and the maximum magnitude

Figure 5.1 Epicenters (A) and hypocenters (B) of the earthquakes recorded from 17 to 22 February 1969 at Kilauea volcano. Crosses show the events recorded between 00:42 h of 17 February and 00:49 h of 22 February (inflation swarm); the circles show the events recorded from 05:23 to 16:33 of 22 February (eruptive swarm). The outline of the caldera is shown, as well as the new ERZ eruptive fissures according to Klein et al. (1987). Data from the catalog of Hawaiian Volcano Observatory.

Figure 5.2 Temporal distribution of total earthquakes recorded before and during the 22 February 1969 ERZ eruption (A) and a structure of the eruptive swarm of 22 February (B). The single arrow indicates the maximum earthquake; the double arrows indicate the period of the opening of the eruptive vents. The dashed line with crosses shows the summit earthquake variations; the line with diamonds shows the ERZ earthquake variations. Data from the catalog of Hawaiian Volcano Observatory.

event (M=3.5) of this swarm were recorded in 3 h after onset of the swarm. The number of events then gradually decreased. Intrusion began 5 h after the swarm beginning. Three systems of eruptive vents were opened along new fissures during about 5 h (Klein et al., 1987).

Seismic events of the both swarms occurred simultaneously in two zones: beneath the summit caldera and near the new eruptive vents in the ERZ (Figure 5.1). The events that occurred near the new eruptive vents during the first swarm were more distant to the south than the events of the second swarm that clustered very close by. Analysis of the temporal variations of both groups of events during the second swarm (Figure 5.2B) shows that the ERZ earthquakes dominated during the first 5 h prior to the opening of the eruptive vents. With the opening of eruptive vents, the summit and ERZ earthquakes were comparable in number and accompanied the 5-h opening of fissures.

The 1974 summit caldera eruption with a rift-zone intrusion occurred on 19 July and lasted for 4 days. The 9-h swarm of about 200 located earthquakes (Figure 5.3) was recorded at depths between 1 and 10 km. The events clustered between the eruptive vents that were formed within the summit caldera and along two fissures opened within the ERZ (Figure 5.4). Magma reached the surface about 2 h after the swarm had ceased.

Figure 5.3 Epicenters (A) and hypocenters (B) of the earthquakes recorded on 19 July 1974 at Kilauea volcano. Crosses show the events recorded between 03:02 and 12:28 h of 19 July prior to the magma intrusion. The outline of the caldera (solid lines) and the new eruptive fissures (dashed lines) are shown according to Klein et al. (1987). Data from the catalog of Hawaiian Volcano Observatory.

Figure 5.4 Temporal distribution of the earthquakes before the 19 July 1974 Kilauea-ERZ eruption. The single arrow indicates the moment of the maximum earthquake; the double arrow indicates the beginning of magma intrusion.
Data from the catalog of Hawaiian Volcano Observatory.

Figure 5.5 Temporal distribution of the earthquakes before the 30 April 1982 Kilauea caldera eruption. The single arrow indicates the moment of maximum earthquake; the double arrow indicates the beginning of the eruption.
Data from the catalog of Hawaiian Volcano Observatory.

The swarm developed in two stages. The first peak was recorded during the first hour of activity; the next peak, more intensive, was observed 5 h after the beginning of the swarm. The maximum magnitude event ($M=3.4$) was recorded during the final part of activity, when the number of events sharply decreased.

The 1982 summit caldera eruption occurred on 30 April and continued for 1 day. The short swarm of earthquakes began about 2.5 h before the eruption and continued up to the beginning of the volcanic manifestations (Figure 5.5).

The majority of earthquakes were recorded during the first hour of activity; then the number of events gradually decreased. The maximum magnitude event ($M=3.2$) was recorded just before the beginning of the eruption.

The epicenters of all 47 events recorded between 08:54 and 11:41 h were located within the summit caldera, where the eruptive fissures opened (Figure 5.6A). Earthquake depths were estimated within the range of 1 to 2 km beneath the caldera (Figure 5.6B).

5.2 Volcano-Tectonic Earthquakes Associated with Stratovolcanoes

A stratovolcano is a volcano constructed of alternating layers of lava flows and pyroclastic rocks. Stratovolcanoes are steeper than shield volcanoes and tend to a conical form. Among the most active basaltic stratovolcanoes are Etna in Sicily, Oshima in Japan, and Klyuchevskoy in Kamchatka, Russia. Many stratovolcanoes have cinder cones and craters of parasitic vents on their flanks (Walker, 2000).

5.2.1 Mount Etna, Sicily

Mount Etna is a basaltic stratovolcano with an elevation of 3350 m. It is the most active volcano in Europe. Etna activity is characterized by persistent activity at the

Figure 5.6 Epicenters (A) and hypocenters (B) of the earthquakes recorded on 30 April 1982 at Kilauea volcano. Crosses show the events recorded between 08:54 and 11:41 h of 30 April prior to the caldera eruption. The outline of the caldera (solid lines) and the new eruptive fissures (dashed lines) are shown according to Klein et al. (1987). Data from the catalog of Hawaiian Volcano Observatory.

summit craters (central, northeast, and southeast craters) as well as episodic flank eruptions (Gasperini et al., 1990). Geodynamic behavior of the eastern and western sides of the volcano is different (Gresta et al., 1990). Earthquake swarms preceded and accompanied the majority of flank eruptions. The summit eruptions as usual were not preceded by any seismic crisis (Gresta and Patané, 1987). The author selected three episodes of volcano-tectonic activity that preceded the small 1983 lateral eruption and two major lateral eruptions occurring during 1991 and 2001. The author presented also the 1984 summit eruption as an example of the absence of volcano-tectonic earthquakes before the summit eruptions.

The lateral eruption of 28 March 1983 occurred on the southern flank of the volcano from a short single fracture at the altitude of 2450−2250 m (Gresta and Patané, 1987). It continued for 132 days. About 75×10^6 m^3 of volcanic products (mainly lava) were erupted (Gasperini et al., 1990).

A swarm of volcano-tectonic earthquakes began about 31 h before the eruption and continued for 28 h (Patané et al., 1984; Figure 5.7). It consisted of three impulses. The number of events gradually increased from impulse to impulse during the first 18 h and reached its maximum about 10 h before the eruption. Then the number of events sharply decreased for 6 h; the eruption began when seismic activity ceased.

Seismic foci were distributed below the southeast slope of the volcano at the depth range from 3 km above sea level (a.s.l.) to 15 km below sea level (b.s.l.) but mainly within the volcanic edifice (Figure 5.8). The majority of events were

Figure 5.7 Temporal variations of number of earthquakes before the 28 March 1983 Mount Etna lateral eruption. The number of earthquakes recorded at least by three stations during a 2-h interval is shown. The double arrow shows the beginning of eruption. Data from Patané et al., 1984.

Figure 5.8 Epicenters (A, B) and hypocenters (C, D) of the earthquakes recorded before the 28 March 1983 Mount Etna lateral eruption. (A) The epicentral distribution of the shallowest and smallest earthquakes; (B) the epicentral distribution of the largest earthquakes; (C and D) the foci distribution along the southwest–northeast and northwest–southeast cross sections, respectively.
From Cristofolini et al. (1988).

concentrated in an area along and on the east of the eruptive fissure (Patané et al., 1984). Patané et al. (1984) consider that only the shallowest events within the volcanic edifice can be related to fracturing processes linked to the opening of eruptive fissures.

The eruption of 27 April–16 October 1984. "As usual for subterminal eruptions, no seismic crisis preceded the long 1984 SE crater eruption" (Gresta and Patané, 1987). It produced the total lava volume of 15×10^6 m^3.

The eruption of December 1991–March 1993 was a remarkable event in the recent eruptive history of Mount Etna, considering both its duration of 473 days and the great volume of emitted products (about 250×10^6 m^3). This eruption took place in the southern flank of the volcano (Lombardo et al., 1996; Figure 5.9). On 14 December 1991 at about 2:00 h, the uppermost segment of a 7-km-long radial fracture that was formed in 1989 (Ferrucci and Patané, 1993) began to reopen on the flanks of the southeast crater, accompanied by a swarm of earthquakes (Figures 5.9 and 5.10). The majority of earthquakes (about 200 events with magnitude from 1 to 3.3; 17 of them were located by S. Gresta; Figure 5.9) were recorded during the first 2 h of the propagation of the fissure below the crater and the eastern slope of the volcano at the depth from 0 to 10 km. Then the seismic activity remained at a low level of zero to three events per hour.

Early in the morning of 15 December, eruptive vents opened at the lower termination of the fracture at the height of 2400 m. A slight increase in seismic

Figure 5.9 Epicenters (A) and hypocenters (B) of the earthquakes associated with the 1991 lateral eruption of Mount Etna. Filled circles show the events recorded during the reopening of the 1989 fissures (02:01 h to 18:41 h, 14 December); crosses show the events recorded during the lava flow from the vents (from 05:52 h, 15 December to the end of December). The fissure system on the western slope of the volcano is shown by heavy line; its position on (B) is shown by the double arrow. The position of the largest earthquake of the swarm ($M = 4.4$; catalog of S. Gresta, personal communication, 2002) is shown, as are the contours of the volcanic relief (1000, 2000, and 3000 m).
Data from the catalog prepared by S. Gresta, personal communication, 2002.

activity was observed during the night of 15 December when a sequence of earthquakes occurred below the crater and the western slope of the volcano at the depth of 0–15 km. This hypocentral zone continued its activity up to the end of December. The largest earthquake of the 14–15 December swarm ($M = 4.4$) was recorded during the lava flow from the new vents, 42 h after the beginning of the swarm.

During the 2001–2005 sequence of lateral eruptions at Etna (Allard et al., 2006), the preceding volcano-tectonic earthquake activity was very different. The first, 17 July 2001 lateral eruption was preceded by the 4-day-duration volcano-tectonic swarm including the $M=3.9$ event. The second, 26 October 2002 lateral eruption was preceded by the swarm of only 2 h duration. The third lateral eruption, beginning on 7 September 2004, was not preceded by any significant seismic activity at all.

The lateral eruption of 17 July 2001 occurred along the 6-km-long submeridional system of fissures that were opened on the southern side of the southeast summit crater at about 3000 m a.s.l. (Figure 5.11). It lasted for 24 days and produced a total of 25×10^6 m^3 of lava. In addition, about 5×10^6 m^3 of tephra were emitted, making this an unusually explosive lateral eruption. The eruption occurred on the background of the sequence of 15 Strombolian bursts occurring from 7 June to 17 July (Alparone et al., 2004; Allard et al., 2006).

Figure 5.12A shows the temporal variations in volcano-tectonic activity during June–July 2001 recorded by the seismic network of the National Institute of Geophysics and Volcanology, Section Catania (Patanè et al., 2003). The threshold of the catalog is $M \geq 1.5$. The epi- and hypocenters of these seismic events are shown in Figure 5.11. It is seen that during the Strombolian bursts activity the

Figure 5.10 Temporal variations of the earthquakes associated with the 1991 lateral eruption of Mount Etna. The arrow shows the onset of the largest earthquake ($M = 4.4$; catalog of S. Gresta) of the swarm. Data from the catalog of Ferrucci and Patané, 1993.

Figure 5.11 Epicenters (A) and hypocenters (B) of the earthquakes associated with the 2001 lateral eruption of Mount Etna. The largest earthquake ($M = 3.9$) epicenter is shown in (A) with the larged circle. Its projection is shown also in Figure B with the arrow. The fissure zone is shown according to Alparone et al. (2004).
Data from the catalog of the National Institute of Geophysics and Volcanology, Section Catania.

daily number of volcano-tectonic earthquakes was between 0 and 7. Seismic activity sharply increased on 12 July, reaching 173 events on 13 July; one of them was the largest of the seismic swarm events with magnitude 3.9 (it occurred in 5 h after the beginning of the swarm; Figure 5.12). During 14–18 July, the number of earthquakes gradually decreased, and the lateral eruption began on 17 July from six eruptive fissures, accompanied by strong explosive activity and destructive lava flows. After the beginning of eruption and until the end of

Figure 5.12 Temporal variations of the earthquakes associated with the 2001 lateral eruption of Mount Etna. (A), All earthquakes recorded during June–July 2001; (B), earthquakes recorded during 12–17 July 2001 within the eastern and western epicentral areas. The single arrow shows the onset of the largest earthquake ($M = 3.9$) of the swarm and the double arrow shows the beginning of eruption.
Data from the catalog of the National Institute of Geophysics and Volcanology, Section Catania.

eruption on 9 August, daily seismicity stayed at low level (zero to two events with magnitude $M \geq 1.5$).

Seismicity associated with the background Strombolian activity, recorded from 7 June to 17 July, was located along the broad east-northeast–west-southwest zone going through the summits of Etna and distributed at the depth range from 5 to 20 km b.s.l. (Figure 5.11). The preliminary 12–17 July swarm of volcano-tectonic

earthquakes clustered into two compact groups of events, western and eastern, separated by submeridional fissure zone. Figure 5.12B shows that the activity began on the eastern side but 1 day later, the western and eastern groups of earthquakes developed simultaneously. At the same time, the events of the eastern group were more numerous. The foci of both groups of events were distributed at the depths from 5 km b.s.l. to 2 km a.s.l., mainly within the volcanic edifice. The largest $M=3.9$ earthquake was located between these two groups of earthquakes within the fissure zone, at the depth of the sea level. The earthquakes, recorded during the lateral eruption at the depths from 5 km b.s.l. to 2 km a.s.l., were located within the eastern cluster and on its continuation in a south-southeast direction. No visible fissures were reported for this direction, but perhaps they were formed at the depths without opening on the surface.

5.2.2 Oshima Volcano, Izu Islands

Oshima volcano is a basaltic stratovolcano that forms an island in Sagami Bay, about 100 km south-southwest from Tokyo. The recent eruption of Oshima volcano started on 15 November 1986 within the central cone of Oshima caldera (Mihara-yama). A new cone, A, was formed on the wall of the caldera producing a lava flow. On 21 November, a fissure opened on the caldera floor at the northeast foot of Mihara-yama (vent B); 1.5 h later, a new series of vents (vents C) were opened outside the caldera rim, on the northern slope of the volcano. The activity of new

Figure 5.13 Epicenters of the earthquakes located before the 15 November 1986 summit eruption at Oshima volcano (A, from 20 April to 15 November 1986), during the eruption (B, from 15 to 20 November), and just before the 21 November 1986 fissure eruption (C) of Oshima volcano.
From Yamaoka et al., 1988.

craters continued with some interruptions up to 18 November 1987 (Aramaki, 1988).

Volcano-tectonic earthquakes were recorded during all stages of this 1-year sequence of summit-flank eruptions (Watanabe, 1988; Yamaoka et al., 1988). Figure 5.13 shows the epicenters of earthquakes located before the summit eruption of 15 November (A), during this eruption (B), and just before the lateral fissure eruption of 21 November (C).

Before the beginning of the summit eruption, the seismic activity of the caldera region was very low and stayed at the same level during the eruption. No preceding swarm was recorded. In contrast, 2 h before the fissure eruption, a swarm of very shallow (depth < 5 km) earthquakes began at the northern part of the caldera, and the seismic foci migrated toward the north (Figure 5.13C). The chain of craters B and C opened along the western part of the swarm area.

A very intensive swarm accompanied fissure eruptions (Figures 5.14 and 5.15). It occupied large area of about 40 km^2. The earthquake foci were distributed at shallow depths (0−5 km) below the island and formed clusters at the depths from 5 to 10 km to the southeast (about 10−15 km from the volcano) and to the northwest (about 5−10 km from the volcano). Yamaoka et al. (1988) noted that the foci of the events located in these two clusters propagated during the first day of the fissure eruption simultaneously to the southeast and northwest directions from the eruptive site with the velocity of 11−16 m/min. These two hypocentral zones were active up to the end of March 1987. The largest earthquake of the swarm ($Mw=5.9$) occurred on 22 November, about 18 h after the beginning of fissure eruptions, off the southeast coast of the island.

5.2.3 Klyuchevskoy Volcano, Kamchatka

Klyuchevskoy stratovolcano is the most active basaltic volcano of the Kurile−Kamchatka volcanic zone. It is the highest active volcano in Eurasia; its altitude is 4750 m. Its activity is represented by paroxysmal and small summit eruptions and frequent flank eruptions. The summit eruptions of Klyuchevskoy occur with the formation of cinder cones within the summit crater followed by explosions and lava flows.

The paroxysmal eruption of 1945 was preceded by significant seismic activity; the shocks were felt at a distance of 32 km from the volcano a few hours before the eruption of 1 January (Piip, 1956). The small summit eruptions that occurred during 1949−1986 were not preceded by seismic activity that could be recorded at a distance of 16 km where the nearest seismic station was installed; only volcanic tremor was observed before and during the summit eruptions (Tokarev, 1966; Khrenov et al., 1991).

The 12 lateral (flank) eruptions that occurred on the flanks of Klyuchevskoy from 1932 to 1987, at the altitudes of 450−3900 m (Khrenov et al., 1991), were preceded by seismic activity. We discuss here the seismic activity associated with two flank eruptions, in 1951 and 1966.

Figure 5.14 Seismic activity accompanied the fissure eruption of Oshima volcano from 16:15 on 21 November to 30 November 1986. (A), epicenters of earthquakes; (B), hypocenters of earthquakes. I–II is the zone of cross section of B.
From Yamaoka et al., 1988.

The 19 November 1951 Bylinkina flank cone eruption occurred at the height of 950 m forming two new cones followed by lava flow (Tokarev, 1981). Seismic activity began 6 days before the beginning of the eruption (Figure 5.16). The number of earthquakes increased rapidly during the first 30 h, and then gradually decreased for 3½ days. The strongest earthquake with magnitude 3.1 occurred in

Figure 5.15 Temporal variations of epicenters from 21 to 22 November in the region shown in Figure 5.14A. The focal depths are shown by circles (0–5 km), triangles (5–10 km), and squares (>10 km). The difference in magnitude is shown by the difference in the size of the symbol. Arrows B and C denote time of the B and C craters eruptions.
From Yamaoka et al., 1988.

Figure 5.16 Temporal distribution of 12-h number of the earthquakes ($M > 1$) prior to the 19 November 1951 eruption of the Bylinkina flank cone of Klyuchevskoy volcano, recorded at the seismic station KLY (see Figure 5.18), at a distance of 22 km. The moment of the largest event with magnitude $M = 3.1$ is shown.
Data from Tokarev, 1981.

2 days after the beginning of seismic activity. The eruption began 1 day after the ceasing of the seismic activity. Only one station KLY was in operation, so no locations were made.

The 6 October 1966 Piip flank cone eruption occurred at the altitude of 1900–2100 m with the formation of a sequence of explosion craters (Tokarev et al., 1968). Seismic swarm began 5 days before the eruption with three magnitude 3.5–3.7 events located at depths of about 20 km during the first hour of activity. All remaining events were very shallow. Thus, the epicentral zone was divided into two parts, a southeast deep zone and a northwest shallow zone, with the site of eruption situated between them (Figure 5.17).

Figure 5.17 Epicenters (A) and hypocenters (B) of earthquakes ($M > 2$) located before the 5 October 1966 eruption of the Piip flank cone of Klyuchevskoy volcano. Star shows Klyuchevskoy volcano; cross shows the site of the new eruptive center. Two nearest seismic stations are shown by triangles.
From the Kamchatka Regional Earthquake Catalog.

Two peaks in the seismic activity were recorded during the first 3 days of the swarm (Figure 5.18). Then the number of events decreased rapidly, and the eruption occurred above the background of the termination of seismic activity.

5.3 Volcano-Tectonic Earthquakes Associated with Fissure Eruptions

A fissure eruption is an eruption that takes place from an elongate fissure, rather than from a summit vent (Schmidt and Schmincke, 2000). The most famous recent fissure eruptions were at Parícutin in Mexico (1943–1952), Krafla in Iceland (1975–1983), and New Tolbachik volcanoes in Kamchatka (1975–1976). Well-documented episodes associated with dyke injections along rift zones occurred during 2005–2009 in the southernmost Red Sea rift near the Afar triple junction.

Figure 5.18 Temporal distribution of 12-h number of earthquakes ($M > 1$) prior to the 6 October 1966 eruption of the Piip flank cone of Klyuchevskoy volcano recorded at the seismic station KLY (Figure 5.17) at a distance of 25 km. The moment of the largest event with magnitude $M = 3.7$ is shown.
Data from Tokarev, 1981.

5.3.1 New Tolbachik Volcanoes, Kamchatka

A chain of new volcanoes was formed from 6 July 1975 to 10 December 1976 along the 12-km-long fissure that was opened at a distance of about 18 km southwest from Plosky Tolbachik volcano (Figure 5.19). The new eight volcanoes were basaltic cinder cones with the height up to 300 m; 1.2 km^3 of lava was produced by the cones, which covered an area of about 45 km^2. The total volume of pyroclastic material was about 1 km^3.

The preceding swarm of earthquakes (73 events with magnitude $M > 2$) began on 27 June. It is possible to discriminate two stages in the seismic activity (Figure 5.20): from 27 June to 30 June and from 1 July to 5 July. During the first stage, a sharp increase in the number of events was observed on 27 June, and then the number of events gradually decreased for 3 days. The epicenters were distributed within an ellipse-shaped area just to the north from the site of forthcoming eruption of the cone I (Figure 5.19A) and covered the space where the following six cones appeared. The depths of earthquakes varied from 15 km to the surface (Figure 5.19B).

The second impulse in seismic activity was characterized by lower number of events and ceased 1 day before the 6 July eruption. The epicenters occupied more compact area than during the first stage, nearer to the forthcoming eruption site. They were shallower and were distributed within the depth range from 8 km to the surface. Two largest earthquakes of the swarm (magnitude $Mw = 4.9$) were observed on 2 July at a depth of 11 km. Their epicenters were located below the cones II and III. During the following eruptive activity with the formation of the new cones in August–September 1975, a few seismic sequences were observed (Figure 5.21). One sequence (11 events with $M > 2$) occurred from 6 to 8 August and preceded the birth of the cone II on 9 August. The epicenters were distributed within the southern part of epicentral zone of the first swarm of 26 June–6 July, on the level of the forthcoming site of the cone II. The next sequence of small earthquakes (35 events with $M > 2$) occurred from 10 to 14 August, before forming the third and the fourth cones on 17 August. These seismic events connected the new eruptive zone with the old basaltic Plosky Tolbachik volcano (Figure 5.21).

Figure 5.19 Epicenters (A) and hypocenters (B) of the earthquakes ($M > 2$) preceding the 6 July 1975 eruption of the first cone of New Tolbachik volcanoes. The epicenters of the events recorded from 26 to 30 June are shown by filled circles; the events recorded from 1 to 6 July are shown by crosses. The largest earthquake is shown by diamante. Active volcano Plosky Tolbachik [PT in (B)] is shown by large star; new cones I, II, II, and VIII are shown by stars.
Data from the Kamchatka Regional Earthquake Catalog.

Figure 5.20 Temporal variations of daily number of earthquakes ($M > 2$) preceding the 6 July 1975 eruption of the first cone of New Tolbachik volcanoes. The moment of the largest event is shown.
Data from the Kamchatka Regional Earthquake Catalog.

The caldera collapse, occurring at Plosky Tolbachik in the middle of August (Dvigalo et al., 1984), marked a clear relation between the old volcano and the new active cones.

The last seismic swarm (12 events with magnitude $M > 2$) was recorded from 16 to 18 September 1975, preceding the birth of the cone VIII (Figure 5.21). The location of these events continued the active line from Plosky Tolbachik

Figure 5.21 Epicenters of earthquakes ($M > 2$) that were observed during the June–September 1975 eruption of the new Tolbachik cones. Four groups of events are shown: 26 June–6 July (crosses); 6–8 August (diamonds); 10–14 August (black triangles); and 16–18 September (open triangles). Cones I–VIII are shown by gray stars; active volcano Plosky Tolbachik is shown by black star. The fissure zone is shown by dashed line. Data from the Kamchatka Regional Earthquake Catalog.

volcano through the north zone of cones I–VII (they finished their activity on 15 September) to the south up to the site of the cone VIII that was born on 18 September.

The total process of migration of epicenters along the 30-km line Plosky Tolbachik—Cone VIII from 6 August to 18 September, is seen in Figure 5.22. During the initial part of the post-I-cone seismicity (6–8 August), the epicenters were clustered on the position of the forthcoming cone II. Then, after forming cone II on 9 August, a northeast migration of foci was observed to Plosky Tolbachik volcano, that continued from 10 to 14 August and coincided with the caldera collapse of Plosky Tolbachik. The last southwest focal movement was from the site of the cones I–VII to the site of the forthcoming cone VIII.

After the end of activity of the eighth cone on 15 December 1976, seismic events began to spread along the line of new volcanoes, to the north and south from the active centers (Figure 5.23). During 2 weeks, a line of 90-km length was outlined by the epicenters. In 1977 and 1978, seismic activity was recorded at the periphery of the active volcanic zone at a distance of 40–60 km from the new volcanoes.

Figure 5.22 Migration of earthquake foci ($M > 2$) along the NE line Plosky Tolbachik volcano—Cone VIII from 6 August to 18 September. Stars along the latitude axis show positions of the new cones (I, II, III, VIII) and Plosky Tolbachik (PT) volcano. Data from the Kamchatka Regional Earthquake Catalog.

5.3.2 The 2005–2009 Ethiopia Rifting Episode

The 2005 Dabbahu, Ethiopia, rifting episode occurred in the southernmost Red Sea rift near the Afar triple junction (Figure 5.24). Afar region was affected by seismotectonic crisis during 1978 and 1989 (Ayele et al., 2007). Dabbahu is a Holocene volcanic massif forming an axial range of the Afar depression. The obsidian flows, lava domes, and pumice cones form the summit and upper flanks of the volcano, which was built over a base of basaltic-to-trachytic lava flows of a shield volcano. Late-stage Holocene basaltic fissure eruptions occurred at the northwest base of the volcano. Abundant fumaroles are located along the crest of the volcano and extend northeast. The first historical eruption of Dabbahu volcano took place from 26 to 30 September 2005 from a fissure vent on the northeast flank of the volcano. It was a small explosive eruption that produced ash fall deposits. The 2005 event ruptured the length of a magmatic segment within the Dabbahu–Manda Hararo segment of the Manda Hararo rift zone. It was preceded by at least two preliminary earthquake swarms recorded in 1978 and in 1989 (BGVN (1990–2011); Ayele et al., 2007; Ebinger et al., 2010).

The seismic crisis began on 4 September 2005 (Ayele et al., 2009). Activity subsided but then restarted on 20 September and continued until 4 October (Ebinger et al., 2010). The 2005 epicentral zone was developed between AVC and DVC (Figure 5.24). Two impulses of seismic activity were observed: from 20 to 23 September and from 23 to 26 September (Figure 5.25). Nine earthquakes with magnitudes Mw between 5.0 and 5.5 (USGS, 2011; Harvard CMT, 2011) were recorded. The largest $Mw = 5.5$ event occurred on 24 September at the peak

Figure 5.23 Post-eruption (from 16 December 1976 to 1978) seismic activity in the region of New Tolbachik volcanoes. (A), Epicentral distribution of earthquakes; (B), Hypocentral distribution of earthquakes. Open diamonds are earthquakes recorded during active period (27 July 1975–15 December 1976); filled circles are earthquakes recorded from December 16, 1976, to the end of 1978. Five-ray stars show new Tolbachik volcanoes; active volcanoes of Klyuchevskaya group (KL, Klyuchevskoy; BZ, Bezymianny; and PT, Plosky Tolbachik) are shown by six-ray stars. NT is site of New Tolbachik volcanoes.
Data from the Kamchatka Regional Earthquake Catalog.

Figure 5.24 Epicenters of earthquakes in the Dabbahu–Manda Hararo segment from 19 October 2005 to 2010, with a 6-month gap between October 2006 and March 2007. Squares are Seis-UK seismic stations operational during 2005–2006 and IPGP seismometers deployed in 2007; triangles are continuous GPS sites. The composite digital elevation model is constructed from Spot 5 imagery; vertical resolution <10 m. Ado'Ale volcanic complex (AVC) is a chain of rifted silicic centers at the segment center. Semera is the Afar region's capital. Stars show the sites of the 26 September 2005 and the 14 August 2007 eruptions. The epicenter of the largest $Mw = 5.4$ 20 September 2005 earthquake is shown by cross. Its Harvard CMT focal mechanism is shown.
From Ebinger et al., 2010, with modifications.

Figure 5.25 Temporal distribution of earthquakes before the 26 September 2005 eruption during the Dabbahu rifting episode. The single arrow indicates the moment of maximum earthquake; the double arrow indicates the beginning of magma intrusion.
Data from the catalog of U.S.G.S. http://earthquake.usgs.gov/earthquakes/eqarchives/epic/.

activity of the second seismic impulse. The 26 September eruption began when the number of earthquakes decreased (Figure 5.25). The September 2005 eruption vent was situated approximately 5 km east-northeast of the Dabbahu volcano summit (Figure 5.24) (Ayele et al., 2007).

Subsequent fissure basalt intrusions, displayed by continuous earthquake swarms (Figure 5.24), were recorded during June 2006–June 2009 along the Dabbahu–Manda Hararo rift segment, in its central and southern parts (Ebinger et al., 2010). The site of the 14 August 2007 mafic eruption in the Karbahi graben south of AVC (Ayele et al., 2009) is shown in Figure 5.24.

Along the 120-km-long rift segment, 420 from the 2005–2009 earthquake hypocenters clustered between 2 and 12 km. The largest of them show rift-normal opening (Ayele et al., 2009; see also Figure 5.24). The post-dyking earthquakes mimic the geodetically determined distribution of the 2005 dyke (Ebinger et al., 2008).

5.4 Volcano-Tectonic Earthquakes Associated with Caldera Collapse

A caldera is a gigantic basin with steep walls at the summit of a volcano; larger than a crater. It is usually formed by collapse (Decker and Decker, 1999). The multiple stages of collapse following evacuation of underlying magma chamber are characteristic for Kilauea (Hawaii) and Fernandina (Galapagos Islands). The great 1883 eruption of Krakatoa (Indonesia) and the 1912 eruption of Katmai (Alaska) are the best geologically documented examples of caldera-forming events.

5.4.1 Fernandina Volcano, Galápagos Islands

Fernandina is a large basaltic shield volcano on the western edge of Galápagos platform (Figure 5.26). At its summit, some 1500 m a.s.l., an elliptical caldera of $4 \times 6.5\ km^2$ of area is situated. On 21 May 1968, the eruption started, and a lava flow was observed on the east flank of Fernandina. A sequence of strong explosions began on 8 June and then declined on 12 June. The aerial reconnaissance of 21 June revealed that the 8-km^2 caldera floor subsided in a series of short drops along an

Figure 5.26 Epicenters of the 1968 Galapagos swarm earthquakes located by NOS network. Large distance of the nearest stations from the epicenters (about 400 km) produced large error in locations; therefore, great area of distribution of earthquakes may not reflect the real situation.
From Filson et al., 1973.

elliptical boundary fault. The maximum subsidence was approximately 350 m with a resulting collapse volume of 2 km^3 (Filson et al., 1973; Francis, 1974).

Minor seismic activity ($Ms < 4.2$) was recorded during the first week of June (Figure 5.27B). On 12 June, sharp increase in seismic activity began and reached its maximum on 19 June. Then the number of earthquakes declined rapidly, and the swarm practically stopped after 23 June. This seismic swarm was characterized by a great number of high-magnitude events that had practically the same form (Filson

Figure 5.27 Distribution of maximum magnitudes *Ms* (A) and daily number of earthquakes (B) from 20 May to 30 June 1968.
Data from Filson et al., 1973.

et al., 1973; Kaufman and Burdick, 1980). Seismic activity developed in two stages. The first stage was associated with the eruptive activity of the volcano (20 May−11 June). The maximum magnitude of events recorded during this stage was 4.2. The second stage began on 12 June, when the magnitude and number of events began to sharply increase. The maximum daily magnitudes 5.1−5.2 were recorded during 4 days, 12−15 June (Figure 5.27A). The total number of events with magnitude $Ms \geq 5.0$ was 5 (Filson et al., 1973), which is very large for any volcanic earthquake swarm. Filson et al. (1973) suggested that the caldera collapse occurred during the second stage of seismic activity, and the earthquakes were resulted from the dropping of the caldera block. All epicenter locations were made by the global seismic network (Figure 5.26) that produced large errors in location. It made it difficult to do any tracing of epicenters and comparison with the volcanic structures.

5.5 Volcano-Tectonic Earthquakes Associated with Submarine Eruptions

Submarine eruptions vary between its birth on the deep seafloor and its emergence above sea level. Evidence of deep submarine eruptions is rarely visible at the surface because the steam they produce is quickly condensed as it rises through cold seawater. At the same time, if the eruption is vigorous and the vent is within 500 m or so to the surface, the water may become turbulent and cloudy (Bardintzeff and McBirney, 1999). The best-documented seismological examples are two eruptions occurring on the volcanic chain of Izu islands, Japan: the 1989 Teishi Knoll eruption and the 2000 Miyakejima eruption.

5.5.1 Teishi Knoll Volcano, Izu Islands

On 30 June 1989, an earthquake swarm started off eastern Izu Peninsula in Japan (Figures 5.28 and 5.29). This was then culminated in the submarine eruption on 13 July. Hourly numbers of earthquakes (Figure 5.30) that were observed at the station KMT, situated at a distance of 6.8 km from the eruption site, show (Figure 5.28; Yamasato et al., 1991) that three main stages in development of the swarm were observed: (1) from 30 June to 4 July; (2) from 4 July to 8 July; and (3) from 8 July to 13 July.

During the stage 1, seismic activity was very low; a great increase in the number of events began from 4 July, and then gradually decreased during the next 3 days (stage 2). The next large impulse in seismic activity (stage 3) began on 8 July and reached its maximum on 9 July, when the $M=5.5$ earthquake, the largest in this sequence, occurred. Then the seismic activity gradually declined and the eruption began on 13 July with a low background level.

Figure 5.28 demonstrates that regularization of the epicentral zone of events during the swarm was observed. During the first and the second stages, the epicentral zone had an elliptical form (Figure 5.28B). At the third stage, the epicenters

Figure 5.28 Spatial-temporal development of the 1989 Teishi Knoll earthquake swarm. (A), Seismic activity of the zone of eruption before the swarm; (B), seismicity during the first and the second stages of the swarm; (C), seismicity during the third stage of the swarm; (D), seismicity just before the eruption; (E), seismicity during the eruption. Inset in (A), shows the area of study in (B–E). Cross in (D), shows the site of eruption. KMT is a seismic station. From Yamasato et al., 1991.

Figure 5.29 Depth distribution of the 1989 Teishi Knoll earthquake swarm events along the eastwest direction. Black circle shows the site of eruption.
From Matsumura et al., 1991.

clustered in a lineal submeridional zone including the site of forthcoming eruption (Figure 5.28C and D). After the beginning of the eruption, seismic events continued to cluster within this lineal zone. The majority of seismic events of the swarm were located within the depth range from 5 to 10 km. Two hypocentral subzones were formed: the initial large subzone just below the eruption site that was active during

Figure 5.30 Temporal variations of the number of earthquakes during the 1989 Teishi Knoll swarm according to the data of the station KMT (see Figure 5.28D).
From Yamasato et al., 1991.

stages I and II, and a narrow smaller size subzone that was formed during stage III of the swarm activity. A thin lineal zone of a few events connects the large hypocentral subzone with the site of eruption (Figure 5.29).

5.5.2 Miyakejima Volcano, Izu Islands

Miyakejima is an active volcanic island consisting of a basaltic stratovolcano with a 9-km basal diameter and a maximum elevation of about 750 m (Figure 5.31). The volcanic island has been active for more than 10,000 years. After 17 years of quiescence, on the evening of 26 June 2000, small volcanic earthquakes began to be recorded at the stations west of the summit. Seismic activity developed into two successive swarms of volcano-tectonic earthquakes associated with two eruptive events: the flank submarine eruption and the following caldera collapse in the crater of the volcano. The swarms continued for about 3 months. Figure 5.31 shows the main seismo-volcanic elements of the eruption: the site of the submarine eruption, the area of the collapsed caldera, and the area of the earthquake swarm. The development of the June–August 2000 seismic activity is shown in Figures 5.31 and 5.32 (Geshi et al., 2002; Nakada et al., 2005).

Seismicity associated with the submarine eruption. The first swarm of volcano-tectonic earthquakes began beneath the summit at the depths from 1 to 4 km

Figure 5.31 The position of Miyakejima volcano in Izu Islands (A) and the main seismo-volcanic elements of the 2000 eruption (B). Outline of the caldera formed during the 2000 eruption, the location of submarine eruption on 27 June, and the approximate area of the earthquake swarm during July and August are shown.
From Geshi et al., 2002.

Figure 5.32 The 2000 Miyakejima earthquake swarm evolution in map view (top panels) and cross sections (bottom panels). Arrows show the epicenters and hypocenters of the $Mw > 5.7$ earthquakes (1 July, $Mw = 6.2$; 9 July, $Mw = 5.9$; 15 July, $Mw = 6.0$; 30 July, $Mw = 6.4$; and 18 August, $Mw = 5.7$). Earthquake magnitudes were estimated by NIED, Tsukuba.
From Toda et al. (2002).

Figure 5.33 Epicenters of the Miyakejima earthquakes (filled circles) occurring from 26 to 27 June 2000 and the submarine eruption site (A). Craters of old eruptions and new submarine craters are shown by small open circles. Radial dyke patterns are shown also by lines with small open circles. The multibeam sides-scan sonar images of the new craters formed during this eruption are shown in B.
From Kaneko et al., 2005.

(Figure 5.32A) on 26 June 2000 but the summit eruption did not occur. Hypocenters of earthquakes moved westward indicating intrusion of a dyke-oriented east-northeast−west-southwest and originating at the west coast of Miyakejima. A flank submarine eruption (Figure 5.33) occurred on the morning of 27 June, coinciding with a sharp increase in the number of earthquakes (Figure 5.34), probably soon after the seismic swarm passed beneath the eruption site. Four craters

Figure 5.34 The rate of caldera volume growth (A) and the daily variations in the number of Miyakejima earthquakes (B). Appearance of large ($M > 6$) earthquakes is shown by red diamonds.

From Nakada et al., 2005.

of 20−30 m diameter aligned east-west in a 100 × (10−20) m^2 area on the seafloor were revealed (Figure 5.32B) (Kaneko et al., 2005; Nakada et al., 2005).

The earthquake activity continued westward, filling the space between Miyakejima and Kozushima islands and between the depths from 0 to 15 km (Figure 5.32A). The activity resumed on 28 June (Figure 5.34). The single $Mw = 6.2$ (NIED) earthquake occurred on the western end of the epicentral area on July 1.

Seismicity associated with the caldera collapse. On 8 July, the collapse of the summit area suddenly began being accompanied by a small explosion. Subsidence of the summit area continued through early August, simultaneously with the intense second earthquake swarm. The swarm occurred at the depths between 0 and 20 km located between Miyakejima and Kozushima islands and to the north from Kozushima island (Figure 5.32 B−D), and including four strike-slip $Mw > 5.7$ (NIED) earthquakes. The 9 July earthquake ($Mw = 5.9$) occurred at the same place as the first July event from the first swarm. Three other earthquakes occurred at the ends of the swarm area ($Mw = 6.0$, 15 July, and $Mw = 6.4$, 30 July) and in its central zone ($Mw = 5.7$, 30 August). The swarm gradually declined in the second half of August (Nakada et al., 2005; Uhira et al., 2005). A collapsed caldera, 1.6 km in diameter and 450 m in depth, was formed. Its final volume was about 6×10^8 m (Geshi et al., 2002). During subsidence, phreatomagmatic explosions occurred on 14 and 15 July. Nakada et al. (2005) suppose that the subsidence occurred due to lateral migration of magma under the ground.

Toda et al. (2002) discussed the model of seismicity compatible with a laterally propagating dyke intrusion during this eruption. They assumed that a vertical dyke propagated during the first week (Figure 5.32A), and then opened continuously for 7 weeks. The geometry of the inferred dyke proposed by Toda et al. (2002) is shown in Figures 5.32C and D. They consider that the process of dyke expansion and associated "dog-bone" pattern of seismicity was controlled by stress transferred from an expanding dyke to vertical faults in the surrounding crust with the mean stressing rate of 32 bars/year.

6 Volcano-Tectonic Earthquakes at Andesitic Volcanoes: Case Studies

Andesitic volcanoes produce lava with viscosity in the range 10^3-10^5 poises at temperatures of about 1000–1300°C; the volcanic activity is of explosive and effusive types; they are characterized by higher level of volatiles. Andesitic volcanoes usually have the form of strato-volcanoes. The most famous among them are Bezymianny in Kamchatka, Merapi in Indonesia, and Volcán de Colima in Mexico.

The eruptions of andesitic volcanoes occur mainly from the summit crater of the volcano and rather seldom on its flanks. The magmas are rich in volatiles, mobile, able to produce phreato-magmatic or phreatic explosions, and can form lava domes in the crater of the volcano. These eruptions are accompanied by lava and pyroclastic flows and numerous rockfalls. The normal cycle for andesitic volcanoes consists of the formation of a lava dome and then its destruction with its collapse or an explosion. The rock fracturing during the formation of the passageway for the magma to the surface produces volcano-tectonic earthquakes. Lava effusion from the summit and flank (fissure) vents may be accompanied by volcanic tremor.

The author discusses here the volcano-tectonic seismicity associated with three types of activity of andesitic volcanoes: seismic activity related (1) to the great "directed blasts" at volcanoes; (2) to the phreatic or phreatio-magmatic explosions; and (3) associated with block-lava extrusions. These three types of eruptions may form also subsequent stages of the eruptive activity of an andesitic volcano.

6.1 Volcano-Tectonic Earthquakes Associated with Volcanic "Directed Blasts"

The "directed blasts" of andesitic and dacitic volcanoes are within the most "large-scale" eruptive events. During the eruption of this type, a great part of volcanic cone can be exploded horizontally and end up distributed over a great area (Gorshkov and Bogoyavlenskaya, 1965). Typical eruptions: Bezymianny, Kamchatka, 1956; Sheveluch, Kamchatka, 1964; Mount St. Helens, Cascades, USA, 1980.

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00006-2
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6.1.1 Bezymianny Volcano, Kamchatka

The 30 March 1956 explosion of Bezymianny volcano became a classic example of a "directed blast" eruption (Gorshkov and Bogoyavlenskaya, 1965). This strato-volcano with a height of 3085 m was considered as dormant, but on 29 September 1955, the first earthquake in the region of the volcano was recorded. From October 6, a swarm of volcano-tectonic earthquakes (Figure 6.1) began, and the first explosion was recorded on 22 October 1955. Strong ash eruption continued for 1 month, and then it changed to moderate explosive activity that culminated in paroxysmal explosion of 30 March 1956. This explosion destroyed the SE part of the volcanic cone and pyroclastic material was erupted to the SE at the angle of 30−40° to the horizon. Ash cloud rose up to 40 km. An area of 500 km^2 at a distance of 30−32 km to the southeast of the volcano was covered by about 3 km^3 of erupted material. The height of the volcano was diminished by about 180 m. Later, during

Figure 6.1 Daily distribution of maximum magnitudes (A) and the number of earthquakes (B) associated with the October 1955−June 1956 eruption of Bezymianny volcano (station KLY, 42 km from the Bezymianny volcano). St 1, St 2, and St 3 indicate the three stages in the development of activity. Arrows show the appearance of the first and largest explosions.
Data from Tokarev, 1981.

April 1956, a new dome began to grow in the crater. This process of the formation and destruction of the dome continued for many years; after 1965 it was accompanied by extrusion of block lava (Bogoyavlenskaya and Kirsanov, 1981; Bogoyavlenskaya et al., 1991).

Figure 6.1 shows the development of seismic activity during 1955–1956. The only seismic station KLY was situated at the distance of 42 km from the volcano and was equipped by short and mediate period three-component seismometers with magnification of 5000 and 300, respectively. It failed to localize the events but it was used to count the number of events and to estimate their magnitude. One month before the eruption, the number and energy release of seismic events gradually increased up to the first explosion. During the next stage, when moderate explosions were observed, the number of events gradually decreased during November and stayed at a stable low level (about 10 events daily) during December 1955–March 1956. The daily earthquake magnitude level during this stage was rather high, reaching 3.5. The paroxysmal explosion of 30 March occurred on the background of the low level in seismic activity and was associated with the magnitude *Mw* 5.3 earthquake. The postexplosion stage of dome-building was characterized by a gradual decrease in the number and magnitude of seismic events.

Gorshkov and Bogoyavlenskaya (1965) noted that the variations in the number of seismic events correlated well with the manifestations of eruptive activity while the variation in the maximum magnitude of earthquakes had no such dependence. Larger earthquakes formed a permanent background from 18 October 1955 (before the eruption) to the end of February 1956; their occurrence did not depend on the eruptive manifestations.

After the paroxysmal explosion, the growth of lava dome in the crater was observed. The partial destruction of the dome by explosions was preceded by an increase in seismic activity 2–5 days before the eruption (Tokarev, 1966, 1981; Bogoyavlenskaya et al., 1991).

Two typical variations of seismic activity before the explosions of the dome during 1959 and 1960 are shown in Figure 6.2. The number of events was counted at seismic station KLY (Tokarev, 1981). During the 1959 event, an increase in seismic activity began on 26 March (Figure 6.2A). The number of events sharply increased up to the moment of the 28 March explosion. After the explosion, the number of events gradually decreased. The majority of seismic events had magnitudes from 0 to 1. The maximum event with magnitude 1.3 was recorded on 28 March.

A similar situation was observed before the 13–14 April 1960 eruption (Figure 6.2B). An increase in seismic activity began 3 days before the explosion; the explosion occurred at the maximum level of seismic activity. After the explosion, the number of events gradually decreased. The majority of seismic events had magnitudes from 0 to 1. The maximum event with magnitude 1.7 was recorded on 12 April.

6.1.2 Sheveluch Volcano, Kamchatka

Sheveluch volcano is the northernmost active andesitic volcano of Kamchatka. It is located at the junction of the Kurile-Kamchatka and Aleutian subduction zones.

Figure 6.2 Variations in the number of earthquakes before two explosions of Bezymianny volcano that occurred during 1959 (A) and 1960 (B) and were observed at KLY station at a distance of 42 km from the Bezymianny volcano.
Data from Tokarev, 1981.

It consists of two parts: the old edifice with the altitude of 3283 m and the young (active) edifice with the altitude of 2800 m. During the last 10,000 years, the volcano has been characterized by two types of eruptions: the great eruptions of the "directed blast" type (they have continued from hours to a few days) and the slight to moderate eruptions associated with the growth and destruction of the crater dome (they have continued for several years) (Bogoyavlenskaya et al., 1991).

1964 eruption. The "directed blast" eruption took place on 12 November 1964 after 15 years of calm. It continued for only 70 min. As a result, about 1.5 km^3 of resurgent material and 0.8 km^3 of juvenile material were ejected; the summit crater and five periphery domes were destroyed by the explosion. A new crater was formed. Figure 6.3 shows the seismic history of the eruptive process (Tokarev, 1967; Gorshkov and Dubik, 1970).

The first alert pronouncing the volcanic unrest was initiated by the April–May 1964 swarm of 37 small shallow earthquakes located near the volcano. This swarm was not followed by any volcanic manifestations. The next swarm of

Figure 6.3 Characteristics of seismic activity at Sheveluch volcano in 1964. The strain release during the seismic activity (as a cumulative value of quadratic root of seismic energy in erg) and variations in the number of earthquakes according to the records at seismic station KLY situated at a distance of 45 km to the south of the volcano are shown. (A–C) The stages in seismo-volcanic activity: (A) the preliminary stage (January to the middle of October 1964); (B) the preliminary seismic swarm (17 October to 12 November); (C) the paroxysmal eruption on 12 November; and (D) the posteruptive stage.
From Tokarev, 1967.

volcano-tectonic earthquakes began on 17 October (Figure 6.4). About 500 earthquakes with magnitude $M\ 2 = 5$ were recorded. Epicenters of all the earthquakes were located by three seismic stations of Klyuchevskaya Volcano Station and occurred within the volcanic edifice at the depths of less than 5 km. Number and energy of earthquakes (according to the data of the nearest seismic station KLY situated at the distance of 45 km to the south of the volcano) gradually increased up to the moment of the paroxysmal explosion of 12 November. The maximum magnitude earthquakes were associated with the 12 November paroxysmal eruption; their magnitudes *Mw* were 5–5.2 (Tokarev, 1967; Zobin, 2001). The airwaves of the 12 November explosions were recorded by barographs and microbarographs at the distance up to a few thousand kilometers. After the 12 November explosion, seismic activity sharply decreased and ceased in 4 days.

1993 eruption. A new dome began to grow during 1980 within the crater that was formed by the 1964 explosion, without any preceding seismic events. Its 1980–1993 growth was accompanied by infrequent Vulcanian and phreatic explosions and

Figure 6.4 Variations in the number of earthquakes before the 11 November 1964 eruption of Sheveluch volcano.
Data from the Kamchatka Regional Earthquake Catalog.

Figure 6.5 Variations in the number of earthquakes calculated for every 48 hrs before the 1993 eruption of Sheveluch volcano according to the records of seismic station SVL situated at a distance of 8 km from the crater. A double arrow shows the beginning of eruption.
Data from Gorelchik et al., 1997.

culminated in the 21 April 1993 paroxysmal explosion of the dome that produced about 0.5 km^3 of pyroclastic material (Gorelchik et al., 1995, 1997).

Figure 6.5 shows the temporal variations in the number of earthquakes preceding the eruption. A seismic swarm began three weeks before the eruption. The majority

of earthquakes occurred within the distance of less than 5 km from the crater dome; their depths did not exceed 3 km below sea level (b.s.l.). The number of earthquakes gradually increased up to the moment of the eruption. After the eruption, the number of earthquakes sharply decreased.

6.2 Volcano-Tectonic Earthquakes Associated with Phreatic and Phreato-Magmatic Explosions

6.2.1 El Chichón Volcano, Mexico

El Chichón volcano is an andesitic strato-volcano in southern México. It erupted on March 1982, after about 600 years of quiescence (Simkin and Siebert, 1994), and produced a sequence of six large phreato-magmatic explosions, occurring from 29 March to 4 April, that destroyed the summit dome and formed a 1-km-wide crater.

The eruption was preceded by a 28-day seismic swarm (Figure 6.6), which began on 1 March 1982. Seismic activity was recorded by the 14-station seismic network of UNAM that was installed in this region during 1979–1980 (Havskov et al., 1983; Yokoyama et al., 1992; Zobin and Jiménez, 2008). The swarm began with a sharp increase in the number of earthquakes during the first 5 days of activity. The maximum magnitude event ($M = 4.0$) was recorded on the fifth day of the swarm and was followed by a gradual decrease in the seismic activity during next two weeks. The activity increased again on 19 March but practically disappeared about 2 days before the eruption. The majority of preceding earthquakes (80%) were located within the area of 10×20 km^2 with the volcano in the center of this area (Figure 6.7). They occurred at the depths ranging from 2 to 8 km; a small group of foci was located also at the depths between 10 and 18 km (Figure 6.8).

During the explosive activity from 20 March to 5 April, the earthquakes clustered densely within the volcano crater zone and to the north of the crater (Figure 6.7). After the last explosion, seismic activity continued for one more month. The depth of events went down up to 20–25 km (Figure 6.8). The earthquake foci filled the depth interval from 0 to 25 km (Figure 6.7) and stayed at this depth in April (Jimenez et al., 1999).

6.2.2 Volcán de Colima, México

The andesitic strato-volcano Volcán de Colima (height of 3860 m) is one of the most active volcanoes in Mexico. The eruptions of Volcán de Colima during the last 480 years involved a wide spectrum of eruption styles, including small phreatic explosions, major block-lava emissions, and large explosive events (Breton González et al., 2002).

Figure 6.6 Daily distribution of maximum magnitudes (A) and the number of earthquakes (B) associated with the 1982 eruption of El Chichón volcano. St 1, St 2, and St 3 indicate the three stages in the development of activity. Arrows show the appearance of the first and largest explosions.
From Zobin and Jiménez, 2008.

The phreatic explosion of 1994 occurred at 20:13 on 21 July. It destroyed the 1991 crater dome and produced avalanches and ash fall. The explosion formed a shallow crater approximately 140 m in diameter and 50 m in depth. The explosion was heard at the distance of 20−35 km to the south of the volcano (Jiménez et al., 1995). The eruption was preceded by a seismic sequence from 13 to 21 July that was located within the volcanic edifice.

About 450 small seismic events were recorded by the nearest seismic station (the distance from the crater of 1.7 km) during the 1994 swarm. Figure 6.9 shows the temporal variations in the number of events. The seismic activity developed in two stages, from July 12 to 15, and from July 17 to 22. The maximum magnitude Mw 1.2 event was recorded during the peak activity of the second stage.

Figure 6.7 Epicenters (above) and hypocenters (below) of the earthquakes associated with the 1982 eruption of El Chichón volcano. Open circles indicate the earthquakes that occurred before the first explosion of 29 March; crosses indicate the earthquakes recorded from 29 March to 4 April; filled circles indicate the earthquakes that occurred after the explosion of 4 April up to the end of April. The position of El Chichón volcano is shown by the star. From Zobin and Jiménez, 2008.

6.2.3 Popocatépetl Volcano, México

Popocatépetl volcano is situated within the Trans-Mexican Volcanic Belt. It is a 5452-m high andesitic strato-volcano. Its most recent unrest began with a short swarm of volcano-tectonic earthquakes on 21 December 1994 after nearly 70 years of quiescence. A series of moderate phreatic explosions at the crater occurred in about 45 min after the swarm beginning (De la Cruz-Reyna et al., 1995). The explosions produced emissions of ash which fell on several towns to the east and northeast from the volcano, including the large city of Puebla. This unrest then developed into a sequence of steam-and-ash and fumarolic emissions up to March 1996 when the repetitive dome-building and dome-destroying episodes began. Eruptive activity continues up to the present time (De la Cruz-Reyna et al., 2008).

Figure 6.8 Temporal variations of earthquake focal depths during March and April 1982. The interval between the first and last explosions is shown by arrows.
From Jiménez et al., 1999.

Figure 6.9 Variations in the number of events calculated for every 12 hrs before the 1994 explosion at Volcán de Colima.

The 21 December 1994 swarm began at 07:03 (GMT; 01:03 L.T.) on the background of very low seismic activity of the volcano (Figure 6.10A). Based on the catalog of volcano-tectonic earthquakes prepared by (Lermo-Samaniego et al., 2006), it continued for about 8 h (Figure 6.10C) but the majority of earthquakes occurred during the first 45 min. The earthquakes were rather small; only 14 volcano-tectonic earthquakes were located by a network consisting of 4 short-period seismic stations. The maximum magnitude event ($M = 3.0$) occurred in about 45 min after the swarm beginning.

Figure 6.10 Daily number of volcano-tectonic earthquakes recorded at Popocatépetl volcano during December 1994 (A) and hourly maximum magnitude (B) and the number of volcano-tectonic earthquakes (C) recorded at Popocatépetl volcano during 21 December 1994.
The catalog of earthquakes was prepared by J. Lermo-Samaniego (personal communication, 2003).

Figure 6.11 shows the distribution of the seismic events occurring during December 1994. All the earthquakes were located within the volcano edifice. The rare events, recorded before the 21 December crisis, were distributed within the depth range between 2.5 km above sea level (a.s.l.) and 2 km b.s.l. The events of the swarm clustered along the northwest-southeast direction across the summit area. They occurred at the depths interval between 2 and 5 km a.s.l., reaching the summit elevation. During the last 10 days of December, seismic events occurred at the depths between 0 and 4 km a.s.l.; their epicentral zone spread to the southwest from the swarm epicenters.

Figure 6.11 Distribution of epicenters (A) and hypocenters (B) located at Popocatépetl volcano during December 1994. The contours of 3500 and 2900 m show the relief of Popocatépetl volcano, in meters.
The catalog of earthquakes was prepared by J. Lermo-Samaniego (personal communication, 2003).

6.2.4 Soufriére Hills Volcano, Montserrat

Andesitic strato-volcano Soufriére Hills (SHV) with the height of 915 m is situated in the southern part of Montserrat Island, West Indies. Its recent eruption, after about 370 years of repose (Simkin and Siebert, 1994), began on 18 July 1995 with the steam and ash venting from the northwest crater. The eruption activity passed through the phreatic and magmatic explosions, growth, and collapse of lava domes accompanied by pyroclastic flows and a great directed explosion. It continued for more than 15 years.

About 10,000 volcano-tectonic and hybrid earthquakes with magnitudes from −0.5 to 2 were located from 28 July 1995 to the end of February 1997 with a high precision (Aspinall et al., 1998; Miller et al., 1998). They followed all the main events in the volcanic activity. Earthquakes occurred within three epicentral zones, beneath Soufriére Hills volcano, along the structure extending to the NE of the volcano, and beneath St. George Hill (SGH), that is situated about 5 km to the northwest from the volcano. All the events were recorded at shallow depths, mainly from 5 km b.s.l. to the surface. Figure 6.12 shows the temporal variations in the depths of seismic events during 1995−1997 for the epicentral zone of the volcano together

Figure 6.12 Depth variations of earthquakes located beneath Soufriére Hills volcano from 28 July 1995 to 28 February 1997. Depth 0 km marks the sea level. Arrows show the main episodes in volcanic activity (Ph. Exp., phreatic explosion of 21 August 1995; Dome Growth and Beginn. Dome Collap., the initial growth and the beginning of dome collapses; Mag. Exp., magmatic explosion of 17 September 1996).
Data from the catalog of Montserrat Volcano Observatory (MVO).

with the main events in the volcanic activity. The tendency in shallowing of earthquake foci during the first year of activity from the phreatic explosion up to the beginning of the dome collapses is clearly seen.

We discuss here the variations in seismic activity preceding two important moments in the volcanic activity of SHV: the phreatic explosion of 21 August 1995 and the magmatic explosion of 17 September 1996.

Phreatic explosion of 21 August 1995. It was the first large phreatic explosion during the 1995–1997 eruptive sequence. It blanketed the nearest city Plymouth, situated at the distance of about 6 km from the volcano, in a thick ash cloud and caused darkness for about 15 min (Chronology. . ., 2002).

Figures 6.13 and 6.14 show the temporal variations in earthquake number and the distribution of epicenters and hypocenters of earthquakes from 1 to 25 August. An increase in seismic activity began with a 2-day swarm of earthquakes along the northeast extend of the SHV from 5 to 6 August, then a new 2-day swarm was recorded beneath SGH that continued from 12 to 13 August. The seismicity beneath the volcano was low up to 20 August. During 2 days, 20 and 21 August, about 140 earthquakes were recorded. The explosion occurred at 08:03 a.m. (L.T.) of 21 August (B. Voight, personal communication, 2002) during the maximum level of seismic activity below the volcano. Just after the explosion, the number of seismic events sharply decreased.

The seismic events recorded beneath the volcano were clustered within a vertical column shape volume extending from 3 km b.s.l. to the summit of the volcano. The foci tended to be shallower before the phreatic explosion. Figure 6.15 shows that the focal depths of earthquakes changed from 1 to 3 km b.s.l. on 1–19 August to the depth ranging from 2 km b.s.l. to the summit of the volcano on 20–21 August.

Figure 6.13 Variations in daily earthquake number before the 21 August 1995 phreatic explosion at Soufriére Hills volcano. I, II, and III are the earthquake sequences that occurred within the northeast extend from the volcano (I), beneath SGH (II), and beneath the volcano (III). The largest earthquake is shown by a single arrow; the phreatic explosion is shown by double arrow. Data from the catalog of MVO.

Figure 6.14 Distribution of epicenters (A) and hypocenters (B) of earthquakes located before the 21 August 1995 phreatic explosion at Soufriére Hills volcano. Square is SGH, star is SHV. The zero-line of sea-level depth is shown.
Data from the catalog of MVO.

The events associated with the extension of the northeast structure of the volcano, as well as the SGH earthquakes, were more deep and located at the depths from 5 to 3 km b.s.l.

Magmatic explosion of 17 September 1996. A series of dome collapses that began in July 1996 (see Figure 6.12) led to the 17 September (23:46 LT) magmatic explosion. Rocks and pumice fell in the south of the island, and houses were destroyed in the nearest village. A major ash plume rose to about 12 km, and about 600,000 tons of ash were deposited in southern Montserrat (Chronology. . ., 2002).

Figure 6.15 Temporal variations in earthquake depths beneath the volcano before the 21 August 1995 phreatic explosion at Soufriére Hills volcano. The phreatic explosion is shown by double arrow. The beginning of the 20–21 August swarm is shown by a single arrow. Data from the catalog of MVO.

The seismic activity that preceded the 17 September explosion is shown in Figures 6.16–6.18. The earthquake foci located during about 40 days of activity are clustered mainly beneath the SHV forming a vertical column from the depth of 3 km b.s.l. to the summit of the volcano. The zones that were active before the 21 August 1995 phreatic explosion now were calm.

The swarm of earthquakes had no clear beginning; it began on the background of continuous seismic activity due to the process of partial collapsing of the dome that began in July 1996. Nevertheless, we can mark the gradual increase in earthquake numbers beginning from 29 August; therefore, the swarm continued for about 23–24 days. All the earthquakes were of low magnitude; the largest event with magnitude 1.9 occurred just before the explosion of 17 September. The explosion occurred during the apparent decrease in number of seismic events; after the explosion, the number of earthquakes sharply decreased.

Figure 6.18 shows that the seismic process developed permanently at the depths between 1 and 2 km. No focal depth migration to the surface was noted before the explosion.

Figure 6.16 Variations in daily earthquake numbers before the 17 September 1996 magmatic explosion at Soufriére Hills volcano. The largest earthquake is shown by a single arrow; the magmatic explosion is shown by double arrow. Data from the catalog of MVO.

6.3 Volcano-Tectonic Earthquakes Associated with Lava Extrusions

6.3.1 Volcán de Colima, México

An eruption occurred on 20 November 1998, when andesitic block lava began to extrude in the summit crater. The dome quickly spilled over the southwest crater rim and fed block-lava flows that descended in three branches on the southern slope of the volcano, to the maximum distance from the crater of 3.8 km; the advance of the block-lava flows was accompanied by pyroclastic flows (Zobin et al., 2002a).

Figure 6.19 shows temporal development of seismic activity before the 1998 eruption. About 15,000 small earthquakes were recorded. The eruption was preceded by five earthquake swarms from 28 November 1997 to 20 November 1998. Each swarm consisted of a sequence of volcano-tectonic earthquakes.

The distribution of epicenters and hypocenters of earthquakes with magnitude Mw from −0.5 to 2.7 recorded from November 1997 to November 1998 is shown in Figure 6.20. We plotted here only the earthquakes located within the area of the seismic network, as those outside the network had larger location uncertainties. The epicenters were located within the Colima volcanic complex (CVC) and covered an area of about 50 km^2. The majority of events clustered in the area of Colima volcano's active crater and between the volcanic edifices of Colima and Nevado volcanoes.

The majority of hypocenters were located at depths ranging from 10 km (mainly from 5 km) b.s.l. to the crater zone of the volcanoes, or up to heights of

Figure 6.17 Distribution of epicenters (A) and hypocenters (B) of earthquakes located before the 17 September 1996 magmatic explosion at Soufriére Hills volcano. The square marks St. George's Hill (SGH), the star marks Soufriére Hills volcano (SHV). Data from the catalog of MVO.

about 4000 m. A dense clustering of hypocenters was observed within a narrow depth interval of 1–4 km a.s.l. within the volcanic constructions. Figure 6.20B shows a space free of earthquake foci just below the volcano at the depths of 0–10 km b.s.l. This free space could correspond to a zone of possible magma storage.

Figure 6.21 shows the depth variations of earthquakes in time. It is difficult to note any variations in depth for the events within the individual swarms. However, there were substantial variations in foci depth from swarm to swarm. It is possible to distinguish two stages in depth variations. The first stage includes the first three swarms from November 1997 to May 1998; the depths gradually decreased from swarm to swarm. During June–July (swarm 4), the depths again increased, with their distribution similar to the depth distribution of the first November–December 1997 swarm. The last swarm (swarm 5), that

Figure 6.18 Temporal variations in earthquake depths beneath the volcano before the 17 September 1996 magmatic explosion at Soufriére Hills volcano. The magmatic explosion is shown by double arrow.
Data from the catalog of MVO.

Figure 6.19 Variations in number of events recorded during 1997–1998 before the 20 November 1998 lava extrusion at Volcán de Colima. Counted events are those recorded by a minimum four stations over 5-day intervals. Numbers 1–5 identify individual swarms. From Zobin et al., 2002b.

Figure 6.20 The distribution of epicenters (above) and hypocenters of earthquakes (below) located during the seismic crisis of 1997–1998 at Volcán de Colima. The contours of 3000, 3500, and 4000 m show the relief of Colima Volcanic Complex. VC is Volcán de Colima; NC is Nevado de Colima. The seismic stations are shown as triangles. An earthquake-free zone is outlined in the cross section.
From Zobin et al., 2002b.

occurred just before the eruption, was shallower, with events occurring only from 1 to 4 km a.s.l.

The eruption of Volcán de Colima continues up to the date of preparing this book. New lava extrusions were formed in 2001–2002, 2004, and the recent lava dome is growing in the crater beginning from 2007. No more swarms of volcano-tectonic earthquakes prior to the lava extrusions were recorded.

Figure 6.21 Temporal variations of event depths during the five swarms (numbers 1–5) at Volcán de Colima. Events in each swarm are marked by particular symbols.
From Zobin et al., 2002b.

6.4 Volcano-Tectonic Earthquakes Associated with Flank Eruptions

6.4.1 Sakurajima Volcano, Japan

Sakurajima volcano is an andesitic strato-volcano (altitude of 1110 m) that is situated in the southern part of Aira caldera, Southern Kyushu. The volcano had repeated flank and summit eruptions during historic times (Ishihara, 1990). The 1914 eruption of the volcano was one of the largest volcanic events in the twentieth century in Japan and it was one of the first volcanic eruptions that were studied using instrumental seismic observations (Omori, 1914; Yokoyama, 1997).

The eruption began on 12 January 1914 at around 08:00 h (local time) with the emissions of white smoke from the summit crater, and 2 h later, explosions occurred at the western and eastern slopes, where the vents were opened. Just after midnight of 13 January, the strongest explosions were observed and at about 20 h lava began to effuse from the craterlets on both the eastern and western slopes. The eruption continued until May 1915 (Yokoyama, 1997).

Seismic activity near the volcano was recorded in Kagoshima by local seismic station beginning from 11 January (Omori, 1914). The recording drum of the low-sensitivity seismographs (magnification 5 to 10) was triggered into motion by

Figure 6.22 Temporal variations of hourly number of volcano-tectonic earthquakes of Sakurajima volcano recorded by the Gray-Milne-Ewing seismograph in Kagoshima at a distance of 10 km from eruptive centers. A single arrow shows the largest earthquake; double arrows show the eruptive events. From Omori, 1914.

intense shocks over a certain threshold (Figure 6.22). Their magnitudes were 4.8−5.2, according to Yokoyama (1997). The higher sensitivity seismic station in Nagasaki (epicentral distance was about 148 km, but the magnifications were of 20 and 120) began to record the swarm at the lower threshold on 9 January. Earthquakes increased in number and magnitude, with the maximum activity at midnight of January 12. After the opening of flank vents in the morning of January, earthquake activity rapidly decreased. Then, at 18:30 on 12 January, with a background of seismic calmness, the $Mw = 7.0$ (Zobin, 2001) shock occurred. About 26 h, later the great lava flank extrusion began. About 1.6 km^3 of lavas overflowed from the volcano, which connected the volcanic island with Oosumi Peninsula (Yokoyama, 1997).

7 Volcano-Tectonic Earthquakes at Dacitic Volcanoes: Case Studies

Eruptions of *dacitic volcanoes* may occur from the summit crater and the new vents on the flank of a volcano. The high-viscosity and relatively low-temperature magmas with great amount of volatiles are characterized by low mobility and high explosivity. The magmas may construct a lava plug dome, a cryptodome, or a lava spine in the crater of the volcano. Large volume of volatiles can produce numerous explosions of different intensity. Volcano-tectonic earthquakes can be observed during the movement of magma to the surface before lava extrusion. Lava effusion from summit and flank vents may be accompanied by volcanic tremor. Dacitic volcanoes have low altitude and stratovolcano shape.

7.1 Volcano-Tectonic Earthquakes Associated with Summit Eruptions

Summit eruption of dacitic volcanoes may consist of the growth of a dome in the crater with formation of lava flows and (or) a great explosion or a sequence of numerous small explosions. The most famous recent summit eruptions at dacitic volcanoes occurred at Mount St. Helens in the United States, Usu in Japan, and Pinatubo in the Philippines. Volcano-tectonic earthquakes are related to lava dome eruptions and magmatic explosions.

7.1.1 Mount St. Helens, Cascades

Mount St. Helens (MSH) is a stratovolcano of 2950 m of height. It is situated within the Cascade Range in the western state of Washington, USA. The eruptive history of MSH began about 40,000 years ago with dacitic volcanism. The products of eruptive activity during the past 2500 years were represented mainly by dacite–andesite–dacites. The volcano had been apparently dormant since 1857 (Mullineaux and Crandell, 1981).

Eruption of 1980. The first 1980 steam-blast eruption from its summit crater was observed on 27 March, and was followed by a sequence of small explosive eruptions up to the climactic eruption on the morning of 18 May. This explosive eruption of dacitic magma formed a vertical Plinian column more than 25 km high. The northward-directed lateral blast of the bulging north flank of the volcano erupted about 2 km^3 of cold and juvenile material and devastated an area of

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00007-4
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approximately 600 km^2. A new crater, with an area of 1.5×3.0 km^2 and of depth of about 500 m, was formed in the northern flank of the volcano. The height of the volcano diminished by about 400 m. Several smaller magmatic eruptions occurred then during 1980 producing pyroclastic flows, ash falls, and several lava domes (Christiansen and Peterson, 1981). Lava domes began to grow in the crater beginning from June 1980 (Moore et al., 1981). The eruption terminated in 1986.

The seismic activity of MSH was monitored by the regional Pacific Northwest Seismic Network (PNSN). Figure 7.1 shows the temporal variations in daily number of events and their maximum magnitudes during three stages of activity: preeruptive, before the great 18 May explosion, and after the great 18 May explosion up to the end of August 1980. The 1980 swarm of volcano-tectonic shallow earthquakes began on 16 March with small events; the first significant $ML = 4.2$ event was recorded beneath Mount St. Helens on 20 March. The number and magnitude of earthquakes sharply increased starting 25 March with peak number on 27 March, when the first explosion occurred. After the beginning of the eruption, the number of earthquakes gradually decreased during 1 month; then their number was more or less stable up to 18 May, when a cataclysmic explosion occurred. The largest earthquake with $Mw = 5.7$ (Zobin, 2001) occurred on 18 May on the background of low-level seismic activity, practically simultaneously with the great directed blast of the volcano.

Figure 7.2 shows the location of earthquakes during March–May 1980. Before the 27 March explosions, 93 earthquakes were located beneath MSH and to its SE. The

Figure 7.1 Variations in daily maximum magnitude (A) and the number of earthquakes (B) located during the seismo-volcanic activity of MSH during March–August 1980. Double arrows show eruptive events.
Data from the PNSN catalog.

Figure 7.2 Epicenters (A) and hypocenters (B) of earthquakes located during the seismo-volcanic activity of MSH during March–May 1980. The earthquakes located before 27 March (first explosion) are shown by circles; the earthquakes located from 27 March to 18 May are shown by crosses. The position of volcano MSH is shown by the 1500-m contour.
Data from the PNSN catalog.

seismic events formed two clusters: shallow, at the depth range of 0–5 km, and deep cluster, at the depth range from 18 to 25 km. During the active period from 27 March to 18 May, 627 located earthquakes occurred beneath the volcanic edifice, mainly at the depth range from 0 to 10 km; the deep cluster disappeared. The largest earthquake of 18 May was located at the depth of 2.8 km beneath the northwest flank of MSH.

After the great 18 May explosion, seismic activity beneath the crater continued, and was associated with a few magmatic eruptions of the volcano between May and August 1980 (Weaver et al., 1981; Malone et al., 1981). At the same time, the regions to the north and to the south of the volcano became seismically active (Figure 7.3). The 50-km-length submeridional linear structure with the volcano in its middle section was activated. Seismic foci to the south of the volcano clustered at the depth range from 20 to 1 km; the northern foci formed a cluster with the length of about 15 km and within the same depth range from the surface to 20 km.

Figure 7.3 Post-18 May 1980 seismic activity in the region of MSH observed from 19 May 1980 to 31 August 1981. (A) Epicenters of earthquakes; (B) hypocenters of earthquakes (north−south cross section). The earthquakes, recorded before the 18 May explosion, are shown by diamonds; the earthquakes recorded after the 18 May explosion are shown by circles. The 1500-m contour of MSH is shown.
Data from the PNSN catalog.

The seismic foci just beneath the volcano occupied space between deep (16−22 km) and shallow (from depths of 10 km to the surface) groups of earthquakes recorded before the 18 May explosion.

Eruption of 2004. During the period between 1980 and 1986, 20 small eruptions were recorded. Then the active period changed with the quiescence between 1987 and 2004. Figure 7.4 shows the change in the seismic activity at late 1987, when the first of several years-long swarms of "deep" (>3 km) volcano-tectonic events began. The depth of these swarms contrasted sharply with the exclusively shallow (<3 km) precursory swarms associated with 19 of the 20 post-18 May 1980, dome-building eruptions.

In September 2004, a new large eruption began. Rapid onset of unrest at Mount St. Helens on 23 September 2004, initiated an uninterrupted lava-dome-building eruption. About 3 weeks of intense seismic unrest and localized surface uplift, punctuated by four brief explosions, constituted a vent-clearing phase. The largest of

Figure 7.4 Time versus depth plot showing all earthquakes (circles) with well-constrained locations occurring at Mount St. Helens between 1980 and 2005. Depths are relative to a datum corresponding to average altitude of stations in Mount St. Helens network, roughly 1.1 km above sea level.
From Moran et al., 2008.

these explosions, rating VEI of 2, occurred on 5 October, generating an ash plume that reached about 2 km above the crater rim. The third week exhibited lessened seismicity and only minor venting of steam and ash, but rapid growth of the uplift, or welt, south of the 1980–1986 lava dome proceeded as magma continued to push upward. Crystal-rich dacite (~65 wt.% SiO_2) lava first appeared at the surface on October 11, 2004, beginning growth of a complex lava dome of uniform chemical composition accompanied by persistent but low levels of seismicity, rare explosions, low gas emissions, and frequent rockfalls. Growth and disintegration of spines continued through the first 8 months of 2005. Two notable explosions occurred on 16 January and on 8 March 2005 (Moran et al., 2008; Scott et al., 2008).

Figures 7.5 and 7.6 show the development of the seismic activity during September and October 2004. The reawakening of MSH after 17 years and 11 months of slumber was heralded by a swarm of shallow (depth < 2 km) volcano-tectonic earthquakes on 23 September 2004. After initial decline on September 25, seismicity rapidly intensified, reaching a peak level on 29 September, when $M > 2$ earthquakes were occurring at the rate of about one per minute. The first explosion occurred on 1 October 2004, 8.5 days after the first earthquakes, and was followed by three other explosions over the next 4 days. Seismicity declined after each explosion. Following the last and the strongest explosion of this series on 5 October, seismicity declined significantly. The largest earthquake of $Mc = 3.8$ occurred on 2 October. With the beginning of lava dome growth on 11 October, the number of located volcano-tectonic earthquakes gradually decayed during 5 days, and ceased in the last days of October.

Figure 7.5 Variations in daily maximum magnitude (A) and the number of earthquakes (B) located during the seismo-volcanic activity of MSH during September–October 2004. Double arrows show explosive events; the single arrow shows the maximum magnitude earthquake. The beginning of the lava dome growth in the crater is shown by the horizontal arrow.
Data from the PNSN catalog.

Figure 7.6 shows the location of earthquakes during 20 September–31 December 2004. Before 1 October, 422 earthquakes with magnitudes *Mc* from 1 to 3.3 were located by PNSN within the crater area of MSH (Figure 7.5A and B). The hypocenters formed a vertical structure filling the depth interval from 0 to 2.1 km. Six largest events with magnitude 3.3 were located at the depths between 0.5 and 1.2 km. After the beginning of the eruption on 1 October, the position of earthquake active volume practically had no change (Figure 7.5C and D). The largest $Mc = 3.8$ event was located at the depth of 1 km beneath the crater zone.

The comparison of the seismic activity, preceding the 1980 and 2004 eruptions, shows that the 2004 seismic events were of smaller magnitudes. They were shallower, and clustered within a small vertical volume beneath the crater area while

Figure 7.6 Epicenters (A) and hypocenters (B) of earthquakes located before the first explosion of the 2004 eruption at MSH that occurred on 1 October 2004, and the epicenters (C) and hypocenters (D) of earthquakes located after the first explosion of the 2004 eruption during the seismo-volcanic activity of MSH in the period of September–October 2004. The position of volcano MSH is shown by the 1500-m contour.
Data from the PNSN catalog.

the 1980 events (see Figure 7.2) developed along the submeridional regional fault including the volcanic edifice and their hypocenters were distributed till the depths of 20–25 km.

7.1.2 Usu Volcano, Hokkaido

Eruption of 1977–1978. The eruption occurred in the southeastern part of the summit crater (Figure 7.7). Four new craters were formed during a sequence of magmatic explosions on 7–14 August. The volume of ejected pumice material was about 0.01 km^3 (Yokoyama et al., 1981).

The eruption was preceded by 32 h of seismic activity (Figure 7.8). The earthquake swarm consisted of about 3000 small (magnitude 1–2) volcano-tectonic earthquakes that were located beneath the summit crater at a depth from 0 to 2 km (Figure 7.7). The number of earthquakes gradually increased over time, and the

Figure 7.7 Epicenters of earthquakes located just before the 7 August 1977 eruption of Usu volcano. MS is the 1910 cryptodome Meiji-shinzan; SS is the 1943–1945 lava dome Showa-shinzan. Solid squares show the position of seismometers; the cross indicates the position of crater 1. From Yokoyama et al., 1981.

Figure 7.8 Temporal variations in hourly number of earthquakes recorded at the station JMA-A located on the somma of Usu 32 h before the 1977 eruption. The double arrow shows the beginning of the eruption; the single arrow shows the maximum magnitude earthquake.
From Yokoyama et al., 1981.

eruption began when the frequency of events reached its maximum. The maximum magnitude $M = 3.7$ earthquake was observed half an hour before the beginning of eruption (Yokoyama and Seino, 2000; Yokoyama, personal communication, 2001). The new eruptive craters were situated within the epicentral zone of earthquakes. The alarm was issued by JMA, and there were no victims (Yokoyama et al., 1981).

During the 8–14 August stage of pumice explosions, strong explosions were preceded and followed by a sharp decrease in seismic activity (Figure 7.9). The stage of calm after the explosions continued for 3–10 h, and then seismic activity gradually increased again (Suzuki and Kasahara, 1979).

7.1.3 Unzen Volcano, Kyushu

Unzen volcano is a composite volcano of 1359 m of height and consists of multiple volcanic cones of dacite to andesite. Following 198 years of dormancy, a small phreatic eruption started at its summit on 17 November 1990. During February–April 1991, a sequence of phreatomagmatic eruptions followed a resumed phreatic eruption on 12 February; they developed into a dacitic dome eruption on 20 May 1991. The growth of the dome continued for about 4 years accompanied by repeating partial collapses and pyroclastic flows. The total volume of erupted materials was 2.1×10^8 m^3 (Nakada et al., 1999).

Figure 7.9 Variations in hourly earthquake frequency at the station SSU situated at a distance of 2.5 km east from the crater ($M > 0.8$) from 6 to 19 August 1977. It is clearly seen that there is an absence of seismic events just after the large explosions Big-I, II, III, IV, and SB.
From Suzuki and Kasahara, 1979.

Seismic activity began 1 year before the phreatic eruption (Nakada et al., 1999; Umakoshi et al., 2001). Figure 7.10 shows the temporal variations in the number of earthquakes. Five swarms of volcano-tectonic earthquakes were recorded up to the moment of lava extrusion. The first two swarms were not followed by volcanic activity. Swarm 1 occurred at a distance of about 20 km to the west of the volcano at depths from 10 to 20 km below the ancient Chijiwa caldera (Figure 7.11A). It began on 21 November 1989 and continued for 4 days. About 600 earthquakes were located with the maximum $M = 3.3$ event. Swarm 2 that occurred during July–August 1990 was shallower; it was located within the depth range from 15 to 5 km. The largest earthquake with magnitude 3.9 was observed on 7 July. This swarm occupied the territory between the first swarm epicenters and the volcano (Figure 7.11A).

The next three impulses of seismic activity occurred in November 1990 and during February–March and May 1991. They were associated with eruptive events. Their hypocenters were clustered, practically repeating the locations from swarm to

Figure 7.10 Temporal variations of located volcano-tectonic earthquakes (number of events per 5 days) before and during the 1990–1991 eruption sequence at Unzen volcano, Japan. Arrows show the main eruptive events. Numbers correspond to the swarm of earthquakes. Data from the catalog of Shimabara Earthquake and Volcano Observatory.

Figure 7.11 Epicenters and cross sections for the swarms of volcano-tectonic earthquakes that occurred during 1989−1991 at Unzen volcano, Japan. The swarms that were not followed by volcanic activity are shown in (A) (circles, swarm 1; triangles, swarm 2). The swarms associated with volcanic eruptions are shown in (B) (swarm 3), (C) (swarm 4), and (D) (swarm 5). Chijima caldera (CC) is shown by dashed line.
Data from the catalog of Shimabara Earthquake and Volcano Observatory.

swarm, within the long inclined zone that began at the depth of 15 km at a distance of about 20 km from the volcano and terminated at the surface about 2 km from the summit of the volcano (Figure 7.11C and D).

Swarm 3 (Figure 7.11B) preceded the phreatic eruption of 17 November 1990. It began on 13 November with a sharp increase in the number of events and continued for about 25 h (Figure 7.12A). The phreatic eruption occurred after the cease of seismic activity. The resumed eruption, that began on 12 February 1991 and was represented by a sequence of phreatic and phreatomagmatic explosions, was not preceded by volcano-tectonic seismicity. Swarm 4 began after the phreatic explosion and accompanied a sequence of phreatomagmatic explosions.

Swarm 5 occurred just before the lava extrusion of 20 May 1991. It began on 18 May and reached its peak in the number of events on 19 May 1991. The lava extrusion was observed just after ceasing of the swarm (Figure 7.12B). This swarm was the smallest of five seismic sequences; it was practically at the background level of seismicity (Figure 7.10). It was characterized by a small number of shallow seismic events; the majority of earthquake foci were located at depths from 6 to 15 km (Figure 7.11D).

The characteristic feature of the Unzen seismicity was the absence of volcano-tectonic events just beneath the volcano crater. Umakoshi et al. (2001) suggested that it was a zone of high-temperature, ductile bodies, or a shallow magma chamber.

After the termination of lava activity in February 1995, volcano-tectonic earthquakes were observed along the submeridional structure crossing the focal zone of swarm 1 at a distance of about 20 km from the volcano (Figure 7.13). They were located at the depth range from 5 to 17 km.

7.1.4 Pinatubo Volcano, Luzon

Mount Pinatubo is a composite volcano of 1745 m elevation and is situated in the Luzon volcanic arc. Its recent unrest began after about a 500-year calm on 2 April 1991 after about 20 days of locally felt earthquakes related to steam explosions. The unrest culminated 10 weeks later in the great explosion of 15 June that produced about 10 km^3 of porous, pumiceous deposits and formed a caldera of 2.5 km in diameter. Within the twentieth century, it was the second largest eruption after the 1912 Novarupta–Katmai (Alaska) eruption (Wolfe and Hoblitt, 1996).

Seismic observations began only in May; therefore, the preceding earthquake swarm was not recorded. Nevertheless, the seismic activity that preceded the 7 June shallow lava extrusion at the summit and the following 15 June great explosion was located by a seismic network. It allowed forecasting the 15 June climactic eruption (Harlow et al., 1996).

Figures 7.14–7.16 show the development of the seismic activity from 7 May to 12 June 1991. Seismic events were recorded within two clusters: just beneath the summit (I) and 5 km northwest of the summit (II) (Figure 7.14). The second group of seismic foci occurred mainly within the depth range from 7 to 3 km while the second group of events was more shallow, within the range 4 km b.s.l. to 1 km a.s.l. The events

Figure 7.12 Variations in the daily number of earthquakes before the 17 November 1990 phreatic explosion (A) and the 20 May 1991 lava extrusion (B) at Unzen volcano, Japan. Data from the catalog of Shimabara Earthquake and Volcano Observatory.

Figure 7.13 The epicenters (A) and hypocenters (B) of the posteruption volcano-tectonic earthquakes (circles) at Unzen volcano, recorded from February 1995 to the end of 1999. Crosses show the earthquakes recorded during the eruptive activity.
Data from the catalog of Shimabara Earthquake and Volcano Observatory.

from the distant zone were dominant from 21 May to 5 June. Seismicity of zone I beneath the volcano began to increase from 30 May, and an intensive swarm of low magnitude (from 0 to 1) was recorded there from June 6 to 7 (Figure 7.15). This swarm culminated in the 7 June lava dome extrusion. Just after the beginning of dome growth, the number of earthquakes sharply decreased, and the 12 June explosion occurred on the background of low-level seismicity for both seismic zones.

Figure 7.14 Epicenters (A) and hypocenters (B) of earthquakes located from 7 May to 12 June 1991 near Pinatubo volcano. The 3-D velocity model was used for location (Harlow et al., 1996). PV is Pinatubo volcano. I and II mark two clusters of earthquakes. The 1000-m contour of the volcano's relief and the zero-line of sea-level depth are shown. Data from Mori et al., 1996b.

Figure 7.16 shows that the seismic foci, which occurred beneath the volcano from 8 May to 5 June, became shallower before the 6−7 June swarm (and the lava dome extrusion), changing the depths from 3−5 km to 0−2 km. The intense seismic activity, which occurred from June 12 to 15, was practically off the scale. Several of the seismic stations were destroyed, the remaining were very often

Figure 7.15 Temporal variations in the daily number of earthquakes located from 7 May to 12 June 1991 beneath the Pinatubo volcano (I) and 5 km to northwest from the volcano (II). Double arrows show the eruptive events; the single arrow shows the maximum magnitude earthquake. (I) and (II) are two clusters of earthquakes.
Data from Mori et al., 1996b.

saturated (Harlow et al., 1996). The 9-h climactic sequence of explosions of 15 June was accompanied by large volcano-tectonic earthquakes. Two of them, recorded at 18:41 and 19:15, reached magnitude $Mw = 5.7$ and 5.6, respectively (Harvard CMT).

The location of seismic events was again possible only from 29 June. The post-explosion distribution of seismic foci sharply changed from the previous one that was shown in Figure 7.14. The seismically active zones extended in space and depth (Figure 7.17). The zone beneath the volcano spread to the west within volcanic edifice and occupied all beneath-volcano space filling it up to the depths of 25 km, about 20 km deeper than before the 15 June explosion. The northwest seismic zone became also deeper and larger in area.

7.2 Volcano-Tectonic Earthquakes Associated with Flank Eruptions

We have only three cases; all are related to the activity of Usu volcano, Hokkaido.

Figure 7.16 Variations in the focal depths of the events from cluster II situated 5 km to the northwest from the volcano (A) and from cluster I beneath the volcano Pinatubo (B) from 7 May to 12 June 1991. Zero-line of sea-level depth is shown. The beginning of the 6−7 June swarm is shown by the arrow.
Data from Mori et al., 1996b.

7.2.1 Usu Volcano, Hokkaido

Eruption of 1910. The eruption of the cryptodome Meiji-shinzan began at the northern foot of the volcano on 25 July (see Figure 7.7), after 4 days of precursory earthquakes (Figure 7.18). Nearly 40 craterlets were formed along the arcade line approximately 3 km long during 10 days of phreatic explosive activity. The eruption caused an upheaval of approximately 90 m above the original terrace (Yokoyama and Seino, 2000).

The felt shakes began on 21 July. On 22 July, 25 shocks were felt at a distance of 8 km from the volcano. Figure 7.18B shows the 6-h frequency of earthquakes recorded by horizontal pendulum seismograph (magnification 30) that was installed at the meteorological observatory of Sapporo, at a distance of 70 km from the volcano. The number of earthquakes increased rapidly and reached its maximum 13 h before the beginning of the eruption. The strongest earthquake ($Mw = 5.4$; Zobin, 2001) was recorded 30 h before the beginning of the eruption. The eruption began

Figure 7.17 Post-15 June 1991 seismic activity in the region of Pinatubo observed from 26 June to 18 August 1991. (A) Epicenters of earthquakes; (B) hypocenters of earthquakes (north−south cross section). The earthquakes, recorded before the 15 June explosions, are shown by diamonds; the earthquakes, recorded after the 15 June explosions, are shown by circles. The 1000-m contour of Pinatubo (PV) is shown.
Data from Mori et al., 1996b.

Figure 7.18 Variations in the number of earthquakes preceding the 1910 flank eruption of Usu volcano. (A) Obtained from macroseismic data at a distance of 8.5 km from the new volcanic site; (B) recorded by a seismic station at a distance of 70 km from the new volcanic site. Data from Omori, 1911–1913.

on the background of a gradual decrease in the number of earthquakes (Omori, 1911–1913). Figure 7.18 shows also the number of earthquakes that were felt at Nishi-Monbets village, at a distance of 8.4 km from the volcano (A). The remarkable similarity between the curves is seen.

Eruption of 1944. It was a birth of a new volcano Showa-Shinzan on the flank of Usu (see Figure 7.7). The seismic activity and deformations began about 6 months before the first explosion of 23 June 1944. Figure 7.19 shows the variations in the number of felt earthquakes recorded at a distance of about 3 km from the new volcano site by M. Mimatsu (Mimatsu, 1995). The author considers the list of Mimatsu-san complete enough for the analysis of temporal variations in seismicity.

Figure 7.19 Variations in the daily number of Usu earthquakes, associated with the 1943–1944 flank eruption, that were felt at a distance of 3 km from the volcanic site. The double arrow shows the beginning of the eruption; the single arrow shows the maximum magnitude earthquake.
Data from Mimatsu, 1995.

Seismic activity began with numerous earthquakes; the maximum number of events was recorded during the first 3 days of the activity on 29–31 December 1943. Then the number of earthquakes began to gradually decrease during January–February 1944 and stayed at the background level up to the first explosion occurring on 23 June. The maximum magnitude event ($M = 5$; Yokoyama and Seino, 2000) occurred in 12 days after the beginning of the activity on 9 January and was located by a three-station network at the depth of 11 km (Kizawa, 1958).

Eruption of 2000. The 2000 eruption began on the northwestern flank of the volcano (Figure 7.20) on 31 March at 13:07 (local time), following about 4 days of intensive seismic activity and a high-rate deformation of the volcano surroundings (Ichiyanagi, 2002; Takahashi et al., 2002). More than 60 explosion craters were formed in March and April by the sequence of phreato-magmatic and phreatic explosions within two new eruptive centers: Nishi-yama (N in Figure 7.20) and Kompira-yama (K in Figure 7.20). The initial phreato-magmatic eruption on 31 March was the largest and produced a tephra fall with a small pyroclastic flow. The volume of dacitic ash and pumice was about 0.1 km^3. The eruption was

Figure 7.20 The epicenters (A) and hypocenters (B) of earthquakes (JMA catalog) that occurred at Usu volcano from 28 March to 1 April, 2000. The events of 28−29 March are shown by dark symbols; the events of 30−31 March are shown by light symbols. The 200-m, 400-m, and 600-m contours of the Usu volcano are shown by lines. The centers of flank eruptions of Usu volcano are shown by the ovals with letters. *N* and *K* are the 2000 centers; *MS* and *SS* are the centers of eruptions during 1910 and 1944, respectively; the centers *W* and *E* were not dated. Broken lines show the lateral extent of volcanic activity during the 1910 and 1943−1944 flank eruptions. The positions of flank eruptive centers and the lines of lateral extent of volcanic activity are shown according to Yokoyama and Seino (2000). From Zobin et al., 2005a.

accompanied by upheaval process that reached of about 75−80 m at the Nishi-yama cryptodome (Jousset et al., 2003). After the summer of 2000, the intensity of upheaval decayed gradually.

About 1200 earthquakes were located by Japan Meteorological Agency (JMA) from 28 March to 1 April; 339 of them with magnitudes ranging from 2.5 to 4.6 (Ueno et al., 2002a). The regional seismic network of JMA and the Institute of Seismology and Volcanology of Hokkaido University (ISV) comprised 16 short-period stations, situated at a distance of 1−100 km from the volcano (Ueno et al., 2002b). The network allowed locating the earthquakes with mean errors for latitude ± 0.6 km and for longitude ± 0.7 km. The mean depth errors were equal to ± 2.2 km. The earthquake magnitudes M_{JMA} (or M) were estimated for 948 events.

Seismic activity, preceding the 2000 Usu eruption, began on 27 March at 20 h (local time) with unlocated low-magnitude earthquakes ($M < 1$). The first located earthquake of magnitude $M = 1.0$ occurred on 28 March at 00:10 at the depth of 5.7 km. During the first 40 h of activity, the frequency of events was rather low, not exceeding 10 located events per hour (Figure 7.21). The maximum magnitudes of earthquakes were 3.2−3.4. The number of earthquakes and their magnitudes increased sharply at the end of 29 March and reached their maximum between 9 and 12 h of 30 March. Then seismic activity gradually decreased and practically ceased on 2 April. Three largest earthquakes of the swarm occurred during the

Figure 7.21 Temporal developments of the 2000 earthquake swarm at Usu (JMA catalog). Single arrows show the position of largest earthquakes; double arrows show the beginning of the eruption in the Nishi (N) and Kompira (K) centers of the eruption.
From Zobin et al., 2005a.

maximum seismic energy release ($M = 4.4$, 30 March, 09:12 and 17:13) and at the final stage of the swarm, 14 h after the beginning of the eruption ($M = 4.6$; $Mw = 4.9$, 1 April, 03:12) (Zobin et al., 2005a).

The earthquake epicenters occupied an area of about 30 km^2, covering the summit zone and its southern and southwestern flanks of the volcano (Figure 7.20). Initially, the earthquakes occurred within the zone situated beneath the summit and southern slope of the volcano. Then the seismic foci clustered along two linear zones intersecting to the south of the initial epicentral zone. The foci were recorded within the depth range between 0 and 13 km, but clustered mainly from 4 to 7 km. The earthquake epicenters stopped at a distance of about 1−3 km from the sites of forthcoming eruption.

The alert of forthcoming eruption was issued by JMA in the morning of 29 March; by that night, more than 15,000 people from the towns around the volcano were evacuated (JMA, 2003).

The 31 March phreato-magmatic explosion from the north vents occurred on the northwestern flank of the volcano about 4 km from the summit of Usu, at the end of seismic swarm (Figure 7.21) and just beyond the zone of the earthquake epicenters (Figure 7.20).

8 General Properties of Volcano-Tectonic Earthquake Swarms

This chapter discusses the general properties of volcano-tectonic earthquake swarms. At first, we generalize the earthquake swarm properties inferred from the data presented in Chapters 5–7. Then these data are supplemented and compared with the published data in this field.

8.1 Properties of Volcano-Tectonic Earthquake Swarms Inferred from the Data of Chapters 5–7

We passed through a sequence of seismo-volcanic episodes discussed in Chapters 5–7. Table 8.1 collects the characteristics of seismic sequences of volcano-tectonic earthquakes that were observed in the twentieth century at 20 volcanoes of the world during their 40 eruptions. These data are not representative for the Earth's volcanic activity during the past 100 years; it is only a small sample showing the most studied or (and) most remarkable examples of seismic activity at volcanoes. Nevertheless, we can discuss the general properties (or tendencies) of seismic swarms, associated with volcanic activity, using these observations. We select for discussion the following characteristics: temporal variations in seismic activity, spatial distribution of seismic foci, posteruption seismicity; duration of seismic activity before volcanic event, position of volcanic event in seismic sequence, and dependence of these characteristics on the type of magma feeding the eruption. The last section of this Chapter is devoted to some regularity in the seismic swarm activity associated with reawakening andesitic and dacitic volcanoes.

8.1.1 Temporal Variations

The temporal variations in seismic activity were of three types: one-peak (I); multi-peak sequences with the maximum peak at the beginning of the sequence (II); or with maximum peak at the end of sequence (III) (Figure 8.1A). The first type of sequences is rare (Figure 8.1B); it was observed only for 9 from 35 cases. The second and the third types are distributed in the equal manner.

For basaltic and dacitic volcanoes, the most popular is a seismic sequence with the maximum peak in the beginning of the sequence (Type 2), while for andesitic

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00008-6
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Table 8.1 Data on Volcano-Tectonic Earthquake Sequences

Volcano, Region	Type of Magma	Year	Type of Erupt.	VEI	Duration of Prel. Earthq. Seq., h	Depth, km	Range of Magnitudes	Type of Eart. seq.	Position of Erup. Event
Kilauea Hawaii	Basaltic	1969	4	0	5	2–10	0–3.5	1	4
		1974	1,4	0	9	0–4	0–3.4	3	4
		1982	1	0	2.5	1–3	0–3.2	1	3
Etna, Sicily	Basaltic	1983	2	1	28	(–)3–15		3	4
		1984	1	2	0			–	–
		1991	2	2	14	0–10	0–4.4	2	1
		2001	2	2	120	(–)2–5	0–3.9	1	3
Oshima, Izu Isl.	Basaltic	1986	1	3	0				
		1986	2	3	2	0–5	0–5.9	–	3
Klyuchevs-koy, Kamchatka	Basaltic	1951	2	2	96		0–3.1	2	4
		1966	2	3	84		0–3.7	2	4
New Tolba-chik, Kamchatka	Basaltic	1975	4	4	216	0–16	0–4.9	2	4
Dabbahu, Ethiopia	Basaltic	2007	4	3	340–600	2–12	0–5.4	2	4
Fernandina, Galapagos Isl.	Basaltic	1968	1	4			0–5.2	–	–
Teishi Knoll, Izu Isl.	Basaltic	1989	5		216	0–12	0–5.5	2	4
Miyakejima, Izu Isl.	Basaltic	2000	2, 5		70	1–4		1	2
		2000	1a, 5	3	1200	0–20	1–6.4	1	
Bezymianny, Kamchatka	Andesitic	1955	1	5	408			3	2
		1956	1	5	4320		0–5.2	2	3
		1959	1	1	48		0–1.3	1	2
		1960	1	2	70		0–1.7	1	2

Sheveluch, Kamchatka	Andesitic	1964	1	4	216		0−5.2	3	2
		1993	1	3	500			3	2
El Chichon, México	Andesitic	1982	1	5	620	0−20	0−4.0	2	3
Volcán de Colima, México	Andesitic	1998	1	3	600	(−)4−0	0−2.7	3	3
Soufriére Hills, Montserrat	Andesitic	1995	1	3	30	(−)0.5−4	0−1.4	3	3
		1996	1		600	(−)0.5−3	0−1.9	3	3
Popocatép-etl, México	Andesitic	1994	1	2	8	(−5)−2	1−3.0	2	2
Sakurajima, Kyushu	Andesitic	1914	2	4	80		0−7.0	3	4
Mount St. Helens, Cascades	Dacitic	1980	1	5	80	0−20	0−4.4	2	2
		1980	1	5	1300	0−10	0−5.7	2	3
		2004	1	2	570	0−2	1−3.8	2	3
Usu, Hokkaido	Dacitic	1910	2	2	108		0−5.4	3	3
		1944	2	2	700		0−5.0	2	3
		1944	2	2	4300			2	4
		1977	1	3	32	0−2	0−3.7	3	2
		2000	2	2	90	0−15	1−4.6	1	3
Unzen, Kyushu	Dacitic	1990	1	1	25	1−15	0−3.0	1	4
		1991	1	1	48	0−15	0−3.0	1	3
Pinatubo, Philippines	Dacitic	1991	1	6	160	(−)1−1.6	0−1.6	3	2

Note: Data about the type of eruptions and their VEI are taken from Simkin and Siebert (1994) and BGVN: 1, central eruption; 1a, caldera collapse; 2, flank eruption; 3, radial fissure; 4, regional fissure; 5, submarine eruption. For the codes of earthquake sequence (1−3) and of the position of volcanic events (1−4), see in text. Symbol (−) before the depth means the depth above sea level.

Figure 8.1 Distribution of three types of volcano-tectonic earthquake swarms for all events of Table 8.1 (B) and for basaltic, andesitic, and dacitic volcanoes (C). The three types of volcano-tectonic earthquake swarms are the one-peak (I) and multipeak sequences with the maximum peak in the beginning of sequence (II) or with maximum peak at the end of sequence (III). The typical curves of swarms are shown in the upper part of (A).

volcanoes, the maximum number of seismic events for the majority of seismic sequences is observed at the final stage of the swarm (Figure 8.1C, Type 3).

8.1.2 Spatial Distributions

Generally, the seismic events cluster within or below the volcanic edifice within a depth range from 20 km to the summit of the volcano. The seismic events of preliminary swarms occupy a greater area than the site of forthcoming eruption. At the initial stage of seismic activity, the events occurring at the depths 10–20 km were observed together with the shallower events. Before an eruption, the earthquakes

cluster at shallower depths. It was observed for New Tolbachik volcanoes (Figure 5.17); Mount St. Helens (Figure 7.2); Pinatubo (Figure 7.14); and Volcán de Colima (Figure 6.20), among others. Very often the earthquakes originate not only beneath the volcano, but along a system of local faults adjusting the volcano also. The examples of well-located earthquakes show that during the initial stage of seismic activity before an eruption, the seismic events were observed along the local faults, such as St. George's Hill for Soufriére Hills volcano (Figure 6.14); the fault 5 km northwest of Pinatubo volcano (Figure 7.14); and the fault southeast of Mount St. Helens (Figure 7.2). The adjacent fault events had a greater depth than the events originating beneath the volcano. Just prior to an eruption, seismic activity began to concentrate at shallower depths beneath the volcanic edifice.

The important feature of spatial distribution of volcano-tectonic earthquakes is the migration of seismic foci (or seismic clusters) for the distance of 10−20 km. The focal migration was observed before the forming of the eighth cone of New Tolbachik volcanoes (Figure 5.21). Earthquakes migrated during 6 days along the fault at the distance of about 15 km from the site of the formation of the first seven cones to the site of the eighth cone.

Multipeak temporal variations in the number of seismic events may be considered as a result of the migration of seismic clusters. Before the submarine eruption of Teishi Knoll, the two peaks in temporal variations of the number of earthquakes corresponded to the clustering of seismic events within two separated zones (Figures 5.28 and 5.30). One of these zones was active from 4 to 7 July 1989, and produced the first peak in frequency of events; the second cluster developed in a longer and narrower zone, and produced the second peak of frequency. The same was observed at New Tolbachik volcanoes (Figures 5.19 and 5.20). During the period from 27 to 30 June 1975, the earthquake source zone was larger and deeper, and produced the first peak in earthquake frequency; during the period from 1 to 6 July 1975, the earthquakes occurred within a long narrow zone, and produced the second peak of earthquake frequency.

8.1.3 Posteruption Seismic Activity

Very often, we have no volcano-tectonic seismicity with the beginning of the eruption. Nevertheless, the posteruption seismicity was observed after the eruptions of New Tolbachik volcanoes, Mount St. Helens, Oshima, Unzen, and Pinatubo. The spatial distribution of the posteruption volcano-tectonic earthquakes overfills the areas that were active before and during the eruptive events. For New Tolbachik volcanoes (Figure 5.23), the epicenters of posteruption earthquakes, that were observed during 2 years after the eruption, outline the 100-km-length fault that includes the 12-km new-eight-cones fissure. They occurred at the same depths as the events of preliminary earthquake swarm.

For Mount St. Helens, the activated posteruption fault, whose central part was occupied by the volcano, was outlined to the north and to the south of the volcano. The free space between two seismic clusters that occurred separately at the depths

of 0−10 km and from 16 to 22 km before and during the eruption was filled by the postvolcanic seismic events during 1 year of activity (Figure 7.3).

For Unzen volcano, the posteruption events traced a 30-km-long structure that intersects at a depth of 20 km the seismic volume where the first preliminary swarm occurred 1 year before the eruption (Figure 7.13).

In the case of Pinatubo volcano, the posteruption seismic events significantly expanded the seismic volume of preliminary events. The epicentral area and the depths of earthquakes were both increased by a factor of 2 (Figure 7.17).

The volcano-tectonic earthquakes, which followed the eruption of Oshima volcano, also significantly expanded the seismic volume that was active before the eruption (Figures 5.13 and 5.14).

These spatial characteristics of the posteruption seismicity suggest a relationship between the collapses of magma chambers after the magma release during the eruption.

8.1.4 Duration of Seismic Swarms Prior to Eruption

The duration of seismic swarms of the majority (70%) of seismic sequences of Table 8.1 ranges between 10 and 1000 h (a geometrical mean of 74 ± 11 h) (Figure 8.2A). These values mainly reflect the properties of andesitic and dacitic volcanoes (Figure 8.2B); for the majority (70%) of basaltic volcanoes, the durations of seismic activity prior to eruption are shorter (from 1 to 100 h; a geometrical mean of 25 ± 6 h).

8.1.5 Position of a Volcanic Event According to the Stage of Volcano-Tectonic Earthquake Swarm

Volcanic events may occur during the different stages of seismic sequence (Figure 8.3A): at the initial stage of seismic sequence (I), at the maximum frequency of seismic events (II), at the ceasing of seismic events (III), and after ceasing of seismic sequence (IV). The majority of the discussed eruptions occurred while the earthquakes were ceasing (stage III) (Figure 8.3B). For basaltic volcanoes, the majority of eruptions occur after the end of the seismic sequence, while for andesitic volcanoes, the majority of eruptions occur during the maximum number of volcano-tectonic earthquakes (Figure 8.3C).

8.2 Additional Data About Volcano-Tectonic Earthquake Swarm Properties

8.2.1 Size of Volcano-Tectonic Earthquake Swarm Area

The epicentral area of earthquake swarm is defined as the area of volcano-tectonic earthquake epicenters (magnitude 1−2 and greater) during eruptive activity. Areas of epicentral zones vary for different volcanoes from 25 km for Mount St. Helens

Figure 8.2 Distribution of duration of volcano-tectonic earthquake swarms for all events of Table 8.1 (A) and for basaltic (1), andesitic (2), and dacitic (3) volcanoes (B).

to 9450 km for northern Tonga (Zobin, 1996). Zobin (1996) showed that the areas are larger for volcanoes on small islands and smaller for volcanoes on large islands or on continents. Assuming that the crust is thinner for oceanic volcanoes than for continental volcanoes, it was found that a significant negative correlation exists between the epicentral areas of volcanic earthquakes, S, and the crustal thickness, H, under the volcano (Figure 8.4):

$$\log S(\mathrm{km}^2) = 3.6 \times 10^5 H^{-3.1}(\mathrm{km}) \tag{8.1}$$

8.2.2 Earthquake Swarm Duration

Benoit and McNutt (1996) have compiled a Global Volcanic Earthquake Swarm Database, 1979–1989. The database gives the data about 191 swarms preceding eruption activity, as reported in the *Bulletin of Volcanic Eruptions* of the Volcanological Society of Japan (Figure 8.5).

Figure 8.3 Position of volcanic event according to the different stages in volcano-tectonic earthquake swarms for all events of Table 8.1 (B) and for basaltic, andesitic, and dacitic volcanoes (C). The stages are illustrated in (A): the initial stage of seismic sequence (I), stage at the maximum frequency of seismic events (II), at the ceasing of seismic events (III), and after ceasing of seismic sequence (IV).

Figure 8.5 shows that the main part of earthquake swarms preceding eruptive activity had durations from 1 to 100 days (or from 24 to 2400 h). A geometric mean of preliminary swarm duration is 8 days (132 h). This value is twice longer than the value obtained from the data of Table 8.1.

Benoit and McNutt (1996) made an interesting observation about the relationship between the swarm duration and eruption repose time. Figure 8.6 shows that the long swarms (longer than 4 months) have a good correlation of their duration with the time of repose; the longer the period between eruptions, the greater duration of seismic activity observed before the eruption. The authors guess that longer

Figure 8.4 Relationship between the epicentral area of volcano-tectonic earthquake swarms and the crustal thickness for eight selected eruptions.
From Zobin, 1996.

Figure 8.5 Histogram of the duration of 191 volcanic earthquake swarms preceding eruptive activity. The data form a nearly log-normal distribution with a geometric mean of 8 days.
From Benoit and McNutt, 1996.

lasting swarms may be the result of ascending magma encountering a cooler and stronger crustal environment.

8.2.3 Magnitude—Frequency Relations of Events in Volcano-Tectonic Earthquake Swarms

The frequency of the occurrence of earthquakes in a given region and for a given period can be represented by the equation $\log N = a - b\,M$, where N is the number of earthquakes with magnitudes of a fixed range around magnitude M, and a and b are constants. The constant b is called the b-value and depends on the relation between

Figure 8.6 Relationship between the duration of 65 precursory swarms and eruption repose time. The regression line shown is for 12 swarms with a duration of 100 days or greater. The correlation coefficient is 0.94, which is significant at the 99% confidence level.
From Benoit and McNutt, 1996.

large and small events in the earthquake sequence. For tectonic earthquakes, *b*-values are between 2/3 and 1 and do not show much regional variability.

For volcanic earthquake swarms, *b*-values were estimated as close to 1 or greater, up to 2.6 (Figure 8.7), indicating that volcano-tectonic earthquake swarms tend to have an anomalously large number of small events compared to the usual seismic sequence (Gresta and Patané, (1983); McNutt, 1986; Zobin et al., 2002b; and others). Results of McNutt (1986) for Pavlof volcano, Alaska, show that *b*-values of volcano-tectonic microearthquake ($M = 0-1$) swarms do not depend on the stage of volcanic activity.

At the same time, when larger earthquakes participate in the swarm, the magnitude−frequency distribution can vary during the volcanic activity or be non-log-linear (Gresta and Patané, 1983; Okada, 1982). This reflects the unstable relationship between small and large events during the volcano-tectonic earthquake swarm. Gorshkov (1959) has noted that during a course of an earthquake swarm related to the 1956 eruption of Bezymianny volcano, Kamchatka (see Chapter 6), there were two independent sequences of large and small events observed. The large events (magnitude about 4) occurred as a background without any relation to the course of eruption, while the smaller events reflected the variations in volcanic activity. Okada (1982) demonstrated the same characteristics for the earthquake swarm preceded the 1980 cataclysmic eruption of Mount St. Helens, Cascades volcano (see Chapter 7). Figure 8.8 shows that the large and small earthquakes develop in time distinctively. There exists an intrinsic lack of earthquakes over a particular magnitude range.

The effect of the intrinsic lack of earthquakes within the volcano-tectonic earthquake swarms may be a reason for the non-log-linear magnitude−frequency relations. Okada (1982) pointed out this effect for three volcano-tectonic earthquake swarms associated with the great eruptions of Fernandina, Galapagos (1968), Usu, Hokkaido (1977−1978), and Mount St. Helens, Cascades (1980) (Figure 8.9). It is

Figure 8.7 A log-linear magnitude–frequency plot for 1316 B-type volcano-tectonic earthquakes recorded 1–8 December 1974 during the warning stages of the period of explosive eruptions at Pavlof volcano, Alaska. This plot may be approximated with the relationship $\log N = 2.53 - 2.6\,M$. From McNutt, 1986.

Figure 8.8 Temporal variations in magnitudes of volcano-tectonic earthquakes preceding the May 18, 1980, directed blast at Mount St. Helens volcano, Cascades. From Okada, 1983.

seen that for all three cases there is a minimum in magnitude–frequency relations. Okada (1982) guessed that larger earthquakes are related to the large amount of surface deformation in a form of doming, caldera collapse, or bulging. Therefore, the volcano-tectonic earthquake swarm may consist of two or more separate swarms reflecting the different-scale processes during great manifestations of volcanic activity.

Figure 8.9 Non-log-linear magnitude–frequency plots for volcano-tectonic earthquakes associated with the 1979–1980 lava dome activity at Usu, Hokkaido (A), the 1968 caldera collapse at Fernandina, Galapagos (B), and the 1980 directed blast at Mount St. Helens, Cascades (C). N is the number of earthquakes; M_{sap} is the magnitude determined by Sapporo station records. The plots are shown for different stages of volcanic activity indicated by dates. From Okada, 1983.

8.3 Some Regularities in the Volcano-Tectonic Earthquake Swarms Proclaiming Reawakening of Andesitic and Dacitic Volcanoes

The general cycle of activity of andesitic and dacitic volcanoes (Chapters 6 and 7) may be described as follows:

1. Quiescence
2. Phreatic explosive activity with the formation of a new crater
3. Repeated dome building and destruction accompanied by explosions
4. Quiescence.

The period of quiescence may continue for a few years or a few centuries. Usually, the largest explosive eruptions occur when the period of quiescence was at least 100 years. Among them are the explosions at volcanoes Bezymianny (1956, Kamchatka), Mount St. Helens (1980, Cascades), El Chichón (1982, México), and Pinatubo (1991, the Philippines). All these large eruptions were preceded by swarms of volcano-tectonic earthquakes described in Chapters 6 and 7. The volcano-tectonic earthquakes followed the main stages of the volcanic eruptions and may serve as a good tool for reconstruction of the stress release during eruptive events.

Six significant eruptions at volcanoes with a long quiescence that occurred during the past 50 years were selected (Table 8.2): Bezymianny (Kamchatka, 1956), Mount St. Helens (Cascades, 1980), El Chichón (México, 1982), Unzen (Japan, 1991), Pinatubo (the Philippines, 1991), and Soufrière Hills (Montserrat, West Indies, 1995).

Table 8.2 Characteristics of Volcano-Awakening Process at Andesitic and Dacitic Volcanoes

Volcano	VEI	Time of Quiescence, Years	Duration of the First Stage (Seismic Swarm Before the First Explosion), Days	Duration of the Second Stage (Before the Great (or Large) Explosion), Days
Bezymianny, Kamchatka, 1955–1956	5	900–1000	24	160
Mount St. Helens, Cascades, 1980	5	120	11	52
El Chichón, México, 1982	5	600	28	7
Unzen, Kyushu, 1991	1	200	360	87
Pinatubo, Philippines, 1991	6	500	20	70
Soufrière Hills volcano, Montserrat, 1995	3	370	180–250	33

Note: The values of VEI and the time of quiescence are given according to the Global Volcanism Program catalog (http://www.volcano.si.edu/, 31 May 2006). The durations of stages are taken from Zobin (2003).

The main criteria for the selection of this group of volcanoes were a long period of quiescence (more than 100 years) and the presence of seismic data associated with the different stages of their activity. The selected eruptions were characterized by different values of VEI, ranging from VEI 1 for the eruptions of Unzen in 1991 to VEI 6 for Pinatubo in 1992.

Figure 8.10 illustrates the development of seismic activity during the reawakening of Bezymianny volcano during 1955–1956 (see Section 6.1.1). It is possible to discriminate three stages in the reawakening process: (1) the preliminary stage that was characterized by a sharp increase in seismic activity near the volcano up to the first phreatic explosion of 22 October 1955; (2) the stage between the first phreatic explosion and the large explosion of 31 March 1956 was characterized by a low in number but high in magnitude seismic events; (3) the postexplosion stage of dome-building with a gradual decrease in the number and magnitude of seismic events.

These three stages are common for reawakening volcanoes (see Chapters 6 and 7) and may be a basis for the analysis of any peculiarities in the awakening process. Table 8.2 summarizes the characteristics of the three stages of seismo-volcanic activity for the six volcanoes studied. Relationships between the durations of stages 1 and 2 and the VEI of the forthcoming large explosion are illustrated in Figure 8.11.

Figure 8.10 Daily distribution of maximum magnitudes (A) and the number of earthquakes (B) associated with the 1955–1956 eruption of Bezymianny volcano. St 1, St 2, and St 3 indicate the three stages in the development of activity. Arrows show the appearance of the first and largest explosions.
Data from Tokarev, 1981.

8.3.1 Relationship Between the Duration of Stage 1 and the VEI of Forthcoming Explosion

Two groups of events are discriminated visually in Figure 8.11A: the group of low-VEI and long-duration stage 1 events (Unzen, 1991 and Soufrière Hills volcano, 1995) and the group of high-VEI and short-duration stage 1 (Bezymianny, 1956;

Figure 8.11 Relationships between the duration of stage 1 (A) and stage 2 (B) with VEI. The volcano indices are UNZ, Unzen; SHV, Soufrière Hills volcano; CHI, El Chichon; BEZ, Bezymianny; MSH, Mount St. Helens; PIN, Pinatubo.
From Zobin and Jiménez, 2008.

Mount St. Helens, 1980; El Chichón, 1982; and Pinatubo, 1991). This discrimination shows that long-lasting seismic activity (about or more than a year) before the reawakening of a volcano signifies a low probability that the volcano will produce a large explosion. On the contrary, rapid development (less than a month) of seismic activity may proclaim a forthcoming large magnitude explosion.

8.3.2 Relationship Between the Duration of Stage 2 and Postexplosion Dome Building

Generally, it is impossible to see in Figure 8.11B any relationship between the duration of stage 2 from the first phreatic explosion at a reawakening volcano to its largest explosion and VEI value. One significant exception should be noted: the only volcano whose sequence of explosions did not culminate in a dome-building period was El Chichón. Its 1982 eruption was very short, its stage 2 was of 7 days only. It is the lowest value for the duration of stage 2 in Figure 8.11B. All other eruptions, whose stage 2 continued for more than 20 days, produced significant lava domes.

8.3.3 The Conceptual Model of Reawakening Process

Zobin and Jiménez (2008) proposed the conceptual model for the three stages in reawakening of dormant volcanoes based on models of lava dome formation (McLeod and Tait, 1999; Melnik and Sparks, 2002).

Stage 1 is the first announcement of the forthcoming eruption at a dormant volcano. Therefore, it may be considered as a reflection of the process of deep dyke nucleation. Considering dyke growth following chamber pressurization, two alternatives may be proposed (McLeod and Tait, 1999): sudden chamber pressurization or gradual chamber pressurization. Dyke growth that is controlled by sudden chamber pressurization may be more rapid than that is controlled by gradual chamber pressurization.

Seismic activity is indicative of a passage way for magma to reach the surface and would be a function of the type of dyke growth. In this case, we may consider slow development of seismic activity as a result of gradual chamber pressurization. On the contrary, rapid development of seismic activity may be considered as a result of sudden chamber pressurization. In both cases, decompressive degassing of the intruding magma during final seismic activity leads to shallow pressurization that produces the first explosion (Gardner and White, 2002).

The stage 2 may be attributed, in the frame of the model by Melnik and Sparks (1999), to the nonlinear effects of crystallization and gas loss by permeable flow within magma. They note that degassing results in rheological stiffening due to gas exsolution and microlite crystallization from an undercooled melt. Viscosity can increase by several orders of magnitude when silicic melts, containing a few percent dissolved water, decompress and degas. This stage is characterized by a process of gas exsolution from the melt that can produce a sequence of ash explosions, forming a new crater. Only a small portion of the fresh magma material is expelled. A wide range of stage 2 durations was observed for different eruptions, from 7 to 160 days (Table 8.2). The duration of the eruption development from the first small phreatic explosion to a large phreatic or phreato-magmatic explosion may be proportional to the rate of gas exsolution and dissolved water liberation from magma melt. The intensity of a large explosion is supposed to depend on the type of magma chamber pressurization (or on the duration of stage 1): sudden magma chamber pressurization leads to a very large explosion, gradual magma chamber pressurization leads to only a large explosion.

Stage 3 is characterized by an important volcanological aspect: whether there follows the construction of a new lava dome or not. This aspect is supposed to be constrained by the duration of stage 2. The only case where no dome was built was the eruption at El Chichón volcano, and this eruption was characterized by the shortest duration of stage 2. Another feature of this eruption was the high intensity of all the explosions. Considering that the viscosity of silicic magma containing only a few percent dissolved water can increase by several orders of magnitude during decompression and degassing (Melnik and Sparks, 1999), it may be supposed that a high rate of decompressing and degassing during a short time had increased the viscosity within the dyke up to a level when it was impossible to push it to the surface. As a result, a dyke was formed as a degassed intrusive structure beneath the volcano without building a superficial lava dome.

It is important to consider the three separate stages during the development of a strategy of alarms when a volcano begins to awake. The first stage is the stage for the monitoring of the awakening process. The second stage is important for understanding that the first explosion of the awakening volcano is not the strongest. The strongest one will occur after some time.

9 Source Properties of Volcano-Tectonic Earthquakes

In this chapter, the author discusses the focal mechanisms and source properties of volcano-tectonic earthquakes. As was shown in Chapters 5–8, volcano-tectonic earthquakes occur within and around the volcanic edifice and reflect the interaction of two general geological processes: the magma migration to the Earth's surface and the crustal tectonic activity. Therefore, it is important to study the focal mechanisms of volcano-tectonic earthquakes and stress release during volcano-tectonic earthquake swarms, and the relationship between the stress situation within the volcanic edifice and the regional stress field.

9.1 Focal Mechanisms of Volcano-Tectonic Earthquakes: Double-Couple and Non-Double-Couple Models

The complete tensor of volcano-tectonic earthquake source may be described as a sum of its isotropic and deviatoric parts (see Section 4.3). The deviatoric part may be decomposed into two source types: a double-couple (DC) and a compensated linear vector dipole (CLVD). The application of these two source types to describe the fault nature of volcano-tectonic earthquakes is discussed.

9.1.1 Double-Couple Model

During 1962–1964, the global seismic networks were installed. It allowed collecting enough *P*-wave first motions to construct the focal mechanism solution on the stereographic projection for volcanic earthquakes of magnitude about 5 or greater. The first *P*-wave motion focal mechanism solutions for volcano-tectonic earthquakes (Figure 9.1) were constructed for two magnitude 5 earthquakes preceding the 11 November 1964 directed-blast eruption at Sheveluch volcano, Kamchatka (Zobin, 1970). They indicated a strike-slip faulting for these earthquakes. Later Zobin (1972, 1979a, b) constructed the focal mechanism solutions for the magnitude 5–6 volcano-tectonic earthquakes associated with the eruptions at Miyakejima, Tori-shima, Beerenberg, and New Tolbachik volcanoes. All of them were of strike-slip faults. It was shown that one of two nodal planes of the focal mechanism solution of large volcano-tectonic earthquake coincides with the strike of the fissure, where new volcanic centers form (New Tolbachik volcanoes, Figure 9.2). Zobin (1972) showed the similarity between focal faulting of these

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00009-8
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Figure 9.1 Focal mechanism solution for the volcano-tectonic earthquake that coincided with the 11 November 1964 directed blast at Sheveluch volcano, Kamchatka. The focal mechanism diagram is an upper hemisphere plot. I and II are possible fault planes; *P* is compression; *T* is tension; solid circles, compressions; open circles, dilatations. Numbers indicate the seismic stations. From Zobin, 1970.

Figure 9.2 Comparison of the focal mechanism solution for the *Mw* 4.9 2 July (04:07 hours) volcano-tectonic earthquake preceding the 6 July 1975 New Tolbachik fissure eruption, Kamchatka with the direction of the eruptive fissure. It is seen that the fault plane I of the solution coincides with the direction of the 12-km-long fissure where the chain of new cones formed. The focal mechanism diagram is a lower hemisphere plot. I and II are the possible fault planes; *P* is compression; *T* is tension; symbol (+) shows compressions; symbol (−) shows dilatations.
From Gorelchik et al., 1990.

volcano-tectonic earthquakes and that of tectonic earthquakes surrounding the volcanic regions.

These first results were supported by studies of focal mechanisms that were developed for volcano-tectonic earthquakes associated with different eruptions in the world. Figure 9.3 shows the focal mechanism solutions that were constructed for some volcano-tectonic earthquakes originating beneath the dacitic Mount St. Helens volcano, Cascades (Weaver et al., 1981) after the directed-blast eruption of 18 May 1980 during the stage of magmatic eruptions (see Section 7.1.1). These earthquake focal mechanisms indicated the thrust (A−C, E) and strike-slip faulting at the volcano-tectonic earthquake's foci. All focal mechanisms have steeply dipping fault planes that strike northeast.

The study of the focal mechanisms of earthquakes that occurred during the same year (1980) within the tectonic structures to the north and south of the volcano (Figure 9.4) shows the similarity in focal mechanism of the earthquakes, originated beneath the volcano in the active eruptive stage, and the earthquakes, originating in tectonic ranges far from the volcano. They were the same thrusts (indexes F, I in Figure 9.4) and strike-slips (G, J, K, L), as well as normal faults (H).

Figure 9.3 Focal mechanism solutions of volcano-tectonic earthquakes associated with the 1980 magmatic activity of Mount St. Helens volcano, Cascades and the epicenters of these earthquakes. Epicenters of earthquakes are shown as small symbols (magnitudes less than 2.8) and large symbols (magnitudes > 2.8). Depths are indicated as symbol + (0–5 km), × (5–10 km), □ (10–15 km), and ◇ (greater than 15 km). Posteruption 1500-m contour is shown. The location of lava dome is shown as *D*. The focal mechanism diagrams are upper hemisphere plots, with darkened quadrants indicating compression and white quadrants indicating dilatations; solid circles, compressions; open circles, dilatations. Mechanism *B* is a composite solution; all other solutions are single-event solutions.
From Weaver et al., 1981.

Focal mechanism solutions for volcano-tectonic earthquakes preceding the 1989 Teishi Knoll submarine basaltic eruption (see Section 5.5.1) were obtained by two methods: from *P*-wave polarity distribution (Matsumura et al., 1991) and from the inversion method (Takeo, 1992). Figure 9.5 shows the *P*-wave first motion focal mechanism solutions for three earthquakes (Matsumura et al., 1991). The largest earthquake (magnitude JMA 5.5) revealed a strike-slip nature with a normal fault component; two magnitude 3.1 earthquakes indicated thrust faulting. The solution obtained by the inversion method (Takeo, 1992) for the same magnitude 5.5 earthquake is consistent with the *P*-wave first motion solution. Figure 9.6 shows the source time functions of the strike-slip and dip-slip components on each subfault of the earthquake fault model marked by arrows. It is seen that a right-lateral strike-slip (SS) motion dominated during the rupturing.

The focal mechanism solutions that were determined by NEIC for seven magnitude *mb* 4.9–5.7 earthquakes, which occurred during the paroxysmal stage of the eruption during 12–15 June 1991 of the dacitic volcano Pinatubo, the Philippines, revealed the predominantly strike-slip movement with a minor vertical (normal or reverse) component; only one earthquake was from normal faulting with a large strike-slip component. These solutions suggest that the largest volcano-tectonic

Figure 9.4 Seismicity and focal mechanism diagrams for some events that occurred within the tectonic environment of Mount St. Helens between 18 May and 31 August 1980. Legend is the same as for Figure 9.3.
From Weaver et al., 1981.

earthquakes at Pinatubo were consistent with the east-west maximum regional stress (Bautista et al., 1996).

Okada et al. (1981) determined the composed focal mechanism solutions for small volcano-tectonic earthquakes, which occurred at Usu volcano, Hokkaido during the different stages of dacitic dome activity during October 1977–June 1979. Dip-slip

Figure 9.5 Focal mechanism solutions (lower hemisphere projection) for three volcano-tectonic earthquakes preceding the 13 July 1989 submarine Teishi Knoll eruption, Izu Islands. *P* is compression; *T* is tension; solid circles, compressions; open circles, dilatations. From Matsumura et al., 1991.

type of faulting with slight variations within the summit zone was dominant. The strike of the steeply dipping fault planes was roughly parallel to the U-shaped major fault zone within the crater.

During the 1986–1987 activity of basaltic volcano Oshima, Izu islands, focal mechanism solutions were determined for earthquakes that occurred before and after the fissure eruption. A zone to the north of the volcano was characterized by right-lateral strike-slip faultings; a zone to the south of the volcano was characterized by normal faults (Figure 9.7). The strike-slip fault motions within the first zone are consistent with the common tectonic stress field around Oshima Island; the peculiar normal fault motions at the second zone may reflect triggering by the fissure eruptions in the tensional stress field (Yamaoka et al., 1988).

Therefore, the fault nature of volcano-tectonic earthquakes may be various: thrust, normal fault, or strike-slip. It may be constrained by the regional or local tectonic stress systems. The focal mechanisms of relatively large earthquakes (magnitude about 5–6), related to fissure eruptions, revealed the coincidence of its one fault plane with the strike of the eruptive fissure.

9.1.2 Non-Double-Couple Model

The non-double-couple sources, or the sources with values $0.2 \leq \varepsilon \leq 0.5$ (ε is the relationship between the smallest and largest eigenvalues of the seismic moment tensor), were determined for a few volcano-tectonic earthquakes located at ring fault structures in Iceland (Ekström, 1994; Nettles and Ekström, 1998) and at Etna and Vesuvius, Italy (Panza and Saraó, 2000; Saraó et al., 2001). Figure 9.8 shows the example of focal mechanism determined for the 29 September *Mw* 5.6 main shock of earthquake sequence preceding the 30 September 1996 eruption of the subglacial volcano Grimsvøtn, Iceland, using broadband teleseismic *P*-waveforms inversion. The ε value for this moment tensor solution was equal to 0.30. Similar solutions were obtained for

Figure 9.6 Spatial variation of source time functions with strike-slip (SS) and dip-slip (DS) components for the magnitude 5.5 earthquake (9 July 1989) preceding the 13 July 1989 submarine Teishi Knoll eruption, Izu Islands. The seismic stations whose records were used in the inversion analyses are shown by filled circles. The arrows show the location of six point sources, which represents a 2 km × 2 km subfault.
From Takeo, 1992.

nine earthquakes that occurred from 1976 to 1994 around the Bárdarbunga caldera (Figure 9.9). The ε values for these earthquakes vary from 0.24 to 0.41, the magnitudes *Mw* range from 5.1 to 5.4. They occurred within the range of depths from 3.3 to 6.7 km. Nettles and Ekström (1998) consider that these non-double-couple solutions may reflect the faulting on an outward dipping cone-shaped ring fault, beneath the Bárdarbunga caldera. The model of rupture propagation on a cone-shaped fault is shown in Figure 9.10 and explains the rupturing as a series of subevents propagating around the surface of the cone.

Figure 9.7 Focal mechanism solutions of volcano-tectonic earthquakes that occurred before and after the 1986 fissure eruption of Oshima, Izu volcano. *M* marks the summit crater Mihara-yama; solid circles, compressions; open circles, dilatations. Arrows show the pressure and tension axis.
From Yamaoka, 1988.

Saraó et al. (2001) studied the variations in percentage of DC and CLVD components in the sources of volcano-tectonic earthquakes preceding the 1991 eruption at Etna, Sicily (See Section 5.1.2). Figure 9.11 shows that between August 1990 and 9 December 1991 the percentage of non-double-couple components of earthquakes, statistically significant at the 95% confidence level, increased before the beginning of the 15 December 1991 flank eruption. Panza and Saraó (2000) consider that this fact can be related to fluid movements.

Figure 9.8 Comparison of the calculated synthetic seismograms (dashed) with the broadband teleseismic *P*-waveforms (solid) for the 29 September 1996 volcano-tectonic earthquake preceding the September 30 flank eruption of Grimsvøtn volcano, Iceland. The focal mechanism corresponds to the full moment tensor solution, and the source time function is determined by the inversion. Maximum amplitudes (in microns) are shown to the right of the seismograms; arrows show picked first arrivals, and the thin vertical bars show the time window included in the inversion.
From Nettles and Ekström, 1998.

Therefore, the non-double-couple nature of the volcano-tectonic earthquake source may reflect the ring (or other complex) structure of the fault generating the earthquake (Nettles and Ekström, 1998). At the same time, every complex fault may be approximated by a linear structure, and for the majority of the events with the non-double-couple solution we have an adequate double-couple solution (Figure 9.12).

The presence of non-double-couple components can also depict a process in which a depressurizing magma body and crack formation leads to magma injection toward the shallower portions of the volcano (Panza and Saraó, 2000).

9.2 Source Spectral Characteristics of Volcano-Tectonic Earthquakes

The spectra of volcanic earthquakes are used generally in two types of applications: the total spectra of seismic records may be used for classification and monitoring of seismic activity and the body-wave (or surface wave) spectra may be used for study of source parameters of earthquakes.

Figure 9.9 Map showing CMT solutions for 10 earthquakes that occurred near Bárdarbunga volcano, Iceland. The earthquakes are highly non-double-couple, and have vertical *T* axes, indicative of horizontal compression.
From Nettles and Ekström, 1998.

Figure 9.10 Model of rupture propagation on a cone-shaped fault: a sub-source with reverse-faulting geometry (short arrows) occurs on each fault segment in succession (long arrow), around a specified portion of the cone.
From Nettles and Ekström, 1998.

Figure 9.11 Temporal variation of the percentage of DC (A) and CLVD (B) components of the earthquakes preceding the 15 December 1991 flank eruption at Etna, Sicily. The dashed lines represent the limits below which the solutions may be due to numerical noise. Numbering corresponds to temporal appearance of earthquakes.
From Saraó et al., 2001.

9.2.1 Spectra of Total Records of Volcano-Tectonic Earthquakes

Figures 9.13 and 9.14 show the seismograms, spectrograms, and velocity spectra of records for the four main types of volcano-tectonic earthquakes recorded during the active period at Redoubt, Alaska during 1989–1990 (Lahr et al., 1994). Each seismogram was of 40.6 s duration; the spectra were normalized for better comparison. The long-period events are characterized by a peak frequency near 1.5 Hz corresponding to *S* and surface waves. The short-period events that occurred at depths

Figure 9.12 Standard Harvard CMT solution showing two possible interpretations of the solution for the 29 September 1996 Grimsvøtn earthquake: a double-couple source (marked by stars) or a non-double-couple source (marked by filled ellipse area). *T* is tension; *P* is pressure; *N* is neutral axes for the best double-couple.

greater than 5 km below the crater floor are characterized by a peak between 6 and 8 Hz corresponding to the body-wave record and a low-frequency maximum between 1 and 3 Hz generated by short seismic coda. Shallower short-period events have two peaks: one, more intensive, between 1 and 3 Hz, and the second, lower, between 7 and 9 Hz. The first peak is produced by surface waves; the second, by body waves. A spectrum of a hybrid earthquake is characterized by a peak between 1 and 2 Hz.

These differences in spectral content allow a visual discrimination of different types of volcano-tectonic earthquakes for monitoring of volcano activity.

9.2.2 Spectral Source Characteristics of Volcano-Tectonic Earthquakes

The spectra of body and surface waves of volcano-tectonic earthquakes, as well as synthetic seismogram modeling, may be used for estimating of earthquake source parameters, such as the seismic moment M_0, stress drop $\Delta\sigma$, and spectral corner frequency fo. These parameters describe the variations of physical properties within the seismic volume below the volcano and can serve to discriminate between the

Figure 9.13 Spectrograms (A, C, E, and G) and seismograms (B, D, F, and H) for the typical volcano-tectonic earthquakes of Redoubt volcano, Alaska. *LP* are the long-period earthquakes; deep *VT* are the short-period events that occurred at depths greater than 5 km below the crater floor; shallow *VT* are the short-period events that occurred at depths less than 5 km below the crater floor; and Hybrid are the hybrid earthquakes.
From Lahr et al., 1994.

seismic events generated during the different types of volcanic activity. In this chapter, the author uses the estimations of source parameters from the Harvard centroid-moment tensor (CMT; Harvard database) and from body-wave spectra (circular dislocation Brune's model; Brune, 1970).

Brune's model (1970) allows the calculation of the radius of circular dislocation r, the seismic moment M_O, and stress drop $\Delta\sigma$ from the S-wave spectrum using the following formulas:

$$Mo = \frac{4\pi\rho\beta^3 S_o}{G(d)C(\omega)R(\theta,\varphi)} \tag{9.1}$$

Figure 9.14 Normalized velocity spectra of the typical volcano-tectonic earthquakes of Redoubt volcano shown in Figure 9.13.
From Lahr et al., 1994.

$$r = \frac{2.34\beta}{2\pi fc} \tag{9.2}$$

$$\Delta\sigma = \frac{7M_o}{16r^3} \tag{9.3}$$

Here $\pi = 3.14$; ρ is the assumed average density of the medium; β is the S-wave velocity; S_0 is the spectral density at zero frequency; fc is the corner frequency of the intersection of the low- and high-frequency branches of the spectrum; $G(d)$ is the function of geometrical spreading and taken as the inverse hypocentral distance d; $C(\omega) = 2$ is the local transfer function; the S-wave radiation pattern $R(\theta, \varphi)$ may be taken as 0.632 (the RMS average of the shear displacement radiation pattern over entire focal sphere). The values of S_0 and fc may be taken from the spectra of S-wave as is shown in Figure 9.15. This methodology may also applied to the spectra of P-waves (Hanks and Wyss, 1972).

Seismic Moment-magnitude Relationship

Figure 9.16 shows a "seismic moment−magnitude" dependence for the volcano-tectonic earthquakes related to central and lateral eruptions. In Figure 9.16A, the

Figure 9.15 Examples of *S*-waveforms and corresponding Fourier spectra of the 2000 Usu volcano-tectonic earthquakes recorded at station NBBT (vertical component). Corner frequencies *fo* and zero-frequency spectral density S_0 are shown. Arrows show the sample windows for spectra.
Courtesy of Japan Meteorological Agency.

parameters were taken from the Harvard CMT database for large 1980−1996 volcano-tectonic earthquakes with magnitude $Ms \geq 5.0$. It is seen that the seismic moments of volcano-tectonic earthquakes related to the central eruptions of Mount St. Helens and Pinatubo volcanoes are larger at the same magnitudes than the seismic moments of volcano-tectonic earthquakes related to the flank eruptions of Oshima, Miyakejima, Teishi Knoll, and Grimsvøtn volcanoes. The same tendency was observed at the basaltic volcano Alaid, Kurile Islands (Figure 9.16B) for smaller volcano-tectonic earthquakes related to its 1972 flank and 1981 central eruptions (Zobin, 1988).

This greater amount of long-period radiation expressed in larger values of seismic moments may be attributed to the difference in formation of the earthquake

Figure 9.16 Seismic moment−magnitude relations (A) for large ($M \geq 5$) volcano-tectonic earthquakes related to the central eruptions of volcanoes Pinatubo (1, 2) and Mount St. Helens (3) and the flank eruptions of Teishi Knoll (4), Grimsvøtn (5), Oshima (6), and Miyake-sima (7) volcanoes and (B) for the 1972 flank and 1981 central eruptions of Alaid volcano. The data for central eruptions are marked by diamonds and for flank eruptions, by stars. Source parameters for (A) are taken from the Harvard CMT database; for (B) from Zobin, 1988.

source for the two types of eruptions. Apparently, during a central eruption, the magma opens rather "smooth" fractures in a medium, which is probably already broken and possibly also hotter from repetitive intrusion of magma along the same path. During a flank eruption, a new eruptive fracture representing a branch of the main central channel is formed every time before the eruption, which causes the generation of asperities or barriers on the rupture. These may cause a higher amount of short-period waves (Zobin, 1988).

Fault area-seismic moment relationship. Figure 9.17 shows the relation between the circular fault area and seismic moments (circular dislocation Brune's model) for volcano-tectonic earthquakes associated with the eruptions at Volcán de

Figure 9.17 Relation between seismic moment and fault area for volcano-tectonic earthquakes associated with the eruptions at Volcán de Colima (1977–1978), Etna (1991 and 2002), and Usu (2000) volcanoes. The lines of equal stress drop are shown. Data were taken from Patanè et al., 1995; Zobin et al., 2002b, 2005a; Giampiccolo et al., 2007.

Colima, Etna, and Usu volcanoes (Patanè et al., 1995, Zobin et al., 2002b, 2005a, Giampiccolo et al., 2007). This plot demonstrates the absence of a linear relation between these two source parameters. The foci are characterized by a large range of stress drops changing from 0.01 to 100 MPa. Analysis of the frequency distribution of the logarithm of stress drops of the events presented in the largest sample of the events, presented in Figure 9.17, which was associated with the 2000 eruption of Usu volcano (Zobin et al., 2005a), showed a significant bimodal distribution, with stress drops lower or higher than 10 MPa (Figure 9.18). The high-stress-drop (HSD) events are more numerous (210 events) and are characterized by the mean geometrical value of log $\Delta\sigma = 1.59 \pm 0.24$. The low-stress-drop (LSD) 32 events are characterized by the mean geometrical value of log $\Delta\sigma = 0.37 \pm 0.43$.

Figure 9.18 Stress drop distribution for volcano-tectonic earthquakes associated with the 2000 eruption of Usu volcano. LST, low stress drop events; HST, high-stress-drop events. Based on the catalog of the Usu earthquakes prepared by V. Zobin.

9.3 Temporal Variations of the Source Spectral Characteristics and Focal Mechanisms of Volcano-Tectonic Earthquakes in the Course of Volcanic Activity

9.3.1 Corner Frequencies Variations

The detailed study of the body-wave spectra of volcano-tectonic earthquakes was carried out prior to the November 1998 andesitic lava extrusion at Volcán de Colima, México (Domínguez et al., 2001; Zobin et al., 2002b). Five earthquake swarms occurred during 1 year of seismic activity, from November 1997 to November 1998 (See Section 6.3.1). For *P*-wave spectra, two groups of events were observed: low-frequency events (LF) with values of *fo* from 2 to 7 Hz and high-frequency events (HF) with values of *fo* from 8 to 15 Hz (Figure 9.19B). These two groups were separated at the significance level 99%. They were observed simultaneously during all five swarms (Figure 9.19A). The same two groups of LF and HF events were observed also for *S*-wave spectra during the June–July 1988 swarm 4 when a three-component seismic station was in operation.

Figure 9.19 Variations in abundance of LF and HF events from swarm to swarm (A) and distribution of events with different *P*-wave corner frequencies during swarms 1, 3, 4, and 5 (B). Arrows show the eruptive events.
From Zobin et al., 2002b.

The relative abundances of LF and HF events changed from swarm to swarm (Figure 9.19A). HF events dominated swarms 1 and 2, whereas LF events were twice as abundant as HF events in swarm 4. The numbers of LF and HF events were practically equal for swarm 5, as they were for swarm 3. As was noted by Zobin et al. (2002b), a sharp decrease of LF events was observed 2 days before the Colima eruption.

The two groups of HF and LF events were observed also for volcano-tectonic earthquakes related to the 1975–1976 Tolbachik (see Section 5.3.1) and the

Figure 9.20 Temporal variations of *P*-wave corner frequencies for volcano-tectonic earthquakes associated with the 1975 fissure eruption of New Tolbachik volcanoes, Kamchatka (A), and of *S*-wave corner frequencies for volcano-tectonic earthquakes associated with the 1991 Etna flank eruption (B). Arrows show the eruptive events. Data from Zobin and Chirkova, 1984 and Patané et al., 1995.

1991–1993 Etna (see Section 5.2.1) eruptions (Figure 9.20). For New Tolbachik volcanoes, the HF events ($fo > 2.5$ Hz) were observed only before the opening of the fissure when the first cone was formed, 6 days before the beginning of the eruption. During the two swarms that occurred prior to the eruptions of the third and eighth cones, only the LF events were observed.

For the December 1991 Etna events, the two groups of LF and HF ($fo > 2.5$ Hz) events occurred simultaneously during the sharp increase in seismic activity at the initial stage of swarm activity on 14 December (Figure 9.20B). With the end of peak activity and the opening of the eruptive fissure, LF events disappeared for 2 h. Then both groups appeared again up to the beginning of the lava flow on 15 December.

Zobin et al. (2002b) suggested that a moving magma dyke could produce HF seismic events at distances less than 70 m from the dyke tip along existing fractures and trigger the fracturing of local faults at distances greater than 70–100 m from the dyke tip as LF events. Therefore, the occurrence of the HF events may indicate the moving of magma from depths, and the LF events may reflect the process of passageway constructing.

From this point of view, a sharp decrease of LF events 2 days before the Colima 1998 lava eruption indicates a magma uplift before the 1998 lava eruption

Figure 9.21 Temporal variations of the 2000 Usu volcano-tectonic earthquakes with different stress drops. I–IV, the stages of deformation, according to Takahashi et al. (2002). Stage I: 27 March, 20:00 hours–29 March, 18:00 hours, no deformations. Stage II: 29 March, 18:00 hours–30 March, 06:00 hours, beginning of slow deformation. Stage III: 30 March, 06:00 hours–30 March, 18:00 hours, sharp increase in deformation rate. Stage IV: 30 March, 18:00 hours–31 March, 13:00 hours, slow deformation near the volcano and fast deformation near the site of forthcoming eruption (KON); the eruption began at the end of the stage. Double arrows show the beginning of the eruption. The single arrows show the day duration. Based on the catalog of the Usu earthquakes prepared by V. Zobin.

along the prepared passageway. For the New Tolbachik volcanoes, magma may be considered as arriving close to the surface on 30 July 1975, when the HF events disappeared; during the following 6 days, a passageway was prepared along the new fissure, then the eruption at the first cone began. During the formation of the third and eighth cones, magma did not rise from depth; it moved laterally from the site of the first cone and from Plosky Tolbachik (See Chapter 5). Therefore, only LF events that ruptured the fissure for new cones were recorded.

9.3.2 Stress-Drop Variations

Figure 9.21 showed that the 2000 Usu earthquakes (See Chapter 7.2.1) were separated into two groups as characterized by high stress drops (HSD) and low stress drops (LSD). The temporal variations of the volcano-tectonic earthquakes with different stress drops were studied corresponding to the four stages in crustal

Figure 9.22 Variations of the LSD (<1 bar) and HSD events (A) and the azimuth of *P*-axis of the focal mechanism solutions (B) for volcano-tectonic earthquakes preceding the 1975 Tolbachik eruption. The time of occurrence of the two 2 July *Mw* 4.9 earthquakes is shown by the vertical line. The horizontal lines show the separation of the LSD and HSD events and the events with the *P*-axis azimuth less and more than 150°, respectively.
Data from Zobin and Chirkova, 1984 and Zobin and Ivanova, 1982.

deformation. These were recorded simultaneously with the seismic activity by the GPS stations situated at a distance of 2−3 km from the central crater. During the first stage (without any deformation), only HSD events occurred. The LSD events began to occur about 30−40 h later with the beginning of deformation of volcanic edifice. Then the two groups of volcano-tectonic earthquakes were recorded during the third and fourth stages. At all times, the HSD events were predominant in number (Zobin et al., 2005a).

Figure 9.22A shows the variations of LSD (< 1 bar) and HSD volcano-tectonic earthquakes preceding the 1975 Tolbachik eruption. It is seen that the both LSD and HSD events were recorded before the occurrence of two *Mw* 4.9 earthquakes on 2 July. Then HSD events disappeared and only LSD events were observed till the beginning of the eruption (Zobin and Gorelchik, 1982).

Zobin et al. (2005a) attributed the HSD earthquakes as the probable product of magmatic heat fracturing of host rocks. The relatively long LSD ruptures may be attributed to the slips along the preexisting and forming due to occurrence of large earthquakes fractures during magma movement.

Therefore, the knowledge of temporal variations of the spectral source parameters may allow the description of the main events during the eruptive process.

9.3.3 Stress Field Rotations

The stress field rotations obtained from focal mechanism solutions were recorded during the swarms of volcano-tectonic earthquakes associated with the 1975 Tolbachik fussure eruption in Kamchatka (Zobin, 1979b; Zobin and Gorelchik, 1982); the 1992 Crater Peak eruption in Alaska (Roman et al., 2004); and the 2004–2005 Mount St. Helens eruption (Lehto et al., 2010).

During the 1975 Tolbachik swarm, this rotation was observed just after two *Mw* 4.9 earthquakes had occurred on 2 July (Figure 9.22B). The stress field rotation was recorded simultaneously with the disappearance of HSD earthquakes (Figure 9.22A). It continued up to the beginning of the eruption on 6 July.

The 2004–2008 eruption of Mount St. Helens was preceded by a swarm of shallow volcano-tectonic earthquakes that began on 23 September 2004 (Section 7.1.1). The focal mechanism solutions were calculated for shallow volcano-tectonic earthquakes recorded during a background period (January 1999–July 2004) and during the first days of the eruption (early vent-clearing phase, 23 to 29 September, 2004). Figure 9.23 shows the normal and strike-slip faulting during the background period and on 23 September; strike-slip and reverse faulting on 24 September; and a mixture of strike-slip, reverse, and normal faulting on 25–29 September. The orientation of σ_1 beneath the volcano, as estimated from stress tensor inversions, was found to be sub-horizontal for all periods and oriented northeast–southwest during the background period, northwest–southeast on 24 September, and northeast–southwest on 25–29 September. It was suggested that the ephemeral approximately 90° change in σ_1 orientation was due to intrusion and inflation of a northeast–southwest-oriented dyke in the shallow crust before the eruption onset (Figure 9.24).

9.4 Seismo-Tectonic Deformations in the Volcanic Region

Knowledge of the focal mechanisms and scalar seismic moments of volcano-tectonic earthquakes, associated with an eruption, allows the determination of seismo-tectonic deformations (STD) within the volcanic edifice and surroundings. It is produced by the process of magma deep migration before, during, and after the eruption. Seismo-tectonic deformation ε_{ik} of the volume V that contained the earthquake foci can be estimated (Riznichenko, 1977) using the equation

$$\varepsilon_{ik} = \frac{\Sigma Mo(\theta_{ik})}{2\mu V} \tag{9.4}$$

where μ is accepted to be 3.2×10^{11} dyne-cm^{-2}; V is equal to 30 km^3; θ_{ik} is the unit tensor of seismic moment Mo; and Σ is the symbol of sum.

The components θ_{xx}, θ_{yy}, and θ_{zz} of the unit tensor of seismic moment are estimated using focal mechanism parameters through formulae

Figure 9.23 Variations in the fault nature of volcano-tectonic earthquakes during the background period and the initial vent-clearing phase of the 2004 Mount St. Helens eruption. (A) Time−depth plot of hypocenters for all well-located volcano-tectonic earthquakes (black crosses) and for well-located volcano-tectonic earthquakes with well-constrained focal mechanism solutions (gray circles) for the background period. (B) Time—depth plot of hypocenters for all well-located volcano-tectonic earthquakes and for those with well-constrained focal mechanism solutions for the vent-clearing phase, 23 September through 29 September, 2004. (C) Histograms of faulting types as defined in text. Note the sharp decrease in normal faulting on 24 September and its reappearance on 25 September. Percentages of faulting type for each time period are also shown (e.g., four 23 September events show 100% normal faulting). The background period covers a time span of about 5 years, while the vent-clearing period covers a time span of 7 days. All depths are relative to a datum of 2.2 km a.s.l.
From Lehto et al., 2010.

Figure 9.24 A proposed model of intrusion and inflation of a magma-filled dyke into the shallow subsurface during the initial phase of the 2004 Mount St. Helens eruption shown in both cross-section and map view for (A) 23 September 2004; (B) 24 September 2004; (C) 25 September 2004; and (D) 26–29 September 2004. Gray arrows represent either upward movement (vertical arrows) or inflation/deflation (horizontal arrows) of the dyke, while black arrows show orientation of maximum compression. *N*, *R*, and *SS* stand for normal faulting, reverse faulting, and strike-slip faulting events, respectively. Dots labeled *P* and *T* represent the positions of the *P*- and *T*-axes, respectively. Dashed box represents the approximate location of the seismogenic volume hosting the volcano-tectonic events. Lower hemisphere projections of focal mechanisms shown are schematic and do not necessarily reflect the exact hypocenter locations.
From Lehto et al., 2010.

$$\theta_{xx} = \sin^2 \sigma_T \sin^2 \alpha_T - \sin^2 \sigma_P \sin^2 \alpha_P \tag{9.5}$$

$$\theta_{yy} = \sin^2 \sigma_T \cos^2 \alpha_T - \sin^2 \sigma_P \cos^2 \alpha_P \tag{9.6}$$

$$\theta_{zz} = \cos^2 \sigma_T - \cos^2 \alpha_P \tag{9.7}$$

where σ_P and σ_T are the angles of P and T axes with the vertical; α_P and α_T are the azimuthal angles of the P and T axes. The X-axis is oriented to the east, the Y-axis to the north, and the Z-axis upward. The STD was calculated for two volcanic zones: for the 1975–1978 volcano-tectonic activity in the region of Tolbachik fissure eruption, Kamchatka (Zobin, 1990), and for the 1997–1998 volcanic prelava unrest at Volcán de Colima, México (Zobin et al., 2002a). Figures 9.25 and 9.26 illustrate the results.

The 1975–1978 seismo-volcanic activity in the region of Tolbachik fissure eruption was characterized by strong 1975–1976 volcanic activity along the new fissure and by the following posteruptive seismic activity within the surroundings (see Section 5.3.1). The magnitude 4–5 seismic events were recorded at the site of the cones just before the beginning of the eruption and at the distance of 3–50 km from the active cones just after the ceasing of the eruption. Figure 9.25 shows the spatial distribution of the three components ε_{xx}, ε_{yy}, and ε_{zz} of extensional deformations that were calculated for all recorded earthquakes of magnitude $mb > 3.7$ that occurred within the upper crust layer from 0 to 20 km. In the eruptive zone I, the largest compressional deformations dominated with a value of -5.4×10^{-5}; while the largest extensional deformations were observed in the latitudinal direction with the maximum value of 3.6×10^{-5}. The vertical deformations reached 1.6×10^{-5}. In the periphery, the maximum deformations were observed in zone IV situated 40 km to the north from the eruptive sites, where the compression was dominant reaching a value of -1.8×10^{-6} while the vertical deformation was of 3.4×10^{-6}.

Generally, the comparison of seismic activity development (Section 5.3.1) and the scheme of Figure 9.25 allow the reconstruction of the process of total crustal deformation within the seismic volume in the region of the fissure eruption. The seismic activity started in zone I, the eruption zone. This had led to considerable bulging of the earth's surface producing extensional deformations in latitudinal and compressional in meridional directions. Later, after the termination of the eruption the seismic activity spread radially from Plosky Tolbachik (zones II and III in Figure 9.25). Zone II extends along the Upper Pleistocene-Holocene extrusive domes to the east of Plosky Tolbachik. Zone III extends along the western border of distribution of the Upper Pleistocene-Holocene cinder and lava cones. Zones II and III were notable for subsidence of different amplitudes that can be caused by discharge of hydraulically connected magma chambers (Fedotov, 1984b). If this assumption is correct, then the outlines of the subsidence zone in Figure 9.25C can serve as the boundary of the zone of the near-surface magma chambers that feed the Klyuchevskaya group of volcanoes. Collapses of the near-surface rocks along the margin of the magma chambers could obviously produce compensational

Figure 9.25 Distribution of extensional deformations in the region of the 1975–1976 Tolbachik fissure eruption, Kamchatka in latitudinal (east-west) (A), meridional (north-south) (B), and vertical (C) directions. Kl is Klyuchevskoy, Bz is Bezymianny, Pl. T is Plosky Tolbachik, and N.T.V is New Tolbachik volcanoes. N and S indicated the north and south groups of new cones of New Tolbachik volcanoes. 1, the isolines of deformations; 2, the regions of negative deformations; 3, the volcanoes; and 4, a boundary of zones I to V. From Zobin, 1990.

bulging in zones IV and V that resulted in the upsurge of seismic activity in 1977–1978 (Zobin, 1990).

The 1997–1998 prelava eruption unrest of Volcán de Colima, México (see Chapter 6) was more local and produced lesser deformations, mainly within the volcanic edifice from sea level up to the crater at an altitude of 3860 m (Zobin

et al., 2002a). Figure 9.26 shows the temporal variations of three components of extensional deformations 1 year before the eruption. The calculated deformations vary from 1.3×10^{-11} to 2.7×10^{-9}. The maximum deformation was observed in June–July 1998 (swarm 4), and the minimum deformation was observed in March 1998 (swarm 2). The process of deformation was rather complicated. The first earthquake swarm of November–December 1997 resulted in a north-south horizontal elongation that is twice greater than the east-west horizontal shortening. The volume was also subject to vertical shortening. In March and May 1998 (swarms 2 and 3), uniform horizontal north-south and east-west elongations, accompanied by vertical elongation of the volume were observed. In June–July 1998 (fourth

Figure 9.26 Variations of principal components ε_{xx}, ε_{yy}, and ε_{zz} of STD from swarm to swarm 1 year before the November 1998 lava eruption at Volcán de Colima. 1–5, the numbers of swarms according to their temporal appearance. Zero-line is shown by the dashed line.
From Zobin et al., 2002a.

swarm), the situation of November–December 1997 (first swarm) was repeated with the north-south horizontal elongation greater than the east-west horizontal shortening of the volume and with an intensive vertical shortening. During the last swarm of October–November 1998, the slight east-west elongation of the seismic volume was accompanied by strong north-south shortening and a very slight vertical shortening.

Quantification of the deformation process within the volume for seismic foci recorded before the November 1998 eruption of Volcán de Colima enables interpretation of the process, thus giving volcanological meaning to the deformation terms. The seismic volume can be considered to be the volcanic edifice. The process of horizontal elongation is interpreted as the process opening a magma passageway; vertical shortening in the seismic volume is interpreted to reflect magma ascent. The north-south component of deformation coincides with the orientation of the Colima rift and the east-west component of deformation can be associated with the adjusting northeast-southwest Tamazula fault system (Zobin et al., 2002a).

Based on these assumptions, in November–December 1997, magma began to rise and to form a passageway along the Tamazula fault system. In March and May 1998, the vertical movement of magma practically stopped, and magma began to fill existing fractures in both the Colima rift and the Tamazula fault system. In June–July 1998, a new strong phase of magma ascent began and the passageway was formed along the Tamazula fault system. During the last swarm in October–November 1998, the final passageway was formed, with a north-south orientation within the Colima rift system, and final ascent of magma to the surface was completed (Zobin et al., 2002a).

The study of STD gives us the possibility for a quantitative comparison of two eruptions. The larger magnitude of events recorded during the seismic activity preceding the Tolbachik eruption yielded larger values of STD that reached 10^{-5}. This difference in four orders of magnitude with the Colima deformations agrees well with the difference in the magmatic volume of erupted products (2.2 km^3 for the Tolbachik eruption versus 0.04 km^3 for the 1998–1999 Colima eruption) (Fedotov, 1984a; Zobin et al., 2002c).

10 Significant Volcano-Tectonic Earthquakes and Their Role in Volcanic Processes

Omori (1912) wrote that a volcanic earthquake could never be equal to a large tectonic shock. He considered that the zones of active volcanoes "may be regarded as a sort of safety valves to, or lines of least resistance in, the Earth's crust, preventing an accumulation in their localities of underground stress sufficient to result in a gigantic earthquake. Hence it is that extremely violent or very extensive earthquakes never originate from under volcanoes."

Long-term observations showed that volcanic earthquakes are characterized by significantly lower M_{max} and a lower frequency of large earthquake appearance than tectonic earthquakes. The majority of volcanic earthquakes are small; their magnitudes do not exceed 2 or 3. At the same time, significant (moderate-size and large) volcanic earthquakes may occur and produce a local effect of high intensity (Abe, 1979; Zobin, 2000; Yokoyama, 2001).

Machado et al. (1962) described the macroseismic effects that were produced by the 13–14 May 1958 volcano-tectonic earthquake sequence at Fayal Island during the flank eruption of Fayal volcano. About 450 earthquakes were felt on Fayal; many of them had a Modified Mercalli (MM) scale intensity of X (Figure 10.1A). Of course, these effects were very local, covering an area of about 2–3 km^2, and they were strongly multiplied by a surface layer of soft volcanic rocks on the flanks of the volcano. Figure 10.1B shows also the macroseismic effect produced by the M_L 3.6 earthquake occurring within the edifice of Vesuvius volcano on 9 October 1999 (Cubellis et al., 2007). It was felt over a wide area within about 25 km from the crater and caused fear and anxiety among the people. A macroseismic study of this event was carried out from an analysis of questionnaires sent out to all secondary schools in each municipality of the Vesuvian and Neapolitan area and surrounding towns. The contours of felt index areas illustrate that maximum intensities, estimated as VI MCS (Mercalli-Cancani-Sieberg scale, VI MCS = V MM, Tiedemann, 1992), correspond to the volcanic edifice from which they decrease radially.

Blong (1984) wrote about at least 29 deaths associated with the secondary effects (landslides, building collapses) of volcanogenic earthquakes during the eruptions of Sakurajima, Kyushu (1914), and Api Siau, Indonesia (1976). Therefore, a seismic hazard of volcano-tectonic earthquakes for the settlements on the slope of volcano may be a reality.

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00010-4
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Figure 10.1 Macroseismic effect of the volcano-tectonic earthquakes observed (A) at Fayal volcano during the 13–14 May 1958 seismic crisis associated with the flank eruption of the volcano and (B) at Vesuvius volcano during the largest event of the August–October 1999 seismic sequence. (A) The numbers VI to X are the Modified Mercalli scale intensities; C indicates the caldera. Data from Machado et al. (1962). (B) The MCS intensity VI was estimated as I_0 for the magnitude M_L 3.6 earthquake occurring on 9 October 1999. Felt index map is shown. The felt index is the percentage response to the question *Did you feel the earthquake*? The area of maximum value (75%) covers the whole Vesuvian area. The felt index weakens strongly in northeast and southeast directions.
Data from Cubellis et al., 2007.

10.1 Selection of Significant Volcano-Tectonic Earthquakes that Occurred in the Twentieth Century

Zobin (2001) listed the significant ($M \geq 4.5$) earthquakes, which occurred near volcanoes from 1910 to 2005 (Table 10.1). They are very rare, rarer than magnitude 7.5–8.5 tectonic earthquakes. Only three magnitude *Ms* 7.0–7.1 volcanic earthquakes were recorded during the century (Abe, 1979, 1992) while the annual average number of $Mw > 7.0$ tectonic events is about 15 (Lay and Wallace, 1995).

The catalog of Table 10.1 contains 32 events. Three of them are with magnitude $Mw \geq 7.0$, four with magnitude from 6.0 to 6.9, and 17 events with magnitude from 5.0 to 5.9. This catalog was prepared for significant seismic events associated with 32 eruptions of 26 volcanoes (Figure 10.2). The earthquakes were associated with the eruptions of different types, and they occurred at different stages of the eruptions. It is unlikely that it is the complete catalog for the century, but it includes all available published data. Unfortunately, the large amount of information about the seismic events of the large eruptions before 1960s was out of publication.

Only the largest earthquakes for each eruption are listed in Table 10.1. These selected earthquakes had different magnitudes, such as *Ms*, *mb*, *MLH*, *Mw*, *Ks*, and so forth. It was necessary to prepare the homogeneous catalog according to a single magnitude. The magnitude *mb* was not representative, because it became saturated at the value of magnitude $mb = 5.8$ (Figure 10.3). From two other magnitudes, *Ms* and *Mw*, the magnitude *Mw* was used as a basic. The moment magnitudes *Mw* estimated by Harvard Centroid-Moment Tensor (CMT) group and by the author (Zobin, 1979a, 1988; Zobin and Levina, 1998) and the International Seismological Centre (ISC) magnitudes *Ms* for 10 events were used for the construction of the relationship between *Mw* and *Ms* for volcanic earthquakes (Figure 10.4). It was obtained from the equation

$$Mw = 0.86Ms + 1.00 \tag{10.1}$$

which was used then for the classification of earthquakes with an unknown seismic moment. The final magnitude limit for the catalog is $Mw \geq 4.5$.

10.2 Focal Rupturing of Significant Volcano-Tectonic Earthquakes and Its Role in Volcanic Processes

The rupture process of three significant volcano-tectonic earthquakes (Nos. 23, 27, and 28 of Table 10.1) was investigated by Takeo (1992), Zobin and Levina (1998), and Zobin (1999). These reconstructions of rupture process were realized by the inversion of the near-field strong-motion records for event No. 23 (Teishi Knoll volcano) and the broadband teleseismic *P*-waveforms for events No. 27 (Akademia Nauk volcano) and No. 28 (Grimsvøtn volcano). They were interpreted in terms of

Table 10.1 List of Significant (Moderate-Size and Large) Volcanic Earthquakes, 1910–2005

#	yyyymmdd	hhmm	Lat. (deg)	Long. (deg)	Dep. (km)	*mb*	*Ms*	*Mw*	Volcano
1	19100624	0749					5.1[1]	5.4[a]	Usu
2	19120608	0736				6.8[2]	7.0[2]	7.0[a]	Katmai
3	19140112	1028					7.0[1]	7.0[a]	Sakurajima
4	19430215	0139					4.5[3]	4.9[a]	Parícutin
5	19560330	0611					5.0[4]	5.3[a]	Bezymianny
6	19620826	0648						6.1[5]	Miyakejima
7	19641111	1906				5.5[7]		5.2[6]	Sheveluch
8	19680613	0733				5.4[7]		5.4[6]	Fernandina
9	19680717	0623				5.1[7]	4.8[b]	5.1[a]	Arenal
10	19700918	0206				5.1[7]	5.3[12]	5.7[5]	Beerenberg
11	19730711	2247				5.3[7]	5.0[12]	5.9[8]	Curacoa Reef
12	19730714	0115				4.2[7]		4.6[5]	Tiatia
13	19740129	1334	[9]37.77	15.08	15	4.5[13]	4.3[9]	4.7[a]	Etna
14	19750702	0734	[15]55.41	160.15	15	4.7[7]		4.9[6]	Tolbachik
15	19770106	1833				5.2[7]		5.1[10]	Nyiragongo
16	19771004	1339[c]					4.3[1]	4.7[a]	Usu
17	19800518	1532	[16]46.21	−122.19	3	4.6[7]	5.2[7]	5.4[6]	St. Helens
18	19830719	1044				5.2[7]	4.5[12]	4.9[a]	Una Una
19	19831003	1333	[17]34.00	139.52	15	5.8[7]	6.2[7]	6.0[10]	Miyakejima
20	19850107	2153	[7]71.13	−7.52	10	4.9[7]	4.3[7]	5.0[10]	Beerenberg
21	19861122	0041	[17]34.53	139.36	15	5.8[7]	5.9[7]	5.9[10]	Oshima
22	19890224	1236				5.2[7]	4.8[7]	5.2[10]	Lonquimay
23	19890709	0209	[17]34.99	139.11	3	5.0[7]	5.2[7]	5.2[10]	Teisi Knoll
24	19910615	1115	[12]15.12	120.35	10	5.7[7]	5.4[7]	5.6[10]	Pinatubo
25	19910815	1216				5.3[7]	5.6[7]	5.8[a]	Hudson
26	19911215	2000	[18]37.74	14.97	2	4.3[7]	4.3[13]	4.7[a]	Etna
27	19960101	0957	[14]53.90	159.43	0	5.8[7]	6.7[7]	7.1[14]	Akademia Nauk
28	19960929	1048	[19]64.68	−17.49	4	5.2[7]	5.4[7]	5.6[10]	Grimsvøtn
29	20000331	1812	[17]42.50	140.83	7	4.9[7]	4.6[7]	4.9[20]	Usu
30	20000701	0701	[17]34.21	139.22	15	5.9[7]	6.1[7]	6.2[20]	Miyakejima
31	20000730	1225	[17]33.96	139.40	18	5.9[7]	6.5[7]	6.4[20]	Miyakejima
32	20050924	1924	[7]12.39	40.57	13	5.0[7]	5.1[7]	5.5[10]	Dabbahu

References: (1) Abe (1979); (2) Abe (1992); (3) Yokoyama and de la Cruz-Reyna (1990); (4) Okada (1983); (5) Zobin (1979a); (6) Zobin (1988); (7) ISC; (8) Gibowicz et al. (1974); (9) Bottari et al. (1975); (10) Harvard CMT; (11) Wano and Okada (1980); (12) NEIS; (13) MOS; (14) Zobin and Levina (1998); (15) Zobin (1979b); (16) University of Washington, Data Base; (17) JMA; (18) Ferrucci and Patane (1993); (19) Icelandic Meteorological Office; (20) NIED, Tsukuba.

[a] *Mw* was calculated from magnitude *Ms* using the relation $Mw = 0.86\ Ms + 1.00$.

[b] *Ms* was calculated from the magnitude mb using the relation $Ms = 2.5\ mb - 8.0$ (Zobin, 1999).

[c] The first earthquake of magnitude *Ms* 4.3 Usu earthquakes from a sequence of 16 events of the same magnitude which occurred from 4 October 1977 to 22 August 1978.

Figure 10.2 The map of volcanoes under study. The volcanoes are shown by black diamonds with numbers according to Table 10.1.

barriers or asperities destruction (Ruff, 1983) along the fault plane. The barriers and asperities may be considered as the ancient magmatic bodies that could be destructed by forming the magma intrusion passageway.

10.2.1 Rupturing of the Magnitude Mw *5.2 Earthquake Preceding the 1989 Teishi Knoll Submarine Eruption*

This earthquake occurred at the final stage of an earthquake swarm 4 days before the eruption (see Chapter 5). Its focal mechanism shows that one of its nodal planes coincides with the linear distribution of epicenters of earthquakes following this event (Figure 10.5). This nodal plane striking in the direction of N86°W was taken as the fault plane. Takeo (1992) studied the rupture process of the source of the *Mw* 5.2 earthquake by the inversion of digital three-component accelerograms recorded at near seismic stations (Figure 10.5). Figure 10.6 shows the comparison of synthetics with records.

Takeo (1992) showed that the *Mw* 5.2 shock (No. 23) was followed 45 s later by the *M* 4.7 earthquake; he studied these two events together as event 1 and event 2 (Figure 10.7). The rupture of event 1 initiated in the eastern side subfault as a right-lateral strike−slip motion. Then the rupture stopped due to a barrier; Takeo assumes that this barrier could be an old dyke formation. The excitation of long-period seismic waves stopped for 1.5 s after the first stage, but the high-frequency seismic waves excited during the second stage are larger than during other stages.

Therefore, it can be suggested that only small-scale fractures took place in the barrier during the second stage. Because the rupture in the third stage involved a little reverse component with the dominant right-lateral strike−slip motion, the

Figure 10.3 Relationship between the magnitudes *Ms* and *mb* for volcanic earthquakes. The values of *Ms* and *mb* are taken from *Bulletin of ISC*. It is seen that the saturation of magnitude *mb* at value *mb* = 5.8.
Data from Zobin, 2001.

barrier seems to have been subjected to a vertical shear stress due to the magma intrusion beneath the source zone. The rupture extended westward and the source area of event 1 reached 4 km in length and 2 km in width. The third stage rupture was prevented by the barrier, which consisted of the old dyke formation existing around the western end of event 1. This barrier ended the rupture process of event 1, but a high stress was concentrated in the barrier. A new fracture nucleated 45 s later (event 2) and spread out westward.

Therefore, the rupturing, produced by the *Mw* 5.2 earthquake, destroyed the old dyke formation and opened the passageway for the magma intrusion. The large stress drop on the east side subfault of event 1 indicates a high stress concentration before its nucleation, due to the magma intrusion. Takeo (1992) guesses that the large stress drop just beneath the Teishi Knoll would assist the magma ascent and may have triggered the eruption.

Figure 10.4 Relationship between the magnitudes *Ms* and *Mw* for volcanic earthquakes. The line of regression is shown. The coefficient of correlation is 0.927.
Data from Zobin, 2001.

10.2.2 Rupturing of the Magnitude Mw *7.1 Earthquake Preceding the 1996 Akademia Nauk Volcano Subaqual Eruption*

The explosive eruption in caldera lake of Akademia Nauk volcano, Kamchatka (see Chapter 2), occurred after 4800 years of dormancy and was preceded and accompanied by an intensive sequence of earthquakes at depths from 0 to 50 km, beginning on 1 January (Figure 10.8). This sequence spread from Akademia Nauk volcano to the southwest. The largest earthquake of magnitude $Mw = 7.1$ (No. 27 of Table 10.1) was recorded on 1 January, about 4 h after the beginning of the seismic sequence and about 14 h before the beginning of the eruption.

This earthquake had a strike−slip nature, with the N14°E fault plane, oriented along the regional fault connecting the epicenter with the new eruptive site in Karymskoe lake (Zobin and Levina, 1998). This direction of the fault plane fitted well with the distribution of the epicenters of events that occurred before the large shock.

A finite-fault, broadband teleseismic *P*-waveform inversion (Figure 10.9) allowed the reconstruction of a dislocation model of the faulting process. The

Figure 10.5 Focal mechanism solution for event 23 and the epicenters of earthquakes following this event for 13 h. Large circles show four seismic stations whose records were used in the inversion. The nodal plane II was chosen as the fault plane.
Data from Takeo, 1992.

rupture model is presented in terms of a contour map of the final slip of the fault (Figure 10.10). Two main asperities, outlined by the slip contour of 125 cm, were destroyed; one at depths between 0 and 15 km and the second at depths between 25 and 35 km. The majority of small earthquake foci were located away from the maximum asperities and filled a space between two zones of asperities at depths from 12 to 20 km. Probably this large seismic event represented two shocks, one shallow and the second at a depth of about 30 km (Zobin and Levina, 1998).

The great amount of juvenile calc-alkaline basalt, representing about 95% of the ejected material (Belousov and Belousova, 2001), supports the inference of the deeper seismic source. It seems possible that during the large earthquake, an ancient deep-seated magmatic column was destroyed and a new system of faults was formed that would act as a magma passageway.

10.2.3 Rupturing of the Magnitude Mw *5.6 Earthquake Preceding the 1996 Grimsvøtn Volcano Subglacial Eruption*

This earthquake (No. 28 in Table 10.1) occurred as the mainshock of preceding earthquake sequence. It occurred 36 h before the eruption (see Chapter 2). The earthquake had an inverse-type nature, with the sub-meridional fault planes (Harvard, 1977–1999). This direction of the fault planes fitted well with the

Figure 10.6 Comparison of observed ground displacements of the *Mw* 5.2 Teishi Knoll earthquake inferred from accelerograms by integrating twice with synthetics. Data from Takeo, 1992.

distribution of the epicenters of events that occurred just after the mainshock (Figure 10.11).

A finite-fault, broadband teleseismic *P*-waveform inversion allowed the reconstruction of a dislocation model of the faulting process. The rupture model is presented in terms of a contour map of the final slip on the fault (Figure 10.12).

Figure 10.7 Schematic rupture process of the *Mw* 5.2 earthquake (event 1) and the following *M* 4.7 earthquake (event 2) that preceded the 1989 submarine eruption of Teishi Knoll, Izu Islands.
Data from Takeo, 1992.

The inversion showed that the rupture had developed downdip from the mainshock hypocenter and the main asperity was ruptured at depths between 4 and 6 km. A comparison with the deep distribution of aftershocks showed that their hypocenters were situated above the destroyed asperities at depths between 0 and 4 km. These space–time seismic patterns and the results of the finite-fault inversion may be interpreted in the following way. There was a magma chamber under Bárdarbunga volcano covered by the strong rocks, or asperities. The mainshock of 29 September broke the asperity and formed a path for magma transport. Then a fissure formed connecting the opened magma chamber with the feeding structures of Grimsvøtn volcano, and magma was injected laterally to the site of eruption. The continuous seismic activity during the third period was related to the collapse of rocks around the released parts of the magma chamber. The role of the *Mw* 5.6 earthquake in this sequence was decisive.

Figure 10.8 Seismicity preceding the $Mw = 7.1$ earthquake related to the eruption of Akademia Nauk Volcano (No. 27). The foreshocks (filled circles) that were recorded from 06:20 to 09:57 h of 1 January 1996 are shown. The volcanoes Karymsky (KV) and Akademia Nauk (AN) are shown as open stars. The epicenter of the $Mw = 7.1$ earthquake is shown by the filled star. The focal mechanism solution for this event is shown in the left corner. Data from the Kamchatka Regional Earthquake Catalog. The focal mechanism was published in Zobin and Levina, 1998.

10.3 The Magnitude 7 Volcano-Tectonic Earthquakes in Volcanic Processes

For more ancient large volcano-tectonic earthquakes we had no rupture reconstructions. At the same time, their participation in eruptive processes is of interest. Here

Figure 10.9 The comparison of observed (solid line) and synthetic (dashed line) *P*-waveforms of the *Mw* 7.1 earthquake of 1 January 1996 preceding the subaqual Akademia Nauk volcano eruption. The epicentral distance (in degrees) and azimuth from epicenter (in degrees) are indicated for each seismogram pair.
Data from Zobin and Levina, 1998.

the author discusses two of them, the largest events with magnitude $Mw = 7.0$ (Nos. 2 and 3 in Table 10.1).

10.3.1 Event No. 2, Katmai, Alaska

The 1912 Katmai eruption was the most voluminous eruption of the twentieth century, producing about 30 km^3 of products. The formerly cone-shaped summit of Mount Katmai vanished, leaving in its place a caldera 3 km in diameter and approximately 1 km deep. The collapse was apparently caused by withdrawal of supporting magma from beneath Mount Katmai toward Novarupta, the actual vent of eruption, 10 km west of Mount Katmai (Abe, 1992). There was no information about previous activity of Katmai in historical time.

The seismic activity was recorded beginning about 5 days before the eruption, but the strongest earthquakes were observed during the eruptive activity. The

Figure 10.10 The comparison of the dislocation model, obtained for the *Mw* 7.1 earthquake preceding the January 1996 Akademia Nauk volcano eruption, with seismic and volcanic characteristics. Solid circles denote the hypocenters of earthquakes with magnitude $M > 3.0$ recorded before the beginning of the eruption. The hypocenter of the mainshock is indicated by a solid hexagon. Cumulative slip is contoured at 75-cm intervals beginning from 125 cm. A.N.V. is the Akademia Nauk volcano; K.V. is the Karymsky volcano. The border between the crust and mantle is shown by the dashed line.
Data from Zobin and Levina, 1998.

eruption began on 5 June 1912 with frequent minor explosions. Then a series of big explosions was observed from 13 h of 6 June to 8 July. The eruption continued until 21 July (Simkin and Siebert, 1994), but without significant explosions. The sequence of strong earthquakes ($Ms \geq 6.0$) was recorded beginning from the second tremendous explosion of 6 June, and the largest event with magnitude $Ms = 7.0$ occurred at 21:26 h (local time) on 7 June, at the final part of the violent phase of the eruption, when from 21 to 22 h, 42% of the total seismic energy was released (Abe, 1992).

Abe (1992) considers that a series of substantial collapses of Mount Katmai occurred during this seismic climax. Therefore, the author speculates that the $Mw = 7.0$ seismic event, which was recorded at the seismic climax and before the third last great explosion, was the decisive one in the Mount Katmai caldera collapse, opening the fault as passageway for withdrawal of supporting magma from beneath Mount Katmai toward Novarupta.

10.3.2 Event No. 3, Sakurajima, Japan

The 1914 Sakurajima eruption was also one of the largest eruptions of the century (see Chapter 6). The eruption began on 12 January at about 08:00 h (local time) with the emissions of white smoke at the South Peak, and 2 h later explosions

Figure 10.11 The 29 September 1996 mainshock epicenter (solid triangle), its focal mechanism solution and the epicenters of its aftershocks (open circles), which preceded and accompanied the subglacial eruption of Grimsvøtn volcano, Iceland. The linear distribution of the aftershocks that occurred during the first 19 h of activity coincides with the directions of two nodal planes of the focal mechanism solution. The black heavy line shows the volcanic fissure, which was formed during the eruption. The volcano calderas are shown by heavy-line ovals.
Data from Zobin, 1999.

occurred at the western and eastern slopes, where vents had opened. Just after midnight on 13 January, lava began to effuse from the craterlets on both the eastern and western slopes (Yokoyama, 1997). It was the first flank eruption from vents situated on the west and east flanks of the volcano in historical time. Earlier eruptions were recorded only from the summit crater Minami-dake or from the flank vents situated on the northeast, southwest, or south slopes of the volcano (Simkin and Siebert, 1994).

Seismic activity near the volcano was recorded beginning 9 January. Earthquakes increased in number and magnitude, with the maximum activity on 11 January. After the opening of vents in the morning of 12 January, earthquake activity rapidly decreased. Then, at 18:30 h of 12 January, the $Mw = 7.0$ shock occurred. About 26 h later the great lava extrusion began (Yokoyama, 1997). The author speculates that this large earthquake opened the passageway for magma and was decisive in the development of eruption from rather calm flank explosions to a greater volume lava eruption.

The analysis of the five largest volcanic earthquakes shows that all events were associated with eruptions of quasi-dormant volcanoes (Katmai and Akademia Nauk) or with a new type of activity in historical time (Teishi Knoll, Grimsvøtn and Sakurajima). The fault nature (event Nos. 23, 27, and 28) and the temporal position of the events in the eruption process allow us to speculate that these earthquakes may have formed new passageways for magma movement.

Figure 10.12 Cross-section beneath the caldera of Bardarbunga. The circles are hypocenters of the aftershocks ($M_L \geq 3.0$); the star is the hypocenter of the mainshock. The slip contours are shown for 20 and 40 cm and outline the asperity broken during the rupturing of the mainshock fault. The zone of a possible magma chamber is shown below the asperities. Data from Zobin, 1999.

10.4 Seismic Hazard of Significant Volcano-Tectonic Earthquakes

The traditional estimation of seismic hazard includes (1) the estimation of maximum magnitude M_{max}; (2) its recurrence time; and (3) the law of earthquake intensity decrease with distance for a variety of magnitudes (e.g., Basham et al., 1982). There was no possibility of a long-term sequence of volcano-tectonic earthquakes for a volcano, so we discuss the seismic hazard for a type of eruption. Table 10.2 describes the types of eruptions and the position of the seismic event in the eruptive process. Classification of 27 of total 32 eruptions was taken from the catalog of volcanoes of the world by Simkin and Siebert (1994) and GVN (2011). The position of the largest seismic event in the eruptive process was estimated from publications about these eruptions.

Significant earthquakes that occurred near volcanoes but without relation to eruptive activity (Kubotera, 1988; Abe, 1979; McNutt, 1996) were not included in

this study. Among them were the *Ms* 6.3 1916 earthquake near Asama volcano and the *Ms* 6.1 1975 earthquake to the north of Aso caldera. There was a problem with the magnitudes estimated by near stations for rather small events, because as a rule, they were overestimated. For example, the magnitude of the earthquake of 7 July 1990, related to the eruption of Unzen volcano, was determined by near stations as *M* 4.8 (Nakada et al., 1999). At the same time, ISC gave the magnitude *mb* 4.4 and no estimation of *Ms*. These doubtful events were also not included in the catalog. The 1975 *Ms* 7.1 Kalapana, Hawaii, earthquake had triggered the very small summit eruption of Kilauea (Klein et al., 1987). This eruption was very insignificant and short; the real relation between the earthquake and eruption is unclear. McNutt (1996) did not include it in his list of large seismic events near volcanoes, but only mentioned it among the large regional earthquakes that could trigger the eruptions. This earthquake was not included in Table 10.1.

The data for recurrence time were taken from the catalog by Simkin and Siebert (1994). The data for calculation of the law of decreasing of intensity with distance were taken for three Japanese earthquakes from *Bulletin ISC* (Table 10.3).

For all regressions calculated in this study, the correlation was significant at the 95% confidence level.

10.4.1 Maximum Magnitude M_{max}

The problem of maximum earthquake magnitude for any region or local zone is very important in the evaluation of seismic hazard. The size of the maximum magnitude is decisive in the prediction of the destructive effects of seismic activity. The maximum magnitude may be estimated (1) by consideration of the maximum fault area that can break in a single event; (2) by the magnitude truncation in the observed seismicity for source zones in which the return period for the maximum magnitude is shorter than the observation period (e.g., Basham et al., 1982); or (3) by statistical study of the catalog (e.g., Kagan, 1997). However, for our case, these types of evidence are not available. We work not with a single volcano, but with a group of volcanoes all over the world. Therefore, we estimate the maximum magnitude $M_{\max}$ for this set of volcanoes combining them in two ways, based on the data of Tables 10.1 and 10.2: for (1) the type of eruption and (2) the stage of eruption.

The seismicity of volcanoes depends strongly on the type of eruption. Central and flank eruptions are characterized by different development of the swarms of earthquakes (Gorelchik et al., 1990; Zobin, 1988). Earthquake swarms, observed in Kamchatka, preceding central eruptions are characterized by continuously increasing numbers and magnitudes of earthquakes up to the beginning of eruption. Earthquake swarms preceding flank eruptions are characterized by the presence of a maximum number and magnitude of earthquakes at the beginning of the swarm, and a decrease of both just before the eruption.

Gresta et al. (1990) showed that the eastern and western flanks of Mount Etna are characterized by different morphological, structural, volcanic, and seismic data. They recognized the higher "degree" of tectonization on the eastern slope from the morphological and structural data. The eastern slope also has shallower and more

Table 10.2 Short Volcanological Information About the Significant Earthquakes (SE) of Table 10.1

No. (Table 10.1)	Volcano, Year	Type of Eruption	Mw of SE	Before Eruption	Initial Stage	Parox. Stage	Final Stage	*T* (years), Remarks
1	Usu, 1910	3	5.4	*1				North flank, $T = \infty$
2	Katmai, 1912	1a	7.0		*2			$T = \infty$
3	Sakurajima, 1914	2b, 3	7.0		*3			The west and east flanks, $T = \infty$
4	Paricutin, 1943	1, 2a, 4	4.9	*4				New volcano, $T = \infty$
5	Bezymianny, 1956	1	5.3			*5		Directed blast, $T = \infty$
6	Miyakejima, 1962	2a, 3	6.1				*6	The NE flank eruption with radial fissure, $T = \infty$
7	Sheveluch, 1964	1	5.2			*7		Directed blast, $T = 110$ [29]
8	Fernandina, 1968	1a	5.4			*8		Caldera collapse, $T = \infty$
9	Arenal, 1968	1, 2a	5.1	*9				The west flank and summit, $T = \infty$
10	Beerenberg, 1970	1, 2a, 4	5.7	*10				The NE and SW flanks and summit, $T = \infty$
11	Curacoa Reef, 1973	5	5.9	*11				Submarine, $T = \infty$
12	Tiatia, 1973	2b	4.6	*12				$T = \infty$
13	Etna, 1974	3	4.7	*13				West flank, $T = 211$
14	Tolbachik, 1975	1, 2a, 4	4.9	*14				New volcano, $T = \infty$
15	Nyiragongo, 1977	1, 3	5.1	*15				The simultaneous eruption of N, S, and W flanks, $T = \infty$
16	Usu, 1977	1	4.7				*16	$T = 124$
17	St. Helens, 1980	1, 2a	5.4			*17		Directed blast, $T = \infty$
18	Una Una, 1983	1	4.9		*18			$T = 45$
19	Miyakejima, 1983	2a, 3	6.0				*19	SW flank, $T = 271$
20	Beerenberg, 1985	2a, 4	5.0		*20			NE flank, $T = 12$
21	Oshima, 1986	1, 3	5.9				*21	

(*Continued*)

Table 10.2 (Continued)

No. (Table 10.1)	Volcano, Year	Type of Eruption	Mw of SE	Before Eruption	Initial Stage	Parox. Stage	Final Stage	*T* (years), Remarks
								The eruption of central crater with formation of fissure in N and NW of caldera, $T = \infty$
22	Lonquimay, 1988	2a, 3	5.2		*22			The flank crater with radial fissure, $T = \infty$
23	Teishi Knoll, 1989	5	5.2	*23				Submarine eruption, $T = \infty$
24	Pinatubo, 1991	1, 2a	5.6			*24		$T = \infty$
25	Hudson, 1991	1, 3	5.8			*24		The radial fissure, $T = \infty$
26	Etna, 1991	3	4.7		*26			SE flank, $T = 23$
27	Akademia Nauk, 1996	4[30]	7.1	*27				Subaqual caldera eruption, $T = \infty$
28	Grimsvøtn, 1996	2a, 4[28]	5.6	*28				Subglacial eruption, the fissure eruption on the N flank, $T = 58$[31]
29	Usu, 2000	2a[32]	4.9				*29	NW flank, $T = \infty$
30	Miyakejima, 2000	2a, 5[33]	6.2				*30	Submarine W flank eruption, $T = 165$
31	Miyakejima, 2000	1a[34]	6.4				*31	Caldera collapse, $T = \infty$[34]
32	Dabbahu, 2005	4[35]	5.5	*35				Rift intrusion, $T = \infty$[35]

Note: The data about eruptions are taken from Simkin and Siebert (1994), except of any special marked cases.
Type of eruption: (1) Central eruption; (1a) caldera collapse; (2a) one-side flank eruption; (2b) two-side flank eruption; (3) radial fissure; (4) regional fissure; (5) submarine eruption. *T* (in years) is the recurrence time.
References: (1) Okada (1982); (2) Abe (1992); (3) Yokoyama (1997); (4) Yokoyama and de la Cruz-Reyna (1990); (5) Gorshkov and Bogoyavlenskaya (1965); (6) Minakami (1964); (7) Tokarev (1981); (8) Filson et al. (1973); (9) Minakami et al. (1969); (10) Siggerud (1972); (11) Gibowicz et al. (1974); (12) Zobin (1976); (13) Bottari et al. (1975); (14) Zobin (1979); (15) Hamaguchi et al. (1992); (16) Wano and Okada (1980); (17) Lipman and Mullineaux (1981); (18) SEAN, 08-07 (1983); (19) Tada and Nakamura (1988); (20) Havskov and Atakan (1991); (21) Yamaoka et al. (1988); (22) Barrientos and Acevedo-Aranguiz (1992); (23) Matsumura et al. (1991); (24) Harlow et al. (1996); (25) Gonzalez-Ferran (1995); (26) Ferrucci and Patane (1993); (27) Zobin and Levina (1998); (28) BGVN, 21-09 (1996); (29) Melekestsev et al. (1991); (30) Belousova (1999, personal communication); (31) Gudmundsson (1996); (32) Zobin et al. (2005); (33) Uhira et al. (2005); (34) Geshi et al. (2002); (35) Ayele et al. (2007).

Table 10.3 Macroseismic Data for Three Volcano-Tectonic Earthquakes

Site	Event 19			Event 21			Event 23		
	I_{JMA}	I_{MM}	R (km)	I_{JMA}	I_{MM}	R (km)	I_{JMA}	I_{MM}	R (km)
Miyakejima	5	7	17	4	6	66			
Oshima	3	5	85				3	5	36
Hacijojima	3	5	102						
Tateyama	3	5	109	4	6	46			
Niijima	3	5	45	4	6	43			
Ajiro	2	4	112	3	5	44	4	6	6
Katsuura	2	4	143	2	4	102	2	4	114
Yokohama	2	4	152	3	5	85	3	5	72
Tokyo	2	4	180	3	5	113	3	5	42
Irozaki				3	5	51	3	5	42
Kofu				3	5	127	2	4	92
Mishima				2	4	61	3	5	23
Shizuoka				2	4	94	2	4	71
Chiba				2	4	125	2	4	117

Note: Numbers of events correspond to Table 10.1. I_{JMA} is the earthquake intensity of Japan Meteorological Agency scale; I_{MM} is the earthquake intensity of Modified Mercalli scale; R is the hypocentral distance. Data were taken from *BISC*.

intensive seismicity than the western slope. They attribute these differences to higher extensional stresses on the eastern slope.

Figure 10.13 demonstrates the tendency of the change of the maximum magnitude in volcanic earthquake swarms, preceding the flank eruptions with the altitude of the newborn cones, of the 4750 m height Klyuchevskoy volcano, Kamchatka, in 1953–1987. The values of M_{max} of swarms preceding the flank eruptions at lower altitudes (1360–1900 m) were generally greater than the M_{max} of swarms preceding the birth of new cones at higher altitudes (2300–3340 m). This tendency may depend on at least two factors: (1) the distance between the new flank cone and the central conduit of the volcano and (2) the mechanical quality of the rocks where earthquake rupture was formed. Both factors could constrain the length of the earthquake rupture and the size of M_{max}. The first factor, that is the decrease of the distance between the central conduit and the new eruptive site with the increase of the altitude of the site, would be decisive in the length of rupture for producing the earthquake with M_{max}. At the same time, the decreasing of mechanical properties of rocks at high altitudes could prevent large ruptures compared to those at lower altitudes. This example shows that the conditions of forming an eruptive site could be an important factor in the size of maximum seismic event related to the eruption.

Therefore, we divided our data into five groups: earthquakes related to (1) central eruption; (2) caldera collapse; (3) central eruption with flank or fissure eruption; (4) flank and (or) fissure eruption without central eruption; and (5) submarine eruptions.

Figure 10.13 Dependence of maximum magnitude of volcanic earthquakes on the eruption site altitude for Klyuchevskoy volcano. The mean altitudes of flank eruption sites were taken from Khrenov et al. (1991). The magnitudes *Ms* were calculated from the energy class *Ks* (Zobin, 1996b). The values of energy class *Ks* were taken from Tokarev (1981), Zobin (1979a), and Gorelchik (1989; personal communication, 1999). The numbers near crosses indicate the date of flank eruption.
Data from Zobin, 2001.

Figure 10.14 shows the magnitude distribution for largest volcanic earthquakes associated with these five groups of eruptions. The M_{max} was taken as the maximum magnitude of earthquake recorded in the century for each type of eruption. Due to this distribution, central eruptions are characterized by the appearance of earthquakes of M_{max} of 5.4. It is the smallest M_{max} value among all types of eruptions. The central eruptions following by the appearance of flank and (or) fissure eruptive centers are characterized by slightly higher M_{max} of earthquake (5.9).

The earthquakes with the largest M_{max} were recorded for the caldera collapses (7.0) and the flank and (or) fissure eruptions without central activity (7.1). The M_{max} for submarine eruptions was equal to 6.2.

The next set of M_{max} values was obtained for the earthquakes occurring at different stages of eruptions (Figure 10.15). We can see that the largest volcanic earthquakes occur at the preliminary or initial stages of eruption (M_{max} 7.1 and 7.0,

Figure 10.14 Distribution of magnitudes *Mw* for five types of volcanic eruptions (C, central eruption; C + C, caldera collapse; C + F, the central eruption with flank and [or] fissure eruption; F, the flank and (or) fissure eruption; and SM, the submarine eruption).

respectively). The paroxysmal stage is characterized by relatively low magnitude earthquakes (5.6), and at the final stage of eruption rather strong earthquakes can be observed (M_{max} 6.4).

Cubellis et al. (2007) estimated the maximum magnitude for the volcano-tectonic earthquakes occurring at Vesuvius volcano using the 30-year seismic observations around the volcano as well as the historical data. Comparison of instrumental seismicity and historical data reveals two significantly different energy levels: a lower earthquake level with $M_{max} = 4.5$, corresponding to current seismicity, and that which accompanied volcanic activity in the eruptive period from 1631 to 1944; an upper level with $M_{max} = 5.4$, represented by the AD 62 earthquake. Authors suggest that the two levels correspond to two stress states and different seismogenetic structures. Considering the central eruptions as characteristic for Vesuvius, these estimations of M_{max} coincide with ours (M_{max} 5.4) suggested for central eruptions.

The problem of the maximum magnitude of volcanic earthquakes was also discussed by Okada (1983). He examined the earthquakes related to "direct blast" or "sector collapse" eruptions of Unzen (1792); Bandai-san (1888); Bezymianny

Figure 10.15 Distribution of magnitudes *Mw* for four stages of eruptions (PREL, the preliminary stage; INI, the initial stage; PAROX, the paroxysmal stage; and FIN, the final stage).

(1956); Sheveluch (1964); and Mount St. Helens (1980). In addition to the data listed in Table 10.1, Okada (1983) estimated the magnitude between 5 and 5.5 for the earthquake related to the Bandai-san eruption, according to its macroseismic effect and to its record in Tokyo (epicentral distance 220 km), with amplitude of 0.2 mm on the Gray-Milne seismograph. He also considered, according to macroseismic data, the magnitude 5−5.5 for the earthquake that accompanied the 1792 explosion of Unzen. Therefore, he concluded that there tends to exist a maximum limit in magnitude at around 5, except those in which regional tectonics are required for their direct causes.

Benoit (1998) has studied the dependence of the M_{max} of volcanic earthquakes upon the type of eruption and the Volcanic Explosivity Index (VEI). He analyzed 180 swarms of earthquakes that occurred near volcanoes, mainly from 1979 to 1989. The M_{max} values ranged from 0.5 to 6.2. The relations between M_{max} and VEI did not demonstrate any significant correlation. At the same time, Benoit (1998) showed that the eccentric and radial fissure eruptions were preceded by larger shocks than central vent eruptions. This result coincides well with ours.

The reason for the significant difference between the M_{max} of earthquakes associated with the central eruptions and of earthquakes related to more complex eruptions, such as forming of regional or flank fissures, may be the difference in magma transport (Zobin, 1988). It was suggested that large earthquakes formed new fissures for magma discharge in zones of regional faults. The "renovation" of the volcano conduit before and during the central eruption does not require such a large effort. For central eruptions, we suggest that the mechanism of failure is a result of local dyke intrusion within the volcanic edifice, which can involve magmatic driving forces and fluid pressurization mechanisms (e.g., Ellsworth and Voight, 1995). This mechanism could trigger moderate-size earthquakes, whose magnitudes would be limited by the size of volcanic edifice and which could be accompanied by a failure of volcanic flanks or sector collapses.

10.4.2 Attenuation of Earthquake Intensity with Distance for Volcanic Earthquakes

The *Bulletin of ISC* published the macroseismic data for three events of magnitude interval 5.2–6.0 (Nos. 19, 21, and 23 in Table 10.1) that were related to the eruptions of Izu Islands, Japan volcanoes. They are collected in Table 10.3. These data were prepared by Japan Meteorological Survey (*JMA*) according to the *JMA* macroseismic scale and then were transformed into more popular MM macroseismic scale using the comparative table published by Tiedemann (1992). The intensity–distance equation for 31 points shown in Figure 10.16 was calculated in the following form:

$$I = aMs - b\,log\,R - cR + d \tag{10.2}$$

where $R = (\Delta^2 + h^2)^{1/2}$ is the hypocentral distance in km, Δ is the epicentral distance in km, h is the focal depth in km, the coefficients b and c are the coefficients of geometric spreading of seismic waves and absorption, respectively, and d is a free member. This form of the intensity–distance equation was recommended for macroseismic studies by the *New IASPEI Manual of Seismological Observatory Practice* (Bormann and Bergman, 2000) and allows the expression of the intensity attenuation as a function of magnitude and distance, as well as the estimation of the inelastic properties of rocks in the region.

The following law for earthquake intensity MM attenuation with the epicentral distances >5 km was obtained:

$$I = 0.66Mw - 1.13\,log\,R - 0.0072R + 3.73 \tag{10.3}$$

The curve, calculated according to Eq. (10.3) for the *Mw* 5.5 earthquake, which occurred at focal depth of 0 km just below the volcanic crater, is shown in Figure 10.16 and can predict a possible macroseismic effect.

At the same time, the epicenter of a significant volcanic earthquake may not coincide with the volcanic crater position. Zobin (1996) showed that the position of

Figure 10.16 Decrease of earthquake intensity of volcano-tectonic earthquakes with distance. The curve calculated from Eq. (10.3) for the earthquake with magnitude *Mw* 5.5, focal depth 0 km, is shown. The data are marked by different symbols for three earthquakes, Nos. 19, 21, and 23 from Table 10.1. Intensity data were taken from Bulletin of *ISC*. Data from Zobin, 2001.

the largest earthquake depends strongly on the crustal thickness H beneath the volcano. Figure 10.17 illustrates the situation for continental and oceanic volcanoes. The epicentral area of volcanic earthquakes is the surface projection of the faulted volume around the magma conduit. For any given amount of erupted material, there would be a narrow vertical faulted volume beneath the continental volcano and more widespread and more horizontal faulted volume for the oceanic volcano. Therefore, for using Eq. (10.3) it is necessary to know the possible deviations of the earthquake epicenter from the volcanic crater. For this purpose, the best estimated locations of 11 significant volcanic earthquakes from Table 10.1 were used (Figure 10.18). The following equation was obtained (Zobin, 2001):

$$\log D(\text{km}) = 3.67 - 2.14 \log H(\text{km}) \tag{10.4}$$

Eq. (10.4) allows correct calculation of possible seismic intensity from Eq. (10.3) by moving the position of the initial point from the crater to distance D, which is characteristic of the volcano under study. For thick oceanic crust, we have larger distance between the crater and earthquake epicenter than for thin continental crust.

The law of decreasing of earthquake intensity with distance for significant volcanic earthquakes may serve for the prediction of a possible seismic hazard of eruptive activity. This law was derived based on data from rather deep-seated foci of volcanic earthquakes (focal depth 3–15 km) and therefore it could predict

Figure 10.17 Illustration of the difference in the position of the seismic volumes beneath the volcanoes situated above the continental (left) and oceanic (right) Earth's crust. (1) "crust–mantle" boundary; (2) active seismic volume beneath the volcano; (3) occurrence of the largest volcanic earthquakes.
Data from Zobin, 1996.

maximum intensity. More shallow seismic sources situated within the volcanic edifice would produce earthquakes of higher attenuation and their intensity would decrease more rapidly.

10.4.3 Recurrence Time

The author considers the recurrence time T as the recurrence time between similar types of eruption of the same volcano that occurred at the same location. The data for the estimation of T were taken mainly from the catalog of Simkin and Siebert (1994). As was shown earlier, the volcano-tectonic activity could be different for different slopes of a volcano (Gresta et al., 1990). Therefore, for the 1914 eruption that occurred on the western and eastern flanks of Sakurajima (event No. 3), it was taken at a time $T = \infty$ but not $T = 135$ years, counting from the 1779 eruption that occurred on the northern and southern slopes of the volcano.

For the 1962 Miyakejima eruption (event No. 6) that occurred on the northeastern flank with the formation of a radial fissure, it was taken $T = \infty$ because the previous fissure-flank eruptions had all occurred on the other flanks (the 1835 eruption occurred on the western flank and the 1643 eruption occurred on the southwestern flank). At the same time, the 1983 eruption of Miyakejima has a recurrence time $T = 271$ years because the similar 1712 eruption occurred on the same southwestern flank. The 2000 submarine flank eruption (event No. 30) occurred on the western flank and has a recurrence time $T = 165$. The information about the magnitude of earthquakes associated with the ancient eruptions is

Figure 10.18 Dependence of distance between the crater and the largest earthquake epicenter on the crust thickness in the volcanic region. The line of regression is shown. The coefficient of correlation is −0.746. The numbers near the black diamonds correspond to the numeration of eruptions in Table 10.1. The position of epicenters is shown in Fig. 10.2. The data about the Earth's crust thickness for events Nos. 13, 14, 17, 19, 20, 21, 23, 26, and 27 are collected in (Zobin, 1996); for the remaining events, they are taken from Mooney et al. (1998) (No. 24) and Staples et al. (1997) (No. 28).

absent; therefore, the seismic aspect of the eruption was not taken into account for determination of T.

The majority of information about volcanoes is collected for historical time, which is different for different parts of the world. Nevertheless, it is possible to hope that for the volcanoes of study, listed in Table 10.2, we have consistent information during the historical time for the last 100−150 years. For our 32 eruptions with significant seismic events, the value of $T = \infty$ (the first event for historical time) is available for 23 events out of 32. What does it mean practically? Three volcanoes (No. 4, Paricutin; No. 14, Tolbachik; and No. 32, Dabbahu) were newborn volcanoes. Two submarine eruptions (No. 11, Curacoa Reef and No. 23, Teisi Knoll) were also born as new volcanoes. Several volcanoes (No. 2, Katmai; No. 5, Bezymianny; No. 9, Mount St. Helens; No. 24, Pinatubo; and No. 27, Akademia Nauk) were considered to be dormant volcanoes. For 12 volcanoes, the $T = \infty$

means it was the first recorded type of such activity for historical time. For example, for Usu volcano, the flank eruption of 1910 (No. 1) was the first mentioned flank eruption after a series of central eruptions beginning from 1663. For Oshima volcano, the eruption of 1986 (No. 21) was the first historical eruption with the formation of a fissure on the north flank after a series of central eruptions of 1783–1974 and previous central and flank eruptions.

The remaining nine eruptions, with the T values less than ∞, have these values in the range between 110 and 271 years (five events) and between 12 and 58 years (four events).

The association between the large recurrence periods for eruptions of the same type at the volcano with significant earthquakes is a very important factor for constraining seismic hazard. Our results show that quasi-dormant volcanoes situated within active volcano-tectonic structures and volcanoes with long-term central crater activity represent higher seismic hazards. They may need to form new passageways for magma ascent. At the same time, the practical application of the recurrence time is problematic. We have the real estimates of recurrence time (not ∞, or less than 200 years) only for six volcanoes (Sheveluch, Usu, Una Una, Beerenberg, Etna, and Grimsvøtn).

10.4.4 Estimation of the Seismic Hazard of Volcanic Activity of Volcán de Colima, Mexico

This volcano is the most active in Mexico. The history of its volcanic activity (Simkin and Siebert, 1994) shows that it is characterized by central or central+flank eruptions.

The last central + flank eruption occurred in 1869–1875, or 135 years ago. We have no information about any significant earthquake related with the eruption and cannot estimate the recurrence time of the event. The M_{max} for a possible future significant volcanic earthquake could be taken as $Mw = 5.9$ for the central + flank eruptions.

The second type of eruption, the central one, was observed here in 1998–1999 without any significant earthquake. M_{max} for possible significant earthquakes associated with central eruptions of Colima could be taken as $Mw = 5.4$.

The thickness of the Earth's crust in the Colima volcanic region is about 36 km (Pacheco et al., 1997). Therefore, the position of the seismic event from the central crater, according to Eq. (10.4), would be about 2 km, or insignificantly far from the crater. We can propose a focal depth of 5 km.

In the case of central + flank eruption, the M_{max} seismic event could produce, according to the macroseismic equation (10.3), a V–VI MM intensity (V–VI means that the intensity MM would be 5.5 grades) in Colima city, situated at 32 km from the volcano, VI MM intensity in the village San Marcos, situated at 14 km from the volcano, and VI–VII MM intensity in the village of La Yerbabuena, situated at 8 km from the volcano.

In the case of a central eruption, the intensities MM would be V–VI in Colima city, VI in San Marcos, and in La Yerbabuena.

11 Origin of Eruption Earthquakes

Eruption earthquakes are related to the movement of magma within volcanic conduits and to subsurface or surface manifestations of volcanic activity. They may be produced by magmatic, phreatic, and phreato-magmatic processes within the conduit or the crater of a volcano as well as by volcanic explosions and the propagation of density currents along the slopes of the edifice (lava flows, pyroclastic flows, and lahars). Therefore, eruption earthquakes reflect the variety of volcanic processes and understanding them is crucial for effective eruption monitoring and the quantification of the associated phenomena.

11.1 Volcanic Processes Generating Seismic Signals of Eruption Earthquakes

During its ascent within the volcanic conduit, magma undergoes exsolution and fragmentation, thus, changing from a continuous melt with a dispersed gas phase to a two-phase mixture of melt fragments within a continuous gas phase (Cashman et al., 2000). This process may lead to effusive or (and) explosive eruptions depending on bubble number densities and the magma flow rate (Gonnerman and Manga, 2007). All these processes represent potential sources of seismic signals.

11.1.1 Processes within the Volcanic Conduit

The transport of magma from deep reservoir to the surface within the conduit can produce different types of seismic signals. Networks of cracks can be formed within the magma body. The cracks, filled by fluid phases, can be excited into resonance by a pressure transient (Chouet, 2003) and generate long-period seismic signals. The upward movement of the fluid–particle mixture above the fragmentation surface can trigger vibrations in the conduit structures (Ripepe et al., 2001). Explosions near or within the crater are possible sources of seismic signals, as well. The fracturing of lava dome rock can generate seismic signals also (Tuffen et al., 2008; see Chapter 3).

11.1.2 Volcanic Flows

The subaerial massive sediment motions at volcanoes are divided into two groups: the phenomena in which the particles are dispersed in the flowing body and the phenomena in which the moving bodies are mostly the agglomerates of soil and

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00011-6
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rocks (Takahashi, 2007). The pyroclastic flows are included in the first group; lahars belong to the second group.

When magma emerges from the conduit, it may form lava domes and (or) lava flows. The type of lava construction depends on the nature of the feeding magma. Basaltic volcanoes form rather fluid low-viscosity lava flows. Andesitic and dacitic volcanoes form lava domes within the crater and (or) high-viscosity lava flows. The gravitational collapse (Merapi type) or the explosive disruption of a growing lava dome (Peléean type) and the collapse of an eruption column (Soufriére type) can generate pyroclastic flows and rockfalls (Francis, 1998). Pyroclastic flows are distinguished from rockfalls by their larger size, longer runout (≥0.5 km), greater production of fine ash, and appreciable buoyant hot ash clouds (Calder et al., 2002).

The model of Takahashi and Tsujimoto (2000) illustrates that pyroclastic flow propagation can be represented as a two-stage process: the collapse of the lava dome (or the rocks from the eruptive column) induces a granular flow, which later changes to a fluidized flow (Figure 11.1A). As was shown by Yamasato (1997) and

Figure 11.1 Mechanical models of pyroclastic flows (A) and lahars (B). Details are discussed in text.
After Takahashi and Tsujimoto, 2000, and Takahashi and Satofuka, 2002.

Zobin et al. (2009b), the main part of the seismic signal is associated with the fluidized flow stage of pyroclastic flows.

Lahars along the slopes of volcanoes are usually triggered by a heavy rain, especially after the intensive fall of fresh ash. At Merapi volcano, for example, about 20 mm h^{-1} is supposed to be critical for lahar triggering if it falls on unconsolidated materials such as ash, sand, or gravel (Lavigne et al., 2000a). In some instances, lahars may have been triggered by steam explosions that generated strong ground vibrations, which induced failure of partially water-saturated pyroclastic materials (Rodolfo et al., 1996). The typical 1991 posteruption lahars at Pinatubo were estimated to be 2–3 m deep, 20–50 m wide, and moving at 4–8 m s^{-1} (Pierson et al., 1996). The frontal mean velocity of lahars at Merapi ranges between 2.5 and 5.9 m s^{-1}; the peak flow velocity may reach 6–15 m s^{-1} (Lavigne et al., 2000a). Small- to moderate-sized lahars at Pinatubo remobilized between 2×10^5 and 2×10^6 m^3 of sediment; larger events lasting up to 4 h have transported about 10 times this amount (Rodolfo et al., 1996).

Takahashi and Satofuka (2002) describe lahars as hybrid-type debris flows between turbulent-muddy and stony debris flows. The hybrid flow in the inertial range consists of two layers: the upper layer of depth h_1 is the particle suspension layer, and the lower layer of depth h_2 is where particles move with frequent collision or enduring contact with each other (Figure 11.1B). The dominance of stony or turbulent-muddy type of flow depends on the relative thickness of these two layers h_1/h_2. The movement of the lower stony layer along the volcanic slope ravine (angle of inclination θ in Figure 11.1B) produces the main part of the seismic signal of lahar.

11.2 Source Mechanisms of Eruption Earthquakes

Source mechanisms of eruption earthquakes may be represented by a combination of single forces and the seismic moment tensor described in Section 3.3.

11.2.1 A Force System Equivalent to a Volcanic Eruption

The model (Kanamori and Given, 1982; 1983; Kanamori et al., 1984) is based on the assumption that seismic signals are produced by a volcanic eruption as Lamb pulses, excited by a nearly vertical single force that represents the counterforce of the eruption. Lamb (1904) computed the transient response of a homogeneous elastic half space to a single force with various geometries. He showed that a pronounced pulse propagates along the free surface with the Raleigh-wave velocity. This pulse is preceded by minor signals, which propagate with *P*- and *S*-wave velocities (Figure 11.2).

The comparison of observed seismograms for small eruptions at Mount St. Helens with synthetics constructed as Lamb's pulses excited by a vertical force (Figure 11.3) shows a good agreement for the Rayleigh-wave signal. Forward

Figure 11.2 Lamb pulses computed for a downward vertical single force applied at the free surface of a homogeneous half-space at a distance of 67 km. Here α and β are *P*- and *S*-wave velocities, I_3 and I_4 correspond to vertical and radial components of displacement, respectively. I_3 and I_4 are convolved with the time derivative of a source time function of the form $s(t) = p/(p^2 + t^2)$ as proposed by Lamb (1904). Corresponding traces given by Lamb (1904) are shown below for comparison. I_3 and I_4 correspond to w_0 and q_0. From Kanamori and Given, 1983.

Figure 11.3 Comparison of the synthetic for the vertical force with the observed long-period record of the 13 June 1980 Mount St. Helens explosion at a distance of 67 km from the volcano. The synthetics are computed using a distance of 67 km, $\mu = 1.9 \times 10^{11}$ dyn cm^{-2}, and f_m (peak value of the force) = 5.6×10^{15} dyn. From Kanamori and Given, 1983.

modeling allows estimating the counterforce of the eruption by comparing the amplitude of the synthetic with the recorded signal. Modeling of more complex signals can be performed assuming that a volcanic eruption may be considered as a combination of a single force, representing the thrust of the eruption, and an

isotropic source corresponding to a sudden pressure release in the magma chamber (Figure 11.4). For a simple model of eruption, the amplitude ratio of seismic waves excited by a single force to those excited by an isotropic source is approximately equal to the ratio of the seismic wave velocity to the particle velocity of the fluid or gas in the magma chamber. Because this ratio is about 10, a single force can adequately model the source (Kanamori and Given, 1983).

Modeling of the complex seismic signal produced by the great 18 May 1980 Mount St. Helens eruption is shown in Figure 11.5. The eruption consisted of separate events that contributed to shape the overall seismic signal. A chronology of the eruptive sequence is shown in Figure 11.5B. The sequence began with a magnitude $Ms = 5.2$ earthquake. Within 7–20 s of the earthquake, a 2.3-km^3 landslide began (event 1). During the landslide, an explosion (the lateral blast) occurred (event 2). A second explosion occurred just after the termination of the landslide. Two minutes after the first earthquake, a second earthquake occurred. The sequence lasted about 3 min.

The time history of the single force inferred from the long-period waves (Figure 11.5D) characterizes the entire eruption sequence as a single event equivalent to an earthquake of magnitude $Mw = 6.2$. The value of this force reached its maximum when the landslide had a maximum velocity. The source time function

Figure 11.4 Force system equivalent to a volcanic eruption. (A) A pressure release model for a volcanic eruption. (B) Forces acting on the top, side, and bottom walls. (C) Decomposition of the force into a vertical single force and an implosive force. From Kanamori et al., 1984.

Figure 11.5 Schematic diagrams showing the chronology of events associated with the Mount St. Helens eruption of 18 May 1980. Time is taken positive downward. The time history of the single force determined from surface waves is shown in the right panel, and the magnitude and the time history estimated from the reported sequence are shown in the left panel.
From Kanamori and Given, 1982.

inferred from the *P*-wave deconvolved record (Figure 11.5C) allows interpreting the sequence of events. There are two separate events of the same magnitude, 5.3, that might be associated with the first earthquake + first explosion and the second earthquake + second explosion (Figure 11.5B).

Figure 11.5A shows the comparative modeling of time history of the eruption with the magnitude of the forces caused by the landslide (1) and the lateral blast (2). According to this reconstruction, the most energetic process was the landslide; the resulting force peaked at about 10^{18} dyn. The lateral blast was less intense; its resulting force reached a maximum of 3.4×10^{17} dyn.

11.2.2 Seismic Moment Tensors of Some Non-Double-Couple Sources of Eruption Earthquakes

Seismic moment tensors describing volumetric mechanisms are thought to represent magma-filled sources of eruption earthquakes. While the injection of a volcanic fluid or a thermal expansion of a fluid-filled crack occurs, a sudden volume

Figure 11.6 System of coordinates and geometry for a crack. Here ξ_1, ξ_2, and ξ_3 correspond to the east, north, and vertical directions, respectively; λ and μ are the elastic parameters of the host rocks. From Kumagai, 2009.

increase ΔV is observed and the fracture acts as a radiating seismic source. The seismic moment tensors of these sources can be represented in a Cartesian system of coordinates (Figure 11.6).

Tensile crack source (Kumagai, 2009). Source coordinates and geometry for the tensile crack are shown in Figure 11.6. Here $\boldsymbol{v}$ is the direction of opening−closing described by the angles θ and φ, where θ is the angle measured from the up direction ξ_3 and φ is the angle of horizontal projection measured from direction ξ_1. The moment tensor for a vertical tensile crack $(\theta = \pi/2)$ undergoing a volume change ΔV is:

$$\vec{M} = \mu\Delta V \begin{pmatrix} \lambda/\mu + 2\cos^2\phi & 2\sin\phi\cos\phi & 0 \\ 2\sin\phi\cos\phi & \lambda/\mu + 2\sin^2\phi & 0 \\ 0 & 0 & \lambda/\mu \end{pmatrix} \tag{11.1}$$

and for a horizontal crack is

$$\vec{M} = \mu\Delta V \begin{pmatrix} \lambda/\mu & 0 & 0 \\ 0 & \lambda/\mu & 0 \\ 0 & 0 & \lambda/\mu + 2 \end{pmatrix} \tag{11.2}$$

If $\lambda = \mu$ and $\phi = 0$, the amplitude ratios of the diagonal terms are 3:1:1 for a vertical crack, and 1:1:3 for horizontal crack.

Cylindrical source (Kumagai, 2009). The surface of a cylinder can be regarded as a cylindrical crack, where radial expansion, $D_c = (d_c + \Delta_c)$, occurs (Figure 11.7). The moment tensor for the radial expansion of a vertical cylinder of length L and radius R is

$$\vec{M} = \Delta V \begin{pmatrix} \lambda + \mu & 0 & 0 \\ 0 & \lambda + \mu & 0 \\ 0 & 0 & \lambda \end{pmatrix} \tag{11.3}$$

Figure 11.7 Vertical cylinder of length L and radius R. The cylinder surface can be regarded as a cylindrical crack, where radial expansion $D_c = (d_c + \Delta_c)$ occurs. Here, the inner wall of the crack moves inward by d_c and the outer wall moves outward by Δc.
From Kumagai, 2009.

Figure 11.8 Spherical crack surface of radius R where a constant radial expansion $D_s = (d_s + \Delta_s)$ occurs. Here, the inner wall of the crack moves inward by d_s and the outer wall moves outward by Δ_s.
From Kumagai, 2009.

where ΔV is the volume change given as

$$\Delta V = 2\pi RLD_c = 2\pi RL\,(d_c + \Delta_c) \tag{11.4}$$

If $\lambda = \mu$, the amplitude ratios of the diagonal terms are 2:2:1.

Spherical source (Kumagai, 2009). The moment tensor for a spherical source (Figure 11.8) is given as

$$M = (\lambda + 2\mu)\Delta V_s \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix} \tag{11.5}$$

where $\Delta V_s = 4\pi R^2 \Delta_s$ is the volume change caused by the displacement Δ_s of a spherical surface of radius R. The diagonal terms are 1:1:1.

11.3 Models of the Eruption Earthquake Sources

11.3.1 Models Based on the Vibration of Magma-Filled Structures

The models have been developed in two main directions: (1) those based on the resonance of fluid-filled cavities with various geometries (sphere, pipe, or crack);

and (2) those based on the oscillation of a number of vibrating conduits. A seismic signal is generated by pressure fluctuations in the magma system. The choice of the model is mostly subjective and depends on the knowledge of specific characteristics of the volcano, as well as on the experience of scientist.

The Fluid-Driven Crack Model

The model (Chouet, 1986; 1996a) consists of a rectangular fluid-filled crack contained within an infinite, linear elastic medium (Figure 11.9). The thickness of the crack is much smaller than the seismic wavelengths of interest. The fluid-driven crack is a composite of a trigger element (the energy source) and a resonator (the fluid-filled crack). A pressure transient is applied over a small area of the crack wall triggering the acoustic resonance of the crack. Repetitive triggering of resonance in the crack by fluid flow produces continuous tremor.

The resonance characteristics of the fluid-filled crack depend on the crack geometry, the physical properties of the fluid and solid, the impedance contrast between fluid and solid, the spatiotemporal characteristics of the pressure fluctuations driving the crack, and the boundary conditions in effect at the crack perimeter. The impedance contrast affects both the frequency and duration of the signal radiated by the crack.

The ground response to a buried source may be synthesized after calculating the source space-time function. Variations in the model characteristics enable the optimization of the synthetics modeling. Figure 11.10 shows the synthetics calculated for different crack excitations and crack depths. The synthetics shape strongly depends on the model depth (compare the synthetics calculated for the depths of 0 and 105 m). Figure 11.10 demonstrates that the single-excitation synthetics give a long-period event waveform, while the synthetics obtained by delaying and superposing multiple copies of the same signal give the tremor record. Figure 11.11 illustrates a good fit of the synthetics to the observed record of a long-period event recorded at Galeras volcano, Colombia.

Figure 11.9 Geometry of the fluid-filled crack model. See the text for details.
From Chouet, 2003.

Figure 11.10 Dependence of synthetics on crack depth and crack excitation. All synthetics are for a 100-m-wide, 200-m-long, vertical crack excited by a pressure transient applied at the center of the crack. Each synthetic represents a vertical component of ground velocity calculated at a distance of 1 km and azimuth of 25°. A, the synthetics of long-period (LP) event and volcanic tremor (T) calculated for the crack depths of 0 and 105 m. B, the Fourier spectra of the same records.
From Chouet, 1996b.

The Conduit-Vibration Model

This model (Seidl et al., 1981; Schick et al., 1982b; Gresta et al., 1991) may be applied to volcanic tremor. It is hypothesized that transient flows within volcanic conduits can generate tremor with certain dominant frequencies, depending on the conduit geometry. Vapor and gas escaping through open conduits are considered responsible for turbulent motions in the magma; consequently the magma-filled pipes in the upper structures of the volcano behave as oscillators. Figure 11.12 shows a model of conduit oscillators for Etna volcano, Sicily. It consists of a set of vibrators with characteristic frequencies from 0.9 to 4 Hz. These characteristic

Figure 11.11 Comparison between the vertical ground velocity recorded at a distance of 1 km from the crater (A) and the synthetics calculated at the same location for the excitation of a vertical fluid-filled crack buried at a depth of 1 km beneath the vent of a volcano (B) for the long-period event that preceded the 14 January 1993 eruption of Galeras volcano, Colombia. The normalized spectra for data (thin line) and synthetics (dotted line) are shown in (C).
From Chouet, 1996b.

Figure 11.12 Schematic model of main magma conduits at Etna. Bocca Nuova and Chasm are active craters of Etna.
From Schick et al., 1982a.

Figure 11.13 Histogram of the first three dominant peaks of tremor recorded from January 1973 to January 1985 at Etna. From Cosentino et al., 1989.

frequencies are comparable with the peak frequencies that were observed for volcanic tremor at Etna (Figure 11.13).

11.3.2 Models Based on Degassing Process of Magma

The following two models consider the deep trigger of an explosion associated with the vertical magma ascent from the conduit bottom to the surface.

Vertical Expansion Source Model

This model (Iguchi, 1994) was developed for volcanic earthquakes of the andesitic Sakurajima volcano in Japan and may be applied to the modeling of low-frequency and explosion earthquakes. This model may be associated with the expansion of a gas pocket in the conduit. Figure 11.14 shows a scheme of the state of the volcanic conduit before an explosion as proposed by Iguchi (1994). The conduit is filled with a mixture of viscous magma and gases, and the top of the conduit is closed by lava dome. The release of high pressure accumulated in a large gas pocket may produce an explosion that generates an explosion earthquake.

The evidence for the vertical expansion model was obtained from the analysis of seismic signals of explosion earthquakes. The polarities of first motions for these signals indicate compressional arrivals for all directions consistent with an isotropically expanding (Figure 11.15). Application of polarization analysis shows that significant *SV*-waves arrive at the expected arrival time of *S*-waves, and *SH*-waves components appear a few seconds later (Figure 11.16). This indicates the dominance of *SV*-waves. The moment acceleration tensor calculated for the signal shows that the vertical vector dipole corresponding to the tensor *zz*-component is dominant (Figure 11.17). All this evidence supports the vertical expansion of a gas pocket as a source mechanism for explosion earthquakes.

Figure 11.14 State of the magma conduit prior to the occurrence of an explosion earthquake at Sakurajima volcano. Arrows indicate that gas pockets expand vertically along the conduit. Hexagons and ellipses in the conduit represent solidified lavas and gas pockets, respectively. From Iguchi, 1994.

Figure 11.15 Polarity distribution of the explosion earthquake of 15 April 1988 at Sakurajima volcano, Japan. Filled circles show compressional arrivals.
From Iguchi, 1994.

Figure 11.16 Seismograms of explosion earthquakes of 4 June 1988 at Sakurajima volcano, Japan, by applying a polarization filter. The traces Lon., Per., and Tran. correspond to the motions of *P*-wave, *SV*-wave, and *SH*-wave, respectively. The symbols S_{est} and S_x indicate arrival times of *S*-waves estimated from focal distance and from the onset time of the *SV*-wave component, respectively.
From Iguchi, 1994.

Figure 11.17 Components of moment acceleration tensor (in units of 10^{14} N-m s^{-2}) calculated for the explosion earthquake of 15 April 1988 at Sakurajima volcano, Japan. (A) the values of six tensor components; (B) the average of all components normalized by the vertical vector dipole. Bars denote errors. The dominance of vertical *zz*-component is obvious.
From Iguchi, 1994.

Figure 11.18 Comparison of the broadband seismic record (EZ5 station) of an explosion at Volcán de Colima (21 February 2007) with the video images of the same event. 1–4 corresponds to the timing of images and seismic record; t_1, t_2, and t_3 are the times of the onset and the end of pre-explosion (P_E) pulse, and the end of co-explosion (C_E) pulse, respectively; the arrow shows the start of ejection of explosive material from the crater. The digital infrared video camera was installed at a distance of 15 km from the crater and records a frame per second. The timing at seismic station and for the video camera was synchronized by the world atomic time survey.
From Zobin et al., 2009b.

Two-Stage Model of Volcanic Explosion

The contemporary records of broadband seismic signal and video images of a Vulcanian explosion at Volcán de Colima (Zobin et al., 2009b) showed that the first superficial manifestation of the explosion (outlet of magma material from the crater) appears only about 15 s after the onset of the seismic record, on the code of the second pulse (Figure 11.18). This indicates the ascending process occurring within the volcano conduit before an explosion. The model proposed by Zobin et al. (2006a, b; 2009a) describes the volcanic explosion process as consisting of the ascent of fragmented magma to the surface, triggering an explosion (stage 1), and a subsequent explosion (stage 2) (Figure 11.19A). Based on this model, the seismic record of an explosion (Figure 11.19B) may be

Figure 11.19 Illustration of the two-stage model of a Vulcanian explosion (A) and its broadband seismic record (B). Photo of the Volcán de Colima eruption column (December 2004) was taken by C. Navarro.
From Zobin et al., 2009b.

Figure 11.20 Illustration of the processing of seismic signals. The broadband unfiltered seismic signals (vertical component of velocity) corrected for instrument response are shown. The time indices t_1, t_2, and t_3 are the same as in Figure 11.18. The modeling of the pre-explosion signals for the seismic records of the Popocatépetl explosions is shown in (A). The dashed line shows the synthetic seismogram; the solid line is the observed seismogram. The seismic record of the Colima explosion and the Fourier spectrum of the co-explosion impulse are shown in (B). The length of the Fourier vector taken from the minimum to maximum frequencies of the spectrum frequency is shown by the double hollow arrows.

considered as consisting of two phases, pre-explosion (between t_1 and t_2) and co-explosion (between t_2 and t_3).

This model may be applied for quantification of the sources of Strombolian and Vulcanian explosions. The initial seismic pulses, generated by Strombolian explosion at basaltic volcano, are supposed to be produced by rapid expansion of gas in the magma conduit (Ripepe et al., 2001). For more viscous magmas, typical for Vulcanian explosions at andesitic and dacitic volcanoes, the initial seismic pulse is supposed to be produced by gas filtration through a system of interconnected bubbles (Melnik and Sparks, 1999). The co-explosion pulses are supposed to be produced by an explosion for both Strombolian and Vulcanian activity.

The four source parameters of an explosion may be inferred from broadband seismic records: the force, F, acting in the conduit before an explosion and governing the ascent of fragmented magma to the surface, the time, $D_1 = t_2 - t_1$, corresponding to the migration of fragmented magma in the conduit before an

Figure 11.21 Illustrations of the dome collapse sequence at Unzen volcano, Japan (left) and the corresponding vertical components of seismic signals (right). A sketch of the dome from the seismic station, situated about 3 km to south of the dome, is shown based on a video record taken by a portable 8 mm video recorder. SP indicates the short-period records; LP, the long-period records. The records are arranged according to the horizontal distance from the source. E3, C, O, K6, and K8 are seismic stations.
From Yamasato, 1997.

explosion, the duration of explosion in the conduit, $D = t_3 - t_2$, and the energy, E, of the explosion. The power of an explosion P may be obtained as E/D_2.

The Rayleigh nature of seismic waves representing the pre-explosion seismic pulses shown by (Zobin et al., 2006) allows the use of the methodology of modeling the first pulse as a Lamb's pulse with the estimation of the force F (Kanamori and Given, 1983; Nishimura, 1995) (Figure 11.20A). The energy of explosion E may be calculated from the spectrum of co-explosion pulse (Figure 11.20B). The theorem of Parseval (Weaver, 1983) states that it is possible to calculate the total amount of energy E_i (for the unit of mass) in a continuous signal v (velocity) from the Fourier spectrum of the signal:

$$E_i = \int_0^{2\pi} v^2(t)\,\mathrm{d}t \tag{11.6}$$

According to this theorem, the energy E_i (in joules) of the co-explosion pulse may be calculated as the integral (11.6) where v is the length of the Fourier vector taken from the minimum to maximum frequencies of the spectrum (Figure 11.20B). To estimate the total energy of the explosion, the energy of the co-explosion impulse, E_i, is multiplied by the coefficient of the seismic portion of the total energy of the

explosion (the ratio between seismic and total energy) p and the coefficient of effective attenuation of seismic energy with distance including the effects of geometrical spreading, energy absorption, focal mechanism, and station conditions, k. The coefficients p and k were approximately estimated for the Colima explosions recorded by the broadband seismic station at a distance of 4 km from the crater: the coefficient of the seismic portion of the total energy of an explosion p was estimated equal to 10^{-5} and the coefficient of effective attenuation of seismic energy with distance k was equal to 10^{-13}. As a result, the following equation for estimation of the energy of explosions at Volcán de Colima at the distance of 4 km from the crater was proposed:

$$E = 10^{18} \times E_i \tag{11.7}$$

Equation (11.7) provides the values of E with a precision of about half an order of magnitude. At the same time, the estimations of E for a sequence of explosions at the same volcano can give a resolution of about ± 0.1 of log unit.

Figure 11.22 Schematic source time functions of the vertical and horizontal components of a single force exciting pyroclastic flow according to the dome collapse sequence shown in Figure 11.21.
From Uhira et al., 1994.

11.3.3 Modeling of Seismic Signals Generated by Pyroclastic Flows and Rockfalls

Pyroclastic flows (or pyroclastic density currents) are the inflated mixtures of hot volcanic particles and gas that flow in variable concentrations and varying velocity along the ground (Schmincke, 2004).

Three-Stages Dome Collapse Model

Modeling of the seismic signals produced by pyroclastic flows followed the three elements noted by Takahashi (2007): fall, flow, and slip (Yamasato, 1997; Uhira et al., 1994). The first element is variable: it may be the fall of the massive sediments caused by the collapse of a lava dome segment or the fall of the rocks from a volcanic eruption column. The flow and slip elements would be similar.

The observations of the generation of pyroclastic flows at Unzen volcano, Japan (Yamasato, 1997; Uhira et al., 1994) illustrated the three stages in the formation of a pyroclastic flow with a volume equal to 5.7×10^4 m^3, recorded by video and seismic instruments (Figure 11.21): the lava dome collapse (stage A), the downfall of lava blocks onto the slope (stage B), and the generation of pyroclastic flow (stage C). This process may be described by a function representing the single-force system that is shown in Figure 11.22 (Uhira et al., 1994). The use of the shape of this function for construction of a seismic source time function allows the calculation of synthetic seismograms that reproduce the observed seismic signals associated with the propagation of the pyroclastic flows (Figure 11.23).

Figure 11.23 Comparison of observed (OBS) and synthetic (SYN) waveforms of a seismic signal produced by the 22 June 1991 pyroclastic flow at Unzen volcano, Japan. The inversion was performed with a source time function of the type shown in Figure 11.22. The single force components F_x, F_y, and F_z used for the inversion are shown below. The inversion was realized for the short-period record of station B and the long-period record of station O.
From Uhira et al., 1994.

Figure 11.24 Examples of three types of rockfall signals recorded by a broadband seismic station at a distance of about 5 km from the dome of Soufriére Hills volcano, Montserrat. In every case, the top trace is the raw ground velocity on the vertical component and the bottom trace the same signal after low-pass filtering at 2 Hz to show the long-period component. (A) A rockfall signal with little energy below 2 Hz; (B) A rockfall that contains energy below 2 Hz; and (C) A long-period rockfall signal with energy below 2 Hz present as a precursory wavelet, in addition to throughout the rest of the signal.
From Luckett et al., 2002.

An inversion gives the magnitudes of the vertical component of the single force corresponding to the collision of fragments with the slope and the horizontal component representing the deceleration of the flowing fragments along the slope.

Gas-Venting + Rock-Moving Model

This model (Luckett et al., 2002) is based on the broadband seismic observations at Soufriére Hills volcano, Montserrat. It consists of two subsources that produce the seismic signal during the propagation of pyroclastic flow. The first is the resonance of a fluid-filled conduit, which contributes rather long periods (frequencies between 1 and 2 Hz). The long-period resonance is associated with gas escaping into the atmosphere. The second is a combination of forces resulting from rocks moving on the surface of the dome, which contributes the higher frequency signal. Figure 11.24 illustrates the presence of long-period (1−2 Hz) component in seismic signals.

12 Volcanic Tremor

Volcanic tremor is the very popular seismic signal recorded at volcanoes. McNutt (1992) wrote about 137 volcanic areas where tremors were recorded; he wrote that over 600 (about 1000 to date) scientific publications have described various features of volcanic tremor. Here, the general characteristics of volcanic tremor and its relationship with volcanic processes are given.

12.1 Seismograms and Spectra

Volcanic tremor represents a continuous record of monotonic (harmonic) or nonharmonic vibrations that may continue for minutes to months (Figure 12.1). The spectral characteristics of the monotonic and nonmonotonic tremors, obtained by Kawakatsu et al. (1992) at Sakurajima volcano, Japan, with a broadband seismometer during the 1991 explosive activity of the volcano at a distance of 4.4 km from the active crater Minami-dake, are shown in Figure 12.2. It is seen that the short-period spectrum for monotonic tremor is characterized by peaks within a narrow frequency range at about 2.5 Hz and its long-period spectrum is characterized by the absence of any predominant peaks. For nonmonotonic tremor, the opposite situation is observed: no unique peaks for the short-period spectrum but a narrow frequency range of peaks at about 0.2 Hz for the long-period spectrum.

Volcanic tremor may be represented by surface waves or body waves (Figures 12.3 and 12.4).

12.2 Location of Volcanic Tremor

The sources of volcanic tremor are located at shallow depths beneath the crater floor.

12.2.1 Oshima Volcano, Izu Islands

Yamaoka et al. (1991) estimated the position of the source of the wide-frequency tremor that preceded a small explosive eruption on 16 November 1987 of Oshima volcano from the cross-correlate analysis of the 0.7 Hz component and the propagation time of the tremor recorded by a dense 13-station seismic network (Figure 12.5). They obtained a source location at a depth of about 350 m above sea

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00012-8
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Figure 12.1 Volcanic tremor recorded at Volcán de Colima, México, with short-period instrument at a distance of 1.6 km from the crater during the 2002 effusive eruption. Courtesy of Colima Volcano Observatory.

level (a.s.l.), which is slightly deeper than the crater bottom before the 1986 Oshima eruption (see Section 5.2.1).

12.2.2 Etna Volcano, Sicily

The tremor source during the 2008–2009 Etna eruption was estimated by two methods: using the seismic amplitude decay and by the comparison of the seismic signal envelopes (Cannata et al., 2010). Signals were recorded by 16 stations, belonging to the broadband permanent seismic network and located at distances ranging between 1.5 and 9 km from the center of the summit area (maximum elevation of 3300 m a.s.l.). The first method assumed the propagation in homogeneous medium and seismic amplitude decay with distance. Three different frequency bands, 1–5, 5–10, and 10–15 Hz, were analyzed (Figure 12.6). In the first case (Figure 12.6A), the tremor sources were located in a volume between the eruptive fissure and the southeast crater (SEC) at an altitude of 1.0–2.8 km a.s.l. The source of the seismic signal, filtered in the bands 5–10 and especially 10–15 Hz (Figure 12.6B and C), was located below the SEC at very shallow depth (about 3 km a.s.l.). The second method, based on the comparison of the seismic signal envelopes, gave the locations that generally are consistent with the locations obtained by the first method (Figure 12.6A–C). At the same time, a regular shift of these source estimations is seen, compared to the sources obtained by the first method.

12.3 Volcanic Tremors in Eruptive Process

Volcanic tremor reflects many important variations in eruptive activity. Some cases of these relationships are shown below. For proper comparison between episodes

Figure 12.2 Vertical velocity broadband records of monotonic (A) and nonharmonic (B) tremors observed at Sakurajima volcano, Japan. The initial time of the records and peak-to-peak amplitude (in 10^{-8} m s^{-1}) of each seismogram are given at the upper-right and lower-right corners. Vertical scales of spectra are arbitrary. From Kawakatsu et al., 1992.

of tremor at different volcanoes, the standardization of tremor amplitudes to a common reference is used. For surface waves, the following formula is used:

$$\text{R.D.(reduced displacement)} = \frac{A\sqrt{\lambda}r}{2\sqrt{2}M} \tag{12.1}$$

where A is amplitude in cm peak-to-peak, r is a distance from source to seismic station in cm, M is seismograph magnifications at the tremor frequency, and λ is the wavelength in cm (McNutt, 1992).

12.3.1 Etna Volcano, Sicily

Gresta et al. (1991) demonstrated a good relationship between the course of the 1984 activity at Etna volcano, Sicily, and the variations in the amplitude of volcanic tremor. They showed that this relationship is clearer for the spectral amplitudes than for overall amplitudes. The most sensitive were the amplitudes of a 1.2 Hz peak (Figure 12.7); it was the only one that displayed coincidence with the more or

Figure 12.3 Example of particle-motion analysis on a time series of tremor recorded during 1995 at Stromboli volcano, Italy, at a distance of 1.8 km from the crater. (A) Three-component short-period seismic records filtered with a Butterworth bandpass between 1.55 and 2 Hz; (B) particle motion. The north–south-directed particle motion with sub-horizontal incidence indicates a source close to the surface and the P-wave nature of volcanic tremor. From Falsaperla et al., 1998.

less remarkable events of volcanic activity at all four summit craters of the volcano. At the same time, it can be seen that an increase in the spectral amplitudes of the tremor does not depend on the type or intensity of volcanic activity; it may only indicate a surface eruptive manifestation. The spectral amplitudes of the peaks of 0.95, 1.45, 1.65, 1.8, and 2.4 Hz, that were also dominant during the eruptive activity, were characterized by a lower sensibility to eruptive manifestations.

12.3.2 Pavlof Volcano, Alaska

The eruption was continuous from 16 September to 4 December 1996, declined, and then had two stronger bursts on 10–12 December and 27–29 December (McNutt, 2000b). Figure 12.8 demonstrates the ability of tremor to reflect the

Figure 12.4 Example of particle-motion analysis on a time series of tremor recorded during 1961 at Klyuchevskoy volcano, Kamchatka, at a distance of 9 km from the crater. (A) The east−west component of short-period seismic records; (B) particle motion. The ellipse-shaped particle motion indicates the Raleigh-wave nature of volcanic tremor.
From Tokarev, 1966.

explosive phases of the eruption by an increase in amplitude. The two largest peaks in amplitude correspond to the stronger December bursts.

12.3.3 Kilauea Volcano, Hawaii

The 1984 East-Rift eruption that produced about 11×10^6 m^3 of lava, began on 19 September at 16:00 h and continued for 13.5 h. This eruption was accompanied by harmonic volcanic tremor recorded at a distance of 6 km west from the site of the eruption and at the summit of Kilauea volcano. A volcano-tectonic earthquake swarm did not precede the eruption. Figure 12.9 shows the variations of the tremor amplitude recorded at volcanic summit and in the East-Rift zone, near the site of the eruption. The tremor near the site of the eruption was more intensive than at the summit of Kilauea. The tremor began with the start of the eruption, gradually reached its maximum value over 7 h, and stayed at this level up to the end of the eruption. A rapid decrease of tremor accompanied the end of sustained high fountaining. The summit tremor began in 6 h after the start of the eruption and reached its maximum amplitude in 2 h after the cease of the lava emission (Koyanagi et al., 1987).

Figure 12.5 Source location (+) of volcanic tremor in the summit area of Oshima volcano. Seismic network is shown by the black circles with numbers.
From Yamaoka et al., 1991.

12.3.4 Klyuchevskoy Volcano, Kamchatka

The 8 September 1994 summit eruption of Klyuchevskoy volcano was preceded and accompanied by intense volcanic tremor (Ivanov, 2010). The eruption began on with lava fountaining and ash plumes and continued for about 1 month. It culminated in the paroxysmal explosion on 1 October with the ash column estimated at 15−20 km altitude a.s.l. and extended >100 km southeast (BGVN, 1994).

Figure 12.10A shows the variations of the intensity of seismic signals, expressed in relation of mean daily amplitude of tremor to the mean period of tremor signals during this day and recorded at the distance of 11 km from the crater (Ivanov, 2010). It is seen that the intensity of tremor sharply increased with the beginning of eruption on 8 September. The gradual increase of tremor intensity continued up to 12 September. On 12 September, a large gas-and-ash explosion occurred. The ash plume reached the altitude of about 8 km a.s.l. The explosion was accompanied by a 1-km-long lava flow on the southwest slope of the volcano; mudflows were also noted. Beginning from 12 September, the intensity level of continuous volcanic tremor was high and relatively stable till 20 September. During this period, gas-and-ash bursts rose 500−700 m above the crater. The eruption column reached 1.5−2.0 km above the crater. Lava flows extruding from two vents 200 m below the crater rim had moved down to 2800 m elevation on the northwest and southwest flanks. Phreatic explosions were occurring at the contact of the northwest lava flow and the glacier. Lava fountains in the central crater reached heights of 300−500 m.

From 20 September, a slow growth of tremor intensity began (Figure 12.10A). During this period, gas-and-ash bursts increased in height to 800−1000 m above the crater. The eruption columns continued to reach ~2 km above the crater. Lava

Figure 12.6 Source locations of Etna volcanic tremor obtained by amplitude decay in the frequency bands 1–5 Hz (A), 5–10 Hz (B), and 10–15 Hz (C) on 29 August and 2–3 October, and by the comparison of the seismic signal envelopes during 27 August–3 September and 2–3 October 2008. EF is the eruption fissure; SEC is the southeast crater. From Cannata et al., 2010.

flows on the northwest and southwest flanks remained active, and fountains in the central crater increased to heights of 500–700 m. Two days before the paroxismal explosion of 1 October, the level of volcanic tremor began to increase significantly up to moment of this explosion. Volcanic tremor was continuous with a maximum amplitude of 8.4 μm. After the paroxysmal explosion, volcanic tremor practically disappeared (BGVN, 1994).

The detailed analysis of the tremor intensity shows that about 1 week before the 8 September eruption there were recorded the precursors to this eruption (Figure 12.10B). The low-level background tremor that began about 2 months before the eruption (stage I in Figure 12.10B), gradually increased from 2 to 8 September (stage II). Then it sharply increased during 8–12 h before the eruption (stage III) (Ivanov, 2010).

Figure 12.7 Amplitude variations of spectral peak at 1.2 Hz of volcanic tremors recorded at Etna volcano during its 1984 activity. The types of volcanic activity are marked by A (ash ejection), V (emission of vapor and/or gas), and S (Strombolian activity). The intensity of eruptive activity is shown by 1 (moderate), 2 (strong), and 3 (very strong). N corresponds to the absence of any activity. A.U. means the arbitrary units.
After Gresta et al., 1991.

12.4 Relationship Between the Intensity of Volcanic Tremor and Volcanic Events

Figure 12.11 shows a comparison between volcanic tremor amplitude and video camera frames of the 29 May 2000 lava fountain episode at Etna volcano (Alparone et al., 2003). A good correlation between the amplitudes of volcanic tremors and the height of lava fountains and pyroclastic material was noted for many volcanoes (Figures 12.12 and 12.13). The data obtained for two volcanoes (Kilauea, Hawaii, and Pavlof, Alaska) overlap (Figure 12.12), suggesting similar relationships for both volcanoes. McNutt et al. (1991) noted that the Kilauea data are from the first phase of the 1959 eruption; later data showed stronger tremor for the same fountain heights.

Tokarev et al. (1984b) showed that the time of output of incandescent material coincided with the time of origin of an intensive seismic impulse in the sequence of volcanic tremors recorded during the November 1976 Strombolian explosions at the southern site of New Tolbachik volcanoes in Kamchatka (Figure 12.13A). The amplitude of volcanic tremor was well correlated with the height of pyroclastic material rising out of the volcanic crater. With an increase in the explosion frequency, the tremor impulses interfere with the decreasing of the amplitude (Figure 12.13B).

Figure 12.8 Variation of normalized amplitude of volcanic tremor during the 3.5 months of the 1996 Pavlof eruptions. Stronger tremor accompanied more explosive phases (shown by filled points) on 18 October, 4 and 23 November, and 10 and 27 December.
From McNutt, 2000b.

Figure 12.9 Variation in the tremor peak-to-peak amplitudes (in micrometers) during and after the September 1984 East-Rift lava eruption at Kilauea volcano, Hawaii, at the volcanic summit (A) and at a distance of 6 km west from the site of lava emission in the East-Rift zone (B). The period of lava emission is shown.
From Koyanagi et al., 1987.

A more general relationship between the maximum amplitude of volcanic tremor and the volcanic explosivity index (VEI) was obtained by McNutt (1994) based on the data from 21 eruptions at 14 volcanoes (Figure 12.14). A lineal regression of amplitudes (in R.D. units) plotted versus VEI yielded the following regression:

$$\log \text{R.D.}(\text{cm}^2) = 0.48(\text{VEI}) + 0.08 \tag{12.2}$$

Figure 12.10 Variations in the intensity of volcanic tremor before and during the 1994 eruption of Klyuchevskoy volcano, Kamchatka, during September–October 1994 (A) and the detailed representation of preliminary variations of tremor intensity from 17 August to 12 September 1994 (B). In (B), I is the stage of the background level of tremor; II is the stage of linear increase in tremor intensity; III is the stage of exponential growth of tremor intensity coinciding with increase of degassing activity of the volcano; IV is the eruption stage. After Ivanov, 2010.

12.5 Special Cases of Volcanic Tremors

12.5.1 Isolated Tremors

Large-amplitude volcanic tremor was observed before the 1989 Teishi Knoll, Izu islands eruption. Figure 12.15 shows the position of these signals in the total sequence of seismotectonic events preceding the eruption (Ukawa, 1993).

Figure 12.11 Comparison between volcanic tremor amplitude and video camera frames of the 29 March lava fountain episode at Etna volcano.
From Alparone et al., 2003.

The sequence commences with a swarm of volcano-tectonic earthquakes (see Section 5.5.1) that continued for about 7 days from 4 July and was accompanied by a large increase in crustal tilt. After cessation of the volcano-tectonic activity and during the maximum tilt, a sequence of low-frequency earthquakes occurred. At a coda of the sequence of low-frequency earthquakes, the continuous tremor began. It culminated with the large-amplitude isolated tremor that took place for about 1 h and its conclusion was coincident with the beginning of the submarine eruption.

Figures 12.16 and 12.17 show the waveforms of the volcanic tremor and their spectra. Here F1 and F2 indicate the episodes of continuous tremor occurring on 11 and 12 July, the large-amplitude tremor of 13 July is shown as the main phase, and the isolated events are shown in the lower part. The power spectra of these events demonstrate that every tremor has its spectral peak at around 0.8 Hz. The spectral shape of the main phase is similar to that of the isolated events. The spectra of the continuous tremor F1 and F2 are different in shape from those of the large-amplitude tremors and are characterized by narrower peak bands. These spectral properties indicate that the F1 and F2 tremors were produced by smaller-scale events compared with the large-amplitude tremors (Hashimoto et al., 1991).

Oshima et al. (1991) considered that the continuous tremors F1 and F2 of 11 July were associated with the formation of a new volcanic knoll (cryptodome). The large-amplitude tremor of 13 July was believed to be caused by the final moments

Figure 12.12 Relationship between the fountain height at basaltic volcanoes Kilauea, Hawaii (round dots) and Pavlof volcano, Alaska (triangles), and tremor amplitude. Both data sets were normalized to units of reduced displacement. The line represents a lineal regression fit to Kilauea data.
From McNutt et al., 1991 and Eaton et al., 1987.

Figure 12.13 Dependence of the amplitude of volcanic tremor upon the height of pyroclastic material rising out during the explosions at New Tolbachik volcanoes, Kamchatka. The shape of pyroclastic cones (A) and a seismogram (vertical component) of volcanic tremor (B) were recorded simultaneously by filming and by short-period seismometer, respectively, at a distance of about 3 km from the site of the eruption. The seismogram is moved to the left for 2.5 s to correct for the seismic wave propagation from the site of eruption to seismic station.
From Tokarev et al., 1984b.

Figure 12.14 Volcanic tremor reduced displacement versus volcano explosivity index (VEI, Newhall and Self, 1982). *R* is the correlation coefficient. The least square regression line is shown.
From McNutt, 1994.

Figure 12.15 Sequence of the seismotectonic events associated with the 1989 Teishi Knoll, Izu islands eruption. Earthquake number is shown for every 4 h; the epochs J1 to J5 reflect the increase of crustal tilt from J2 to J5 that exceeded 20 mrad.
From Ukawa, 1993.

Figure 12.16 Seismograms of the isolated tremors preceding the 1989 Teishi Knoll, Izu islands eruption. Recorded at a distance of about 8 km from the site of submarine eruption. From Hashimoto et al., 1991.

of pre-eruptive and eruptive magma movements. The hypocenters of the isolated events are located near the position of the submarine eruption (Goto et al., 1991).

A study of the particle motions for the large-amplitude tremors (Hashimoto et al., 1991) indicates their surface-wave nature and shows that the radial and vertical components of the motions are represented by Rayleigh waves while the transverse components are Love waves (Figure 12.18).

The newborn New Tolbachik cones in Kamchatka were formed along a freshly opening fault on 6 July 1975 (see Section 5.3.1). The 16-min record of low-frequency volcanic tremor was observed just before the beginning of the preceding swarm of volcano-tectonic earthquakes (Figure 12.19). It represented the continuous vibrations with a gradual increase in the amplitude. Maximum amplitudes were four times larger than the initial amplitudes. After reaching a maximum, the

Figure 12.17 Power spectral density (in μm^2 s) curves of NS component of isolated tremors preceding the 1989 Teishi Knoll, Izu islands eruption shown in Figure 12.16. Dashed lines indicate the frequency of 0.8 Hz.
From Hashimoto et al., 1991.

Figure 12.18 Waveforms of the isolated event E2 recorded at a distance of about 8 km from the site of submarine eruption of Teishi Knoll volcano (A) and the particle motions (B) at 2-s intervals. The arrows on $R-Z$ plane indicate retrograde motions that characterize the Rayleigh-type surface waves.
From Hashimoto et al., 1991.

amplitudes began to gradually decrease. With the onset of the volcano-tectonic earthquake swarm, the tremors ceased and were not observed for the following 2-week seismic activity up to the beginning of the explosive activity of the new-formed cone (Tokarev et al., 1984b).

Figure 12.19 Low-frequency isolated volcanic tremor recorded by a short-period seismic station at a distance of about 50 km from the site of a newborn cone of New Tolbachik volcanoes, Kamchatka, 10 days before the beginning of the 6 July eruption.
Seismic record is courtesy of the Kamchatkan Methodical Seismological Department of the RAS Geophysical Survey, Russia.

12.5.2 Banded Tremor

A particular kind of volcanic tremor characterized by regular cyclic increases of amplitude was observed in several volcanic and hydrothermal areas. Because of its peculiar seismic signature, forming evident stripes on seismograms, this kind of seismic signal is called a "banded" tremor (Figure 12.20).

The study of banded tremor was realized at Miyakejima volcano after its 2000 eruption (see Section 5.5.2). As was noted by Fujita (2008), the banded tremor was not observed throughout the entire period of the 2000 volcanic activity but only intermittently between 2001 and 2006. The active periods of banded tremor were occasional and each period lasted about 1–2 days on average, ranging from a few hours to 10 days. Remarkable banded tremor was recorded during the periods of April and August 2002; March, April, August, and September 2003; March–April, May–June, July–August, and November–December 2004; and April and August 2005. It seems that the banded tremor had seasonal activity. The durations of the tremor episodes in each active period were similar. The durations of signals during spring season (March and April) were shorter than those in summer (May–August). Figure 12.21 shows the histograms of durations and intervals observed during the whole 2001–2006 period of observations and this clearly shows that there were typical values for both characteristic times about 6.5 and 25 min, respectively.

The duration of the tremor episodes and the intervals between them were studied also at Etna volcano during its low degassing activity occurring just after the end of the 13 May to 6 July 2008 effusive–explosive eruption from an eruptive fissure opened east of the summit crater (Cannata et al., 2010). It was found that the first studied period, 20 August to 9 September 2008, was characterized by 620 clear episodes of banded tremor with fairly steady duration for the entire period and equal to about 25 min (Figure 12.22A). The interval between the end of an episode and the onset of the following one was also quite steady and roughly equal to the time duration of the episodes (Figure 12.22C). During the second period of 1–27 October 2008, 650 episodes of banded tremor occurred, with average episode duration of about 30 min and intervals between the end of an episode and the onset of the following one roughly equal to 25 min (Figure 12.22B and D).

Figure 12.20 Examples of waveform of banded tremor waveforms recorded at Miyakejima volcano on (A) 15 August 2002, (B) 11 August 2003, and (C) 29 March 2004. Each plate includes a 1-day seismic record consisting of 24 traces of 1-h records of vertical components of a velocity seismometer installed at about 100 m depth at the distance of about 1.8 km from the summit. Spindle-shape packets, called "banded tremors" are recorded intermittently at similar intervals.
From Fujita et al., 2008.

Figure 12.21 The histograms of (A) duration and (B) intervals of banded tremor at Miyakejima volcano.
From Fujita et al., 2008.

This type of distribution of the banded tremor episode durations and the intervals between them, obtained for Etna and Miyakejima, suggest the existence of some threshold to trigger the banded tremor (Fujita, 2008). The important factor for understanding the nature of banded tremor is also its absence during the eruption activity. Fujita (2008) proposed the shallow hydrothermal system as the most plausible mechanism for the generation of banded tremor. The shallow water aquifers appear to be capable of maintaining continuous gas emission and triggering oscillations, while magma plays only a passive role specifically as a heat source. Fujita (2008) modeled

Figure 12.22 Durations of the banded tremor episodes observed at Etna volcano (A and B) and the length of the intervals between the end of one episode and the start of another (C and D). The tremor was recorded during the periods 20 August–9 September (A and C) and 1–27 October 2008 (B and D), respectively.
From Cannata et al., 2010.

Figure 12.23 Broadband records of long-period seismic events recorded at Aso volcano, Kyushu, without surface activity and during the phreatic explosion. Both types of records filtered at 15 s have comparable waveforms.
After Legrand et al., 2000.

the source of this phenomenon with a two-phase hydrothermal instability flow model, consisting of sub-cool and superheated water regions. The interval between banded tremor episodes can be interpreted in terms of this two-phase flow oscillation with a heat supply from the magma beneath the hydrothermal system.

12.5.3 Long-Period Tremor

Long-period (longer than 7 s) signals of rather short duration (several tens of seconds) have been observed at Aso volcano, Kyushu, by a broadband network in 1984 (Kawakatsu et al., 2000). These long-period signals occurred at the volcano during the small phreatic explosions or without surface activity (Figure 12.23). The most energetic period of 15 s is always clearly recorded.

These events were located within the depth range from 1.2 to 1.6 km above the volcanic crater. Legrand et al. (2000) used a linear inversion to estimate the moment tensor of the events. The inversions showed a similar isotropic mechanism for all 43 long-period tremors and the 6 phreatic explosions.

Long-period volcanic tremors were observed also at Usu volcano by (Yamamoto et al., 2002) during the first weeks of the 2000 eruptive activity just after the 31 March and 1 April phreatic explosions (see Section 7.2). Long-period volcanic tremors were recorded as isolated wave packets. The period of these tremors varied gradually from 10 to 12 s during the first 2 or 3 days after the first eruption, and became steady at 12 s. The source of long-period tremors was located by the waveform semblance method. The estimated source region was about 1.5 km southeast of the summit crater at a depth of 5–6 km (Figure 12.24). The best fitting seismic moment tensor, estimated for the 7 April 2000 isolated wave

Figure 12.24 The source location of long-period tremor that was observed just after the 2000 phreatic explosions from craters N and K at Usu volcano, Japan. ABT, ZEN, and SNZ are broadband seismic stations.
From Yamamoto et al., 2002.

packet of long-period volcanic tremor, corresponded to a spherical source and consisted of an isotropic component (the amplitude ratios M_{xx}:M_{yy}:M_{zz} are 1:1:1) and a compensated linear vector dipole (CLVD) component, with the ratio of these two components equal to (3.75):(−1) (Figure 12.25).

Figure 12.25 Source mechanism of the 7 April 2000 long-period signals generated by volcanic tremor at Usu volcano. CLVD is the compensated linear vector dipole. From Yamamoto et al., 2002.

Figure 12.26 Locations of the 1973−1977 deep tremor sources under Kilauea volcano, Hawaii (A) and their depth distribution (B).
Data from Aki and Koyanagi, 1981.

12.5.4 Deep Tremor

Deep tremors may be distinguished from shallow ones by their amplitude distribution. For Kilauea, Hawaii, deep tremor shows a more uniform amplitude for many stations over a large area as compared to the shallow events, which show the concentration of large amplitudes only at a few nearby stations. Deep Hawaiian tremor demonstrated an identical period at almost all stations but it had a tendency to change with time. Figure 12.26 shows the source locations of 24 deep tremor episodes that were determined using the onset times. The tremor sources were distributed at a distance from 13 to 40 km to the southwest of Kilauea volcano within a depth range of 27–75 km (maximum number of events were recorded between the depths from 35 to 45 km) with the mean error in-depth estimation of about 6 km (Aki and Koyanagi, 1981).

There exists a dependence of the duration of deep volcanic tremor episodes on their reduced displacements. Figure 12.27 shows that the duration decreases exponentially with an increase in reduced displacement. Aki and Koyanagi (1981) consider that it may indicate a unique length scale involved in the source process of deep volcanic tremor such as the average size of the conduits.

Aki and Koyanagi (1981) noted that the deep magma transport inferred from the deep tremor data is relatively unaffected by volcanic activity near the surface. At

Figure 12.27 The duration of tremor per year recorded with the reduced displacement for deep tremors under Kilauea volcano, Hawaii during 1962–1979. The duration decreases exponentially with the increase in reduced displacement.
From Aki and Koyanagi, 1981.

the same time, numerous publications that appeared beginning from 2002 (Obara, 2002; Nadeau and Dolenc, 2005; Payero et al., 2008) show the existence of deep (between 20 and 40 km of depth) nonvolcanic tremor on some subduction zone thrust faults and on the San Andreas Fault. A number of studies associate this deep tremor as a manifestation of the process on the transition zone between the seismogenic coupled and deep free-slipping segments of the subduction interface. Therefore, it is possible that the deep tremor beneath Hawaii is also of the same nature but not of volcanic origin.

13 Seismic Signals Associated with Pyroclastic Flows, Rockfalls, and Lahars

Volcanic hazard is significantly related to the propagation of density currents, such as pyroclastic flows and lahars. These flows represent the motion of natural sediments generated along volcanic slopes that propagate with high velocity and can devastate everything in their path. Both types of volcanic flows have produced destructive effects during human history. In the list of the 27 largest volcanic catastrophes between 1700 and 1986 that were responsible for at least 1000 deaths (Bardintzeff and McBirney, 1999), 11 were associated with pyroclastic flows and 11 with lahars. Therefore, the ability to monitor and track these phenomena in real time is important for the mitigation of volcanic risk.

13.1 Occurrence of Pyroclastic Flows, Rockfalls, and Lahars During Volcanic Eruptions

Figure 13.1 shows the typical occurrence of pyroclastic flows, rockfalls, and lahars during the 2004–2006 effusive-explosive volcanic activity at Volcán de Colima, Mexico. The extrusion of lava generated numerous pyroclastic flows and rockfalls in October 2004 due to partial collapses of the summit dome. The next, and more voluminous, portion of pyroclastic flows was produced by the fall of volcanic column material associated with large explosions in 2005. Lahars were observed during the rainy seasons (from June to October) in 2005 and 2006. Their number was larger in 2005 just after lava extrusion and the large explosions. In 2006, their number decreased with decreasing eruptive activity.

13.2 Seismic Signals Associated with Pyroclastic Flows and Rockfalls: Waveforms and Spectra

Pyroclastic flows and rockfalls are common for both explosive and effusive activity at andesitic and dacitic volcanoes. They may be generated by a number of different mechanisms. Among them is the gravitational collapse of a growing lava dome, the

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00013-X
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collapse from an eruption column, and the explosive destruction of growing lava dome (Francis, 1998; Carey and Bursik, 2000). As noted by Calder et al. (2002), pyroclastic flows are distinguished from rockfalls by their larger size, longer runout (≥ 0.5 km), greater production of fine ash and appreciable buoyant hot ash clouds. The fall-and-flow of fragments of lava dome or flow fronts and the numerous rockfalls generate seismic signals.

Figure 13.1 Sequence of volcanic events during the 2004–2005 effusive-explosive episodes at Volcán de Colima. (A) Variations in the rate of lava extrusion (Zobin et al., 2006). (B) Seismic activity associated with the occurrence of pyroclastic flows and rockfalls (PF&RF) and with small gas-and-ash explosions (S_EX). The period of large explosions is shown. (C) Occurrence of lahars.
From Zobin et al., 2009a.

Pyroclastic flows and rockfalls are associated with similar emergent, cigar-shaped waxing and waning seismic signals with frequencies of 1–10 Hz, which differ only in their durations and/or amplitudes (Calder et al., 2002). The signals produced by pyroclastic flows and rockfalls may represent the main component of seismicity during block-lava emission, subsequent lava dome destruction, and explosive activity.

13.2.1 Seismic Signals of Pyroclastic Flows Produced by the Partial Collapse of Lava Dome

Figure 13.2A and B shows broadband seismic record of pyroclastic flow associated with the partial lava dome collapse occurring at Soufrière Hills volcano on 8 January 2007 and recorded at a distance of about 2 km from the crater. De Angelis et al. (2007) give the detailed characteristics of the observed time series. These signals can be divided into three main phases. Phase 1 is characterized by a steady increase in amplitude with time and is supposed to represent the progressive failure of slip zones with variable orientation along the boundaries of an unstable portion of the dome as the result of increased internal pressurization. Phase 2 consists of a strong spindle-shaped pulse; the onset is emergent and no clearly defined seismic phases could be recognized in the seismic records. This phase is thought to represent the fall of the unstable portion of the lava dome, its collision with the slopes of the volcano and resultant fragmentation, and the descent of the flows down the flanks of the edifice. Phase 3 includes a sequence of seismic pulses possibly due to the remobilization of material deposited along the same path by the pyroclastic flow during phase 2.

Figure 13.2 Seismic envelope (A) and broadband velocity seismogram (vertical component) (B) for the signals associated with pyroclastic flow at Soufrière Hills volcano, recorded on 8 January 2007 at a distance of about 2 km from the crater. The seismogram is divided into different phases according to variable observed activity (see text).
After De Angelis et al., 2007.

Figure 13.3 Video image of the event 2004_10010100 (A) and broadband seismic records (vertical component) of pyroclastic flows produced by the partial collapse of lava dome at Volcán de Colima and recorded at a distance of 4 km from the crater (B). The index near the seismic record represents the date (yyyy_mmddhhmm) of the events. The rectangle on the seismogram of 2004_10011405 shows the selected sample for Figure 13.4.
Courtesy of Colima Volcano Observatory.

Figure 13.3 shows broadband seismic records associated with the pyroclastic flows produced by the partial collapse of a lava dome during the 2004 lava extrusion at Volcán de Colima, México and recorded at a distance of 4 km from the crater. They correspond to phase 2 in Figure 13.2. The waveforms begin with a low amplitude signal which gradually increases in amplitude and then more slowly

Figure 13.4 Details of seismic record 2004_10011405 of the pyroclastic flow that was shown in Figure 13.3 in rectangle. Arrows with numbers show different seismic impulses (see text).
Courtesy of Colima Volcano Observatory.

decreases. Figure 13.4 shows the details of the initial part of the record. It consists of an initial relatively short-period impulse (1) of duration of about 2 s, followed by a long-period impulse (2) of duration of about 9 s, and then a long-duration short-period impulse (3).

Comparison with the seismograms and video images observed for the Unzen volcano pyroclastic flows (Yamasato, 1997; see Chapter 11) showed that the first impulse may be produced by the lava dome collapse, the second corresponds to the downfall of the lava blocks onto the slope, and the third is associated with the generation of pyroclastic flows. The durations of the first two impulses are comparable with those reported for the Unzen pyroclastic flows (Yamasato, 1997).

13.2.2 Seismic Signals of Pyroclastic Flows Produced by the Collapse of Eruption Column

Figure 13.5 shows broadband seismic records associated with the pyroclastic flows produced by the collapse of an eruption column during the four large 2005 Vulcanian explosions at Volcán de Colima, México. Pyroclastic flows entered the Colima ravines and reached distances up to 3–6 km from the summit. Satellite imageries identified the ash clouds produced during these eruptions at heights between 7.6 and 8.5 km a.s.l. Seismic station was situated at a distance of 4 km from the crater.

The records of this type of pyroclastic flows are more complex comparing with those produced by the partial collapse of dome. The detailed structure of the seismic record "explosion–pyroclastic flow" is shown in Figure 13.6. The seismic signal begins with the large-amplitude records of the explosions; then the collapse of volcanic material from an eruption column generates pyroclastic surge. The generation of pyroclastic flow begins 20–25 s after the onset of the seismic signal. The

Figure 13.5 Photo of the 2005_03132128 explosion (A); broadband seismic records (vertical component) of explosions and the following pyroclastic flows produced by the collapse of volcanic material from the volcanic column at Volcán de Colima and recorded at a distance of 4 km from the crater (B). The bold-line rectangle on the seismogram of 2004_10011405 shows the selected sample for Figure 13.6. The thin-line rectangles show the selected samples for Figure 13.7.
Courtesy of Colima Volcano Observatory.
Photo of the 2005_03132128 explosion was taken by Sergio Velasco.

record of an explosion has significantly larger amplitudes than the signal of a pyroclastic flow. We can identify at least 5 impulses. As was proposed in (Zobin et al., 2006b), the first two impulses are generated by movement of fragmented magma above the fragmentation surface in the shallow conduit (1) and by an explosion in

Figure 13.6 The structure of the seismic record of explosion 2004_10011405 at Volcán de Colima and the following pyroclastic flow shown in Figure 13.5. Arrows with number shows the arrivals of the seismic impulses (see text).
Courtesy of Colima Volcano Observatory.

the conduit (2). The impulses 3, 4, and 5 are equivalent to the impulses 1–3 in Figure 13.3 for the pyroclastic flows originating from a lava dome collapse: impulse 3 is generated by the rockfall from the eruptive column, impulse 4 corresponds to the movement of the rocks along the slope and impulse 5 is generated by the pyroclastic flow. The waveform of the impulse 5 is more heterogeneous than the impulse 3 in Figure 13.3. The fall of volcanic rocks from the eruptive column and generation of pyroclastic flow begins, therefore, 20–25 s after the onset of the seismic signal. The continuous fall of the rocks from the column and their following movement adds the low-frequency component in this phase.

Figure 13.7 shows a 100-s portion of the seismic records of Figure 13.5 following the explosive onset. To remove the low-frequency noise, they were high-pass filtered at 2.5 Hz. After filtering, they are similar enough to the records of pyroclastic flows generated by the collapse of a lava dome. At the same time, their amplitudes are one order higher than for the seismic records of the block-and-ash flows initiated by the partial collapse of a lava dome.

13.2.3 Seismic Signals of Pyroclastic Flows Produced by the Explosive Destruction of Growing Lava Dome

Figure 13.8B shows short-period seismic record associated with pyroclastic flow that followed the 9 May 2006 explosive activity at Bezymianny volcano, Kamchatka and was recorded at a distance of 40 km from the crater. This eruption consisted of two phases recorded by video and seismic stations (Droznin and Droznin, 2007). During the first phase, represented by three explosions, gas-and-ash plumes with the height up to 6 km a.s.l. were produced from the summit during about 3 min. The second phase, that began about 5–7 min after the end of the first phase, was represented by the propagation of pyroclastic flow, generated due to the

Figure 13.7 Seismic records (samples in rectangles selected in Figure 13.5) of pyroclastic flows following explosions and produced by the collapse of volcanic material from an eruptive column at Volcán de Colima.
Courtesy of Colima Volcano Observatory.

explosive destruction of the lava dome, and by the generation of the co-ignimbrite eruption plume produced from a pyroclastic flow and reaching the height of 11 km a.s.l. The umbrella of the eruption plume is shown in Figure 13.8A. The seismic record allows discriminating the initial seismic signal associated with the explosive events (shown in ellipse), and the secondary group of seismic signals associated with the propagation of pyroclastic flow (shown in rectangle). Duration of the seismic signal, associated with pyroclastic flow, was about 16 min.

Figure 13.8 Photo of the umbrella of the 9 May 2006 co-ignimbrite plume at Bezymianny volcano produced by pyroclastic flow (A), short-period seismic record (horizontal component) associated with the gas-and-ash plumes (shown in ellipse) and the following pyroclastic flow (shown in rectangle) (B), and (C) Fourier spectrum of the signal produced by pyroclastic flow. The seismic signal was recorded at a distance of 40 km from the crater. Seismic record is courtesy of the Kamchatkan Methodical Seismological Department of the RAS Geophysical Survey, Russia.
Photo was taken by Yu. Demenchuk.

13.2.4 Seismic Signals Produced by Rockfall

Figure 13.9 shows the seismic records associated with rockfalls during the 2004–2005 effusive-explosive episodes at Volcán de Colima, México. The waveforms are similar to those produced by pyroclastic flows although shorter. The

Figure 13.9 Video image of the event 2005_02071657 (A) and broadband seismic records (vertical component) of rockfalls observed at Volcán de Colima during the 2004–2005 effusive-explosive episodes and recorded at a distance of 4 km from the crater (B). Courtesy of Colima Volcano Observatory.

same observation emerges from the comparison of the seismic signals produced by rockfalls and pyroclastic flows at Soufrière Hills volcano (Figure 13.10) (De Angelis et al., 2007). The amplitudes of the rockfall seismic records obtained at Volcán de Colima may be comparable with the amplitudes of the pyroclastic flows produced by the partial collapse of a lava dome.

Figure 13.10 Short-period seismic records associated with the pyroclastic flow (A) and rockfall (B) observed during the 2007 activity at Soufrière Hills volcano and Fourier spectra of these signals.
After De Angelis et al., 2007.

13.2.5 Spectral Characteristics

Fourier spectra of the seismic signals, associated with the propagation of pyroclastic flows and recorded at a distance of 4 km from the crater of Volcán de Colima, are shown in Figure 13.11. They were calculated for broadband seismograms including the seismic records of pyroclastic flows initiated by a partial lava dome collapse (PF_collapse); the seismic records of an explosion and following pyroclastic flow (Explosion + PF) and of the pyroclastic flow only (PF_explosion); and the seismic records of rockfalls. The spectra of both types of pyroclastic flows are similar with close peak frequencies (between 2.4 and 4.4 Hz).

Fourier spectrum of the seismic signal, produced by the pyroclastic flow that followed the 9 May 2006 explosive destruction of the lava dome at Bezymianny volcano and recorded at a distance of 40 km (Figure 13.8C), had its maximum amplitude at 0.7 Hz. Its spectral characteristics is close to the peak frequencies of

Figure 13.11 Spectra of broadband seismic signals recorded during the effusive-explosive activity at Volcán de Colima (shown in Figures 13.3, 13.5, 13.7, and 13.9). Heavy red lines show spectra of the signals associated with the pyroclastic flows produced by the partial collapse of lava dome (PF_collapse); thin blue lines show spectra of the signals including the seismic record of the explosion and of the following pyroclastic flow produced by the collapse of volcanic material from the volcanic column during the explosion (Explosion + PF); heavy violet lines show spectra of the part of these signals associated with pyroclastic flows only (PF_explosion); and thin black lines show spectra of the signals associated with rockfalls (Rockfalls).

the "explosion + PF" seismic records at Volcán de Colima (0.8–1.3 Hz). Duration of the seismic signal at Bezymianny was significantly longer than the signals recorded during the propagation of the pyroclastic flows observed at Volcán de Colima.

Spectra of the seismic signals associated with rockfalls occurring at Volcán de Colima (Figure 13.11) are characterized by a tendency to be of lower frequency for shorter seismic records and their spectral peaks vary between 1 and 3 Hz.

13.3 Occurrences of Earthquakes Associated with Pyroclastic Flows and Rockfalls

The earthquakes associated with pyroclastic flows and rockfalls occur in dense groups, sometimes completely filling the seismogram (see Figure 2.13). The total number of these events recorded during the 80-day andesitic block-lava emission at Volcán de Colima, México in November 1998–February 1999 exceeded 6000 (Zobin et al., 2005b). Pyroclastic flow and rockfall signals replaced the volcano-tectonic earthquakes from the preceding swarm just after the start of lava extrusion. The rate of seismic energy release during the seismic activity of pyroclastic flows and rockfalls was comparable with the rate of seismic energy release during the precursory swarm of volcano-tectonic earthquakes (Figure 13.12).

Malone (1983) reported that the typical 1980–1981 dome-building eruption episodes at Mount St. Helens, Cascades begin with a few volcano-tectonic earthquakes several weeks before the actual eruption. At the onset of the eruption, the volcano-tectonic events are replaced by avalanche signals and explosion events. Both types of events may last for several days, slowly decreasing in number.

However, pyroclastic flow and rockfall seismicity may begin without any preliminary volcano-tectonic activity as was observed during the 2002, 2004, and 2007 extrusions at Volcán de Colima (Zobin et al., 2008).

13.4 Relationship Between the Pyroclastic Flow and Rockfall Earthquakes and Seismo-Volcanic Activity During Lava Emission

Relationship between the seismic signals, produced by pyroclastic flows and rockfalls, and eruption dynamics has been discussed by several authors. Neuberg (2000) reported that these signals correlated visually with the avalanches of material from the lava dome at Soufriére Hills volcano. Ratdomopurbo and Poupinet (2000) noted that the duration of the seismic signals ("guguran" in Indonesian) at Merapi volcano, Java is several minutes. Their frequencies were higher than that of a tremor. Purbawinata et al. (1997) suggested that the occurrences of "guguran" may indicate the level of stability of the lava dome. When the dome is unstable, "guguran"

Figure 13.12 Cumulative RSEM (real-time seismic energy measurement) curve of variations in seismic energy release at Volcán de Colima during October 1998–February 1999. It is seen that the rate of seismic energy release, generated by volcano-tectonic earthquakes and recorded before the eruption, did not significantly change with the beginning of the lava eruption when the seismic signals were generated mainly by rockfalls. From Zobin et al. (2005b).

occurrence may be greater than 500 events per day, whereas, when the dome is stable, only about three events per day occur. A rate of more than 50 events per day is usually considered as an indication that the dome is in an unstable state.

A dacitic lava dome at Unzen volcano, Japan, emerged during May 1991, and its collapse resulted in the generation of pyroclastic flows. Yamasato (1997) studied the seismic signals produced by the collapse of lava blocks from the dome, the fall of material along the slope, and the migration of pyroclastic flows along the flanks of the edifice. Comparison of the seismic signals with video footage allowed identifying seismic signals associated with the different stages of the formation of a pyroclastic flow. Small-amplitude seismic waves with a predominant frequency of 2–3 Hz were excited almost simultaneously with the collapse of the lava dome. Within a few seconds, when lava blocks fell onto the slope, the amplitude of seismic signal become large with the appearance of a low-frequency (about 0.5 Hz) component. Then the fragmented pyroclastics started to flow generating high-frequency (greater than 2 Hz) seismic waves.

Systematic variations in the number of rockfalls estimated from seismic data may be used in monitoring of eruptive activity. Figure 13.13 shows variations in the number of rockfalls during the 2002–2003 effusive eruption at Volcán de

Figure 13.13 Sketch of lava flows produced during the 2002–2003 lava extrusion at Volcán de Colima (A) and the variations in the number of seismic events produced by rockfalls during January 2002–June 2003 (B). In (A), bold dashed line outlines the crater formed during the 1999 activity. 1–5, numbers of lava flows according to their appearance. In (B), I–III, mark the stages of eruption described in the text. The sketch was prepared by C. Navarro, as well as the count of events.
Courtesy of Colima Volcano Observatory.

Colima. The lava dome filled the crater up until earn February 2002, and two new lava flows began their movement on 14 February (Figure 13.13A). The formation of lava flows was preceded and accompanied by rockfall activity starting on the evening of 4 February and peaking on 24 February (Stage I). At the end of February effusive activity significantly decreased, as reflected by the sharp decrease in the number of rockfalls on Figure 13.13 (Stage II). In May 2002, a new stage of effusive activity began, and three additional lava flows (Figure 13.13A) were formed during June–December 2002 and evolved until February 2003

Figure 13.14 Correlation of the number of rockfalls with the rate of lava emission during the 2004 lava extrusion at Volcán de Colima. *R* is the coefficient of correlation. Figure was prepared using the data collected by C. Navarro.
Courtesy of Colima Volcano Observatory.

(Stage III). This stage corresponds to a sharp increase in the number of rockfalls until November 2002 and followed by a gradual decrease in their number until March 2003.

During the 2004 lava extrusion at the same volcano, the number of rockfall events was good correlated with the rate of lava emission (Figure 13.14).

13.5 Quantification of Pyroclastic Flow and Rockfall Earthquakes

Pyroclastic flow and rockfall signals can be characterized by their amplitude and duration. Figure 13.15 illustrates the seismic signals recorded by short-period seismograph at Volcán de Colima, México. Their durations and amplitudes vary from event to event. Uhira et al. (1994) demonstrated that long-period amplitude of the seismic signals, produced by pyroclastic flows and rockfalls, may vary by about a factor of ten. Simultaneous observations with short- and long-period seismographs

Figure 13.15 Seismic signals recorded during the 1998 lava eruption at Volcán de Colima, México. The records of short-period seismic station at a distance of 4 km from the crater are shown. M1, M2, and M3 denote the records of earthquakes with magnitude 1, 2, and 3. Courtesy of Colima Volcano Observatory.

at Unzen volcano, Japan, showed that long-period records of the pyroclastic flow and rockfall earthquakes are significantly shorter than short-period records of the same signals (Uhira et al., 1994). On the other hand, the durations of broadband and short-period seismic records, produced by pyroclastic flows and rockfalls at Volcán de Colima occurring due to the 1998 partial collapses of the lava dome, were well correlated (Figure 13.16).

Norris (1994) reported that rockfall sequences (Mount St. Helens, Cascades; Makaopuhi crater, Hawaii) show a poor correlation between the maximum signal amplitude recorded by a short-period seismometer and the volume of the rockfalls.

Figure 13.16 Correlation between the durations of broadband and short-period seismic records of pyroclastic flows and rockfalls observed during the 2004–2005 efffusive-explosive episodes at Volcán de Colima. *R* is the coefficient of correlation.

Mills (1991) used the RSAM (real-time seismic amplitude measurement) to investigate rockfall activity at Mount St. Helens and reported a similar poor correlation; rockfalls that differed in volume by several orders of magnitude could be distinguished by their seismic amplitude or duration, but those closer in volume could not.

Comparison of the seismic data with the visual observations of pyroclastic flows at Soufriére Hills volcano, Montserrat showed that short, sharp RSAM peaks represented discrete pulses of large-volume, energetic flows (Calder et al., 2002). Uhira et al. (1994) showed a good correlation between the maximum signal amplitude recorded by long-period seismometers and the volume of pyroclastic flows at Unzen volcano, Japan. As noted by Calder et al. (2002), the relationship between the volume of pyroclastic flow and signal amplitude is not simple.

13.5.1 Quantification of Pyroclastic Flow and Rockfall Earthquakes Occurring Due to Partial Collapse of the Lava Dome and Recorded by Short-Period Instruments at Volcán de Colima, México

Amplitude of short-period seismic signal is technically less reliable parameter of earthquake than its duration; its record may be saturated for large event. It was a

Figure 13.17 Relationship between the number of rockfalls and the duration of their signals during the 1998 extrusion (circles) and the 2004 extrusion (squares) at Volcán de Colima. The least-square regression line log $N = 15.7 - 6.2 \log D$ (s) for the 1998 events is shown. Here N is the number of events.
From Zobin et al., 2008.

reason to apply the coda magnitude *Mc*, based on the duration of the seismic signal, in seismic routine (e.g., Bakun, 1984). Therefore, the duration of seismic signals is convenient to be used as a magnitude parameter of pyroclastic flows and rockfalls.

A duration-frequency plot for the seismic signals associated with pyroclastic flows and rockfalls during the 1998 and 2004 lava eruption at Volcán de Colima was constructed as shown in Figure 13.17. Gutenberg and Richter (1956) showed that the relationship between the magnitude of seismic events and their frequency characterizes the physical conditions of material fracturing. Pyroclastic flow and rockfall earthquakes are products of the destruction of lava bodies. A log-log plot of the duration intervals versus the number of events of this duration evidences a robust linear dependence between the parameters. This relationship indicates a power-law distribution of the seismic signals related to pyroclastic flow and rockfall earthquakes occurring due to partial collapse of the lava dome. This relationship is typical for fractal sets (Scholz, 1990) and may indicate the self-similarity in the dimensions of the pieces that were broken off during the gravitational destruction of the lava body.

13.5.2 Relationship Between the Magnitude of Pyroclastic Flow and Rockfall Earthquakes and the Volume of Pyroclastic Flows

The results presented in the previous sections suggest that the duration of the short-period records of pyroclastic flow and rockfall earthquakes may provide a way to assess the magnitude of these earthquakes. The duration of the signal was measured beginning from the onset of the signal and till the end of its coda. The following magnitude scale was proposed based on the durations of earthquakes observed during the 1998–1999 effusive eruption at Volcán de Colima and recorded by a short-period station situated at a distance of 4 km from the crater:

magnitude 1 for events with durations of seismic signal less than 100 s;
magnitude 2 for events with durations between 100 and 200 s; and
magnitude 3 for events with durations greater than 200 s (see Figure 13.15).

The events of magnitude 1 can be attributed mainly to rockfalls; the magnitudes 2 and 3 events may be associated with pyroclastic flows and large rockfalls. Self-similarity in the dimensions of the pieces that were broken off during the gravitational destruction of the lava body allows us to estimate the mean values of the material that was transported by pyroclastic flows of different magnitude.

The volume of pyroclastic flows that occurred between 25 and 26 November, 1998 (7.2×10^5 m^3) and of the total volume of pyroclastic flows that occurred during the eruption in November 1998–January 1999 (1.6×10^6 m^3) were estimated by Navarro-Ochoa et al. (2002) and by Saucedo et al. (2002), respectively. Thus the approximate mean volume of flows corresponding to seismic records of different durations was estimated. For the 25 and 26 November observations, three events with magnitude 3, 123 events with magnitude 2, and 219 events with magnitude 1 were recorded. For the total period of the eruption, four events of magnitude 3, 577 event with magnitude 2, and 4939 events with magnitude 1 were recorded. Field observations showed that the magnitude 1 events correspond to rockfalls with a mean volume of 50 m^3. A system of two linear equations can be solved in order to measure the cumulative volume of material associated with events of magnitude 2 and 3:

$$577V_2(\mathrm{m}^3) + 4V_3(\mathrm{m}^3) = 1.6 \times 10^6(\mathrm{m}^3) - 50(\mathrm{m}^3) \times 4939 \tag{13.1}$$

$$123V_2(\mathrm{m}^3) + 3V_3(\mathrm{m}^3) = 7.2 \times 10^5(\mathrm{m}^3) - 50(\mathrm{m}^3) \times 219 \tag{13.2}$$

where V_2 and V_3 are the mean volumes of material corresponding to the pyroclastic flows of magnitudes 2 and 3, respectively. From these two equations, a mean volume of 2.0×10^5 m^3 was obtained for events with magnitude 3, and a volume of 1.0×10^3 m^3 as a mean value for the events with magnitude 2.

This method of estimating volumes is approximate but may be useful for eruption monitoring. The values of V_2 and V_3 calculated at Volcán de Colima are in agreement with the estimates of the volume of pyroclastic flows made at Unzen volcano, Japan (0.6×10^4–5.7×10^4 m^3; Uhira et al., 1994; Yamasato, 1997) and

Soufriére Hills volcano, Montserrat (small pyroclastic flows, 10^3-10^4 m^3; medium-sized pyroclastic flows, 10^4-10^6 m^3; Calder et al., 2002). The validity of this relationship was tested when geologists (Muñoz Martínez et al., 2006) estimated the volume of the 17 July 2003 pyroclastic flow at Volcán de Colima. They obtained a volume of 2.4×10^5 m^3. The duration of the seismic signal was 320 s that corresponds to magnitude 3 events, or to a pyroclastic flow with mean volume of 2.0×10^5 m^3, a value very close to the field estimation.

13.5.3 Relationship Between the Amplitude of Long-Period Records of Pyroclastic Flow and Rockfall Earthquakes and the Volume of Pyroclastic Flows

Uhira et al. (1994) showed that the maximum amplitude of long-period seismic records, A_{max}, produced by pyroclastic flows is proportional to the volume of the blocks fallen from the dome, V (Figure 13.18). The seismograms were recorded at a distance of about 4 km from the lava dome.

Figure 13.18 Relationship between the volume of pyroclastic flows and the maximum amplitude of long-period records (vertical component) of the earthquakes produced by pyroclastic flows at Unzen volcano, Japan, in 1991. The seismic station was situated at a distance of about 4 km from the crater.
From Uhira et al., 1994.

The author calculated the least-square regression between these two parameters based on Figure 22 from Uhira et al. (1994):

$$V(10^5 \mathrm{m}^3) = 0.22 A_{\max}(\mu m) + 0.06 \tag{13.3}$$

The correlation coefficient is equal to 0.93; therefore, the maximum amplitude of the long-period signal may also be used for the quantification of pyroclastic flows.

13.6 Location Pyroclastic Flows Using the Amplitude Signals of Earthquakes

The amplitude of the signals recorded by a high dynamic range seismograph network may be used for tracking the trajectory of pyroclastic flows and rockfalls (Yamasato, 1997; Jolly et al., 2002). Figure 13.19 shows typical seismograms and spectrograms for an earthquake recorded at Soufriére Hills volcano, Montserrat, and associated with the propagation of a pyroclastic flow.

The signal observed at one or more seismic stations reflects the onset of the flow at a single location, followed by multiple moving sources spread over a large area. The amplitudes of these sources can be exploited to determine an average location for the event. Jolly et al. (2002) used the spectral amplitudes of the signal at frequency of 7–9 Hz in order to track pyroclastic flow propagation. The seismic record was divided into four segments (Figure 13.19) and locations were calculated for each segment. Figure 13.20 shows locations for a pyroclastic flow on August 12, 1999 at Soufriére Hills volcano. It can be seen that the pyroclastic flow propagated from the dome to NE with runout of about 1.4 km.

The comparison with the video camera record of the pyroclastic flow (Figure 13.21) strongly supports the seismic amplitude-based reconstruction of the movement of pyroclastic flow.

13.7 Seismic Signals Associated with Lahars: Waveforms and Spectra

A lahar, or volcanic mudflow, is a turbulent-muddy debris flow or a hyper-concentrated flood flow on volcanic slopes irrespective of temperature, in which heavily ejected fine ash or pyroclastic materials are the main constituent (see Section 11.1.2). A lahar starts as a debris flow for the first few kilometers, but then evolves into a hyper-concentrated flood flow (Takahashi, 2007).

The duration of seismic records produced by lahars is substantial, up to a few hours (Zobin et al., 2009b). Lahar discharge at Merapi was well correlated with the RSAM (real-time seismic amplitude measurement) amplitudes. Within a single

Figure 13.19 Seismograms (A and C) and spectrograms (B and D) of pyroclastic flow recorded on 12 August 1999 at Soufriére Hills volcano, Montserrat at a distance of about 2 km northeast from the dome (station MBLG), and about 2 km southwest from the dome (station MBBY).
From Jolly et al., 2002.

lahar event, SSAM (spectral seismic amplitude measurement) signals are higher at the flow front than at the flow tail for a given discharge. The best frequency band for monitoring the Merapi lahars ranges from 8.75 to 9.25 Hz (Lavigne et al., 2000b). The observations at Pinatubo showed that the energy from lahars was concentrated within 10−100 Hz (Marcial et al., 1996).

Figure 13.22 shows 1-hour seismic records produced by lahars passing along the ravines La Lumbre (Figure 13.22A) and Montegrande (Figure 13.22B) situated on the flanks of Volcán de Colima. These two ravines differ in their morphology and activity. La Lumbre contains a perennial stream; while Montegrande is filled

Figure 13.20 Amplitude-based locations by a network of five seismic stations for the pyroclastic flow recorded on 12 August 1999 at Soufriére Hills volcano, Montserrat. The event is divided into four time intervals to show propagation through time (see Fig. 13.19). The first segment is marked by circles, the second by crosses, the third by plus signs, and the fourth by stars. The amplitude-based locations suggest that the event initiated on the southern part of the dome and subsequently propagated along the Tar River Valley.
From Jolly et al., 2002.

Figure 13.21 Schematic diagram extracted from video footage recorded on 12 August 1999 at Soufriére Hills volcano, Montserrat, showing pyroclastic flow location through time. The view of the video camera is to the north from the dome.
From Jolly et al., 2002.

during rainy seasons only. La Lumbre is characterized by a more straight drainage system compared to the strongly curved Montegrande ravine. Both ravines are characterized by the vertical sides with widths changing from about 10 m (gradient about 32°) at a distance of about 1.6 km from the crater of Volcán de Colima to about 20 m (gradient about 19-7() at a distance of about 10 km. Then the width of their channels increases from 40 to 100 m (Davila et al., 2007; Abel Cortes,

Figure 13.22 Broadband seismic records (vertical component) of lahars propagating along the La Lumbre (A) and Montegrande (B) ravines on the flanks of Volcán de Colima. The index near seismic record represents the date (yyyy_mmddhhmm) of the events. Courtesy of Colima Volcano Observatory.

personal communication, 2008). The peak flow velocity of some lahars was estimated from video records as about 10 m s^{-1}. They reach about 8−12 km from the volcanic cone. A rain gauge, installed in 2006 near the Montegrande ravine at an altitude of 2,450 m a.s.l., showed that the triggering of lahars occurred if the rain intensity reached 60−120 mm h^{-1}.

Figure 13.22 (*Continued*)

It can be seen that both groups of records have similar waveforms. They represent a sequence of impulses that gradually increase in time, reaching their maximum amplitude, and then gradually, but more slowly, decrease. A detailed analysis of the initial records of lahars is practically impossible due to the large noise that hides the small-amplitude initial impulses produced by the initial stage of a lahar. Nevertheless, it is possible to separate some groups of impulses. For the case of the 2006_07150414 La Lumbre lahar (Figure 13.23A), the first low amplitude impulse has a duration of about 8 min. The onset of the second impulse is sharp; it has larger amplitude than the first impulse. The second impulse gradually decreased over about 40−50 min. The record of the 2005_0506300500 Montegrande lahar

Figure 13.23 Structure of broadband seismic records (vertical component) of the 2006_07150414 La Lumbre lahar (A) and the 2005_0506300500 Montegrande lahar. Arrows with number shows the arrivals of the seismic impulses (see text).
Courtesy of Colima Volcano Observatory.

(Figure 13.23B) consists of at least three impulses with increasing amplitude. The final long impulse 3 is similar to the final long impulse 2 of the La Lumbre event. The presence of sharp, large-amplitude, long-duration final impulses may indicate a change in the nature of lahars during their propagation or a change in their path conditions.

Figure 13.24 shows Fourier spectra calculated for the broadband seismic records of two groups of lahars generated along two ravines, La Lumbre and Montegrande. They are similar; no special features due to different relief of the ravines are seen. Their peak frequencies are between 6.3 and 7.6 Hz.

13.8 Seismic Signals as a Source of Information About the Lahar Structure

Doyle et al. (2010) carried out the multidisciplinary (visual, video, seismic, particle concentration, grain size distribution, rheological properties, and pore pressure)

Figure 13.24 Spectra of the seismic records of lahars propagating along the La Lumbre (thin black lines) and Montegrande (heavy red lines) ravines on the flanks of Volcán de Colima shown in Figure 13.22.
Courtesy of Colima Volcano Observatory.

observations of the 5 March 2008 lahar at Semeru volcano, Indonesia. Their data (Figure 13.25) illustrate a progression from dilute through to hyper-concentrated flow expressed in the seismic signals. The pore-pressure sensors buried in the hard-rock beds, allowed the separation the arrivals of several individual events (lahar packets) which constitute the lahar (Figure 13.25A). The onset of each packet was identified by an emergent and rapid increase in stage and wetted area calculated from pore-pressure stage records (Figure 13.25A, D). The arrivals were also interpreted from video and visual observations, and from sharp increases in the seismic signal (Figure 13.25B), and the approximate spectral energy of Fast Fourier Transform spectra, where the latter is calculated throughout the lahar with 50% overlapping windows (Figure 13.25C). The samples taken to estimate the particle concentration, grain size distribution, and rheological properties, illustrate that the sediment maxima lag the lahar front by approximately 33 min (Figure 13.25D, packet 3).

Each packet was characterized by the front velocity V_f and the sampled particulate concentration. Packet 1 has a sampled particulate concentration downstream of about 26 vol.% and travels at $V_f = 1.5 \pm 0.1$ m s^{-1} between sites, packet 2 has about 48 vol.% and $V_f = 2.9 \pm 0.2$ m s^{-1}, packet 3 has about 60 vol.% and

Figure 13.25 Multidisciplinary observations of the 5 March 2008 lahar at Semeru volcano, Indonesia: (A) 20 s smoothed wetted area; (B) cross-channel component of seismic ground motion, filtered <5 Hz to remove volcanic noise; (C) approximate spectral energy calculated from integrations of 512 point FFT velocity amplitude spectra; (D) downstream site data; (E) Averaged 512 point FFT velocity amplitude spectra for each packet in (B). Arrows in (A) illustrate the equivalent sample times at the 'lava' site. In (D), the 20 s smoothed wetted area (line) and sampled volume concentrations (circles) are shown. Vertical lines indicate packet arrival times.
From Doyle et al., 2010.

$V_f = 4.0 \pm 0.3$ m s^{-1}, and packet 4 has about 40% and $V_f = 1.8 \pm 0.1$ m s^{-1}. The wetted area of the higher concentration packets 2 and 3 increases downstream, from peaks of 6.1 and 6.5 m^2 to 10.3 and 8.9 m^2, respectively (Figure 13.25A and D). A correlation between particle concentration and flow behavior was also observed in the video footage.

Spectral content of the seismic signal is supposed to provide an indication of the particle concentration and rheological behavior of the flow, where laminar high concentration flows are commonly characterized by lower frequencies ($\ll 50$ Hz) than turbulent hyper-concentrated flows (>50 Hz). For this lahar, the largest seismic energies occur between 8 and 20 Hz, with additional peaks between 28 and 32 Hz, 40 and 45 Hz, and the highest at 55 and 60 Hz (Figure 13.25E). There was

little energy in the highest frequencies of packet 1 (Figure 13.25E), as is expected for a turbulent, but dilute, flow. In general, there was an increase in the signal amplitude and energy as the particulate volume concentration increases. However, while packet 3 has a larger wetted area (Figure 13.25A) and a higher concentration than packet 2 (60 vol.% versus 48 vol.%), it induces lower overall seismic energies (Figure 13.25C and E). However, while packet 3's seismic response is less energetic, it still has 62% of its energy above 30 Hz. This broad frequency seismic response suggests that both low frequency frictional and high-frequency collisional processes are occurring in the concentrated phases of these flows.

13.9 Seismic Tracking of Lahars

Kumagai et al. (2009) used the seismic waveforms, recorded by five seismic broadband stations installed around Cotopaxi volcano, Ecuador, to track the source of a lahar that occurred at the volcano on 10 May 2007. They estimated the source of the lahar as temporal changes of the best-fit locations in the residual distributions. The seismic records are shown in Figure 13.26. The 5-s time windows used for locations (between 5075 and 5080 s and 5810 and 5815 s) are shown by two

Figure 13.26 Vertical velocity signals, associated with lahar on the flanks of Cotopaxi volcano, Ecuador, band-passed filtered between 1 and 3 Hz. Time in seconds along the horizontal axis starts from 18:00 hours on 10 May 2007 (UTC). Two vertical lines indicate two 5-s time windows used in Figure 13.27 for lahar location.
From Kumagai et al., 2009.

vertical lines. As a result, the region with small residual, obtained from the first window data, can be seen on the NE flank of the volcano (Figure 13.27A). In the second time window, small residuals were broadly distributed between the northeast and northwest flanks (Figure 13.27B). The field survey team of Instituto Geofísico, which worked in 1 and 2 days after the event, found fresh lahar deposits

Figure 13.27 Spatial distributions of lahar locations in the term of normalized residuals estimated in the time windows (A) 5075–5080 s and (B) 5810–5815 s shown in Figure 13.26.
From Kumagai et al. (2009).

on both the northeast and northwest flanks of the volcano in different river systems that support the seismic locations.

13.10 Comparison of the Seismic Characteristics of Pyroclastic Flows and Lahars

Figure 13.28 represents the seismic characteristics of pyroclastic flow and lahars, recorded at a distance of 4 km from the crater of Volcán de Colima, in terms of the

Figure 13.28 Duration of seismic signal versus spectral peak frequency for lahars and pyroclastic flows. Characteristics of the seismic signals associated with pyroclastic flows at Volcán de Colima, recorded at a distance of 4 km from the crater and produced by the collapse of eruption column, are shown with black circles and those of pyroclastic flows initiated by the partial lava dome collapse are shown with gray circles. Characteristics of the seismic signals associated with Vulcanian explosions and following pyroclastic flows are shown with open stars. Characteristics of pyroclastic flows at Bezymianny volcano, produced by the explosive destruction of growing lava dome and recorded at a distance of 40 km from the crater, is shown with a red star. Characteristics of the seismic signals recorded at a distance of 4 km from the crater of Volcán de Colima and associated with lahars, propagating along the La Lumbre ravine, are shown with gray stars; those of them, propagating along the Montegrande ravine, with black stars.

relationships "duration of seismic signal versus spectral peak frequency." It is clearly seen that the group of characteristics for each type of volcanic flows are separated. The second-to-minute durations of the seismic signals of pyroclastic flow significantly differ from the 10-minute-to-hours durations of the seismic signals of lahars. The peak frequencies of Fourier spectra of pyroclastic flows are significantly lower than the same of lahars.

This observation allows a simple recognition of these events in real time. A spindle-shape signal, completely recorded on a seismogram during the first 3–5 min, would indicate a pyroclastic flow but if a signal gradually increases in amplitude during the same time interval, it may refer to the beginning of lahar movement. It may be used for early diagnostics of the type of volcanic flow in real time.

14 Seismic Signals Associated with Volcanic Explosions

The explosive events occurring during magmatic, phreatic or phreato-magmatic eruptions generate seismic signals. As noted in Chapter 1, the main types of volcanic magmatic explosions are Strombolian (VEI 1–2), Vulcanian (VEI 2–4), Plinian (VEI 4–8), and Ultra-Plinian (VEI 5–8). Strombolian explosions are short (seconds to minutes), discrete, and produce eruptive columns or lava fountains; they are characteristic for basaltic volcanoes and occur at quasi-regular intervals producing three to five explosions per hour with the eruption columns of tens or hundreds of meters in height (Vergniolle and Mangan, 2000). Vulcanian explosions are also short (seconds to minutes) discrete explosions with a small amount (less than 1 km^3) of erupted products but they are characterized by high eruption columns, which can reach 10–20 km altitude, and occupy the intermediate position in the explosion hierarchy between low-rise Strombolian and high-rise Plinian explosions whose eruption columns may reach up to 45 km altitude (Francis, 1998; Morrissey and Mastin, 2000; Schmincke, 2004). Vulcanian and Plinian explosions are characteristic for andesitic and dacitic volcanoes. They may be violent, sometimes destroying a part of the volcano edifice and producing highly fragmented ash (Francis, 1998; Morrissey and Mastin, 2000).

A phreato-magmatic explosive eruption is driven primarily by the volumetric expansion of external water after it has been rapidly heated by contact with magma. The most commonly recognized styles of explosive events occurring during the phreato-magmatic eruptions are the explosive jets of ash accompanied by billowing clouds of steam (so-called Surtseyan explosion) or Vulcanian-type explosions (Morrissey et al., 2000).

Explosion earthquakes may occur as a single event or as an earthquake sequence. Tokarev and Firstov (1967) studied seismic activity of Karymsky volcano, Kamchatka, in 1965, when small explosions occurred, accompanied by a slow emission of dacitic lava. During 3 weeks of observation, 727 explosion earthquakes were recorded. During eruptive activity within a summit lava lake at Mount Erebus, Antarctica, the common eruption styles ranged from passive degassing to Strombolian explosions. They typically occurred several times daily, and occasionally in swarms of up to 900 per day (Rowe et al., 2000). Short-period 24-h seismogram, recorded at Volcán de Colima, México, during the 2007 lava dome growth (Figure 14.1), illustrates the numerous seismic signals of explosive degassing.

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00014-1
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Figure 14.1 Short-period 24-hour seismogram (velocity, vertical component) recorded during the 2007 lava dome growth at Volcán de Colima, México, and illustrating the occurrence of the seismic signals generated by small explosions. Seismic station was situated at a distance of 1.7 km from the crater.
Courtesy of Colima Volcano Observatory.

14.1 Waveforms and Spectra

The seismic records of explosion earthquakes are characterized by a great variety. At the same time, they may form the families of similar waveforms (Johnson et al., 1998; Rowe et al., 2000) depending on the type and style of explosive activity. The seismic records of explosive events may be complicated by following signals of pyroclastic flows (see Chapter 13) or acoustic signals (discussed in Chapter 17).

14.1.1 Strombolian Explosions

Figure 14.2 shows a typical seismogram of Strombolian explosion. It was recorded during the 1992 activity at Stromboli volcano at a distance of 150 m from the crater. The three components of ground velocity show a distinctive preliminary low-frequency (LF) signal with dominant frequency near 2 Hz and the main phase of higher-frequency with frequencies up to 10 Hz starting roughly 1.4 s after the onset of the event (Chouet et al., 1997). Martini et al. (2007) described the similar seismic signals associated with 2007 Strombolian explosions at Stromboli volcano. They consider that the two components of the seismic radiation, noted by Chouet et al. (1997), are linked to two different source processes during an explosion (see also Section 3.2.2, Figure 3.6).

Figure 14.2 Three-component short-period velocity record of the seismic signals associated with Strombolian explosion at Stromboli volcano, Italy (7 May 1992, 20:27). Seismic station was situated at a distance of 0.15 km from the crater. LF and main phase indicate preliminary low-frequency phase and higher-frequency main phase of the seismic signal. After Chouet et al., 1997.

14.1.2 Vulcanian Explosions

Figures 14.3 and 14.4 show four typical broadband seismic waveforms (three components) associated with Vulcanian explosions occurring at Volcán de Colima, México, and Fourier spectra of the vertical components. All these waveforms are characterized by the existence of large-amplitude, high-frequency signal of the main phase. The main phase is preceded by low-frequency signal LF (Type 1), or high-frequency signal HF (Type 2), or by both high-frequency + low-frequency signals (Type 3). These preceding phases are better recognized on vertical components but may not be so well expressed on horizontal components. Seismic signal associated with an explosion may be recorded without any preceding signal also (Type 4). Domination of the main phase on the seismic records of explosive events constrains the peak frequency between 1 and 2 Hz for all these signals (Figure 14.5).

14.1.3 Phreato-Magmatic Explosions

Figure 14.6 shows seismic waveforms of phreato-magmatic explosion recorded during the lateral 2000 eruption at Usu volcano (see Chapter 7) and their Fourier spectra. Broadband waveform (Figure 14.6B), recorded at a distance of 5 km from the crater, is characterized by a sequence of high-frequency and low-frequency pulses supposedly associated with the sequence of frequent explosive

Figure 14.3 Video images (A and C) and three-component broadband seismic velocity records of Types 1 and 2 (B and D) associated with Vulcanian explosions at Volcán de Colima during the 2003–2007 activity. Seismic station was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 15 km south from the crater. LF and HF indicate preliminary low-frequency and high-frequency signals. Courtesy of Colima Volcano Observatory.

events during a phreato-magmatic eruption. The same groups of pulses but of higher-frequency content are observed on short-period seismogram (Figure 14.6C) recorded at a distance of 21 km from the crater. The main phase, similar to that observed at the seismic records of Strombolian and Vulcanian explosions, is absent. Fourier spectra of these seismic records (Figure 14.6D) are very similar for the interval between 0.1 and 5 Hz and represent a stable amplitude level within this spectral interval.

Similar broadband waveforms were reported for phreatic explosions at Aso volcano, Japan, during the Aso94 experiment (Kawakatsu et al., 2000).

Figure 14.4 Video images (A and C) and three-component broadband seismic velocity records of Types 3 and 4 (B and D) associated with Vulcanian explosions at Volcán de Colima during the 2004–2007 activity. Seismic station was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 15 km south from the crater. LF and HF indicate preliminary low-frequency and high-frequency signals. Courtesy of Colima Volcano Observatory.

14.2 Nature of the Seismic Signals of Explosive Earthquakes

14.2.1 Comparison of the Contemporary Video and Seismic Records During an Explosion

Strombolian Explosions

Figure 14.7 shows three-component broadband seismic records of an explosion that occurred at Stromboli volcano in January 2011, together with the contemporary

Figure 14.5 Fourier spectra of four types of the seismic signals (vertical component) associated with Vulcanian explosions at Volcán de Colima during the 2003–2007 activity and shown in Figures 14.3 and 14.4.

video record. It is seen that the preliminary low-frequency pulse (better seen in horizontal components, Figure 14.7C and D) is recorded before the outlet of the volcanic products from the crater and supposedly corresponds to the process occurring within the conduit before the explosion. The high-frequency and large-amplitude signals of the main phase, which arrived 5 s later, were recorded just before and during the outlet of volcanic products from the crater.

Vulcanian Explosions

Figure 14.8 shows the contemporary records of broadband seismic signals and video images of Vulcanian explosion (Type 3) at Volcán de Colima, México. Type 3 seismic signals associated with Vulcanian explosions were selected because contains both preliminary phases, HF and LF. It is seen that the first superficial manifestation of explosion, marked by an arrow on the image 2, appeared 26 s after the onset of the seismic signal. Therefore, the initial part of the seismic record, containing

Figure 14.6 Video image (A), broadband seismic record (velocity, vertical component) (B) and short-period seismic record (velocity, vertical component) (C) associated with the 31 March 2000 (13:08) phreato-magmatic explosion at Usu volcano (lateral cone Nishi-yama). Broadband seismic station was situated at a distance of 5 km from the crater; short-period station, at a distance of 21 km from the crater. (C) Fourier spectra of seismic records of this explosion recorded at a distances of 5 km (broadband, solid line) and 21 km (short-period, broken line). LF and HF indicate low-frequency and high-frequency signals.
Video image was taken from JMA (2003). Broadband seismic record is a courtesy of Tohoku University; short-period seismic record is a courtesy of Japan Meteorological Survey.

Figure 14.7 Contemporary video and three-component broaband velocity seismic records during Strombolian explosion (13 February 2011, 02:57) at Stromboli volcano. Seismic station was situated at a distance of 0.5 km from the crater. Video station was situated at a distance of about 1 km from the crater. LF and HF indicate low-frequency and high-frequency signals. 1, 2, and 3 correspond to the timing of images and the seismic record. Courtesy of Istituto Nazionale di Geofisica e Vulcanología, Sezione di Catania.

the preliminary phases LF and HF, may be considered to be generated by the process occurring within the volcano conduit before an explosion.

14.2.2 Type of Waves Composing the Seismic Signal of an Explosion

Strombolian Explosions

Figure 14.9 shows the particle motions for the preliminary low-frequency signal of the seismic record associated with the explosion at Stromboli volcano (see

Figure 14.8 Contemporary video and broadband velocity seismic records (vertical component) during Vulcanian explosion at Volcán de Colima (22 February 2007, 16:32). Seismic station was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 15 km south from the crater. LF and HF indicate low-frequency and high-frequency signals. 1, 2, and 3 correspond to the timing of images and the seismic record. Courtesy of Colima Volcano Observatory.

Figure 14.6). To avoid the influence of high-frequency noise, the records of vertical and radial components were low-pass filtered at 0.5 Hz. It displays a retrograde elliptical particle motion indicating Rayleigh wave. The same retrograde elliptical particle motion was shown by Rowe et al. (1998) for the initial 15-s signals of broadband seismic records of explosive events at the lava lake of Erebus volcano.

Vulcanian Explosions

Figure 14.10 shows the particle motions for the preliminary low-frequency signal of the seismic record associated with the explosion at Volcán de Colima. To avoid the influence of high-frequency noise, the records of vertical and radial components were low-pass filtered at 0.5 Hz. The record is short, but it is possible to see that the

Figure 14.9 Particle motions of preliminary low-frequency signal of seismic records associated with Strombolian volcanic explosion shown in Figure 14.7. The seismic records ((A) vertical component and (B) radial component) were low-pass filtered at 0.5 Hz. (C) the trajectory of retrograde particle motion.
The seismic records are courtesy of Istituto Nazionale di Geofisica e Vulcanología, Sezione di Catania.

retrograde motion of particles forms well-expressed ellipses in the vertical plane that is characteristic for Rayleigh waves (Zobin et al., 2006). Tameguri et al. (2002) studied the particle motions of Sakurajima explosions derived from broadband records. They also obtained that low-frequency phase LP (equivalent to LF in Figure 14.4) shows an elliptical orbit and a retrograde rotation in a vertical cross-section along the direction of wave propagation. The same result about the Rayleigh nature of the preliminary LF phase was obtained by Zobin and Martínez (2010) for the seismic records associated with Vulcanian explosions of Popocatépetl volcano.

Tameguri et al. (2002) studied the wave nature of preliminary HF signals recorded by short-period borehole seismometers during the 1999 Sakurajima Vulcanian explosions and concluded that they are *P*-waves.

Figure 14.10 Particle motions in the seismic record of preliminary low-frequency signal (LF) of earthquake associated with Vulcanian explosion at Volcán de Colima (27 January 2005, 13:39). (A) The seismogram (velocity, vertical component); the LF signal is marked by arrows. (B and C) The vertical and radial components of the LF impulse were low-pass filtered at 0.5 Hz. (D) the trajectory of retrograde particle motion. Seismic station was situated at a distance of 4 km from the crater.
From Zobin et al., 2006b.

14.3 Sources of Explosion Earthquakes and Their Quantification

The complex structure of the seismic signals associated with volcanic explosions as well as the problem of identification of the nature of waveforms, composing these seismic signals, and unclear arrivals of the seismic phases make difficult the study

of the source process of explosion earthquakes. To solve the problem, many research scientists study the source of the explosion event as a source that generates long-period and very-long-period seismic signals neglecting short-period components. They band-pass filter the original broadband seismic records, leaving only long-period and very-long-period vibrations (as is shown in Figure 14.11), and then perform the waveform inversion obtaining the characteristics of a centroid as the source of these vibrations (e.g., Arciniega-Ceballos et al., 1999; Chouet et al., 2003, 2005). This approximation gives the integral characteristics of the source of explosion events, neglecting the details of the process.

In this chapter, the author shows the results that were obtained using the fine structure of the seismic signal as the information about the development of ascending magmatic process within the conduit before an explosion and during the explosion

Figure 14.11 Broadband seismic record (velocity, vertical component) and video image (A) of Vulcanian explosion at Volcán de Colima (23 March 2005, 03:41). (B) Unfiltered record; (C) the same record but band-filtered at 2−30 s. Seismic station was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 5 km north from the crater. LF indicates preliminary low-frequency signal.
Courtesy of Colima Volcano Observatory.

(e.g., Tameguri et al., 2002; Zobin et al., 2006b, 2009). The results that were obtained using long-period and very-long-period approximations of the seismic signals associated with volcanic explosions will be discussed in Chapter 15 together with other applications of long-period signals generated by volcanic processes.

Tameguri et al. (2002) discriminated four phases of the seismic records of Vulcanian explosions at andesitic Sakurajima volcano and estimated the moment tensors for the sources generating these phases. Zobin et al. (2006, 2009) developed a two-stage model of explosive process for explosions at Volcán de Colima and Popocatépetl that allows quantification of the volcanic explosions in the units of force (Newtons) and energy (Joules) from unfiltered broadband seismic records (see Chapter 11). Kumagai et al. (2011) proposed the ascending seismic source for the explosion at Tungurahua volcano, Ecuador, considering that a pressure wave propagates along the magma conduit triggering fragmentation of magma at shallow depths.

14.3.1 Multiple Source of Explosions

Tameguri et al. (2002) discussed the nature of broadband seismic records of Vulcanian explosions at Sakurajima volcano occurring during August–December 1999. Figure 14.12 shows broadband seismic record of the 1999 Sakurajima explosion obtained on the seismic station situated at a distance of 2.75 km. This seismic record corresponds to Type 3 of Vulcanian events shown in Section 14.1.2. Tameguri et al. (2002) analyzed three phases: initial phase *P* and the second phase *D* within the first group of short-period pulses (or phase HF in Figure 14.4), and the phase LP corresponding the group of long-period pulses (or phase LF in Figure 14.4). The phases *P* and *D* were identified as *P*-waves, and LP phase, as composed of Rayleigh waves.

Moment tensors of the seismic sources, generated by these phases, were estimated from waveform inversions for 14 explosions. The source of phase *P* was characterized by three diagonal components with the same positive values for all events. It corresponds to the isotropic expansion source (Figure 14.13). The sources

Figure 14.12 Broadband seismic signal (velocity, vertical component), associated with Vulcanian explosion at Sakurajima volcano, Japan, and recorded at a distance of 2.8 km from the crater. P, D, LP, see in text.
After Tameguri et al., 2002.

Figure 14.13 Broadband waveform inversion of the *P* and *D* phases. The explosion earthquake occurred on 15 September 1999, 04:14. Solid and broken traces indicate observed and synthetic waveforms, respectively. *V*, *R*, and *T* indicate vertical, radial, and tangential components. Two triangular source time functions 1 and 2 corresponding to *P* phase and following *D* phase are displayed in the upper left and obtained moment tensor components are shown in the right side. Bars with symbols "*P*" and "*D*" under the waveforms represent time windows of waveform inversion. The observed and synthetic waveforms were low-pass filtered at 5 Hz.
From Tameguri et al., 2002.

of phase *D* (Figure 14.12) were characterized by horizontal dipoles twice large than the vertical that is in accordance with the theoretical ratio among the three diagonal components caused by a volume change of cylindrical source (2:2:1). The sources generated *P* and *D* phases were located at the depth of 2 km beneath the crater. LP phase was excited by two sources: the first source with an isotropic expansion and the following horizontal contraction source at shallow depths of

Figure 14.14 Broadband waveform inversion of preliminary LP phase. The explosion earthquake occurred on 27 August 1999, 19:21. Solid and broken traces indicate observed and synthetic waveforms, respectively. *V*, *R*, and *T* indicate vertical, radial, and tangential components. Two triangular source time functions 3 and 4 corresponding to LP1 phase and following LP2 phase are displayed in the upper left and obtained moment tensor components are shown in the right side. Bars with symbols "LP1" and "LP2" under the waveforms represent time windows of waveform inversion. The observed and synthetic waveforms were low-pass filtered at 1 Hz.
From Tameguri et al., 2002.

0.25−0.5 km beneath the crater bottom (Figure 14.14). The values of seismic moments obtained for the sources varied from 10^{11} to 10^{12} N m.

Tameguri et al. (2002) found that the origin times of LP phases coincided with those of the air-waves associated with the explosions. More recent studies carried out by Yokoo et al. (2009) showed that the origin times of LP phases of the Sakurajima explosions coincided with the preceding low-amplitude phase of

Figure 14.15 Time sequence of origin times and depths of four sources generating *P*, *D*, and LP phases of explosion earthquake. Solid and broken lines indicate source time functions of horizontal and vertical dipoles, respectively. Mechanics of explosive eruption from seismic data is schematically shown in lower part. (A) An isotropic expansion occurs at a depth of 2 km beneath the crater. Then, a pressure wave generates and propagates through the conduit up to the crater. (B) A cylindrical contraction is generated by pressure decrease of source region due to escape gases upward in the conduit. (C) Outbreak of gas pocket beneath a lava dome is started by shallow expansion, which is excited by a pressure wave propagating from the deep source. At the same time, air-shock is ejected from the crater bottom. (D) Gases are released from the uppermost part of conduit. Horizontal contraction starts 1− 2 s after the beginning expansion.
From Tameguri et al. (2002).

the infrasonic wave. Yokoo et al. (2009) showed that the origin of impulsive compression phase of the infrasonic wave coincided with the peak time of the moment velocity reconstructed for the LP phase. Therefore, the explosion occurs after the beginning of the radiation of LP phase.

Based on the conceptual model of the Sakurajima explosions proposed by Ishihara (1985, 1990), the sequence of seismic sources was supposed (Tameguri et al., 2002) as corresponding to the following eruptive process (Figure 14.15). The

gas pocket, created by the exsolution of volatiles in the upper part of shallow magma chamber at the depth of about 2 km (phases *P* and *D*), abruptly moves up through the conduit to the surface generating the phase LP. Then the explosion, accompanying the outflow of gas, ash, and volcanic blocks, occurs (the coda of LP signal and the second high-frequency group of the seismic record).

14.3.2 Two-Stage Model of Explosive Process

The two-stage model, proposed by Zobin et al. (2006, 2009b) for Type 1 Vulcanian explosions (see Section 11.3.2), describes the volcanic explosion process as consisting of the ascent of fragmented magma to the surface, triggering an explosion (stage 1), and a subsequent explosion (stage 2). The model may be applied also to Type 2 and 3 explosions (Figures 14.3 and 14.4), considering as a trigger of an explosion the fracturing of lava body within the conduit with generation of high-frequency signal (Tuffen et al., 2008; Zobin et al., 2010; see also Section 3.2.1). Based on this model, the seismic record, associated with an explosion, may be considered as consisting of two phases: pre-explosion (LF, HF, or LF + HF; Figures 14.3 and 14.4) and co-explosion (main phase). For all of these seismic records, as shown in Section 3.2.1, the counter force may be estimated from the modeling of LF pre-explosion phase and the energy of explosion may be calculated from the main co-explosion phase.

The source parameters, obtained for 120 Colima explosions and 26 Popocatépetl explosions (Zobin et al., 2009b), varied over the following ranges: the time D_1 of the movement of fragmented magma in the conduit before the explosion varied from 1 to 13 s; the counter force of the eruption F acting in the conduit before an explosion and governing the fragmented magma movement to the surface spanned three orders of magnitude ranging from 8.3×10^7 to 8.5×10^{10} N; the duration of explosion in the conduit D_2 varied from 11 to 65 s; and the energy E of explosion spanned six orders of magnitude ranging from 4.6×10^7 to 1.5×10^{13} J. The scaling relationships of these parameters are discussed below.

Counter Force versus Energy

The regression between the energy of explosions E and the counter force of the eruption F obtained for the explosions occurring on Volcán de Colima and Popocatépetl (Zobin et al., 2009b)

$$\text{Log}\, E = 1.87\, \text{Log}\, F - 7.29 \tag{14.1}$$

shows (Figure 14.16) a strong dependence of the energy of the explosion on the counter force (coefficient of correlation $R = 0.87$). In this case, both groups of large and small explosions demonstrate the same scaling of the magnitudes of the force controlling the magma movement before the explosion and the energy of the explosion.

Figure 14.16 Relationship between the counter force of the eruption and the energy of the explosion for the Vulcanian explosions of Volcán de Colima (circles) and Popocatépetl (black diamonds).
From Zobin et al., 2009b.

Counter Force versus Source Duration

The first scaling law for counter force of explosion earthquakes was proposed by Nishimura and Hamaguchi (1993). They obtained the relationship between the peak amplitude of the single counter force F and the source duration τ of the single force using the seismic records of large volcanic explosion of the world. The obtained forces varied from 2.6×10^{12} N for the great 18 May 1980 explosion at Mount St. Helens, Cascades (Kanamori et al., 1984) to 10^7 N for small explosions at Mount Tokachi, Japan (Nishimura, 1995). For large events ($F > 10^9$ N), the empirical law

$$\text{Log } F(N) = 2 \text{ Log } \tau(s) + 9.6 \tag{14.2}$$

was obtained (Nishimura and Hamaguchi, 1993) (Figure 14.17A). For smaller events, the dependence of the peak force upon the source duration is too weak; τ was almost a constant value. Nishimura (1995) considered that this difference

Figure 14.17 Relationship between the counter force of the eruption and the source duration of a single force. (A) Scaling law obtained by Nishimura and Hamaguchi (1993) for different volcanoes of the world; (B) relationship for the explosion earthquakes of Volcán de Colima (circles) and Popocatépetl (black diamonds). The regression lines are shown separately for large and small explosions. Dashed line with index *N−H Law* shows the scaling law proposed by Nishimura and Hamaguchi (1993).
After Nishimura and Hamaguchi, 1993 and Zobin et al., 2009b.

was caused by the difference in a "critical magma radius" of the hypothetical spherical gas-rich magma batch that produced the volcanic explosion.

The results of Zobin et al. (2009b), obtained for the Volcán de Colima and Popocatépetl explosions, support the scaling law of Nishimura and Hamaguchi having the same tendency of $F \sim \tau^2$. Two types of scaling were observed again (Figure 14.17B). The first cluster joins the large events with the counter force greater than 4×10^9 J; the second group consists of smaller events. The large events of both

Colima and Popocatépetl explosions were correlated with the source duration with the coefficient of correlation $R = 0.64$ ($R_{crit99\%} = 0.44$) and their regression equation was

$$\text{Log } F = 1.59 \text{ Log } \tau + 9.90 \tag{14.3}$$

The group of small events in Zobin et al. (2009b) was more numerous than the group of small events shown by Nishimura (1995) with the greater difference in τ. It allowed Zobin et al. (2009b) to calculate the regression line for small events also

$$\text{Log } F = 1.77 \text{ Log } \tau + 8.31 \tag{14.4}$$

with the coefficient of correlation $R = 0.60$ ($R_{crit99\%} = 0.25$).

Size Parameters of Explosions Versus the Durations of Two Explosion Stages

The durations of two explosion stages, including the fragmented magma movement in the conduit and the explosive process, in the framework of the conceptual model of Zobin et al. (2009b), are described by the parameters of D_1 and D_2, respectively.

The time of fragmented magma movement D_1 does not depend on the counter force of the eruption F (Figure 14.18A, coefficient of correlation $R = 0.08$). Similarly, the energy of the explosion is not correlated with the time of the fragmented magma movement D_1 (Figure 14.18B, coefficient of correlation $R = 0.06$). At the same time, Figure 14.18B shows the presence of two groups of explosions, large and small, separated at the energy level of 1×10^{11} J. The high-energy explosions ($E > 10^{11}$ J) occurred when the D_1 values range between 3.5 and 7.5 s while the low-energy explosions ($E < 10^{11}$ J) occurred without any dependence on the value of D_1 and ranged between 0 and 13 s.

The duration of the explosion, D_2, has a negative correlation with the size of a counter force F (Figure 14.19A; coefficient of correlation $R = -0.49$) and with the energy of the explosion E (Figure 14.19B; coefficient of correlation $R = -0.39$):

$$\text{Log } D_2 = -0.14 \text{ Log } F + 2.89 \tag{14.5}$$

$$\text{Log } D_2 = -0.07 \text{ Log } E + 2.20. \tag{14.6}$$

These regressions are characterized by low coefficients of correlation but they are significant at the 99% confidence level ($R_{crit99\%} = -0.22$). The duration of the magma movement D_1 does not correlate with the duration of an explosion D_2 (coefficient of correlation $R = 0.18$).

14.3.3 Relationship Between the Seismic Moment of Preliminary LP Phase and the Energy of Explosions for Sakurajima Volcano

Broadband seismic records of 14 explosions of Sakurajima volcano, discussed in Section 14.3.1, were used to obtain the relationship between the seismic moment

Figure 14.18 Relationship between the counter force of the eruption and the duration of the pre-explosion impulse (A) and the energy of the explosion and the duration of the pre-explosion impulse (B) for the Vulcanian explosions of Volcán de Colima (circles) and Popocatépetl (black diamonds). The dashed line that divides large and small explosions in (B) is shown.
From Zobin et al., 2009b.

Figure 14.19 Relationship between the counter force of the eruption and the duration of the preliminary pulse (A) and the energy of the explosion and the duration of the preliminary pulse (B) for the Vulcanian explosions of Volcán de Colima (circles) and Popocatépetl (black diamonds). The regression lines are shown.
From Zobin et al., 2009b.

Figure 14.20 Relationship between the components M_{zz} of moment tensors of Sakurajima explosions estimated in Tameguri et al. (2002) and the energy of the same events. The small explosions are shown within the rectangle. These parameters are correlated with the coefficient of correlation $R = 0.81$. The correlation of these parameters for micro-explosions ($E < 10^8$ J) is absent.
The seismic record is courtesy of Sakurajima Volcano Research Center.

component M_{zz} of preliminary LP phase estimated by Tameguri et al. (2002) and the energy of explosions estimated according to methodology by Zobin et al. (2006). The seismic records of station KUR situated at a distance of 4.1 km from the crater were used for the estimation of the energy. Figure 14.20A shows the typical seismic record of the Sakurajima explosion at station KUR and its main phase signal used for energy calculation. The relationship M_{zz} versus E is shown in Figure 14.20B. The components M_{zz} of moment tensor, estimated by the inversion of LP phase, are well correlated with the energy of small ($E > 10^8$ J) explosions of Sakurajima. It allows obtaining the following equation relating these two parameters:

$$\text{Log } M_{zz} = 0.38 \text{ Log } E + 8.4 \tag{14.7}$$

For micro-explosions with energy $E < 10^8$ J, the correlation between these two parameters was not observed.

14.3.4 Volcanic Explosive Process as a Source of Hybrid Earthquakes

According to the definition of hybrid earthquakes, observed during the 1989–1990 eruption of andesitic Redoubt volcano, Alaska (Power et al., 1994), the seismic signals of this type have a high-frequency onset and an extending low-frequency coda. Lahr et al. (1994) suggested the hybrid events may result from brittle faulting in zones of weakness intersecting a fluid-filled crack and thus involve both double-couple and volumetric source components. Benson et al. (2010) reported (see Chapter 3) that hybrid seismic events are likely to be produced by the dual process of crack nucleation and deformation producing the high-frequency onset and, once these fluid pathways are created, fluid moving through the damage/crack network producing low-frequency resonance is seen in the coda of the waveforms.

Simultaneous seismic and video observations at Volcán de Colima showed that the seismic signals with high-frequency onset and an extending low-frequency coda were recorded during some of Vulcanian eruptions (Type 2, see Figure 14.3). This observation allows the suggestion that the initial high-frequency signal is generated by a brittle fracturing in hot magma rocks and serves as a trigger for the following explosive (volumetric) event. Based on this point of view, the seismic records of the Type 3 Vulcanian explosions (see Figure 14.4) are supposed to be generated by the complex process including the brittle fracturing in hot magma rocks (HF signal), triggering the ascent of magma to the surface (LF signal) and the subsequent explosion. The events generating seismic records of Type 2 are thought to be shallower than the Type 3 events.

14.4 Location of Explosion Earthquakes

As seen in Section 14.1, the seismic records of explosive events are rather complicated and their source may be considered as a sequence of sub-events. Therefore, it is important to locate these sub-events, at least the position of the initial seismic source, preceding an explosion, and the position of the source of the explosion. The absence of good-expressed body waves in the seismic signal makes impossible the traditional hypocenter determination using onset arrival times.

14.4.1 Location of the Initial Sub-Events from Waveform Inversion

The waveform inversion, performed for the whole seismic record, gives the position of a centroid corresponding to the integrated source and cannot be applied

Figure 14.21 Locations of broadband seismic and infrasonic stations (triangles) at Tungurahua volcano, Ecuador, and contour plots of horizontal and vertical distributions of the normalized residuals obtained by waveform inversion of an explosion event on 11 February 2010. Observed (black lines) and synthetic (red lines) particle motions obtained from displacement seismograms are also shown. Dots indicate the nodes used in the grid search to estimate the source location. (B) Source locations of the explosion event estimated by using vertical seismic amplitudes in a frequency band of 5–10 Hz. The circles showing source locations are scaled by the initial amplitudes. The horizontal dashed line in B (top) is the horizontal location of the vertical profiles shown in A (bottom) and B (bottom). From Kumagai et al., 2011.

here. Therefore, only the results obtained for the initial phases of the seismic signal are discussed. Kumagai et al. (2011) performed the source location of the 11 February 2010 explosive event occurring at Tungurahua volcano, Ecuador, using waveform inversion for the initial part (about 5–7 s) of broadband seismic records. They obtained the source position at a depth of about 5 km beneath the summit (Figure 14.21) attributing it to the position of the initial ascending seismic source before eruption.

14.5 Explosion Sequences

The seismic records of explosion earthquakes form the families of similar waveforms (Johnson et al., 1998; Rowe et al., 2000). Strong repeatability of the gross waveform characteristics in the records of lava lake Strombolian explosions at

Figure 14.22 Illustration of the similarity of seismic records for explosion earthquakes. Seventeen lava lake Strombolian explosions of Erebus volcano, Antarctica produced these short-period seismograms of explosion earthquakes that were recorded at a distance of 10 km from the lava lake. Traces have been band-pass filtered from 1 to 7 Hz to reduce incoherent signals.
From Rowe et al., 2000.

Erebus volcano (Figure 14.22) is evident among events up to 20 s into the coda, demonstrating a repetitive source process.

Size–frequency relationship for explosion earthquakes sequences differs from the lineal regression common for volcano-tectonic earthquakes. Figure 14.23 shows the size (maximum ground velocity) distribution of 364 explosions of Erebus volcano, Antarctica (Rowe et al., 2000). Note that the sharp bend in the cumulative curve of Figure 14.23B corresponds to the peak in the histogram of Figure 14.22A. For 727 Karymsky explosion earthquakes, the magnitude (energy)–frequency relation had a parabolic form (Figure 14.24). Tokarev and Firstov (1967) consider that the maximum in Figure 14.24 corresponds to the optimal energy for explosions in the crater, according to the properties of the magmatic agents and conduit

Figure 14.23 Size (maximum ground velocity) distribution of 364 explosions of Erebus volcano, Antarctica recorded between October 1997 and July 1998 at a distance of 10 km from the lava lake. (A) Histogram of explosion counts versus $\log_{10}$ of seismic amplitude, representative of explosion seismic magnitude. (B) $\log_{10}$ of monthly cumulative counts (for each size bin and higher) versus $\log_{10}$ seismic amplitude.
From Rowe et al., 2000.

dimensions ("critical magma radius"; Nishimura, 1995); all remaining events represent the deviations from this optimal case.

The size–frequency distributions for large and small explosions, recorded at Volcán de Colima and defined in Section 14.3.2, display the difference in their source properties. Figure 14.25 shows the energy–frequency distributions obtained for two databases: (1) 15 large explosions ($E > 10^{11}$ J) occurring during the March–September 2005 explosive sequence at Volcán de Colima and (2) 236 small explosions ($E < 10^{11}$ J) of the November 2004 sequence at the same volcano. Application of the Kolmogorov–Smirnov test (Figure 14.25B), considering as a null hypothesis that the two databases, n_1 and n_2, belong to the same sample, demonstrated that the null hypothesis had to be rejected and these two databases belong to different samples. To reject the null hypothesis at the significance level α, $D_{n1,n2}$ had to be greater than the critical value equal to $k_{1-a}\sqrt{(n_1+n_2)/n_1n_2}$, where $k_{1-\alpha}$ is the quantile of the Kolmogorov–Smirnov distribution at the significance level α (Muller et al., 1979). The maximum difference between two cumulative distributions, $D_{n1,n2}$, was 1.00, which is larger than the 0.01-critical value of 0.4334. Therefore, the large and small explosions at Volcán de Colima may be considered as having the different nature of generation.

Figure 14.24 Magnitude (energy in J)—frequency dependence for explosion earthquakes of Karymsky volcano.
Data from Tokarev and Firstov, 1967.

14.6 Explosion Earthquakes in Eruptive Process

Figure 14.26A shows the 10-year variations in the number of explosion earthquakes at Karymsky volcano. The mean rate of lava emission during each eruptive episode is shown in Figure 14.26B. It is seen that the episodes with lava emission are characterized by an increase in the frequency of explosive earthquakes. At the same time, no dependence between the volume of lava emission and the maximum number of seismic events was observed. These earthquakes were also observed without lava emissions (Khrenov et al., 1982).

Tanaka et al. (1972) studied the explosion earthquakes that occurred during the 1970—1971 eruption of Akita-Komaga-take volcano, Japan. The eruption began with a sequence of Strombolian explosions and basaltic lava emission on 18 September 1970 and continued until 25 January 1971. About 33,000 explosion earthquakes were recorded. They were the main seismic signals for this eruption. Single volcano-tectonic earthquakes (Figure 14.27B) were innumerous. Figure 14.27 shows the variation in the number of explosion earthquakes that were observed from the very beginning of the eruption. These earthquakes were directly

Figure 14.25 Distribution in the energy for large explosions of the March–September 2005 sequence (1) and small explosions of the November 2004 sequence (2) at Volcán de Colima (A) and the Kolmogorov–Smirnov test comparison cumulative fraction plot for the sequences 1 and 2 (B). The statistic *D* is the maximum vertical deviation between two curves.

connected with volcanic explosive activity and may serve for monitoring of this type of eruption.

Figure 14.28 shows the effusive–explosive activity of Volcán de Colima during 2002–2006. During this period, two episodes of lava flow activity were recorded, from February 2002 to February 2003 and from September to December 2004. The increase in the number of small explosions, recorded at a distance of 1.7 km from the crater by short-period seismic station, was observed during these two episodes.

Figure 14.26 Variations in the monthly number of explosion earthquakes (A) and in the mean rate of lava emission (in kg/s) during eruptive episodes of Karymsky volcano, Kamchatka, during 1970–1980.
From Khrenov et al., 1982.

Figure 14.27 Daily number of explosion earthquakes (A) and volcano-tectonic earthquakes with $S-P$ times less than 3.0 s (B) recorded during the eruption of Akita-Komaga-take volcano, Japan, at the seismic station situated at a distance of about 4 km from the crater. Open circles (curve 2) represent the number of explosion earthquakes with maximum amplitude larger than 200 μkine.
From Tanaka et al., 1972.

Figure 14.28 Development of effusive–explosive activity of Volcán de Colima during 2002–2006. Variations in the daily number of small explosions are shown together with the intervals of lava extrusions (EFF) and large Vulcanian explosions (LE).

Figure 14.29 The energy of Vulcanian explosions observed at Volcán de Colima during 2004–2005 (shown as bars) and the appearance of pyroclastic flows accompanying largest of them (with energy $> 10^{11}$ J).

Large explosions (energy greater than 10^{11} J) occurred when the effusive activity was terminated and the activity of small explosions significantly decreased.

Zobin et al. (2006) showed that the explosions at Volcán de Colima were accompanied by pyroclastic flows only if their energy was greater than 10^{11} J. It is illustrated in Figure 14.29.

15 Long-Period and Very Long-Period Seismic Signals at Volcanoes

This chapter is devoted to the seismic signals which are recorded at volcanoes without any association with a particular type of eruptive event but they are common for volcanic activity in general. They are the so-called long-period and very long-period seismic signals. The seismic signals recorded at volcanoes with an emergent onset of *P* waves, no distinct *S* waves, and a harmonic signature in the range of periods between 0.2 and 2 s (0.5 and 5 Hz) are defined by Chouet (2009) as long-period (LP). They are characterized by simple decaying harmonic oscillations except for a brief time interval at the event onset (Chouet, 2003). Seismic signals in the range of periods between 2 and 100 s are defined as very-long-period (VLP) (Chouet, 2009). These signals became a subject of intense study with the beginning of broadband seismic observations at volcanoes in the 1990s. They were recognized at different volcanoes and during different stages of eruptions.

15.1 Waveforms and Spectra

15.1.1 Long-Period Seismic Signals

Long-period seismic signals, derived from volcanic seismicity and used by scientists, may represent the natural LP seismic records or LP component of raw signals of volcanic earthquakes obtained after filtering. Figure 15.1 shows the typical waveforms of the natural LP seismic signals and their Fourier spectra.

The seismic signals, demonstrated in Figure 15.1, were recorded during different stages of eruptive activity. The first of them (Figure 15.1A) was observed simultaneously with a swarm of volcano-tectonic earthquakes preceding the 2000 flank eruption at Usu volcano (see Section 7.2). It has an emergent high-frequency onset; the body waves are not identified. The amplitude of the signal slowly increases, then gradually decreases. These LP signals began to appear within the swarm when the deformation of the site of forthcoming flank eruption became significant indicating a magma intrusion (Zobin et al., 2005).

The second signal is the so-called "tornillo" (screw) (Figure 15.1C); it was recorded on 1 January 2010 before the 2 January explosive eruption at Galeras volcano (BGVN, 2010). Signals of this type show emergent onsets, but occasionally

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00015-3
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Figure 15.1 Examples of long-period seismic waveforms (velocity, vertical component) and their Fourier spectra. (A) Seismic signal recorded at Usu volcano on 30 March 2000 (courtesy of Japan Meteorological Agency) and its Fourier spectrum (B). (C) Seismic signal recorded at Galeras volcano on 1 January 2010 (courtesy of INGEOMINAS, Colombia) and its Fourier spectrum (D).

may have slight impulsive onsets. They are characterized by slowly decaying coda waves. These signals are characteristic for different volcanoes of the world and may be recorded before eruptions, during the swarms of volcano-tectonic earthquakes, after eruptions, or during quiescent periods (Gómez and Torres, 1997).

All these events are characterized by a peak spectrum between 0.8 and 2 Hz (Figure 15.1B, D, and F).

15.1.2 Very Long-Period Seismic Signals

VLP signals at volcanoes are characterized by impulsive signatures; their common characteristic is a period longer than 2 s (Figures 15.2–15.4). VLP signal may be

Figure 15.2 VLP seismograms of the January 2009 explosion at Fuego volcano, Guatemala, extracted from the Guralp CMG 40T (0.02–30 s) broad-band seismic records. The seismic stations were situated at the distances of 800 m (F900) and 1.2 km (F9A) from the crater.
After Lyons and Waite, 2011.

Figure 15.3 Broadband velocity record (top trace) and associated filtered signal of VLP events (bottom trace) recorded at Kilauea volcano during a surge in the magma flow feeding the east rift eruption on 1 February 1996.
After Dawson et al., 1998.

seen frequently only after filtering of the raw signal. Figure 15.2 shows the VLP signals obtained during the January 2009 explosions at Fuego volcano, Guatemala. These signals were extracted from the Güralp CMG 40T sensors (0.02–30 s) records by deconvolving the instrument response and then filtering with a two-pole, zero-phase-shift Butterworth band-pass filter with corners at 30 and 10 s. The highest-energy portion of the VLP signal lasts at least 30 s (Lyons and Waite, 2011). Figure 15.3 shows the repetitive VLP signals recorded at Kilauea volcano (the fourth

Figure 15.4 Broadband velocity VLP seismic record accompanied by LP signal (A) recorded at Cotopaxi volcano on 26 June 2002 without any reported surface manifestation of volcanic activity, its spectrogram (B), and Fourier spectrum (C).
From Molina et al., 2008.

trace). These signals, consisting of pulses with periods of about 20 s, were recognized only when a low-pass filter ($f < 0.125$ Hz) was applied. VLP signals may be followed by LP signals (Figure 15.4).

VLP signals, shown in Figures 15.2–15.4, were recorded during different stages of eruptive activity such as explosive activity at Fuego volcano (Figure 15.2), a surge in the magma flow at Kilauea volcano (Figure 15.3); or without any superficial activity at Cotopaxi volcano (Figure 15.4). Figure 15.5 illustrates the dependence of the amplitude of VLP signals on the intensity of explosive process that was observed at Santiaguito volcano, Guatemala (Sanderson et al., 2010). Two volcanic events, a small degassing and Vulcanian explosion, occurred on 1 January 2009 (Figure 15.5A). The seismic signals, associated with these two events, were recorded by broadband seismic station at a distance of about 500 m from the crater (Figure 15.5B). These signals were band-pass filtered between 600

Figure 15.5 Dependence of the VLP (or ultra-long-period ULP) seismic signal amplitude on the intensity of explosive process at Santiaguito volcano. (A) Photos of small degassing (left) and Vulcanian explosion (right) occurring on 1 January 2009. (B) Raw seismic broadband record. (C and D) Band-pass filtered VLP seismic signals. Vertical bars mark the onset of eruption manifestations for each of the event.
After Sanderson et al., 2010.

and 30 s (Figure 15.5C) and between 10,000 and 30 s (Figure 15.5D). The filtered seismic records displayed VLP signals associated with two volcanic events.

15.1.3 Occurrences of LP and VLP Events

LP and VLP events often occur in sequences where the events are very similar and the time separating events remains relatively constant. Figure 15.6 shows the occurrence of LP (1−2 Hz) events observed at Shishaldin volcano, Aleutian Islands (Peterson et al., 2006). These events began to occur at the volcano from September 1999, a few months after the April−May 1999 eruption. They continued until April

Figure 15.6 Occurrence of LP events recorded at Shishaldin volcano, Aleutian Islands, on 21 November 2002. The section of the short-period helicorder record shows 8 h of data recorded at a distance of 6.3 km from the summit.
From Petersen et al., 2006.

2004. This activity consisted of many hundreds to thousands of events per day, occurring every 1–2 min. The LP waveforms formed the families consisting of nearly identical signals for time spans of days to months (Figure 15.7A).

Long-term sequences of LP events with similar waveforms have been observed at many volcanoes. At Mount Pinatubo, there were three periods with occurrences of similar waveforms: 1 week prior to the June 1991 cataclysmic eruption, a day before a new 1992 dome began to grow, and during the 1992 dome growth (Ramos et al., 1999).

Numerous LP signals (Figure 15.7B) were recorded during the 1989–1990 Redoubt, Alaska eruption (Chouet et al., 1994). The initial eruption on 14 December 1989 was heralded by a 23-h swarm of LP events that ended abruptly with the eruption. Another swarm, with LP events similar to those of the first, began on 26 December and ended in a major tephra eruption on 2 January 1990. A significant but less-intense swarm preceded the 15 February 1990 eruption. From 4 March to the end eruption on 21 April, six swarms of LP were recorded. All of them preceded the eruption episodes.

LP events accompanied the eruptive activity at Popocatépetl volcano during December 1994 through May 1995 (Arciniega-Ceballos et al., 2003), they were

Figure 15.7 Examples of repeating events at Shishaldin volcano recorded at a distance of 6.3 km from the summit (3 January 2002) (A) and at Redoubt volcano recorded at a distance of 4 km from the summit (B). The cluster at Shishaldin volcano was extracted by cross-correlation. The spectral-coherence values for these events were ≥0.9. The upper part of the plot in (A) shows a stack of all events in the lower part. The similar waveforms in (B) are indicated by arrows.
From Petersen et al., 2006 and Chouet et al., 1994.

recorded in 1991 during the emplacement of a lava dome in Galeras volcano (Gil Cruz and Chouet, 1997), and they were reported during the 2004 observations at Etna volcano (Lokner et al., 2007). The number of references may be continued.

Families of VLP signals have been recorded at different volcanoes also: Stromboli (Chouet et al., 2003), Popocatépetl (Arciniega-Ceballos et al., 1999), Erebus (Aster et al., 2008), among others. Figure 15.8 shows three families of VLP seismic records from Strombolian eruptions of Erebus volcano. Group 1 eruptions (Figure 15.8A) are characterized by positive initial vertical motions, and group 2 eruptions (Figure 15.8B) show negative initial vertical motions. Group 3 eruptions (Figure 15.8C) were very rare and exhibit a relatively simple pulse-like shape.

15.1.4 Nature of LP and VLP Seismic Signals

The main properties of LP and VLP signals may be summarized as the following: they are repetitive; they form the families consisting of similar waveforms, which can be sustained for a long time. The presence of long-live families of similar

Figure 15.8 Three families of VLP seismic signals recorded during Strombolian explosions at the lava lake of Erebus volcano.
From Aster et al., 2008.

waveforms demonstrates the stable and nondestructive nature of their sources. Chouet (2003) wrote that the LP waveforms are characterized by simple decaying harmonic oscillations may be interpreted as oscillations of a fluid-filled oscillator in response to a time-localized excitation and can be thought of as the manifestation of acoustic resonance. Unlike them, VLP signals may be thought of as the result of inertial forces associated with perturbations in the flow of magma and gases through conduits (Chouet, 2009). Both these types of sources may be considered as multiphase systems.

Seismic Waves Generated Within the Two-Phase System

The fluid-filled oscillator may be represented as a solid-fluid composite (Kawakatsu and Yamamoto, 2007). The presence of ample fluid (gas, vapor, magma, or their mixtures) in volcanic edifices introduces an additional class of vibration/wave phenomena to the well-studied elastic formulations. There exists a class of waves in the two-phase system of a solid−liquid composite that travel slower than any of the sound velocities of the pure material constituting the two-phase system (i.e., the wave speed can be slower than that of the liquid phase). Rayleigh wave and Stoneley wave, as well as the crack wave found by Chouet (1986), may be included in this class. The simplest example of slow waves may be the plane wave propagation in a medium which consists of a fluid layer sandwiched by two elastic half-spaces (Ferrazzini and Aki, 1987).

Figure 15.9, taken from Kumagai (2009), illustrates the relationship between the phase velocities of the acoustic and crack waves (Figure 15.9B), generated in a solid−fluid composite. Here a compressive fluid layer is sandwiched between two elastic half-spaces, which mimics a dyke containing basaltic magma with a gas-volume fraction of 15% (Figure 15.9A). The dispersion relationship in Figure 15.9B clearly indicates that the phase velocity of the fluid-solid coupled crack wave depends on the wavelength. Kumagai (2009) showed that the normal displacement, symmetric with respect to x_3, causes the deformation of the fluid-solid boundary. The deformation has the effect of lowering the phase velocity of the crack wave.

Source of LP and VLP Seismic Signals

Waveform inversions of LP and VLP seismic signals show that their sources may be modeled by the superposition of six moment-tensor and three single-force components (Kumagai et al., 2005; Ohminato et al., 2006; Waite et al., 2008; Chouet, 2009). The modeling is applied at a common point within the volcano, and therefore, as noted by Aster et al. (2008), is a highly idealized representation of a spatially distributed source. However, this point approximation may be good enough when scaled to the wavelength of long-period seismic waves.

An example of the source-time functions obtained for the 2 July 2005 LP event recorded during lava dome growth at Mount St. Helens (Waite et al., 2008) is shown in Figure 15.10. The moment tensor is dominated by the three diagonal

Figure 15.9 Fluid layer of thickness d sandwiched between two elastic half-spaces bounded at $x_3 = -d/2$ and $x_3 = d/2$ (A) and the phase velocities of the acoustic wave (dashed line) and crack wave (solid line) generating within this fluid–solid composite (B). These are normalized by the fluid acoustic velocity, and plotted as a function of the wavelength normalized by the thickness of the fluid layer.
From Kumagai, 2009.

terms (dipole components) of the seismic moment tensors (M_{xx}, M_{yy}, and M_{zz}) and a vertical single-force component F_z. The dipole components are in phase and there is very little energy on the shear components, indicating the mechanism represents a volumetric source. The amplitude ratio F_z/M_{zz} is roughly 2.8×10^{-4} m^{-1}. The

Figure 15.10 Source time function of the 2 July 2005 LP event recorded at Mount St. Helens. It is dominated by the volumetric components of the moment tensor (A) and a vertical single force (B).
From Waite et al., 2008.

moment-tensor and single force components contribute equally to the waveform amplitudes when this ratio is 10^{-3} m^{-1} (Ohminato et al., 1998). Therefore, the F_z contribution is equal to 28% of the wave amplitudes. Waite et al. (2008) consider that the vertical-force component may be explained by oscillations of the growing spine and possibly the entire dome.

An example of the source-time functions obtained for the Popocatépetl VLP source during the 23 May 2000 Vulcanian degassing (Chouet et al., 2005) is shown in Figure 15.11. Volumetric components of the moment tensor clearly dominate the solution, with the source time functions characterized by almost identical time histories. Accompanying the volumetric source components are two clear components of a single force in the east and vertical directions contributing up to 16% and 13% of the wave amplitude, respectively.

This dominance of the components of the moment tensor above the single force components is characteristic for the majority of LP and VLP sources obtained from an inversion of the seismic signals.

15.2 Geometry of the Sources of LP and VLP Seismic Signals

LP and VLP seismic signals originating from the vibration of the stable nondestructive fluid-filled sources (dykes, conduits, cracks) carry information about the

Figure 15.11 Source time function of the 23 May 2000 VLP event recorded at Popocatépetl volcano. It is dominated by the volumetric components of the moment tensor and the eastern and vertical components of a single force.
From Chouet et al., 2005.

geometry of these sources. Inversion of the seismic waves of the LP and VLP signals gives us the seismic moment tensor of the source. As was mentioned in Section 3.1.3, the ratios of three diagonal terms of the seismic moment tensors M_{11}, M_{22}, and M_{33} (or M_{xx}, M_{yy}, and M_{zz}) allow the discrimination of the geometry of the nondouble-couple sources of seismic waves indicating a crack, a cylindrical pipe, or a sphere (Table 15.1).

The source of the 1 February 1996 LP event, that occurred just 12 h before the volcanic crisis at Kilauea volcano, was modeled as a horizontal crack (Kumagai et al., 2005). The source time functions of the six moment-tensor components calculated from the waveform inversion are shown in Figure 15.12. A domination of the M_{zz} component over the M_{xx} and M_{yy} components is seen (M_{11}:M_{22}:$M_{33} \cong 1$:1:3) reflecting a horizontal crack source.

A cylindrical pipe was obtained as a model of the source of LP signals for the 30 October 1986 event associated with the explosion at Sakurajima volcano (Uhira and Takeo, 1994). The source time functions of the six moment-tensor components calculated from the waveform inversion are shown in Figure 15.13. The amplitude ratios M_{xx}:M_{yy}:M_{zz} are about 1.8:1.8:1. This nearly corresponds to a contraction of a cylindrical region almost vertically embedded in an elastic medium, equivalent to the conduit.

Sometimes the moment tensor solutions give the ratios of three diagonal terms of the seismic moment tensors M_{11}, M_{22}, and M_{33}, which differ from the simple situations shown above. For example, the ratios of three diagonal terms for

Table 15.1 Ratios of Three-Diagonal Terms of the Seismic Moment Tensors M_{11}, M_{22}, and M_{33} for Some NonDouble Couple Sources of Volcanic Earthquakes (the Elastic Modulus Is Taken as $\mu = \lambda$)

Type of source	M_{11}:M_{22}:M_{33}	Reference
Spherical	1:1:1	Kumagai (2009)
Vertical cylinder	2:2:1	Kumagai (2009)
Vertical tensile crack	3:1:1	Kawakatsu and Yamamoto (2007)
Horizontal tensile crack	1:1:3	Kumagai (2009)

Figure 15.12 Source-time functions of the six moment-tensor components obtained from waveform inversion of the LP waveforms recorded at eight stations on 1 February 1996 at Kilauea volcano.
From Kumagai et al., 2005.

the 2000 Vulcanian degassing event at Popocatépetl was 1.20:1.45:2.00 (see Figure 15.11), that, as considered by Chouet et al. (2005), indicates the possibility of a dual mechanism consisting of two elements: a shallow dipping crack (sill) and a northeast trending quasi-vertical crack (dyke).

15.3 Type of Fluid Within the Fluid-Filled Cracks

The harmonic oscillatory signatures of LP events are interpreted as oscillations of fluid-filled resonators at the source of these signals. The type of fluid filling these resonator systems may be identified from the comparison of the complex frequencies of decaying harmonic oscillations in the tail of the seismogram of the LP seismic signals and the model crack data.

Figure 15.13 The result of inversion of the LP waveforms recorded on 30 October 1986 at Sakurajima volcano at three stations (ARIM, KOME, and KAGO) and the estimated source-time functions of the six moment-tensor components. Thin lines show the error bound of the estimated time histories. The observed and calculated seismograms are abbreviated as OBS and CALC. The vertical, radial, and transverse components are denoted by UD, R, and T, respectively.
From Uhira and Takeo, 1994.

15.3.1 Crack Model

The rectangular crack model with a width W, length L, and aperture of the crack d was used for the simulation of the dynamics of a fluid-filled crack (see Section 11.3.1, Figure 11.9). This model, proposed by Chouet (1988), is characterized by the independent parameters α/a and ρ_f/ρ_s, where a is the speed of sound propagating within the fluid in the crack and α is the P wave velocity of the rock

matrix, and ρ_f and ρ_s are the densities of the fluid and rock matrix, respectively. The waveform, obtained from the simulation of the crack excitation in the form of a pressure step applied over a small area of the crack wall at the center of the crack, can be regarded as the velocity impulse response of the crack.

15.3.2 Complex Frequencies of the LP Seismic Signal for Different Fluids

Complex frequencies refer to the frequencies and quality Q values of decaying harmonic oscillations of a LP signal. The complex frequency is defined as $f - ig$, where f is the frequency, g is the growth rate, and $i = \sqrt{-1}$. Accordingly, the factor Q of the signal is defined as $-f/2g$. The crack model simulations predict Q due to the radiation loss, or $Q = Q_r$ (Kumagai and Chouet, 2000).

Figure 15.14 demonstrates the contour diagrams of Q_r and the dimensionless frequency $\upsilon = fL/\alpha$ and plots of α/a versus ρ_f/ρ_s curves for various types of fluids calculated for the synthetic waveforms obtained from the simulation of the crack excitation (Figure 15.15). As a result, it is possible to discriminate various mixtures of gas, liquid, and solid, including gas−gas mixtures (H_2O-CO_2, H_2O-SO_2), liquid−gas mixtures (water−H_2O and basalt−H_2O), and dusty and misty gases (ash−SO_2 and water droplet−H_2O) (Kumagai and Chouet, 2000).

15.3.3 Identification of the Type of Fluid from LP Seismic Signals

To estimate the complex frequencies f, a Sompi analysis (Hori et al., 1989), based on an autoregressive model and assuming harmonic decaying oscillations, may be applied. In this method of spectral analysis, a signal is deconvolved into a linear combination of coherent oscillations with decaying amplitudes (called wave elements) and additional noise. An amplitude value also is calculated for each wave element. Generally, in order to represent a set of complex frequencies, they are plotted on a two-dimensional plane with the frequency f and growth rate g axes, called the "$f-g$ diagram." Wave elements scattering widely in the plot are considered to be noise, while wave elements densely populated on the theoretical frequency lines are considered to be the dominant spectral components of the signal (Hori et al., 1989).

Figure 15.16A shows a LP seismic signal recorded near Kilauea volcano, Hawaii (Kumagai et al., 2005). To determine the complex frequencies from the waveform, the onset portion of the LP seismic signal, which contains the inhomogeneous part of the signal, was deleted. The Fourier spectrum of this signal (Figure 15.16B) was characterized by two dominant peaks near 0.6 and 1.3 Hz. The Q values of the two peaks are in the range of 10−20 (Figure 15.16c). Comparison with the diagrams of Figure 15.14 gives a bubbly water fluid filling the crack.

The Fourier and Sompi frequency domain analyses were applied to the tornillo-type LP events (Figure 15.17A) recorded at Vulcano volcano, Italy during 2007−2008 (Miluzzo et al., 2010). Recent volcanic activity of this volcano, since the last eruption in 1888−1890, produces fumarolic emissions. The Fourier

Figure 15.14 Contour diagrams of Q_r (A) and dimensionless frequency $\upsilon = fL/\alpha$ and plots of α/a (B) versus ρ_f/ρ_s curves for various types of fluids. Dots indicate points where the synthetic waveforms are calculated. Dashed, solid, and dash-dotted lines indicate α/a and ρ_f/ρ_s calculated for fluids at pressures of 5, 10, and 20 or 25 MPa, respectively. From Kumagai and Chouet, 2000.

Figure 15.15 Synthetic far-field compressional waveforms produced by the excitation of a crack containing different types of fluids. The crack is embedded in a rock with fixed properties $\alpha = 2{,}700$ m s^{-1}, $\rho_s = 2{,}500$ kg m^{-3}. (A) Waveform for a crack at a depth of 200 m containing bubbly water with 5% gas-volume fraction at a temperature of 100 °C. (B) Waveform for a crack at a depth of 500 m containing a bubbly basalt with 5% gas-volume fraction at a temperature of 1,200 °C. (C) Waveform for a crack at a depth of 200 m containing a 50% mixture of either H_2O-CO_2 or H_2O-SO_2 gases at a temperature of 500 °C. (D) Waveform for a crack a depth of 500 m containing an ash-laden gas with gas-volume fraction of 60% at a temperature of 500 °C.
From Chouet, 2003.

Figure 15.16 Seismic record of the 1 February 1966 LP event recorded at Kilauea volcano (A), its Fourier amplitude spectrum (B) and the plots of the complex frequencies estimated for this waveform. The clusters of points within ellipses represent clear signals. The solid lines represent lines along which the Q factor is constant.
From Kumagai et al., 2005.

analysis showed that the most of the tornillo signals are characterized by two sharp spectral peaks at frequencies of about 6 and 10 Hz (Figure 15.17C). To obtain information about the decay of the tornillo signals, the Sompi analysis was applied. The amplitude information was added by using a color scale for the dots (Figure 15.17D). Such amplitude information is useful to further discriminate the noise and dominant spectral components of the signal. The former and the latter are generally characterized by small and large amplitudes, respectively (Hori et al., 1989). A 10-s window from the recording of the vertical component of short-period

Figure 15.17 Vertical component waveforms (A), short-time Fourier transform spectrogram (STFT) (B), Fourier spectra (C) and $f-g$ plots (D) of four examples of tornillos recorded at Vulcano volcano during 2007–2008. The spectra and the $f-g$ plots were calculated by using 20-s-long windows. The color of dots in (D) indicates the amplitude values of the wave elements (see color scale). Clusters of dots in (D) (encircled with solid ellipses) indicate dominant spectral components of the signal, scattered dots represent noise.

From Miluzzo et al., 2010.

seismic station IVCR, situated at a distance of about 2 km from the crater, was analyzed for each tornillo event. The signal was filtered in two frequency bands, 5–8 Hz and 8–12 Hz, and the Sompi method was applied to each band. The frequency values obtained confirm the results of the FFT analysis. It was noted that most of the quality factor values ranged between 20 and 400. According to Figure 15.14, the fluid-filled resonator could be an ash–gas or a water droplets–gas mixture.

A swarm of LP seismic events was recorded in December 2001 before heightened eruptive activity at Tungurahua volcano, Ecuador (Molina et al., 2004). The waveforms of two of them are shown in Figure 15.18A. Sompi analysis was applied to estimate the complex frequencies and Q for 57 LP events. Figure 15.18B and C shows that both the frequency and Q systematically changed during the swarm. The factor Q gradually increased from 100 to 400, while the frequency gradually decreased from 3.5 to 3 Hz. Molina et al. (2004) considered that the

Figure 15.18 Seismic records of two December 2001 LP events recorded at Tungurahua volcano (A) and the temporal variations in Q (B) and frequency (C) determined for LP events recorded at Tungurahua volcano during the December 2001 seismic swarm. The solid lines show the best fits of frequency and Q calculated for a crack containing the ash-gas mixture at various pressures between 25 and 75 MPa corresponding to the source depths between 1 and 3 km below the summit, in which crack lengths of 200, 230, and 240 m at depths of 1, 2, and 3 km, respectively, were assumed.
After Molina et al., 2004.

observed temporal variations may be explained by a gradual increase of the ash content in the crack at the source of the LP events.

15.4 Location of the Sources of LP and VLP Events

This type of seismicity cannot be located by conventional means due to the lack of impulsive body wave phases in the records. The source locations of LP and VLP events are obtained in a probabilistic sense using such methods as frequency-slowness analysis (Almendros et al., 2001), waveform inversion (Chouet et al., 2003), cross-correlation (De Barros et al., 2009), and waveform semblance (Almendros et al., 2002). The majority of the sources generating LP and VLP signals at volcanoes have been located within the depth interval between 100 and 400 m beneath the active crater or dome (Merapi, Hidayat et al., 2002; Stromboli, Chouet et al., 2003; Mount St. Helens, Waite et al., 2008; Kilauea, Kumagai et al., 2005; Etna, De Barros et al., 2009); however, some of them were located much deeper, between 1 and 8 km (Cotopaxi, Molina et al., 2008; Popocatépetl, Chouet et al., 2005; Aso, Legrand et al., 2000; Kilauea, Almendros et al., 2002).

Bottaglia et al. (2003) developed the methodology of locating LP events using seismic amplitudes instead of using arrival times. The method is based on seismic amplitudes corrected for station site effects using coda site amplification factors. Once corrected, the spatial distribution of amplitudes shows smooth and simple contours for long-period events. The density of the contours also gives an estimation of the depth of the source. On the basis of the simplicity of these distributions, the inversion methods for locating their origins were developed. To achieve this, the decrease of the amplitude as a function of the distance to the source was approximated by the decay either of surface or of body waves in a homogeneous medium. Amplitudes at each station were calculated using RMS on 20.28-s windows starting shortly before the onset of the signal.

Figure 15.19A shows the location of LP event occurring at Kilauea volcano in January 1998 using seismic amplitudes (Battaglia et al., 2003). The best source is plotted as a white star on the plane projection showing the summit of the volcano, as well as on the east−west and north−south cross sections. This methodology was applied to the location of 94 LP events occurring during January 1998 below Kilauea (Figure 15.19B). The foci of LP events clustered at depths between 2 and 6 km b.s.l.

Figure 15.20 shows the location of the source centroid obtained by waveform inversion of VLP seismic signals generated during the 23 May 2000 explosive activity at Popocatépetl volcano, México. A network consisting of 15 broadband seismic stations was used. The source centroid that best fit the data was located at an elevation of 3,600 m, approximately 1,500 m below the western perimeter of the crater floor (Chouet et al., 2005).

The methodology of cross-correlating signals, obtained from 50 broadband seismic stations, was used to locate LP seismic events at Etna volcano during June 2008 (De Barros et al., 2009). To extract the long-period events, a STA/LTA

Figure 15.19 Location of LP events, occurring during January 1998 below Kilauea volcano, using seismic amplitudes. (A) Location of an LP event and the east–west and north–south cross sections. (B) Location of 94 LP events and the east–west and north–south cross sections. The decays (observed and modeled) of the amplitude as a function of the distance are shown in insert. In (B) the routine catalog locations are plotted in the background as gray points. See text for more explanations.
After Battaglia et al., 2003.

Figure 15.20 Source location (white dot) of the VLP event occurring during the 23 May 2000 explosive activity at Popocatépetl volcano from waveform inversion. A south-west looking cutaway view of the volcano is shown at the top to help identify the location of the VLP source within the volcanic edifice. Contours represent 7.9%, 8.1%, and 8.3% errors. From Chouet et al., 2005.

method (2 s over 20 s window lengths, with a threshold of 2.5) on the band-pass filtered data (0.2–1.5 Hz) was used. Selected 500 events were classified using a cross-correlation analysis between all pairs of signals. The events that gave a correlation coefficient greater than 0.9 with all events on at least three out of the four permanent stations close to the summit formed two families with a similar number of events (63 and 66, respectively). The location of the LP events was computed in two steps. At first, the mean position for each family was found, and then the individual events were located using this first position.

Figure 15.21 shows the locations of 129 sources that generated two families of LP seismic signals. The hypocenters were found at shallow depths (20–700 m deep) below the summit craters. The very high location resolution allowed the detection of the temporal migration of the LP events along a dyke-like structure (De Barros et al., 2009).

The location of the sources of VLP seismic signals at Kilauea volcano during the period 9–16 September 1999 was performed using a method based on analysis of semblance (Almendros et al., 2002). Semblance is a measure of the similarity of multi-channel data weighted according to the rectilinearity of the particle motions. The method determines the point for which the radial components of the signals recorded at different receivers, aligned to correct for the wave propagation delay, show maximum similarity. The procedure, used to obtain a source location by the semblance method, consists of three steps: (1) determination of the spatial extent of the region

Figure 15.21 Location for the sources of family 1 (left) and 2 (right), generating LP seismic signals at Etna, performed by cross-correlating signals for pairs of stations, with colors indicating temporal evolution. Color scale (days) is common for both families. Views are from above, south and west. Triangles are some of the broad-band stations. ''⊥'' and ''→'' symbols represent the planar and pipe-shape structures, with dip and plunge angles, respectively.

From De Barros et al., 2009.

of study by defining a three-dimensional grid of assumed source positions; (2) fixation of the interval of signal considered by selecting the start time of the analysis window and its duration; and (3) calculation of semblance using the arrival times calculated for each node of the source grid. This procedure gives a three-dimensional distribution of radial semblance.

Two source regions for 16 VLP events were identified (Figure 15.22). One of them was located at a depth of about 1 km beneath the northeastern edge of the Halemaumau pit crater. The second region was located at a depth of about 8 km below the northwest quadrant of Kilauea caldera (Almendros et al., 2002).

15.5 Conceptual Models of the Relationship Between the Sources of the LP and VLP Seismic Signals and Their Role in Eruptive Process

A study of the sources of LP and VLP seismic signals recorded during the period from November 2003 to May 2006 at Etna using radial semblance showed that these two types of signals are not always generated by a common source, but by

Figure 15.22 Source locations of the 16 VLP events recorded at Kilauea volcano during September 1999 and located by a method based on analysis of semblance and particle motion. The sources are shown by large open circles. The topography of Kilauea caldera is shown at the top of the figure. H is Halemaumau, UB is Uwekahuna Bluff. Contours on the side planes are projections of the error regions for the source locations. The gray surfaces represent the outline of the error regions for all events.
From Almendros et al., 2002.

different sources connected to each other. It was obtained that a deeper dyke, at about 2.5 km a.s.l., was the source of the VLP events, while a shallower crack, situated 500 m above the dyke, generated LP signals (Cannata et al., 2009a). The same relationship between depth and the generation of LP and VLP seismic signals was reported for Cotopaxi and Mount St. Helens volcanoes (Molina et al., 2008; Waite et al., 2008).

The relative source process for the 2 July 2005 Mount St. Helens LP and VLP events located using waveform inversion was discussed by Waite et al. (2008). In the seismic record, the main part of the VLP signal occurred after the initiation of an LP event. The position of the LP and VLP sources is shown in Figure 15.23. The LP source is very shallow, at about 2,000 m elevation and directly beneath the growing dome. The VLP source is 250 m deeper and 400 m north–northwest, beneath the southwest end of the old 1980s lava dome. The LP source does not seem to be within the eruption conduit.

A possible scenario for the LP events at Mount St. Helens involves the repeated pressurization of a sub-horizontal crack that lies beneath the southern crater and new lava dome. Steam generated by the heating of groundwater, in addition to exsolved steam from the magma, fed into the crack, gradually expanding and pressurizing it. The southern, higher altitude end of the crack must be sealed to prevent

Figure 15.23 Relative position of the sources that generated the 2 July 2005 LP and VLP seismic signals during lava dome growth in the crater of Mount St. Helens. The east-northeast, three-dimensional, perspective view of the LP crack (blue) and VLP sill-dyke (green) sources are shown. Topography is shown at the bottom as a surface and is contoured at 20 m. Distances are in m.
From Waite et al., 2008.

steam from escaping. When a pressure threshold is reached, some steam is forced out of the crack through fractures that eventually lead to the surface or out to the glacier-crater floor interface. Waveform inversion of the VLP events revealed a north-dipping sill plus a near-vertical dyke or pipe. The VLP source sill was roughly in the plane of the LP source crack. The main VLP signal was generated by deflation−inflation of the sill and near-simultaneous inflation−deflation of the dyke. Large, sub-horizontal forces that accompanied the crack volume changes were interpreted as reaction forces on the Earth resulting from transient mass advection in the conduit. It was proposed that the collapse of the LP crack and related sagging of the dome triggers a response in the magma conduit generating the VLP band.

Molina et al. (2008) proposed the following conceptual model of the source processes of VLP and LP events recorded at Cotopaxi volcano and located from the waveform inversion (Figure 15.24). Temporal changes of both the VLP and the LP signals were observed, in which a harmonic signature gradually appeared in the LP signal and became dominant over the VLP signal (see Figure 15.4). Molina et al. (2008) consider that the process began with a magma intrusion beneath Cotopaxi in November 2001, which triggered a swarm of VT earthquakes. The magma intruded into the shallower edifice up to a depth of 2−3 km a.s.l. beneath the northeastern

Figure 15.24 Conceptual model for source processes of the VLP and LP events observed at Cotopaxi volcano. See text for details.
From Molina et al., 2008.

flank of the volcano (Figure 15.23A). Supersaturated conditions were reached within the intruded magma with a resulting pressure increase through decompression-induced degassing and vesicle growth, while the outer part of the magma solidified due to chilling from the surrounding rock (Figure 15.23B). As stress due to the pressurization reached the shear strength of the magma, brittle fracture occurred at the top of the magma body, releasing small magma particles with gases. The released particle-laden gas opened a crack in the rock matrix just above the magma system (Figure 15.23C). The LP events can be interpreted as the resonance of a crack above the magma system while the VLP events represent the volumetric expansion during the fracturing in the magma and opening of the crack. At the beginning of the VLP/LP activity, the LP source region above the magma system consisted of a dendritic system of cracks so that the LP waveforms displayed complex non-harmonic oscillations (Figure 15.23D).

16 Swarms of Microearthquakes Associated with Effusive and Explosive Activity at Volcanoes

Swarms of microearthquakes (majority of them with magnitudes of about 0) have been recorded on different volcanoes of the world (e.g., Núñez-Cornú et al., 1994; Orozco-Rojas, 1994; Neuberg et al., 1998; Carniel et al., 2006; Iverson et al., 2006; Macedo et al., 2008; Waite et al., 2008; Varley et al., 2010; Zobin et al., 2010). They have been associated with effusive and explosive eruptions at volcanoes of all types: basaltic, andesitic, and dacitic. These swarms occur frequently just before large explosions or lava extrusions. While these seismic signals have been given many names (hiccups, pulgas, drumbeats, low-frequency, long-period, and hybrid), in this chapter these are referred to as "microearthquakes" and discusses their waveforms, their nature, and their relationship to eruptive events.

16.1 Waveforms and Spectra

Figure 16.1 shows two seismograms typical of the microearthquake swarms recorded at Volcán de Colima. One seismogram (top panel) was recorded just before the beginning of the 30 September 2004 lava extrusion, the second seismogram (bottom panel) was recorded just before the 30 April 2005 large explosion. These seismic signals are numerous, low-amplitude, and form families of similar records (Figure 16.2). Cross-correlation test performed on the initial 10-s of the seismic records of microearthquakes observed during the March–September 2005 explosive activity at Volcán de Colima (Zobin et al., 2010) shows that the waveforms have a good intercluster correlation.

Three types of microearthquake waveforms were discriminated at Volcán de Colima. The seismic signals and spectral characteristic of the first type are shown in Figure 16.3, together with the waveforms of microearthquakes recorded at volcanoes Stromboli, Ubinas, and Mount St. Helens. The waveforms consist of two signals: an initial low-frequency and small-amplitude pulse (between t_1 and t_2) and a second pulse characterized by larger amplitude and higher frequency. The spectral peaks are between 2 and 7 Hz. They are usually referred to as "low-frequency" or "long-period" earthquakes.

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00016-5
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Figure 16.1 Twenty-four-hour seismic records of microearthquake swarms that were observed at Volcán de Colima, México before the 30 September 2004 lava extrusion (top panel) and before the 30 April 2005 explosion (bottom panel). Short-period seismic station EZV4 (velocity, vertical channel), 1.7 km from the crater.
Courtesy of Colima Volcano Observatory.

Figure 16.2 Eight seismic records (velocity, vertical component) of micro-earthquakes at Volcán de Colima with adjusted relative time are shown. The coefficients of inter-cluster correlation for these eight records vary between 0.55 and 0.95.
From Zobin et al., 2010.

The second type of microearthquake waveforms is shown in Figure 16.4. The waveforms consist of two signals: an initial high-frequency and small-amplitude pulse (between t_1 and t_2) and a second pulse characterized by larger amplitude and lower frequency. The spectral peaks are between 3 and 4 Hz. These waveforms are usually referred to as "hybrid" earthquakes.

The third type of microearthquake waveform observed at Volcán de Colima is shown in Figure 16.5. These waveforms are characterized by a spindle-shaped seismic signal with a gradual increase–decrease in amplitude, often without a clear signal onset. They may be long (Figure 16.5A) or short (Figure 16.5C). The spectral peaks are between 2 and 4 Hz.

16.2 Structure of Microearthquake Swarms

Martini et al. (2007) studied the microearthquake swarms recorded before and after explosions that were observed during the 2007 eruption at Stromboli volcano. They showed that the swarms consisted of two types of signals defined as long-period events and hybrid events (Figure 16.6).

The swarms of microearthquakes (named "drumbeats"), that were observed during the 2004–2005 effusive–explosive activity at Mount St. Helens volcano, consisted of repetitive low-frequency (peak frequencies in the 2–3 Hz range) and

Figure 16.3 Seismograms (velocity, vertical component) of microearthquakes of the first type recorded at volcanoes Stromboli (24 February 2006, 300 m from the crater, broadband record); Ubinas (29 May 2006, 2.5 km from the crater, broadband record); Mount St. Helens (8 March 2005, a few hundred m from the crater, short-period record); and Volcán de Colima (14 May 2005, 1.7 km from the crater, short-period record) are shown on the left side of the figure and the corresponding Fourier spectra of the full records are shown on the right side. Arrows show the first pulse of the seismic record.
Courtesy of Sezione di Catania, Istituto Nazionale di Geofisica e Vulcanologia, Instituto Geofísico de Perú, Arequipa, and Colima Volcano Observatory, respectively.

hybrid (peak frequencies in the 8−16 Hz range) earthquakes and accompanied the regular rate of extrusion of several dacite spines. The first swarm of these microearthquakes that occurred on 16 October 2004 is shown in Figure 16.7. Most of these seismic signals occurred in families of similar events (Figure 16.8).

Moran et al. (2008a) noted also the occasional occurrence of larger, or "big" ($M > 2$, $M_{max} = 3.4$) shallow earthquakes within the microearthquake swarms

Figure 16.4 Seismograms (velocity, vertical component) of microearthquakes of the second type recorded at Volcán de Colima (A and C) during the 2005 activity and corresponding Fourier spectra of the full records (B and D). Arrows show the first pulse of the seismic record.
Courtesy of Colima Volcano Observatory.

Figure 16.5 Seismograms (velocity, vertical component) of microearthquakes of the third type recorded at Volcán de Colima (A and C) during 2005 activity and corresponding Fourier spectra of the full records (B and D).
Courtesy of Colima Volcano Observatory.

Figure 16.6 Seismic signals (velocity, vertical component) and corresponding spectrograms of long-period (A) and hybrid (B) microearthquakes recorded at Stromboli volcano by a broadband seismic station during 2007 activity.
From Martini et al., 2007, with modifications.

(Figure 16.9A). Despite their size, the larger earthquakes were similar in several ways to the drumbeat events. Like the drumbeats, they were classified as either long-period or hybrid events. Spectra for drumbeats and the larger events were also similar (Figure 16.10). Epicenters of larger events were located within a small area 1 km by 1 km centered on the vent; they clustered more densely than drumbeat locations (Figure 16.9B). The part of small earthquakes were rockfall seismic signals (Moran et al., 2008).

Figure 16.7 Short-period seismic record (velocity, vertical component; 2.5 km from the crater) of the microearthquake swarm that occurred at Mount St. Helens from 15 to 17 October 2004 (A) and characteristic seismic records (B) of low-frequency and hybrid events composing this swarm.
After Moran et al., 2008.

Figure 16.8 Example of a family (multiplet) of waveforms of microearthquakes occurring within the 16 October swarm at Mount St. Helens (see Figure 16.6). Top trace is seed event for multiplet, and last trace is stack of all waveforms within multiplet at the given station. Date and time on first trace are when seed event occurred. Arrow in bottom trace shows first arrival with polarity (*up* is compressional, *down* is dilatational). Cross-correlation value of each trace with respect to seed event is shown on the *y*-axis. Waveforms are unfiltered and chosen at constant time intervals within the lifespan of their respective multiplets.
After Thelen et al., 2008.

Zobin et al. (2010) showed that the microearthquakes of the first and the second types, which were observed during the 2005 explosive activity at Volcán de Colima and described in Section 16.1, represented from 84% to 89% of total 966 identified events while the microearthquakes of the third type represented from 11% to 16%.

16.3 Microearthquake Swarms in Eruption Process

As mentioned above, microearthquake swarms have been observed during effusive and explosive activity at volcanoes of different types all over the world.

16.3.1 Kizimen volcano, Kamchatka

The 2376-m-high Kizimen is an isolated, conical andesitic stratovolcano situated in the Eastern Kamchatka, Russia. The latest eruptive cycle began about 3000 years ago with a large explosion and was followed by lava dome growth lasting intermittently about 1000 years. Only a single explosive eruption, during 1927-28, had been recorded in historical time. New volcanic unrest began in July 2009 with the swarms of volcano-tectonic earthquakes. Beginning from October 2010, the strong gas-and-steam activity was recorded. Strong explosions with the ash plumes rising up to 6−7 km a.s.l. were observed during December 2010 to May 2011. Numerous

Figure 16.9 Short-period seismic record for 48 h between 17:00 PST on 28 November and 17:00 PST on 30 November 2004, showing 10 "big" earthquakes occurring along with regular drumbeats (A) and the map showing epicenters of drumbeat events (blue dots) and impulsive "big" events (red dots) occurring between 1 November 2004 and 31 December 2005 (B). Dot sizes vary with magnitude. Drumbeat locations include events with *P* arrivals on at least seven stations, the nearest station being within 2 km, and an azimuthal gap ≤135° (all impulsive "big" event locations meet these same criteria). Yellow triangles correspond to seismic station locations as of July 2005.

After Moran et al., 2008a.

Figure 16.10 Instrument-corrected short-period seismograms and frequency spectra for typical drumbeat (A) and "big" (B) seismic events. Stations are arranged in increasing distance from vent, with distance indicated for each station. Spectra are for 10.24-s windows starting with *P*-wave arrival. Thin dotted line below each spectral plot is noise estimate based on 10.24-s window taken before first arrival. Vertical thick dashed lines indicate corner frequencies (F_c) for each event. (A) Drumbeat event, M_L 1.2, on 20 August 2005 at 13:45 PDT. (B) "Big" event, M_L 3.0, on 26 March 2005 at 19:28 PST.
From Moran et al., 2008a.

pyroclastic flows and a small lava flow were formed (Malik and Ovsyannikov, 2011; GVP, 2011).

A sequence of micro-earthquakes began on 12 May 2011 after the large explosion of 2 May (Pavel Firstov, personal communication). Figure 16.11 shows the sequence of micro-earthquakes recorded by the short-period seismic station at a distance of 2.6 km from the crater. These short (duration of 20–30 s) seismic signals are characterized by the long-period initial phase followed with larger-amplitude high-frequency phase.

16.3.2 Stromboli Volcano, Aeolean Islands

The 924-m-high basaltic stratovolcano Stromboli, situated in the Southern Tyrrhenian Sea on the island of Stromboli, is an open-conduit volcano characterized by the permanent emission of a gas plume from the craters and persistent activity during at least the last 1000 years (Rosi et al., 2000). The swarms of

Figure 16.11 Sequence of micro-earthquakes observed during the 2011 explosive-effusive activity at Kizimen volcano (above) and the detailed seismic record of two of them (below). Courtesy of Institute of Volcanology and Seismology, Kamchatka, Russia.
From Carniel et al., 2006.

microearthquakes, represented by hybrid and long-period waveforms, have been observed at different stages of explosive and effusive activity of this volcano.

On 5 April 2003, a powerful explosion covered a large part of the normally tourist-accessible summit area with bombs (Carniel et al., 2006). The event started with a reddish ash emission related to collapses within the craters and was soon replaced by emissions of darker juvenile material from the northeast crater and then similar emissions from the southwest crater, with the appearance of a dark, mushroom-shaped cloud rising about 1 km above the summit. Bombs, ash, and blocks affected most of the western sector; the volcano top above 700 m above sea level (a.s.l.) was completely covered by bombs, some being meter-sized. This explosion was preceded by a swarm of microearthquakes.

Significant effusive eruption interrupted the persistent explosive activity at Stromboli during the period 27 February to 2 April 2007. Microearthquakes were recorded before, during, and after the effusive episodes (Martini et al., 2007). Figure 16.12 shows the seismograms filled with microearthquakes recorded just before and after the 27 February effusive eruption. Its onset changed the visible activity of microearthquakes; before the onset of effusive eruption, they were of larger amplitudes.

Figure 16.13 shows the location of long-period microearthquakes during the period from January 1 to 22 May 2007 (Martini et al., 2007). The temporal range is divided into four intervals: (1) 1–31 January; (2) 31 January–26 February; (3) 27 February–31 March; (4) 31 March–22 May. There were no significant changes between the (1) and (2) periods (before the effusive phase), whereas a depth increase is evident in (3) and (4) (after the beginning of the effusive phase).

Figure 16.12 Microearthquake swarms observed during 27–28 February 2007 at Stromboli volcano just before and after the 27 February effusive eruption. The circle indicates the start of the effusive eruption.
From Martini et al., 2007.

16.3.3 Mount St. Helens, Cascades

The 2549-m-high dacitic stratovolcano Mount St. Helens began erupting on 1 October 2004 after 18 years of quiescence (see Section 7.1.1). The newly extruded dacite formed a series of spines beginning on 11 October. These spines were repeatedly formed and partially destroyed (Iverson et al., 2006). A sequence of explosions was recorded during 2004–2005 (Moran et al., 2008a, 2008b).

A series of low-frequency and hybrid microearthquakes with similar waveforms began to appear from 26 September and accompanied the 1–5 October vent-clearing phase of small explosions and the initial stage of the lava dome growth that commenced on 11 October. Beginning on 16 October, the microearthquake events appeared to be regularly spaced (1–3 min intervals) which inspired them to be named "drumbeats." Significant variations in drumbeats size and spacing (Figure 16.14B and C) were observed during the 2004–2005 growth of lava spines 1–7, indicating the correlation of the drumbeats characteristics with changes in the

Figure 16.13 The location (above) and the histograms of the hypocentral elevation of microearthquakes (below) before and after the beginning of the effusive eruption (27 February 2007) at Stromboli volcano.
From Martini et al., 2007.

style of extrusion at the surface. At the same time, changes in drumbeat character did not correspond to variations in magma flux at the conduit (Figure 16.14A), indicating that drumbeat size and spacing may be more a function of the mechanics of extrusion than of the extrusion rate (Moran et al., 2008a).

During the 2004–2005 eruption at Mount St. Helens, six explosions were recorded: four of them occurred during the vent-clearing phase from 1 to 5 October 2004, before the lava dome growth, and two explosions occurred on January 16 and March 8 2005, during the dome-building phase (Moran et al., 2008b). The ends of two "big" earthquakes clusters coincided with explosions on 18 January 2005, and 8 March 2005. Figure 16.15 shows a sequence of micro- and big earthquakes, preceding the largest of them, the 8 March 2005 explosion. Some of clusters of "big" earthquakes from the drumbeats swarms occurred in association with the breakup of lava spines and during steady extrusion of one of them (Moran et al., 2008a).

16.3.4 Ubinas Volcano, Perú

The 5672-m-high andesitic-to-rhyolitic stratovolcano Ubinas has been the most active Peruvian volcano for the last 450 years. Its historical activity, documented since the sixteenth century, has consisted of intermittent minor-to-moderate explosive eruptions. The recent unrest at Ubinas began on 25 March 2006 with small ash explosions. A new lava dome in the crater began to grow during April 2006. A sequence of large explosions accompanied by the ash columns and reaching to 7–8 km a.s.l., occurred during May–October 2006; the ash eruptions and steam emissions of different intensity continued to at least January 2009 BGVN (1990–2011); Macedo et al., 2008. Forty-three of 134 explosions were preceded

Figure 16.14 Plot showing the variations in the volumetric lava dome growth rate (A) together with the interevent spacing (B), and peak amplitudes (C) for detected microearthquakes during the 2004–2005 eruption at Mount St. Helens volcano recorded at a distance about 3 km from the crater between 23 September 2004 and 31 December 2005. The intervals of the spines 1–7 (Sp1–Sp7) growths are shown.
Collated using data from Moran et al., 2008a and Vallance et al., 2008.

by sequences of microearthquakes (Macedo et al., 2008). Figure 16.16 shows the fragment of a sequence of microearthquakes preceding the 29 May 2006 explosion. The microearthquakes were quite similar, with the dominant frequencies between 2.8 and 3.6 Hz (Figure 16.17).

16.3.5 Volcán de Colima, Mexico

Microearthquake sequences at andesitic Volcán de Colima were recorded first during its 1991 eruption (Figure 16.18). This earthquake sequence continued for 1 day only, 28 days after the beginning of lava extrusion. The dominant frequency of their seismic records was about 3 Hz (Núñez-Cornú et al., 1994; Orozco-Rojas, 1994). No microearthquake sequences were recorded before or during the 1994 phreatic explosion (Jimenez et al., 1995). During the 1999–2005 sequence of large Vulcanian explosions at Volcán de Colima, the appearance of microearthquake

Figure 16.15 The 8 March 2005 explosion at Mount St. Helens volcano (A) and a sequence of drumbeat microearthquakes recorded by a short-period seismic station at a distance of a few hundred meters from the crater just before the explosion (B).
Photo is taken from Moran et al., 2008b; the seismic records were provided by Stephan Malone (Pacific Northwest Seismic Network).

Figure 16.16 The velocity short-period record (vertical component) of the sequence of long-period microearthquakes before the large 29 May 2006 explosion at Ubinas volcano, Peru (2.5 km from the crater).
The seismic records were provided by Orlando Macedo (Instituto Geofísico del Peru).

Figure 16.17 A family of the 2006 long-period microearthquakes of Ubinas volcano recorded by short-period seismic station (velocity, vertical component) at a distance of 2.5 km from the crater.
From Macedo et al., 2008.

swarms became regular. The microearthquake sequences were recorded a few hours before the largest explosions of 1999 on 10 February, 10 May, and 17 July (Zobin et al., 2002; Vargas-Bracamontes et al., 2009). Most numerous and long-acting microearthquake activity at Volcán de Colima was observed during the March−September 2005 explosion sequence, when 15 explosions with energy $\geq 10^{11}$ J were recorded (Zobin et al., 2006); all these explosions were preceded and accompanied by microearthquake swarms over several days.

Figure 16.19 shows a set of seismograms recorded at station EZV5 during 14 to 16 May 2005 prior to the large ($E = 2.1 \times 10^{12}$ J) explosion of 16 May 2005. The moment on 14 May when the micro-earthquakes began can be seen. In total, 151 events were recorded. The sequence was composed of different types of small seismic signals; the majority of micro-earthquakes lasted between 20 and 30 s. Their rate of appearance increased about 20 hrs after the beginning of sequence (Figure 16.19B). The cumulative number continued to grow gradually until the moment of explosion that occurred on 16 May at 02:06, about 50 hrs after the

Figure 16.18 Seismic activity at Volcán de Colima during February to April 1994. One-day-duration swarm of microearthquakes was recorded 28 days after the beginning of the lava extrusion.
From Orozco-Rojas, 1994.

beginning of the micro-earthquake sequence. The rate of the occurrence of micro-earthquakes strongly decreased after the explosion.

Figure 16.20 shows the cumulative number of microearthquakes during the March and May–June 2005 explosive sequences. It can be seen that every time, some hours before the large explosions, the cumulative number of events sharply increased. The durations of these preliminary intervals varied from 6 to 14 h without any correlation with the energy of the forthcoming explosions. Between the large explosions, the rate of microearthquakes occurrence decreases.

Figure 16.21 shows a set of short-period seismograms recorded during 26–30 September 2004 prior to the 30 September 2004 lava extrusion. The rate of occurrence of microearthquakes slightly increased about 18 h after the beginning of sequence (Figure 16.21B). The cumulative number continued to grow gradually until the beginning of the lava extrusion about 70 h after the beginning of the microearthquake sequence.

16.4 Nature of Microearthquakes

The repetitive waveforms of microearthquakes and their occurrence in families imply a non-destructive, repetitive source mechanism that has a stationary source location. The type of these mechanisms is discussed.

16.4.1 Similarity Between the Microearthquake Waveforms and the Seismic Signals Well-Associated with the Volcanic Events

Simultaneous video images and seismic records of eruptive processes have allowed the characterization of seismic signals here described as microearthquakes (Zobin

Figure 16.19 Seismic records during three days of micro-earthquake activity at Volcán de Colima (short-period seismic station, 4 km from the crater) (A) and the curve of the cumulative number of micro-earthquakes (B) before the 16 May 2005 explosion. The moment of explosion is shown by an arrow. The seismic record of the explosion is marked by an ellipse.
Courtesy of Colima Volcano Observatory.

et al., 2010). Through comparison between the seismic signals, generated by volcanic explosions and rockfalls, and the simultaneous video images (Figures 16.22A and B, 16.23A and B, and 16.24A and B), it was possible to characterize the events. The comparison of identified seismic records of explosions with the seismic

Figure 16.20 Temporal variations in the cumulative number of microearthquakes during the March (A) and May–June 2005 explosive sequences at Volcán de Colima. The moments of explosions are shown by the arrows.

records of microearthquakes of the first and second types shows that the both waveforms, have either a low-frequency first pulse (Figure 16.22; long-period events) and/or a high-frequency first pulse (Figure 16.23; hybrid events), which may be associated with explosion-type sources. Martini et al. (2007) also noted the similarity among the waveforms and spectrograms of volcanic explosions and long-period microearthquakes occurring at Stromboli volcano (Figure 16.25). They suggested that these two types of the seismic signals are actually related to the same source mechanism (gas-slug bursting).

The difference in the spectral content of the initial pulses may be related to different types of trigger of the microexplosions within the crack system, which generates microearthquakes of both types. It may be, for example, the vibration of a crack, containing magmatic fluids, during its filling by gas and steam exsolved from ascending magma. This process is comparable with low-frequency vibrations generated by gas bubbles flowing upward within the pipe (foam model; see Section 3.2.1). In other turn, the initial high-frequency pulse may be a result of brittle fracturing of high-temperature rocks (Tuffen et al., 2003, 2008; see Section 3.2.1) surrounding the fluid-filled crack. The following explosion (or gas-slug burst) would be the result of a piston-like action of the gas-steam jet escaping with a high velocity from the crack. This volumetric process produces the second high-amplitude, high-frequency pulse whose spectral amplitude gives the energy of explosive event.

Figure 16.21 Seismic records during four days of microearthquake activity at Volcán de Colima (short-period seismic station, 1.7 km from the crater) (A) and the curve of the cumulative number of microearthquakes (B) before the 30 September 2004 lava extrusion. The beginning of extrusion is marked by a swarm of rockfall earthquakes on the seismogram of 30 September and is shown by an arrow on the curve.
Seismograms are courtesy of Colima Volcano Observatory.

Figure 16.22 The comparison of the seismic records of an explosion (A), identified with a video image (B), with two records of microearthquakes of the first type (C and D) occurring at Volcán de Colima. Seismic station was situated at a distance of 1.7 km from the crater. Code is yyyy_mmddhhmm. Arrows show the duration of the first pulse. The velocity records are unfiltered and corrected for instrument response.
Courtesy of Colima Volcano Observatory.

A comparison between the seismic records of a rockfall with the seismic records of the microearthquakes of the third type (Figure 16.24) shows a similar spindle-type waveform. Therefore, the third-type microearthquakes may be considered as produced by small rockfalls.

16.4.2 Quantification of Microearthquakes

According to these conclusions, the methodologies developed for quantification of the seismic signals associated with explosive events (Chapter 14) and rockfalls (Chapter 13) were applied to quantify the 2005 microearthquakes occurring at Volcán de Colima (Zobin et al., 2010).

Figure 16.23 The comparison of the seismic records of an explosion (A), identified with a video image (B), with two records of microearthquakes of the second type (C and D) occurring at Volcán de Colima. The seismic station was situated at a distance of 1.7 km from the crater. Code is yyyy_mmddhhmm. Arrows show the duration of the first pulse. The velocity records are unfiltered and corrected for instrument response.
Courtesy of Colima Volcano Observatory.

16.4.2.1 Quantification of Microexplosions

The energy of 418 microexplosions, calculated from the Fourier spectra of high-amplitude, high-frequency pulse (Zobin et al., 2010), ranged over about four orders varying from 1×10^4 to 7.2×10^7 J. Figure 16.26 shows the distributions of the energy for the groups of micro-explosions, occurring during March, 2005 (curve 1) and May–June, 2005 (curve 2). The Kolmogorov-Smirnov test was applied (Figure 16.26B) to show that these two distributions belong to the same sample with the characteristic mean values and, consequently, occurring in the same conditions of their generation. The maximum difference between two cumulative distributions, $D_{n1,\ n2}$, was 0.1644. We checked the significance of D at the level of $a = 0.01$ ($k_{1\text{-}a} = 1.628$) obtaining the critical value of 0.1830, that is greater than the observed

Figure 16.24 A comparison between the seismic records of a rockfall (A), identified with a video image (B), and two records of microearthquakes of the third type (C and D) occurring at Volcán de Colima. The seismic station was situated at a distance of 1.7 km from the crater. Code is yyyy_mmddhhmm. The velocity records are unfiltered and corrected for instrument response.
Courtesy of Colima Volcano Observatory.

$D_{n1,\ n2}$. It allows us to consider that the both databases of micro-explosions belong to the same sample and may be considered as having the same nature of generation.

The comparison of energy distributions of micro-explosions with the energy distributions obtained for 236 small explosions, associated with the September–November, 2004 lava extrusion (Fig. 16.26A curve 3), demonstrates (Figure 16.26C) that they, in their turn significantly differ. The maximum difference between the cumulative distributions, D, is 1.00 (Figure 24C), that is larger than the 0.01-critical value of 0.1402.

Considering the result (Chapter 14, Figure 14.25) showing the difference in the samples of energy distribution for small and large explosions, this suggests that micro-explosions occurred in conditions that differ from those of small and large explosions, and may be generated by different processes within the volcanic

Figure 16.25 Seismic signals (velocity, vertical component) and corresponding spectrograms of explosion earthquake (A) and long-period microearthquake (B) recorded at Stromboli volcano by broadband seismic station during the 2007 activity. After Martini et al., 2007.

conduit. The separation between large and small explosions, as it was shown in Chapter 14) is at a level corresponding to 10^{11} J and the separation between small and micro-explosions is at a level corresponding to 10^{7} J. The maximum difference between the cumulative distributions, D, is 1.00 (Figure 16.26D), that is larger than the 0.01-critical value of 0.4334.

This suggests that microexplosions occurred in conditions that differ from those of small and large explosions, and may be generated by different processes within the volcanic conduit. The separation between large and small explosions is at a level corresponding to 10^{11} J and the separation between small and microexplosions is at a level corresponding to 10^{7} J.

16.4.2.2 Quantification of Rockfall-Type Microearthquakes

Rockfall signals generally represent a small part of microearthquake records. These events occur as a result of the partial collapse of active lava domes or

Figure 16.26 Energy distributions (top panel) obtained for micro-explosions occurring at Volcán de Colima during the March 2005 (1) and the May-June 2005 (2) sequences, and for small explosions of the November 2004 sequence (3) at Volcán de Colima. In the bottom panel are shown the Kolmogorov-Smirnov test comparison cumulative fraction plots for the sequences 1 and 2, and 2 and 3, respectively. The statistic *D* is the maximum vertical deviation between two curves.

flows, or other instabilities on the flanks. The size of rockfall deposit may be approximately estimated from the duration of seismic signal (Chapter 13). Zobin et al. (2005) studied the dependence of the duration of short-period seismic records at a distance of 4–7 km from the crater on the volume of the corresponding pyroclastic flow deposit. It was shown that the short-period seismic signals can be divided into three categories based on their duration: short events with durations less than 100 s; intermediate events with durations between 100 and 250 s; and long events with durations longer than 250 s. It was inferred that long events correspond to pyroclastic flows with mean deposit volume $\sim 2 \times 10^5$ m^3, and intermediate events represent pyroclastic flows with mean deposit volume $\sim 1 \times 10^3$ m^3. Field observations suggest that short events correspond to rockfalls with a mean volume of ~ 50 m^3.

Figure 16.27 shows the distribution in durations of the rockfall-type microearthquake seismic signals recorded during the 2005 explosive activity at Volcán de Colima (Zobin et al., 2010). All of them were characterized by durations less than 100 s, settling them within the category of short events that correspond to rockfalls with a mean volume of ~ 50 m^3.

Figure 16.27 Distribution of the durations of rockfall microearthquakes occurring at Volcán de Colima during the 2005 explosive activity and their position as compared with the durations of short-period seismic signals for different volumes of pyroclastic flow deposits and rockfalls. The mean volume of deposits of pyroclastic flows and rockfalls associated with the seismic records of different durations is shown according to Zobin et al. (2005).
From Zobin et al., 2010.

16.4.3 Nature of Microearthquakes Resolved from Waveform Inversion

Waite et al. (2008) modeled the source mechanisms of low-frequency microearthquakes, occurring at Mount St. Helens, in the 0.5−4 s band using data recorded in July 2005 with a 19-station temporary broadband network. The source mechanism of the low-frequency events, obtained from the waveform inversion, included a volumetric component modeled as a resonance of a gently north-northwest-dipping, steam-filled crack located directly beneath the actively extruding part of the new dome and within 100 m of the crater floor, and a vertical single force attributed to movement of the overlying dome.

In total, study of 68 microearthquakes demonstrated best fit locations and source time functions indicating the repeated activation of the same source. In each of these mechanisms, the moment tensor was dominated by a volumetric component that was consistent with a northerly dipping crack, and F_z dominates the single-force components. Waite et al. (2008) considered that the source of low-frequency microearthquakes does not seem to be within the eruption conduit. The modeled crack could be along a preexisting zone of weakness that separates two older lava flows, for example. The most likely fluid in the crack was steam, owing to the abundant water in the glacier and exsolved from the magma.

17 Acoustic Waves Generated by Volcanic Eruptions

"An earthquake in the air." This was the first reaction of people watching the sudden and unanticipated jerk on the barograph records produced by the cataclysmic 27 August 1883 explosion of Krakatau volcano. The barograms were obtained from 16 observatories situated in various parts of Europe, from St. Petersburg in the east, to Valentia off the Irish coast. The barograph records allowed us to see that the acoustic wave from this explosion had traveled around the Earth not once but seven times. This explosion was distinctly heard up to a distance of about 5,000 km (Scott, 1883–1884; Winchester, 2003). Acoustic waves are generated also during small Strombolian and Vulcanian eruptions and phreatic eruptions or by simple degassing, during the movement of pyroclastic flows and rockfalls. Sustained acoustic signals resembling seismic tremor are also recorded.

In this chapter, the author presents the general characteristics of the acoustic waves observed during different types of volcanic activity together with the simultaneously recorded seismic waves. Making acoustic observations in conjunction with seismic observations at volcanoes allows one to separate shallow or subaerial processes from deeper (purely seismic) processes. The second advantage is that atmospheric acoustic waves travel with different propagation paths than seismic waves and can therefore provide constraints on source versus propagation effects. Another advantage is that infrasonic waves and acoustic-gravity waves can travel far from volcanoes and can be detected by remote stations.

17.1 Infrasonic Acoustic Waves from Small Volcanic Explosions (VEI 1 and 2)

This type of acoustic wave is the most famous to volcanologists and is observed at many volcanoes. Infrasonic is the portion of the atmospheric acoustic wave spectrum below the low-frequency threshold of human hearing (~20 Hz) (Beer, 1974). The commonest source of volcanic infrasonic is the atmospheric perturbation caused by the explosive outfluxes of volcanic volatiles (Johnson, 2003). Infrasonic pressure traces represent the time history of atmospheric pressure perturbations relative to background atmospheric pressure. This excess pressure is usually very small compared to ambient atmospheric pressure (~10^5 Pa). Thus, much volcanic infrasonic can be treated as linear elastic waves rather than nonlinear shock waves.

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00017-7
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Much volcanic infrasonic appears to be dominated by frequencies between 0.5 and 10 Hz. The infrasonic traces that are recorded several kilometers from the explosion source have peak excess pressures ranging from 10^1 to 10^2 Pa. An excess pressure of 10 Pa, that is, the typical infrasonic amplitude recorded 1 km from a Strombolian source, corresponds to the sound pressure level of 115 dB. If this sound were in the audible frequency band, this level of sound would be heard as the noise equivalent of a pneumatic riveter (Johnson, 2003).

17.1.1 Waveforms and Spectra

Figure 17.1 demonstrates the simultaneous acoustic and seismic signals generated by the 15 April 1991 small explosion at Arenal volcano, Costa Rica, and their

Figure 17.1 Seismic (A) and infrasonic (B) signals generated by the 15 April 1991 explosion at Arenal volcano. The Fourier spectra are shown for the seismic signal (C) and the infrasonic signal (D). Rectangle shows the infrasonic signal in B.
Courtesy of W. Melson, Smithsonian Institution.

Fourier spectra. The signals were recorded at the same site at a distance of 2.8 km from the crater. The acoustic signal represents a single impulsive compression with a peak frequency of 7.3 Hz contrasting with the peak frequency of the seismic signal of 1.1 Hz. The airwave arrival corresponding to this signal is seen on the seismic record also.

The infrasonic waveforms may be more complicated depending on the style of the explosive event (Johnson, 2003).

17.1.2 Families of Infrasonic Signals

Infrasonic signals, as with seismic signals from explosions, can form families of similar waveforms. Figure 17.2 shows the family of the seismic signals associated with the September 2003 degassing explosions at Shishaldin volcano (Petersen and McNutt, 2007). The acoustic signals accompanying these seismic signals are characterized by similar waveforms also. The seismic events have a mean cross-correlation coefficient of 0.8, while the acoustic events have a value of 0.9.

Cannata et al. (2009b) applied the cross-correlation analysis to 987 infrasonic events recorded during the September–November 2007 Etna eruption, characterized by lava-fountaining episodes and degassing explosions. The events with correlation coefficients greater than 0.8 were grouped into nine families, which comprised about 95% of all considered events (Figure 17.3). Tuning the cross-correlation coefficient threshold to 0.9, three clusters were recognized. Families from 1 to 5 were grouped in cluster 1, families 6 and 7 in cluster 2, and families 8 and 9 in cluster 3. The infrasonic signals of the three clusters were not only similar in their waveforms, but also grouped in the terms of their quality factor and spectral peak frequency (Figure 17.4). Cannata et al. (2009b) considered that the differences between the waveforms families and especially the clusters were linked to the different source vents and activities as also observed at other volcanoes (Ripepe and Marchetti, 2002).

The existence of waveform families and clusters suggests the repetitive excitation of stationary sources of infrasonic signals.

17.1.3 Source Location of the Infrasonic Events

The source of infrasonic waves may be approximated as a point source and its surface position can be very accurately located by examination of infrasonic arrivals which tend to be impulsive and possess self-similar waveforms across an array (Figure 17.5). The acoustic propagation velocities in the atmosphere are an order of magnitude slower than seismic velocities in the Earth. For arrays, deployed at the epicentral distances within 5 km of the vent, explosion sources can be located with only a few meters error. Figure 17.5 shows that the source of infrasonic waves generated by an explosion at Karymsky volcano coincides with the crater of the volcano (Johnson et al., 2003).

The location of infrasonic events occurring in families may be done using the data of the stacked events belonging to the different families. In the case of the

Figure 17.2 Waveforms of similar seismic and corresponding similar acoustic signals produced during the explosive activity at Shishaldin volcano and recorded at a distance of 6.3 km from the crater. Seismic traces are high-pass filtered (above 0.8 Hz), acoustic traces are bandpass filtered (0.02−3 Hz). The bottom two panels show an overlay of the six events. From Petersen and McNutt, 2007.

nine families of the Etna infrasonic events, described above (Cannata et al., 2009b), the location of the infrasonic source was estimated by a grid-searching procedure over a surface of 1.5×1.5 km, with space of 25 m, centered on the volcanic edifice. Infrasonic signals at the different stations were delayed and compared with the semblance function. The source was located in the node where the delayed signals showed the largest semblance value. As a result, the sources of the infrasonic

Figure 17.3 Correlation matrix of the 987 infrasonic events (A) and the waveforms obtained by stacking the events belonging to nine families of infrasonic events, produced by the explosive activity at Etna volcano and recorded at a distance about 1.5 km from the active craters.
From Cannata et al., 2009b.

Figure 17.4 Relationship between the peak frequency and the quality factor (calculated by Sompi method) for three clusters of infrasonic events of Etna formed by grouping of the nine families of infrasonic events shown in Figure 17.3.
From Cannata et al., 2009b.

Figure 17.5 Source location of acoustic signals determined from acoustic arrivals across an infrasonic array at Karymsky volcano.
From Johnson, 2003.

events of all nine families were associated with two Etna summit craters that were active during September–November 2007 (Figure 17.6).

Among other methods of infrasonic source locating can be mentioned application of the semblance-based method (Ripepe and Marchetti, 2002) and the techniques based on the time lags between station pairs calculated by cross-correlation functions (Johnson, 2005).

All these direct methods located the surface projection of the infrasonic sources but did not allow constraining the infrasonic source depth. To do that, the visual–seismic–infrasonic time delays could be very useful (e.g., Ripepe et al., 2001; Johnson, 2007; Yokoo et al., 2009). At the same time, the estimations of the depth of infrasonic sources, based on the time delay between simultaneous seismic, acoustic, visual, and thermal records, strongly depend on the physical model of the sources chosen by authors and on uncertainties in the velocity models in air and solid within the near zone to the volcanic crater.

17.1.4 Relationship Between the Amplitudes of the Seismic and Infrasonic Signals

Amplitudes of the seismic and infrasonic signals have a general tendency to be correlated, but this relationship may depend on the style of explosive event and its

Figure 17.6 Location of the stacked events of the nine families of infrasonic events of Etna. From Cannata et al., 2009.

surface manifestations. Figure 17.7 shows the plots of the seismic and acoustic parameters for the explosive events observed at Sakurajima and Suwanosejima volcanoes (Iguchi and Ishihara, 1990) and at Karymsky and Erebus volcanoes (Johnson et al., 2003). The explosive events observed at Sakurajima and Suwanosejima volcanoes (Figure 17.7A) demonstrate the general correlation between the maximum amplitude of the compressional phase of infrasonic waves and the maximum amplitude of the seismic signals. Iguchi and Ishihara (1990)

Figure 17.7 Relationships between the amplitudes of infrasonic and seismic waves that were produced by explosion earthquakes at Sakurajima and Suwanosejima volcanoes (A) and Karymsky and Erebus volcanoes (B). In Figure 17.7B, 427 Karymsky 1998 explosions are shown with the gray symbols, 189 Karymsky 1999 explosions with the open symbols, and 64 Erebus 1999–2000 explosions are shown with the black symbols.
Compiled from Iguchi and Ishihara, 1990, and Johnson et al., 2003.

showed that this relationship depends on the style of an eruption. They compare the acoustic and seismic signals produced by large explosions at Sakurajima and Suwanosejima and small explosions at Sakurajima. Within the same amplitude range of the seismic signals [$(5-7) \times 10^{-5}$ m s^{-1}], the amplitudes of infrasonic waves produced by the explosions at Suwanosejima are larger than by small eruptions at Sakurajima and smaller than those produced by the large Sakurajima explosions.

Figure 17.7B displays a relatively fixed relationship between seismic reduced displacement and acoustic reduced pressure for the Erebus explosion events. However, the intensity of Karymsky infrasonic appears to have little correlation with seismic amplitude. Johnson et al. (2003) attribute this variability in energy partitioning to changeable source locations within the conduit and/or muffling of the infrasonic pulse due to an overlying plug of material. They believe that degassed andesite choking the upper portion of the Karymsky conduit possesses a considerably higher viscosity than the Erebus phonolitic lava lake. Later Johnson and Aster (2005) estimated the volcano acoustic–seismic ratio during Strombolian eruptions (which is the ratio of elastic energy propagated through the atmosphere and into the earth) as equal to 8 for Erebus eruptions and equal to 0.2–1.5 for Karymsky eruptions.

Seismo-acoustic study of the explosions that occurred at the andesitic Tokachi-dake volcano, Hokkaido (Okada et al., 1990) showed that the amplitudes of the seismic signals may not correlate with the amplitudes of the airwaves (Figure 17.8) when the surface manifestations of explosive events are different. Okada et al. (1990) recognized that the explosions accompanied by pyroclastic flows were characterized by larger seismic amplitudes with a similar intensity of the airwaves.

Figure 17.8 Relationship between the amplitudes of airwaves and seismic waves that were produced by explosion earthquakes at Tokachi-dake volcano, Hokkaido, during 1988–1989. The amplitude of seismic waves is peak-to-peak maximum amplitude recorded by a short-period seismograph at a distance of about 7 km from the crater; the amplitude of the airwave was measured at a distance of about 3 km from the crater. The filled circles indicate the explosions accompanied by pyroclastic flows. From Okada et al., 1990.

17.2 Long-Period Acoustic and Acoustic-Gravity Waves from Large Volcanic Explosions (VEI 4–6)

Large volcanic eruptions, which form high eruption columns with a great heat injection into the atmosphere, generate large-scale atmospheric pressure perturbations as long-period acoustic waves with periods greater than 270 s or acoustic-gravity waves that are driven by buoyancy forces and propagate in a gravitational field of atmosphere. These long-period waves are characterized by low attenuation and may propagate for thousands of kilometers traveling around the Earth (Beer, 1974; Johnson, 2003).

17.2.1 Near-Field Waveforms of the Long-Period Acoustic Waves

Figure 17.9 shows the long-period 3-h records of the atmospheric pressure perturbations associated with four great Plinian and sub-Plinian (VEI 4–5; Newhall and Self, 1982) explosions at volcanoes Mount Pelèe (1902); Bezymianny (1956); Sheveluch (1964); and Mount St. Helens (1980) collected by Firstov (2007). These waveforms were recorded at the distances from 22 to 54 km from the crater. The maximum period of these records ranges between 580 s for Bezymianny volcano and 1,200 s for Sheveluch volcano, the range of maximum double amplitudes of pressure is between 350 Pa for Mount Pelèe and 2,100 Pa for Bezymianny volcano.

Audible sounds were also reported in association with these great explosions. About the 1902 Mont Pelée explosion: "A dreadful scream, then another much

Figure 17.9 Long-period 3-h records of the atmospheric pressure perturbations associated with four great Plinian explosions at volcanoes Mount Pelèe (A), Bezymianny (B), Sheveluch (C), and Mount St. Helens (D).
From Firstov, 2007.

weaker groan, like a stifled death rattle" (Zebrowski, 2002). Gorshkov and Bogoyavlenskaya (1965) reported that the acoustic signal generated by the 30 March 1956 gigantic explosion of Bezymianny volcano was not heard at near distances from the volcano, at least within 45 km from the crater; only a weak boom was reported. During the 1980 Mount St. Helens great explosion, it was noted that the blast seemed to have been surprisingly quiet in the vicinity of the mount. At the same time, loud noises due to the eruption were heard hundreds of kilometers away (Rosenbaum and Waitt, 1981). A similar phenomenon was reported for the 1883 climactic eruption of Krakatoa, which was heard 5,000 km away, but not on neighboring islands (Symons, 1888).

Firstov (2007) noted that the waveforms for the explosions of Mount Pelèe and Bezymianny volcanoes were similar and were characterized by two compressive and one rarefaction phases. The acoustic records of the Sheveluch and Mount St. Helens explosions were more complicated. Tokarev (1967) considers that the complicated acoustic waveform of the Sheveluch explosion was produced by not a single explosion but by a sequence of explosive events.

17.2.2 Far-Field Registrations of the Long-Period Acoustic Waves

The barograph records allow us to see that the acoustic waves from the explosions travel around the Earth not once but several times. As mentioned in the beginning of this chapter, the acoustic waves from the Krakatoa explosion traveled around the Earth seven times. The acoustic waves from the Bezymianny explosion traveled around the Earth 1.5 times (Passechnik, 1958).

Far-Field Signals of Acoustic Waves

Figure 17.10 shows the copies of the microbarographic records of the long-period acoustic waves produced by the 1980 explosion at Mount St. Helens volcano obtained at the acoustic stations of North America (distances from 927 to 3,950 km) and Japan (distances from 6,970 to 8,240 km). Two impulses in the beginning of all records are seen. The duration of them is about 5–7 min. Mikumo and Bolt (1985) considered that these two impulses are related to the initial stage of the paroxysmal eruption. The first of them was attributed to the generation of the rockslide avalanche and the northward horizontal blast of ash and debris, and the second to a vertical Plinean eruption column.

Tahira et al. (1996) describe the far-field signals of acoustic waves produced by the 15 June 1991 Pinatubo eruption. They discuss the records obtained at Kariya Observatory, Japan, 2,770 km to the northeast of the volcano. Waves that arrived 2 h, 45 min to 54 min after major explosions are interpreted as the waves that propagated along the shorter great circle path in the atmosphere. The strong acoustic signal (the parts of it are shown in Figure 17.11) lasted for almost 10 h. The signals propagated with the velocity between 265 and 280 m s^{-1}; their maximum amplitudes varied between 0.17 and 1.73 Pa.

Figure 17.10 Long-period far-field acoustic perturbations associated with the 1980 explosion at Mount St. Helens volcano and recorded at the acoustic stations of North America (Reed, 1987) and Japan (Sawada et al., 1982).
From Firstov, 2007.

Far-Field Signals of Acoustic-Gravity Waves

A long train of acoustic-gravity waves produced by the 15 June 1991 Pinatubo eruption and associated with extremely large movements of air parcels was recorded at five microbarographic stations in Japan (Tahira et al., 1996). The high-pass filtered waveforms and their power spectra are shown in Figure 17.12. The group velocity of the leading part of the wave is 354 m s^{-1}. The duration of the acoustic-gravity waves is 3 h 50 min that is comparable to the duration of the

Figure 17.11 Acoustic signals recorded at Kariya Observatory from the explosions of Pinatubo volcano on 15 June 1991. The three traces in each panel show the output of three microphones of the tripartite array.
From Tahira et al., 1996.

strong infrasonic signals at Kariya (Figure 17.11). The spectral peaks obtained from the power spectra were noticed at the frequencies of 1.2, 1.9, 2.4, and 4.4 mHz, respectively; the longest period is 13.9 min.

17.3 Acoustic Waves Produced by the Lava Dome Collapse and the Propagation of Pyroclastic Flow and Rockfalls

17.3.1 Dome Collapse

Yamasato (1997) carried out the studies of the acoustic waves produced by pyroclastic flows that followed the dome collapse at Unzen volcano. Figure 17.13 shows the simultaneous seismic and infrasonic records obtained during the dome collapse occurring on 18 October 1992. A small impulsive infrasonic signal was excited at the time of the dome collapse and larger infrasonic waves were excited when the lava blocks fell onto the slope.

Oshima and Maekawa (2001) also performed similar observations at Unzen volcano during the October 1992 pyroclastic flow activity. Their instruments were deployed at a distance of 4 km from the new lava dome. They revealed from the correlation of infrasonic waves with video images, that during the collapse of lava blocks falling from the dome, which generated a pyroclastic flow, the successive impulsive waves and the low-frequency wave (<0.8 Hz) were excited (Figure 17.14). Oshima and Maekawa (2001) supposed that the impulsive wave

Figure 17.12 Acoustic-gravity waveforms from the explosions of Pinatubo volcano, observed at five stations in Japan on 15 June 1991 (A), and their power spectra (B). The data were high-pass filtered to eliminate the incoherent noise with periods longer than 34 min and retain periods shorter than 17 min.
From Tahira et al., 1996.

Figure 17.13 Example of simultaneous seismic (left) and infrasonic (right) signals from a dome collapse at Unzen volcano. Seismic records were obtained by short-period (SP) and long-period (LP) vertical seismographs. Infrasonic records were obtained by low-frequency microphones. The phases A and B are assumed to correspond to the start of dome collapse and the fall of lava blocks onto the slope, respectively.
From Yamasato, 1997.

was excited by an explosive blowout of the pore-gases which were released by the breakdown of the falling blocks and the low-frequency wave was excited by the volumetric change which resulted from expansion of the pore-gases released by pulverization of the falling blocks.

17.3.2 Pyroclastic Flow Propagation

Observations in Near Zone

Figure 17.15 shows the simultaneous seismic and infrasonic records obtained during the pyroclastic flow propagation at Unzen volcano occurring on 5 May 1993 at distances of 1−6 km from the crater (Yamasato, 1997). It is seen that the amplitude variation of the infrasonic signal recorded at these distances was similar to that of the seismic signal, and the infrasonic signals became attenuated when the seismic waves decayed.

Simultaneous seismic and infrasonic records, obtained at a distance of 16 km from the crater during the 13 October 1984 pyroclastic flow propagation at Bezymianny volcano (Firstov and Tristanov, 2010), show the impulsive acoustic signal that arrived in 35 s after the beginning of the seismic signal (Figure 17.16). No similarity in the amplitude variation of the infrasonic and seismic signals recorded at this distance was observed. The spectral peaks for the acoustic signal

Figure 17.14 Comparison of infrasonic and seismic signals with visible events for the lava collapse that generated pyroclastic flow at Unzen volcano on 11 October 1992 at 15:04. The lowest trace indicates filtered infrasonic waves with low-pass Butterworth filter. Thick broken lines (A), (B), (C), and (D) correspond to visible events: (A) a lava block was detached from the dome; (B) the basal part of the falling block landed on the slope; (C) the block violently crumbled from below and explosively discharged the gas–ash–debris complex on the slope; and (D) the block broken into pieces and the complex started to rush down.
From Oshima and Maekawa, 2001.

were at the periods of 2.1, 2.8, and 4.3 s. Firstov and Tristanov (2010) reported that the generation of intense infrasound signal began in about 25 s after entering of the pyroclastic flow from the steep narrow trough into gently sloping wide ravine.

An infrasonic array, installed at 3.5 km from Soufriere Hills Volcano, has successfully detected the infrasonic waves associated with large Vulcanian explosions and the propagation of energetic pyroclastic fronts during May, July, and December 2008 (Ripepe et al., 2010). The array recorded clear signals with durations of 400–1,500 s (Figure 17.17). Coherent infrasonic signal across the array started with an impulsive onset and reached peak amplitudes of 40–80 Pa during the most vigorous phase of the joint explosion–pyroclastic flow event. It is difficult to separate the acoustic signals produced by the explosion and by the pyroclastic flow on these records, obtained near the volcanic crater. Only for the 3 December event, the signal between 1,500 and 2,500 s is large enough to be identified as a record of pyroclastic flow. The infrasound, recorded at the array for this explosion–pyroclastic flow event, was relatively broadband, ranging from 0.4 to 7 Hz.

Figure 17.15 Illustration of an advancing pyroclastic flow at Unzen volcano and the corresponding seismic (SP) and infrasonic (MIC) signals. Stars indicate the source locations of the infrasonic signals.
From Yamasato, 1997.

Figure 17.16 Seismic (A) and infrasonic (B) records, obtained during the October 1984 pyroclastic flow movement at Bezymianny volcano at a distance of 16 km from the crater and the Fourier spectrum of infrasonic record (C). Arrow shows the beginning of the seismic signal. Vertical line shows the beginning of the infrasonic signal.
From Firstov and Tristanov, 2010.

Figure 17.17 Infrasound associated with Vulcanian eruptions and associated pyroclastic flows on 29 May, 29 July, and 3 December 2008 recorded by the infrasonic array installed at 3.5 km from the Soufrière Hills volcano (A) and Fourier spectra of the infrasound waveforms (B). Note the 29 July (blue) and 3 December (black) spectra are peaked at 0.9 Hz, while the smaller event on 29 May (red) shows the largest energy at 2 Hz. From Ripepe et al., 2010, with modifications.

Distant Observations

Firstov and Tristanov (2010) presented the observations of the acoustic signals produced by the 9 May 2006 eruption at Bezymianny volcano when the propagation of pyroclastic flow generated the eruptive plume that reached the height of 11 km a.s.l. (Figure 17.18A and B). These signals were recorded at two acoustic stations situated at the distances of 340 and 361 km from the volcano. For comparison, the seismic signal obtained at a distance of 190 km from the volcano is shown also (Figure 17.18C). Two phases of the eruption, which represented the explosive activity and the propagation of pyroclastic flow (see Section 13.2.3), are well indicated on the infrasound and seismic records. Infrasound record allows to discriminate at least three phases (i_1^1, i_1^2, and i_1^3) within the activity of explosive gas-and-ash plumes.

17.3.3 Large Rockfall Propagation

Very long period (VLP) infrasonic signals (Figure 17.19) were produced by unusually large rockfall that occurred on 29 May 2006 at Mount St. Helens (Moran et al., 2008a). The field observations showed that no explosive activity or lava

Figure 17.18 Acoustic (A, broadband channel, 0.003–0.3 Hz; B, band-filtered record at 0.4–8.0 Hz) and seismic (C, short-period velocity channel) records of the 9 May 2006 explosive event at Bezymianny volcano and associated pyroclastic flow. The phases associated with the gas-and-ash plume activity are marked with $i_1^{1,2,3}$, the arrivals of the pyroclastic flow phase are marked with i_2 at the seismic signal and with e_2 at the infrasonic signal.
From Firstov and Tristanov, 2010.

dome collapse was observed before this event. However, an atmospheric plume had risen to 6,000 m a.s.l. and the seismic signal produced by this event was equivalent to the $M = 3.1$ earthquake (Figure 17.19A and C).

Two large-amplitude high-frequency infrasonic pulses, marked as A1 and A2 in Figure 17.19, were recorded. Accompanying the higher frequency infrasonic signals was a ≈50-s-period VLP infrasonic pulse that can be seen in the raw and low-pass filtered versions of the CDWR microbarometer records (Figure 17.19D and F). The onset of VLP signal is coincident with the arrival of the A1 pulse, suggesting that it is related with the initial collapse of the rock mass. VLP pulse featured an initial pressure increase followed by a pressure decrease before returning to zero (Figure 17.19F).

17.4 Acoustic Waves Produced During Volcanic Microearthquake Swarms ("Drumbeats")

Seismicity during the 2004–2008 eruption of Mount St. Helens has been characterized by a sustained sequence of microearthquakes occurring beneath the dome and

Figure 17.19 Seismic and infrasonic signals associated with large rockfall that occurred on 29 May 2006 at Mount St. Helens. (A) Low-gain vertical-component record from station SEP (0.4 km from the vent). (B) Infrasonic signals recorded by a microphone at station SEP. A1 and A2 labels indicate timing of notable high-frequency infrasonic signals. (C) Broadband vertical component record from station CDWR (13.4 km from the vent). (D) Unfiltered infrasonic signals recorded by a microphone at station CDWR. (E) Same as (D), but filtered with a 1.0 Hz high-pass filter. A1 and A2 labels indicate timing of notable high-frequency infrasonic signals. (F) Same as (D), but filtered with a 0.1 Hz low-pass filter.
From Moran et al., 2008.

named "drumbeats" because of their highly repetitive and regular nature and constant delay time between successive events (see Chapter 16). Matoza et al. (2007) demonstrated that the majority of these seismic events had intermittent infrasonic twins, the infrasonic arrivals with waveforms mimicking the seismic event waveforms and arriving after the seismic event with a velocity and time-delay appropriate for an acoustic wave (Figure 17.20). Matoza et al. (2009a) showed that the acoustic signals were generated by rapid degassing associated with a fluid source of microearthquakes.

Figure 17.20 Seismic and infrasonic recordings of "drumbeats" events observed on 11 November 2004 at Mount St. Helens at a distance of about 13 km from the summit. One hour of vertical seismic (top, red) and beam-formed infrasonic (bottom, black) recordings of "drumbeats" events are shown in (A). Both series are filtered 2–4 Hz. Box indicates the time range covered in (B). (B) displays 500 s of vertical seismic (top, red) and beam-formed infrasonic (middle, black) events from the time indicated in (A).

From Matoza et al., 2007.

17.5 Utility of the Acoustic Signals for Volcano Activity Monitoring

17.5.1 Estimation of the Energy of Eruptive Events

Large Explosions

The source energy from the microbarographic records was estimated for several large volcanic eruptions. For the 1980 Mount St. Helens and the 1991 Pinatubo

explosions, the energy was obtained from the far-field records of acoustic-gravity waves (Donn and Balachandran, 1981; Tahira et al., 1996). An empirical equation proposed by Posey and Pierce (1971) was used to estimate the source energy from the microbarographic records:

$$E = \frac{\{13P[R\ \sin(l/R]^{0.5}H(c\tau)^{1.5}\}}{(4.22 \times 10^{15})} \tag{17.1}$$

where E is the energy release in explosive equivalent of trinitrotoluene (TNT) in megatons (Mt), P is the first peak-to-peak amplitude (Pa), R is the radius of the Earth (m), l is the great circle between the source and receiver (m), c is the sound speed (m s^{-1}), and τ is the time interval (s) between the first and second peaks. The energy of the 1980 Mount St. Helens was estimated as 35 Mt of TNT and the energy of the 1991 Pinatubo explosions amounts to about 70 Mt of TNT.

Gorshkov and Bogoyavlenskaya (1965) calculated the energy E (in ergs) of airwaves generated by the 1956 Bezymianny explosion from the near-field barograph records (45−780 km from the volcano) using the following equation:

$$E = 1.25 \times 10^{20} \sin\varphi \sum \frac{A^2}{2} \tag{17.2}$$

where φ is the distance from the explosion source in degrees, and A is the amplitude of airwave in mbar. The energy of airwaves was estimated as 3×10^{22} ergs (about 1 Mt of TNT). Gorshkov and Bogoyavlenskaya (1965) suppose that the energy of airwave represents about 0.1% of the total energy. Then the total energy of this explosion was estimated as 1.2×10^{24} ergs, or about 30 Mt of TNT.

For comparison, the Little Boy atomic bomb dropped on Hiroshima on 6 August 1945 exploded with energy of about 15×10^{-3} Mt of TNT. The energy of the 1908 Tunguska meteoroid explosion was estimated as 30 Mt of TNT.

Small Explosions

The acoustic energy of small explosions was estimated for some explosive events considering that the acoustic waves were radiated spherically and that the total acoustic energy was proportional to the time-integrated squared excess pressure trace. Using the following equation:

$$E_{ac} = \frac{2\pi r^2}{\rho_a c} \int \Delta P^2 \ \mathrm{d}t \tag{17.3}$$

where r is the distance between the source and receiver (m), ρ is the air density (kg m^{-3}), and c is the sound speed (m s^{-1}), Johnson (2003) estimated the acoustic energy from the infrasonic signals recorded at the distances from 0.3 to 15 km for 12 small Strombolian explosive events recorded at 9 worldwide volcanoes. The

acoustic energy, representing a very small portion of the total explosion energy (about 0.1%, as noted by Gorshkov and Bogoyavlenskaya (1965)), ranges between 10^4 and 1×10^8 J. The estimations of the acoustic energy for some Strombolian and Vulcanian explosions at Villarica, Stromboli, Santiagito, and Fuego volcanoes spans five orders of magnitude, varying from 10^2 to 10^8 J (Marchetti et al., 2009).

Pyroclastic Flows

Eq. (17.3) may be used also for the estimation of the energy of infrasonic waves produced by moving pyroclastic flow. Yamasato (1997) obtained the energy of infrasonic waves from the 1993 Unzen pyroclastic flows of more than 3 km duration. It ranged between 1.2×10^6 and 4.7×10^7 J and was almost the same order as the seismic energy.

17.5.2 Reconstruction of the Process of Dome Collapses and Pyroclastic Flow Movement

Figure 17.21 shows the distribution of infrasonic source locations for dome collapses and for following pyroclastic flows of Unzen volcano for the interval from June 1992 to November 1993 (Yamasato, 1997). The sources of the infrasonic waves were estimated from waveform correlation and amplitude distribution. Pyroclasts began to excite infrasonic and seismic waves when they were fragmented by collision. Both sources were located near the front of pyroclastic flows. The

Figure 17.21 Location of the sources of infrasonic signals from pyroclastic flows at Unzen volcano that occurred from June 1992 to November 1993. Closed circles indicate the sources due to dome collapses and open circles indicate the sources of later phases that arise from the flow of pyroclastic. The hatched area indicates the locus of the lava dome. Stars indicate the locations where the amplitudes of the later phases of the seismic and infrasonic signals reached a peak.
From Yamasato, 1997.

Figure 17.22 Examples of infrasonic waveforms of the 2000 Usu phreatic explosions (B) recorded by the network (A) on 18 May 2000.
From Yamasato et al., 2002.

infrasonic sources corresponding to pyroclastic flows show migrations along the pyroclastic flows pass trajectories. It allows determining the direction of a pyroclastic flow from analysis of infrasonic data even if visual observation is impossible.

17.5.3 Monitoring of Phreatic and Strombolian Explosions

The sequences of the infrasonic signals were recorded during the 2000 phreatic explosions at the Usu flank craters K and N (Yamasato et al., 2002). The pulses were excited at intervals of several seconds (Figure 17.22A). Using the records from the infrasonic network, consisting of 10 stations and installed around the volcano (Figure 17.22B), source location of the pulses generated from two active craters K and N were identified. The amplitude of the infrasonic pulses reflected the activity of the phreatic eruption from these craters and they became smaller with the decline of the phreatic explosions at the active craters (Figure 17.23).

Figure 17.23 Amplitude variations of the infrasonic pulses observed at ABUT station (see Figure 17.22B) during the 2000 Usu phreatic explosions. Closed and open circles indicate the infrasonic phases located at the N and K craters areas, respectively. From Yamasato et al., 2002.

Figure 17.24 Acoustic records of three types of Strombolian explosions at Karymsky volcano during 1997 activity and their spectrograms. (A) Simple impulse event, (B) high-frequency event, and (C) "chugging" event. Spectrograms are calculated with 10-s moving windows at 2-s increments and are bandpassed between 0.25 and 12.5 Hz. Code of date is yy:ddd:hh:mm, ddd is a day from the beginning of year.
From Johnson et al., 2003.

Figure 17.25 Infrasonic signals recorded during Vulcanian–Plinian volcanic eruptions: A, the 8 March 2005 Mount St. Helens eruption recorded at 13.4 km; B–D, the Tungurahua eruptions recorded at 36.9 km (B, 14–15 July 2006; C, 16–17 August 2006; and D, 6 February 2008). Each trace represents a time-domain beam across four array elements with unit gain. Discrete explosion occurrences are marked by "e." The signal labeled "M" on the third trace is the Mount St. Helens eruption [see (A)] plot at the same scale for comparison. From Matoza et al., 2009b.

Johnson et al. (2003) showed the utility of the infrasonic monitoring of Strombolian explosions at Karymsky, Sangay, Tungurahua, and other volcanoes. The infrasonic signals recorded at Tungurahua volcano indicated a continuous style of degassing, consisting of both explosive pulses and quasi-continuous "jetting." At Karymsky volcano, where three types of acoustic signals were recognized (Figure 17.24), the successive explosions, whether they were simple impulse, high-frequency, or chugging events, tended to vary according to the ease with which gas is able to escape to the free surface. The chugging at Karymsky and Sangay volcanoes was associated with visible degassing.

Matoza et al. (2009b) investigated long-duration broadband infrasonic signals associated with sustained phreatic and magmatic Vulcanian, sub-Plinian, and Plinian explosions. Signals last many hours in duration in some cases (Figure 17.25). Signals were attributed to a low-frequency form of jet noise. That is, the sustained jet flow in the gas-thrust region at the base of eruption columns is thought to radiate sound in a similar way to man-made jet flows, such as those that come from jet engines or rockets. Because volcanic jet flows were very large in comparison to man-made jet flows, frequencies of radiation are much lower, hence Matoza et al. (2009b) considered them as infrasonic.

18 Seismic Monitoring of Volcanic Activity and Forecasting of Volcanic Eruptions

Chapters 5–10 of this book described and discussed the properties of volcano-tectonic earthquakes, and demonstrated the temporal and spatial regularities in the earthquake sequence development before volcanic eruptions. It allows forecasting of volcanic eruptions using seismic data. This chapter discusses the special aspects of seismic monitoring of volcanoes and the methodology of forecasting of volcanic eruptions.

18.1 Methodology of Seismic Monitoring of Volcanic Activity

Seismic monitoring of volcanic activity serves to forecast an eruption and to issue an alert of a volcanic hazard to the public (Shimozuru, 1971; McNutt, 1996, 2000b; McNutt et al., 2000). Therefore, the system of monitoring must be capable of precisely locating seismic events in the region of a volcano and able to quickly process the obtained information. To resolve this problem, a seismic network has to consist of a minimum of three to five seismic stations, with a spacing of a few kilometers and they need to be installed within 1–10 km from the active crater. Digital and analog seismic records are usually telemetered to a center of monitoring for processing.

18.1.1 Seismic Networks Around Volcanoes

A typical seismic network around an active volcano, Volcán de Colima, México, is shown in Figure 18.1. It consists of five short-period vertical-component seismic stations distributed at a distance from 1.3–7.0 km and covering a broad azimuthal range. The stations are situated within the altitude range from 1,700–4,000 m; it is possible to locate earthquakes of magnitude 0–4, which occur within the edifice of this volcano. The center of telemetrical monitoring is situated in Colima city at a distance of about 30 km from the stations. The analog data are transmitted by radio to the center where they are recorded on drums and stored as digital time series (Castellanos and Jiménez, 1995). This seismic network is of minimal capability; the

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00018-9
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Figure 18.1 The position of seismic stations (A) and their amplitude responses (B). The stations EZV3–EZV7 are shown as triangles; the center of telemetrical monitoring is shown by cross. Shown are the contour lines for 2,500 and 3,000 m. VC is Volcán de Colima; NC is Nevado de Colima. Lines labeled 1 and 2 in B are two types of seismic station responses. From Zobin et al., 2002b.

destruction of any seismic stations during an eruption could stop the monitoring. The absence of three-component stations lowers the precision of locations.

A good illustration of the importance of three component stations and the use of both *P*- and *S*-waves for volcanic earthquake location is shown in Figure 18.2. The 40 best-constrained earthquakes recorded at Etna volcano, Sicily are relocated in the absence of some essential data (Ferrucci, 1995). The seismic network consists of 13 stations, 11 of them are of three components. Location of events by all stations (Figure 18.2B) constrains seismicity to within a focal volume of less than 12 km^3 and confines it between sea level and a depth of 2 km. In Figure 18.2C, the seismic cluster is shown after excluding of the two three-component stations nearest to the crater, that were situated above the seismically active volume. It causes a

Figure 18.2 Illustrations of the importance of the use of three-component stations in seismological routine and the adequate positioning of the stations according to the seismic volume. (A) seismic network (triangles, three-component stations; circles, one-component stations); (B) epicenters (left) and hypocenters (right) computed using all the *P*- and *S*-wave data; (C) epicenters (left) and hypocenters (right) computed without the two three-component stations nearest to the crater that were situated above the seismically active volume; and (D) epicenters (left) and hypocenters (right) computed using only the *P*-wave data.
From Ferrucci, 1995.

dramatic spreading of the epicenters toward the southeast and an increase in the depth interval. In Figure 18.2D, all stations are included but only *P*-wave data have been used; the result being only a weak effect on the epicentral distribution, but it strongly increases the depth interval.

Recently, many seismic networks at volcanoes are equipped with three-component digital broadband seismometers that give greater potential for detailed investigations of volcanic earthquakes (Chouet, 1996b; Aster et al., 2000).

18.1.2 Initial Processing of Seismic Data

The processing of seismic data is realized in real time and afterward the earthquakes are localized. For real-time processing, the digital and drum-recorded seismograms and spectrograms are used, as well as the compact presentation of seismic information with the RSAM (Endo and Murray, 1991) or RSEM (De la Cruz-Reyna and Reyes, 2001) methodologies. Figures 18.3–18.5 show these three types of presentation of seismic information. A seismogram of volcanic earthquakes in an active period (Figure 18.3) gives the general impression of the different types of the seismic signals associated with volcano-tectonic earthquakes, explosions, rockfalls, or tremor. At the same time, it can be overfilled by a sequence of numerous and saturated signals. The spectrogram (Figure 18.4) helps to separate low- and high-frequency events, to see the temporal variations of these signals, as well as the period of volcanic tremor.

Figure 18.3 Seismogram recorded at the distance of 1.7 km from the crater of Volcán de Colima, México during the 2002 lava eruption. The numerous seismic signals were produced by rockfalls and pyroclastic flows.
Courtesy of Colima Volcano Observatory.

Figure 18.4 Spectrogram of the same record as in Figure 18.3. Courtesy of Colima Volcano Observatory.

Figure 18.5 Variations of RSEM during the 1997–2006 unrest at Volcán de Colima. The episodes of volcano-tectonic earthquake swarms (VT), lava extrusions (EFF), and Vulcanian explosions (EXP) are discriminated.
Courtesy of Colima Volcano Observatory.

The RSEM (real-time seismic energy measurement) (Figure 18.5) or RSAM (real-time seismic amplitude measurement) (Figure 18.6) curves give the characteristics of strain release associated with seismic activity. The RSAM (RSEM) continuously samples the absolute amplitude (or square amplitude) of seismic signal; the average of each signal is obtained summing the measurements made in a certain time interval and dividing by the number of measurements (or unit time).

18.1.3 Automatic Classification of the Seismic Signals

Automatic detection and classification problems play a critical role in the branch of seismology devoted to the continuous monitoring of seismogenic zones and active volcanic areas. The classification of seismic events should be done as close to real time as possible in order to reduce the response time in case of a volcanic alert. During the past decade, a new generation of automatic systems has been developed that allows very advanced analysis of the seismic records from the telemetered data in near real time (Esposito et al., 2006; Ibañez et al., 2009).

The techniques developed to automatically discriminate between different kinds of seismic signals range from statistical analysis to cross-correlation and wavelet analyses. Limitations of deterministic models have led to more sophisticated methods that incorporate artificial neural networks and discrete hidden Markov (HMM) modeling tools to

Figure 18.6 The 4-h average RSAM values showing the marked division between the metastable and pre-explosive buildup seismic phases of the 1991 eruption of Pinatubo volcano, Philippines (see Chapter 7).
From Harlow et al., 1996.

implement the classifier. These methods offer adaptation mechanisms to learn from the signals and assign model parameters (Ibañez et al., 2009; Curilem et al., 2009).

A supervised neural system was applied to discriminate among landslide, explosion-quake, and volcanic microtremor signals during the 2002 activity at Stromboli volcano (Esposito et al., 2006). The data were pre-processed to obtain a compact representation of the seismic records. Both spectral features and amplitude-versus-time information have been extracted from the data to characterize the different types of events (Figure 18.7). As a second step, they have set up a supervised classification system, trained using a subset of data (the training set) and tested on another dataset (the test set) not used during the training stage. The number of hidden units and training cycles has been chosen empirically by trial and error.

The automatic system was able to correctly classify 99% of the events in the test set for both explosion-quake/landslide and explosion-quake/microtremor couples of classes, 96% for landslide/microtremor discrimination, and 97% for three-class discrimination (landslides/explosion-quakes/microtremor). A useful tool for analyzing classification results is the confusion matrix (Figure 18.8), which contains information not only about the system performance, but also about misclassified versus correctly classified data in the test set. For the explosion/landslide discrimination (Task A), all explosions were correctly classified, whereas just one landslide was

Figure 18.7 Three typical signals used for automatic classification of the Stromboli seismic events are represented both in the time and in the time-frequency (spectrogram) domains. Whereas the explosion-quake (A) and the microtremor (C) signals exhibit a limited frequency content (mostly in the 1−6 Hz range), the landslide related signal (B) has a broader spectrum. From Esposito et al., 2006.

Task A
Network CM99.61%

	Explosion-quakes	Landslides
Explosion-quakes	156	0
Landslides	1	104

Task C
Network CM99.40%

	Explosion-quakes	Microtremor
Explosion-quakes	167	2
Microtremor	0	165

Task B
Network CM97.43%

	Landslides	Microtremor
Landslides	101	3
Microtremor	4	165

Task D
Network CM98.39%

	Landslides	Explosion-quakes	Microtremor
Landslides	97	0	4
Explosin-quakes	0	167	1
Microtremor	2	0	164

Figure 18.8 Confusion matrix on the test set for best network realization for each type of classification task at Stromboli.
From Esposito et al., 2006.

misclassified (or confused) as an explosion. Similarly, for Task C (explosion/microtremor), two explosions were confused, whereas all microtremor signals were correctly identified. In Task B (landslide versus microtremor), three landslides were confused and four microtremor events were misidentified. The network performed well for these two-class tasks, and it also did well in the three-class case, where explosions, microtremor, and landslides are jointly discriminated (Task D).

A HMM-based seismic-event recognition system decomposes the incoming seismic signal as a sequence of feature vectors. The length of the analysis window is such that the seismic waveform can be considered as stationary. The recognition system performs a mapping between a sequence of feature vectors and the corresponding sequence of seismic events. To obtain a useful dataset, the continuous seismic waveform must be processed in two stages. The pre-processing stage is the selection of samples corresponding to the vertical velocity channel and conversion into a one-dimensional stream of 16-bit binary. The following feature extraction stage converts the volcano seismic signal into a parametric representation with less redundant information (Benitez et al., 2009).

Cortés et al. (2009) had applied the automatic continuous HMM-based recognition system to the seismic events recorded at Volcán de Colima. A dataset was collected at two short-period monitoring stations plus one broadband station mainly during the period of 2004–2006, with some events from 1998. The acquisition system was attached to the widely used Earthworm software, developed by the USGS (U.S. Geological Survey), which continuously digitizes the seismogram waveform. In order to set up the system and check database reliability, continuous close and blind isolated event recognition tests were performed: close tests used the same datasets to train the models and to evaluate them, while in blind tests the training dataset was not used for the evaluation. To improve the reliability of the results, and minimize the negative effect in the HMM from a lack of events, blind recognition is carried out, averaging the results of three tests where alternatively two datasets were used to train models and the third to evaluate. The recognition results are evaluated using the %correct and %accuracy rates (%Corr, %Acc), number of correctly recognized, substituted, and deleted events.

Table 18.1 Confusion Matrix for Colima Events and Continuous Blind Test Results for Dataset 1

	SIL	EXP	LP	REG	T1	T2	T3	COL	LAH	VT	*DEL*	%COR	%ACC
SIL	*321*	0	1	1	1	10	0	4	2	0	47	85.15	81.96
EXP	1	*186*	0	1	0	0	0	1	0	2	0	97.38	91.10
LP	1	2	*173*	0	0	0	3	0	0	7	3	91.53	89.42
REG	0	3	0	*111*	0	0	1	1	0	0	0	95.69	89.66
T1	0	2	0	10	*50*	0	0	0	0	1	13	65.79	57.89
T2	1	1	0	2	0	*6*	0	0	0	0	0	60.00	60.00
T3	2	1	5	1	0	0	*194*	1	0	1	29	82.91	79.91
COL	1	0	0	0	0	0	1	*183*	4	3	36	80.26	75.44
LAH	0	0	0	0	0	0	0	0	*26*	0	0	100.00	42.31
VT	0	2	10	1	0	0	2	0	0	*112*	1	87.50	85.94

From Benitez et al., 2009.

The following basic classes of the seismic events were identified: volcanic tremor (T) of different types (T1–T3), volcano-tectonic earthquakes (VT), regional earthquakes (REG), collapses (COL) or rockfalls, lahars (L), long-period events, and explosions (EXP). The confusion matrix, obtained for one of three sets of data (Table 18.1) during continuous blind recognition of the shape of the seismic signals, was close to a diagonal matrix (which corresponds to a 100%Corr). The total number of recognized events in this set was 1,575; the number of correctly recognized events was 1,362; the number of substituted events was 84, and the number of deleted events was 129. The diagonal component of the matrix achieves a successfully 81.65%Acc for Colima. Figure 18.9 plots the continuous event classification carried out by the whole system during September 2008 to January 2009.

18.1.4 Location of Seismic Events

For more detailed analysis of the situation, we need the maps and cross-sections of the focal distribution of the earthquakes. Figure 18.10 illustrates these characteristics of seismic regime for Volcán de Colima during the May–June 2001 initial stage of lava dome growth. The events that occurred in May and June are marked by different symbols. This map and cross-section show that the June events became shallower and may reflect the intensification in lava dome growth.

18.2 Applications of Volcanic Seismicity to the Forecasting of Volcanic Eruptions and Predicting of Volcanic Hazards

The case studies of Chapters 5–7 show that volcano-tectonic seismicity occurs before the majority of volcanic eruptions, indicating the site of the forthcoming

Figure 18.9 Automatic continuous classification of events at Volcán de Colima between 2 September 2008 and 14 January 2009 showing long period (blue line), explosions (red), and tremor (green) events activity.

From Cortés et al., 2009.

event, F. Omori wrote in May 1911, after the first historical instrumental study of seismic and volcanic activity during the 1910 Usu, Hokkaido eruption: "I believe the problem of forecasting of great volcanic eruption is, in some cases, not very difficult" (Omori, 1911–1913). He wrote later (Omori, 1914): "Unlike large destructive earthquakes, which originate along seismic zones, but are not repeated from one and the same center, the eruptions of a given volcano take place usually from the central crater or from new side vents, being anyhow thus localized to the mountain itself. Hence, it may be that the great outbursts from a given volcano at widely different epochs resemble each other more or less, both in the eruptive phenomena and in the precursory events."

18.2.1 Methods Based on the Statistical Variations in the Parameters of Volcano-Tectonic Earthquakes

Minakami was a pioneer in the development of the seismic statistics method (Minakami, 1960, 1974b). The studies were based on the monitoring of seismo-volcanic activity at the andesitic Asama volcano, Honshu. His famous diagram showing the

Figure 18.10 Epicenters (A) and hypocenters (B) of small volcano-tectonic earthquakes located at Volcán de Colima, México during May (filled circles) and June 2001 (double circles) at the initial stage of lava dome growth. The stations EZV3 and EZV4 are shown as triangles. The contour lines of 3,000; 3,500; and 4,000 m are shown. VC is Volcán de Colima; NC is Nevado de Colima.

relationship between the number of B-type volcano-tectonic earthquakes and the daily frequency of volcanic events observed at this volcano (see Figure 2.2) became classic. It was a typical example, in which premonitory earthquakes increased in number before an eruption. Minakami proposed the empirical laws (1960 and 1966) for

Figure 18.11 The empirical curves formulae derived in 1960 and 1966 for predicting the Asama eruptions based on the 5-day frequencies of B-type earthquakes. *P* is the probability of eruptions in the next 5 days; *N* is the seismic frequency of past 5 days. From Minakami, 1974b.

predicting the Asama eruptions based on the 5-day frequencies of B-type earthquakes (Figure 18.11).

These laws predict that with an increasing number of B-type earthquakes, there is an increasing probability of an eruption. The practical application of the 1960 formulae is shown in Figure 18.12. The problem of forecasting was dealt with using the seismic frequencies of 5 successive days, as follows:

$$\begin{aligned} n_1 + n_2 + n_3 + n_4 + n_5 &= N_5 \\ n_2 + n_3 + n_4 + n_5 + n_6 &= N_6 \\ n_3 + n_4 + n_5 + n_6 + n_7 &= N_7 \end{aligned} \tag{18.1}$$

where n_1, n_2, n_3, etc., indicate seismic daily frequencies F_2 of earthquakes and N_5, N_6, N_7, etc., indicate seismic frequencies N of 5 successive days. So if the value of N on first day is, for example, 800, the curve 1960 of Figure 18.12 gives the probability P of an eruption within the following 5 days as 0.85, showing a very dangerous state of the volcano (Minakami, 1974b).

Tokarev has developed this methodology for the forecasting of volcanic eruptions in Kamchatka using, not only the earthquake frequency, but also the seismic strain release (Tokarev, 1963, 1971, 1978, 1983, 1985). For andesitic volcanoes, he proposed the hyperbolic law of cumulative strain release ε:

$$\varepsilon_t = \frac{at + b}{t - c} \tag{18.2}$$

where a, b, and c are the coefficients for the given eruption and the time t is expressed in days. Eruptions occur in a period Tu$\approx c$. Figure 18.13 illustrates this methodology. Coefficient c that was large at the initial stage of earthquake swarm strongly diminished with the approach of the eruption. In practice the date of the forthcoming eruption was estimated on the 40th day of activity, 1 week before the eruption.

Recently, this statistical methodology was strongly improved by including the material failure model (FFM) as a theoretical basis for the forecasting of an eruption

Figure 18.12 Practical application of the 1960 formula to eruptions that took place in 1961, based on the daily frequency of B-type earthquakes. Arrows indicate the eruptions; P is the probability of eruption given from the 1960 formula; F_2 is the daily frequency of B-type earthquakes originating from Asama.
From Minakami, 1974b.

(Voight, 1988; Voight and Cornelius, 1991; Cornelius and Voight, 1994, 1995). This FFM method is based on the empirical law describing the failure of materials. It is proposed that the acceleration (the second derivative) of any variable of state Ω, which could be a possible precursor such as seismicity, deformation, or gas emission, is proportional to the rate (the first derivative) of this variable to the order of α (Eq. (18.3)):

$$\Omega'' = Ax\Omega'^{\alpha} \tag{18.3}$$

The empirical coefficient of proportionality A and the constant α are dimensionless. For $\alpha > 1$, there is a time of singularity, or the time at which the rate theoretically becomes infinite, and is therefore an upper bound to the time of failure (or the onset of the eruption, in our case).

The graphical technique of forecasting is based on a plot of the inverse rate of the variable of state versus time (Figure 18.14). The cumulative values of RSAM or SSAM as well as cumulative coda duration or seismic amplitudes can be used as a seismic variable. When the cumulative seismic variable begins to increase sharply, the inverse rate of variable goes to zero. The estimation of a time of the onset of an eruption is based upon successive estimations of the apparent failure times from the approximation of the inverse rate curve according to Eq. (18.4):

$$(\Omega''*)^{\beta} = Ct + D \tag{18.4}$$

Figure 18.13 Practical application of Tokarev's method for forecasting eruptions of andesitic volcanoes. The variations of cumulative strain release (A) and coefficient c (B) before the 12 April 1960 Bezymianny volcano, Kamchatka eruption are shown. Here t_0 is the time of the first earthquake in the swarm. The strain release is expressed in units of $[10^3] \times J^{1/2}$; c is expressed in days. Arrows show the time of beginning of the eruption Tu. From Tokarev, 1983.

where $\beta = 1 - \alpha$; C is a slope of the inverse rate curve, and D is the intercept. The time of singularity, that is, the occurrence of an infinite rate, is found by extending the inverse rate data toward the time axis. A finite peak rate at the time of failure may be assumed, or estimated from experience from earlier eruptions. This finite rate, plotted as a horizontal line in the inverse rate plot, can be used for the intersection with the extending trend of the inverse rate. The interval between the apparent time of failure and time of eruption is termed the delay interval.

Figure 18.15A–D shows the applications of the method for three types of seismic variables. These seismic data were observed before the 22 October (Julian day

Figure 18.14 Schematic inverse rate curve. Open circles define inverse downward (accelerating) rate trends, solid dots decelerating jogs. Decelerating jogs may cyclically alternate with accelerating jogs (dashes and arrows), which may be steeper than the overall trend (heavy solid line). Data points beyond the apparent peak rate were excluded from analysis (crosses). Inverse rate trends can be fitted numerically and a data envelope (curved lines of small dots) can be constructed at a specified confidence level. The intersection of the data envelope with a range of expected rates at time of eruption (horizontal lines of small dots), as derived from previous eruptions, may be used to define an eruption window (vertically dashed lines). Alternatively, the time of eruption may be defined in terms of an expected apparent peak rate, plus a delay interval. Time of singularity t_s is defined by the intersection of the curve fit with the abscissa, implying infinite rate (square).
From Cornelius and Voight, 1994.

295.5) 1986 eruptive event at the lava dome of Mount St. Helens. The inverse rates for all variables (coda duration (A), seismic wave amplitude (B), and RSAM (D)) were fitted to Eq. (18.4) using the linearized least-squares technique. The α-values derived for the inverse rate curves are comparable (1.87 for coda accumulation; 1.67 for amplitude accumulation; and 1.38 for RSAM).

The rates used for the curve fitting of A and B (solid dots in A and B) were calculated over periods of 2 days, as daily accumulations. Calculating the rates over shorter periods will usually increase data scatter but is feasible when rate becomes larger. Examples (circles) in Figure 18.15 (A and B) are for accumulations per day calculated over periods of 1 day between October 14 and 19 (Julian days 288−293). Distinct downward trends in the inverse plot are thus recognizable 2.5 days before the extrusion. Eruption began around day 295.5. The forecasting windows were $294.6 < 294.9 < 296.1$ for A curve and $295.0 < 297.2 < 301.6$ for

Figure 18.15 Illustration of the FFM method applied to eruption prediction (the 22 October 1986 Mount St. Helens volcano lava dome extrusion). Plots of cumulative coda duration (A), cumulative amplitude (B), and cumulative RSAM (C) are shown with their inverse rate curves (A, B, and D, respectively). Dashed lines in (A) and (B) show the fitted curves. St. Helens West and Garden are seismic stations of the Cascade Volcano Observatory. Solid dots are the rates calculated over period of 2 days; open circles are the mixed dataset of rates calculated over periods of 2 and 1 days. Dotted vertical lines mark onset of extrusion. From Cornelius and Voight, 1995.

B curve. The forecasting window is based on 97.5% confidence level for linearized fits according to FFM.

For the RSAM system (Figure 18.15C and D), the data were arbitrary averaged over 3-h periods. The data window chosen for the RSAM analysis ends at the same time as one data window used for the analysis of coda and amplitude. The forecasting window for RSAM is $296.3 < 296.8 < 297.6$.

These methodologies work if the eruption occurs at the moment of maximum seismic strain release during volcano-tectonic earthquake swarm. The cases described in Chapters 5–7 show that this type of activity is common for andesitic and dacitic volcanoes, but it is rare for basaltic volcanoes when the eruption may begin after the end of seismic activity.

The FFM methodology may be applied also to the development of micro-earthquake swarm (Chapter 16) before an explosion. Arámbula-Mendoza et al. (2011) applied this method to the micro-earthquake signals, preceding the 16 September 2005 explosion at Volcán de Colima, using Seismic Spectral Energy Measurement (SSEM) as a variable parameter. Figure 18.16 illustrates the methodology. The time

Figure 18.16 Application of the material failure forecast method to Colima seismic data before the explosion of 16 September 2005. It is shown the time evolution of the inverse of the SSEM (A) and the seismic record of the sequence of micro-earthquakes (B). The arrow in A and ellipse in B indicate the actual occurrence of the explosion.
From Arámbula-Mendoza et al., 2011.

evolution of the inverse of SSEM, that was computed for the spectral band between 1 and 3 Hz using a second order Butterworth bandpass filter, was studied using 10-min-long windows. The explosion was preceded by a 3 day-long swarm of micro-earthquakes. During this period, the inverse SSEM displays a clear decreasing tendency. The adjusted straight line intersects the axis about 10 h after the explosion (Figure 18.16A).

18.2.2 Chronicle of Some Forecasting of Volcanic Eruptions Based on Seismic Monitoring

The methodologies described above are of interest, but they are very instable and may be applied for the limited number of well-known volcanoes. The problem is that the eruptive activity of a volcano changes the statistical regularities in its seismic activity without any announce. As a result, the methodology, serving well during 3–5 years, may be wrong with the change in the eruptive behavior of the volcano. For example, the regularities in the seismic stress release before the 1957–1970 eruptions at Bezymianny volcano, Kamchatka, associated with the lava dome growth episodes (Tokarev, 1981; Figure 18.13 shows this regularity before the 1960 eruption) were changed in the beginning of 1970s. Chubarova (1995) wrote that the magnitudes of seismic events preceding the volcanic eruptions

significantly decreased during 1971–1994 (they were even completely absent before some of them), and the forecasting of eruptions at Bezymianny volcano became impossible. Therefore, it is difficult to propose a general theory for volcanic eruption forecasting.

At the same time, Omori pointed out an advantage in predicting of volcanic eruption: we know the place of the event. Therefore, in short, the practical forecasting of volcanic eruption is based on a simple fact that with a sharp increase of seismic activity near a volcano, we may have an eruption. The most effective forecasting of volcanic eruptions has had no special methodology; more important has been the human factor of experience. Of course, the forecasting will benefit from the use of additional types of monitoring such as deformation, chemical parameters, etc. The author presents here chronicles of successful monitoring and forecasting of volcanic eruptions.

The July 6, 1975 eruption of New Tolbachik volcanoes in Kamchatka (see Chapter 5) was forecasted officially, with publication of the date of eruption in the regional newspaper (Tokarev, 1978). This forecasting was based on unusual seismic activity in an unusual place. The Tolbachik Valley, a region to the southwest from Plosky Tolbachik volcano, where the swarm of volcano-tectonic earthquakes was localized, has been calm at least since 1960. Therefore, the appearance of a seismic swarm could have been a signal of possible eruption. The swarm continued for 2 weeks giving time for the analysis of the situation and for correct decision. The forecasting of forthcoming eruption was made by P. Tokarev, according to the similarity in the seismic process near Tolbachik volcano with the seismic activity observed prior to flank eruptions of Klyuchevskoy volcano.

The first seismic signals appeared on the records of seismic stations (the nearest of them was situated at a distance of about 50 km from the epicentral zone) on 27 June at 00:53, as a sequence of tremor-type vibrations (see Figure 12.19). The first volcano-tectonic earthquake was recorded the same day at 01:09, and an intense swarm began at 09 h (Figure 18.17). During 28 and 29 June, hundreds of small earthquakes were recorded. The *P*- and *S*-wave arrivals of these events were sent from the seismic stations of Kamchatkan network to the Institute of Volcanology, Petropavlovsk-Kamchatsky, every few hours by radio, and the seismic events were located in near-real time. The epicentral area was estimated about 10 km southwest from Plosky Tolbachik volcano. The development of seismic activity was similar to the activity preceding the lateral eruptions of Klyuchevskoy volcano. On 30 June, the administration of the Institute of Volcanology was notified that a new lateral eruption is expected south-southwest from Plosky Tolbachik during the period from 30 June to 5 July (first alarm).

The number of seismic events decreased on 30 June; the forthcoming eruption was predicted to begin on 2–3 July. At the same time, two large ($M = 4.9$) earthquakes occurred on 2 July, and a new pulse of seismic activity was recorded during 2–4 July. On 3 July, Tokarev had sent a message to the head of the Far Eastern Division of the USSR Academy of Sciences about the imminent eruption 10 km to SSW from Plosky Tolbachik volcano (second alarm). A sharp decrease in the number of earthquakes on 5 July indicated that the eruption may begin during the nearest hours. It occurred on 5 July at 21:45 (GMT); local time 6 July, 09:45 a.m.

Figure 18.17 A sequence of alarms before the 1975 Tolbachik eruption. The variation in the number of earthquakes is shown according to the Kamchatka Regional Earthquake Catalog. The periods of alarms are shown based on Tokarev, 1978.

The information for the citizens of Kamchatka about the forthcoming eruption was sent by Tokarev to the regional newspaper "the Kamchatskaya Pravda" on 1 July; the first short article was published on 3 July. More detailed forecasting was published in the same newspaper on 5 July.

The importance of this forecasting for the scientific study of the eruption was high. The first expedition of the Institute of Volcanology had arrived at the site of the forthcoming eruption on July 1; their camp was only 400–500 m from the place where a fissure opened on July 6. The eruption was being observed by the volcanologists a few minutes after the opening the fissure.

No people live within the Tolbachik Valley; no evacuation was made.

Forecasting of the 15 June 1991 large explosions of Mount Pinatubo, the Philippines was carried out also according mainly to a continuous increase in seismic activity beneath the volcano, after moderate manifestations of volcanic activity (Harlow et al., 1996). During April to May, seismic activity was recorded mainly within a cluster located about 5 km northwest of the volcano, and it was only 50% probable that a large eruption may occur. During the first week of June, the seismic activity beneath the volcano strongly increased, and the lava extrusion occurred on 7 June. From 8 to 12 June, the majority of earthquakes occurred beneath the new dome; the tremor steadily increased from 9 June. These seismic data together with

the observations of an increase in SO_2 flux from the volcano allowed the forecasting of the possibility of an increase in volcanic activity. This information was communicated to the officials, and the communities close to volcano, as well as nonessential personnel from Clark Air Base, were evacuated on 10 June, 2 days before the first explosive eruption. The continuous increase in seismic activity beneath the volcano, including long-period and shallow volcano-tectonic earthquakes, and volcanic tremor showed that a climactic eruption was possible. During the morning of 15 June, at 07:30 the monitoring team and the remaining military personnel abandoned Clark Air Base. The climactic eruption began at 13:42.

This forecasting of the volcanic eruption was realized only because the USGS volcanological team had a great experience in the study of volcanic activity in other volcanic regions. They had no fixed methodology for this case, but they were prepared to try and understand the situation and take a decision.

Forecasting of the 20 November 1998 summit eruption of Volcán de Colima, México was based on the complex analysis of seismological and geochemical precursors (Reyes-Dávila and De la Cruz-Reyna, 2002; Zobin et al., 2002a). The monitoring of Volcán de Colima was coordinated by the Scientific Advisory Committee (SAC), which was organized by Colima University from the very beginning of this volcanic unrest on October 1998 and later authorized by the Colima State Government. The SAC consisted of scientists from Colima University, UNAM, CENAPRED, and Guadalajara University, as well as state civil protection authorities, military officers, and foreign specialists, and served to advise the Colima State Government with regard to volcanic hazard mitigation. The SAC had meetings practically every week beginning in the spring of 1998; three of the meetings were held with the Governor of Colima State. The SAC was directed by the Head of Scientific Investigations of Colima University and by the Technical Secretary of the Civil Protection of Colima State. At each meeting, the SAC discussed the development of volcanic activity and prepared a report for public dissemination. The Bulletins of the SAC were published regularly on the website of the CVO.

During 1998, two groups of SAC members saw indications in monitoring data for a future eruption of Volcán de Colima, and expressed those concerns first in a long-term forecast, and later in a short-term prediction. The long-term forecast was based on observed increases of S/Cl and δD in gases from two summit fumaroles beginning from the late 1997. Calling attention to those trends at the January 1998 session of the SAC, Y. Taran and J.C. Gavilanes forecast an effusive eruption "sometime in the nearest future, maybe this year." On 13 November 1998, during the fifth seismic swarm (see Chapter 6), the SAC received a Memorandum prepared by two other members, S. De la Cruz-Reyna and G. Reyes-Dávila, in which they announced a high probability of a new eruption at Volcán de Colima during the interval of 16−18 November. Their short-term prediction was based mainly on seismic data and the material-failure model of Cornelius and Voight (1995). The predicted range of dates was estimated from three different models (Figure 18.18). The authors of the prediction proposed three scenarios for the eruption, with the most probable scenario involving formation of a new dome in the summit crater followed by lava flows and pyroclastic flows (Reyes-Dávila and De la Cruz-Reyna, 2002).

Figure 18.18 The three material-failure models used to support the forecast of an eruption at Volcán de Colima for the period 16–18 November 1998. The thick irregular line represents cumulative 2-min averages of real-time seismic energy measurements (RSEM: expressed as the square of the seismic signal amplitude $\times 10^{-7}$) obtained during the first 20 days (480 h) of November 1998 from station EZV7 Vulcancito (see Figure 18.1), the nearest to the crater. The initial forecast, using the visco-elastic model, was issued late on 12 November 1998 (marked by "X"). Based on these 288 h of data, three sets of model parameters were used to define a forecast range for the failure time (times marked with a square, a circle, and a diamond). Around hour 371 (midday on 16 November), the strain accumulation rate decreased for about 1 day, and then resumed its increase. The regime of strain release changed again after hour 408 (late on 17 November). A new lava dome was first seen growing in the summit crater at 07:30 h on 20 November (hour 463).
Modified from Reyes-Dávila and De la Cruz-Reyna, 2002.

This Memorandum received the support of the SAC members, and on this basis, recommendations for the State Government were elaborated. In response, on the morning of 18 November, the Government of Colima State evacuated the ~180 inhabitants of Yerbabuena, a village located 8 km southwest from the crater of the volcano, and moved them to special pre-arranged shelters. That same day, Jalisco State evacuated the ~120 residents of Juan Barragan, located 10 km southeast of the crater. The inhabitants of Yerbabuena, Juan Barragan, and other proximal communities had been prepared for this action during many previous interactions with the Social Response Group of the CVO. These scientists had many discussions with the villagers during 1997–1998 about the hazards associated with future eruptions of Volcán de Colima. As a result, the evacuation of 18–19 November took place without complications. The lava eruption began during the night of 19–20 November.

Figure 18.19 Sequence of alarms before the 2000 Usu eruption.
From Okada, 2002.

The 31 March 2000 lateral eruption of Usu volcano, Hokkaido (see Chapter 7) was the first case in Japan when a formal alert was issued by the Japan Meteorological Agency (JMA) before the eruption (Okada, 2002; JMA, 2003). Figure 18.19 illustrates the process of the forecasting. The number of volcano-tectonic earthquakes recorded at the volcano increased clearly during the morning of 27 March. The Muroran Local Meteorological Observatory prepared the first report about this situation early on 28 March. Two hours later, a Volcanic Advisory about the warning attention was issued.

Emergency meetings of the Coordinated Committee for Prediction of Volcanic Eruption of JMA were held on 28 and 29 March to evaluate all data and information. GPS observations indicated that the volcano had inflated since March 29 (Takahashi et al., 2002). The Committee concluded that the eruption of Usu volcano should be considered "imminent" within a few days. Based on this conclusion, a "Volcanic Alert" was issued at 11:10 (Japan Standard time) 29 March by JMA. This alert signified that volcanic activity reached extremely high levels and urgent measures to prevent and mitigate danger to human lives were required. Based on Volcanic Alert of 29 March, local governments established a headquarters for volcanic disaster prevention and issued evacuation orders. Fifteen thousand residents and tourists were evacuated from the area of volcanic danger by local governments without any confusion. After 30 March, cracks and faults were also found in the northwestern foot and summit, indicating that crustal deformation had increased (JMA, 2003). The last Alert was issued on 31 March, a few hours before the eruption.

Forecasting of the 1 October 2004 volcanic eruption at Mount St. Helens, Cascades, and the prediction of associated volcanic hazards. This description of the volcano monitoring before an eruption is given according to the review article by Scott et al. (2008). It is of a particular interest because it presents not only the analysis of variations in seismic activity, but also it shows the cooperation between the scientific, governmental, and media teams, and shows the system of notification of ash hazards to aircraft.

A 2-day swarm of tiny (mostly $Md < 1$), shallow volcano-tectonic earthquakes at the volcano began early on 23 September 2004 (see Chapter 7; Figure 18.20). The

Figure 18.20 Sequence of alarms before the 2004 Mount St. Helens eruption. Real-time seismic amplitude measurement (RSAM) during the first 6 weeks of unrest and eruption at Mount St. Helens served as proxy for rate of seismicity (modified from Moran et al., 2008a). Times of explosions and periods of tremor are shown by symbols.
From Scott et al., 2008.

Information Statement by USGS–CVO and Pacific Northwest Seismic Network (PNSN) was issued at 18:00 PDT on 23 September. The statement discussed the swarm's similarity to previous ones, surmised that this swarm might reflect heavier-than-normal precipitation in the preceding 4 weeks, but considered eruption unlikely without significant further precursors. Beginning on the afternoon of September 25, earthquakes increased in magnitude ($Md \leq 2.8$), and by the following morning a total of 10 $Md > 2.0$ earthquakes had been recorded—the most in a 24-h period since lava-dome building ended in 1986. This increase in seismicity, along with the appearance of low-frequency and hybrid earthquakes, prompted reassessment of the probability of hazardous activity. At 15:00 PDT on 26 September, after notifying the Washington State Emergency Management Division and Giffort Pinchot National Forest, home of the Mount St. Helens National Volcanic Monument (GPNF), USGS–CVO and PNSN released a Notice of Volcanic Unrest (Figure 18.20, Alert level 1) indicating that seismicity had surpassed background level and that the volcano was in a state that could evolve toward eruption.

The increase in alert level spurred several actions. The GPNF activated their Emergency Coordination Center and closed the south-flank climbers' trail and other trails near the crater. Field crews took advantage of uncommonly clear autumn weather to begin installation of additional seismometers and global positioning system (GPS) receivers around the volcano. Aerial observations of the crater at this time showed no obvious changes. By 29 September seismicity had intensified to about three events per minute with maximum magnitudes of $Md = 2.4–2.8$, rising RSAM values, and hours-long periods during which repetitive earthquakes of similar waveform dominated the records. The well-known

association of such seismicity with lava-dome-building eruptions of the past few decades at Mount St. Helens and elsewhere raised eruption concerns further and prompted issuance of a Volcano Advisory midmorning of 29 September. The aviation color code was raised to Orange (Alert level 2, Figure 18.20) because of increasing concern that explosions could send ash to altitudes where air traffic would be affected. The Volcano Advisory signified that processes were underway that could lead to hazardous eruptive events, but not imminently.

The alert-level rise to Advisory triggered the GPNF to activate a local Incident Management Team and to begin discussions with other agencies regarding a Joint Operation Center, which would be necessary if unrest continued to escalate or eruptions began. Intense media interest, directed chiefly toward USGS−CVO and the Monument, was beginning to adversely affect operations, so local and state emergency management agencies joined with USGS−CVO and the GPNF to begin organizing a Joint Information Center.

Real-time seismic amplitude values increased again on the evening of September 29, reflecting an increase in maximum earthquake magnitudes (Md = 2.8−3.3) and rate of earthquakes of Md > 2 to about one per minute. The RSAM values fluctuated over the next 1.5 days. Flights during this time period again failed to detect significant amounts of volcanic gas. Data from a GPS receiver on the west side of the 1980s lava dome showed northward movement of several centimeters per day, consistent with a shove from a rising mass south of the dome.

The inaugural flight at Mount St. Helens of a helicopter-mounted, forward-looking infrared radiometer midday on 1 October was rewarded with close observation of the first explosion of the current eruption, which bored an ice-walled crater through the western part of the welt. An ash and vapor cloud rose about 2 km above the crater rim and drifted southwestward. Neither this nor other explosions over the next few days had any recognizable precursors. The rate of earthquakes rapidly increased again during the evening, and by late evening, RSAM values exceeded those prior to the explosion.

For the next 2 days (2−3 October), seismicity remained at a high level of several thousand events per day, punctuated by earthquakes of Md = 3.5−3.9. Midday on 2 October, a 50-min episode of energetic, relatively broadband seismic tremor, coupled with hints of volcanic gas during the previous few days and a continued high rate of growth of the welt, prompted USGS−CVO and PNSN to issue a Volcano Alert at 14:00 PDT. Because of concern over an imminent and potentially hazardous event, the GPNF evacuated the Johnston Ridge Observatory, 8 km north of the crater; the Washington State Department of Transportation closed State Route 504 at the Coldwater Ridge Visitor Center; and land and air space was closed within 8 km of the crater. The increased possibility of ash clouds reaching high altitudes where they could affect air traffic warranted raising the aviation color code to Red (Alert level 3, Figure 18.20).

Volcano monitoring did not give the exact date of the first explosion, but the Alert level 2, pronounced on 29 September, signified that the small explosive eruption is imminent. Thus its occurrence on 1 October was expected.

19 Seismic Activity at Dormant Volcanic Structures: A Problem of Failed Eruption

Numerous seismic events are located below a dormant volcanic structure where no seismic activity was observed during many years. What does it mean: a forthcoming eruption or only a seismo-tectonic episode? A volcano can be dormant for hundreds or thousands of years and then become active again; a new volcano may be born. We have a seismic sequence which frequently leads up to the eruption. But once in a while we have a seismic sequence in a volcanic zone without any eruptive activity. The problem is how to recognize if the seismic swarm announces the forthcoming eruption or is it only a seismic crisis without volcanic manifestations, a failed eruption, an instance in which magma has intruded to shallow depths (generally < 2 km to 3 km below the surface), accompanied by anomalous seismicity, deformation, degassing, but ultimately fails to reach the surface, as was defined by Moran et al. (2011). In this chapter, the author gives some of the best-documented cases of the seismic activity that were recorded at dormant volcanic structures and were not followed by an eruption; then some results of modeling of magma propagating to the surface and its arrest near the surface; and finally, a short discussion of the problem of monitoring of the seismic activity at dormant volcanoes.

19.1 Failed Eruptions: Case Studies

The failed eruptions were recorded at different volcanic centers. Here the author discusses the occurrence of seismic swarms recorded without eruptions at large calderas, at strato-volcanoes, and in rift settings.

19.1.1 Failed Eruptions at Large Calderas

Large resurgent calderas are characterized by slow post-eruption upheaval of its floor. They form a broad topographical depression with a central elevated massif. Vertical uplift of more than 1 km often takes place (Francis, 1993). In this section, two cases of the seismic activity associated with the failed eruptions at large ancient calderas are described.

Introduction to Volcanic Seismology. DOI: 10.1016/B978-0-444-56375-0.00019-0
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Long Valley Swarms

The large 17 × 32 km Long Valley caldera (California, USA) was formed as a result of the voluminous Bishops Tuff eruption about 730,000 years ago. Resurgent doming in the central part of the caldera occurred shortly afterward, followed by rhyolitic eruptions from the caldera moat and the eruption of rhyodacite from outer ring fracture vents, ending about 50,000 years ago. The caldera remains thermally active, with many hot springs and fumaroles (GVP, 2002).

Seismic swarms began in October 1978 and were accompanied by a broad uplift of the caldera floor and local increase in fumarolic activity. Seismic activity culminated in a sequence of four magnitude *M* 6 earthquakes that occurred beneath and to the south of the caldera from 25 to 27 May 1980. Since a swarm in July 1984, most seismicity has been outside the caldera (Newhall and Dzurisin, 1988).

The majority of seismic events are typical tectonic earthquakes, but also episodes of spasmodic tremor and long-period events were observed (Cramer and McNutt, 1997). Figure 19.1 shows the temporal characteristics of the 1989 swarm.

This swarm occurred beneath Mammoth Mountain that is situated at the southwest rim of the caldera. It continued for about 6 months; more than 1500 events of magnitude 3 and less were recorded within the depth range of 2−11 km. Only 30 of them were identified as low-frequency events. Fourteen strong bursts of spasmatic tremor were also recorded. No regularities in the occurrence of low-frequency events and tremors were observed, and no volcanic eruption occurred during this period of caldera unrest.

Figure 19.1 Seismic events recorded during the 1989 swarm beneath Mammoth Mountain near Long Valley caldera, USA. The occurrences of magnitude 3 events are shown. From Cramer and McNutt, 1997.

Campi Flegrei Swarms

Campi Flegrei is a volcanic complex located near Naples, Italy. Eruption of about 80 km^2 of the Campanian Tuff about 35,000 years ago initiated formation of the Phlegrean Fields, a complex of cones and craters within a 13 km diameter caldera (Francis, 1993). The last eruption was in 1538; it was preceded by uplift that lasted over 30 years. After this eruption, the caldera has been predominantly sinking. A new period of uplift began near Pozzuoli during 1969. It continued for 4 years and reached 1.5 m. After a short gap, the uplift continued in 1982 to 1984 with the maximum value of 1.8 m (Figure 19.2). The uplift was accompanied by numerous small earthquakes (Barberi and Carapezza, 1996). Figures 19.3 and 19.4 show the temporal and spatial characteristics of the 1982 to 1984 seismicity. During the period January 1983 to September 1984 more than 10,000 events were recorded, with magnitude ranging between 0.1 and 4.2. The earthquakes clustered within the depth range between 0 and 4 km; their epicenters were distributed within the uplift zone. Since January 1985, the area

Figure 19.2 Uplift of Campi Flegrei, Italy during the 1982–1984 unrest. The curves are isolines of vertical uplift in cm. The cross near 180 cm indicates the point of maximum uplift. From Barberi and Carapezza, 1996.

has undergone a phase of general subsidence interrupted by short-duration uplifts. The uplift phases were accompanied by seismic activity while the subsidence phases occur aseismically (Del Pezzo et al., 1987). No volcanic eruption has occurred.

Figure 19.3 The epicenters (A) and hypocenters (B) of earthquakes located during the 1982–1984 unrest at Campi Flegrei, Italy.
From Barberi and Carapezza, 1996.

Figure 19.4 Temporal variation in daily number of earthquakes recorded during the period May 1983 to May 1984 at Campi Flegrei, Italy.
From Del Pezzo et al., 1987.

19.1.2 Failed Eruptions at Strato-Volcanoes

In this section, three cases of the seismic activity associated with failed eruptions at ancient strato-volcanoes are presented. The seismic activity was of different duration, from days and weeks to years.

Matsushiro Swarms

The Matsushiro earthquake swarms (Central Honshu, Japan) began on 3 August 1965 with the earthquakes clustering within a 5-km radius zone around Mount Minakami that was not among the known active volcanoes of Japan. The swarm activity strongly increased in 1966 (Figure 19.5), reaching about 7000 small earthquakes per day. During February–June 1966, strong deformations accompanied by ground fissures and spring water were observed. During July to December 1966, the spring water and the expansion of ground fissures caused numerous landslides.

Area of the epicentral zone spread significantly during the first 2 years of activity (Figure 19.6). After June 1967, activity quickly dissipated and completely stopped by the end of 1970. More than 700,000 events were recorded during this swarm; more than 60,000 of them were felt. The largest of these earthquakes had a magnitude $M = 5.4$; the total energy of the swarm events was equivalent to a magnitude $M = 6.4$ earthquake. No eruption was observed (Seismic Activity in . . ., 1998).

Figure 19.5 Daily frequency of earthquakes during the Matsushiro earthquake swarm. Data from Nagano Local Meteorological Observatory.

Figure 19.6 Spreading of the source region during the Matsushiro swarm (1) up to February, 1966; (2) up to July 1966; (3) up to December 1966; and (4) up to May 1997. From Seismic Activity in . . . (1998).

Aag Swarm

The swarm of 113 earthquakes occurring within the edifice of dormant, 2319-m high, andesite-basaltic Aag strato-volcano, Southern Kamchatka, began on 10 August 1966 and continued for 65 h (Gorelchik, 1969). Aag belongs to the Avachinsky-Koryaksky group of Middle Pleistocene-Holocene volcanoes (Masurenkov et al., 1991).

A total of 46 events were located during the swarm (Figure 19.7A). Epicenters clustered within the compact area of about 80 km^2. They occurred at a depth between 5 and 10 km. Maximum event with magnitude $mb = 3.5$ was recorded during the peak in the number of seismic events, 5 h after the beginning of the swarm (Figure 19.7B). Seismic records displayed the typical waveforms of volcano-tectonic earthquakes with clear arrivals of *P*- and *S*-waves (Figure 19.8). No eruption occurred. No new swarm was observed during the last 40 years.

Asacha Swarm

Intense swarm of volcano-tectonic earthquakes occurring within the edifice of the andesitic, 1910-m high, complex volcano Asacha, Southern Kamchatka, was

Figure 19.7 Earthquake epicenters (A) and variations in the number of events (B) during the 1996 Aag swarm, Kamchatka. Stars show the volcanic centers; crosses mark the epicenters. Contours of the volcanic edifice of Aag volcano are shown by the solid line. Data from the Kamchatka Regional Earthquake catalog and from Gorelchik, 1969.

recorded between 7 and 30 March, 1983 (Zobin et al., 1986). Asacha represents the complex massif of four Pleistocene-to-Holocene strato- and shield volcanoes located within a distinctly fault-bounded crustal block. Ten lava domes dot the flanks of the Asacha volcanoes. Most of them were formed during the Pleistocene, but some may be early Holocene in age. Also during the Holocene, basaltic cinder cones and lava flows related to regional volcanism were erupted along the western and southern flanks of the Asacha complex (GVP, 2011).

A total of 390 events were located during the swarm (Figure 19.9A). They filled the area extending to east from the main volcano summit zone and covering the central part of the volcanic complex edifice. The majority of events (77%) occurred within the depth interval between 5 and 20 km. Three peaks in the swarm activity

Figure 19.8 Seismic record (displacement) of one of the largest earthquakes of the Aag swarm recorded by short-period station at a distance of 17 km.
From Gorelchik, 1969.

were observed (Figure 19.9B). During the peak activity of 13 March, five earthquakes with magnitudes *mb* (USGS) between 4.5 and 5.0 were recorded. These large events were located along the western border of the epicentral area, including the main summit zone of the volcano. No eruption occurred. No new swarm was observed during the last 30 years.

19.1.3 Failed Eruptions in Rift Settings

In this section, the seismic activity associated with two failed eruptions in rift setting is described. One of the cases was observed in the northern volcanic rift zone in Iceland, and other, in the Arabian rift zone.

Icelandic Rift Zone Swarm

Since February 2007, over 10,000 micro-earthquakes, ranging in size from magnitude $M_L = 0.9$ to $2.\tilde{2}$, have been detected near Mount Upptyppingar (Figure 19.10). Upptyppingar is located within the northwestern edge of the fissure swarm that

Figure 19.9 Earthquake epicenters (A) and variations in the number of events (B) during the 1983 Asacha swarm, Kamchatka. The epicenters of largest earthquakes with magnitude *mb* = 4.5−5.0 are shown with filled diamonds. A star show the volcanic center; crosses mark the epicenters. Contours of the volcanic edifice of Asacha volcano are shown by the solid line.
Data from the Kamchatka Regional Earthquake catalog and USGS catalog.

occupies part of the Kverkfjöll volcanic system, within the Northern Volcanic Zone of Iceland. Rising to ~400 m above the surrounding landscape, the mountain comprises hyaloclastic deposits formed in a sub-glacial eruption toward the end of the last glaciation, over 10,000 years BP.

The seismic records were characterized by impulsive P- and S-wave onsets. Over an 11-month period, the seismically active region migrated from Upptyppingar, spreading ~12 km east-northeast to the Álftadalsdyngja lava shield. Starting from a depth of 18 km b.s.l., the seismicity migrated in both up- and down-dip directions and also laterally before stopping at 13 km depth. In addition to sustained earthquake activity, significant horizontal displacement was registered at two permanent GPS stations in the Upptyppingar region (Jakobsdóttir et al., 2008; Geirsson et al., 2009; White et al., 2011).

The seismicity was supposed (White et al., 2011) to be produced by progressive melt intrusion of a dyke moving upward from a sill at 18 km depth in the mid-crust of the volcanic rift zone (Figure 19.11). A total of 545 events, recorded from the

Figure 19.10 Spatial and temporal view of seismicity at Upptyppingar and Álftadalsdyngja, Iceland. (A) earthquake epicenters; (B) temporal variation of hypocenters. Different colors correspond to temporal development of the swarm. The nearest seismic station MKO is shown with triangle. From Geirsson et al., 2009, with modifications.

period 6–24 July 2007, across the array, consisting of 27 seismometers, were selected for detailed study of earthquake foci clustering. The most remarkable feature of the seismicity was that hypocenters shallower than 17.5 km, which were attributed by White et al. (2011) to intrusion along a dyke, lie on a single plane dipping at 50° and striking at 74° with a root-mean-square misfit of only 115 m. No surface intrusion was observed.

Arabian Rift Zone Swarm

During April–June 2009, a swarm of more than 30,000 earthquakes struck the Harrat Lunayyir, situated in the northwestern end of the Saudi Arabian lava fields, east of the Red Sea. Harrat Lunayyir is one of the smallest Holocene lava fields of Saudi Arabia, but individual flow lobes radiate long distances from the center of the Harrat, and flows reached the Red Sea in two places. Harrat Lunayyir contains about 50 volcanic cones that were constructed over Precambrian crystalline rocks along a north-south axis. Lava flows are basaltic to basanitic in composition, and the Holocene flows are alkali olivine basalts. One of the cones may have erupted around

Figure 19.11 Hypocenter locations and the possible position of the dyke viewed along strike of the dyke during the 6–24 July 2007 period of supposed melt injection at Upptyppingar. Dots show earthquakes with fault plane solutions: red are reverse faults and blue are normal faults. Inset shows fault planes from 229 double-couple reverse solutions viewed along strike and projected onto an equal area projection orientated vertically and orthogonal to the dyke. Broken gray line shows speculative location of a feeder sill.
From White et al., 2011.

the tenth century AD or earlier (GVP, 2011). The system of Saudi Arabian lava fields, situated along the western coast of Arabian Peninsula, is connected with the Main Ethiopian Rift in northeastern Africa.

The seismic swarm began on 18 April 2009 (Al-Amri and Al-Mogren, 2011). The joint KSU and KACST seismic network, consisting of 21 seismic stations, was installed around the volcano, and 398 earthquakes with $M > 3$ were located from 29 April to 27 June 2009, mainly from 1 to 21 May. Two epicentral areas were formed, to north and south from Harrat Lunayyir, leaving a free space beneath the volcano (Figure 19.12). During the period from 29 April to 12 May, seismicity developed to south from the volcano. Three earthquakes at the depth between 13 and 15 km were recorded in the very beginning of this sequence on 30 April and 1 May, and then the events clustered within the depth range between 2 and 7 km (Figure 19.13). Earthquake frequency reached its maximum on May 12 and then sharply ceased. Maximum magnitude of earthquake was 3.7.

Beginning from 12 May, the seismic activity was recorded to north from the volcano with a significant increase in the number and magnitude of earthquakes. The seismic activity reached its peak on 19 May. Total of 12 earthquakes with $mb \geq 4.0$ (USGS, 2011) were recorded this day, including the $Mw = 5.7$ event.

Figure 19.12 Epicenters of earthquakes with $M > 3.0$ located by the joint KSU and KACST seismic network at Harrat Lunayyir, Saudi Arabia, during April–June 2009. A star shows the volcano; filled circles show the earthquake epicenters. The epicenters of three large earthquakes with Harvard CMT solutions are shown with crosses; the largest cross indicates the position of the largest earthquake with magnitude $Mw = 5.7$. The seismic stations are shown with triangles. The position of surface ruptures (according to Pallister et al., 2010) is shown with a broad line. Insert shows the satellite photo of Harrat Lunayyir (NASA Space Shuttle image STS26-41-61, 1988, http://eol.jsc.nasa.gov/). The catalog of earthquakes was taken from Al-Amri and Al-Mogren, 2011.

These seismic events clustered at a depth between 6 km and 11 km, deeper than the initial southern group of earthquakes.

A northwest trending 8-km-long surface fault rupture with 91 cm of offset was formed (Pallister et al., 2010) near the epicenter of the $Mw = 5.7$ earthquake (Figure 19.12). The focal mechanism solutions published by Harvard University (Harvard CMT, 2011) for three events, occurring on 19 May (Figure 19.12), indicated the normal fault rupturing with the fault planes coinciding with the direction of the surface rupture and with the general direction of the surface structures of Harrat Lunayyir volcano (see inset in Figure 19.12). INSAR analysis of interferograms acquired for the period 17 to 28 June (Pallister et al., 2010) showed that the deformation field at the volcano region was best modeled by intrusion of a 10-km-long, northwest-trending dyke, with a top at less than 2 km depth and volume of about 0.13 km^3. No surface intrusion was observed during 2 years after the seismic episode.

Figure 19.13 Temporal development (A) and the meridional cross section of earthquake swarm at Harrat Lunayyir (B). In A, the solid line shows the variations in the number of the southern group of events and the broken line shows the variations in the number of the northern group of events. A star shows the volcano.
The catalog of earthquakes was taken from Al-Amri and Al-Mogren, 2011.

19.2 Modeling of Magma Ascent Resisting

As we can see, there is no difference in spatial-temporal distribution between the cases described above and the preceding swarms of volcano-tectonic earthquakes described in Chapters 5–7. No difference with volcano-tectonic earthquake records was observed in the shape of the seismic signals associated with failed eruptions. Crustal deformation recorded for majority of these cases allows to consider that this seismic activity beneath the volcanic constructions may be associated with any subsurface movement of magmatic matter that was arrested in its way to the surface. So the question is what determines whether or not a buoyant magma-filled crack actually reaches the Earth's surface or halts as a shallow intrusion.

Several mechanisms have been proposed for dyke propagation arrest, including magma solidification, reduction of magma-chamber overpressure, barriers related to sharp gradients in mechanical properties, and density stratification. Chen et al. (2011) noted that the dyke arrest may be dominated by a single mechanism in some geological settings, or may be caused by the simultaneous action of multiple mechanisms. Magma solidification may dominate when a dyke propagates slowly for a long period of time. Magma chamber overpressure reduction may be a major mechanism for dyke arrest when the dyke volume is significant compared with the volume of the magma chamber. Mechanical properties barriers and density stratification may act simultaneously when a dyke propagates near the level of neutral buoyancy as the dyke experiences marked changes in buoyancy force and mechanical resistance to propagation.

Below some results of laboratory and numerical experiments are presented. They give the ideas about the process of dyke arrest below a volcano and, as a result, a case of failed eruption.

19.2.1 Experimental Study of the Ascent of a Fixed Magma Volume

Taisne et al. (2011) studied the downward propagation of a fixed volume of dense liquid through a lighter elastic-brittle medium made of gelatin. A fixed volume of dense liquid was injected in a small pre-existing cut at the top of a large tank. The initial cut was deep enough to be unstable, such that the buoyancy force was sufficient to induce fracturing at the tip. The experiments were focused on cracks issuing from a source of small breadth B_o in the horizontal direction, such that the crack extended both horizontally and vertically.

Taisne et al. (2011) analyzed different phases of crack propagation. In a first phase, the crack grew larger in both breadth and length, driven by a lateral pressure gradient due to the liquid overpressure. It maintained an approximately circular shape. The propagation was characterized by continuous thinning of the nose region because of loss of liquid to the growing tail region. The dyke eventually reached a limit width and extended only in the vertical. In a second phase, the crack maintained a constant breadth and extended only in the vertical direction with a parabola-shaped nose region. The final phase saw a rapid decrease of velocity and the arrest of propagation. In all the experiments, propagation was driven by buoyancy and resisted by forces associated with fracturing at the tip, elastic deformation, and viscous flow through the thin crack aperture. The cracks were characterized by the crack length in the vertical direction, noted L, crack breadth in the horizontal direction, noted B, and thickness of fluid within the crack, noted δ. The relevant physical properties were the fracture toughness K_c and shear modulus G of the host solid and the buoyancy Δ and viscosity μ of the liquid.

Experimentally derived scaling laws (Eqs (19.1) and (19.2)) allowed us to estimate the dimensions and volume of the largest dyke, associated with a magma source at a depth D, that can stall below Earth's surface. Equation (19.1) shows

the dependence of the final crack length L_f on the fracture toughness K_c and the buoyancy $\Delta\rho$:

$$K_c = \gamma \Delta\rho g L_f^{3/2} \tag{19.1}$$

where γ, the shape coefficient, depends on the final aspect ratio, L_f/B_f, and g is gravitational acceleration. Figure 19.14A shows how γ varies as a function of the final crack aspect ratio.

The crack achieves its final breadth B_f during the first propagation phase. The balance between elastic and buoyancy forces leads to the following reasonable agreement with the experimental data (Figure 19.14B):

$$B_f = \left(\frac{GV}{(1-\nu)\Delta\rho g}\right)^{1/4} \tag{19.2}$$

According to experimental results, the successful propagation of a dyke to the surface or its stalling is conditioned by the initial crack breadth, the final length of the crack, and the magma volume available. One condition for the stalling of a dyke below Earth's surface is that the final length of crack $L_f \leq D$. A stronger constraint on eruption comes from the magma volume available. It was found that the successful propagation of a dyke to the surface requires that the magma volume exceeds a threshold value. If the initial crack breadth is B_o, the threshold volume $V_{\min}$ is therefore

$$V_{\min} = C_o B_o L_o \delta_o \tag{19.3}$$

where C_o is a proportionality coefficient of order 1. The minimum magma volume that is required for an eruption is shown as a function of source depth D in Figure 19.14C, which gives the final maximum vertical extension that a dyke can attain given a finite amount of magma. If the value found is smaller than the depth of the source, the magma cannot reach the surface, and there is no eruption.

19.2.2 Arrest of Propagating Dyke Due to Mechanical Barriers and Density Stratification in an Upper Crustal Horizon

The crust above the magma chamber is composed of layers with large and abrupt contrasts in stiffness. Because injection and arrest of dykes and sheets occurs primarily in the layers above the source chamber of the volcano, the focus is on stiffness variation between those layers. In the uppermost parts of many active volcanic zones and central volcanoes, layers of soft sediments and pyroclastic rocks offer strong mechanical contrasts to the stiffer lava flows. A soft, low-density layer may suppress the tensile stress associated with the tip of a dyke or sheet intrusion so much as to make it impossible for the intrusion to continue its extension.

Figure 19.14 Relationship between the parameters used for modeling of magma ascent to the surface. (A) Geometrical shape factor γ for the stress intensity factor of experimental cracks in gelatin as a function of the crack aspect ratio, *Lf*/*Bf*. In the limit of *Lf*/*Bf*→0, one

Gudmundsson (2002) used boundary-element models to study the effects of a large variation in stiffness between layers on the stress field inside a volcano. He showed that the soft layer would act as a stress barrier where many sheets and dykes could become arrested. Gudmundsson (2002) noted also that many layered rocks, such as in active central volcanoes and volcanic zones, contain sharp contacts between layers, and these contacts are often mechanically weak. When a horizontal discontinuity, such as a contact, a joint, or a fault, is approached by the tip of a vertically propagating dyke, the discontinuity may open up and contribute to arresting the dyke.

A series of numerical calculations, performed by Taisne et al. (2011), showed that a low density crustal layer can be an effective barrier in the path of the vertical dyke propagation. When magma accumulates in a swollen nose region at the interface between the low-density layer and the dense basement, the magma overpressure is largest at this interface and increases with increasing penetration into the upper layer. It may become large enough to induce horizontal fractures in the dyke walls and lateral intrusion of a sill, which prevents eruption. This requires that the thickness of the low-density layer exceeds a threshold value that depends on the density contrast between magma and host rock. Figure 19.15 shows the obtained dependence of the minimum magma volume on the thickness of low density rocks and density contrast between magma and host. If the magma volume is smaller than a threshold value, neither sill intrusion nor eruption is possible, and magma gets stored in a horizontal blade-shaped dyke straddling the interface.

Chen et al. (2011) numerically modeled the arrest of dyke propagating into the region above the level of neutral buoyancy (the depth where the host rock density becomes equal to the magma density) applying the finite difference/perturbation method. Numerical results showed that dyke propagation velocity initially increased with dyke length, reached a peak value when the dyke tip reached a position well above the level of neutral buoyancy, and then rapidly decreased to zero after the dyke propagates further into the region above the level of neutral buoyancy (Figure 19.16). Chen et al. (2011) consider that these results indicate the density gradation as an effective mechanism for dyke arrest.

Figure 19.14 (Continued) retrieves the 2-D crack theory, such that $\gamma o = 1.12\sqrt{\pi/2}$, which is indicated by the thick bar at the left of the plot. The dashed curve was drawn through the data for illustration purposes. (B) Measurements of the final fracture breadth as a function of values predicted by a simple balance between buoyancy and elastic stresses in a penny-shaped crack, which is appropriate for the first propagation phase (Eq (19.2)). (C) Relationship between the volume of magma and the vertical length of a buoyancy-driven crack that stalls beneath Earth's surface (values for the physical properties are taken as $\Delta\rho 0 = 10$ and 100 kg m^{-3}, $G = 109$ and *1010 Pa*, and $\nu = 0.25$). This relationship can be understood as the minimum magma volume required for an eruption as a function of the depth of the magma source. The two curves were drawn for the two extreme values of ratio $\Delta\rho g/G$ that are indicated. After Taisne et al., 2011.

Figure 19.15 Results of the numerical investigation of the minimum volume needed for a sill or for an eruption. Top panel: minimum volume needed for a sill as a function of the thickness of low density rocks and density contrast between magma and host. Due to the negative buoyancy of magma, a dyke can only penetrate over a finite distance before reaching the rupture threshold for intrusion of a horizontal magma sheet. The blank region on the left is such that the interface is not deep enough to allow sufficient pressure build up within the dyke. The two black horizontal lines correspond to volumes of 10^4 and 10^5 m^3/m. Bottom panel: minimum magma volume for an eruption. The blank region on the right is such that the dense basement is too deep for the dyke to extend to the surface. Calculation for a value $\Delta\rho o = 100$ kgm^{-3} of magma buoyancy below the interface. Black curves correspond to 10^{-4}, 10^{-2}, 10^0, 10^2, and 10^4 m^3/m.
From Taisne et al., 2011.

Figure 19.16 Dike propagation velocity versus propagation length. (A) dike is 1 km below the level of neutral buoyancy (LNB); (B) dike is 1 km below the LNB.
Collated after Chen et al., 2011.

19.3 Monitoring of the Seismic Activity at Dormant Volcanoes

As was shown in Chapters 5–8, the swarms of volcano-tectonic earthquakes occurring at active andesitic and dacitic volcanoes and at basaltic volcanoes are characterized by different features of their development. Therefore, the author discusses here separately the monitoring of seismic activity at the dormant volcanoes of these two groups of volcanic structures.

19.3.1 Monitoring of Andesitic and Dacitic Dormant Volcanoes

The problem of re-awakening of andesitic and dacitic dormant volcanoes was discussed by Zobin and Jimenez (2008) (see also Chapter 8.3). It was shown that is possible to discriminate three stages in the re-awakening process: the preliminary stage that was characterized by a sharp increase in seismic activity near the volcano and continues till the first phreatic explosion; the stage between the first phreatic explosion and the large explosion; and the post-explosion stage of dome-building with a gradual decrease in the number of seismic events (Figure 19.17B and C).

Seismic activity, associated with failed eruptions at andesitic and dacitic volcanoes discussed in Section 19.1.2, was also characterized by a sharp increase in seismic activity near the volcano. The first impulses of seismic activity recorded before an eruption and without eruption are similar. The only difference is an appearance of a phreatic explosion as a signal of forthcoming eruption. If a phreatic explosion was recorded at the moment of peak frequency of seismic events, the scheme of Figure 19.17B may be taken as a more probable continuation of the seismo-volcanic activity. If phreatic explosion was recorded during continuous seismic activity,

Figure 19.17 Variations in the number of seismic events of swarms observed at Bezymianny volcano, Kamchatka (A) and Unzen volcano, Japan (B). *St1*, *St2*, and *St3* indicate the stages of seismic activity. The data were taken from Tokarev (1981) (Bezymianny) and the catalog prepared by Shimabara Earthquake and Volcano Observatory (Unzen).

the scheme of Figure 19.17C may be taken as a more probable continuation of the seismo-volcanic activity.

For the majority of well-documented events, the duration of the first seismic impulse was no longer than 1 month. Therefore, the final solution may be taken in such manner: if the first month of seismic activity was finished with a complete cease of seismic events, this seismic episode with a high level of probability was associated with failed eruption. If the seismic activity continues for a longer time, once can only wait and continue monitoring until the end of seismic episode.

This may be illustrated for the case of the 1983 Asacha swarm (Figure 19.9). This swarm reached maximum frequency of events in 7 days after the beginning of activity. The high level of activity continued for 6 days, and then gradually decreased. The swarm ceased after 24 days of activity. According to characteristic scenarios of re-awakening of andesitic and dacitic dormant volcanoes shown in Figure 19.17, there are two ways of re-awakening: a phreatic explosion may occur between the seventh and thirteenth days of maximum activity, and then activity can develop according to the scenario shown in Figure 19.17A. No phreatic explosion was recorded, and this scenario is rejected. Now, beginning from the twentieth day of activity, the scenario of Figure 19.17B may be applied for monitoring. But the swarm was over after 24 of activity. Therefore, in about 1 month after cessation of the swarm, the decision about failed eruption may be made.

19.3.2 Monitoring of Basaltic Dormant Volcanoes

Activation of basaltic dormant volcanic structures may occur in three ways: re-awakening of dormant volcano, the birth of new volcano, and associated with dyke injections along rift zones. The author knows only one case of re-awakening of basaltic volcano that occurred at Academia Nauk volcano, Kamchatka, on 2 January 1996. It was the first historic eruption of Academia Nauk volcano after 4800 years of repose (Belousov and Belousova, 2001). At the same time, the birth of new volcanoes (Kamchatka, México, submarine events) and surface dyke injections along rift zones (Iceland, Afar) are frequent enough. Below I give some characteristic features of the seismic activity associated with these eruptive events at dormant structures.

Re-Awakening of Basaltic Volcano

The re-awakening of Akademia Nauk volcano in Kamchatka after 4800 years of repose that occurred on 2 January 1996 (Belousov and Belousova, 2001; see also Chapter 2.3) may serve as an example of this type of activity. During 24 years before the eruption, the seismic activity developed at depths of 0 to 22 km, with the majority at depths between 5 and 15 km. The main preliminary swarms were recorded in 1978, 1992, and 1995. The maximum magnitude of 1972–1995 events was $Ms = 5.5$ (Zobin et al., 1983).

A sharp increase in seismic activity was observed on 1 January 1996 (Figure 19.18A). The process began with the 2-h sequence of 199 very small

(magnitude about 0) events. The second sequence included 58 events of magnitude greater than 2. Majority of them occurred at the depth between 2 and 20 km; but 9 events were recorded between 40 and 60 km of depth. This sequence continued for 4 hours and culminated in the $Mw = 7.1$ earthquake.

The third sequence of seismic events occupied the area of the second sequence epicenters and spread also to the southwest along the regional fault. This sequence consisted of 261 events with magnitudes greater than 3 and continued for about 14 h. The eruption of new volcanic center in Karymskoye Lake situated within the Academia Nauk caldera occurred about 14 h after the beginning of aftershock

Figure 19.18 Variations in the number of seismic events of swarms observed at Academia Nauk volcano, Kamchatka (A), New Tolbachik volcanoes, Kamchatka (B), and Dabbahu volcano, Ethiopia (C). The data were taken from the Regional Catalog of Kamchatka Earthquake (Academia Nauk and New Tolbachik volcanoes) and from the USGS catalog (Dabbahu).

sequence. The seismic activity continued then for about a month, then gradually decreased (Zobin and Levina, 1998).

The main difference in the seismic activity just before the beginning of the 1996 eruption and the seismic activity of 1972–1995 was in the deeper distribution of the foci of the January 1996 earthquake sequence and in the appearance of large $Mw = 7.1$ earthquake.

Birth of New Volcanoes

The birth of new volcanoes at basaltic volcanic structures was recorded within the Michoacán-Guanajuato Volcanic Field, México (Jorullo, 1759; Parícutin, 1943); within the Tolbachik Dol field of cinder cones, Kamchatka (New Tolbachik volcanoes, 1975); during submarine eruptions at Izu volcanic arc (Teishi-Knoll, 1989); and CoAxial Segment volcano (1993) within the Juan de Fuca submarine ridge, among others. Seismic activity associated with some of them is described in Chapter 4.

The best-documented cases of the 1975 Tolbachik eruption (Chapter 5.3.1) (Figure 19.17B) and the 1989 Teishi-knoll eruption (Chapter 5.5.1) allow us to formulate the following characteristics of this type of activity: the seismic swarm continued for about 2 weeks, the largest earthquake occurred during the second peak in the seismic activity, and the eruption of new volcanoes began in a few hours after ceasing or at low background level of the final stage of seismic activity.

Dyke Injections Along Rift Zones

The best-documented case was recorded during the 2005 dyke injection along the Dabbahu-Manda Harraro rift segment (Ethiopia) of the Afar triple junction zone (see Chapter 5.3.2). It was preceded by at least two preliminary earthquake swarms recorded in 1978 and in 1989 (Ayele et al., 2009).

The seismic crisis began on 4 September 2005. Activity subsided but then restarted on September 20. Figure 19.18C shows the development of this main part of the swarm according to the USGS catalog. It continued for about 2 weeks. The largest earthquake occurred during the first peak in the seismic activity. The eruption of new volcanic center began when the seismic activity gradually decreased during its final stage.

Possible Scenarios for Monitoring of Basaltic Dormant Volcanoes

Data presented above allow us to formulate some regularity in the development of seismic activity associated with basaltic eruptions. Periods of seismic activity may be separated into three stages:

Stage 1, preliminary. A sequence of seismic swarms without eruptions; may continue during a few years.

Stage 2, preceding. Seismic swarms occurring just before the eruption; may continue during time interval from a few hours to a few weeks.

Stage 3, continuous. A sequence of seismic swarms occurring after the initial eruption with formation of new volcanic centers; may continue for a few years.

Table 19.1 Characteristic Durations of Three Stages in Seismic Activity Associated with Basaltic Eruptions

Type of eruption	Stage 1	Stage 2	Stage 3
Eruption at active volcano		0–120 h	
Newborn volcano		300–500 h	Two–three months
Re-awakening volcano	Some years	20–30 h	
Rift intrusion	Some years	400–600 h	Some years

Table 19.1 shows the presence of these stages (or part of them) for different types of eruptions.

Now we can discuss a possible strategy for monitoring of basaltic dormant volcanoes. We have a seismic swarm in a volcanic zone where no seismic activity was recorded for a long time. According to the characteristics of seismic activity described above and shown in Figure 19.18 and in Table 19.1, it may be a swarm preceding the birth of new volcano, or preliminary swarm before re-awakening of dormant volcano or before the dyke injection, or simply it may be that the earthquake swarm was associated with failed eruption.

Our strategy may be illustrated for the case of the 2009 Saudi Arabian swarm at Harrat Lunayyir (Figures 19.12 and 19.13). The first option may be that this seismic swarm is preliminary, representing the stage 1 from Table 19.1. In this case, the probable scenario would be the process similar to the 2005 dyke injections along rift zones in Afar (Figure 19.18C). It is a structure very similar to the Saudi Arabian volcanic rifts and having preliminary seismo-tectonic crisis during 1978 and 1989. In this case, the 2009 Harrat Lunayyir episode would have a continuation of the seismic activity (stages 2 and 3) within Saudi Arabian harraat adjacent to Harrat Lunayyir and accompanied by the surface dyke intrusions with formation of new eruptive centers. Monitoring is needed for the following few years.

The next probable scenario, also stage 1, may be the re-awakening of the old Harrat Lunayyir volcano. This process, observed at Akademia Nauk volcano (Figure 19.18A), was characterized by three preliminary low-magnitude (up to $Ms = 5.5$) and shallow earthquake swarms during 8 years before the volcano re-awakening. The appearance of new earthquake swarms within the Harrat Lunayyir lava field may be a signal of this type of scenario. Monitoring is needed for the following several years.

The birth of a new volcano at Harrat Lunayyir is less probable. This activity (Figure 19.18B) would be characterized by a 2- to 3-week earthquake swarm with the maximum magnitude earthquakes of about 5–5.5 and culminated in the birth of a new volcano. From this point, the 2009 swarm had finished already without the appearance of a new volcanic structure within Harrat Lunayyir lava field, and the process was over.

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