# Volcanology

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## GLOSSARY

**Ash** Volcanic ejecta composed of fragments of glass, crystals, and lithics finer than 2 mm in diameter.

**Base surge** Highly turbulent, dilute cloud of volcanic ash, air, and steam, which expands radially away from the base of an eruption column at high velocity.

**Caldera** Circular depression, generally 5 to 20 km in diameter, formed by collapse or subsidence of the central part of a volcano.

**Column collapse** Condition that occurs when an explosive eruption column, composed of volcanic gases, tephra, and air, becomes denser than the ambient atmosphere and collapses to ground level.

**Dome** Roughly symmetrical and hemispherical mound of lava formed during extrusion of magma that is so viscous that it accumulates over the vent.

**Eruption column** Vertical jet of tephra and gases formed during an explosive volcanic eruption that mixes with air and may rise to stratospheric heights.

**Fallout** Settling and deposition of tephra from the atmospheric plume of an explosive eruption.

**Glowing avalanche** Incandescent and turbulent mixture of pyroclastic particles, volcanic gases, and air that flows under gravity down the flanks of a volcano.

**Hawaiian eruption** Eruption of highly fluid and low-viscosity magma that forms thin and widespread lava flows and very minor pyroclastic deposit around the vent or fissure.

**Nuee ardente** Density current consisting of hot pyroclastic fragments, volcanic gases, and air that flows at high speed from the crater region of a volcano.

**Phreatomagmatic eruption** Volcanic eruption that results partly from the interaction of external water and magma, resulting in explosive conversion of water to steam and disruption of the magma.

**Pillow lava** Type of lava flow resulting from the eruption of magma under water, with formation of bulbous masses of lava due to the chilling of magma in water.

**Plinian eruption** Explosive eruption that generates a high and sustained eruption column producing fallout of pumice and ash.

**Pumice** Low-density fragments of silicate glass containing a large fraction of voids due to exsolution of gases from magma during eruption.

**Stratovolcano** Volcanic landform produced by alternating explosive and lava eruptions resulting in a volcano with alternating layers of lavas and pyroclastic deposits.

**Tephra** Collective term for all fragmentary rock material produced by volcanic eruptions.

**Vesicle** Cavity formed in a volcanic rock as a result of exsolution and expansion of gas phase when the rock was a magma.

**VOLCANOLOGY** is the study of the ascent of magma (molten rock) through the planet's mantle and crust and its eruption on the planet's surface. It deals with the chemical and physical evolution of magmas and their transport and eruption. The evolution of volcanos through geologic time is studied by volcanic stratigraphy, whereas physical volcanology is the study of the mechanisms of magma ascent and eruption. The monitoring and prediction of volcanic eruptions is primarily carried out by geophysical measurements.

## I. INTRODUCTION

The science of volcanology is the study of the transport of magma and its eruption on the surface. It is thus closely allied to the science of petrology, which is the study of the generation of magmatic liquid during melting and its modification and eventual disappearance during crystallization or quenching. The terrestrial planets are generally composed of a thin outer crust; a solid mantle, which typically forms the dominant mass of the planet; and a dense iron-rich core, which may be liquid as in the case of the earth. Owing to a variety of chemical and physical processes, localized melting may occur in the outer mantle or lower part of the crust, leading to formation of molten rock or magma. Volcanism is the dominant process of transfer of matter from planetary interiors to the surface and thus plays a crucial role in geochemical cycles and the origin and evolution of continents, oceanic crust, hydrosphere, and atmosphere. Volcanic activity has both a beneficial and a harmful impact on human ecology. Soils in volcanic regions are unusually fertile because of their open texture and high solubility of nutrientforming elements. Other direct benefits of the ascent of magmas into the planetary crust are hot springs and other geothermal reservoirs that can form important energy resources. Because of their high heat flux and well-developed hydrothermal circulating systems, the root zones of volcanoes are regions of mineralization and often the sites of economic deposits of certain ores. The best-known aspect of volcanic activity is, however, the hazard related to volcanic eruptions, when the sudden or explosive release of heat and energy can have devastating effects and lead to large loss of life and property in regions close to the erupting volcano. Due to the emission of aerosol-forming volcanic gases, which lead to screening of the earth from solar radiation, volcanic activity can have a significant impact on the earth's climate over a period of months or years following very large eruptions.

Progress in the understanding of the causes of volcanic activity and the formation of volcanic rocks was not made until late in the 18th century, when the fierce debate between two schools of geologic thought led to the birth of volcanology. The Neptunists advocated a theory of earth history in which all rock formations, including the rocks later identified as of volcanic origin, were believed to be formed by precipitation from a primeval universal ocean. They denied that the earth had a great store of internal heat and regarded volcanoes as the minor surface manifestation of the combustion of coal within the earth. The Plutonists, on the other hand, recognized the volcanic origin of some ancient rock formations because of their similarities to lavas flowing from active volcanoes and proposed a hot interior of the planet to account for many geological phenomena. The true source of heat was not revealed until 1909, however, with Joly's discovery of radioactivity and his proposal of heat generation by radioactive decay in the interior of the earth. In the early part of the 19th century another controversy arose over the origin of volcanoes as landforms. Proponents of the theory of "craters of elevation" held that volcanoes were formed by cataclysmic upheaval of the earth's crust due to the upwelling of molten rock or magma from depth, whereas those of the opposing view maintained that volcanoes were built up by the gradual accumulation of lavas and pyroclastic deposits around the vent. This debate was terminated by a submarine eruption in the Mediterranean in 1831, which led to the buildup of the volcanic island of Ferdinandea, or Graham Island, by deposition of tephra around the vent.

## II. GENERATION AND ASCENT OF MAGMA

The generation of magma may be attributed to changes in pressure, temperature, or composition within the earth. It is evident from the highly systematic distribution of volcanoes on the earth's surface that conditions for magma generation are restricted to well-defined zones of the planet's interior. Volcanoes occur principally in three structural settings (at divergent and convergent plate margins and at intraplate hot spots), and the specific magma generation processes operating in each setting are highly dependent on plate tectonics (Fig. 1).

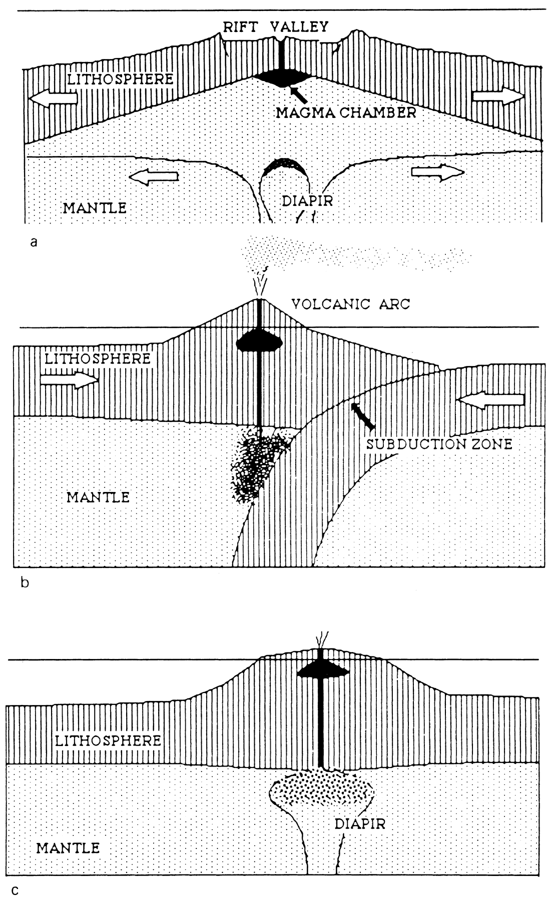


**FIGURE 1** The three principal geologic settings of volcanism on earth: (a) midoceanic ridge, (b) convergent margin, and (c) hot spot. At divergent plate margins, such as midoceanic ridges, the lithospheric plates are separating at rates of 1 to 10 cm/year, resulting in upwelling of mantle material and generation of magma. At convergent margins, the subduction of a lithospheric plate into the mantle leads to magma generation and eruption of volcanics, which form island arcs. Melting anomalies or hot spots in the earth's mantle lead to uprise of partially molten mantle diapirs and magma generation, which leads to buildup of oceanic volcanic islands on the overlying plate. The movement of the lithospheric plate with respect to the stationary mantle hot spot results in the formation of an island chain, such as the Hawaiian Islands.

### A. Midoceanic Ridges

The great majority of volcanism occurs at divergent plate margins, where volcanic activity leads to the buildup of the midoceanic ridges and accounts for about 80% of all volcanic activity on earth. Most of this activity goes unobserved by humans, however, because of the submerged character of the midoceanic ridges. The global midoceanic ridge system is about 60,000 km in length and generates new oceanic crust at a rate of 3 $km^2$/year. Thus volcanism along this type of plate boundary produces an estimated 3 to 6 $km^3$ of rock annually (layer 2). One consequence of the divergence of plates at midoceanic ridges is the upwelling of asthenospheric mantle material from depth, which in turn leads to generation of magma. Two observations are important for the understanding of magma generation under the midoceanic ridges. First, the solidus or minimum melting curve for the asthenosphere mantle has a positive slope; that is, it increases in temperature with increasing pressure or depth in the earth (Fig. 2). Second, the mantle geotherm (i.e., the observed distribution of temperature with depth) is uniformly lower than the solidus; hence the mantle is solid. In regions of asthenosphere mantle upwelling, however, such as under the midoceanic ridges, hot portions of the mantle are brought upward to regions of lower pressure and thus lower solidus temperature. At depths of 30 to 60 km beneath the midoceanic ridges, the upwelling mantle intersects the solidus and partial melting of the peridotite mantle rock begins at 1300° to 1400°C. The melting process is thus not due to increase

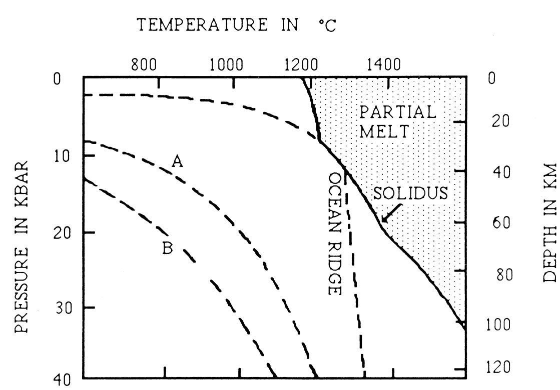


**FIGURE 2** The melting relationship of peridotite, the principal rock type in the earth's mantle. The curve that defines the beginning of melting of peridotite (solidus) is sensitive to pressure and temperature. Curves A and B show the temperature distribution with depth in the earth under the oceans and continents, respectively. They show that the temperature distribution in the earth is such that the mantle is solid. In anomalous regions, such as under ocean ridges, where upwelling of mantle occurs, the geotherm is locally elevated, resulting in partial melting and magma generation. [Adapted with permission from Morse, S. A. (1980). "Basalts and Phase Diagrams," Springer-Verlag, Berlin/New York.]

in temperature, but rather is a consequence of decrease in pressure, which results in a lower solidus temperature of the mantle. The partial melting process of the mantle rock produces basaltic magma that represents melting of up to 20% of the mantle rock involved. The mantle under the midoceanic ridges is thus analogous to a layer of refractory sand that is permeated by a hot intergranular magmatic liquid. When the fraction of melt reaches a few volume percentages of the upwelling mantle region, the mantle becomes permeable and the intergranular films of magma will begin to segregate into larger pockets and rise toward the surface. Ascent and segregation are aided by a density contrast of 0.4 to 0.6 g/cm$^3$ between the lower density basalt magma and the denser peridotite mantle residue. The buoyancy-driven ascent rate of magma along perpendicular cracks in a pressure gradient is approximated by the following equation:

$$v = (\partial - \partial')g\Delta x^2/12n,$$

where $\partial$ and $\partial'$ are densities of mantle and magma, respectively; $g$ is the gravitational constant; $\Delta x$ is the width of the crack through which the magma flows; and $n$ is magma viscosity. Because information on actual crack width is generally lacking, ascent velocities cannot be accurately estimated. Magma ascent velocities can, however, be independently inferred from the size of peridotite mantle xenoliths carried to the surface in magmas, giving values for $v$ of the order 50 cm/sec (1.8 km/hr). Magma ascent is also driven by overburden squeeze (i.e., the difference in pressure generated by the rock column and the magma column). Because of the lower density of magma compared to the surrounding rocks, the overburden squeeze may sustain a magma column that is higher than the rock column. As shown in the example for a magma column of 60 km length in Fig. 3, the magma must rise 4.7 km above the reference surface to equalize the pressure at the base of the columns.

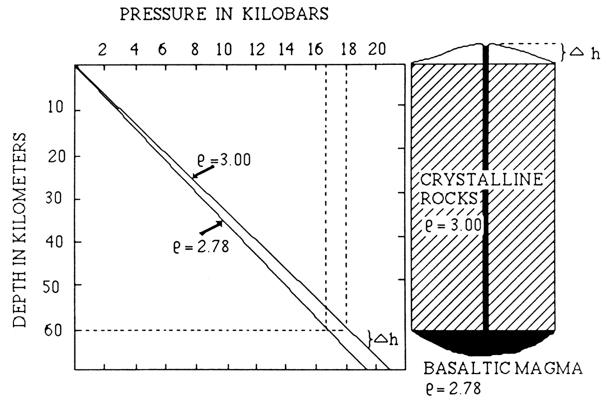


**FIGURE 3** The rise of magma to the earth's surface as a result of hydrostatic pressure. The difference in pressure generated by the rock column and that generated by the magma column is one of the forces that causes magma rise to the surface. At 60-km depth, the pressure at the base of a rock column with density 3.00 g/cm$^3$ is 18 kbar and that at the base of the magma column with density 2.78 g/cm$^3$ is 16.7 kbar. To equalize the pressure at the base of the columns, the magma must rise 4.7 km above the reference level ($\Delta h$) defining the theoretical height of the volcano. [From Yoder, H. S. (1976). "Generation of Basaltic Magma," Adapted with permission of the National Academy of Sciences, Washington, DC.]

## B. Volcanic Arcs

Processes occurring at convergent plate margins lead to generation of magmas and construction of volcanic arcs [Fig. 1(b)]. With a total length of 36,000 km, the subduction zones at convergent plate margins have a mean subduction rate of 8.3 cm/year and consume oceanic crust at a rate of 3 km$^2$/year or equal to crustal generation rate at midoceanic ridges. Magma generation and volcanic processes at convergent margins vary as a function of rate of subduction, composition, and age of the converging plates. At plate margins where the overriding plate is composed of relatively dense oceanic lithosphere, the volcanic arc edifice may be largely submerged, with only the upper part emerging above sea level as a volcanic island arc, such as the Lesser Antilles, Aleutian, and Marianas island arcs. Where the overriding plate is composed of lower-density continental lithosphere, the arc edifice is constructed on a continental margin and is wholly emergent above sea level, such as the Andean, Central American, and Cascades volcanic arcs.

In volcanic arcs the generation and ascent of magma is a consequence of asthenospheric mantle flow and compositional changes induced by the subduction process. Drag forces imposed by the subducting slab induce overturn of the overlying asthenosphere mantle wedge, bringing hotter mantle material upward toward the volcanic arc. Furthermore, dehydration reactions in the oceanic crustal layer that veneers the subducting slab result in migration of volatile components, such as water and $CO_2$, from the slab to the overlying mantle wedge. The resulting partial hydration of the mantle wedge beneath the arc has the important effect of lowering the solidus of the mantle peridotite and induces partial melting. Magma generation in the mantle beneath volcanic arcs may thus be attributed to both ascent of hotter mantle due to subductioninduced overturn and to compositional change of the mantle wedge between the slab and the volcanic arc due to migration of fluxing components from the slab, which suppress the mantle solidus below the geotherm (Fig. 4). In oceanic island arcs the mantle is the dominant source of magmas, which are principally basaltic and similar to midoceanic ridge magmas. Because of the gradual accumulation of a stationary volcanic arc edifice on the earth's surface above

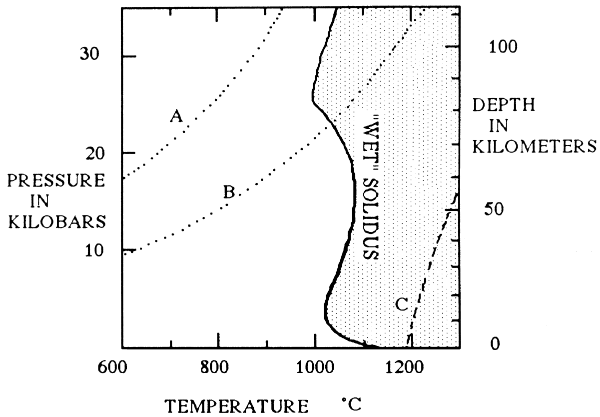


**FIGURE 4** Melting relationships for mantle peridotite in the presence of a few tenths of a percent of $H_2O$, as envisaged for magma generation under volcanic arcs. In the presence of excess water, the beginning of melting curve ("wet" solidus) is suppressed to lower temperature than the "dry" solidus (curve C). Curves A and B show the geotherms for the mantle under the continents and oceans, respectively. It is evident that the subduction of lithosphere into a water-bearing mantle in the oceanic regions would result in intersection of the "wet" solidus and generation of magma. [Adapter with permission from Wyllie, P. J. (1979). *In* "The Evolution of the Igneous Rock" (H. S. Yoder, ed.), p. 503, © Princeton Univ. Press.]

the zone of magma generation, the basaltic magmas become ponded at the base of the lower density crust and differentiate chemically to andesites and other silicic and volatile-rich magmas. Thus mature arcs are dominated by eruptions of andesite and dacite magmas, which are both more viscous and richer in volatiles than basaltic magmas and therefore tend to erupt explosively at the surface. In volcanic arcs where the overlying plate is composed of continental lithosphere, such as the Andean arc, fusion of continental material contributes to the generation of andesitic, dacitic, and rhyolitic magmas, which are typically also viscous and volatile rich.

There are certain segments of convergent margins where subduction is not accompanied by volcanic activity, such as the central Peru region and southwest Honshu, Japan. In these regions the subducted slab is anomalously buoyant and therefore underplates the overlying continental lithosphere without an intervening wedge of hot, mobile asthenospheric mantle.

### C. Hot-Spot Volcanism

Localized melting anomalies in the mantle lead to volcanism in various regions of the earth that are independent of plate margins. These melting anomalies, which are generally referred to as hot spots, are of unknown origin, but are frequently attributed to the upwelling of deep mantle plumes (Fig. 1c). Most hot spots occur beneath lithospheric plates and form intraplate volcanoes in oceanic regions, such as Hawaii, or in continental regions, such as the Yellowstone volcanic complex. The systematic age progression of volcanic islands away from a hot spot, such as in the Hawaiian–Emperor chain, is evidence that the hot spot is sited in the mantle underlying the moving lithospheric plate. More rarely, hot spots are located on midoceanic ridges, such as Iceland and the Azores on the Mid-Atlantic Ridge. Magma generation in hot spots occurs owing to mantle upwelling, analogous to the situation under midoceanic ridges. The ascent of mantle diapirs in a pressure gradient leads to partial melting of mantle peridotite and generation of basalt magma. In intraplate hot spots the driving force of mantle upwelling cannot be attributed to plate divergence, as in the case of midoceanic ridges, and the primary cause of the upwelling or proposed plume activity under hot spots may be due to thermal and chemical anomalies in the deep mantle. The deep origin of hot spots is supported by the observation that they remain relatively stationary in the earth, in contrast to the mobile lithospheric plates. The influence of a single hot spot may extend over a region of several hundred kilometers on the earth's surface and result in the formation of a cluster of volcanoes.

## III. MAGMA CHAMBERS

The ascent of magma toward the earth's surface occurs in response to a pressure gradient that is the result of a density difference between the magma and surrounding rocks. As magma enters the lower density crust, the density difference is diminished or may disappear altogether, resulting in stagnation of magma in the crust or at the crust–mantle interface. In addition to these theoretical arguments for the ponding of magmas within the crust, based on density relations, there is a wealth of evidence for the existence of crustal magma chambers, indicating that the transport of magma from deep source regions to the surface is typically a two-stage process and that magmas may have a long residence time in such shallow reservoirs. Chemical processes operating in magma chambers, such as fractional crystallization, lead to changes in the composition of the magma, from primitive to evolved types. These processes are accompanied by loss of heat from the magma and crystal accumulation at the base and walls of the chamber and may result in a compositionally stratified magma reservoir, with lower density evolved magma residing in the roof zone, above higher density primitive magma. Such differentiation processes are also important in leading to enrichment of volatile components in the evolved magma, which will influence its mode of eruption.

Magma chambers have been postulated at depth of 2 to 5 km under midoceanic ridges, 4 to 8 km wide, and elongated in the direction of the ridge axis, but discontinuous at 40- to 50-km spacing. They receive a steady-state supply of primitive basaltic magma from the peridotitic mantle upwelling under the ridge axis. Under fast-spreading ridges, such as the East Pacific Rise, the magma supply rate is high and the chambers are long and well developed, but truncated at the fracture zones that transect the ridge axis at intervals of 50 to 100 km. Heat is transferred from the magma reservoir by hydrothermal circulation of seawater through the overlying rifted crust and by conduction to the surrounding lithosphere. Such hydrothermal circulation between the ocean and the volcanic crust has important effects on the chemistry of seawater. The heat loss results in crystallization of high-temperature minerals in the magma, crystal nucleation and growth on walls and roof of the reservoir, and crystal settling to its floor. The process leads to modification in the chemical composition of the magma, but is repeatedly interrupted by influx of primitive magma to the reservoir, leading to magma mixing. Because of the size of the reservoir and steady-state character of these open-system processes, the reservoir magma maintains a relatively uniform basaltic composition throughout the global midoceanic ridge system. At slow-spreading midoceanic ridges, such as the Mid-Atlantic Ridge, magma chambers are smaller and discontinuous, resulting in greater diversity of magma types. The majority of magma in midoceanic ridge chambers crystallizes on the walls or accumulates as crystal mush at the base, forming gabbroic and ultramafic igneous rocks, or is intruded as dikes into the overlying volcanic layer. A smaller fraction is erupted from the chamber to form pillow lavas of the volcanic layer of the oceanic crust.

Magma chambers in volcanic arcs and in oceanic islands formed by hot-spot activity are better studied than magma chambers of the submerged midoceanic ridges. The largest chambers probably occur within the crust of the continental magmatic arcs, where the eruption of up to 1000 km$^3$ of magma in some eruptions indicates the presence of magma chambers with volumes in excess of this figure. Magma chambers intervene in the magma supply system between the deep magma source in the mantle and the volcano at the surface. In rare cases, a chamber is absent and magma may ascend directly from the source region to the surface in a steady-state fashion. More commonly, magma rises steady state from the source region up to a high-level chamber, which in turn supplies the volcano intermittently. The evolution of the magma chamber and the characteristics of the chamber–volcano magmatic system can be studied from the eruption history and geodesy of the volcano. These systems are characterized by a steady supply of the magma chamber but intermittent tapping from the chamber to the volcano. Thus periods of no eruptive activity (repose periods) may last several years, during which magma mass and pressure may build up in the chamber. Repose periods are followed by eruptions, when magma is tapped from the chamber and erupted at the surface during a relatively short time span. From the study of the eruptive history of an individual volcano, a cumulative erupted volume curve can be constructed (Fig. 5), which generally shows that the periodic tapping of magma chambers follows a steady-state, long-term trend, with volcanic output rates of the order 0.1 to 5 m$^3$/sec. This long-term eruption rate must be significantly lower than the steady-state supply rate from the source region to the chamber, since some of the magma crystallizes in the chamber or forms shallow intrusive bodies. In general, the longer the repose period between tapping and successive eruptions, the greater the explosivity and volatile content of the subsequent eruption. Volatile-rich magma is thus generated when the system is closed to the upward loss of magma. Volatile enrichment may occur because of upward diffusion through the magma chamber or because of volatile diffusion from the surrounding crust. Many other chemical constituents show systematic vertical zonation in magma chambers, and these are accompanied by gradients in physical properties of the magma, such as increasing temperature and density with depth. Consequently, the magma tapped during a single eruption may show progressive compositional and physical changes, which in turn may affect the mode of eruption.

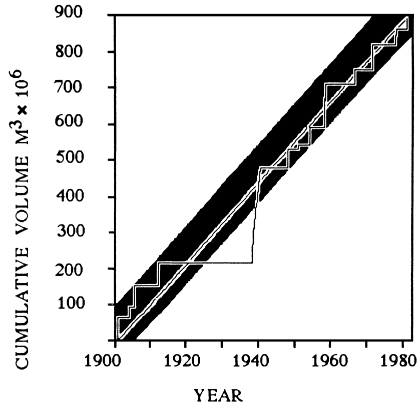


**FIGURE 5** The volume eruption rate of Nyamuragira volcano in central Africa for the period 1900 to 1980. Except for the large eruption of 1938–1940, the behavior of the volcano is characterized by frequent and small eruptions leading to a steady-state volume eruption rate of 0.35 m$^3$/sec. The shaded region shows the maximum eruption volume per eruption likely for the steady-state condition (150 × 10$^6$ m$^3$). [Adapted with permission from Wadge, G. (1982). "Steady state volcanism." *J. Geophys. Res.* **87**(B5), 4037, Copyright by the American Geophysical Union.]

Important information on magma chamber evolution comes from geodesic surveys of the overlying volcano.

Repeated surveys of volcanoes such as Kilauea (Hawaii), Krafla (Iceland), and Campi Flegrei (Italy) have shown major ground uplift of central region of the volcano, up to 4 mm/day, in response to buildup in magma pressure and inflation of the magma chamber due to steady-state flux of magma into the chamber from a deep source region. Inflation episodes may have a duration of months or years, resulting in gradual elevation of the volcano's edifice by several meters. Inflation is suddenly terminated as the confining walls of the chamber fail due to excess magma pressure, leading to eruption or intrusion of magma into the surrounding crust and rapid deflation. In regions where the volcanic edifice is under horizontal tension, such as in rifted volcanic zones, the majority of magma tapped from the magma chamber may flow into tensional fractures in the crust and solidify as dikes, whereas only a small fraction may emerge at the surface. The process is repeated many times, with resumed inflation, followed by attainment of a critical magma pressure and chamber failure.

## IV. ERUPTION MECHANISMS

The tapping of a magma chamber and the triggering of a volcanic eruption may result from a complex interaction of several factors. Due to the steady-state influx of magma into the chamber from depth, there is a gradual buildup in magma pressure within the chamber, which can lead to the triggering of an eruption after the chamber pressure exceeds both the overburden pressure and the lithostatic tensile strength of the surrounding crust. This process is particularly important in rifted volcanic regions, such as the midoceanic ridges, where the buildup in magma pressure is minor before tapping of the reservoir occurs, and the eruptions are consequently frequent and of relatively small mass. Buoyancy of magma is another factor that may trigger an eruption after a density contrast has been established between magma and denser crust. Such a density contrast may arise because of fractional crystallization or other processes that lead to compositional change of the magma and result in the formation of less dense and evolved magma, which accumulates in the roof zone of a magma chamber. Eruption triggering due to oversaturation in volatile components of magma may also occur and lead to exsolution of a gas phase in the magma chamber, with consequent volume expansion, overpressuring, and increased buoyancy contrast.

Magma may rise out of the chamber along preexisting conduits, or it may produce its own fissure by hydraulic fracturing if there is an adequate local excess pressure in the magma. Because of the high-temperature contrast between magma and the cold surrounding crust and the increase in magma viscosity with falling temperature, there is a cooling limitation on the tendency of magma to erupt. A basaltic magma rising at 0.12 m/sec from 5 km depth in a pressure gradient with a density contrast of 200 kg/m$^3$ and with a viscosity of $10^2$ Pa/sec will be erupted if the conduit diameter is greater than 0.44 m; magma rising in a narrower conduit would freeze before reaching the surface. The cooling limitation is more severe for silicic magmas that are more viscous and cooler. For example, a rhyolite magma with viscosity of $10^4$ Pa/sec and same density contrast and depth of origin must rise along a conduit with minimum diameter of 1.4 m and rise velocity of 12.5 mm/sec in order to reach the surface before solidification and stagnation in the conduit. In addition to the great differences in temperature and viscosity between silicic and basaltic magmas, they generally differ also in the quantity of dissolved volatile gases. Because of the effects these differences excert on the behavior of magmas during ascent and eruption, it is useful to consider them separately. Finally, another important mechanism for explosive eruptions is phreatomagmatic volcanism, where the gas propellant is not only primary magmatic volatiles, but also external water encountered by the rising magma.

### A. Silicic Magma

The eruption of silicic magma, such as rhyolite and dacite, is in most cases driven by a combination of buoyancy and exsolution of a gas phase from the magma. The dominant volatile component dissolved in silicic magmas is water, whose solubility is strongly dependent on pressure (Fig. 6). Solubility of water in a rhyolitic magma as a function of pressure is given by the following expression:

$$N = 0.13 P^{0.5},$$

where $N$ is the weight percentage of water dissolved in the magma and $P$ total pressure in bars. Carbon dioxide and $SO_2$ are also present as volatile components in silicic magmas, but in fractions of such low weight that they are generally unimportant in eruption dynamics. The weight fraction of dissolved volatiles (mainly water) is the most important variable in determining the style of eruption of silicic magmas. Water-rich magmas exsolve a vapor phase during ascent, and the progressive volumetric expansion of the vapor phase with decreasing pressure results in acceleration of the magma and gas mixture up the conduit, leading to exit velocities of several hundred meters per second through the volcano's vent. Another important property of magmas is the viscosity. This physical property is strongly dependent on magma composition, water content, temperature, and crystal content. Viscosities of magmas vary over seven orders of magnitude, but silicic magmas are typically $10^4$ to $10^7$ P (Fig. 7). The viscosity influences the rheological properties of magma, such as

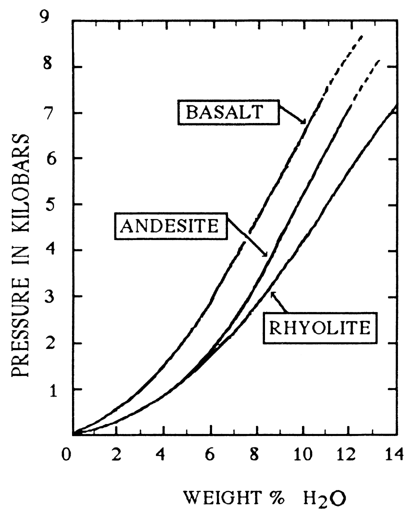


**FIGURE 6** Relationship between total pressure and the solubility of $H_2O$ in basaltic (1100°C), andesitic (1100°C), and rhyolitic (750°C) magmas. A rhyolitic magma at 2 kbar (approximately 7 km deep in the earth's crust), for example, could contain about 7 wt.% $H_2O$ in solution. Ascent of this magma to lower pressure would result in supersaturation and exsolution of the vapor phase. [Adapted with permission from Burnham, C. W. (1974). *Bull. Soc. Fr. Mineral. Cristallogr.* **97,** 224.]

ascent rate, convection, laminar or turbulent flow, and the rate of escape of gases. In low-viscosity or fluid basaltic magma, the exsolving gas phase expands freely, overpressures do not build up within the vesiculating magma, and the eruption is nonviolent. In contrast, the vesiculating gas phase in highly viscous silicic magma is limited in its volumetric expansion because of the inhibiting influence of high viscosity of the liquid. Significant overpressures can thus build up within the gas vesicles, as volumetric expansion of the gas phase does not keep up with the reduction in lithostatic pressure during magma ascent. Bubble bursting occurs when critical internal overpressures have been attained in the gas vesicles, resulting in violent gas expansion and explosive eruption.

The Mount St. Helens eruption of May 18, 1980 is an example of the ascent and eruption of silicic magma (Fig. 8). During the 9-hr main phase of the eruption, dacite magma rose from a reservoir at about 7 km below the volcano, into a 100-m-diameter conduit. The magma, which consisted of a mixture of dacite liquid and about 40 vol% crystals, was at 930°C and with a viscosity of $2.3 \times 10^6$ P. From the observed magma discharge rate of about $2 \times 10^7$ kg/sec through the vent, a magma ascent velocity of 1 m/sec can be inferred. With water content of 4.6 wt.%, the magma

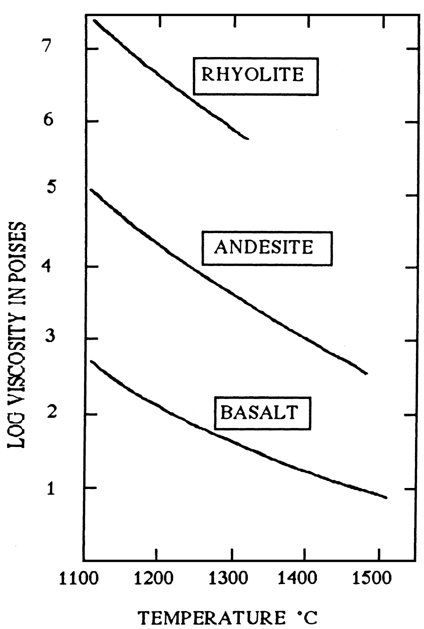


**FIGURE 7** The viscosity of rhyolite, andesite, and basalt magmas as a function of temperature. [Adapted with permission from Murase, T., and McBirney, A. R. (1973). *Geol. Soc. Am. Bull.* **84,** 3563.]

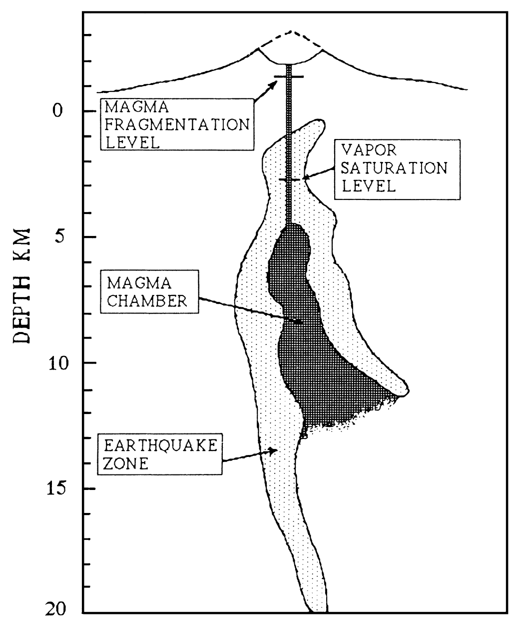


**FIGURE 8** Schematic cross section of the magmatic system of Mount St. Helens volcano in Washington State at the time of the May 18, 1980 eruption. The top of the magma chamber is located at depth of 7 km below the original summit of the volcano. The chamber is surrounded by a zone of weak earthquake activity due to thermal and mechanical stresses from the magma. Water-undersaturated magma flowed out of the chamber and ascended the conduit. With decreasing pressure, the magma reached the vapor saturation level at 4.5 km depth, and exsolution of vapor began. At about 570 m depth the volume fraction of vapor in the magma reached 75%, and the magma was disrupted and explosively erupted from the vent. [Adapted with permission from Carey, S., and Sigurdsson, H. (1985). *J. Geophys. Res.* **90,** 2948.]

in the chamber would be water undersaturated and behave as single-phase flow in the lower part of the conduit. At about 4.5 km deep below the volcano the magma became saturated with respect to water, and exsolution of a gas phase began with formation of minute vapor bubbles in the dacite liquid. Above this saturation level the magma rises in the conduit as a two-phase fluid, consisting of a silicate liquid (the continuous phase) and a gas phase, which is continually expanding in response to the decreasing lithostatic pressure with decreasing depth. Because of the increasing volume of the foaming mixture of magmatic liquid and gas, the magma ascent velocity increases with decreasing depth. When the gas phase, which is contained in millimeter-size bubbles or vesicles enclosed by liquid, constitutes 75 vol% of the magma, the foam disrupts as the gas pressure within the bubbles exceeds the ambient lithostatic pressure and the bubble tensile strength. The exsolution and vapor phase expansion of 4.6 wt.% water with decreasing pressure would have reached 75 vol% of the combined magma—gas froth at a depth of about 580 m in the Mount St. Helens conduit, resulting in explosive disruption of the magma. Above the zone of disruption, the mixture of dacite liquid fragments and the gas is now a two-phase fluid, where the gas is the continuous phase. Within the expanding gas are carried fragments and frothy or vesicular clots of dacite liquid, which will later cool and solidify to pumice and volcanic ash. As the mixture moves up the conduit, gas expansion is further accelerated in the decreasing lithospheric pressure gradient, and this leads to greatly accelerating velocity of the expanding mixture. As the mixture exits the conduit through the vent, it has velocities of 200 to 300 m/sec. During the main explosive phase of the Mount St. Helens eruption, this steady discharge was maintained for 9 hr.

## B. Basaltic Magma

The eruption of basaltic magmas is in general driven primarily by buoyancy due to a density contrast between magma and the surrounding crust and by pressure in the erupting fluid, such as overpressure in the magma chamber in excess of the lithostatic pressure. Volatile exsolution plays a much smaller role in the eruption of basaltic magmas than rhyolitic magmas because of their lower volatile solubility and lower viscosity. The pressure dependence (Fig. 6) of the solubility of water in basalt magma is as follows:

$$N = 0.0215P^{0.7},$$

where $N$ is weight percentage water and $P$ total pressure in bars. The solubility of $CO_2$ is much lower, or

$$N = 0.00023P.$$

The volatile content of most basaltic magmas is in the range 0.2 to 0.5 wt.% and a major part may be carbon dioxide. The influence of this volatile content on eruption mechanics will depend on the geological environment. Thus during ascent and eruption of basaltic magmas at the midoceanic ridges, the site of eruption is at 2 to 3 km deep in the ocean, and basaltic magma will vesiculate only slightly or not at all under such high hydrostatic pressures. Ocean-floor basalts are consequently extruded as single-phase fluids, relatively free of vesicles. Basalt magma ascending through the crust of an oceanic volcanic island will, on the other hand, experience much lower ambient pressures before eruption and consequently may experience both nucleation of a gas phase and magma disruption due to volatile expansion. Eruption of such magma occurs in fire fountains, which are incandescent jets of disrupted magma up to 500 m in height above the vent or eruptive fissure. Basaltic magmas, which may become fragmented during ascent and eruption, will often form liquid lava flows because of the high temperature and low viscosity of the fragments at the surface.

Basaltic eruptions occur from circular vents, but more commonly from fissures, which may be tens of kilometers in length. Mass eruption rates in individual basaltic eruptions may reach $10^7$ kg/sec, which corresponds to a mass eruption rate per meter length of the fissure of $10^3$ to $10^4$ kg/sec m.

## C. Phreatomagmatic Processes

Volcanic eruptions involving the explosive interaction of external water and magma are referred to a phreatomagmatic eruptions. Magma ascending through the earth's crust may intersect a body of water, such as a lake, the ocean, or a porous rock formation with a well-developed aquifer system. If the interaction of magma and water occurs in a confined system, such as a volcanic conduit, and if the magma is already fragmented by exsolution of primary magmatic volatiles, the interaction leads to boiling of the external water and very high gas pressures in the confined but superheated vapor. At ambient pressures corresponding to about 10-m water depth, for example, the volumetric expansion of water vapor is about 1600 times the volume of liquid water. The resulting phreatomagmatic explosion is a short-lived event, but can be repeated continuously as long as the condition for magma ascent and influx of water to the conduit is maintained. The process is dependent on the several-hundredfold volumetric expansion of water vapor upon boiling and superheating at low ambient pressures. The process cannot occur, therefore, in the environment of high ambient pressures at the mid-oceanic ridges, where the hydrostatic pressure is so high that the volumetric expansion of water upon boiling is negligible. At water depths less than 300 m, the process becomes important, however, and phreatomagmatic eruptions are characteristic for the emerging stage of volcanic seamounts (i.e., the transition to a volcanic island). In addition to relatively low ambient pressure, the phreatomagmatic process requires intimate mutual penetration by magma and water and a large surface area for rapid and efficient transfer of heat from the magma. This mixing of magma and external water can be realized only if the magma is already fragmented in the conduit (i.e., due to vesiculation of magmatic gases and bubble bursting). Phreatomagmatic explosions will not occur, on the other hand, during the interaction of lava flows and water, where interpenetration is negligible and confining pressures nonexistent, resulting in a dissipated boiling front around the lava. In shallow-water explosive basaltic eruptions, such as the 1964 Surtsey eruption off the coast of Iceland, it is clear that phreatomagmatic processes drive the explosive eruption and cause extreme fragmentation of the magma. In silicic eruptions, on the other hand, it is difficult to assess the role of magmatic volatiles versus external water in the explosive activity, but some characteristics of the resulting volcanic deposits, such as accretionary lapilli and vesiculated fine ash, can be an indication of phreatomagmatic processes, as discussed in the next section.

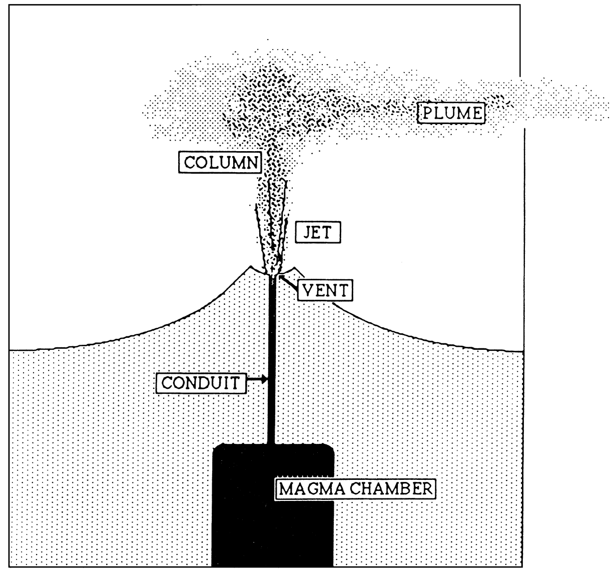


**FIGURE 9** The components of an explosive volcanic eruption, with a mixture of pyroclastic material and volcanic gases emerging from the vent, forming a jet by force of the gas thrust. At height of 1 to 3 km above the vent, the column rise is due to bouyancy as atmospheric air is incorporated and heated by the pyroclastic mixture, thus leading to rise of the column as a hot-air balloon. As the column looses heat, it equilibrates thermally with the high atmosphere and ceases to rise. At this level the column is sheared by the prevailing wind and forms an eruption plume elongated downwind.

## V. ERUPTION COLUMNS

Volcanic explosions are the result of the escape of gases at high velocity from the vent, and they are usually accompanied by the ejection of large quantities of disrupted magma and solid rock fragments. The detailed behavior of the eruption column depends on many factors such as volatile content, magma viscosity, vent radius, and mass eruption rate of the magma. The height attained by an eruption column in the earth's atmosphere is thus primarily determined by the mass eruption rate. For an explosive eruption where magma ascent is largely driven by a buoyancy force, the mass eruption rate (MER) is expressed as follows:

$$\mathrm{MER} = r^4 \partial_\mathrm{m}(\partial_\mathrm{c} - \partial_\mathrm{m})g/8n,$$

where $r$ is radius of the conduit, $\partial_\mathrm{m}$ magma density, $\partial_\mathrm{c}$ the density of the crustal rocks surrounding the conduit, $g$ the acceleration of gravity, and $n$ the viscosity of the magma. Typical values of mass eruption rate during explosive eruptions are $10^5$ to $10^9$ kg/sec, but it is evident from the preceding relationship that small changes in conduit radius can have a large effect on mass eruption rate.

The explosion may be an instantaneous and short-lived event, or it may be sustained as a continuous uprush of gas and magma from the vent for hours or days. Volcanic explosions result in formation of atmospheric eruption columns, typically 15 to 25 km high, but rarely as high as 40 km. Eruption columns consist of two components: the lower component is a jet, formed by gas thrust from the vent (Fig. 9). Its driving force is mechanical energy generated during exsolution and volumetric expansion of volatiles from magma in the conduit, which can lead to exit velocities of the mixture of gas and pyroclasts of 200 to 700 m/sec. Because exit velocity of the jet is largely driven by exsolution of the volatile phase of the magma, it is most sensitive to the magma's original volatile content. As the jet rises above the vent, its velocity decreases due to spreading of the jet, mixing with the atmosphere and, the action of gravity on the pyroclastic particles carried in the jet. Counteracting the force of gravity is the drag force exerted on the particles by the gas rushing upward in the jet, which will maintain smaller particles in suspension in the rising jet, but larger particles will fall out and accumulate around the vent.

The jet phase typically extends 1 or 3 km above the vent, but above that level the eruption column is maintained buoyant by a convective thrust that derives its energy from the heat given off by the jet. During turbulent entrainment and mixing of atmospheric air, which occurs at the periphery of the jet, the air is heated in contact with the pyroclasts and by mixing with magmatic volatiles in the erupted mixture. The thermal energy transfer to entrained atmospheric air and the resulting expansion gives the column buoyancy in the atmosphere and results in convective rise at rates of the order 50 m/sec. The maximum height attained by the eruption column is proportional to the fourth root of the mass eruption rate. The exact relationships between column height and mass eruption rate depend on the efficiency of heat transfer from pyroclasts to the entrained air. In the case where 70% of the heat is used to drive the convective rise of the column, the relationship between eruption column height and mass eruption rate is as follows:

$$H = 236.6m^{1/4},$$

where $H$ is eruption column height and $m$ the mass eruption rate in kilograms per second. As shown in Fig. 10 the observed column height and mass eruption rate for several historic eruptions indicate a high efficiency in utilization of the heat driving the column. High efficiency is generally the case where fragmentation of magma is extreme and a large surface area therefore exists between fragments and the entrained air. In eruptions where the fragmentation is poor, the rate of heat loss from fragments to air entrained in the column is lower. In phreatomagmatic eruptions, magma heat utilization is also inefficient and a lower eruption column will result, as much of the thermal energy is used in vaporizing external water. Recovery of this energy can occur, however, if condensation occurs in the column. Phreatomagmatic eruptions are consequently often accompanied by mud rain, resulting from mixing of condensing water vapor and volcanic fragmentary matter.

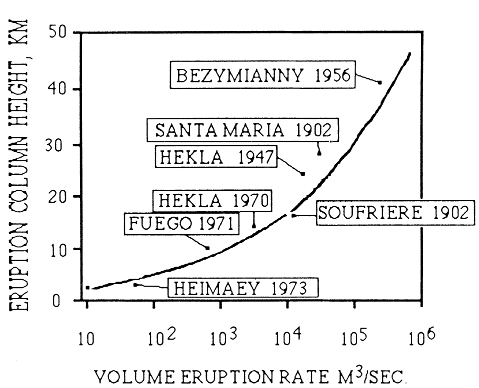


**FIGURE 10** The relationship between eruption column height and the volume eruption rate of magma observed for seven historic eruptions. The curve shows the theoretical relationship when a 70% efficiency of heat usage in the eruption column is assumed. It is evident that in the case of the historic eruptions the heat efficiency is closer to 100%. [Adapted with permission from Wilson, L., Spark, R. S. J., Huang, T. C., and Watkins, N. D. (1978). *J. Geophys. Res.* **83,** 1829.]

The rise of an eruption column will cease when the top of the column enters heights in the atmosphere where it is neutrally buoyant. Although this level is primarily controlled by the mass eruption rate, it is also influenced by the gradients of temperature and density in the atmosphere. At this level the eruption column will pond and spread laterally into a mushroomlike cloud (Fig. 9). The eruption cloud will then become sheared by the prevailing winds, resulting in formation of a ribbonlike eruption plume, which may extend downwind from the column over thousands of kilometers. Because of the complex wind patterns in the earth's atmosphere, the shearing of the eruption cloud and development of the plume is a highly variable process. Most large eruption columns extend above the earth's tropopause at 10 to 15 km high and into the stratosphere. Their eruption plumes are, therefore, sheared and dispersed mainly by easterly or westerly stratospheric wind currents. Velocities of the volcanic plume front in the stratosphere are largely dependent on atmospheric wind velocities and range from 100 to 250 km/hr. In some eruptions, such as the 1982 explosive eruption of El Chichon in Mexico, the eruption column may stagnate near the tropopause, and shearing and dispersal of the eruption cloud are thus dependent on both stratospheric and tropospheric winds. In Mexico these winds had opposite polarity vectors, generating two plumes that were dispersed in opposite directions.

The behavior of eruption columns is strongly influenced by gravity and atmospheric pressure, and the calculations discussed above, therefore, apply only to the terrestrial case. On Mars, where many volcanoes occur although none are known to be historically active, the atmospheric pressure is about 0.006 of that on earth and Martian gravity is about one-third of the terrestrial value. Consequently, the initial expansion of a Martian eruption column will be much greater than on earth, and the eruption column height will be about three times greater than a terrestrial eruption column with the same gas content. On the Moon there is virtually no atmosphere and the gas phase from an erupting volcano will meet no frictional resistance and, therefore, expand indefinitely. Furthermore, the lunar gravity is only one-sixth that of earth. Explosive activity on the Moon would consequently result in an extremely widespread but thin pyroclastic deposit.

The mixture of gas and pyroclasts discharged during an explosive eruption is always initially denser than the surrounding atmosphere. The generation of a stable eruption column from this mixture is totally dependent on the amount of air that can be entrained into the jet and heated in contact with the pyroclastic fragments, leading to expansion of the air and convective buoyant uprise of the column.

Under certain conditions the column remains denser than its surroundings, or conditions during the eruption may change, leading to increased column density and gravitational column collapse. The resulting mixture of pyroclasts, gas, and entrained air then form a dense, falling cloud around the vent, which spreads down the flanks of the volcano under gravity as a concentrated pyroclastic flow or more dilute surge cloud. Column collapse may be continuous, and the explosive eruption then produces a sustained fountain of pyroclasts and gas, which maintains a steady pyroclastic flow. Collapse may also be intermittent, with the condition of the explosive eruption fluctuating between a fully buoyant column that generates pyroclast fall and a collapsing column that generates pyroclastic flow or surges. Column collapse occurs for certain combinations of gas content, gas velocity, and vent radius (Fig. 11). Thus increased vent radius and decreased gas content will lead to column collapse. Similarly, column collapse will also occur at very high mass eruption rates because of the decreased efficiency of mixing with atmospheric air. It is likely that the maximum volume eruption rate for stable convecting eruption column is about $10^6$ $m^3$/sec; at higher rates the column will collapse. Many explosive eruptions have a characteristic pattern, beginning with a stable, sustained eruption column, which later collapses and generates pyroclastic flows and surges. This pattern is most likely related to progressive widening of the vent due to gradual erosion by the erupting mixture and decreasing gas content of the erupting magma, as progressively deeper and volatile-poor levels of the magma chamber are being tapped.

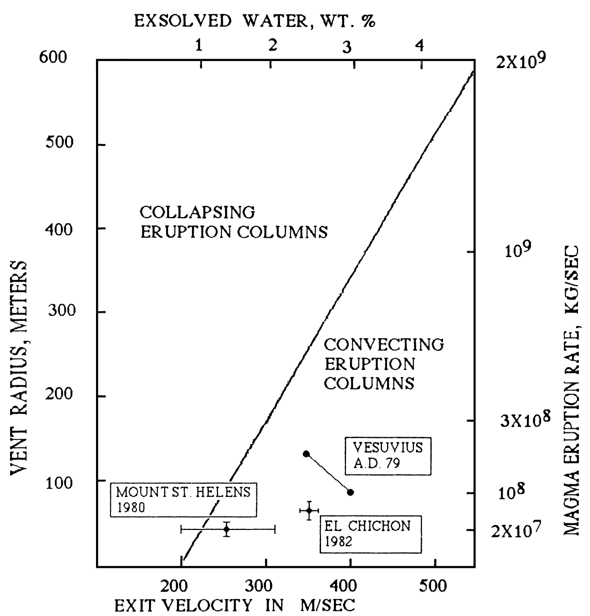


**FIGURE 11** The parameters that define convecting and collapsing eruption columns. Many explosive eruptions begin with convective eruption columns, but changes in one or more parameters—such as decreasing water content of the magma, increasing vent radius (due to erosion), or increasing mass eruption rate—lead to transition to collapsing column, with formation of pyroclastic flows and surges. [Adapted with permission from Wilson, L., Spark, R. S. J., and Walker, G. P. L. (1980). *Geophys. J. R. Astron. Soc.* **63,** 117.]

## VI. VOLCANIC DEPOSITS

The materials brought to the surface by volcanic eruptions accumulate around the vent as volcanic deposits and lead to the construction of a volcanic edifice or a volcano. Two principal types of volcanic deposits are recognized: lavas and pyroclastic deposits. They are both generated by the solidification of magma during eruption, but lavas represent magma that has solidified at the earth's surface from a flow consisting of liquid and suspended crystals, whereas pyroclasts are derived from fragmentation and disruption of magma during explosive eruption.

### A. Lava Flows

During nonviolent eruption of volatile-poor magma, the silicate melt may form a continuous flow from the vent or it may fountain above the vent and contribute to the flow. Lava flows are probably the most common volcanic rock type on earth since they make up most of the volcanic layer of the oceanic crust. Their morphology is varied, however, in response to different rates of discharge, environments, and magma viscosities. Pillow lavas are the most widespread type. They form during subaqueous eruptions, such as on the midoceanic ridges, where basaltic magma at temperatures of 1100° to 1250°C flows onto the ocean floor and comes in contact with cold seawater. The outermost layer of flow is rapidly chilled by the seawater, resulting in the formation of a rigid glass layer that inhibits expansion of the flow. Increased magma pressure leads to the rupture of the glassy skin and breakout of pillow-shaped or sausagelike protrusions of lava typically 0.5 to 1 m in diameter. This budding process is controlled by internal magma pressure in the pillow–tube network and the tensile strength of the glassy pillow skin. Continued activity results in a pillow lava pile around the vent; as the pile steepens, individual pillows break off from the main mass and tumble down slope. When the mass eruption rate in subaqueous eruptions is very high, the magma pressure exceeds the chilling effect of the ambient water and sheet flows are formed instead of pillow lavas. Sheet flows solidify to form smooth, nearly horizontal and pavementlike deposits on the ocean floor that consist of thin multiple units with a glassy surface. The submarine lava flows formed at the midoceanic ridges are produced by relatively small-volume (0.1 $km^3$) and frequent eruptions.

Subaerial basaltic lava flows have volumes ranging up to 20 $km^3$, but flows with volumes up to 1500 $km^3$ have been reported in older geologic formations on earth. Subaerial lava flows are typically 5 to 30 m thick and up to 200 km in length and can be very mobile, as indicated by lavas that have flowed tens of kilometers on slopes less than 1°. Flow length is largely determined by mass eruption rate, although viscosity and heat loss are also important factors. Basaltic lava flows are classified into two principal types according to their surface features. Pahoehoe lava flows are generally thin and have smooth, billowing, glassy, and ropy crusts and a variety of other features that indicate a low magma viscosity. Aa lava flows are thicker and have very rough surfaces composed of a jumble of spiny and vesicular lava blocks. The difference between the two lava types is most likely related to the effect of temperature and gas content on the viscosity of the magma. In many eruptions the magma issues from the vent as a gas-rich pahoehoe, which loses volatiles and heat during flow and changes to more viscous aa flow downslope. The viscosity is also increased by vesiculation and the immersion of blocks of the crust into the flow. When lava flows come to rest and cool, they develop a characteristic internal structure. Pahoehoe flows have a massive lower part and interior, with a prominent pattern of vertical columnar joints that divide the massive rock up into five- or six-sided columns. This structure is due to volume contraction of the lava upon cooling that leads to thermal stresses that propagate joints perpendicular to the cooling surface of the flow. The upper part of pahoehoe flows is gradational to the interior but more vesicular. Aa flows also have a massive, jointed interior, but are characterized by a blocky or rubbly layer both at base and on top. Many pahoehoe flows contain lava tubes or nearly horizontal and hollow tunnels, which may reach tens of kilometers in length. Lava tubes form when a crust develops over a flowing river of lava. With diminishing mass eruption rate, the level of magma in the tube drops and it may drain altogether. Open lava channels or rilles also form long, sinuous and narrow valleys in lava fields on the earth and the moon.

## B. Domes

Magma with high viscosity and low volatile content cannot flow from the vent upon extrusion and will accumulate over the vent as a hemispherical dome (Fig. 12). The rheological behavior of magmas is greatly influenced by the relations between shear stress experienced by the magma and the resulting shear strain or permanent deformation during flow. If a force or stress is applied, the liquid changes shape or suffers strain. The flow behavior of

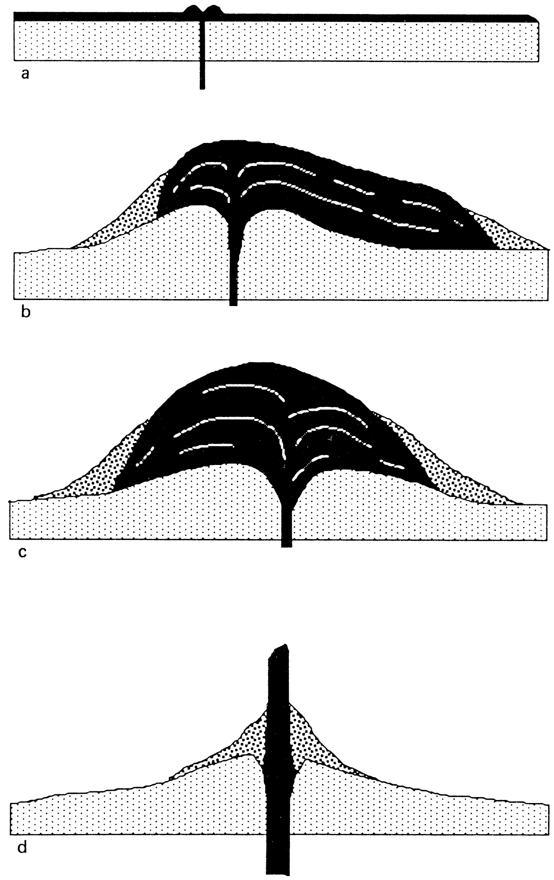


**FIGURE 12** The effect of magma viscosity on the morphologic form of lava extrusions. Low-viscosity basaltic magmas (a) are erupted in highly fluid form and spread widely as thin lava flows. Viscous andesite magmas (b) form short and thick lava flows, often with a rubble of fragmented lava material at the flow front (indicated by large dot pattern). High-viscosity rhyolitic magmas (c) are too viscous to flow and accumulate over the vent as a dome. In rare cases, the viscosity of the rhyolitic magma may be so high that it is extruded in a semirigid state as a spine over the vent (d), surrounded by talus scree or rubble from fragmentation of the spine.

magma in chambers and conduits may be largely Newtonian; that is, the stress is proportional to the rate of strain. In viscous magmas a relatively high shear stress is required to maintain a given rate of strain. In low-viscosity magmas, on the other hand, a low shear stress will maintain the same rate of strain. The rheology of magmas generally changes upon eruption, however, since formation of crystals and vesicles and temperature decrease result in increased viscosity and a change from Newtonian to non-Newtonian behavior. Lavas and other non-Newtonian liquids are characterized by possessing a yield strength; that is, a finite amount of stress (yield stress) must be applied to the lava before flow or deformation will occur. The downhill flow of lava occurs because of the action of gravity. The shear stress at top of the lava is zero and increases with depth in the lava in proportion to the lava thickness. If the shear stress at the base of the lava is lower than the yield stress, no flow will occur and the lava will accumulate over the vent. Thus lavas with high yield stress will form domes or thick flows, whereas lavas with low yield stress will be thinner and longer.

The extrusion of domes typically follows the explosive eruption of silicic magma. During the preceding explosive eruption the volatile-rich upper part of the magma chamber has been tapped, leaving volatile-depleted magma, which rises slowly and nonviolently through the conduit and out of the vent. Extrusion of a dome often continues for several years and may result in a dome structure that is 1 to 2 km in diameter and several hundred meters in height. Gradual expansion and growth of the dome leads to oversteepening of the flanks, which in turn results in avalanching of dome material down the volcano's flanks. In rare cases the magma extruded is so viscous that it emerges at the surface as a plug above the vent. Thus, the plug of Mount Pelee in Martinique grew into a spine that extended 300 m above the vent.

## C. Pyroclastic Deposits

The ascent and explosive eruption of volatile-rich magma leads to disruption and ejection of magma fragments, which are quickly chilled or quenched to glassy material in the atmosphere. The fragmentary material settles out under gravity and forms a pyroclastic deposit on the ground. The formation of pyroclastic material is primarily dependent on the volatile content of the erupting magma, with exsolution of volatiles leading to bubble formation, expansion, and disruption of the magma froth. The formation and bursting of bubbles is, however, related to a number of factors, such as magma viscosity, surface tension, diffusion rate of volatiles, and temperature. Consequently, the resulting pyroclastic fragments show a wide range in vesicularity or porosity and morphology, depending on the amount of volatiles originally in the magma and the degree of expansion of the gas phase before quenching. Glassy and vesicular fragments are dominant component in pyroclastic deposits. Other components that may be important are crystals from the magma and lithics or xenoliths from wall rocks of the conduit or magma chamber. The collective term for all volcanic fragmentary material is tephra. Pyroclastic fragmentary material is classified according to size, as shown in Table I. Pyroclastic fragments that are highly vesicular have density $<1.0$ g/cm$^3$ and are referred to as pumice. They are typically of rhyolitic or dacitic composition. Vesicular pyroclastic fragments of basaltic composition are generally more dense and are termed scoria.

**TABLE I Classification of Pyroclastic Fragments**

| Pyroclast type | Diameter (mm) | Consolidated deposit |
|---|---|---|
| Block or bomb | $>64$ mm | Agglomerate |
| Lapilli | 2–64 mm | Lapillistone |
| Coarse ash | $\frac{1}{16}$–2 mm | Coarse tuff |
| Fine ash | $<\frac{1}{16}$ mm | Fine tuff |

Several types of pyroclastic deposits are recognized that reflect different modes of origin, transport, and deposition. The three most important types are fallout deposits, pyroclastic flow deposits, and pyroclastic surge deposits. Fallout deposits result from the settling of tephra particles through the atmosphere from a high eruption column or plume. They are the most widespread of all volcanic deposits and can extend several thousand kilometers from the source volcano. The distribution of fallout deposits is generally characterized by a circular or fan-shaped pattern and a relatively thin layer around the vent (Fig. 13). The fallout pattern becomes increasingly more ellipsoidal with increasing eruption size, and the fallout pattern from very large eruptions is an elongate ribbon that is distributed in an east–west direction in the middle latitudes. In small eruptions, the fallout pattern reflects the direction of low-level winds prevailing during the eruption, whereas in larger eruptions, the pattern reflects the influence of prevailing winds in the stratosphere, where most of the eruption plume is dispersed. The thickness of a tephra fallout layer can be up to 10 m on the flanks of the source volcano, but the layer thins exponentially as a function of distance from source along the fallout axis. The fallout from individual eruptions can be traced as 0.1- to 10-cm-thick layers in soils and sediments over land and on the ocean floors over great distances. Thus, the tephra fallout from the Toba eruption

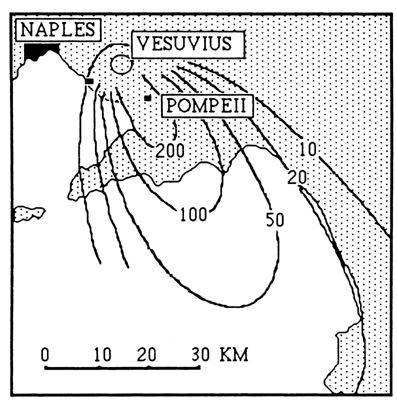


**FIGURE 13** Distribution of the pyroclastic fall deposit from the Plinian explosive eruption of Vesuvius volcano in Italy in AD 79. Contours are lines of equal thickness of the fallout deposit in centimeters.

on Sumatra is present near India, over 2500 km from its source.

The volume of individual tephra fallout deposits, which can be estimated from isopach maps, is typically in the range 0.1 to 10 $km^3$ of tephra, but up to 1000 $km^3$ in the case of the great Toba eruption 75,000 years ago. Tephra volume estimates are typically five times higher than the corresponding volume of erupted magma because of the large volume of void space in the tephra. They can be converted to corresponding magma volume if the density of the tephra deposit (e.g., 0.6 $g/cm^3$) and density of magma (e.g., 2.6 $g/cm^3$) are known. Thus the Toba tephra fallout deposit corresponds to about 230 $km^3$ of magma.

Tephra fallout deposits form uniform layers that mantle the topography and the layers possess certain diagnostic internal structures. The layers are characterized by high sorting (i.e., a small range in grain size of constituent fragments) due to the size sorting in the eruption plume during transport. Thus, all the largest particles will fall out together near the vent, whereas the finest ash will be transported far downrange before settling out of the atmosphere. The high degree of sorting in tephra fall deposits aids in distinguishing them from the poorly sorted pyroclastic flow and surge deposits. Many tephra fallout deposits are graded, that is, show a systematic change in grain size from the top to the bottom of the layer. When the grain size decreases upward in the layer, the grading is termed normal and is usually attributed to decreasing intensity of the eruption with time, resulting in decreasing height of the eruption column and consequently deposition of progressively smaller tephra particles at a given location in the fallout pattern. Tephra layers that show increasing grain size with height are referred to as reversely graded, reflecting progressive increase in eruption energy and column height, resulting in deposition of coarser particles with time during the eruption.

Very fine tephra that is transported to great height during explosive eruptions can remain in the stratosphere for days or weeks before settling out to earth. The long stratospheric residence time is due to the low terminal velocity of fine volcanic ash particles, which is, for example, less than 0.1 m/sec for a 10-$\mu$m particle. Such particles may be supported for expanded periods in the atmosphere by minor wind turbulence and produce a variety of optical effects as well as interact with the earth's thermal radiation budget, as discussed below.

Pyroclastic flows form another important type of pyroclastic deposits, which are produced by hot density currents of pyroclastic fragments, volcanic gas, and admixed air. As discussed in Section V, pyroclastic flows are most commonly produced by eruption column collapse. During several explosive eruptions, witnesses have observed a glowing avalanche, commonly referred to as a nuee ardente, which cascades from the crater or from the eruption column above the crater and down the flanks of the volcano. Such glowing avalanches were observed, for example, during the devastating eruptions of Mont Pelee on Martinique and La Soufriere on St. Vincent in the West Indies in 1902, when large loss of life was caused by hot, glowing pyroclastic material that avalanched down the volcanoes. The glowing avalanches consist of two components: a pyroclastic flow at ground level, with a high concentration of pyroclastic material transported under force of gravity by laminar flow, and an upper highly turbulent and billowing ash cloud that obscures the pyroclastic flow cannot therefore be directly observed during their emplacement, and their characteristics must be based on the study of the deposits. Since they are also derived from explosive eruption columns, pyroclastic flows consist of the same particle constituents as tephra fall deposits [i.e., glassy fragments of quenched magma (pumice, lapilli, ash), crystals from the magma, and lithics or xenoliths]. Most pyroclastic flows result from the eruption of silicic magmas, whereas basaltic examples are unknown.

Pyroclastic flow deposits, also commonly called ignimbrites, have several distinguishing features that reflect their mode of transport and the topographic control of their distribution. Because of their origin as density currents, pyroclastic flows typically follow valleys and other depressions and form long, narrow deposits on sloping ground, such as on the flanks of volcanoes. They spread out on flat-lying terrain to form a wide plain, with a horizontal upper surface of the pyroclastic flow deposit. Because of this topographic control on their mode of transport, these deposits are easily differentiated from tephra fallout deposits, which mantle the topography. Many explosive eruptions produce tephra fallout only, without generation of pyroclastic flows. The volume of pyroclastic flows produced in recent explosive eruptions is typically in the range 0.1 to 10 $km^3$. In the geologic record there are, however, pyroclastic flow deposits of enormous proportions or ranging from 100 to 3000 $km^3$, such as in Long Valley (California), Yellowstone (Wyoming), and the San Juan Mountains (Colorado). Such large-volume pyroclastic flows are in general related to caldera collapes, that is, the subsidence of the central part of the volcano along concentric fractures due to the partial emptying of the magma chamber beneath. In general, the size of the resulting caldera is proportional to the volume of erupted pyroclastic flow material, but caldera diameters may range from 2 to 30 km.

Pyroclastic flow deposits are typically 1 to 10 m thick and massive. Their internal structure is characterized by very poor sorting, with pyroclastic fragments, such as pumice and lithics of great range in size, supported by a fine-grained matrix. Size grading is usually poor or absent, but a concentration of larger pumices often occurs in the uppermost part of each flow. Carbonized tree logs are sometimes present and are also concentrated in upper part of the flow. Vertical pipes, filled with pumice or lapilli clasts, are present in many pyroclastic flows, but they result from the escape of gases such as water vapor from the flow during cooling. The temperature of pyroclastic flows during emplacement can range from 300° to 800°C. In some flows the temperature is high enough to render the vesicular pumice or lapilli fragments plastic, resulting in their deformation, collapse of vesicles, and compaction and welding of the deposit. In welded pyroclastic flows the larger pumices are, therefore, highly flattened and elongate in the direction of flow, but such deformed pumices are referred to as fiamme. Upon cooling, welded flows develop a columnar structure, which is identical to the internal structure of solidified lava flows.

Pyroclastic surge deposits represent the third main type of pyroclastic deposits. They are volumetrically less important than either fallout or pyroclastic flow deposits, but surges are particularly important because of their destructive potential. The density current that is generated during some explosive eruptions is a rather dilute and highly turbulent surge cloud consisting of pyroclastic fragments carried in a mixture of volcanic gases and admixed air. Such density currents differ principally from pyroclastic flows in that the latter have a higher concentration of pyroclastic material and display laminar flow. Surges are broadly divided into two types reflecting their origin: wet and relatively cold base surges generated during phreatomagmatic explosions and dry and hot pyroclastic surges generated from glowing avalanches or nuees ardentes. Surge clouds are highly mobile and travel at higher velocity than pyroclastic flows. It is estimated that surges may advance at rates of 100 to 200 m/sec, and although they may both be generated from the same nuee ardente, the coexisting surge is likely to outdistance the accompanying pyroclastic flow. There are certain features of the resulting deposits that can be used as diagnostic aids in determining the origin of surges, as discussed below. Surge deposits are generally distributed in a circular pattern around the volcano and commonly extend 2 to 10 km from the source. Individual surge deposits are thin compared to pyroclastic flows and decrease in thickness with distance from the source. Owing to the highly turbulent conditions within an active surge cloud, the depositional process is very variable and the resulting deposit consequently shows a complex internal structure. The most common structures in surge deposits are sandwaves. These are bedforms with undulating surfaces or surfaces inclined at an angle to the base of the deposit that result in cross-bedding. Such bedforms may result in formation of dune structures on the surface of the surge deposit. Sandwave dunes of this type show a systematic decrease in both amplitude and wavelength with distance from source, reflecting the decreasing velocity of the surge cloud. Typical thicknesses are 2 to 5 m near the source and 5 to 10 cm in the distal area. Accretionary lapilli, which are pellets composed of concentric layers of fine ash, are often present in base surge deposits. These pellets are an indication of condensed water vapor in the surge cloud, since accretionary lapilli require the presence of water for the adhesion of ash particles into a concentric structure. They are thus a diagnostic feature of relatively cold surge clouds. Hot pyroclastic surges, on the other hand, do not contain accretionary lapilli and their high temperature is furthermore attested to by the presence of carbonized wood.

Lahar is a flow of pyroclastic material admixed with water. Either lahars are formed directly from pyroclastic flows that encounter water during flow down the flanks of a volcano, or they may be formed as a result of very heavy rainfall on the volcano during or following an eruption and resulting mobilization of unconsolidated and unstable pyroclastic material on the steep flanks of the volcano. Also known as mudflows, lahars can range from viscous flows that carry huge boulders to dilute floods that grade into torrid and muddy streams. Lahars present a major volcanic hazard, as demonstrated during the 1985 eruption of Nevado del Ruiz volcano in Colombia, where 23,000 people lost their lives in lahars.

During subglacial volcanic eruptions, enormous floods result when the meltwater bursts forth from underneath the ice sheet, carrying large quantities of volcanic debris. Lahars or floods of this type are common in Iceland and are known as jokulhlaups or glacier bursts. Lahars have extended up to 300 km from source and are known to travel at speeds of 50 to 100 km/hr. Like pyroclastic flows, they are a form of relatively dense and coherent laminar gravity current and, therefore, flow along valleys and into topographic depressions. The resulting lahar deposits are massive, very poorly sorted layers containing a large assortment of particles. Volcanoes with crater lakes are particularly prone to the production of lahars. Similarly, snow- or ice-covered volcanoes typically generate lahars in their early stages of eruption.

## VII. TYPES OF VOLCANIC ERUPTIONS

Several types of volcanic eruptions are recognized and classified according to a variety of features such as the degree of explosive activity, the pattern of explosions, and the role of external water versus magmatic volatiles. Like all such schemes, this classification is only for convenience and it must be kept in mind that a complete spectrum exists

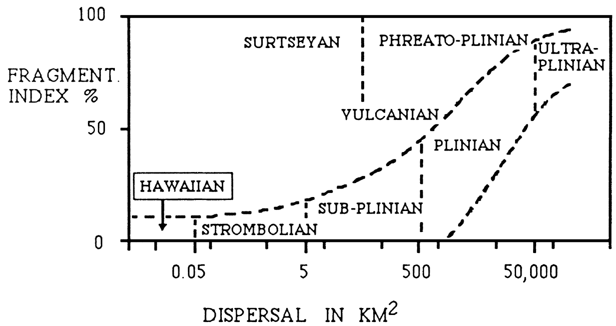


**FIGURE 14** Classification scheme of volcanic eruptions on basis of the dispersal and fragmentation of their pyroclastic fall deposits. Fragmentation index is the weight percentage of the deposits that is finer than 1 mm along the dispersal axis, where it is crossed by the isopach line that is 10% of the maximum thickness of the deposit. Dispersal is the area enclosed by the 1% isopach line. Adapted with permission from Walker, G. P. L. (1973). *Geol. Rundshau.* **62,** 431.

between all the types described here. In one classification scheme (Fig. 14) the eruption type is determined from the characteristics of the pyroclastic fall deposits, that is, the relationship between degree of fragmentation ($F\%$ is the weight percentage of the pyroclastic deposit finer than 1 mm along dispersal axis where it is crossed by the isopach line that is 10% of the maximum deposit thickness) and dispersal ($D$ is the area enclosed by the isopach line which is 1% of the maximum deposit thickness).

### A. Plinian Eruptions

In Plinian eruptions the exsolution of magmatic volatiles in the volcano's conduit leads to disruption and explosive ejection of pyroclastic material and the formation of an eruption column, which is sustained for hours or days above the volcano. In eruptions of this type, the magma is so viscous that exsolved volatiles cannot rise through the liquid magma at an appreciable rate relative to the rate of rise of the magma through the conduit. The volatiles thus escape only by explosive disruption of the magma. In Plinian eruptions the fragmentation of magma is very high and the surface area of fragments exposed to the atmosphere is, therefore, also high. Consequently, eruptions of this type are characterized by a higher efficiency in utilization of thermal energy than other volcanic eruptions. The Plinian eruption produces only pyroclastic material, which exits from the vent at several hundred meters per second and forms a convective column tens of kilometers in height. Plinian eruptions are thus not instantaneous explosions, but rather sustained events involving ejection of fragmented magma at rates of $10^6$ to $10^9$ kg/sec. This type of eruption is named after Pliny the Younger, who witnessed the eruption of Vesuvius volcano in Italy in AD 79 and wrote the first detailed account of a volcanic eruption. This event began about noon on August 24th with the formation of an eruption column that rose to 27 to 33 km above Vesuvius and led to intense fallout of pumice and volcanic ash over districts to the south and east, including the city of Pompeii. These districts were gradually buried under up to 2 m of pumice. Roofs collapsed in Pompeii and other towns, and widespread evacuation was necessary. The extent of ashfall was widespread, including north Africa. After 12 hr of fallout, the style of the eruption changed suddenly about 1 A.M. on 25 August 25th, as the eruption column collapsed, with generation of the first pyroclastic surge. During the next 8 hr the eruption column oscillated six times between a high eruption column producing pumice fallout and collapse with surge and pyroclastic flow generation. The surges swept down the flanks of the volcano in all directions and devastated areas to the north and west, which had not been affected by the earlier fallout, such as the city of Herculaneum, where hundreds of people were asphyxiated by the first surge. The surges were of progressively greater extent and the fourth surge inundated Pompeii and killed an estimated 2000 people who were in the process of evacuating from the city.

The behavior of the eruption column of Vesuvius in AD 79 is typical of many Plinian eruptions, which generally begin with a high and sustained eruption column that produces tephra fallout, followed by one or more column collapse events that generate hot pyroclastic surges and pyroclastic flows. As discussed in Section V, the collapse and generation of surges and pyroclastic flows can be related to several factors, but the most important ones are probably decreasing gas content in the magma and increasing mass eruption rate.

### B. Strombolian Eruptions

In Strombolian eruptions discrete explosions occur in a magma column near the surface at intervals of seconds or minutes. They are characteristic of basaltic and andesitic magmas with moderate volatile content and relatively low viscosity. In these magmas the magma viscosity is so low that the exsolved volatiles form coalescing and large gas bubbles, which can rise relatively rapidly up the conduit at a rate much higher than the ascent rate of the magma. Each bubble burst at the surface blasts off a spray of magma, which is fragmented and quenched in the atmosphere above the vent to form pyroclasts. The ejecta are generally incandescent as they leave the vent, but form scoria, lapilli, and ash. Fragmentation is much less than in Plinian eruptions. Consequently, the transfer of heat from pyroclasts to the atmosphere is highly inefficient, and convective rise of the eruption column is weak and plays a very small role in tephra dispersal. The conversion of thermal to kinetic energy in Strombolian eruptions is, therefore, much weaker than in Plinian eruptions. Because of the low exit velocities of the erupting mixture, the pyroclasts have a restricted range and are deposited around the vent to form a scoria cone or are dispersed in a circular or weakly ellipsoidal fallout pattern with a radius of 1 to 5 km around the vent. The principal products of Strombolian eruptions are thus moderately vesicular scoria and lapilli, and the dominant landform resulting from these eruptions are scoria cones. These eruptions eject solely magmatic pyroclasts, and lithics or xenoliths of wall rocks are rare or absent.

This type of volcanic eruption is named after the volcanic island of Stromboli in the Mediterranean Sea, west of Italy, which has been almost continuously active for 2500 years. Strombolian eruptions are among the most common type of volcanic activity, but are typically relatively small volume events. In these eruptions the initial Strombolian activity is followed by a flow of lava. This pattern is due to a higher gas content in the early erupted magma, whereas magma erupted later is gas poor and, therefore, extruded nonexplosively.

### C. Vulcanian Eruptions

Vulcanian eruptions are characterized by discrete explosions in the volcano's conduit at intervals of minutes or hours with exit velocities of the order 400 m/sec. The explosions eject a mixture of magmatic pyroclasts and lithics derived from explosive erosion of the conduit walls. The lithic component may be half of the total ejecta. Fragmentation of pyroclasts is very high, with production of lapilli and ash dispersed over a wide area. Because of the repeated and discrete explosions, the resulting fallout deposits are multilayered. The eruptions are often accompanied by generation of surges and minor pyroclastic flows. Following a period of Vulcanian explosive activity over a number of days or months, the volcano may extrude a dome or a viscous lava flow as the terminal stage in an eruption sequence. The volume of the lava extrusion is often greater than the total volume of erupted pyroclasts. Vulcanian explosions occur in virtually all types of magmas, but are most common in andesitic volcanoes, where magma viscosity is moderate to high. These events are preceded by ascent of magma into the conduit of a volcano, but the magma is not extruded because of the presence of a plug. The plug may be preexisting rock formations or debris from previous explosions or from cooling and partly solidified and very viscous magma. Pressure buildup under the plug may be due to either exsolution of magmatic volatiles or vaporization of groundwater that flows into the conduit or percolates down through the plug. Over a finite time interval, the buildup of vapor pressure below the plug exceeds the confining pressure, leading to rupture of the plug and explosive clearing of the conduit and the vent, forming a discrete Vulcanian explosion. This type of eruption is named after the 1888 eruption on the island of Vulcano near Sicily in the Mediterranean Sea.

As stated above, the buildup of pressure that precedes the Vulcanian explosion may be due to either exsolution of magmatic volatiles or the vaporization of external water that comes in contact with magma in the conduit of the volcano. There is an entire spectrum between these two extremes and it is generally impossible to establish which of these agents is more important in a specific event. Thus, many Vulcanian explosions are in part phreatomagmatic events; that is, they result from the explosive interaction of magma and external water, such as groundwater from an aquifer traversed by the volcano's conduit.

### D. Hawaiian Eruptions

Eruptions of the Hawaiian type are characteristic of highly fluid basaltic lavas of very low gas content. They give rise to thin and highly fluid lava flows, which gradually build up a large and broad shield volcano. Because of the low viscosity and gas content, the degree of magma fragmentation is minor and production of pyroclasts is restricted to a minor scoria cone around the vent, principally due to fire-fountaining activity. This type of activity has built up the Hawaiian shield volcanoes of Mauna Kea, Mauna Loa, and Kilauea as well as many other shields in Iceland and elsewhere where basaltic volcanism is dominant. The shields are built around a central crater, which is fed by a conduit from a deep magma reservoir. The crater often evolves into a larger summit caldera, which may possess a lava lake during periods of vigorous activity. During Hawaiian eruptions the magma exits the vent as a fountain, which may reach 500 m in height. The fountain is driven by hydrostatic pressure in the magma reservoir and in part by the exsolution and expansion of gases from the magma in the conduit. The vast proportion of the fountain falls to the ground and contributes to a low-viscosity lava flow, but a minor fraction of the fountain is dispersed with the prevailing wind, which draws out the highly fluid magma droplets into delicate needlelike forms called Pele's hairs and Pele's tears, named after the Hawaiian goddess of fire.

Hawaiian eruptions are generally frequent but of small volume. Thus, the typical steady-state, long-term magma output rate of Hawaiian volcanoes is 1 $m^3$/sec, but can be as high as 4 $m^3$/sec as in the case of Kilauea for the vigorous period from 1969 to 1973.

### E. Fissure Eruptions

The most important type of volcanic eruption on the earth's surface is the fissure eruption, when the crust of the planet is rifted along a long and narrow fissure from which lava erupts along its length. This type of activity characterizes volcanism along the 60,000-km length of the midoceanic ridges of the earth. Because the ridge system is dominantly submarine, however, the activity is rarely witnessed except in the rare subaerial parts of the ridge system, such as Iceland and the Afar region of Ethiopia. In the midoceanic ridge environment the fissure eruptions are due to tensional stresses in the crust induced by sea-floor spreading. The process leads on the average to 5-cm extension of the ridges annually, but it is not a continuous process in detail. Locally, the ridge extension and resulting volcanism occur in spurts of rifting episodes, with quiescent intervals of several years. Fissure eruptions also occur on Hawaiian volcanoes, but here the tectonic mechanism is quite different from the midoceanic ridge rifting process. On the island of Hawaii, for example, the gravitational sliding of the large volcanic edifice, which is the highest mountain structure on earth measured from the ocean floor, causes rift zones as the flanks of the volcanoes slump oceanward and adjust to the gravity field. Although the tectonic mechanisms are different, the resulting fissure eruptions are of the same type.

Fissure eruptions are restricted to highly fluid and low-viscosity basaltic magmas, which produce a widespread lava flow, with very minor production of pyroclastic material. They are quite similar to Hawaiian eruptions except for the configuration of the vent, which is a circular one or a short fissure in Hawaiian eruptions, as opposed to a fissure with length of tens of kilometers. During fissure eruptions, the initial activity extends along the total length of the fissure as a continuous curtain of fire fountains. With decreasing mass eruption rate, the active fissure shortens and gradually the activity becomes concentrated in several vents along the former fissure. The final stages of activity involve weak lava emission, mixed with weak Strombolian activity in a few vents, leading to the construction of a linear array of scoria cones along the former fissure. Eruption rates for each meter length of the fissure are of the order $10^3$ to $10^4$ kg/sec m. The largest fissure eruption in historical time was the great Laki eruption in southern Iceland in 1783. During a period of 8 months, the 25-km-long fissure pured out a total of 12.3 km$^3$ of basaltic lava, which covered an area of 565 km$^2$ (Fig. 15). The volume of magma is truly large, as it is about four times the annual production rate of the entire midoceanic ridge system. Such eruptions are not uncommon in Iceland, however, where they relate to the presence of large, shallow magma chambers under the central volcanoes and the lateral flow of magma out of these magma chambers during crustal rifting episodes. Icelandic central volcanoes have magma chambers at depth of 2 to 5 km below the surface, with volumes of tens or hundreds of cubic kilometers. Activity

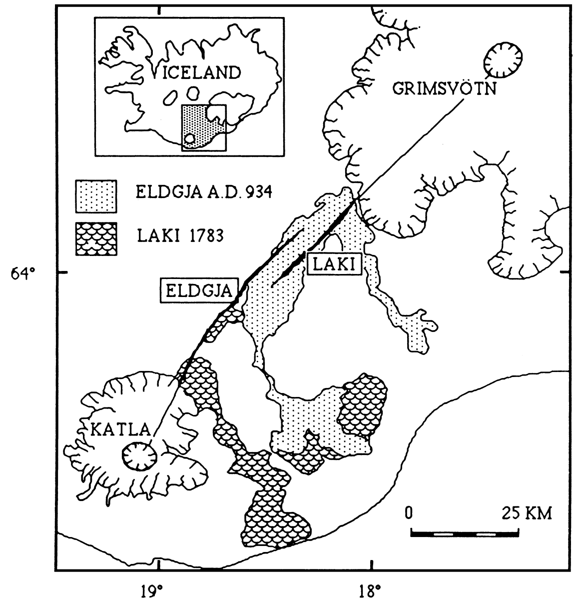


**FIGURE 15** Historic fissure eruptions of basaltic magma in Iceland. The Laki eruption in 1783 occurred along a 25-km-long fissure in the southern lowlands, producing about 12.6 km$^3$ of magma. It is probable that the source of this magma was the magma chamber of Grimsvötn central volcano under the ice sheet to the northeast. Similarly, the adjacent Eldgja fissure eruption of AD 934 derived its 9 km$^3$ of magma from the chamber of Katla central volcano to the southwest. Both examples demonstrate the importance of lateral flow of magma within fissures of the crust.

in these volcanoes is highly episodic and associated with periods of renewed rifting of the fissure swarm, which typically traverses the volcano, and may occur at intervals of 100 to 200 years. During periods of no rifting the magma reservoir is recharged and only minor eruptions may occur within the central volcano. When rifting is renewed due to the episodic extension of the plate boundary, the fissure swarm of the central volcano is reactivated, and the high hydrostatic pressures of magma in the chamber result in lateral flow of magma from the chamber into the dilating fissure swarm. In the case of the Laki fissure eruption and some others, there is evidence that magma from a chamber under the central volcano flowed laterally 50 to 70 km within vertical fissures in the crust. The magma then emerged in a fissure eruption in low-lying parts of the fissure swarm. Fissure eruptions of Hawaiian volcanoes also occur in fissure swarms on the lower flanks of the volcanoes and are fed laterally from reservoirs under the summit region of the volcano. Thus, fissure eruptions illustrate in general that rifted volcanic edifices can have rather complex plumbing systems, with magma chambers feeding both vertical conduits to the summit region (Hawaiian eruption) and lateral flow to the volcano's flanks through fissures. Consolidation of magma in fissures at the close of an eruption leads to the formation of intrusive rock bodies known as dikes.

## VIII. VOLCANIC LANDFORMS

The eruption of magma onto the earth's surface produces a variety of landforms whose characteristics depend primarily on the physical and chemical properties of the magma, such as its viscosity and volatile content. Some volcanic landforms are determined by structural or tectonic features of the region, which influence whether the eruption occurs through a fissure or a circular vent, while other landforms are peculiar to eruptions under water or glaciers.

### A. Stratovolcanoes

Volcanoes built up of alternating eruptions of pyroclastic material and lavas are termed stratovolcanoes. They are one of the most common types of volcanoes, especially in the volcanic arcs above subduction zones. They consist of interbedded layers of tephra, both fallout and pyroclastic flow deposits, and lava flows. Stratovolcanoes are dominantly andesitic or dacitic in composition, and their elevated morphology is a direct consequence of the high viscosity of these magmas. Most of the eruptions originate in a central circular vent, and the accumulation or products around the vent results in the construction of a symmetrical and cone-shaped mountain. The high degree of regularity of this process often results in the construction of a profile of great beauty, such as the famed Fujiyama in Japan, with gradual steepening of the slopes from about 10° on the lower slopes to up to 35° toward the crater at the summit (Fig. 16a). Stratovolcanoes range from a few hundred meters to 400 m in height and may have a basal diameter of 30 km. Fujiyama has a height of 3700 m above the base of the volcano, a diameter of 30 km, and a volume of 870 km$^3$. Plinian and Vulcanian eruptions are typical of the activity of stratovolcanoes, which produce widespread tephra fallout and pyroclastic flows and surges that form deposits on the lower slopes and the surrounding plain. These eruptions may terminate with effusion of lavas or dome extrusion.

During their later stages of activity, stratovolcanoes often loose their graceful symmetrical form as vents become active on their flanks that accompany or replace the activity of the central crater. These subsidiary vents may be aligned in radial rift zones or randomly scattered on the flanks. They result from the increasing complexity of the geometry of the conduit feeding system as well as the progressive increase in height of the volcano.

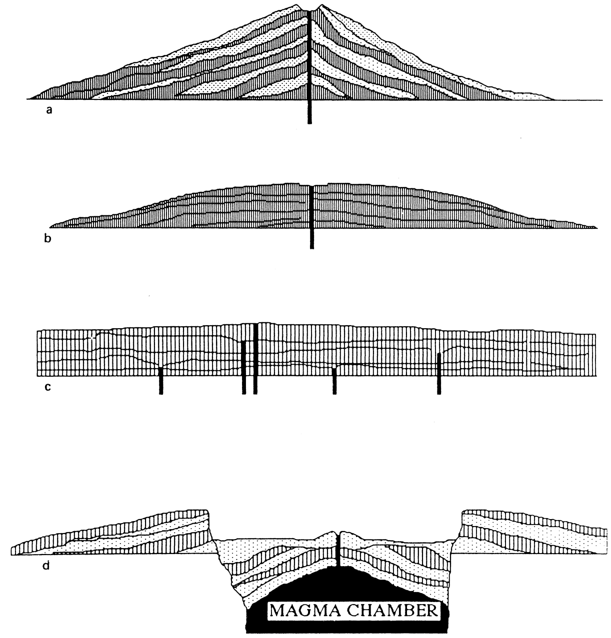


**FIGURE 16** The principal types of landforms produced by volcanic activity: (a) stratovolcano, (b) shield volcano, (c) lava plateau, and (d) caldera. Volcanoes with alternating styles of explosive and effusive eruptions from a central vent produce a stratovolcano, built up of alternating layers of lava flows and pyroclastic deposits. Stratovolcanoes are typical of volcanic arcs, such as Fujiyama in Japan. Volcanoes dominated by the effusion of lavas from a central vent result in the construction of a shield volcano, with a low profile. They are typical of basaltic hot-spot volcanism, such as Mauna Loa in the Hawaiian Islands. In regions of crustal extension and rifting, fissure eruptions produce extensive lava flows, which result in the buildup of lava plateaus. Examples of such activity include Iceland and the Columbia River basalts in Washington State. During very large eruptions, the withdrawal of large volumes of magma from the magma chamber leads to weakening of the volcano's edifice and its collapse, with formation of a caldera. An example of this volcanic landform is Crater Lake in Oregon.

### B. Shield Volcanoes

The effusion of volatile-poor and low-viscosity basaltic magmas from a circular vent results in the formation of a shield volcano consisting of a broad and relatively flat cone made almost entirely of lava flows. These gentle and symmetrical mountains have slopes generally between 2° and 10°, but their profile is typically that of a very flat sigmoidal curve, with the slope decreasing gradually with distance from the central crater (Fig. 16b). Shield volcanoes were originally named in Iceland, where many landforms of this type are typically 5 to 10 km across at the base and about 1000 m in height. Each of the Icelandic shield volcanoes is believed to be built during a single prolonged eruption of highly fluid basaltic magma.

The giant shield volcanoes of Hawaii are similar landforms but of quite different scale. These mountains rise 8 to 9 km above their ocean-floor base and form the tallest volcanic landforms on earth. Their eruptions consist almost exclusively of volatile-poor, low-viscosity basalt magmas, which form highly fluid and mobile lavas that flow far from source. A magma reservoir 2 to 6 km below the summit supplies a central crater via a conduit system. Because of their size, the Hawaiian shields are much more complex than the Icelandic shields and typically possess two or three rift zones, which radiate from the central crater region. Their volcanic activity is thus related to eruptions from both the central crater and the fissure swarms of the radial rifts.

## C. Lava Plateaus

In major rift zones, such as in Iceland, the Afar, and elsewhere on the axis of midoceanic ridges, low-viscosity basaltic magmas are typically erupted from fissure systems. Each fissure is normally active only once, and subsequent fissure eruptions take place parallel to, but at some distance from, the previously active fissure. Because of the high mobility of the basaltic magmas, the products of the fissure eruption form only minor scoria cones around vents on the fissure, and the majority of the products are distributed widely as thin lava flows. Fissure eruptions, therefore, produce only minor local landforms (Fig. 16c), but their major effect on the landscape is the formation of lava plateaus, which may cover areas of $10^4$ to $10^6$ km$^2$ and have thicknesses of 100 to 1000 m. In the subaerial environment, the major recent lava plateaus are in Iceland and the Afar region of Ethiopia. Older lava plateaus include the Deccan basalts in India, formed about 65 million years ago; the Columbia Plateau in the northwestern United States (mid-Tertiary in age); and the Parana basalts of South America. Lava plateaus result from the vast outpouring of basalt magmas onto the earth's surface over a relatively short period. In most cases this activity is associated with a hot spot or a local melting anomaly in the earth's mantle underneath the plateau.

## D. Calderas

The extraction of large volumes of magma from the magma chamber during major volcanic eruptions can lead to instability of the overlying volcanic edifice and its collapse into the chamber below. The resulting circular depression is known as a caldera (Fig. 16d). They are typically circular or ellipsoidal in plan, 2 to 30 km in diameter, and several hundred meters in depth. The walls of calderas are generally steep cliffs that expose the stratigraphy of earlier erupted products of the volcano as cut through by the caldera ring fault. Calderas are typically dynamic features, however, where intermittent subsidence of the caldera floor is balanced by accumulation of new volcanic deposits within the depression. Thus, a caldera may be refilled and buried during a vigorous phase of the volcano's activity.

The formation of a caldera usually occurs at a late stage in the lifetime of a volcano, when a large magma chamber has been established at a shallow depth. The partial draining of the chamber, which results in a weakening of the roof, may occur during major rifting of the volcano's flanks and resulting fissure eruptions in the rift zones. This type of tapping of magma from high-level chambers by lateral magma flow is the typical caldera-forming process in basaltic shield volcanoes, such as in Iceland and Hawaii. Very large explosive eruptions in stratovolcanoes also result in partial chamber evacuation, which may occur by eruption through a central vent or from a series of vents along the ring fault of the caldera.

## E. Seamounts and Subglacial Volcanoes

The most abundant volcanoes on the surface of the earth are seamounts within the ocean basins, which are approximately $10^5$ in number. Because of their submerged nature, they are, however, largely unstudied and remain the most neglected aspect of volcanology. Seamounts are scattered in the ocean basins and do not show a systemic relationship to midoceanic ridges. Several seamounts occur in elongate chains that are several thousand kilometers in length and relate to the relative motion of a hot spot under the earth's lithosphere. Seamounts are most common in the Pacific, where their density is about 100 seamounts per $10^6$ km$^2$ area of sea floor and their average size is 600 km$^3$. They have basal diameters of 5 to 20 km and are 1 to 4 km in height. Most seamounts are basaltic in composition.

The surrounding ocean has two important effects on the growth of seamounts. First, the cooling effect of the ocean results in quenching of the erupting magmas to pillow lavas, which accumulate around the vent and thus prevent the formation of extensive lava fields. Second, the high hydrostatic pressures prevailing at great water depths (up to 1 kbar) inhibit the exsolution of volcanic volatiles and thus restrict the fragmentation of magmas in deep-water eruptions. Magma fragmentation due to volatile expansion can only occur when the shallowing seamount has built up an edifice close to sea level. Because of these effects, seamounts are generally circular in plan and steep sided. During their main stage of growth, seamounts build up a pile of pillow lavas until the vent is within a few hundred meters of sea level. With decreasing water pressure in shallow water, the exsolution and expansion of volatiles from the magma become more important, leading to magma fragmentation and production of some pyroclastic deposits. In addition to the fragmentation induced by magmatic volatiles, explosive interaction of the ocean water and hot magma also occurs in water depths of less than 300 m. Such explosive activity is due to phreatomagmatic activity and explosive expansion of vaporized seawater in contact with magma and resulting fragmentation of the magma. Consequently, the uppermost deposits of seamounts are largely composed of pyroclastic layers interbedded with pillow lavas. These fragmentary deposits are also transported down the seamount's flanks and form a mantle over the lower pillow lavas and spread as volcanogenic sedimentary deposits into the adjacent ocean basin. When seamount volcanoes breach sea level, the magma is no longer in contact with ocean water and water–magma explosive interaction ceases. The magma consequently erupts as fluid lavas, which form a cap over the pyroclastic deposits above sea level.

Seamounts are rarely exposed for study, but their internal structure can be inferred from the features of subglacial volcanoes, which are formed by the same kind of processes (Fig. 17). Subglacial volcanoes occur in Iceland, Antarctica, and elsewhere where volcanic activity has taken place under a glacial cover 1 to 2 km thick. During initial activity, the extrusion of basaltic magma under the ice sheet results in the formation of a large melt

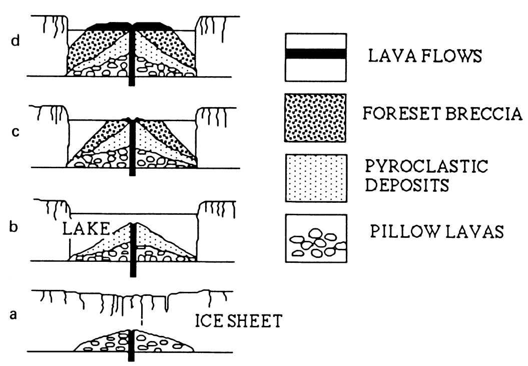


**FIGURE 17** Stages in the formation of a subglacial or submarine volcano. Initial activity takes place at the base of the ice sheet (a), with extrusion of pillow lavas. As the vent reaches shallower water (b), phreatomagmatic explosive activity begins owing to interaction of magma and water, and a mantle of pyroclastic material is deposited over the pillow lava base. When the vent reaches above the water (c), magma–water interaction ceases and lavas flow from the crater. The entrance of the lavas into water around the new island produces foreset breccias on the flanks. With further activity (d), the island grows by flow of subaerial lavas and subaqueous deposition of foreset breccias. [From Allen, C. C. *et al.* (1982). "Subglacial volcanism in north-central British Columbia and Iceland." *J. Geol.* **90,** 713. Adapted by permission of the University of Chicago Press. Copyright © University of Chicago Press.]

cavity in the glacier and buildup of a pillow lava pile in the water-filled cavity. As the volcano grows in height, the overlying water column shallows, the hydrostatic pressure decreases, and phreatomagmatic activity becomes possible, with production of pyroclastic deposits. Finally, when the volcano's crater emerges above the level of the glacial meltwater lake, lavas begin to flow. With the retreat of the Pleistocene ice sheet, the subglacial volcanoes in Iceland have become exposed for study and provide direct insights into the submarine and subglacial eruption processes.

## IX. VOLCANISM AND THE ATMOSPHERE

Throughout the history of earth, volcanic activity has been the main agent of transport of volatile components such as $CO_2$, $H_2O$, $SO_2$, and $H_2S$ from the interior of the planet to the earth's surface. The origin of the earth's atmosphere and hydrosphere is thus intimately linked with degassing of the earth through volcanic eruptions. In addition to this important role of volcanism in the evolution of the atmosphere and, thus, life on this planet, the gases released from volcanoes during large eruptions or volcanic episodes have also had important short-term climatic effects.

Primordial gases trapped during the earth's formation about 4.5 billion years ago are still being released today, and it is estimated that about half of the gas supply in the planet's interior has been released to the atmosphere and hydrosphere. There is evidence that early gases lost from the earth were characterized by low fugacity of oxygen, or $fO_2 \sim 10^{-12}$, which is significantly lower than that of modern volcanic gases ($\sim fO_2\ 10^{-8}$). The early reducing gases released during the first billion years of earth history may have been in equilibrium with metallic iron, whereas modern volcanic gases are primarily in equilibrium with silicate melts. The early gases consisted largely of $H_2$ and $H_2O$, with CO the dominant carbon species, $H_2S$ the dominant sulfur species, and $N_2$ the dominant nitrogen species. In modern volcanic gases, on the other hand, $H_2O$ is dominant and $H_2$ is very subordinate; $CO_2$ is the dominant carbon species and $SO_2$ is the main sulfur species. Because of the high injection rate of $H_2$ during the first billion years, the earth's early atmosphere may have been virtually devoid of free $O_2$. The atmosphere was then mildly reducing as opposed to the highly oxidizing atmosphere today. The geologic evidence for the anoxygenic early atmosphere includes the presence of banded iron formations with high ferrous iron content. The transition to oxygenic atmosphere came about as free oxygen was liberated by dissociation of oxides such as $CO_2$ and $H_2O$, either through photosynthesis in organic reactions or by photodissociation in inorganic reactions. Thus, the transformation of the early atmosphere, dominated by anoxygenic volcanic gases, to the oxygenic atmosphere is directly linked to the generation of lifeforms on the planet that are capable of carrying out photosynthesis reactions. The early photosynthetic organisms were probably prokaryotes analogous to modern blue-green algae.

The volatile input from volcanism to the atmosphere and oceans continues with modern volcanism. The total flux to the earth's surface of $CO_2$ from volcanism is estimated in the range $4 \times 10^{13}$ to $4 \times 10^{14}$ g $CO_2$ yr. Because of the huge size of the crustal, oceanic, and atmospheric reservoirs of water, carbon dioxide, and other components in volcanic gases, the volcanic input is proportionately minute compared to the size of these surface reservoirs. Thus, for example, the annual $CO_2$ input by volcanism is only about $1/10^9$ of the surface reservoir for carbon dioxide. Most of this volcanism is submarine and, therefore, does not influence the atmosphere directly. Certain volcanic events on earth have, however, had important short-term impacts on our atmosphere. These are large-volume subaerial volcanic eruptions that have injected huge quantities of volcanic gases into the atmosphere that resulted in short-term perturbations of the earth's climate. The impact of volcanic gases on the atmosphere is primarily due to sulfur compounds, although halogens may also have an important effect. Sulfur is primarily released as $SO_2$ from the eruption column, but undergoes a gas-to-particle conversion in the atmosphere upon hydration by atmospheric water and forms minute particles or an aerosol of sulfuric acid droplets. In nonexplosive volcanic eruptions, the sulfur emission is largely restricted to the troposphere and the $H_2SO_4$ aerosol droplets fall out with general precipitation as acid rain. The residence time of volcanic gases in the troposphere is, therefore, short. During explosive eruptions and very large fissure eruptions of basaltic magma, the eruption columns extend to much higher levels; this results in injection of volcanic aerosols into the stratosphere. In the absence of convective circulation and weather processes, the residence time of volcanic aerosols in the stratosphere is much longer, and they can have profound effects on the earth's radiation and heat budget, resulting in temporary cooling at the surface for a period of months or years following a major eruption.

The heat budget at the surface of the earth is dominated by transfer of radiation through the atmosphere. Volcanic aerosols can alter the balance between incoming heat with sunlight and outgoing infrared radiation. Thus, a stratospheric temperature increase can occur because of absorption of terrestrial thermal infrared radiation by volcanic aerosols in the stratosphere, and surface cooling can occur because of backscattering of incoming visible light radiation by the aerosol. Sulfuric acid aerosols in the stratosphere strongly backscatter incoming solar radiation and thus increase the optical depth. The importance of this process to the surface temperature of the planet depends primarily on the mass of sulfur injected into the stratosphere during an eruption. It is evident from historical eruptions that the sulfur yield to the stratosphere depends on both the mass of magma erupted and the composition of the magma (i.e., the concentration and type of volatiles the magma carries to the surface). In eruptions of basaltic magma, the yield of $H_2SO_4$ and HCl to the atmosphere is approximately $7 \times 10^{12}$ and $2 \times 10^{12}$ g, respectively, for each cubic kilometer of erupted magma. For explosive eruptions of silicic magma, the corresponding figures are approximately $3 \times 10^{11}$ and $4 \times 10^{11}$ g for each cubic kilometer of erupted magma. It is, therefore, evident that basaltic eruptions have a much greater potential for perturbing the earth's radiation balance compared to silicic eruptions and result in surface cooling. This potential is to some extent offset by the fact that basaltic eruptions are dominantly effusive eruptions of lavas and their volatiles are thus largely confined within the troposphere and quickly precipitated as acid rains. Some very large basaltic eruptions, such as the Laki fissure eruption in Iceland in 1783 (12.6 $km^3$ magma), have generated thermal plumes that have reached stratospheric levels and resulted in significant short-term climate modification. This eruption injected about $9 \times 10^{13}$ g of $H_2SO_4$ into the atmosphere over a 8-month period and led to crop failures in Iceland, which resulted in the loss of about 75% of the country's livestock and the eventual death of 24% of the inhabitants.

Most of the current basaltic volcanism on earth is submarine and the sulfur is, therefore, largely trapped within the quenched pillow lavas on the sea floor or dissolved in the surrounding ocean water. The impact of midoceanic ridge volcanism on the atmosphere today is therefore trivial. In the past history of the earth, however, there have been periods of greatly increased subaerial basaltic volcanism. Most notable in this regard is the episode of volcanism that produced the Deccan lava plateau and contemporaneous volcanism in Britain, Greenland, and elsewhere about 65 million years ago, or at the boundary of the Cretaceous and Tertiary periods. The vast outpouring of subaerial basaltic lavas at this time emitted exceptional quantities of volcanic volatiles, which may have played a role in the short-term global changes in the atmosphere and biosphere that characterize this boundary.

As noted above, the yield of volatiles to the atmosphere during explosive volcanic eruptions of silicic magmas is generally smaller per unit mass of erupted magma than in basaltic eruptions. These differences are largely due to the lower solubility of sulfur and halogens in silicic melts. Explosive eruptions are more important, however, because of their high eruption columns, which inject much of the volcanic volatiles directly into the stratosphere. Several historic explosive eruptions have thus had a notable impact on the earth's radiation budget. In 1963 Agung volcano on Bali in Indonesia erupted explosively and released a large quantity of sulfur, which formed an $H_2SO_4$ aerosol of $3 \times 10^{12}$ g. The eruption resulted in a stratospheric temperature increase of 4° to 5°C, which is attributed to absorption of solar and infrared radiation by the sulfuric aerosol. A concurrent surface cooling of 0.5°C was observed in the tropics because of absorption and backscatter of sunlight by the stratospheric aerosol. The climatic effects of the 1815 explosive eruption of Tambora volcano on the island of Sumbawa in Indonesia were much more severe. Approximately 100 $km^3$ of magma was erupted, and the volatile sulfur release resulted in about $5 \times 10^{13}$ g $H_2SO_4$ stratospheric aerosol and even larger emission of HCl and HF gases. The climatic effects of the Tambora eruption were severe, but poorly documented. Its eruption resulted in a devastating climatic deterioration, which led to the "year without summer" in 1816 in the northeastern United States. The periodicity of eruptions of the size of Tambora 1815 is probably about 300 years, whereas the periodicity of gigantic eruptions of the scale of the 1000 $km^3$ Toba eruption is probably of the order 10,000 years. Such events will recur in the future and severely influence the environment.

## X. GEOTHERMAL PROCESSES

Heat flows from the interior of the earth to the atmosphere at a rate of about 80 $mW/m^2$, corresponding to a heat loss from the whole earth of about $4 \times 10^{13}$ W. The vast proportion of this heat is transported through the upper mantle and the crust by convective and conductive processes, whereas only a small fraction, estimated as $8 \times 10^{11}$ W, is brought to the surface as "volcanic heat" during eruptions. The role of erupted magma is therefore small in the total heat flow budget of the earth, but the cooling of buried magma bodies within the crust plays an important role in heat flow and the formation of geothermal fields. Geothermal fields are regions on the earth's surface characterized by springs of hot water and vents emitting steam and gases. They are located in areas of young volcanism and are intimately associated with volcanic activity. Hydrothermal processes result in surface expressions that have a variety of forms. Hot springs are dominated by water, and the volume of steam and gas is minor. They are the most widespread form of geothermal activity, and three general types are recognized: hot springs, with large amounts of $CaCO_3$ in solution; alkaline springs, typically with high amounts of chlorides; and acid springs rich in sulfates. Fumaroles are geothermal springs that are dominated by steam emission. They have a high thermal emission, but the groundwater influx is relatively low and the available water is, therefore, converted to steam during subsurface boiling. Fumaroles can change seasonally to hot springs as the groundwater level rises or falls. Solfataras are geothermal areas that are dominated by emission of sulfur-rich gases and superheated steam. Geysers are a type of hot spring that jet out a column of boiling water up to several hundred meters in height at frequent intervals. In the geyser reservoir a column of water is heated to near its boiling point throughout the range of depth. Because deeper water is at higher pressure, it will be superheated when transported to shallower depth. If the weight on the water column is reduced, the hydrostatic pressure decreases and an accelerating boiling process propagates down the boiling column, resulting in vapor exsolution and jetting of the column out of the reservoir to the surface.

The surface manifestations of geothermal activity are due to hydrothermal circulation within the crust of the earth. Hydrothermal circulation is driven by the heat loss from magma chambers or other bodies of molten or hot rock that reside at high levels within the crust. Groundwater that comes in contact with hot rock bodies at depth is heated; this results in increased buoyancy and rise of the water toward the surface. The rising water is replaced by colder water from deepseated aquifers, and thus a circulation system is set in motion between the heat source at depth and the heat sink at the surface. Most hydrothermal circulation systems contain water or seawater as the dominant fluid, but there are two types of geothermal systems. In vapor-dominated systems the fluid is hot enough to be largely vapor, whereas in water-dominated systems the pressure-controlling fluid is liquid water. In addition, the fluid contains dissolved solids, water vapor, and gases dissolved in the liquid and free in the vapor. In continental or subaerial systems the dominant source of the hydrothermal fluid is meteoric water and the chemical composition of these waters is largely controlled by the solvent action of the hot water on the surrounding volcanic rocks. Certain chemical components of the water, such as $CO_2$ and $SO_4$, are, however, in part derived from the release of magmatic gases into the hydrothermal system. In submarine hydrothermal systems the dominant source of the hydrothermal fluid is seawater, but here again certain chemical components are also derived from magmatic sources.

The high heat output of geothermal fields requires an efficient heat extraction mechanism and deep downward penetration of water into hot rocks by a process of cooling and cracking. The cracking front in the hot rock around a magma body advances rapidly and the cooling of a large volume of hot rock in a short time provides an intense localized source of geothermal output. On the East Pacific

Rise, "black smokers," or active hydrothermal vents in the volcanic rift zone, jet out water at 380°C at water depth of 2500 to 2900 m. These high temperatures and the high content of sulfides and other metals carried in these jets require hydrothermal penetration of seawater very close to the boundaries of magma chambers under the midoceanic ridge at depths of 2 to 3 km below the surface. Because of the high rate of extraction of heat from the magmatic source, individual hydrothermal systems rarely have a lifetime greater than $10^4$ to $10^5$ years.

Hydrothermal circulation within the crust in volcanic regions of the earth is an important but largely untapped energy source. In some areas, such as Iceland, Italy, New Zealand, and California, these geothermal reservoirs have been tapped with 2- to 3-km-deep drill wells, which provide an inexpensive source of thermal energy for domestic or industrial use and can be converted to electric power with suitable steam turbines. Another important economic aspect of hydrothermal circulation is its role in the concentration of minerals and formation of economic ore deposits in the root zones of volcanoes. Thus, most of the world's supply of copper comes from porphyry copper deposits in the deeply eroded cores of old volcanoes, where hydrothermal circulation has formerly concentrated the ores. The circulating solutions also carry sulfur, lead, zinc, and molybdenum, which also form important ore deposits.

## XI. VOLCANIC HAZARDS

Volcanic eruptions are one of the earth's major natural hazards and have claimed 240,000 lives in the past 380 years. During this period, some fatalities have occurred in 5% of all eruptions, and one of six of the earth's active volcanoes has caused deaths. Excluding the activity of the submerged midoceanic ridge system, it is estimated that 1415 volcanoes have been active in the past 10,000 years, and 6073 volcanic eruptions have been documented from 8000 BC to the present time. Volcanic eruptions are, however, relatively benign when compared with the hazard of earthquake activity, which has caused about 20 times this number of deaths in the same period. Because of the increasing population density of our planet, volcanic hazards are of increasing concern and are likely to take a greater toll in the future.

One of the major problems with volcanic hazard assessment is the identification of active volcanoes. Traditionally, volcanoes that have erupted during the historical period are considered active and those for which there is no record of historic volcanic activity are considered dominant or extinct. This classification is of limited value since the historical period (i.e., the period for which written records exist) varies from a few hundred to 2000 years, depending on the region. Furthermore, many recent and devastating eruptions have occurred in volcanoes for which there is no previous record of activity, and some of these mountains were not even known to be volcanic in nature. If one considers only those 180 volcanoes for which the historic records span 500 years (Fig. 18), then it emerges that the median volcano has an eruption rate of 0.45 times per century; about 20% of these volcanoes erupt less often than once every 1000 years and about 2% less than once every 10,000 years. If we consider the entire Holocene record of volcanism, then the median interval between eruptions at each volcano is about 5 years (Fig. 19), but a significant number of eruptions were preceded by repose periods of 1000 to 10,000 years. Thus, it is certain that many volcanoes have an interval between eruptions that is much longer than the historic period. Prior to its devastating 1982 eruption, there was no historic record of activity of the El Chichon volcano in Mexico, but subsequent studies have shown that its previous eruption was 550 years ago. Similarly, prior to its 1951 eruption, which killed 2900 people, Mt. Lamington in Papua, New Guinea, was not known to be a volcano, and this was the first eruption after a dormancy of at least several thousand years. Of the 529 volcanoes that have erupted in the historic period, the most hazardous ones are located at convergent plate margins. Thus 88% of all eruptions with fatalities have occurred at convergent margins.

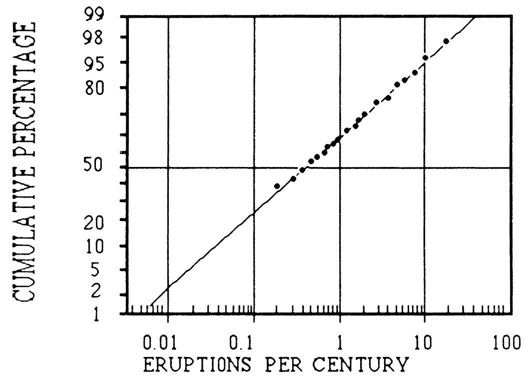


**FIGURE 18** Logarithmic probability plot of the cumulative proportion of the number of eruptions per century in 180 volcanoes that have historic records spanning the past 500 years. The median volcano in this data set has an eruption rate of 0.45 times per century. (After Walker, 1974.)

Although very common and widespread, lava flows present a relatively minor volcanic hazard to humans. Their flow front velocities are generally of the order 200 to 400 m/hr, and thus lava flows can be easily avoided by people in their path. The material damage due to lava flows is, however, significant, since all buildings, agriculture, and other developments that lie in their paths are totally destroyed. An additional hazard presented by very large

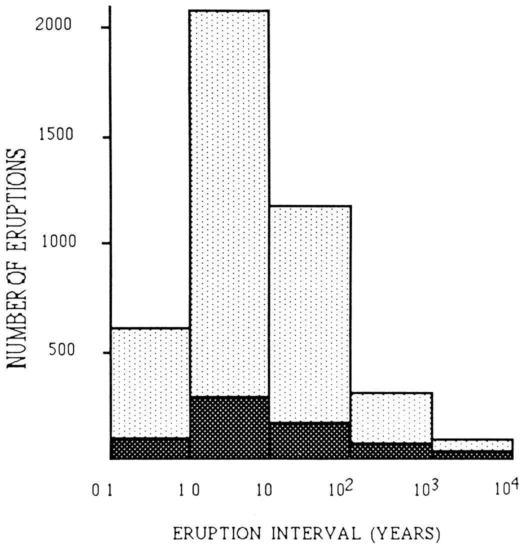


**FIGURE 19** The interval between eruptions in the same volcano for 4246 volcanic are eruptions (light shading) and 629 nonarc eruptions (dark shading). The median eruption interval for both groups is 50 years. [After Simkin, T., and Siebert, L. (1984). "Explosive Volcanism: Inception, Evolution, and Hazards," Adapted with permission of the National Academy Press, Washington, DC.]

lava flow eruptions is the emission of volcanic gases. The best example of this hazard is the Laki fissure eruption of basaltic magma in Iceland in 1783.

Tephra fall from explosive eruptions is a significant hazard for populations near the volcano due to roof collapse, danger from projectiles, and ignition of fires by incandescent pyroclastic fragments over 10 km from the vent. Extreme darkness during tephra falls also creates dangerous situations. The volcanic hazard from tephra falls is relatively minor and rather localized, whereas the beneficial aspects of tephra fall to agriculture over a wide region are considerable.

Pyroclastic flows and surges are by far the most hazardous of volcanic phenomena. These density currents of hot gases and pyroclasts move at velocities of tens to hundreds of meters per second and cover areas of hundreds to a thousand square kilometers. Because of their sudden appearance and high velocity, pyroclastic flows and surges are very difficult to escape. People in their path are generally asphyxiated in the hot, oxygen-poor, and dust-laden cloud or killed by pyroclastic projectiles or by skin burns. Pyroclastic flows generally follow valleys and depressions and their course is, therefore, rather predictable. Pyroclastic surges, on the other hand, have greater mobility and generally spread radially across the terrain around a volcano. Recent studies have shown that most fatalities in explosive eruptions are due to the pyroclastic surges. Thus 2000 people were killed in a surge from El Chichon volcano in Mexico in 1982, where nine villages were destroyed. The 1980 surge blast of Mount St. Helens volcano in Washington State also killed 67 persons and leveled about 500 km$^2$ of forest. Similarly, a surge cloud killed 2900 people around Mt. Lamington in Papua, New Guinea, in 1951. The largest known human disaster of this type is, however, the eruption of Mt. Pelee in Martinique in 1902, where a surge killed 28,000 people in the city of St. Pierre. Surges and pyroclastic flows rarely present a hazard beyond about 20-km radius from the volcano, but within that area total devastation of life and property can be expected.

Lahars or mudflows from volcanoes have been responsible for a significant number of deaths either during or following volcanic eruptions, approximately 10% of all volcano related fatalities. Thus about 5000 people were killed by lahars from the Kelut volcano in Indonesia in 1919, which destroyed 104 villages and covered 130 km$^2$. Kelut erupted again in 1966 and lahar generated from the volcano's crater lake killed an additional 210 people. The loss of 25,000 lives in lahars during the 1985 eruption of Nevado del Ruiz volcano in Colombia is one of the largest volcano-related disasters in history. Heat output from the small explosive eruption led to the melting of part of the extensive ice cap of the volcano, which resulted in the formation of four lahars, which inundated several towns around the volcano, with total destruction of the town of Armero.

The historic record shows that tsunami, or volcano-generated tidal flood waves, are the second-most-important direct cause of death in eruptions, after pyroclastic surges and pyroclastic flows. Tsunami are seismic sea waves that are generated by disturbances of the ocean floor. They may be related to the eruption of a very large mass of material into the ocean around a volcano or to the collapse of the volcanic edifice into the magma chamber. The displacement of water causes a large wave, which rises and crests in shallow water and breaks inland. The explosive 1883 Krakatau eruption in the Sunda Straits in Indonesia generated the largest known volcanic tsunami. The tsunami was 15 to 40 m in height at a distance up to 50 km from the volcano and flooded the adjacent coasts up to 5 km inland, with the resulting loss of 36,000 lives. Eruptions that generate tsunami of this magnitude fortunately are rare and are restricted to island volcanoes or volcanoes near the shoreline.

## XII. PREDICTION OF VOLCANIC ERUPTIONS

The current rapid growth in population density on earth is accompanied by increased volcanic risk, that is, the increased probability of loss of life and property due to volcanic eruptions. There is therefore a growing awareness of the need for volcanic hazard assessment and ultimately the prediction of volcanic eruptions. Volcanic hazard at a given volcano is a function of the probability of the occurrence and the intensity of a volcanic eruption. A general prediction can be made on the basis of a hazard assessment that takes into account the eruptive history of the volcano and thus requires detailed geologic studies of deposits from past eruptions and historic activity. Such general predictions can be considered as long-range forecasts, but they are of only limited value, since the fundamental assumption that past activity is the key to future activity may not always be valid. Volcanoes are dynamic, growing, and changing geophysical systems that do not necessarily follow a regular pattern of behavior. General predictions based on previous activity are thus primarily useful in forecasting the kinds, scales, and likelihoods of activity at a specific volcano but are of little value in predicting the timing of future activity. Furthermore, the database required to carry out a volcanic hazard assessment and a general prediction is unfortunately only available for a handful of volcanoes on earth. The greatest hazard in fact may stem from those volcanoes that are as yet unknown and have not erupted in the historical period.

Another approach to volcanic hazard assessment involves the specific prediction of volcanic eruption based on geophysical and geochemical volcano-monitoring techniques. This approach relies on the detection of precursory events and utilizes the rate of change in precursory phenomena as a basis for predicting an eruption. Eruption predictions that are based on volcano monitoring rely on the fact that eruptions are preceded by the ascent of magma within the crust, which in turn gives rise to a variety of detectable precursor phenomena. Studies of this sort require, however, extensive instrumentation around the volcano and continuous monitoring to establish baseline information. Because of the costs involved, only a dozen of the earth's volcanoes are well monitored today, and consequently, specific prediction of an eruption is only feasible for these volcanoes.

As magma moves within the crust, it leads to deformation of the overlying ground surface. Thus, ground deformation studies have become one of the most reliable volcano-monitoring techniques and form a basis for short-term prediction. The change in shape of a volcano is monitored by repeated trilateration and triangulation, long-ranging distance measurements between benchmarks on the volcano and in the surrounding terrain, and measurement of ground tilt on the flanks of the volcano. Changes in shape can be quite dramatic, as, for example, in the case of Mount St. Helens before the 1980 eruption, when a bulge on the north flank grew by as much as 2.5 m/day owing to intrusion of magma at a high level within the volcano. Renewed volcanic activity is often preceded by influx of magma into a magma chamber, resulting in inflation of the overlying volcano edifice. The inflation is often radial and centered over the reservoir, with ground uplift of the order 1 to 10 mm/day. Such inflation episodes preceded the 1975 to 1984 eruptions of Krafla (Iceland), and similar recent ground inflation in the calderas of Rabaul (Papua, New Guinea) and Campi Flegrei (Italy) led to the issue of volcano alerts, but the level of activity subsided without an eruption.

One of the most useful sources of data for volcanic predictions is seismic monitoring. Earthquakes are generated by a number of mechanisms within the volcanic system preceding an eruption, such as during fracturing of the crust surrounding an inflating magma reservoir or due to thermal stresses in the crust. Volcanic earthquakes are often seen to increase in numbers several weeks or days before an eruption. The sudden increase in seismic energy release (strain release) is thus a valuable basis for short-term prediction. Changes in heat flow from a volcano and changes in the composition and emission rate of volcanic gases can be useful in anticipating an eruption. However, these monitoring techniques are less reliable than seismic and deformation studies because of the relatively slow diffusion rates of heat and gases through the earth's crust compared to the instantaneous transmission of seismic waves.

## SEE ALSO THE FOLLOWING ARTICLES

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