2

Subaerial Terrestrial Volcanism

Eruptions in Our Own Backyard

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2.1. INTRODUCTION

Current understanding of volcanic eruptions is the result of millennia of written observations refined by decades of scientific research. There is still much to learn about the details of how individual volcanoes work, but the existing body of literature about subaerial volcanism on Earth represents the basis against which all other volcanic eruptions are compared. Excellent books summarize the processes and products of subaerial volcanism (e.g., Macdonald, 1972; Williams and McBirney, 1979; Cas and Wright, 1987; Cattermole, 1989; Francis, 1994), and it would be impossible to condense all of this information into one chapter. However, it is important to provide a concise compilation of the most salient aspects of subaerial volcanism to which all other examples of volcanism can be compared.

The purpose of this chapter is to review the basic eruptive styles, landforms, and products that result from volcanism on Earth's surface, where observations and samples are both readily obtainable. At present, documented samples of extraterrestrial lavas have been collected and returned only from the Moon, and these materials are all basaltic in composition (Heiken *et al.*, 1991). However, to these Apollo and Luna samples can be added a suite of basaltic meteorites now thought to come from Mars (McSween, 1994; see Chapter 4), plus various remote sensing data sets that indicate a preponderance of basaltic materials on rocky surfaces throughout the solar system (Basaltic Volcanism Study Project, 1981). Thus, we focus primarily on basaltic volcanism, but information on other compositional types is also included both to illustrate the diversity of volcanism on Earth, as well as to cover the strong

Environmental Effects on Volcanic Eruptions: From Deep Oceans to Deep Space.
Edited by Zimbelman and Gregg, Kluwer Academic/Plenum Publishers, New York, 2000.

likelihood that our present information substantially underrepresents the full volcanic diversity throughout the solar system (see Chapter 8).

This chapter begins by following volcanic magma at depth as it rises toward the surface. The location and style of volcanic eruptions on Earth are strongly influenced by both the plate tectonic setting and the presence of localized, deep-seated hot spots, which guides our understanding of most terrestrial volcanic centers. Effusive eruptions result in a plethora of products, but there are some basic forms and textures that transcend the location or type of individual volcanoes. Explosive eruptions occur when escaping magma is violently disrupted; for basaltic lavas, this tends to involve interaction with ground or surface water, but more evolved magma chemistries fragment through the degassing and violent explosion of magmatic volatiles. Some parts of Earth's surface were the sites of massive outpourings of lava, resulting in localized Large Igneous Provinces (LIPs), apparently special cases of how magma reaches Earth's surface. The chapter closes with a brief discussion of how lava and ash interact with Earth's subaerial environment, including how that environment is altered through magmatic input.

2.2. MAGMA GENERATION AND RISE CONDITIONS

2.2.1. Magma Generation

Many of the processes involved in the genesis of materials that will eventually erupt at the Earth's surface take place at such great depths that little is known about their details. However, the synthesis of many lines of investigation (e.g., geophysical, geochemical, and theoretical) permits formulation of some general models. Basaltic magmas are thought to originate at depths of 50–170 km as a consequence of partial melting of mantle materials. The source rocks most likely are garnet peridotites or lherzolites, which are crystalline aggregates of olivine, orthopyroxene, and clinopyroxene with other constituents (e.g., garnet) (Yoder, 1976; Wright, 1984). Energy to melt the parent rock is derived from a wide variety of sources. The most significant sources for basalt include: (1) pressure-release melting induced by diapiric rise (Ramburg, 1972) or solid-state convection (Verhoogen, 1954) in the mantle; (2) friction related to viscous strain (Shaw, 1973), shear strain along propagating cracks (Griggs and Baker, 1969), or tidal dissipation (Shaw, 1970); (3) reduction in melting temperature by the addition of volatiles (Yoder and Tilley, 1962); (4) conductive heat trapping as a result of changes in thermal conductivity with temperature and pressure (McBirney, 1963); and (5) internal radiogenic heat production. The temperatures at which partial melts form are a function of source depth and composition, but because of the release of latent heat the melting process should be approximately isothermal for any given case. For example, the formation of Hawaiian tholeiite magma at a depth of 60 km will take place at ~1350–1400°C (Decker, 1987). Melting commences along boundaries between mineral grains, initially forming thin films (Yoder, 1976). As melting progresses the films increase in volume until segregation into discrete melt pockets takes place, a process governed by the rheology of the liquid-rock mush and the balance between melt buoyancy, local stresses, and viscous forces (Spera, 1980).

2.2.2. Rise Mechanisms

The accumulation of the melt pockets to form larger magma bodies will eventually cause net vertical motion as a result of increased buoyancy forces. In the asthenosphere, magma may rise as diapirs by deforming the ductile country rock, the ascent rate being a function of diapir size, density contrast, and rheologic properties. However, ascent through the lithosphere must overcome the rigidity of the cooler rock, and magma may stall at the asthenosphere–lithosphere boundary (Pitcher, 1979) before sufficient stresses accumulate to initiate brittle fracture and continue motion through the resultant dikes (Sleep, 1988). Buoyant ascent may also be halted in the shallow crust as the country rock density decreases as a result of the decreasing compaction related to lithostatic overburden. Such levels of neutral buoyancy may therefore define storage zones where trapped batches of melt accumulate to form a magma reservoir at depths of a few kilometers (Rubin and Pollard, 1987; Ryan, 1987). Further dike propagation is precluded until some driving mechanism overcomes the barrier imposed by the density trap. Possible examples include pressurization of the magma reservoir related to continued magma input from below, or redevelopment of buoyancy as a result of volatile exsolution and generation of a zone of low-density bubble-rich magma (Jaupart and Vergniolle, 1989), and/or loss of dense mineral components related to fractional crystallization (Sparks *et al.*, 1980). An increase in reservoir pressure might lead to lateral dike propagation within the edifice, whereas buoyancy-driven motion would be more nearly vertical. The velocity of magma in the dike and consequent eruption rate are governed by the magnitude of the driving forces (pressure or buoyancy), dike width, and magma rheologic properties (Wilson and Head, 1981). Eruption volume may be linked to the size of the subsurface reservoir, where one exists (Blake, 1981). However, not all magmas will erupt: Velocities must be sufficient to avoid excessive cooling and solidification, and pressure forces must be sufficient to drive the magma all the way to the surface.

2.2.3. Vesiculation and Fragmentation

During the final stages of magma ascent, the behavior of magmatic volatiles such as H_2O, CO_2, and SO_2 can critically influence the manifestation of the eruption. Volatile solubility is a function of magma composition and pressure (i.e., depth) (Mysen, 1977). CO_2 and SO_2 are less soluble than H_2O and usually they will exsolve at much greater depths. Consequently, these may be lost during transport or residence in a magma chamber. As the confining pressure decreases during ascent, the volatiles exsolve and gas bubbles nucleate and grow, initially by diffusion of the volatiles through the melt and into bubbles; later, as the pressure drops (i.e., approaching the external ambient pressure), bubble size increases and decompression dominates the growth (Sparks, 1978). If the near-surface magma rise speed is relatively low, bubble rise velocity may exceed that of the melt because the buoyancy forces increase with bubble size. Runaway bubble collision and coalescence may take place to form very large bubbles ascending much faster than the magma, producing intermittent, strombolian-type eruptions (see Section 2.5.2) (Blackburn *et al.*, 1976). Alternatively, such gas pockets might form in shallow reservoirs by coalescence of a bubbly foam. On reaching the surface of the magma column the bubbles burst, ejecting a spray of small magma clots and fragmentation products formed from disrupted bubbles. At magma rise speeds of $\sim$0.5 to 1 m s^{-1} the bubbles remain effectively locked to the magma during ascent. Once the gas occupies a critical fraction ($\sim$75 vol%) of the total magma volume (such that the bubbles are close-packed and further expansion is prohibited) which typically occurs at depths of a few tens to a few hundred meters, disruption of the magma into a collection of magma clots entrained in a gas stream is almost inevitable (Sparks, 1978). These conditions favor the formation of sustained, Hawaiian-type lava fountains or plinian columns (Wilson and Head, 1981). However, if enough gas is lost during either storage or transport, lava will simply effuse passively from the vent.

A number of interrelated factors, associated mainly with magma composition, influence the "explosivity" of an eruption. The low viscosity of basalt means that viscous opposition to bubble expansion is negligible, allowing bubbles to grow relatively large and precluding significant excess pressures in the bubble (Sparks, 1978). Although fragmented to some degree, the resulting eruptive products are coarse-grained. Bubbles in magmas with higher silica contents and viscosities can develop significant excess pressures and undergo sudden, violent decompression on reaching the fragmentation level, disrupting the magma into fine particles that are able to transfer heat efficiently to the surrounding atmosphere. The heated atmosphere is then entrained to drive buoyantly a vigorous eruptive plume (Wilson, 1976). By comparison, basaltic eruptions involve low gas pressures, large pyroclasts and, together with the low total volatile contents, these factors contribute to the low fountains (and lack of a tall convecting plume) that characterize typical explosive basaltic eruptions (Head and Wilson, 1989). More vigorous explosive basaltic eruptions are possible as a result of interaction with external water sources (e.g., Williams, 1983; Walker *et al.*, 1984). The above scenario can be compared to magma rise on other planetary bodies (see Chapters 4, 6, and 8).

2.3. GLOBAL SETTING: PLATE TECTONICS AND "HOT SPOTS"

2.3.1. The Unifying Theory of Plate Tectonics

For many the ears the location and distribution of volcanoes on Earth could not be easily reconciled with existing theories. During the 1960s, geophysical and geologic data led to the "Plate Tectonic Revolution" that was finally successful in presenting a unified theory that accounted for the global distribution of volcanic and seismic phenomena (see Press and Siever, 1974; Summerfield, 1991; Francis, 1994). The strong correlation of active volcanism and seismicity with the margins of crustal plates (Figure 2.1) provided a rationale for why

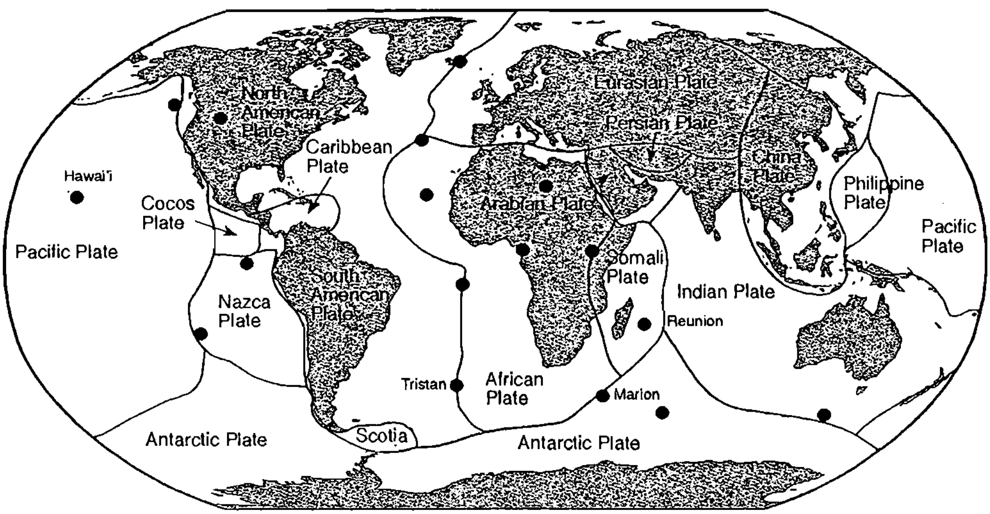


Figure 2.1. Plate margins superposed on the outline of the continents. Selected prominent hot spots are also indicated by dots. Redrawn from Simkin *et al.* (1994).

most volcanic eruptions occurred where they do. Concentrations of volcanoes are found at island arcs and continent-margin subduction zones, mid-ocean ridges, and continental rift valleys, representing either divergent or convergent boundaries.

2.3.2. Divergent Margins: Generation of New Crust

Plates of Earth's crust move away from each other at divergent margins. The spreading apart of two plates causes passive upwelling in the underlying asthenosphere, resulting in partial melting of the asthenosphere and subsequent volcanism. The gross morphologic and volcanologic characteristics of divergent margins vary widely, and depend on whether the rifting is occurring on land (such as the East African Rift) or underwater (such as the Mid-Atlantic Ridge). The vast majority of Earth's divergent margins are found on the ocean floor, where they form a submarine volcanic chain approximately 75,000 km long called the mid-ocean ridge (MOR). Spreading rate has a strong control on morphologic and volcanologic characteristics. For discussion, MORs are classified on the basis of spreading rate: slow ($\lesssim$30 mm a^{-1}, full spreading rate), intermediate ($\sim$60–90 mm a^{-1}), and fast (>90 mm a^{-1}). The northern Mid-Atlantic Ridge, excluding Iceland (the only subaerial portion of a MOR), typifies a slow-spreading center while the Juan de Fuca Ridge and northern East Pacific Rise are type examples of intermediate- and fast-spreading margins, respectively.

At a MOR, the forces of volcanism, magmatism, and tectonism interact to generate the overall ridge morphology, which is reflective of the relative dominance of these processes over the scale of a million years. However, all spreading ridges are characterized by a broad topographic rise capped by a central valley, graben, or collapse trough (Chadwick and Embley, 1998; Fornari *et al.*, 1998). The width and depth of the summit depression depend on the long-term magma supply. For example, the slow magma supply to the Mid-Atlantic Ridge results in a central valley as wide as 40 km that is flanked by faulted walls reaching as high as 2 km above the valley floor (Mutter and Karson, 1992). The floor of the axial valley contains large volcanic constructs termed *axial volcanic ridges* that can be up to several kilometers long and as high as 2000 m (Smith and Cann, 1996). In contrast, the magmatically robust East Pacific Rise near 9°50′N is characterized by a narrow ($\sim$40 m wide), shallow (5–8 m) summit collapse trough formed by repeated near-surface collapse of lava flow crusts (Haymon *et al.*, 1991; Fornari *et al.*, 1998). The southern East Pacific Rise near 17°30′S does not display a summit trough, but instead appears to be covered with a relatively young lava flow that could have obliterated any previous structure on the axial summit (Auzende *et al.*, 1996). These trough-filling eruptions may take place approximately every 5–10 years per 10 km of ridge (Fornari *et al.*, 1998).

The precise frequency and duration of volcanic eruptions at MORs are unknown because technologic and financial constraints preclude constant monitoring of the global MOR system (see Chapter 5). The Juan de Fuca Ridge has been monitored with a SOund SUrveillance System (SOSUS) since 1993, and since that time, four eruptions have been detected (Fox *et al.*, 1995; Chadwick *et al.*, 1998). These events lasted up to 14 days, and emplaced $\sim 10^5$–10^6 m^3 of lava, giving an average effusion rate of $\sim$1 to 100 m^3 s^{-1} (Gregg and Fink, 1995; Fox, 1998; Chadwick *et al.*, 1999). The site of an eruption between 9°46′ and 52′N on the East Pacific Rise was visited shortly after activity ceased (Haymon *et al.*, 1993). Results of recent modeling suggest that here $\sim 10^6$ m^3 of lava was emplaced in <2 h, giving an average effusion rate of 10^4–10^5 m^3 s^{-1} (Gregg *et al.*, 1996). Based on mapping as well as numerical and physical modeling of submarine lava flow behavior, it appears that eruptions at fast-spreading centers are more frequent, of shorter duration, and have smaller volumes than those at slow-spreading centers (Smith *et al.*, 1995; Gregg *et al.*, 1996; Chadwick *et al.*, 1998).

Iceland, located directly on the Mid-Atlantic Ridge, provides the unique opportunity to study MOR volcanism and tectonism in a subaerial environment. The volcanism on Iceland consists primarily of basalt and icelandite (a term recently applied to possible basaltic andesite compositions on Mars, see Chapter 4). The Mid-Atlantic Ridge is visible on Iceland as a series of fissures across which the spreading of the rift has been closely monitored (e.g., Thorarinsson, 1967). Historic fissure eruptions in Iceland have provided valuable insight into divergent margin volcanism; these include the Laki eruption of 1783 (Thorarinsson, 1969; Thordarson and Self, 1993) and a fissure eruption within the Askja caldera, which was closely observed by geologists in 1961 (Thorarinsson and Sigvaldason, 1962). The manifestation of Icelandic (and MOR) volcanism is generally very distinct from that associated with rifting within continents, such as along the East African Rift, where exotic lavas like carbonatite (see Chapter 8) accompany the basalts.

2.3.3. Subduction: Destruction of Old Crust

Lithospheric material descends back into Earth's interior along subduction zones; as the name implies, lithosphere is forced under either oceanic or continental crustal materials. Subduction zones are compressive tectonic regimes, with driving forces resulting primarily from the pulling force of the descending slab, along with the relative motion of the plates on either side of the trench (see Summerfield, 1991, pp. 48–54; Bebout *et al.*, 1996).

Subduction zones formed between two oceanic plates are characterized by a topographic trench along the line where the subducted slab disappears, with a line of volcanic islands paralleling the trench on the overriding slab. Because of the geometry of a spherical planet, volcanoes associated with oceanic subduction zones tend to form broadly curving archipelagos ("arcs") of volcanic islands. The subducted lithosphere consists of both basalts and peridotites, plus a liberal coating of oceanic sediments and trapped seawater. As the slab descends into the hot mantle, water and other volatiles are "boiled" out of the slab. The volatiles ascending from the slab alter the properties of the asthenospheric wedge between the downgoing and overlying plates, leading to extensive partial melting (see Stein and Stein, 1996). These melts rise through the overlying oceanic plate to produce island arc volcanoes, typically found between 70 and 90 km from the associated subduction trench. Erupted materials include basalts, broadly similar to those found at ocean ridges but with a slightly different chemistry distinctive of island arc basalts, as well as minor amounts of silica-enriched basaltic andesites and andesites.

Where an oceanic plate collides with a continent, the resulting subduction zone is broadly similar to that of an oceanic subduction zone described above, but with the added complication of continental crustal materials. The ascending magma can now cause partial melting of the lower continental crust, providing new (usually silica-rich) materials for incorporation into the magma as it rises through this region. Continental margin volcanism, such as that found in the Andes along the western margin of South America, tends to be dominated by more evolved materials such as basaltic andesite, copious quantities of andesite and dacite, but very little basalt, most of which is underplated at the base of the crust (Francis, 1994, p. 38). Fractional crystallization can eventually lead to very silica-rich magma such as rhyolite.

Both island arc and continental margin volcanoes contribute to make one of the largest subaerial volcanic regions on Earth. The margin of the Pacific Ocean, comprised of many hundreds of active or recently active volcanoes, has been termed the "Ring of Fire"; it is the result of subduction processes at the convergent boundaries around the Pacific plate. In comparison with subduction zones, the magmatic sources for intraplate volcanism tend to be much deeper within the Earth.

2.3.4. Hot Spots: Deep Mantle Sources

The deep interior of the Earth accumulates heat at a rate faster than it is lost by the combined processes of conduction to the Earth's surface and of volcanism at oceanic spreading centers and subduction-related volcanoes. This causes large diapirs of hot mantle material known as *mantle plumes* (Figure 2.2) to rise toward the surface, presumably from the deep mantle (see Francis, 1994, pp. 38–47). As the large plume head rises, it remains connected to its source by a much narrower tail, through which additional material also rises.

When the plume head flattens itself against the base of the oceanic or continental lithosphere, it causes doming of the crust via both physical uplift and thermal expansion of the crustal rock (Phipps Morgan *et al.*, 1995), as well as radiating dike swarms (see Chapter 5). The material in the 800- to 1000-km-diameter plume head begins to melt primarily as a result of the decrease in lithostatic pressure. The volume of such a diapiric plume head is great, so the amount of magma it can produce is tremendous. Huge areas of flood basalts are the result, and they have been found across the globe (see Section 2.6). The Columbia River Plateau basalt in the USA's Pacific Northwest is a young and well-studied example; its total area is 164,000 km^2 and total volume is approximately 174,000 km^3, with individual basalt lava flows reaching up to 2000 km^3; 90% of its volume erupted over a period of 1.5 Myr (Tolan *et al.*, 1989; Reidel and Hooper, 1989). Other flood basalt provinces include the Deccan Traps in India and the Paraná Basin in Brazil, both on continental crust. Not all of the

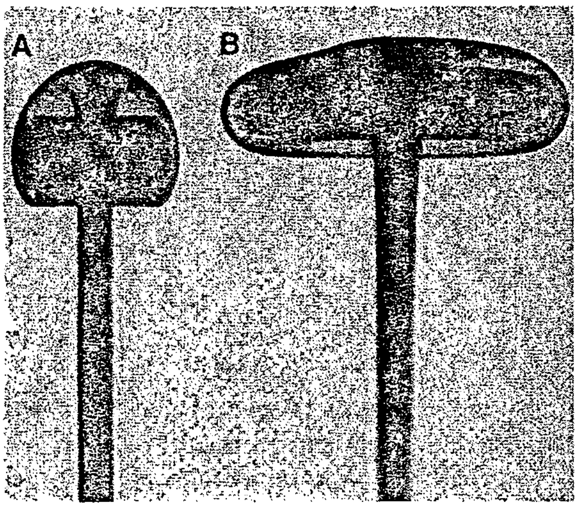


Figure 2.2. Plumes of material rising from the base of the mantle are thought to have a similar morphology and behavior to these plumes of hot glucose syrup (dark) rising diapirically through cooler syrup (light) in a lab experiment at the Australian National University. (A) Rising plume entraining cooler material. (B) Impingement of the plume head against a horizontal barrier (i.e., the base of the lithosphere). After the head becomes stagnant, hot material continues to rise through the tail of the plume. Modified from Hill *et al.* (1992).

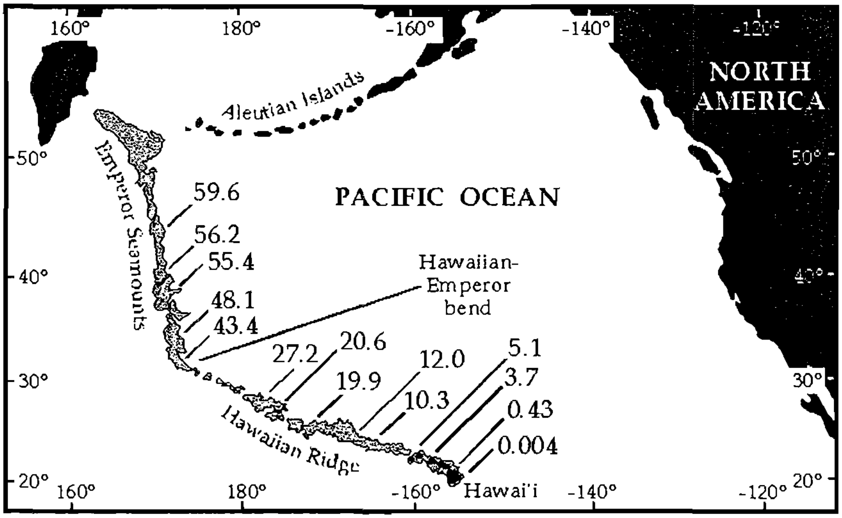


Figure 2.3. The Hawaiian-Emperor chain of seamounts and islands, marking the motion of the Pacific plate over the Hawaiian hot spot. Numbers show the oldest reliable ages (in Ma) of selected basaltic volcanoes. The bend in the chain marks a change in the vector of motion of the Pacific plate about 40 Ma. Redrawn from Calgue and Dalrymple (1987).

magma generated by the plume head is erupted; much may be intruded as dikes and sills within the crust, adding heat to the crust and changing its bulk composition. The plume head eventually equilibrates thermally with the upper mantle and ceases producing magma. However, hot mantle material may continue to rise through the tail for many millions of years. This represents a "hot spot" beneath the crust, so magmatism continues, though at much reduced rates (Richards *et al.*, 1989).

Hot spots are thought to be fairly stationary relative to the mobile tectonic plates that make up the Earth's surface. Thus, as the lithospheric plates move over hot spots, tracks of volcanic centers are formed on the crust, in a manner similar to the way burn holes form in a sheet of paper being moved horizontally just above a candle flame. In the ocean basins, hot spot traces are expressed as linear chains of islands and seamounts, each of which is composed of several basaltic shield volcanoes. The Hawaiian-Emperor island and seamount chain (Figure 2.3), active for at least 80 Myr, is perhaps the most extensive and best-known example (Clague and Dalrymple, 1987).

On the continents, hot spots generally leave tracks of silicic volcanic centers instead of basaltic shields. The Snake River Plain (White *et al.*, in press), in southern Idaho, USA, is believed to be primarily the track of the tail of the hot spot presently under Yellowstone (Figure 2.4). The head of this same plume may also be responsible for the Columbia River basalts. Volcanic activity along the Snake River Plain (SRP) began about 16 Ma in south-eastern Oregon and southwestern Idaho and progressed to the Yellowstone Plateau, which has been volcanically active for the past 2 Myr. Both discrete and less well-defined silicic eruptive centers, up to at least 100 km in diameter, were formed sequentially as the North American tectonic plate moved southwestward over the plume. Basaltic magma generated from the material in the plume tail provides the heat source for SRP plume track volcanism, leading to abundant basalt and silicic products in close proximity. Fractional crystallization of the basaltic magma, plus partial melting of basalts intruded into the deep crust, generated the silicic magmas; melting of silicic crustal rocks can also occur. Hot spot magmatism tends

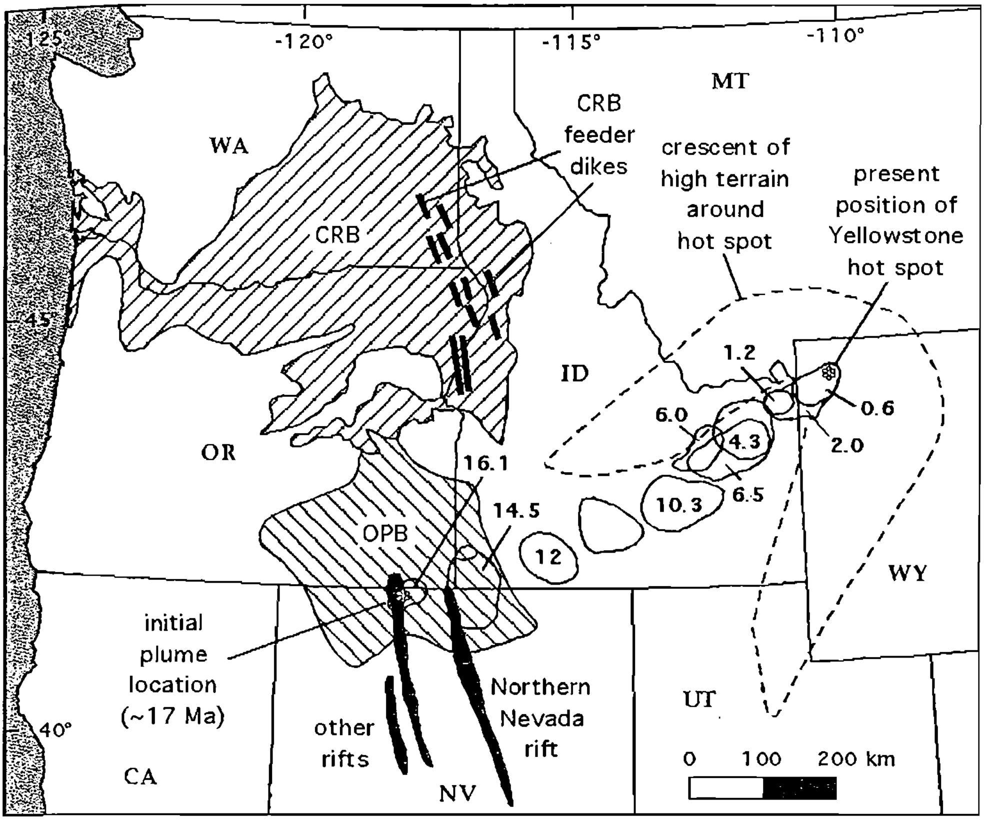


Figure 2.4. The Columbia River Basalt (CRB) and related volcanics in the northwestern United States. Impingement of the head of the Yellowstone plume at ~17 Ma is commonly thought to have led to rifting in northern Nevada, and possibly to the eruption of the CRB and the Oregon Plateau Basalts (OPB), both ~17–13 Ma. After magmatism associated with the plume head waned, the plume tail created a track of silicic eruptive centers (labeled with ages in Ma) as the North American plate moved southwest, relative to the hot spot. The hot spot is still active and is presently located beneath Yellowstone National Park. Redrawn from Pierce and Morgan (1992).

toward silicic magmas with somewhat lower water contents compared with silicic magmas from non-hot-spot-related centers. These lower water contents manifest themselves in extremely voluminous silicic lava flows, which are common features of hot spot centers (Bonnichsen, 1982; Manley, 1995). Elsewhere on the continents are large, long-lived, silicic centers similar to those along hot spot tracks, but which do not seem associated with any hot spot.

2.4. EFFUSIVE ERUPTIONS

2.4.1. Lava Flows

The flow of lava is the primary mechanism through which effusive eruptions build edifices and invade the surrounding terrain. Flow properties are inherently associated with a host of parameters such as viscosity, density, temperature, and crystal content (see Chapter 9), but remarkably most lava flow textures fall within a few broad categories. The following discussion focuses primarily on basaltic lava (see Williams and McBirney, 1979, pp. 106–112), but includes brief references to other significant compositional types. Also note that transitional textural types can be found between these basic types.

A'a. The Hawaiian word *a'a* refers to flows covered with jumbles of rough, clinkery, and spinose fragments, ranging from small chips to blocks meters across (Figure 2.5a). Flows with a'a texture range from one-half to tens of meters in thickness, and are usually comprised of distinct lobes fed by a central leveed channel. Once solidified, a'a lava typically displays rubbly upper and lower surfaces, with a massive central unit where the liquid core of the flow solidified under the insulating effects of the rubble zones. Basalts, basaltic andesites, and andesites all commonly display a'a textures.

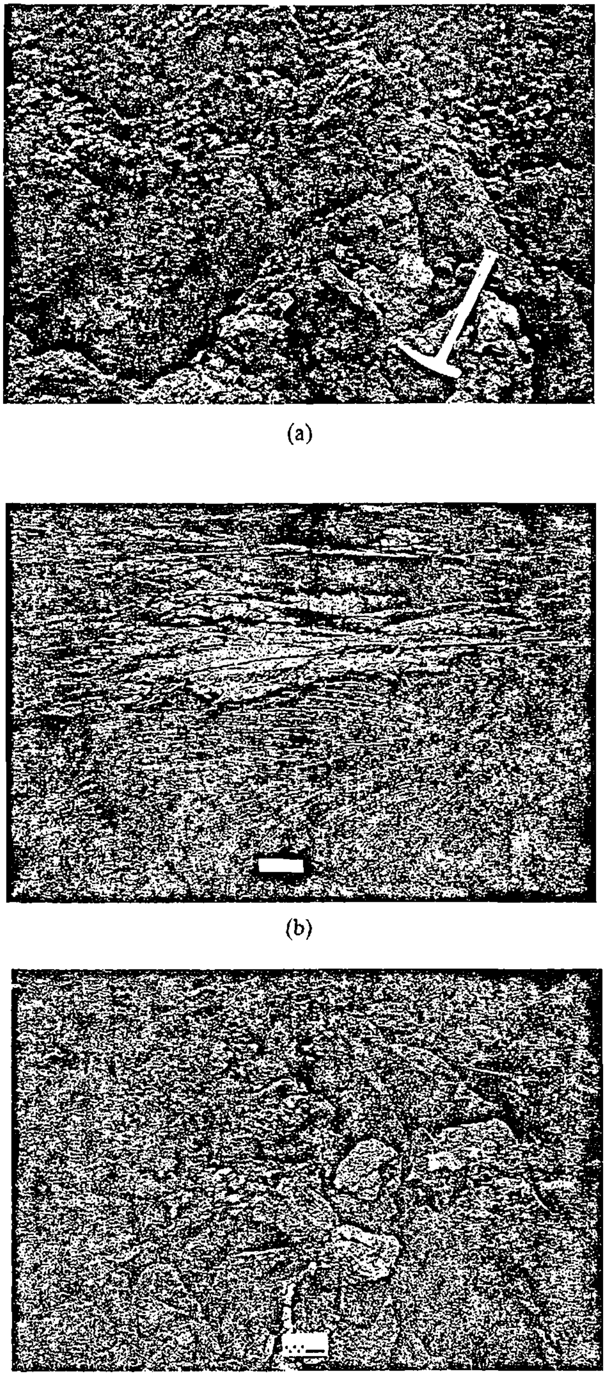
Pahoehoe. The Hawaiian word *pahoehoe* refers to flows with smooth crusts that can occur in a bewildering array of shapes, ranging from twisted ropes (Figure 2.5b) to shelly blisters. Pahoehoe textures result from fluid lava usually emplaced at relatively low effusion rates (see Section 2.4.3); pahoehoe flow fields are composed of many thousands of individual flow units or "toes." These toes are usually <50 cm thick, up to 1 m wide, and up to 10 m long. Individual flows can transition from pahoehoe to a'a, particularly if the localized flow rate increases such as on a steep slope. Adjectives such as *scaly*, *shelly*, *slabby*, and *massive* have been used in field descriptions of pahoehoe flows, along with tumuli and "squeeze-ups" for pahoehoe that is locally extruded along cracks into the flow crust. A glassy pahoehoe surface is readily weathered in temperate environments, leaving vesicular blocks lacking the intricate pahoehoe surface textures. Near-vent basalts commonly have pahoehoe textures, but pahoehoe is usually absent from more silica-rich lavas. Recently portions of the Columbia River Basalt (CRB) sequence have been interpreted to be inflated pahoehoe flows (Self *et al.*, 1996, 1997).

Block Flows. Block flows are sometimes considered synonymous with a'a flows, but the term should be restricted to flows made up largely of detached, polyhedral blocks with planar to curved faces (Figure 2.5c). Block basalt flows are much less common than a'a flows, and they are best developed in intermediate to siliceous flows that formed thick, glass-rich crusts that subsequently fractured into angular blocks. Andesite, dacite, and rhyolite flows display blocky characteristics that grade with increasing volume into domes, discussed in Section 2.4.4 below.

Pillow Lavas. Pillow lavas may be the most abundant volcanic rocks on Earth, since they form where lava is slowly erupted under water (see Chapters 3 and 5); they are essentially the underwater equivalent of pahoehoe. Because this chapter concentrates on subaerial eruptions, we will not discuss them further here. However, the distinctive budded pillow morphology occurs in flows ranging from ultrabasic to rhyolitic in composition; more viscous lavas tend to generate larger pillows.

2.4.2. Constructs

When a centralized volcanic vent is active for a prolonged period of time, substantial constructs are formed through the accumulation of lava flows and pyroclastic material. As with lava flows, volcanic constructs take on a myriad of shapes and forms. Fortunately, two broad types of constructs encompass the majority of volcanic landforms encountered throughout the world.



(a)

(b)

(c)

Figure 2.5. Lava flow textures. (a) A'a flow. Episode 5 of the Pu'u O'o eruptions of Kilauea, Hawaii. (Photo by J. Zimbelman, Jan. 1984.) (b) Pahoehoe flow. 1969–1972 flows from Mauna Ulu, Hawaii. (Photo by J. Zimbelman, Sept. 1983.) (c) Block flow. SP flow near Flagstaff, Arizona. (Photo by J. Zimbelman, June 1998.)

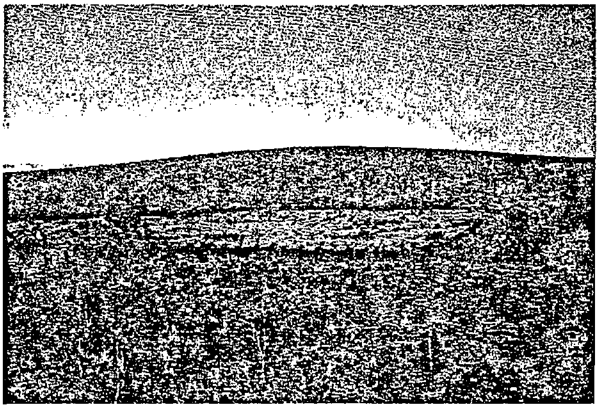
Shield Volcanoes. Repeated eruption of fluid lava from a central vent produces a broad convex-upward volcanic edifice called a *shield volcano* (Figure 2.6a), based on its resemblance to the rounded shields of early Germanic warriors (Macdonald, 1972, p. 271). Shield volcanoes typically have flanks with slopes of 4° to 5°, culminating in a collapse caldera or caldera complex near the summit. The edifice is comprised of countless individual lava flows stacked adjacent to and on top of each other, with only a very small amount of included pyroclastics. Shield volcanoes are predominantly comprised of basaltic lavas, although basaltic andesites are not uncommon. Well-known examples include Mauna Loa and other volcanoes in Hawaii, the smaller but very symmetrical Icelandic shields like Skjalbreiður, and the Galapagos archipelago where some volcanoes have flank slopes >20° (Williams and McBirney, 1979, pp. 197–205). Huge shield volcanoes occur on other planets, including both Mars (see Chapter 4) and Venus (see Chapter 5).

Composite Volcanoes. If effusive flows (generally more viscous than basalt) are intermixed with substantial pyroclastic deposits, a composite volcano is formed (Figure 2.6b), the classic conical form that even nongeologists associate with a volcano. The term *composite* is preferable to the earlier term *stratovolcano* because shields and many domes can also be considered to be stratified (Williams and McBirney, 1979, p. 179). Rarely is the composite structure built of regularly alternating effusive and pyroclastic layers, but the relatively common mixture of flows and pyroclastics (Figure 2.6c) lets the volcano flanks attain slopes of 10° to 12°, more than twice the average slope of a shield volcano. Composite volcanoes are predominantly comprised of andesite and dacite materials, which contributes to the abundance of pyroclastics, although once again basaltic andesites are not uncommon. Composite volcanoes are common throughout the world, but Fujiyama in Japan, several of the Cascade peaks in the western United States, and preeruption Mt. Pinatubo in the Philippines are among the better-known examples.

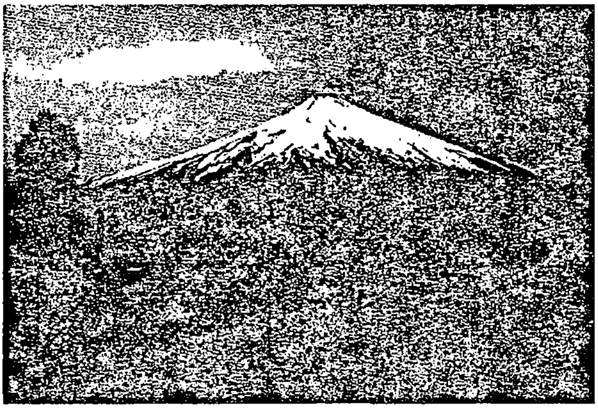
Other Volcanic Centers. Many volcanoes do not fall into either the shield or composite category. Pyroclastic (cinder) cones are monogenetic constructs composed of scoriaceous ejecta (Figure 2.6d). When a cinder cone eruption stops, it rarely erupts again at the same vent; instead, a new cinder cone forms in the general vicinity of the previous eruption, resulting in local concentrations of tens to hundreds of cinder cones. Fissure eruptions likely produce the largest volumes of effusive materials (see Section 2.7), but they rarely result in significant landforms other than low spatter cones or ramparts along the active segment of the fissure. Volcanic domes represent an important volcanic construct distinct from other effusive or explosive volcanoes (see Section 2.4.4).

2.4.3. Flow Effusion Rates

The emplacement of lava flows is governed by the lava rheology (a complex function of composition, temperature, volatile content, and so on), preflow topography (slope and roughness), and effusion rate, perhaps the single most important parameter in characterizing an eruption. An evaluation of documented Hawaiian eruptions led Rowland and Walker (1990) to conclude that effusion at a rate <10 m^3 s^{-1} results in almost exclusively tube-fed pahoehoe flows whereas eruptions at >20 m^3 s^{-1} are almost exclusively a'a flows. The average effusion rate for current eruptions at Kilauea is $\sim$5 m^3 s^{-1} (Wolfe *et al.*, 1989). From 1983 to mid-1986 the Kilauea eruption was characterized by distinct episodes at high effusion rates that produced high fountains and fast-moving a'a flows. Since mid-1986 the eruption has been more or less continuous, producing pahoehoe flows and little or no pyroclastic activity. In contrast, the 1984 Mauna Loa eruption emplaced a 27-km-long a'a flow over 21



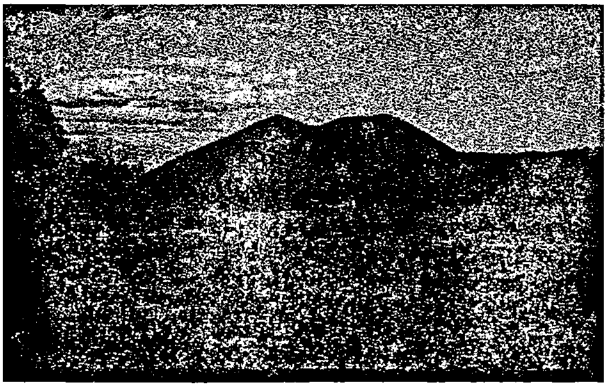
(a)



(b)

(c)

Figure 2.6. Volcanic constructs. (a) Shield volcano, Mauna Loa, Hawaii. (Photo by J. Zimbelman, Sept. 1983.) (b) Composite volcano, Villarrica, Chile. (Photo by J. Zimbelman, Dec. 1985.) (c) Ash interbedded with basaltic andesite lava, on the northern flank of Villarrica volcano, Chile. (Photo by J. Zimbelman, Dec. 1985.)



(d)

Figure 2.6. Volcanic constructs. (d) Cinder cone. Sunset Crater, north of Flagstaff, Arizona. (Photo by J. Zimbelman, July 1992.)

days at an average eruption rate of ~110 m^3 s^{-1} (Lipman and Banks, 1987). The largest effusion rate inferred for a historical flow comes from the Laki eruption of 1783–1785 in Iceland, with a peak effusion rate of 4600 m^3 s^{-1}, equivalent to an effusion rate of ~6 m^3 s^{-1} per meter of active fissure (Thordarson and Self, 1993). Such observations are important considerations when examining the emplacement of large lava flows on other planets (Zimbelman, 1998).

Lava tubes can significantly enhance the delivery of relatively unaltered lava to an active flow front well removed from the vent (e.g., Greeley, 1987). Tubes develop primarily within fluid basaltic lavas, at least in some cases through the roofing over of an active channel (Greeley, 1971). Complex networks of tubes can feed the emplacement of many large flow fields. Some drained lava tubes can be very large; tube sections in the Undara volcanic field in northeastern Australia attain widths of up to 20 m (Atkinson *et al.*, 1975). Calculations of the thermal efficiency of a well-developed tube indicate that lava can be transported many hundreds of kilometers without losing enough heat to alter the lava rheology (Keszthelyi, 1995; Sakimoto *et al.*, 1997). Tubes may play an important role in the emplacement of enormous volumes of basaltic lava found around the world (see Section 2.6).

2.4.4. Volcanic Domes

In the simplest definition, a lava dome is an extrusion of lava with a thickness comparable to its diameter. Lava domes form most often from lavas with high viscosities—andesite, dacite, rhyolite, and trachyte. Several broad types of domes have been distinguished, based largely on gross morphology, but with differences in lava rheology (primarily the yield strength of the lava) as the underlying factor. Domes can display a variety of surface textures (e.g., Anderson and Fink, 1990), but the various emplacement mechanisms lead to the following basic types.

Upheaved Plug Domes. When the lava solidifies in the vent and becomes very strong before reaching the surface, it may be forced slowly upward in a steep-sided, coherent mass of rock that can barely spread under its own weight. Lassen Peak, in northern California, USA, and the Chain of Puys in France are well-known examples.

Peléean Domes. Lava that reaches the surface before solidifying, but which still has an appreciable strength, will form a dome similar to that extruded during 1902–1903 at Mont Pelée, Martinique. Spines (small versions of upheaved plugs) and craggy piles of rubble protrude from the summits of Peléean domes, and their lower portions are usually completely engulfed by aprons of rubbly talus. The spines themselves are transitory, as they often crumble and collapse as they cool. Episodic growth over many years, punctuated by avalanches and explosions, is common with these domes. It was a block-and-ash flow from the Mont Pelée dome that destroyed the town of St. Pierre on Martinique in 1902 (see Section 2.5.2).

Coulees. Lava that would form a circular dome on a flat surface will tend to flow slowly down a slope, forming a coulee (a short but very thick lava flow). Most coulees are covered by a carapace of broken pumiceous or scoriaceous blocks. The Mono Domes in eastern California, USA, contain several classic examples. If a large volume of magma is available to erupt, a coulee may continue to increase its lateral extent, usually with only a minor increase in height; a dome may thus lose its domical aspect and become a lava flow.

Low Domes. Lava sufficiently low in viscosity so as to spread slowly away from the vent on a flat surface will form a generally circular, smooth-profile dome called a *low lava dome* by Blake (1990). Although their upper surfaces can be covered with loose rubble, they lack the craggy relief typical of a Peléean dome. Avalanches and explosions are uncommon. A well-studied example is the 1979 Soufrière dome on the island of St. Vincent (Huppert *et al.*, 1982).

Cryptodomes (Intrusive Domes). In some cases, the magma does not reach the surface but intrudes at shallow depths, doming or lifting up the overlying rock and soil. These intrusive domes have been termed *cryptodomes*, for the magma itself remains hidden.

The Dome Spectrum. A complete spectrum exists from plug domes through low domes to coulees and lava flows (Figure 2.7), because the controlling factors of viscosity and erupted volume also vary smoothly. For example, the dome that grew in the crater of Mount St. Helens from 1980 to 1986 showed aspects of both low and Peléean-type domes. Other domes change character abruptly during their growth. In the Inyo Domes of eastern California, USA, the last lava that erupted was more viscous and had a higher yield strength than the first, so it piled up high over the vent, forming a craggy Peléean-type dome surrounded by a low-dome aureole (Figure 2.8) only half as thick (Sampson, 1987).

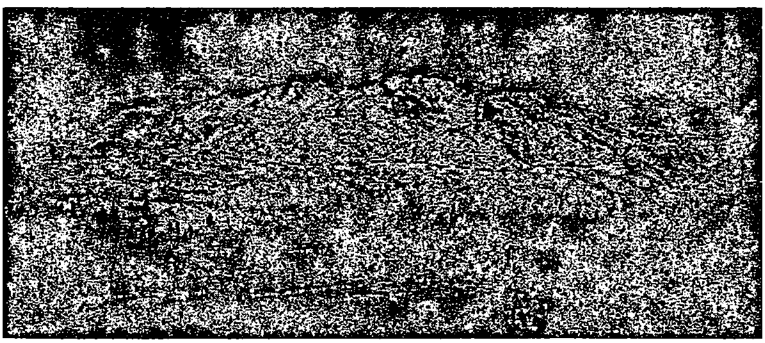


Figure 2.7. South Deadman dome, in the Inyo volcanic chain, eastern California, USA. The dashed line separates the earlier-erupted, low-yield-strength, crystal-poor lava, which makes up a flat-topped aureole that surrounds the tall (Peléean-type) center, comprised of later-erupted, high-yield-strength, crystal-rich lava. Photo courtesy of Allen F. Glazner.

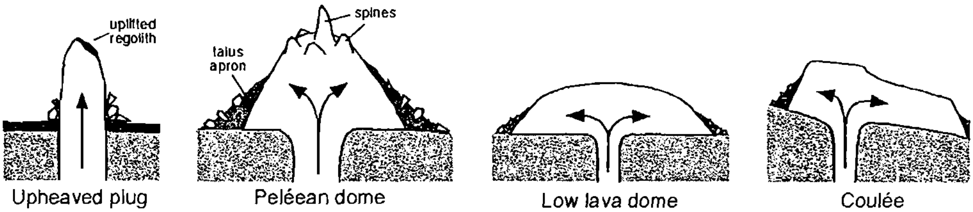


Figure 2.8. The four types of extrusive lava domes. Their morphologies are primarily controlled by the viscosity and yield strength of the lava. Redrawn from Blake (1990) and Francis (1994).

Viscosity and Size. Although the high viscosity of most silicic magmas might seem to limit the size of lava domes and flows, in fact it simply slows the rate of advance of the lava across the Earth's surface. The distance the lava may move from the vent then becomes dependent on the balance between the lava's cooling and crystallization on the one hand, and its rate of supply on the other (Manley, 1992). Voluminous silicic lava flows have now been mapped in many locations on Earth, but the great majority are associated with the trails of mantle plume hot spots across the continents, such as the Snake River Plain hot-spot track across southern Idaho, USA (Bonnichsen, 1982), and the Paraná continental flood basalt (and rhyolite) province of southern Brazil (Garland *et al.*, 1995).

2.4.5. Interactions with Meteoric Water

When a subaerial lava flow advances into wet areas such as a lake bed or a marshy depression, water can become trapped beneath the lava and flash into steam, bursting explosively through the lava to form a pseudocrater or rootless vent (Francis, 1994, pp. 151–152). One of the best-known locations for pseudocraters is around Lake Myvatn in Iceland where dozens of pseudocraters are clustered. Pseudocraters are discussed in more detail in Chapter 3, because lava–ice interactions can also contribute to steam explosions. If ground water within a shallow aquifer is brought into contact with ascending magma, steam explosions can excavate into the preexisting substrate and produce a broad, shallow depression termed a *maar*, after lake-filled craters in Germany generated in this fashion (Francis, 1994, pp. 341–342). As the water–magma ratio increases, phreatomagmatic eruptions can achieve significantly increased mechanical energy release (governed by fuel–coolant reactions) that result in tuff cones or tuff rings as the energy of emplacement increases (Wohletz and Sheridan, 1983). If the water–magma ratio continues to increase beyond the tuff cone/ring stages, eventually pillow lavas result from the rapid chilling of lava lobes in the surrounding water. When lava enters standing water >1500 m deep, the hydrostatic pressure of the water is sufficient to inhibit the formation of steam, leading to the benign generation of pillow lavas (see Chapter 5).

Interaction of basaltic volcanic ash with water shortly after emplacement can lead to alteration of the glassy shards by hydration to form palagonite. The orange-brown color of palagonite is distinctive of this wet alteration process; these materials have reflectance spectra very similar to those of the bright dusty regions on Mars. If palagonite is a significant component of the Martian fines, this would strengthen the likelihood that both basaltic products and access to a hydrating agent such as water were readily present at some point in Martian history (see Chapter 4).

2.5. EXPLOSIVE ERUPTIONS

2.5.1. Pyroclastics

Pyroclastic rocks (Greek, "fire-broken") consist of material that is fragmented and ballistically ejected by expanding gases, and which subsequently accumulates into a deposit (Macdonald, 1972, p. 23). Pyroclastics can encompass other descriptive terms like *pumice* or *ash*, although the latter term also indicates a specific size range. What mechanisms contribute to this breaking of rocks by expanding gases? By far, the most important mechanism is related to the initial fragmentation of magma, either during ascent or at the surface (see Section 2.2). Fragment sizes relate directly to basic properties of the magma itself, such as viscosity and volatile content. The greater the gas content and viscosity, the higher the gas pressure can build up within the magma; the stronger the explosions, the smaller the resulting fragments. After eruption, additional processes may further reduce the size of the erupted products. Some fragmentation results from the tensile stress exerted when the particle cools from a molten state, and abrasion between fragments during transportation can both round the particles as well as generate abundant fine abrasion products. If conditions are right, some pyroclastic particles actually grow in size after eruption, such as accretionary lapilli formed when moist ash particles adhere together, but this requires abundant meteoric water to form.

Pyroclastic materials can be broadly subdivided by the readily observable quantity of particle size. Pyroclastic deposits are termed *ash* when individual particles are <2 mm across, *lapilli* (Latin for "little stones") when between 2 and 64 mm, and *bombs* when >64 mm (Fisher, 1961). If particle size is not part of the intended description, the more generic Greek term *tephra* is preferred (Francis, 1994, p. 185). Deposit thickness measurements can be combined with the areal extent of a deposit; if measurements are sufficiently distributed, this provides a reasonable estimate of the deposit volume. Other important quantities either measured in the field or determined from later laboratory measurements include the maximum clast size in a given deposit at a given location (related to the energy of emplacement), the degree of sorting, the vesicularity of the clasts, and the presence or abundance of either crystal or lithic fragments.

Pyroclastic materials do not in themselves imply any particular composition. Magmas from basaltic to rhyolitic compositions can all produce pyroclastic materials if the eruption is sufficiently vigorous to break apart the erupting liquid. In general, the areal extent of mafic pyroclastics tends to be more restricted than that of either andesitic or especially rhyolitic magmas, which typically produce increasingly explosive eruptions.

2.5.2. Modes of Emplacement

The depositional products from explosive eruptions can be broadly divided between falls and flows where, as the names suggest, the former descend either ballistically or roughly vertically out of the atmosphere while the latter are emplaced by movement across the surface. There is no way to cover the diversity found within such deposits within one summary chapter; the interested reader is referred to standard texts for more detailed treatment (e.g., Williams and McBirney, 1979; Fisher and Schmincke, 1984; Cas and Wright, 1987; Francis, 1994).

Falls. The degree of explosivity of an eruption is a useful way to classify eruptive styles and the pyroclastic products that result from them. Early on these distinctions were assigned to type volcanoes displaying specific attributes, but more recently the Volcanic Explosivity

Index (VEI) has provided a way to quantify some of these eruptive types based on the size and dispersal of pyroclasts (Newhall and Self, 1982). At the low end of the explosive sequence are Hawaiian eruptions, which are primarily effusive but which can be accompanied by fire-fountaining that deposits a wide variety of glassy particles (e.g., spatter, clots, "Pélé's tears," reticulite). Strombolian eruptions launch magmatic cobble-sized vesicular cinders (scoria) that collect around the vent to produce the classic cinder cone; rare larger blocks are termed *volcanic bombs*. Vulcanian eruptions incorporate a large fraction of nonmagmatic lithic fragments derived from the initial "throat-clearing" and subsequent erosion of the vent wall, with products ranging in size from ash to bombs. Strombolian and Vulcanian eruptions discussed thus far deposit materials almost exclusively in the immediate vicinity of the vent, with very limited dispersal, in marked contrast to the very explosive Plinian eruption.

Plinian eruptions derive their name from both the famous Roman naturalist, Pliny the Elder, who died in the catastrophic outburst of Vesuvius in 79 AD, and Pliny the Younger, who meticulously described the eruption (Macdonald, 1972, p. 231). Plinian eruption columns can enter the stratosphere in the most dramatic cases. In Plinian and sub-Plinian explosive eruptions, magma (commonly with preeruptive water contents of 4 to 6 wt%; Lowenstern, 1995) vesiculates to the point of explosive comminution as it rises in the conduit toward the Earth's surface. This produces tephra with a large proportion of fine ash particles, so the eruptive column above the vent becomes a roiling, turbulent mass of gas and dust. Entrainment of air into the column completely cools the fine-grained tephra and carries much of it high into the atmosphere and far from the vent. The prodigious ash produced by these eruptions may travel large distances, blanketing the surroundings with deep layers of bedded deposits. Both nonmagmatic lithics and highly vesiculated magmatic pumice fragments abound; the sizes of these can provide information on the variability of intensity during the course of an eruption (Sigurdsson *et al.*, 1985). A Plinian deposit represents the most explosive product from a single vent, surpassed only by massive caldera collapse eruptions that lead to ignimbrite deposits, discussed below.

Flows. Pyroclastic flow products range from low-density, vesiculated pumice to dense, unvesiculated lava clasts; the following discussion utilizes the broad division of pyroclastic flows derived from this distinction (Fisher and Schmincke, 1984).

Ignimbrites. In broad terms, an ignimbrite is the deposit left behind by a pyroclastic flow, consisting of poorly sorted mixtures of pumice blocks and lapilli in a matrix of fine ash (Williams and McBirney, 1979, pp. 161–165). Ignimbrites are predominantly rhyolitic in composition, although rhyodacites and dacites are also common. An ignimbrite can result from one or more individual pyroclastic flows, but if multiple flows are emplaced rapidly enough, they can result in a single compound cooling unit. The pyroclastic flows that produce an ignimbrite result from many causes, but the most widely held model involves flow by collapse of a Plinian column when the local cloud density can no longer be supported by buoyancy (Sparks *et al.*, 1997, pp. 144–178). Ignimbrite materials do not conduct heat very efficiently, so that the heat remaining after emplacement can partially to completely weld the pumice and ash into a central zone of dense obsidianlike glass (Fisher and Schmincke, 1984). The top of an individual flow within an ignimbrite sometimes consists of fine co-ignimbrite ash, which settles from the cloud of hot gas and ash that rises above an active pyroclastic flow (Sparks *et al.*, 1997, pp. 180–208). The 1815 eruption of Tambora produced the largest accumulation of ignimbrite and co-ignimbrite deposits yet identified for a historic eruption (Sigurdsson and Carey, 1989).

Block-and-Ash Flows. Observers of the eruptions of Mt. Pelée in 1902 introduced the term *nuées ardentes* ("glowing clouds") to describe the pyroclastic block-and-ash flows that periodically swept down the volcano, one of which wiped out the town of St. Pierre. For our purposes, nuées ardentes can be considered pyroclastic flows where the magmatic component is dense rock rather than the vesiculated pumice found in ignimbrites (Francis, 1994, p. 247). The mechanism of generation of such flows can range from explosive disruption of a volcanic dome, as occurred on Mt. Pelée (Boudon and Gourgaud, 1989), to hot avalanches caused by the nonexplosive gravitational collapse of a dome, as at Merapi, Indonesia, and Unzen, Japan (Francis, 1994, pp. 250–253), to eruption column collapse incorporating hot juvenile blocks, such as the 1968 eruption of Mayon (Moore and Melson, 1969). The deposits from all of these pyroclastic flows are generally very poorly sorted, chaotic, and unstratified; these deposits are now termed *block-and-ash flows* because of the distinct particle sizes involved.

Surges. The precise origin of pyroclastic surges remains rather controversial, but their distinctive characteristics include reduced bulk density (relative to ignimbrites and block-and-ash flows) and evidence of turbulent rather than laminar emplacement regimes (Francis, 1994, p. 236) Surges are usually associated with hydromagmatic eruptions, where the interplay of nonjuvenile water and magma is dominant and which leads to a high degree of fragmentation (Sheridan and Wohletz, 1983; Wohletz and Heiken, 1992, pp. 26–37). Surges are also identified with the initiation of many ignimbrites, and they may form crucial parts of the column collapses involved in the AD 79 eruption of Vesuvius (Sheridan *et al.*, 1981; Sigurdsson *et al.*, 1985). Surge deposits often include sections with sand waves and climbing dunes, indicative of emplacement by saltation (bouncing along the ground) rather than full suspension (Sigurdsson *et al.*, 1987), which is directly related to the intensity of energy release associated with a hydromagmatic eruption (Sheridan and Wohletz, 1983).

2.5.3. Fountain-Fed Rhyolitic Eruptions

Fire-fountaining is commonly considered an eruptive mode confined to basalts and other mafic magma compositions with relatively low viscosities, but an increasing number of rhyolitic units are interpreted to have formed by fountaining. They range from small welded tephra rings through moderately sized secondary lava flows to widespread welded airfall tuffs that blanket any preeruptive topography (Christiansen and Blank, 1975; Duffield, 1990; Manley, 1994; Christiansen, in press; Manley and McIntosh, in press), and are known primarily from Yellowstone National Park in Wyoming, southwestern New Mexico, and southwestern Idaho, all in the United States. The fountaining mode of eruption in these units is facilitated not by low viscosities caused by high temperatures, but rather by the behavior of the erupting magma before and during disruption in the conduit. These units all seem to have preeruptive water contents lower than those erupted in a Plinian manner, and many have high fluorine contents. Rhyolitic magma with a preeruptive water content of less than about 3 wt% (Manley, 1994, 1996) becomes less comminuted and produces a much smaller proportion of ash-sized tephra than those with more water. This results in a low-bulk-density eruptive column; entrainment of air is minimized, as is heat loss, and the clasts are emplaced ballistically, as in a basaltic fire-fountain. Because more of the tephra lands nearer the vent and remains hot enough to weld, a welded airfall deposit forms. High magmatic fluorine contents of up to 3 wt% (Webster and Duffield, 1991) may play an important role by decreasing the viscosity (Dingwell *et al.*, 1985) of the rather dry magma sufficiently that it is still able to rise through the conduit and erupt.

2.5.4. Modification of Ash Deposits

Ash deposits can generate distinctive geomorphic terrains. Large ignimbrite deposits fill in preexisting topographic lows to produce remarkably flat units that can bury valleys and embay the surrounding hills, often with only a gradual slope extending away from the vent area. Individual pyroclastic flows can have a planform similar to that of lava flows but which can demonstrate remarkable fluidity in detail in skirting around topographic obstacles. If remobilized by water, ash can produce enormous mudflows (identified by the Philippine word *lahar*) capable of doing considerable geomorphic work and moving at rapid speeds when traveling over steep slopes. Remobilized ash is often difficult to distinguish from a traditional mudflow without close examination of the deposits.

The erosional characteristics of ash deposits also are not uniquely diagnostic of a pyroclastic origin, but they do highlight variations in consolidation that may be difficult to duplicate with other processes. Thick ignimbrites usually show a vertical progression of alteration and competency that results from the slow cooling of such deposits (see Section 2.4.2). A densely welded zone can form near the middle of the deposit, which can make a prominent cliff-forming unit as compared with the overlying alteration zone (where vapor-phase interactions are important) and the underlying weakly consolidated ash, both of which rapidly weather. Large ignimbrites in the Andes are locally sculpted into immense yardang fields where wind and sand supply are sufficient, as along the elevated Altiplano in Bolivia and Chile (de Silva and Francis, 1991). However, yardangs themselves are not diagnostic of ignimbrites, only of a material strong enough to hold steep slopes but weak enough to be scoured by aeolian processes (see Chapter 4 for a discussion of yardang fields associated with hypothesized ignimbrite deposits on Mars).

2.6. LARGE IGNEOUS PROVINCES

2.6.1. Distribution

LIPs represent some of the largest volcanic features on Earth, both in terms of areal coverage and total volume of lava (Table 2.1). LIPs occur across the entire globe (Figure 2.9); several are located on the ocean floor, including the Ontong Java Plateau, which at $\sim 3.6 \times 10^7$ km^3 is the largest LIP on Earth (Coffin and Eldholm, 1993). The most intensely studied LIPs (to date) are found on continents, such as the CRB group in the Pacific Northwest, USA, and the Deccan Traps in western India. The diverse settings for LIPs result in considerable variability in what is known about a given site, yet to date an impressive global data set has been compiled for various LIPs (Macdougall, 1988; Mahoney and Coffin, 1997).

Why do such voluminous outpourings of lava occur in these locations? There are several hypotheses, numbering almost as many as there are identified LIPs. The most widely cited explanation ties together both a mantle plume and the initiation of plate rifting. White and McKenzie (1989) relate the Paraná basalts in southeastern South America (see Renne *et al.*, 1992) and the Etendeka basalts in southwestern Africa (see Renne *et al.*, 1996) to continental rifting and the opening of the south Atlantic. Rifting was associated with the buoyant rise of the mantle into the thinned lithosphere, which led to decompression melting that generated copious quantities of basalt magma. This process may have contributed to the opening of the south Atlantic as well as the formation of a stable mantle plume whose volcanic products can be traced from the South American and African basalts to the Tristan island group, which

Table 2.1. Large Igneous Provinces[a] Emplaced in the Past 300 Myr

Name	Province	Age (Ma)	Area (10^6 km^2)
Snake River	Western United States (SRP)	0.002–15	0.3
Columbia River	Western United States (CRB)	6–17	0.2
Iceland	North Atlantic Tertiary Basalts	7–20	0.04
Traps	Ethiopia	20–40	0.8
Britain/Greenland	North Atlantic Tertiary Basalts	45–65	0.7
Deccan	India	60–67	0.5
Caribbean	Caribbean	70–93	1.0
Ontong Java	Pacific Ocean	90 & 122	1.5
Kerguelen	Southern Indian Ocean	90–115	0.7
Paraná	Brazil	115–135	1.2
Manihiki	Pacific Ocean	115–125	0.7
Karoo	South Africa	110–200	1.4
Patagonia	Argentina	190–210	0.7
Siberia	Russia	250–280	1.5

[a] Information from Coffin and Eldholm (1993) and Mursky (1996).

represents the recent output from the plume source (Ewart *et al.*, 1998) (see also Section 2.3.4). Similar conclusions can be drawn for other LIPs such as those around the north Atlantic, but with volcanic tracks less obvious than the Paraná–Tristan–Etendeka volcanics. Interested readers are referred to Mahoney and Coffin (1997) for a recent compilation of both observations and hypotheses for many LIPs.

2.6.2. Emplacement

There is considerable debate concerning how LIPs were emplaced. Favored mechanisms can be separated into diametrically opposed camps: rapid emplacement of fast-moving flows and slow emplacement with progressive inflation of the individual flows. The rapid

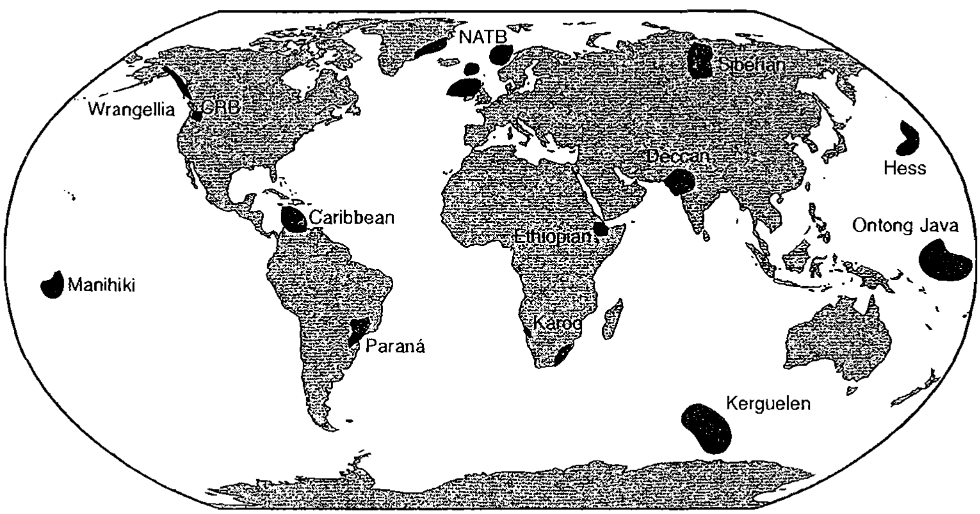


Figure 2.9. Large Igneous Provinces (black), present on both continental (gray) and oceanic (white) crust. See Table 2.1 for information on selected examples.

emplacement scenario was pioneered by Shaw and Swanson (1970), who made the first quantitative estimates of CRB emplacement by relating steady effusion along feeder dikes to turbulently flowing lava powered by the hydraulic head resulting from the gentle slope of the lava surface. Their calculations led to emplacement times for individual CRB flows of days to weeks, an interpretation favored by some more recent investigations (Swanson *et al.*, 1989, pp. 21–26; Tolan *et al.*, 1989; Reidel and Tolan, 1992). In contrast, recent observations of inflated basaltic flows in Hawaii provide a mechanism for slow emplacement of what eventually becomes a thick flow (Walker, 1991; Hon *et al.*, 1994). Inflation features recently have been tentatively identified in CRB flows at several locations, suggesting that individual CRB flows may have been emplaced over years to decades (Self *et al.*, 1996, 1997). An advantage of the slow emplacement mechanism is that required effusion rates are comparable to documented effusion at Hawaii and Laki (see Section 2.4.3), rather than invoking effusion rates that are orders of magnitude larger than any documented eruption. One recent model attempts to incorporate elements of both fast and slow emplacement to explain the CRB (Reidel, 1998).

The inflation mechanism has applications to flows other than just those of basalts or LIPs. Ultramafic komatiite flows have compositions suggestive of very fluid lava (see Chapter 8), but komatiites typically are quite old and preserved evidence of flow emplacement is often difficult to find. Inflation has recently been proposed as a way to reconcile known komatiite characteristics with models for the generation and eruption of ultramafic lavas (Cas and Self, 1998). Clearly all investigators need to look for details that might test the inflation hypothesis, not only in basaltic terrains, but also with more exotic lavas.

2.7. DISCUSSION: VOLCANO/ENVIRONMENT INTERACTIONS

2.7.1. Physical Interaction between Volcanic Products and Earth's Environment

As volcanic materials leave the vent, they will immediately interact with whatever environment is surrounding the volcano. This interaction can take the form of generation of a solid crust, mechanical degradation of solidified materials, or even agradation of fine particles (e.g., accretionary lapilli). Here we will not explore all of the variety of ways this interaction manifests itself, because that is one of the main goals of this book. However, it may be helpful to discuss briefly some of the physical consequences of fresh volcanic materials erupted subaerially on Earth (subaqueous eruptions are discussed in Chapter 5).

The tensile strength of a geologic material is a fundamental physical property that affects not only possible tectonic deformation but also how the material responds to sudden environmental change. Without question, the abrupt transition from conditions within the vent to those at the Earth's surface will be the most dramatic change the magma experiences. Physical properties have been measured for numerous igneous and volcanic rocks (e.g., Shaw *et al.*, 1968; Murase and McBirney, 1973; Robertson and Peck, 1974; Horai, 1991), but the important issue here is the total magnitude of the temperature contrast experienced by a volcanic product and the rate at which the thermal change is imposed. In both effusive and explosive eruptions, the temperature contrast will be very large (up to 1000°C) and rapidly applied (seconds). This combination exerts a substantial mechanical stress on the cooling material.

For effusive eruptions, the temperature contrast and the thermal conductivity of the lava allow crustal growth to be modeled (Crisp and Baloga, 1990; Kilburn and Lopes, 1991;

Manley, 1992; Kilburn, 1993). Crusts on slow-moving pahoehoe flows allow the discrete "toes" (~10 to 30 cm in width) typical of pahoehoe to form and grow at a rate comparable to the average rate of flow advance. This newly formed glassy crust remains sufficiently heated (by the still-molten lava within the flow) to deform plastically, so that intense fracturing of the pahoehoe surface typically does not occur during emplacement. Faster-moving a'a flows disrupt the surface of the active lava channel into centimeter-scale clinkers that aid somewhat in insulating the flowing lava beneath them. However, gaps between solidified portions of an active flow surface, produced either by thermal stress or by the shearing action of the flow, allow incandescent lava to radiate energy more efficiently than does the cooled crust, which can significantly affect the cooling history of the flow (Crisp and Baloga, 1990). When a flow section comes to rest, thermal stresses can still contribute to pervasive fracturing of the outer sections of the flow.

Which is more important: the physical conditions of the magma at eruption, or the external environment into which the eruption takes place? It is unlikely that any one answer will apply to all volcanic eruptions, especially when the discussion extends to the plethora of environmental conditions found throughout the solar system. Instead, the combination of lava properties and the environmental conditions will be playing against each other to constrain the condition of the final volcanic product. In each individual situation, it is important to consider this dynamic interplay of material properties and environment when evaluating volcanic features (see Chapter 9).

2.7.2. Volcanic Effects on the Atmosphere and Climate

Historic eruptions provide the opportunity to look for potential links between volcanism and environmental change (Francis, 1994, pp. 368–379). Probably the first documented case of making the connection between a volcanic eruption and unusual atmospheric conditions is a report by Benjamin Franklin (1784), describing several weather anomalies from summer through winter that he conjectured may have been related to an eruption reported in Iceland. Although Franklin did not know specifics about the eruption, it was in fact the 8-month-long Laki fissure eruption whose consequences were devastating for Iceland and very serious elsewhere in Europe. Recent analysis indicates this eruption released massive quantities of SO_2 and other caustic gases (Thordarson and Self, 1993), which turned into an acid rain and haze in Scandinavia and damaged much of the fall harvest. The consequences were worst in Iceland, where the acid haze demolished the summer crops, resulting in famine that led to the death of 75% of Iceland's livestock and 25% of the human population (Thordarson and Self, 1993).

The massive Krakatau eruption of 1883 also resulted in enormous loss of life (primarily through tsunamis). The estimated 20 km^3 of erupted pyroclastic material, in association with stratospheric droplets condensed from volcanic gases, decreased the intensity of sunlight reaching Earth's surface and resulted in worldwide unusual optical effects (e.g., dramatic sunsets and solar rings). Available evidence supports a cooling in the Northern Hemisphere of 0.25°C for 1 to 2 years (Rampino and Self, 1982). The even larger 1815 eruption of Tambora released >50 km^3 of magma in a series of explosions from April 5 to 10 that produced large Plinian columns, massive pyroclastic flows, and co-ignimbrite ash (Sigurdsson and Carey, 1989). Unusual optical effects also accompanied this eruption. More importantly it produced an observable decrease in both sunlight and starlight around the world, which when combined with documented weather changes (e.g., the "year without a summer" in the Northern

Hemisphere) make a strong case for volcanically induced climate change of more than 2 years' duration (Stommel and Stommel, 1983; Stothers, 1984).

A better understanding of the mechanism behind these volcanic environmental consequences resulted from two more recent eruptions in Mexico and the Philippines. Several Plinian columns and associated pyroclastic surges erupted from El Chichón volcano in Mexico between March 29 and April 4, 1982, releasing ~ 1 km^3 of magma with severe consequences to nearby villages (Sigurdsson *et al.*, 1987). The climatic significance of this eruption was that the tephra contained up to 2 wt% sulfates (including anhydrite crystals), resulting in an eruption cloud unusually rich in sulfuric acid aerosols that quickly encircled the globe, as monitored by orbiting satellites (Rampino and Self, 1984). The importance of sulfur-rich eruptions was highlighted by the larger eruption of Mt. Pinatubo in the Philippines on June 15–16, 1991, which released 3 km^3 of magma (Wolfe and Hoblitt, 1996), and which also contained abundant anhydrite and sulfur-rich gases (Bluth *et al.*, 1992; Hattori, 1996). Sulfur-rich gas, once in the stratosphere, condenses into sulfuric acid droplets that are the primary long-term agents of global cooling. In this situation, volcanic eruptions that place a sufficient volume of the right materials into the stratosphere might affect the environment more than the eruption itself is affected by its surrounding environment.

2.8. CONCLUSION

In this chapter we have reviewed the considerable diversity of volcanic products emplaced within a subaerial environment on Earth. Field observations and theoretical developments have provided a good indication of what happens to magma during its ascent to the Earth's surface. Plate tectonics provides the unifying theory through which the wide range of volcanic styles and locations can be understood. Volcanic materials are derived from both effusive and explosive eruptions, resulting in a wide range of volcanic constructs and deposits. Massive outpourings of basaltic lavas occur in LIPs distributed throughout the world, but most are likely associated with mantle plumes and the initiation of plate rifting. All of the volcanic materials described here have interacted with the subaerial environment found at Earth's surface, and in some cases the environment itself may have been influenced by the eruptions. This information represents the basis against which we can evaluate volcanic eruptions and associated products encountered throughout the rest of the solar system, as discussed in the remainder of this book.

2.9. REFERENCES

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