4.12 Volcanology 101 for Seismologists

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Glossary

dike A tabular igneous intrusion that cuts across planar structures of surrounding rock, often cutting vertically through horizontal strata.

eruption Extrusion or explosive ejection of volcanic material onto the Earth's surface or into the atmosphere – (1) phreatic eruption: steam blast, without new magma; (2) magmatic eruption: eruption purely of magma; and (3) phreatomagmatic eruption: hybrid of two. Many eruptions progress from phreatic through phreatomagmatic to magmatic, and some revert back to phreatomagmatic in their late stages as groundwater regains access to magma.

felsic Adjective describing minerals and rocks rich in silica and aluminum and poor in iron and magnesium. Felsic rocks include dacite, rhyolite, their intrusive equivalents of granodiorite and granite, and a variety of less common rock types with higher or lower contents of Na and K.

groundmass Fine-grained mixture of glass and microlites, forming the interstitial matrix of volcanic rocks containing phenocrysts.

mafic Adjective describing minerals and rocks rich in iron and magnesium, and relatively poor in silica and aluminum. Mafic rocks include basalt, its intrusive equivalent of gabbro, and a variety of less common rock types with higher or lower contents of Na and K. Andesite and its intrusive equivalent, diorite, with slightly higher silica content, are intermediate between mafic and felsic.

magma Molten rock, consisting of up to three phases – melt, crystals, and exsolved volatiles (bubbles).

magma mixing (and mingling) Mixing is mechanical and chemical homogenization of two magmas. Mingling is incomplete mixing and is seen in 'marble-cake' banded rocks.

microlite Very small crystal in volcanic rock, typically <<1 mm in longest dimension, formed in volcanic conduits upon degassing or cooling during or shortly before eruption.

oxygen fugacity (abbreviated f_{O_2}) A measure of the oxidizing or reducing character of a fluid. More precisely, the chemical activity of oxygen in a chemical solution, that is, the amount of oxygen available for chemical reactions. Typically expressed in log units, absolute or relative to the amount of free oxygen associated with an assemblage of minerals, for example, QFM, quartz (SiO_2) + fayalite (Fe_2SiO_2) + magnetite (Fe_3O_4).

petrologic method Estimation of pre-eruption volatile contents from the highest volatile concentrations trapped in glass inclusions within phenocrysts, and of degassing based on the difference between volatile concentrations in glass inclusions minus concentrations in matrix (= groundmass) glass. A variation of the petrologic method approximates the concentration of volatiles in magma as 100 wt.% minus the analytical total of nonvolatile components.

phenocryst Crystal in igneous rock that is notably larger than the average grain size of that rock, probably because it formed over extended periods within magma reservoirs several kilometers or more beneath the surface. No specific size definition, but almost always >1 mm in length.

pumice Rapidly quenched ('frozen') magma foam. Low-density, full of vesicles (bubble holes). Density often <1 $g\,cm^{-3}$ so it floats in water, in contrast to nonvesicular rock densities of approx 2.5 $g\,cm^{-2}$.

scoria ('cinder') Volcanic rock with high vesicularity but less than that of pumice. Although there is a tendency for scoria to be mafic and pumice felsic, composition is not formally a part of the definition of pumice or scoria.
second boiling Process whereby crystallization concentrates volatiles in the remaining melt phase and eventually causes them to exsolve into bubbles, mimicking a boiling process even though temperature may be constant or decreasing.
sill A tabular igneous intrusion that parallels planar structures of surrounding rock, such as horizontal strata.
ultramafic rocks Rocks containing low silica contents and crystallizing large percentages of mafic minerals, for example, olivine, pyroxenes, and amphiboles. Includes peridotite, eclogite, and a variety of less common combinations of mafic minerals.
unrest An anomalous change from the background level of seismicity, ground deformation, gas emission, thermal output, and other measured parameters of a volcano. Some unrest leads to eruptions but much does not, so we use the term 'unrest' rather than 'precursor' unless the temporal or causal association with an eruption is clear. Eruptions are included within the definition of unrest.
volcano A vent or cluster of vents from which molten or solid debris is erupted. Most volcanoes build edifices from erupted lava and fragmental debris; some that erupt very large volumes of magma create depressions in the ground, as in the case of calderas.

4.12.1 Introduction to Volcanic Systems

Active volcanoes are dynamic systems that can be instructively compared to active fault zones. Both are subject to stress and strain, both produce distinctive patterns of seismicity and, in a sense, both exhibit stick-slip behavior with similar power-law magnitude–frequency relations. The distribution of areas at risk from active volcanoes and fault zones can be mapped and future hazards forecast over decades, centuries, or longer. However, beyond the similarities, volcanic systems involve high-temperature silicate melts, significant volumes of dissolved and undissolved volatiles, and important mass transfer before eruptions. Volcanic eruptions have far more pronounced precursors than do tectonic earthquakes, and, accordingly, in addition to long-term forecasts, both hour-day and week-month scale forecasts are often possible. Some of the precursors to eruptions are similar to those sought for earthquakes (e.g., precursory creep and microseismicity) while others are unique on account of the magma (e.g., leaks of high temperature gas).

Some volcanoes are sensitive to static or dynamic tectonic strain, for example, Long Valley and Yellowstone calderas (Hill *et al.*, 2002), and, conversely, magmatic intrusions and hydrothermal pressurization can induce nearby tectonic seismicity (White and Power, 2001; Legrand *et al.*, 2002).

This chapter describes the fundamental matter of volcanoes – molten rock (magma), which is composed of melt (typically silicate melt), crystals (dominantly, feldspars, quartz, and Fe–Mg silicates), and volatiles (mostly water). The volatile species are dissolved or occur as bubbles in the melt. Although magma is generated in the upper mantle and lower crust, this chapter is concerned mainly with its ascent from relatively shallow crustal reservoirs to the surface, and associated pressurization of magma and surrounding groundwater. Volcanic unrest – volcanic seismicity, ground deformation, gas emission, and other changes – reflects intrusion, pressurization, and pre-eruption degassing of magma, and forms the basis for short-term eruption forecasts. Finally, the chapter notes briefly different eruptive phenomena, posteruption erosion of volcanoes, and future directions in volcanology.

Volcanic eruptions range from a few m^3 to thousands of km^3 of magma, minutes to decades in duration, and gently effusive to astoundingly explosive. One important control on the explosivity of an eruption appears to be the 'leakiness' of the system, that is, how much of its magma's gas has leaked out since the previous eruption and before soon-to-erupt magma reaches the surface.

Worldwide, more than 500 million people live within zones that could be seriously affected by volcanic eruptions (Tilling, 1991; Ewert and Harpel, 2004). Millions more fly over or downwind from volcanoes every year. The average annual death toll from volcanic activity between AD 1800 and 2000 was ~1000 yr^{-1}. That average is dominated by just a few particularly sad disasters, such as the eruptions of Montagne Pelée in 1902 and of Nevado del Ruiz in 1985. During the past four decades, roughly 1.5 million people on the ground have been evacuated out of harm's way, and hundreds of thousands of air passengers, possibly more, have been diverted away from dangerous ash clouds. Populations living on and near volcanoes are ever increasing, and volcanologists today are faced with rising expectations for accurate and timely eruption forecasts to keep these populations safe.

Although a review, this chapter shows some biases on the way volcanoes work. Volcanologists including the author are strongly influenced by the volcanoes we have known. This chapter should be considered as a set of working hypotheses, judged likely today but subject to revision tomorrow. What follows is by no means comprehensive. Sigurdsson *et al.* (2000), Francis and Oppenheimer (2004), and Schmincke (2004) are excellent references on many more aspects of volcanology, including volcanic rocks and features, and Dobran (2001) gives mathematical descriptions of many volcanic processes.

4.12.2 Magma and Gas

4.12.2.1 Melt

Molten silicate, or in rare cases molten carbonate, is generated in the upper mantle beneath spreading centers and hot spots, in mantle wedges of arcs, and within the crust. Most mantle melts are basaltic in composition, commonly distinguished by their variable alkali (Na and K) contents. They originate by variable degrees (usually <20%) partial melting of mantle peridotite. Crustal melts tend to be much more silicic, rhyolite or dacite, produced by <20% and often <10% partial melting of (mostly lower) crust or, in some cases, by fractionation of mantle-derived basalt. The percentage of partial melt present in a source region at any given time is considerably less, probably not more than a few percent. Mafic melts can have temperatures up to 1200°C; rare ultramafic melts can be up to 1400°C. Silicic melts can exist down to about 700°C.

The term magma (molten rock) encompasses not only the molten silicate but also any crystals that have formed or remain from incomplete melting, and volatiles that are either dissolved within the silicate melt or occur as bubbles in that melt (**Figure 1**).

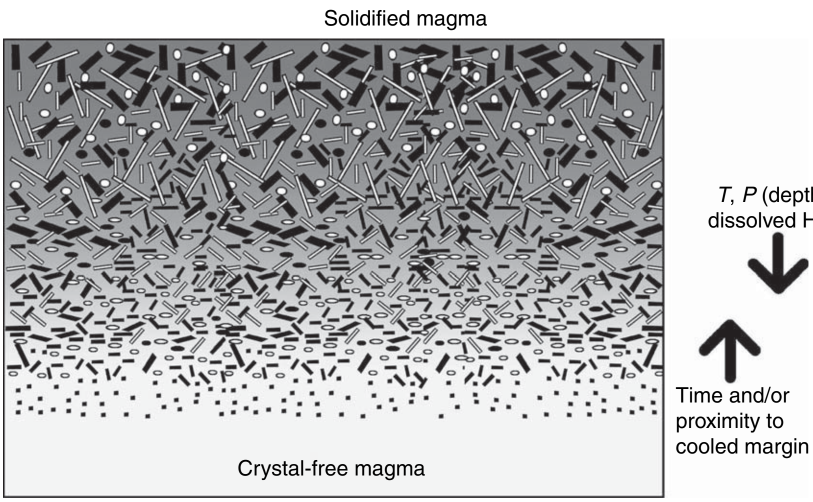

Figure 1 Components and evolution of magma, from melt to solid. Initially, at high temperature and pressure, only melt is present. As temperature declines, crystals will begin to form. As lithostatic pressure declines, water and other volatiles will also begin to exsolved into bubbles, inducing further crystallization. The sequence shown here from bottom to top, symbolizing change as magma moves from depth to the surface, is also repeated from center to margin of dikes and other intrusions. After Marsh BD (2000) Magma chambers. In: Sigurdsson H, Houghton BF, McNutt SR, Rymer H, and Stix J (eds.) *Encyclopedia of Volcanoes*, pp. 191–206. San Diego, CA: Academic Press.

Magmas can further differentiate by fractional crystallization in which early crystallizing phases and remaining melt are separated by crystal-settling or/or by filter-pressing melt out of crystal-rich mush, by assimilation (digestion) of country rock, and by magma mixing.

4.12.2.2 Crystals

Crystals form in melt as it cools or as it loses dissolved volatiles. Decreasing temperature and loss of dissolved water lead to polymerization of silica tetrahedra (SiO_4) in the melt that are the building blocks of common volcanic minerals like quartz (SiO_2), plagioclase feldspars (Ca/Na–Al silicates), alkali feldspars (Na–K–Al silicate), olivine (Mg/Fe silicate), pyroxenes (Ca/Mg/Fe silicates), hornblende and other amphiboles (hydrous Ca/Na/K–Mg/Fe/Al silicates), biotite and other micas (hydrous K/Na–Mg/Fe/Al/Ti silicates) (slash indicates solid solution between these end members, for example, between Ca and Na in plagioclase feldspar). Other important volcanic minerals, small in volume but very useful in tracking petrologic processes, are the Fe–Ti oxide minerals which can be used to determine the temperature and oxygen fugacity of magma.

Crystals occur in two broad size groups, phenocrysts (typically, 1–10 mm) and groundmass crystals (microlites, typically <<1 mm). Phenocrysts form while magma is still in its reservoir, whereas microlites grow during final ascent and eruption of magma.

Volcanic rocks rarely contain more than 60% phenocrysts (Marsh, 1981). The reason is that higher phenocryst contents make magmas too viscous to erupt. The balance, 40%, is melt in the reservoir but can grow a significant percentage of microlites shortly before and during eruptions.

4.12.2.3 Volatile Phases – Both Dissolved and Exsolved

4.12.2.3.1 Gas supply

Volatile components of magma, dominantly H_2O, CO_2, sulfur gases SO_2 and H_2S, halogen gases HCl and HF, hydrogen, and helium in decreasing order of abundance, are incorporated into magma during initial melting of mantle or crust. Water content in magma can range from <0.1 wt.% up to ~8 wt.%. Carbon dioxide content can be as high as 1 wt.% (10 000 ppm) but is usually a fraction of this. Sulfur concentrations (expressed as elemental S) are typically 10–1000 ppm, and occasionally several times higher. Beneath arcs, additional water, and other volatiles come from subducted sediment and from metamorphism of downgoing, hydrothermally altered oceanic crust to eclogite. That water is released upward into the mantle wedge, where it fluxes melting and gets incorporated into magmas.

4.12.2.3.2 Degassing

There are limits to volatile solubility in silicate melts. The limits depend mostly strongly on pressure; the composition, temperature, and oxygen fugacity of the silicate melt; and the specific volatile species in question. Volatiles will form bubbles if they are added to or concentrated in melt in excess of their solubility. One way to add volatiles to magma is by convective and diffusive transfer from fresh magma entering the base of a magma reservoir. Another very common scenario of volatile concentration is called 'second-boiling', in which volatiles, excluded from most crystal structures, are concentrated by magma crystallization into the remaining melt fraction, and eventually exceed their solubilities in that melt. Temperature may be constant or even falling, yet volatiles are 'boiling', exsolving from the melt into bubbles.

Bubbles will be of lower density than their host melt and can rise buoyantly through that melt. Large bubbles (sometimes, up to several meters in diameter) are seen to rise up through fluid magma in Hawaii, at Mount Erebus in Antarctica, and elsewhere. This process ('separated flow') is an effective way to degas fluid magmas (Vergniolle and Jaupart, 1986). An everyday analogy is the bottle of soda pop or champagne from which bubbles rise freely once the confining (lithostatic) pressure is released. Eventually, the soda pop and magma 'go flat' unless there is resupply of volatiles from below. Degassing in viscous magmas is not as well understood. Clearly gases do escape – we can measure them at the surface and we also know that silicic magmas can 'go flat'. However, in viscous magmas simple Stokes' law bubble rise is too slow to explain observed gas fluxes. Three alternate processes may be at play. One is convective transport of gas-rich magma up to a level at which it can foam, release its gas, and then sink in return flow as dense, now-degassed silicate melt and crystals. One can observe this process directly in lava lakes, and it is inferred at many other volcanoes for which the gas flux far exceeds what could have been dissolved in the volume of magma erupted over that same period (e.g., Allard, 1977; Allard *et al.*, 1994; Witter *et al.*, 2004; 2005). A second mechanism by which gas can escape from

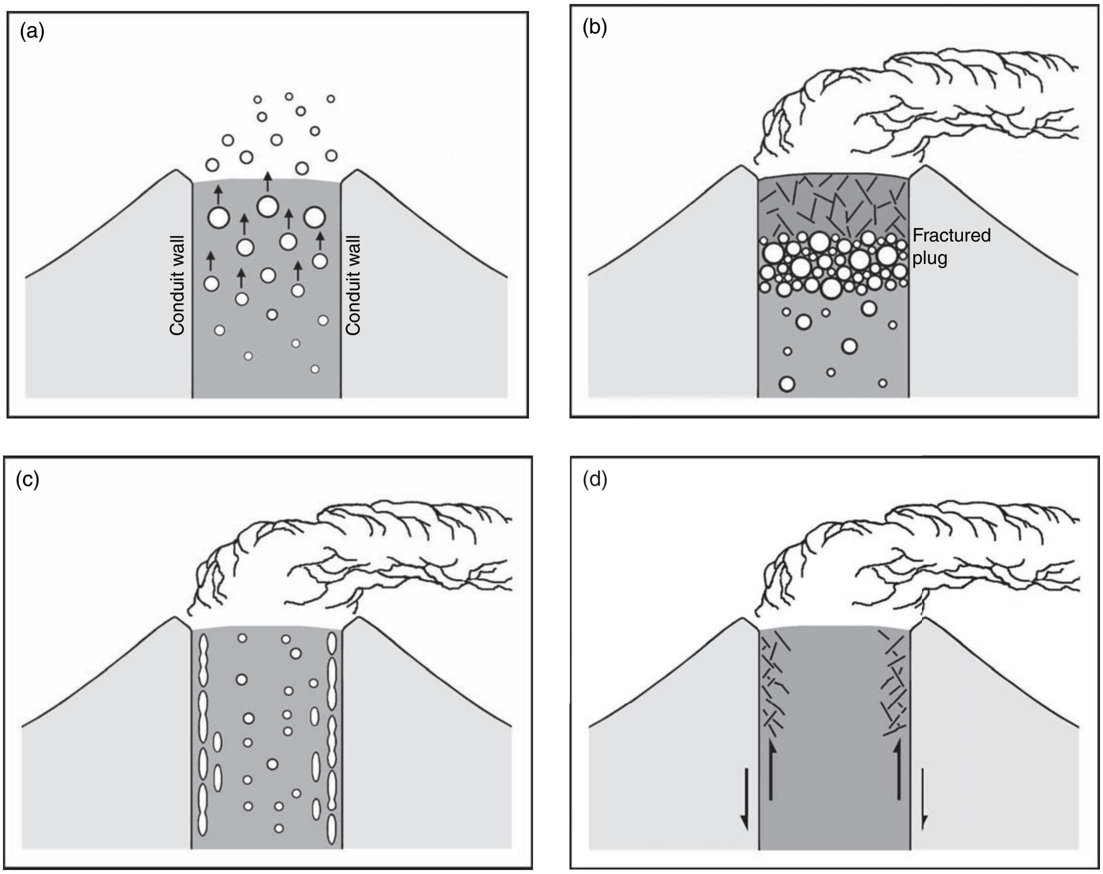

Figure 2 Various modes of gas permeability in magma: (a) separated flow of bubbles through melt; (b) development of a permeable foam at top of magma column; (c) dynamic shear-induced permeability (from bubbles and melt being stretched); and (d) shear-induced fractures along the margins of a volcanic conduit.

viscous magma is by diffusion or bubble transport over relatively short distances to the edge of magma conduits, followed by escape into high-permeability breccias that form along the margins of some conduits by shear (Jaupart and Allègre, 1991; Stasiuk *et al.*, 1996; Rust *et al.*, 2003; Burgisser and Gardner, 2005). Third, some dynamic bubble-to-bubble permeability may also develop within viscous magma as bubbles are brought into contact and interconnected by shear during magma ascent. Various modes of degassing are illustrated in **Figure 2**.

4.12.2.3.3 Gas budgets

The balance between (1) gas influx into crustal magma reservoirs (e.g., arriving dissolved in fresh basaltic magma), (2) continuous, passive degassing into the atmosphere and into hydrothermal systems, and (3) degassing during explosive eruptions is an important measure of the behavior of a volcano (**Figure 3**).

The gas influx term can be roughly estimated from a volume rate of magma supply (from the mantle and lower crust up into a shallow magma reservoir) times the concentration of each volatile species in that magma. A minimum estimate for volume rate of magma supply can come from the long-term eruption rate for that volcano; a better estimate that includes magma which does not erupt (intrusions) can come from long-term growth rate for a whole volcanic edifice or volcanic arc. If magma supply is relatively continuous, these long-term rates may also apply to shorter time intervals. Typical magma supply rates for arc volcanoes are $\sim10^7$ $m^3\,yr^{-1}$, plus or minus an order of magnitude; rates for ocean island and hot-spot volcanoes can be $\sim10^8$ $m^3\,yr^{-1}$ plus or minus.

If one measures gases that are not common in the atmosphere, for example, SO_2, one can quantify passive degassing of magma between times of eruptions. Unfortunately, passive gas flux has been estimated

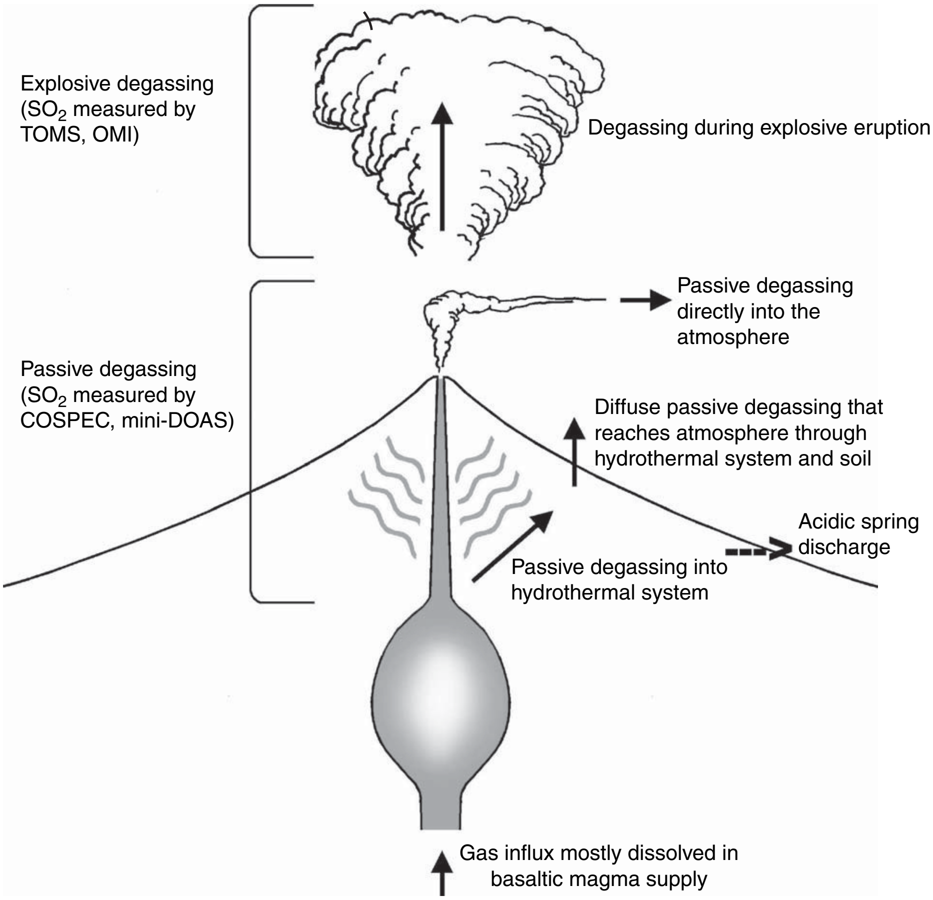

Figure 3 Elements of a simple volcanic gas budget, including input from depth, loss into a hydrothermal system, acidic spring discharge, passive loss to the atmosphere, and explosive loss to the atmosphere.

consistently at only a handful of volcanoes. Elsewhere, measurements have been too infrequent to give reliable estimates. A promising development for systematic, space-based monitoring of passive degassing of SO_2, down to levels of hundreds of tonnes/day, is the new ozone mapping instrument (OMI) which began collecting data in 2004 (Krotkov *et al.*, 2003).

A final term in a gas budget is the mass of gas released during eruptions. Again using SO_2 (and estimating other species in relation to SO_2), the best data come from NASA's total ozone mapping spectrometer (TOMS). Data from the largest eruptions have been gathered since 1977 (Bluth *et al.*, 1997; Carn *et al.*, 2003). OMI data will supplement those from TOMS.

Of the three main terms in a gas budget – input, passive degassing, and eruptive degassing – the input term is the most poorly known. As a result, it is difficult to know the degree to which magma has already degassed before eruption. Yet, pre-eruption degassing may be the key to estimating explosive potential of an impending eruption.

One important insight that is appearing from gas budgets is that most volcanoes lose most of their gas passively, between eruptions. Degassing during explosive eruptions is spectacular, but less noticed, steady leakage dominates discharge. If we do a thought exercise by assuming concentrations of 1000 and 2500 ppm S in $10^7\ m^3\ yr^{-1}$ of fresh mafic magma supply to typically explosive arc volcanoes, multiply by the number of years since the last gas-depleting VEI $\geq$ 3 explosive eruption, convert from S to SO_2, and compare to the SO_2 release of the next VEI $\geq$ 3 eruption (Carn *et al.*, 2003), we find that for most arc volcanoes explosive gas release is <10% of inferred gas input. The difference must be released passively. This has interesting implications for conduit dynamics between eruptions as noted in Section 4.12.3.3, and for explosive potential as noted in Section 4.12.6.5.2.

Volcanoes in Hawaii are classic examples of volcanoes that degas passively. Even within arcs, for the above-mentioned thought exercise, there are sharp contrasts between volcanoes that release almost all of their gas passively (e.g., Mayon, **Figure 4(a)**) and those that retain a substantial percentage of their incoming gas and then release it in large explosive eruptions (e.g., Pinatubo, **Figure 4(b)**) (Newhall, 2004).

Another important insight that has come from gas budgets is that some magmas appear to accumulate volatiles well in excess of saturation, with the excess volatiles accumulating in discrete fluid bubbles (Luhr *et al.*, 1984; Gerlach *et al.*, 1996; Keppler, 1999; Wallace and Anderson, 2000; Wallace 2001). Petrologic comparison of volatile contents in undegassed glass inclusions that were trapped long before eruptions in early formed phenocrysts, versus degassed matrix glass, gives a minimum measure of how much degassing has occurred (Anderson, 1974). For example, undegassed glass inclusions might contain a few hundred or even a few thousand ppm S, whereas matrix glass might contain only a few tens of ppm S; the difference in concentration is attributed to degassing before eruption and multiplying this concentration difference by the volume of magma erupted gives a minimum mass of pre-eruption degassing. When the amount of SO_2 released during large explosive eruptions (as measured by TOMS) is compared to that which would have been predicted from the petrologic method, the former may be 10–100× higher than the latter, and, indeed, much greater than could ever have been dissolved in the volume of magma erupted. Since this gas is erupted all at once it must be assumed to have resided within the magma, but it must have been in discrete fluid bubbles well in excess of saturation in the melt. (Using the petrologic method on products of magmas that had discrete pre-eruption volatile bubbles at depth, one often observes just a small difference between S in glass inclusions and S in the matrix glass, implying minimal pre-eruption degassing. In fact, though, a great deal of degassing from the melt already occurred at depth – 'into' the discrete bubble phase. As soon as CO_2-rich bubbles begin to form at depth, other volatiles including H_2O and SO_2 move preferentially into the volatile phase rather than remain dissolved in the melt.)

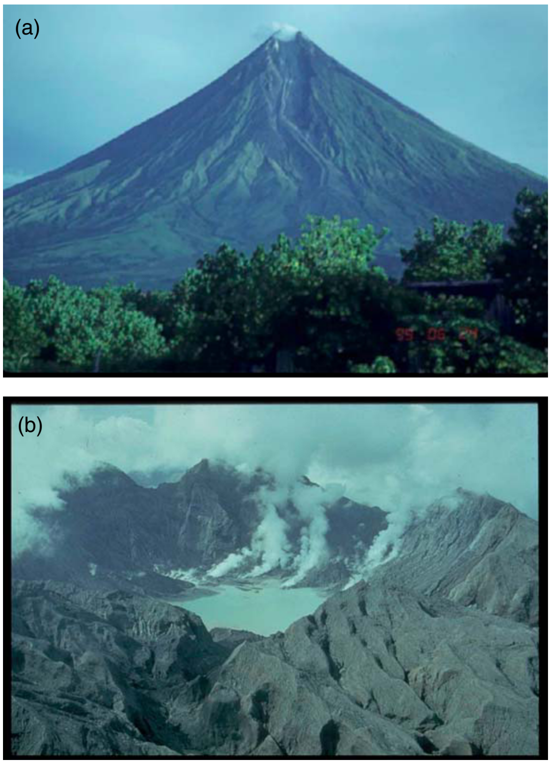

Figure 4 Two arc volcanoes with sharply different patterns of degassing. (a) Mayon, Philippines, which since 1978 has released ~95% of its gas passively and only ~5% explosively. Only 50 kt of SO_2 was recorded by the TOMS sensor, during the explosion of 1984. (b) Pinatubo, Philippines, whose conduit was plugged for approximately 500 years so ~17 Mt of SO_2 (~65% of incoming SO_2) accumulated in the magma and was released in a large explosive eruption in 1991; the remaining 35% was either released passively into the pre-1991 hydrothermal system or remains in the magma. (Assumptions for both volcanoes: continuous deep supply of 10^7 m^3 basaltic magma per year, 1000 ppm S. TOMS data are from Bluth *et al.* (1997) and Carn *et al.* (2003).

4.12.2.4 Viscosity, Density, and Volatile Content

Melt viscosity varies dramatically with magma composition and with cooling and degassing as magma rises from a crustal reservoir and approaches the surface. Higher-silica magmas show greater polymerization of silica tetrahedra, so for similar temperatures, a rhyolite might be 3–7 orders of magnitude more viscous than an alkali basalt (**Figure 5**) (Shaw (1972), Spera (2000) and others). At constant P_{H_2O}, the viscosity of those same magmas will increase by 0.5–2 orders of magnitude per 100°C decrease in temperature (faster in more silicic magmas). Viscosity will increase by 2–4 orders of magnitude as water content drops from 8 wt.% down to < 1wt.%, with changes slightly faster at lower temperatures than at higher temperatures, and especially fast in the last 1 wt.% of degassing. These are profound changes.

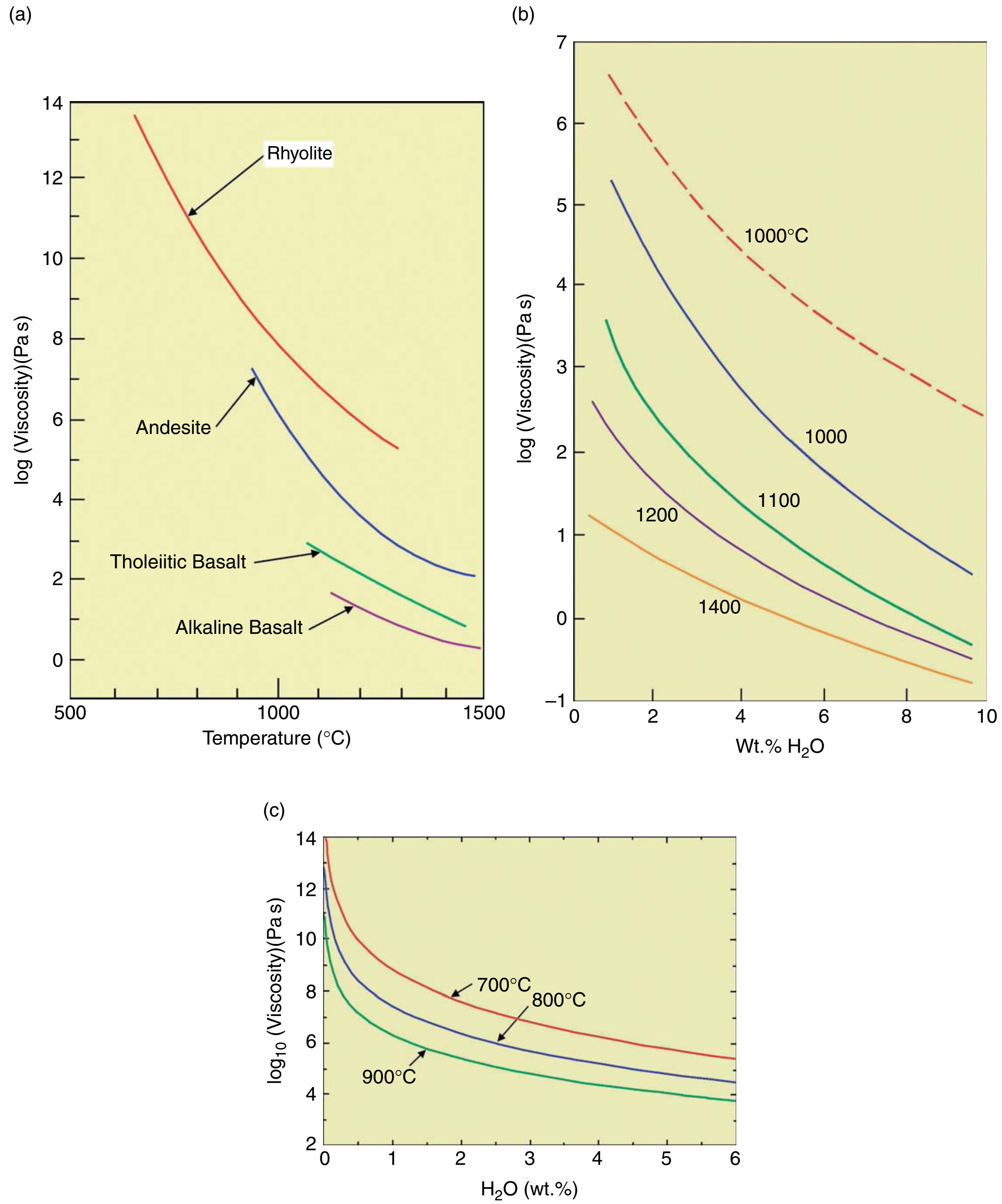

Figure 5 Viscosity of magma varies by orders of magnitude depending on SiO_2 and H_2O content, and can increase by orders of magnitude as it rises to the surface, losing H_2O and growing crystals. (a) Viscosity vs SiO_2 content and temperature; (b,c) viscosity vs H_2O content and temperature, for mafic and rhyolitic melts. Figures from Wallace and Anderson (2000), Dobran (2001), and Hess and Dingwell (1996).

Magma viscosity (i.e., viscosity of the melt + crystal + bubble mix) also increases by orders of magnitude as crystal contents rise from 0% to 60%, following an empirical relation (in lavas) of

$$\mu_{\mathrm{magma}} = \mu_{\mathrm{melt}}(1 - cX)^{-2.5}$$

where $c \sim 1.67$, X = volume fraction phenocrysts (Marsh, 1981).

Magma viscosity can either increase or decrease with bubbles, depending on the 'capillary number' (the ratio of viscous and surface tension stresses in a bubble). Small capillary numbers inhibit stretching of bubbles so an increase in bubbles increases magma viscosity. Large capillary numbers allow bubbles to stretch so an increase in bubbles can lower magma viscosity (Manga and Loewenberg, 2001). In either case, change is usually by less than a factor of 5 (Pal, 2003).

Magma density is a minor function of melt composition (silicic, e.g., rhyolitic magmas are slightly less dense than mafic, e.g., basaltic magmas; 2.4 g cm^{-1} vs 2.6 g cm^{-1}). Even volatiles in bubble form, if above their critical points (for water, 374°C, ~22 MPa; for carbon dioxide, 31°C, ~7 MPa), exert relatively minor influence on magma density. A much stronger influence comes from volatiles that are already in a gas state, in magma at shallow depth at pressures less than a few tens of MPa. Recalling the ideal gas law, $PV = nRT$, pressure and volume are inversely related when temperature and number of gas molecules are constant. If depth and confining pressure decrease, bubbles will expand and the density of magma can decrease dramatically. We can visualize this with the same bottle of soda pop that is considered earlier. The density of soda pop with only dissolved gases is close to 1.0 g cm^{-1}. Once it foams, the density can drop to a small fraction of that.

Volatile contents play into both viscosity and density. Exsolution of water from melt has the effect of increasing viscosity of the remaining melt. Because degassing of melt also induces microlite crystallization, another major contributor to increased viscosity, degassing is especially effective in sharply increasing 'magma' viscosity. To some extent, this is a self-limiting process because the viscosity of the degassed outer carapace of magma will increase so much that it acts to block further degassing, but that carapace is subject to cracking. New bubbles, if they remain in the magma, will increase or decrease viscosity slightly, and have a major effect on its density.

4.12.3 Intrusions and Convection

4.12.3.1 Orientation, Geometry, Dimensions of Intrusions

The shapes and orientations of magma conduits are functions of principal tectonic stresses and pre-existing structures. In cases of intrusion into a homogeneous medium within a tectonic stress field, intrusions will be as roughly vertical planar bodies ('dikes') and perpendicular to the minimum principal stress direction σ_3. As intrusions reach closer to the surface, a vertical axis often becomes the minimum principal external stress direction and there is a tendency for intrusions to become vertical cylinders. However, in regions of extensional stress, σ_3 remains horizontal and intrusions may persist as dikes all the way to the surface. Where magma has limited buoyancy, or encounters an impenetrable horizontal layer during its ascent, it can spread laterally as broadly horizontal planar sheet ('sill').

Basaltic dikes range from a few tens of centimeters to several tens of meters in width, most less than a few meters. Composite dikes (dikes consisting of several intrusions, each one wedging into the previous one) can be wider, as can 'dike swarms' (sets of many subparallel dikes that have intruded the same area over time, each one solidifying before the next). Silicic dikes can be several meters to several tens of meters in width, rarely up to ~100 m wide. Cylindrical intrusions of magma termed plugs, often silicic, can be tens to hundreds of meters in diameter. Sills can vary from several meters to several tens (rarely, hundreds) of meters in thickness. Like composite dikes, sills may inflate by repeated, closely spaced intrusions.

4.12.3.2 Drivers of One-Way Magma Ascent

Conventional wisdom is that magma follows a one-way path upward to eruption. Forces that drive magma upward are (1) buoyancy, (2) lithostatic pressure, and (3) magmatic overpressure. Buoyancy is effective only in fluid–fluid interaction where downward-directed return flow can occur. Lithostatic pressure constantly squeezes the sides of a magma intrusion and that magma will try to move up to a zone of lower lithostatic pressure. The tops of many crustal magma reservoirs are at 3–6 km, where lithostatic pressures are between 100 and 200 MPa; speculative reasons for this fact are raised in Section 4.12.9. Magmatic overpressure is internal pressure in excess of lithostatic pressure, caused by volatile expansion at shallow depths or by addition of magma from greater depth. Like lithostatic pressure, it acts in all directions and not just upward, but the easiest relief of this pressure is usually upward.

In a few cases, lithostatic pressure may be augmented by compressive tectonic strain (literally, squeezing magma upward from cracks or from ductile rock at depth). Compressive strain of about 0.1 MPa beneath Pinatubo as a result of the 1990 Luzon earthquake might have squeezed basaltic melt upward – from either the lower crust or, perhaps more likely, throughout the whole length of a deep magma feeder system – until it intruded a residual dacitic reservoir (Bautista *et al.*, 1996; Newhall *et al.*, 2002).

4.12.3.3 Drivers of Convective Magma Ascent with Return Flow

The conventional wisdom of one-way transport is hard to reconcile with so-called 'excess degassing', (Andres *et al.*, 1991), a common occurrence in which volcanoes emit far more SO_2 and presumably other volcanic gases than could have been dissolved in modest volumes of magma that are erupted over the same period. A combination of field observations and modeling (Kazahaya *et al.*, 1994, 2002, 2004; Stevenson and Blake, 1998; Witter *et al.*, 2004, 2005) suggests that in many, perhaps most of these cases, the extra gas came from magma that is convecting within the conduit. Low confining pressure allows the top of a magma column to develop into a permeable foam, probably, just a few hundred meters below the surface. Gas escapes and the residual degassed, collapsed foam sinks back down through the conduit (**Figure 6**). Gravitational settling of these blebs of degassed magma is what drives convection, by displacing gas-rich magma upward (Bergantz and Ni, 1999). This is not a 'conveyor belt' convection cell with continuity between upwelling and downwelling limbs, where higher viscosities of the degassed magma would stop the convection. Instead it is a modified, 'lava lamp' style of convection in which the viscosity of the downgoing blebs is irrelevant. Sinking blebs can even be solid and the process will still operate.

Magma foams are known from their 'frozen' equivalents, including very low-density, fluffy 'reticulite' pumice, and are also inferred from relic textures in some silicic lava flows that suggest they were erupted as foam (Eichelberger *et al.*, 1986, Eichelberger (1995); for cautions, see Fink *et al.* (1992)). Dimensional modeling by Witter (2003) indicates that mafic magmas not only can but probably will convect in conduits 1–5 m diameter, and relatively hot, volatile-rich silicic magmas will do the same in conduits 5–20 m in diameter. Relatively, high alkali content of silicic magmas will also reduce viscosity and increase the chance for convection. Rates of

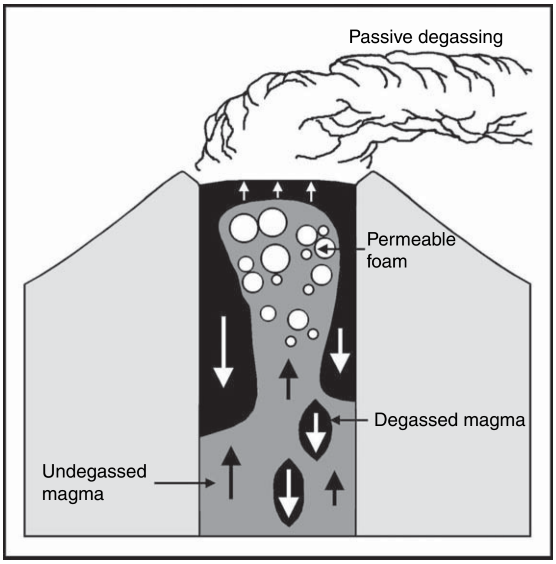

Figure 6 'Lava lamp' convection in a conduit. Gas-rich magma develops a head of foam near the surface. Gas escapes, and the degassing foam collapses. Now dense, it sinks back down through the column, displacing gas rich magma upward. Once the process begins, it will be driven primarily by sinking blebs, unless or until the supply of fresh gas rich magma into the conduit becomes too slow to sustain the process. Adapted from Kazahaya K, Shinohara H and Saito G (2002) Degassing process of Satsuma–Iwojima volcano, Japan: Supply of volatile components from a deep magma chamber. *Earth Planets Space* 54: 327–335.

SO_2 gas emission constrain rates of upward flow during routine, continuous convection to be in the order of mm to cm/sec.

This has important implications for both seismic and deformation precursors of eruptions. Where magma is convecting freely in a conduit, any seismic and deformation precursors to eruption will be subtle and occur only shortly before eruption. Little or no rock needs to be fractured before magma can erupt. Empirically, this is observed at frequently active, relatively open-vent volcanoes, often with mafic to intermediate magmas. Steady-state gas supply to the near-surface, balanced perfectly by gas leakage (degassing), can be tipped in favor of gas supply by a fresh intrusion of magma, gas, and heat at depth and a corresponding increase in the rate of convection. This imbalance, causing gas pressures to build slightly, will produce very subtle seismic and deformation precursors to eruptions – typically recordable only at or near a crater rim just minutes before an explosive eruption, for example, at Sakurajima Volcano (Kamo *et al.* 1994) or Merapi Volcano (Voight *et al.*, 2000). The same subtlety applies at all open conduits, even where magma ascent appears to be one-way upward, for example, at Soufriere Hills, Montserrat (Voight *et al.*, 1999).

In contrast, seismic and deformation precursors where magma in the conduit is neither convecting nor otherwise open can be expected to be more pronounced, with significant magma pressurization and rock breakage as magma breaks a new path to the surface.

Aside from observed degassing, is there other evidence for convection of magma in conduits? Might there be seismic and/or electromagnetic evidence of magma moving through the 'lava lamp' convection 'cell'. The author is not aware if this has been found, but he expects this to be the case. Do we find large phenocrysts that show oscillatory compositional zonation to indicate that they rode a convection elevator up and down the conduit multiple times? Perhaps, as at Erebus volcano, Antarctica (Dunbar *et al.*, 1994), though most other cases of oscillatory zoning are equivocal and might alternatively be explained by fluctuations in magma water content or temperature without convection. We might expect that U–Th ages of plagioclase crystals in convecting magma might span a wide range of ages, and indeed such ages at Mount St. Helens range from 10^3 years to just a few years (Cooper and Reid, 2003), but neither gas emission nor other evidence points to convection in the conduit of Mount St. Helens. Perhaps the strongest evidence will be found in degassed groundmass glass. In products of the 2000 eruption of Mayon Volcano, Arpa *et al.* (2006) document abundant blebs of (apparently) degassed andesite with microcrystalline groundmass wrapped within darker, (apparently) undegassed lava of the same composition. Degassing was inferred from textures and still needs to be confirmed by direct analysis of the glasses. It would be interesting to reexamine more and less crystalline groundmasses of Merapi lavas described by Hammer *et al.* (2000), particularly in light of the findings of Allard *et al.* (1995) about excess degassing which, to this author, strongly suggests convection of magma in the conduit.

4.12.3.4 Magma Mixing – Conditions, Evidence, Effects

Mafic magmas arriving from the mantle and lower crust often encounter more silicic magmas enroute to the surface. This phenomenon is especially common at arc and continental volcanoes at which the long-term supply rates of mafic magma are sufficiently high (10^6–$10^8\,m^3\,yr^{-1}$) to generate and sustain silicic magmas by partial melting of crust and by fractional crystallization and other differentiation processes in long-lived magma reservoirs (Shaw, 1985). Such systems have a thermal memory of previous intrusions, and many have large silicic magma bodies (<5 to 1000's of km^3) at depths ~5–15 km beneath the surface.

In order for magmas to mingle or mix, their viscosities must be within approximately one order of magnitude of each other (Sparks and Marshall, 1986). A number of models of mingling and mixing have been proposed. One mechanism involves underplating of basaltic magma beneath a silicic reservoir, with thermally driven convection in one or both layers and mixing along a subhorizontal interface (Sparks *et al.*, 1977; Sparks and Marshall, 1986). Another mechanism involves intrusion of mafic magma into silicic magma. Rising mafic magma may have sufficient momentum to simply intrude right into overlying silicic magma (Bergantz and Breidenthal, 2001). In a third mechanism, cooling and partial crystallization of mafic magma when it encounters less dense silicic magma will induce vesiculation of the mafic magma, make it less dense than the silicic magma, and buoyancy will let mafic magma 'tunnel' vertically into the silicic body. Mingling will occur along a subvertical interface (Eichelberger, 1980; Bergantz and Breidenthal, 2001).

Mingling and mixing of two magmas is clearly evident in the hybrid products. Blebs of the intruding, vesiculated basalt may be quenched and preserved as inclusions in the hybrid magma, with distinctive open-lattice ('diktytaxitic') textures (Eichelberger, 1980; Pallister *et al.*, 1996). Further shear can cause quenched blebs to mechanically disintegrate. Incomplete mixing ('mingling') produces a marble-cake swirl of the two components. If mafic melt mixes completely with silicic melt, compositionally intermediate melts develop. Both mingled and mixed magmas show disequilibrium juxtaposition of otherwise incompatible minerals, for example, olivine and quartz, and reaction rims on minerals from the higher temperature end member, or resorption of the rims of plagioclase from the lower temperature end member.

Magma mixing is not known to have a direct geophysical signal. However, changes of temperature and oxygen fugacity associated with mixing can substantially change the solubilities of gases in both magmas. If solubilities are decreased enough, bubbles will form and one should expect seismic and deformation signals associated with buoyant rise of magma and expansion or pressurization of bubbles as that magma rises and lithostatic pressure decreases. One of the best-documented instances of magma mixing began at Pinatubo Volcano several months before its eruption in June 1991 (Pallister *et al.*, 1996; M. Coombs, personal communication, 2002); the latest mixing occurred <1 week before eruption (Rutherford and Devine, 1996). Water-rich basalt intruded (tunneled into) and mixed with sluggish dacite magma, early in a chain of events that, by 15 June 1991, led to eruption of $\sim 5\,\mathrm{km}^3$ of dacitic magma, the second largest eruption in the twentieth century.

4.12.3.5 Geospeedometers of Magma Ascent, Convection, and Mixing

Despite what we might expect from a dike propagating toward the surface, there are very few known cases in which earthquake hypocenters have migrated upward. A broad ~ 3 km upward migration of hypocenters may have occurred at Pinatubo (Philippines) in 6 days in June 1991 ($\sim 20\,\mathrm{m\,h^{-1}}$) (Harlow *et al.*, 1996). That migration was apparent in fast playback of hypocenters in VOLQUAKE (Hoblitt *et al.*, 1996), though less obvious in relocated hypocenters (Rowe *et al.*, 2001). In 1975 at Plosky Tolbachik (Kamchatka), Fedotov *et al.* (1980) traced seismicity (and, by inference, basaltic dike intrusion) from 20 km to the surface at an average rate of about $0.1\,\mathrm{km\,h^{-1}}$. In both cases, hypocenters spanned a significant part of the conduit length at any given time, but moved broadly upward. Other reported cases are Kilauea east rift basalt at 0.1–$1.5\,\mathrm{km\,h^{-1}}$ (Klein *et al.*, 1987), East of Izu basalt at $1\,\mathrm{km\,h^{-1}}$ (Hayashi and Morita, 2003), and Unzen dacite at $\sim 3\,\mathrm{m\,h^{-1}}$ (Nakada *et al.* 1999).

In principle, the tip of a dike would also be a point source of deformation (Mogi, 1958) and this, too, could be tracked upward. There are a few instances in which tiltmeters have recorded traces consistent with upward migration. A growing body of InSAR data will surely show some new rates of migration; one early result is that of Lu *et al.* (2003) for magma dynamics at Westdahl Volcano, Alaska.

A different geospeedometer – applicable during and perhaps before eruptions – considers the diameter of dense ultramafic or mafic xenoliths that get erupted, calculates their likely settling rate, and infers that rate as a minimum upward velocity of magma in order to bring these xenoliths to the surface (Spera, 1984). In the 1801 eruption of Hualalai, Hawaii, magma ascent was apparently $>30\,\mathrm{m\,h^{-1}}$.

Several petrologic geospeedometers exist, all relying on the fact that minerals which are taken out of equilibrium will develop disequilibrium reaction rims (e.g., hornblende will develop pyroxene and Fe–Ti oxide rims), and the travel time for that magma from reservoir (equilibrium) to the surface (quenching) is modeled as the time required for the observed thickness of reaction rim to grow in lab experiments (Coombs *et al.*, 2000; Rutherford and Gardner, 2000). Sometimes, the absence of zoning or reaction rims can also be used to make a minimum estimate of ascent time – or maximum diameter of a conduit – on the premise that zoning or rims would have developed given enough time.

In discussion of convection, earlier, we noted that rates of ascent can be judged from gas flux rates and assumptions about conduit diameter and the concentration of gas in the magma. These estimates suggest rates of meters to tens of $\mathrm{m\,h^{-1}}$ (Witter, 2003).

By similar arguments, a known volume rate of extrusion and an assumed conduit diameter will give an ascent rate. At Mount St. Helens in the 1980s, mass eruption rates during explosive events implied ascent rates of 1–$2\,\mathrm{m\,s^{-1}}$ (3600–$7200\,\mathrm{m\,h^{-1}}$). During nonexplosive dome extrusion in the 1980s, and again in 2004–2005, ascent rates were tens to hundreds of $\mathrm{m\,h^{-1}}$.

Finally, the time from magma mixing to eruption can be estimated by the thickness of reaction rims or by the preservation versus loss of disequilibrium mineral pairs that were juxtaposed by mixing. From Arenal Volcano, olivine reaction rims of at least 11 year growth are described by Coombs and Gardner (2004) and, at the opposite end of the time spectrum, preservation of disequilibrium Fe–Ti oxides at Mount St. Helens, Pinatubo, Unzen, and Soufrière Hills or cummingtonite (an amphibole mineral not stable above 800°C) at Pinatubo indicate maximum times from mixing to eruption of only a few days (Rutherford and Devine, 1996; Pallister *et al.*, 1996; Devine *et al.*, 1997, 2003; Pallister *et al.*, 2005). In another approach, Blake and Fink (2000) propose a conceptual framework from which time-frames of mixing might eventually be estimated from the relative sizes and degrees of deformation of mafic enclaves that get ‘frozen’ upon being mixed into cooler silicic magma.

4.12.4 Pressurization within and around Magma

4.12.4.1 Increase in Pressure

Pressures will be increased by the following.

4.12.4.1.1 Volatile accumulation

Volatile accumulation is to eruptive potential as accumulated strain is to tectonic earthquakes. To the extent that there are 'eruption cycles', they are driven by patterns of net volatile accumulation. Recall that under discussion of gas budgets, it is postulated that magma and gas supply to most volcanoes is relatively frequent and in small increments. At some volcanoes, it may be almost continuous at all timescales; at others, it may be intermittent over timescales of days to years but constant over timescales of decades. In either case, the interval of magma supply from depth, into a shallow crustal reservoir, is probably quite a bit shorter than the eruption interval.

Plots of cumulative volume of magma erupted versus time for various volcanoes can show time predictability, volume predictability, or no predictability, that is, completely Poisson behavior (Koyama and Yoshida, 1994). Most volcanoes that have open vents and leak gas and magma fairly freely are time predictable, as the threshold of internal pressure that needs to be exceeded is low and constant. Volcanoes whose conduits freeze up between eruptions are often volume predictable. At such volcanoes, explosivity correlates with volume erupted, both being controlled by the accumulated mass of volatiles that have not leaked out before magma reaches close to the surface. Most of the largest explosive eruptions – 'plinian eruptions' – are of magma that is not only saturated with volatiles but has also accumulated 10×–100× more volatiles (by weight) in bubble form (e.g., Pinatubo; Gerlach *et al.* (1996) and Wallace (2001)). This is the basis for the oft-repeated dictum that a volcano which has been dormant for hundreds of years is likely to produce a large explosive eruption. This dictum is true only if the conduit has been closed and gas has been accumulating, not leaking. Pinatubo was just such a volcano, and accumulated most of the incoming gas for ~500 years. Iwo-jima in Japan has had almost as long a repose period and plenty of magma supply but shows no signs of erupting magma – because it is exceptionally leaky (Newhall *et al.*, 1994, 2003; Newhall, 2004; Ukawa *et al.*, 2005).

If volatiles accumulate in excess of saturation, forming bubble phases, further accumulation will gradually increase pressure on the walls of the container. The effect is minimized at depths where volatiles are most likely accumulated and volatile phases are in a supercritical fluid. However, accumulation of volatile fluid bubbles increases the 'potential' for rapid pressure buildup at shallow depths where confining pressures are low.

4.12.4.1.2 Magma ascent and volatile expansion

As magma ascends and lithostatic pressure decreases, partial pressure of the gas increases and only the strength of viscous magma and country rock will hold it in check. The pressure of this contained gas will drive magma to intrude through country rock. Eventually, if the internal gas pressure exceeds the strength of the container, the container will fail and magma will erupt.

4.12.4.1.3 Heating and/or compression of groundwater

Confined groundwater, hot or cold, can add significantly to pressures in and around a magma body. If heated, it expands and increases pore water pressures (Reid, 2004), though slow diffusion of heat through groundwater will keep effects quite local or slow. Thermally induced hydraulic pressurization may be greatest at the tip of a rising body of magma – that is, a wedge driving open cracks.

If porous host rock undergoes sudden compression due to nearby magma intrusion, its pore water pressures will jump quickly (Sato *et al.*, 2001; Shibata and Akita, 2001; Newhall *et al.*, 2001; Matsumoto *et al.*, 2002; Roeloffs and Linde, 2007). Observations of research wells at Usu Volcano in Japan and Krafla Volcano in Iceland have shown up to 100 m immediate rise in water level, corresponding to a notable 1 MPa increase in pore water pressure. This pressure bleeds off in a matter of days, but can briefly trigger seismicity for 5–20 km around a volcano (White and Power, 2001), and/or sharply decrease slope stability as probably occurred at Mount St. Helens in 1980, Unzen in 1792, and many other volcanoes that have massive sector collapses (Day, 1996; Voight and Elsworth, 1997).

4.12.4.1.4 Self-sealing in hydrothermal systems

Hydrothermal waters often contain dissolved silica and other minerals, and precipitate them near their margins. The effect is to develop an impermeable seal, trapping hot hydrothermal brines beneath (Fournier, 1999, 2007).

4.12.4.1.5 Lithostatic pressure

As described earlier, lithostatic pressure is the main source of pressure within magma bodies at depth, where pressures and temperatures are above the critical point for water. Lithostatic pressure squeezes the sides of a magma body at the same time it 'weighs heavily' on that magma. Squeezing of this body will be greater at its base than at its top, hence creating a driving force for upward intrusion. Furthermore, if a large body of magma feeds upward into a narrow feeder (e.g., a dike), magma in the conduit will act like a manometer (as in the straw of a slightly squeezed juice box), and stress will be concentrated at the tip of that dike that will help the magma break its way up through country rock.

4.12.4.1.6 Regional tectonics – quasistatic and dynamic strain

Any compression added by quasistatic or dynamic strain will enhance the squeezing mentioned above. Conversely, decompression, whether static or transient (dynamic), will tend to lower the solubility of gases in solution and/or allow expansion of bubbles in the magma. Lowering of solubility can increase internal volatile pressures more than the triggering decrease of confining pressure.

4.12.4.2 Decrease in Pressure

The pressure will be diminished by the following.

4.12.4.2.1 Gas leaks

The effect of gas leaks from magma is like that of gas leaks from soda pop: the magma goes flat. For details, see earlier discussion, Giggenbach (1996), and Jaupart (1998).

4.12.4.2.2 Eruptions

Volatiles tend to accumulate in the uppermost parts of magma conduits and reservoirs, and will be depleted by eruptions that, naturally, tap those parts. Eruptions are major 'gas leaks'.

4.12.4.2.3 Groundwater diffusion and dissipation of porewater pressures: Breakage of self-seals

A pressurized hydrothermal fluid surrounding magma will eventually bleed off its pressure through porous flow or fractures. If the cause of pressurization in the first instance was development of silica-rich self-seals, fracturing of those seals from within or by regional earthquakes will rapidly depressurize the source fluid but may briefly pump up pressures along connected, surrounding fractures. The concept is described by Fournier (1999, 2007).

4.12.4.3 Dynamic Balance between Pressurization and Depressurization

In summary, a variety of processes act to increase or decrease pressures within magma and surrounding hydrothermal systems. The increases and decreases may be in delicate balance. This delicate balance often appears during ascent of gas-rich magma from a crustal reservoir toward the surface. Volatile exsolution occurs as solubilities in melt decrease, and both newly exsolved and any bubbles already existing in the reservoir exert increasing internal pressure as lithostatic pressure decreases. If gas is able to leak out and decrease internal pressure as fast at that pressure is increased from ascent, there will be no net increase in gas pressure and magma may arrive at the surface with little or no explosive potential. Conversely, if magma ascent is rapid and pressures build faster than they can bleed off, an explosive eruption will result.

A parallel exists in hydrothermal systems. Confined groundwater can be pressurized by addition of insoluble gas, by heating (from that and from soluble gas), and, most effectively, by mechanical compression of the aquifer by nearby magma intrusion. If these pressurization processes are slow, porous and fracture outflow may prevent any rise of porewater pressure. On the other hand, there are clearly some instances in which water pressures rise sharply, whether at the tip of a rising dike or more broadly, and resulting hydrofracturing may facilitate magma ascent and/or phreatic explosions.

4.12.5 Volcanic Unrest

This section is a brief overview of the nature, magnitude, and frequency of 'observable' physical and chemical changes that occur between eruptions, and especially as magma intrudes toward the surface. Principal measuring techniques are also mentioned. Useful general references include UNESCO (1972), Tilling (1989), Carroll and Holloway (1994), McGuire *et al.* (1995), Scarpa and Tilling (1996), and Sigurdsson *et al.* (2000). Readers might note that gas emissions are described ahead of seismicity and ground deformation, opposite the usual sequence. This is for two reasons. First, gas emissions are often the first potentially detectable sign of unrest. Second, increases of gas and hydrothermal pressures are what drive volcanic seismicity, ground deformation, and magma ascent.

4.12.5.1 Gas Emissions

In situ collection of gas samples from fumaroles, for subsequent analysis in a laboratory, has the advantage of collecting the whole gas and allowing use of a range of laboratory analytical instruments (Symonds *et al.*, 1994). However, for monitoring purposes it has the disadvantages of collections being infrequent and often dangerous. Remote measurements from a safer distance like the foot of a volcano, from aircraft, or even from satellites have the advantage of being more frequent and safer but are usually of just one or a few gas species.

The most easily and widely measured indicator of magma degassing is sulfur dioxide (SO_2). It is much more abundant in magmatic gas than in the atmosphere and can be measured *in situ* or remotely by now-small and relatively inexpensive UV spectrometers. Early work (Stoiber and Jepsen, 1973; Stoiber *et al.*, 1983) showed that SO_2 emission increased before eruptions as its solubility in rising magma decreased. Subsequent work has found that this pattern is sometimes complicated by path effects. One such effect is temporary blockage of SO_2 exit from magma by sharp increase in viscosity of magma and/or by quenching and solidification of a carapace on the magma. This effect is thought to explain brief decreases in SO_2 emission shortly before explosions of Galeras volcano (Fischer *et al.*, 1996; Zapata *et al.*, 1997) and perhaps at Pinatubo as well (Daag *et al.*, 1996). Another path effect is absorption of SO_2 into groundwater, in which it is highly soluble (Doukas and Gerlach, 1995; Symonds *et al.*, 2001). The apparent absence of SO_2 before eruptions, once thought to indicate absence of magma, is now understood in some cases to reflect absorption (scrubbing) of SO_2 into groundwater. Scrubbing persists as long as the gas must pass through water-saturated country rock; it ends and SO_2 emission can rise sharply after hot gas has boiled open a dry chimney through the groundwater.

Increasingly, CO_2 gas is also monitored – directly from fumaroles, by remote sensing (FTIR), and by measurement of CO_2 flux through soil. CO_2 has the lowest solubility of any of the major gases in silicate melt, so it is the first to be released – starting when magma is as deep as ~20 km depth). It is also much less soluble in groundwater than SO_2, so at many volcanoes, it leaks out of the slopes of the volcano (diffuse degassing) and can be measured in soil. At Usu Volcano, an increase in diffuse CO_2 outgassing appeared 6 months before seismic or geodetic changes were detected (Hernandez *et al.*, 2001). At Mammoth Mountain in California, diffuse CO_2 outgassing was first noticed in late 1989 and early 1990, months after seismicity in mid 1989 signaled a magma intrusion at ~2–10 km depth, and rose again during an increase in seismicity at 10–20 km depth throughout 1997 and at 4–8 km depth in late November 1997 (Gerlach *et al.*, 1998; McGee *et al.*, 2000; Hill *et al.*, 2003). The apparent delay in CO_2 in 1989–90 may simply reflect delayed sampling; there was no delay in 1997 (T Gerlach, oral communication).

Because the solubility of SO_2 in silicate melt is greater than that of CO_2 and less than that of HCl or HF, rising magma might produce CO_2 at first, then increasing SO_2/CO_2 ratios, and eventually increasing HCl/SO_2 or HF/SO_2 ratios (e.g., Noguchi and Kamiya, 1963). This simple solubility-related progression is a nice conceptual framework for monitoring and might even offer prospects of estimating magma depth from gas emissions. However, discovery that many magmas are already gas-saturated in their reservoirs, and that other gases remaining in melt will partition preferentially into bubble phases rather than silicate melt, means that there may already be a fairly complete suite of magmatic gases in bubbles at depth, and all can be released together.

4.12.5.2 Seismicity

Because this author is a geologist rather than a seismologist, and because (Chapter 4.13) give readers a modern view of volcanic seismology, I originally thought to skip this section entirely. However, it might be interesting for seismologists to have a glimpse of how this one nonseismologist struggles to understand their field, and to relate it to other topics in this review.

Most volcanic seismicity falls into categories described by Minakami (1960), Shimozuru (1972), McNutt (1996, 2000a, 2005), and (Chapter 4.13). Volcano-tectonic (VT) earthquakes result from brittle fracture (of country rock) beneath, within, and near the volcano, caused by rise and pressurization of magma or induced by related increases in adjacent porewater pressures. Most VT events are normal, high-frequency events and are individually indistinguishable from small tectonic earthquakes. Indeed, some may be nothing more than tectonic earthquakes along faults beneath the volcano. What distinguishes VT events from strictly tectonic events is their common occurrence in swarms rather than in mainshock–aftershock sequences, their location beneath or near active volcanoes, and their common association with other indications of magma intrusion. *b*-values of VT earthquakes tend to be higher than for tectonic events (Sanchez *et al.*, 2004) because of heterogeneous structure and limited fault lengths, and often-elevated pore-pressures in hydrothermal fluids (Sanchez *et al.*, 2004; McNutt, 2005).

A subset of VT earthquakes that occurs around rather than beneath volcanoes has been called 'distal VTs' by White and Power (2001). Some volcanoes have earthquake swarms at distances of a few up to ~20 km from their summit, out beyond the margins of any magma body. Centered 17 km from Guagua Pichincha volcano, the North Quito swarm turned on and off as the hydrothermal system of the volcano pressurized and then erupted, respectively (Legrand *et al.*, 2002). Near Soufrière Hills and Pinatubo volcanoes, swarms 3 and 5 km from the respective summits occurred during early stages of their eruptions, as pressures in the magma reservoir and surrounding hydrothermal system increased. There is no implication that magma itself is present where distal VTs are occurring, nor even in hydraulic connection with the hypocentral area. The correlation can be through propagation of mechanical strain that acts on locally confined aquifers.

Decades ago, Minakami (1960) recognized a general progression from high-frequency, impulsive-onset VT (or, in his terminology, A-type) earthquakes at depths >1 km, upward to low-frequency (his B-type) earthquakes at <1 km depth that were often more emergent as well as having lower frequency content. Minakami's concept of a standard progression of earthquake types as magma rises remains useful, and was nicely updated into a generic earthquake swarm model by McNutt (1996).

Subsequent work suggests that 'B-type earthquakes' – shallow and of relatively low frequency – are actually of multiple origins. Some reflect source processes as discussed in the next paragraph, and others reflect attenuation of high frequencies and other path effects as emphasized by Minakami (1974) and Malone (1983). Ishihara and Iguchi (1989), Iguchi (1989), and Tsuruga *et al.* (1997) distinguish two to three subcategories of 'B-type' earthquakes at Sakurajima, representing both fracturing and subsurface explosions.

Low-frequency earthquakes at volcanoes have several different origins. Some seem to occur by rupture of relatively ductile material, either hydrothermally altered rock or of viscous, nearly solid magma (Ramos *et al.*, 1999; Tuffen and Dingwell, 2004; Moore *et al.*, 2005). If strain rate is high enough, even ductile materials can rupture. A subset of these low-frequency earthquakes ('hybrids') has a high-frequency onset followed by a low-frequency tail, and these commonly occur within or just beneath growing lava domes. Rupture may begin in the brittle carapace and propagate inward to softer material. Alternatively, hybrids might reflect occurrence of an earthquake in brittle rock that sets nearby fluid into oscillation (McNutt, 2000a), or result from complex paths rather than complex sources (Kedar *et al.*, 1996, 1998).

Another class of low-frequency earthquakes is sometimes termed a long-period (LP) earthquake to convey an added character – a relatively monochromatic character, dominantly 1–3 Hz. Some of these events are strikingly monochromatic, such as so-called 'tornillo' earthquakes of Galeras Volcano and elsewhere (Gil Cruz and Chouet, 1997; Gómez and Torres, 1997). Chouet (1992, 1996a, 1996b) modeled these as the result of sudden jetting of pressurized gas through cracks, followed by resonance of the crack that just opened and shut. In some but not all cases, there is visual corroboration of steam jetting shortly after LP events. Initially, occurrence of LP events in the buildup toward explosive eruptions (e.g., at Redoubt, 1989–90) led to an inference that these reflected gas pressures within the magma, and explosive potential (Chouet, 1996a). Subsequent observation of many such events in the aftermath of eruptions, when conduits are relatively open and streaming gas, suggests jetting of gas through cracks but not necessarily under great pressure. Hellweg (2003) suggests an interesting alternative, yet to be fully tested, that the monochromatic or harmonic character of some such events might be a path effect, resulting from scattering polarization of signals.

Some low-frequency (LP) events occur deep beneath volcanoes. At Pinatubo, deep LPs (DLPs) were approximately 35 km deep and were observed in late May and early June, only a week before magma erupted (White, 1996), but evidence for earlier magma mixing suggests that other LPs might have occurred before seismic monitoring was established. Beneath Kilauea, deep LPs are 30–60 km deep (Koyanagi *et al.*, 1987), and beneath Fuji, 10–20 km deep (Nakamichi *et al.*, 2004; Ukawa, 2005). A number of examples are noted by Power *et al.* (2004). DLPs are thought to reflect resupply of magma from depth into shallower reservoirs. They do not by themselves indicate imminent eruption.

With increasing use of broadband seismometers at volcanoes, there are increasing reports of very-long-period (VLP) earthquakes with periods of up to several tens of seconds (e.g., Kumagai *et al.*, 2003; Chapter 4.13). Some are thought to occur when slugs of fluid (magma, water, or gas) pass through constrictions. The very long period reflects the time it takes for fluid head to pass through the constriction.

The last major category of volcanic seismicity is volcanic tremor – periods of continuous vibration lasting from minutes to years (McNutt, 1992; Konstantinou and Schlindwein, 2002; Instituto Geofisico, 2006). The energy of tremor normalized for station distance, magnification, and signal spreading is reported as reduced displacement (Aki *et al.*, 1977; Fehler, 1983). For surface waves, reduced displacement (RD) = $\left(A\sqrt{\lambda r}/2\sqrt{2}M\right)$, where A is their peak-to-peak amplitude, λ is the wavelength, r is the distance in km from source to seismic station, $2\sqrt{2}$ is the root mean square (rms) amplitude correction, and M is the instrument magnification. Some tremor evolves from or devolves into a succession of many small earthquakes, for example, at Mount St. Helens (Malone and Qamar, 1984) or Soufrière Hills, Montserrat (Neuberg, 2000), and is apparently just overlapping of those earthquakes, be they high or low frequency.

Other tremor begins as a low-level continuous signal, grows and is sustained, and eventually fades away as it began. Some is broad spectrum; other tremor can be of remarkably narrow spectrum with harmonics. One suggested origin, in addition to overlapping of discrete events, is nonlinear oscillation induced by flow of magma (or hydrothermal fluid) through conduits (cracks) (Julian, 1994, 2000; Balmforth *et al.*, 2005). A second mechanism is excitation and resonance of a fluid-filled crack, irrespective of whether fluid is flowing (Chouet, 1992, 1996a, 1996b). Both origins are widely inferred.

Generation of tremor by a third mechanism, hydrothermal boiling, gets less attention than it probably deserves. Leet (1988) described experiments on 'subcooled boiling', that is, boiling in which bubbles form where groundwater is in contact with magma or hot gas, but then implode within cold surrounding water. Subcooled boiling appeared to be capable of generating strong tremor. At least some instances of strong tremor with reduced displacement >30 cm^2 are associated with phreatic eruptions and times when we know that magma is intruding and probably boiling groundwater, for example, at Mount St. Helens in April–May 1980 and a during a short but strong episode on 2 October 2004. Some such tremor is also 'banded', turning on for a few hours, off for a few hours, and back on and off for days, strongly suggestive of geyser activity (e.g., at Karkar Volcano; McKee *et al.* (1981)). Elegant studies of tremor at Yellowstone were published by Kieffer (1984) and Kedar *et al.* (1996, 1998).

4.12.5.3 Ground Deformation

Ground deforms around magma in two ways – elastically in response to pressure (magmatic or hydrothermal), and inelastically as it is fractured and physically shoved aside by magma or if the flank of a volcano is collapsing (e.g., Johnson, 1987). Dzurisin (2003, 2007) offers excellent reviews of methods and interpretation. Elastic deformation is a normal consequence of magma intrusion and pressurization at depth, and can be aseismic. Radar interferometric (InSAR) monitoring of volcanoes has shown a number of volcanoes that are being deformed by relatively deep sources (probably, resupply of basaltic magma) without accompanying seismicity, for example, at Westdahl Volcano, Alaska (Lu *et al.*, 2003; Dzurisin and Lu, 2007). Another good example is in the southwest corner of the Three Sisters volcanic complex in Oregon, where deformation began as early as 1996 and was not known to have induced any seismicity until a small, shallow swarm in March 2004 (Wicks *et al.*, 2002; Dzurisin *et al.*, 2004).

The shape and gradients of the deformation field indicate the depth and shape of the deformation source (Mogi, 1958; Okada, 1992; Cervelli, 2000). It is beyond the scope of this chapter to describe the relationship between sources and deformation fields in detail, but a few general relations can be noted. Uplift from a point source (such as the rising tip of a magma column) forms concentric contours. The half-width of uplift, defined as the diameter of the contour of half the maximum uplift, is 3/2 times the depth of the point source. Tilt resulting from a point source shows its maximum value at a radial horizontal distance 0.5 times depth of the point source. Different patterns result from dikes. These models, developed by Mogi and Okada on theoretical grounds, are remarkably robust and consistent with seismic and other evidence for magma depths and shapes.

Inelastic deformation immediately around a magma intrusion is just extreme strain, exceeding the tensile strength of the rock. For example, grabens typically form above a near-surface dike, and the north flank of Mount St. Helens bulged irreversibly northward as a viscous dacite cryptodome (hidden dome) intruded into its north flank.

Iwo-jima Volcano (Japan) shows an interesting combination of elastic and inelastic deformation. Over the past 400 years at least, and perhaps the past 2600 years, the floor of the submarine Iwo-jima caldera has risen more than 120 m above sea level (Kaizuka *et al.*, 1985; Oyagi and Inokuchi, 1985; Ukawa *et al.*, 2005). Episodic magma intrusion of approximately $10^7\,m^3\,yr^{-1}$ is the driver. Some of the deformation is elastic, in which episodes of inflation once every decade or two are followed (partially recovered) over succeeding years by deflation of the central part of the island but, at least recently, by continued inflation outboard of the center (Ukawa *et al.*, 2005). But net deformation since the time of caldera formation is inelastic, and includes broad uplift plus numerous faults with offsets of several meters that cut post-WWII era runways (Oyagi and Inokuchi, 1985). Iwo-Jima is an actively resurging dome, and is also remarkable for its efficient degassing of magma that (at least for now) seems to prevent magmatic eruption (Newhall *et al.*, 2003). Some of the deformation clearly involves intrusion, possible pressurization, and degassing of magma; other deformation probably involves pressurization and depressurization of the surrounding hydrothermal system.

Because some nonvolcanic landslides are known to show accelerating creep shortly before failure, those monitoring the bulge at Mount St. Helens in 1980 were hoping for a similar indication of imminent failure. It did not occur. If anything, creep slowed slightly in the 3 days immediately before the failure (Lipman *et al.*, 1981). Apparently, resistance along the basal plane of the landslide block(s) was overcome suddenly.

4.12.5.4 Other Changes – Gravity, Magnetic, Electrical, Thermal, Hydrologic

Ascent of magma and release of hot gases into surrounding groundwater causes a variety of other changes. The strength of the gravitational field around volcanoes (typically, known at just one or a few monitoring sites) can increase or decrease by tens or even hundreds of microgals (10^{-1}–10^{+1} gu) (Rymer, 1996). An increase could be interpreted as subsidence, intrusion of magma that is denser than the rock into which it intrudes, constant-volume densification of magma by collapse of vesicles and coupled inflow, and/or a compression-induced rise in water table (increasing density just beneath the gravimeter). A decrease could be interpreted as uplift, drainback of magma (rare), vesiculation of magma, or dilatation-induced drop in water table. Independent geodetic measurements and modeling of the combined deformation-gravity field are needed to discriminate between these possible contributors (Rymer and Williams-Jones, 2000). Frequent or, better yet, continuous measurements are needed to capture short-lived, transient events. With GPS and other geodetic instruments we can have continuous measures of deformation, but only a very few volcanoes have had continuously recording gravimeters (Etna, Vesuvius, Merapi, Asama, Iwate, Miyake-jima, and Kilauea), and even fewer still have them.

Magnetic fields are known to change by as much as ten or more nanoteslas at some volcanoes and during some intrusions (Johnston, 1989; Sasai *et al.*, 1990; Zlotnicki, 1995). Heating can slowly demagnetize rock near an intrusion. Stress associated with intrusions can induce piezomagnetic effects, in the order of 1 nT per 10 MPa of overpressure (Zlotnicki, 1995). Still other changes may be of electrokinetic origin, when a conductive fluid (hydrothermal or magma) is moving (Mizutani *et al.*, 1976; Fitterman, 1978).

Magma intrusions can change the distribution of resistivity within a volcano, by changing porosity and permeability of a hydrothermal reservoir or conductivity of a hydrothermal fluid. Lenat (1995) gives a useful review of these changes and ways to measure them. One particularly interesting example is at Kilauea, where active-source electromagnetic monitoring saw change in phase and amplitude of signals during passage of a dike intrusion, and remeasurements of self-potential across a rift zone also showed change after a dike intrusion (Jackson *et al.*, 1985; Jackson, 1988). Similar changes have been recorded at Piton de la Fournaise and Etna volcanoes (Lenat, 1995), and at Miyake-jima (Zlotnicki *et al.*, 2003).

Minor shifts in the locus of thermal activity occur frequently at volcanoes, probably related to self-sealing and changing patterns of permeability within the hydrothermal system. Most of these are insignificant. Increases in ground, water, or gas temperatures by several degrees or even tens of degrees are more significant, potentially caused by increased flux of hot gas. Crater lakes are particularly useful integrators of thermal and chemical flux from hot gas streaming up into the lake (Varekamp and Rowe, 2000), and Bernard (2004) is developing a useful application of ASTER satellite imagery to monitor changes in crater lake temperatures. Fumaroles at or near the boiling temperature of water rarely change temperature because they are buffered by boiling of groundwater. High-temperature, drier fumaroles (some, as hot as 950°C) can have many-degree swings in temperature with increases or decreases in the flux of gas from magma. Large increases in volume of steaming, as at Mount Baker (1975), Pinatubo (1991), Turrialba (1997–present), or Garbuna (2004) can be another indication of hot gas from a fresh magma intrusion heating water of a hydrothermal system.

Finally, changes in water level or spring discharge occur at some volcanoes, principally in response to mechanical compression or dilatation of confined aquifers (Newhall *et al.*, 2001; Roeloffs and Linde, 2007). Most reports of change in water level are anecdotal but good measurements have been obtained in Japan and in Long Valley caldera. As mentioned earlier, increases in porewater pressure can induce seismicity at or near a volcano, cause modest inflation, and destabilize the flanks of volcanoes if pressurization occurs in a widespread, outward-dipping weak layer such as a layer of weathered ash (Day, 1996; Voight and Elsworth, 1997; Reid, 2004).

4.12.6 Volcanic Eruptions

Historically, volcanologists believed that nearly all eruptions were triggered internally, that is, when pressures within magma exceeded the tensile strength of country rock. No doubt this is true for many eruptions. But, especially since the 1992 Landers earthquake, evidence has built that some eruptions are triggered externally (Hill *et al.*, 2002; Schmincke, 2004; Manga and Brodsky, 2006). Here is a list of potential eruption triggers.

4.12.6.1 Eruption Triggering Mechanisms

4.12.6.1.1 Internal

4.12.6.1.1.(i) Intrusion of fresh magma (and its gas) from depth, driven by partial melting and by buoyant forces and/or lithostatic squeezing These drivers exert modest upward and lateral pressures that can overcome tensile strength of country rock and lead to eruption. Also, gas dissolved within that incoming magma is added to pre-existing magma, potentially increasing buoyancy and internal gas pressure to the point of failure.

4.12.6.1.1.(ii) Buildup of gas pressure as magma rises The solubility of gases decreases with decreasing lithostatic pressure, and if gas has no room to expand, its internal pressure will rise. The effect holds true regardless of whether magma is on one-way ascent or in the upwelling limb of two-way convection. Intrusion of fresh magma (Section 4.12.6.1.1.(i)) and any associated increase of gas pressure are probably the most common triggers of eruptions.

4.12.6.1.1.(iii) In situ buildup of gas pressure from crystallization, as in 'second boiling' Crystallization concentrates volatile phases in the residual, interstitial silicate melt. When those concentrations rise above gas solubility, bubbles will form and increase in either volume or pressure. This process, called 'second boiling', may trigger relatively small eruptions.

4.12.6.1.1.(iv) Mixing of fresh magma with pre-existing magma Mixing may add gas to the pre-existing magma, and can also change the oxygen fugacity (f_{O_2}). Sulfur gases in particular have a solubility minimum in temperature-f_{O_2} space, so if mixing pushes magma toward this minimum in gas solubility, more gas will exsolve (Kress, 1997). Addition of fresh mafic magma usually raises the temperature of the pre-existing magma, in turn decreasing the solubility of most gas in that pre-existing magma. Also, thermally or gas-induced density changes in the pre-existing magma cause gas-bearing magma from the base of the reservoir to rise and exsolve still more gas. Any combination of these effects can trigger an eruption.

4.12.6.1.2 External

4.12.6.1.2.(i) Regional earthquakes Following the 1992 Landers earthquake and a number of more recent earthquakes, local seismicity and strain increased at Long Valley and other large calderas. Possible mechanisms for such response are discussed in Hill *et al.* (2002), Manga and Brodsky (2006), and (Chapter 4.09). Many involve geothermal waters.

Although most regional earthquakes do not trigger eruptions, nor are most eruptions triggered by earthquakes, there are several dozen known cases in which a strong earthquake was followed 'shortly' by an eruption, for example, eruptions of Cordon Caulle (Puyehue) in Chile within hours of major earthquakes in 1921 and 1960 (Katsui and Katz, 1967). 'Shortly' is relative to the normal repose periods (interevent times) of the volcano, hours versus decades in the case of Cordon Caulle. In 1917, San Salvador volcano began steaming immediately and erupted for several months after an *M* 6.5 earthquake occurred just west of the capitol city. Another apparent case of earthquake-induced eruption is that of Pinatubo, Philippines, which became restless within weeks and erupted months after a nearby *M* 7.8 earthquake, after (as best as is known) being quiet for the preceding 500 years.

How might regional earthquakes trigger eruptions? Immediate triggering, as in Chile, may involve local faulting that relieves confining pressure (Lara *et al.*, 2004), shaking-induced bubble nucleation (Manga and Brodsky, 2006), or in low viscosity hydrothermal water, shaking-induced rise and pressurization of bubbles (Linde *et al.*, 1994). Eruptions that are delayed by weeks or months may be induced if a dense mat of crystals is dislodged from the roof or walls of a body of magma, sinks, and displaces an equal volume of gas rich magma upward (Bergantz and Ni, 1999; Hill *et al.*, 2002; Manga and Brodsky, 2006). An alternate explanation, invoked in the case of Pinatubo (Bautista *et al.*, 1996), is that volumetric compression (or elsewhere, dilatation) as a result of a nearby earthquake could squeeze (or draw) magma upward. In the case of Pinatubo, +0.1 MPa static compression throughout the crust, ~35 km-deep long period earthquakes and clear evidence of fresh basalt intrusion into dacite beginning ~6 months after the earthquake and 3 months before magmatic eruptions, suggest that basaltic magma was squeezed upward. It might have risen as a diapir (discrete batch) all the way from the lower crust or it might have been squeezed and coalesced from interconnected magma-filled cracks throughout the thickness of the crust.

4.12.6.1.2.(ii) Unloading, rapidly as by a massive landslide off a volcano's flank, or slowly, as by glacial retreat Decrease of confining pressure induces exsolution of volatiles and the related chain of increased buoyancy and/or increased internal pressure. A classic example of an eruption induced by a large landslide is that of Mount St. Helens in 1980. A statistical increase in eruptions following glacial retreat has been postulated in Iceland, but some caution is needed to distinguish between real postglacial increases in volcanism versus better postglacial preservation of deposits (Maclennan *et al.*, 2002; Van Vliet *et al.*, 2005) (see also Jellinek *et al.* (2004)).

Mastin (1994) made a related proposal: that rainfall-induced fracturing of hot rock can induce phreatic and eventually magmatic eruptions, working from the top down.

4.12.6.1.2.(iii) Earth tides Some volcanoes show a slight statistical preference to erupt on or around fortnightly tidal maxima (or, in a few cases, fortnightly minima), and a few show a preference to erupt on or near semidiurnal tidal maxima (or minima) (Dzurisin, 1980; McNutt and Beavan, 1981, 1987; Emter, 1997). Most volcanoes whose eruptions can be triggered by earth tides have relatively open conduits, relatively low-viscosity magma, and frequent, small eruptions, for example, Kilauea, Pavlof, and Mayon. The exact mechanisms by which earth tides trigger eruptions are not known, but possibilities include slightly enhanced exsolution during semidiurnal tidal decompression and/or a slightly increased rate of magma ascent. Hydrofracturing might also be involved, as seismicity in areas of high stress and high hydrothermal pore pressures can also correlate with earth tides (Glasby and Kasahara, 2001).

Statistical preference to erupt on tidal maxima (or minima) does not indicate a significant increase of eruption probability at each tidal cycle. Volcanoes that are not ready to erupt will not be affected by tidal cycles. Only those volcanoes that are ready to erupt, and that require only the slightest additional push, might show an influence from the tides.

4.12.6.2 Explosive versus Effusive Eruptions: Fragmentation of Magma

When magma finally nears the surface, its gas content and viscosity will exert important controls over the style of eruption. An overly simplistic view is that magma with high gas content will explode, producing ash and larger pyroclasts, while magma with low gas content will erupt gently (effusively), producing lava flows. A simple analogy is the initial opening of soda pop or champagne, charged with gas and explosive if opened suddenly. Once open, it starts to 'go flat' and, soon, can produce only a gentle overflow. Viscosity in its simplest role limits how easily gas can escape from magma, that is, how easily it can go flat. While gas is certainly a necessary component of explosive eruptions, and while viscous magma tends to erupt more explosively than fluid magma, neither all gas-rich magmas nor all viscous magmas will explode.

The explosivity of an eruption is an expression of the AMOUNT and RATES of exsolution and bubble growth (Sparks, 1978, 2003). As magma is rising up a conduit, these are controlled by the following:

- Its initial gas concentration and enroute leakage, the latter being controlled in large part by viscosity and the ability of bubbles to separate from melt. Wallrock permeability may also be a factor.
- Rate of confining pressure decrease, determined largely by ascent rate. Since ascent rate is not readily measured, it is inferred from the volume rate of extrusion. At Soufrière Hills volcano, explosivity was directly related to ascent and extrusion rate; faster ascent brought gas-rich but viscous magma to the near-surface faster than it could degas, so explosive eruptions ensued (Sparks and Young, 2002).

When magma reaches within several tens to hundreds of meters of the surface, the dominant controls on rates of exsolution and bubble growth become:

- the remaining volatile content, in melt and as discrete bubbles;
- any sudden rupture and release of confining pressure;
- bubble nucleation rate;
- melt viscosity, which limits
  - how fast dissolved volatiles can diffuse through melt into bubbles. Water in silicic to mafic melts diffuses slowly, from 10^{-1} to 10^{3} μm^2 s^{-1} (Zhang and Behrens, 2000; Zhang *et al*., 2003; Behrens *et al*., 2004);
  - how fast bubbles can expand;
  - whether glass as it is stretched by bubble expansion will fragment or continue to stretch; and
- permeability of any foam that develops.

As magma expands into foam, the melt walls of bubbles are stretched, usually faster than the melt can stretch. Two things can happen: either the magma blows itself apart, or it develops new shear permeability that bleeds off gas pressure before the magma can explode. In laboratory experiments with rapid decompression (intentionally rupturing a membrane in a pressure vessel), fragmentation occurs when bubbles expand rapidly and stretch bubble walls to their breaking point (Eichelberger *et al*., 1986; Cashman and Mangan, 1994; Klug and Cashman, 1996; Dingwell, 1998, 2003; Cashman *et al*., 2000; Spieler *et al*., 2004). A typical vol.% of bubbles before fragmentation is ~70%, reachable just by expansion of 0.1 wt.% gas from 100 MPa (3 km) to 0.1 MPa (surface), and far exceeded with higher wt.%'s of gas. If this vol.% of bubbles is exceeded rapidly, the stretched molten glass walls of those bubbles break quickly and the magma fragments (blows itself to pieces). However, if the decompression rate is slower, equivalent to an ascent rate of magma of less than 0.35 m s^{-1}, shearing-induced permeability may prevent explosions (Burgisser and Gardner, 2005).

Nucleation and diffusion rates are a minor issue if most of the gas is already in separate bubble phase (in extreme cases, magma can have 100 times more volatiles in pre-ascent bubbles than dissolved in the melt). Where there is not already a separate bubble phase, the presence or absence of other impurities like crystals is important. With impurities as bubble nucleation sites, exsolution can begin immediately upon saturation, so there is a long, slow, weakly explosive process of 'heterogeneous' bubble nucleation and growth. Without impurities, magma becomes oversaturated with gas and then bubbles nucleate 'homogeneously', suddenly and explosively (Mangan and Sisson, 2000).

Thus, the explosivity and eruption – reflecting the rates of bubble nucleation, growth, and magma fragmentation – is a function of initial gas content and melt and magma viscosity; presence or absence of tiny microlites to cause heterogeneous bubble nucleation; the presence or absence of a pre-existing discrete bubble phase; and the effective permeability of the magma, magma foam, and wallrock, and hence the efficiency with which magma can bleed off excess gas pressure shortly before and during eruptions.

4.12.6.3 Measures of Magnitude, Intensity (Volume DRE, Mass Discharge Rate, VEI)

The concepts of magnitude ('bigness') and intensity of eruptions are different than magnitude and intensity of earthquakes. Neither term has a precise definition in volcanology, but 'magnitude' refers broadly to the volume of products erupted and 'intensity' is the rate at which products are erupted. To be more precise, volume is usually expressed in terms of m^3 or km^3, with a qualifier to indicate whether that is the volume of deposit ('bulk volume', thickness times areal extent, including bubble and intergranular space), or the volume of magma that erupted ('DRE', or dense rock equivalent, corrected to a bubble free or prevesiculation volume of magma). Tsuya (1955) proposed the first volume-based magnitude scale for eruptions, and applied it to the sum of lavas and pyroclastic deposits.

Conversion from bulk volume back to DRE volume is made by multiplying bulk volume by the ratio of average density of deposit to the density of bubble-free magma. Typically, lavas have densities close to that of magma and pyroclastic deposits have densities around half that of magma.

Mass discharge rate is typically expressed in mass erupted/unit time, for example, $kg\,s^{-1}$. It is estimated by multiplying erupted volume times density and dividing by the duration of the eruption. A theoretical and empirical relation has been demonstrated in explosive eruptions between column height and mass discharge rate (Wilson, 1976; Wilson *et al.*, 1980; Carey and Sigurdsson, 1989), so the instantaneous intensity of an eruption can also be estimated from column height. Another empirical relation, between reduced displacement of eruption tremor and column height, was found to give another estimate of column height during eruptions (McNutt, 2000b).

To estimate volume and mass discharge rate requires mapping of deposits and/or detailed visual, satellite, or seismic observations during an eruption. This is possible for many modern eruptions but difficult to impossible for many older eruptions. To describe the explosive magnitude of older eruptions semiquantitatively, Newhall and Self (1982) proposed a volcanic explosivity index (VEI) that is based (in decreasing priority) on bulk volume of pyroclastic deposit, maximum column height, duration of eruption, and qualitative adjectives. Over all but the smallest eruptions, VEI rises one step with each order of magnitude increase in volume of deposit. As with earthquake and many other magnitude–frequency relations, there is a power-law decrease in frequency with increasing VEI (Simkin, 1993; Simkin and Siebert, 1994, 2000) (**Figure 7**). Not all eruptions that occur after a long repose (e.g., after hundreds to thousands of years) are of high VEI, but nearly all high VEI eruptions follow long reposes. Not counting short intervals between the start and climax of large explosive eruptions, it appears that a long repose is a necessary but not sufficient condition for a large explosive eruption, probably because it allows enough time for gas to accumulate in excess of saturation if the volcano is not leaky.

However handy VEI is as a shorthand among volcanologists, it is not a quantitative parameter, so bulk volume, mass discharge rate, and other details should be reported directly when known. Walker (1973) proposed a more quantitative classification of

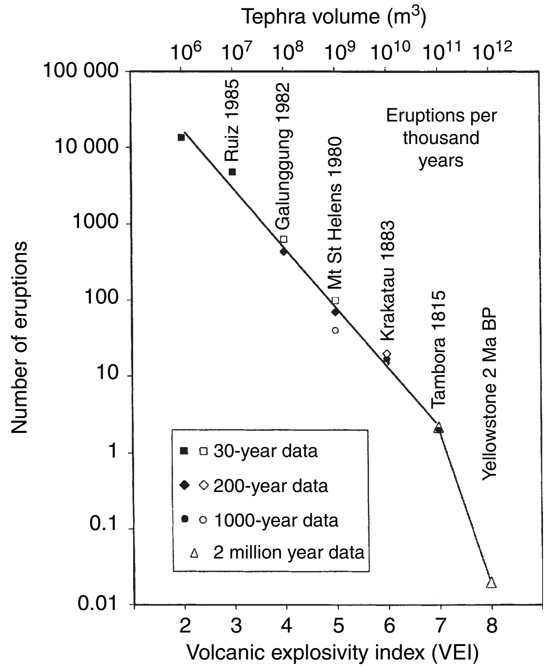

Figure 7 VEI magnitude–frequency plot. Over the range from VEI 2–8, each successively higher VEI is about 6 times less frequent than the next smaller category. A similar magnitude–frequency relation for earthquakes is well known. From Simkin T and Siebert L (1994) *Volcanoes of the World*, 349 p. Tucson, AZ: Geoscience Press.

explosive eruptions, based on two characteristics of their fall deposits:

1. fragmentation (F) = wt.% of pyroclasts <1 mm along dispersal axis, where thickness = 0.1 maximum thickness of deposit ($0.1\,T_{\max}$).
2. dispersal (D) = area enclosed by ($0.01\,T_{\max}$) isopach line of the tephra fall deposit.

Walker's fields of various eruption types on plots of (F) versus (D), as modified by Cas and Wright (1987) and reproduced here in **Figure 8**, correspond roughly to terminology that had already developed over preceding decades (plinian, strombolian, etc.), but define those eruption types more precisely.

Rutherford and Gardner (2000) and Cashman (2006) discuss the controlling effects of magma ascent rate, vesiculation, and degassing on explosivity. Ascent rates of $\sim$0.5 m s^{-1} or higher inhibit degassing enroute to the surface and result in higher explosivity; those of $\sim$0.05 m s^{-1} or lower encourage enroute degassing and result in relatively low explosivity. Cashman (2006) suggests correlation of magma supply rate to column height and dispersal (D), and proportion of erupted magma that is erupted explosively ('%tephra') to explosivity and fragmentation (F).

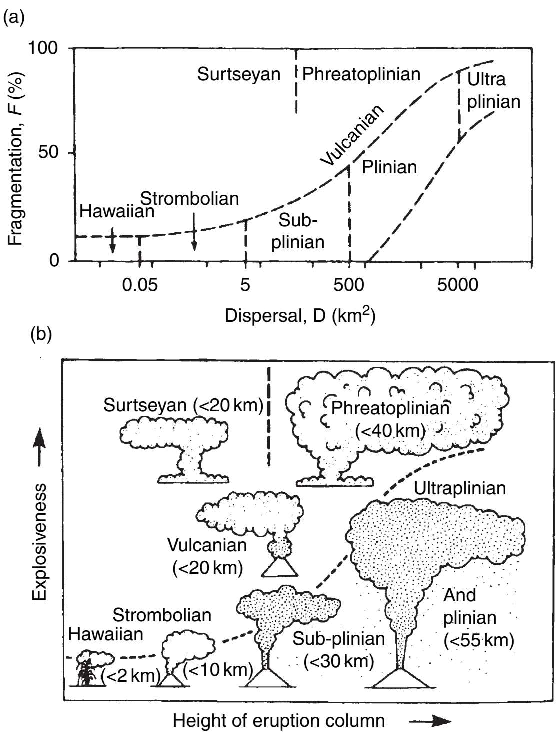

Figure 8 (a) Types of volcanic eruptions plotted in F–D space. F = fragmentation = wt.% of tephra fall that is <1 mm, along dispersal axis where $t = 0.1T_{\max}$; D = dispersal = area enclosed by ($0.01T_{\max}$) isopach line. T = thickness of tephra fall deposit. (from Cas and Wright 1987, after Walker 1973 and Wright *et al.* 1980). (b) Types of volcanic eruptions as functions of column height (itself, a proxy for mass discharge rate) and explosivity (as indicated by degree of fragmentation) (from Cas and Wright, 1987). Plot (a) is strictly descriptive; plot (b) is interpretive, though based on measurable proxies.

Further discussion of various measures of eruption 'bigness' may be found in Walker (1980), Carey and Sigurdsson (1989), and Pyle (2000).

4.12.6.4 Eruptive Phenomena

4.12.6.4.1 Lava flows and lava domes

When magma pours or oozes from the ground, it is renamed 'lava'. Given a slope and a viscosity low enough to flow, lava will form lava flows. Most lava flows are elongate ribbons of lava with thickesses of 10^0–10^1 m, widths of 10^1–10^2 m, and lengths of 10^2–10^4 m. Flows an order of magnitude large are not uncommon. Flood basalts such as the Columbia River Basalt or the Deccan Traps are erupted from dikes but form broad, thick sheets up to 10^6 m long and wide. Basaltic and other low-silica, low viscosity flows can travel considerable distances; more silicic and thus more viscous lava might travel only a few kilometers from its vent.

A number of different terms are used to describe the surface features and flow details of lavas. Two of the most common are Hawaiian terms, *'aa'* and *'pahoehoe'*. *Aa* lava has a blocky, rough surface while *pahoehoe* has a smoother, ropy surface. Various factors contribute to the difference but the main factors are viscosity and shear rate. Higher viscosities and higher shear rates favor *aa*.

Extremely viscous lava – too viscous to flow more than a few tens or hundreds of meters from the vent – mounds up and forms 'lava domes'. Growth by extrusion of one or a succession of viscous flows is called exogenous; growth by inflation of an existing dome is called endogenous. An endogenous dome that grows just beneath and dramatically lifts the ground surface – soil, trees and all – is called a cryptodome ('hidden dome').

Nearly all lava flows travel slowly enough that people can walk out of harm's way. Buildings and other immovable objects are generally destroyed, either by fire or by mechanical bulldozing.

Thick, extensive flows of obsidian (volcanic glass) bear special mention. None has been observed closely during formation, but they pose an interesting paradox. Glass forms by rapid quenching of melt upon eruption – usually, within seconds or minutes of eruption. The glassy character of obsidian implies rapid quenching. But many obsidian flows are so thick that they would have been expected to cool slowly, crystallizing rather than being quenched. Also, most obsidian flows are highly silicic and would have been expected to be more viscous than their broad dimensions suggest. Eichelberger *et al.* (1986) suggested the most likely explanation: that magma was erupted as a permeable foam, not unlike a good foam on a beer. As gas escaped, the foam collapsed into a quenched, dense glass that spread just a short distance farther. This gives the deceptive appearance of obsidian having flowed as lava over the entire distance from its vent, whereas in fact it was a much lower-viscosity foam over most of that travel distance.

4.12.6.4.2 Tephra fall

When an eruption is explosive, magma is blown out as fragments ranging in size from fine dust or sand ('ash') to large blocks or volcanic bombs. These fragments are called 'pyroclasts' (from the Greek for fire and broken) or 'tephra' (from the Greek for volcanic ash). Large fragments behave as ballistics, following parabolic trajectories and landing within a few hundred meters (rarely, up to a few kilometers) from the vent. Smaller fragments can be carried farther aloft by vigorous thermal convection. Eruption clouds from the largest explosive eruptions convect up to 50 km above the vent; small and moderate size eruption clouds (from VEI 1–3 eruptions) rise 0.1–20 km above the vent.

At altitude, ash or tephra-bearing clouds are typically sheared by prevailing winds and tephra rains out as a curtain elongate in the direction of the prevailing wind. For the most part, it rains out in order of decreasing particle size and density, following Stokes' law, but, if moist or electrically charged, the finest grain sizes clot together and fall out earlier than they otherwise would.

Light tephra fall (up to a few centimeters) does not directly threaten the lives of people on the ground, but can cause serious damage to computers and other electronics, electrical distribution grids, heating-ventilation-air conditioning (HVAC) systems, storm drains, and vulnerable crops. Tephra fall in excess of 10 cm, especially if wet, is the threshold beyond which structural damage and roof collapse is possible. Ash clouds pose a special hazard to modern jet aircraft, whose engines run at temperatures high enough to melt ingested ash. In the cooler parts of the turbine, molten ash glazes over fuel and air nozzles, engines shut down, and planes become gliders. Serious encounters, including near crashes of jumbo jets, have led to a new, global 24/7 (around-the-clock) network of Volcanic Ash Advisory Centers (Casadevall 1994, Romero 2004, NOAA 2005).

4.12.6.4.3 Pyroclastic flows and surges

Pyroclastic flows are avalanches of hot fragments and gas down the slopes of volcanoes, typically at 20–50 m s^{-1}. Most form by collapse of an eruption column, collapse of a dense slug of debris erupted just a few hundred meters above a vent, or collapse of the toe of a lava flow or dome growing on a steep slope. Pyroclastic flows have a relatively dense basal avalanche that exhibits laminar flow, and an accompanying, expanding turbulent ash and steam cloud. Ash from the basal avalanche is elutriated just like the dust cloud from a normal rock avalanche, and is almost as hot as the basal avalanche. Basal avalanches tend to follow valleys but can climb over some ridges; the overriding ash clouds are less constrained by topography. Runout distances are typically 1–20 km but in extreme cases can reach >100 km. Because of their high speed and high temperature, pyroclastic flows are generally lethal. The worst tragedy from a pyroclastic flow occurred in 1902 on the island of Martinique, where nearly all 29 000 residents of St. Pierre died in one short-lived but deadly flow.

Pyroclastic surges are relatively dilute cousins of pyroclastic flows, and indeed very similar to the upper, turbulent facies described above. The blast that leveled forest at Mount St. Helens was, in most senses, a pyroclastic surge. Increasingly, the term pyroclastic density current (PDC) is being applied to pyroclastic flows and surges.

4.12.6.5 Eruption Forecasts

Some volcanologists distinguish between forecasts (relatively general, often probabilistic statements based on past history) and predictions (more specific deterministic statements taking special account of monitoring and interpretation of current process) (e.g., Swanson *et al.*, 1983). In contrast, we see a spectrum of statements that are all probabilistic to some degree and thus 'forecasts', based on monitoring, past history, and interpretation of current process.

4.12.6.5.1 Long-range forecasts

Forecasts for eruptive activity within coming months to decades are typically based on eruptive history as seen in the geologic record and in historical documents. The past is the key to the future. In a few cases there might be robust enough clustering or other temporal patterns to infer something other than a Poisson distribution, but these are rare. From geologic and human history, we can outline the types, frequency, and probable extent of future eruptions. Such forecasts are heavily probabilistic. Hazard maps are one form of long-range forecasts, as are estimates of the probability that a particular site (e.g., a critical facility) will be hit by eruptive events within a specified time. Long-range forecasts are used for land-use planning, emergency response planning, facility siting, and other planning decisions.

4.12.6.5.2 Short-range forecasts

Forecasts for eruptive activity within minutes to months are based on a combination of geophysical, geochemical, and visual monitoring plus background information from the long-range forecast. Specific patterns of seismicity, ground deformation, gas emission, and other changes can indicate specific steps along the way toward eruption. In some cases, progress toward eruption gives a monotonic or exponential increase in monitored parameters; in other cases, values of a parameter might suddenly decrease, for example, seismic quiescence shortly before some eruptions. Patterns vary from volcano to volcano and, in some cases, from eruption to eruption, and the best short-range forecasts will be those that weigh empirical experience and real-time interpretation of processes together. Short-range eruption forecasts are analogous to weather forecasts.

When a volcano that is preparing to erupt is known to have exhibited a wide range of explosive scales in the past, can the explosive magnitude (VEI) of its impending eruption be forecast? At present, there is no tried and true method for forecasting VEI. Indeed, precursors of the giant 1991 eruption of Pinatubo were indistinguishable from precursors to much smaller eruptions until well into the eruption, just 24 h before its climax (Harlow *et al.*, 1996; Power *et al.*, 2001; Newhall *et al.*, 2002).

One approach to estimating explosive ‘potential’ (i.e., maximum possible VEI) of an impending eruption is to calculate a gas budget since the last time gas-rich magma was reamed out of the volcano (e.g., since the time of the previous VEI ≥ 3 eruption). If during the latest repose magma has degassed apace with inferred fresh volatile supply, the author’s unpublished estimates suggest that the maximum VEI will be 4 and most probably less, but if volatiles have accumulated into a discrete volatile phase a VEI 3 or higher is almost certain and the maximum potential VEI could be 5 or higher. This approach is being tested as new eruptions permit, but it has not yet been validated. Within the bounds of maximum possible VEI, the actual VEI will depend greatly on magma ascent rate so any indication of rapid magma ascent, including rapid escalation of unrest, will favor higher rather than lower VEI.

Short-range forecasts can be of narrative form (what is expected, when, where, etc.), in the form of alert levels (e.g., color-coded alert levels; Gardner and Guffanti, 2006), or in the form of forecast windows that might be weeks long in an early or intermediate stage of unrest and days to hours long if seismicity and other parameters point to an imminent eruption (Swanson *et al.*, 1983).

One way to integrate short- and long-range forecasts is through use of event trees (Newhall and Hoblitt, 2002; Aspinall *et al.*, 2002; Marzocchi *et al.*, 2004, 2005; Hincks *et al.*, 2005). The thought process to estimate probabilities at each node of an event tree, and make Bayesian updates of these with each new piece of information, can be very helpful in guiding scientific discussion and further work.

4.12.6.5.3 Successes and false alarms

We have already seen that volcanoes give signals that magma is rising or otherwise preparing to erupt. Some particularly gratifying successes, such as at Pinatubo in 1991, have raised expectations. On the other hand, some eruptions still occur without being forecast, from volcanoes that are not monitored at all or at which the precursors were too subtle for current monitoring. Relatively fewer cases occur in which warning signs are detected but underestimated or otherwise misinterpreted.

A bigger challenge for volcanologists is an overabundance of precursors, and how to distinguish those that will lead to eruption from those which will not. Signs that magma is rising toward the surface are similar regardless of eventual outcome. At an early stage of unrest, we may see unmistakable signs of magma rising but not know yet whether it will stall or eventually erupt, and intrusion may continue until practically the ‘last minute’ before stalling, as occurred at Soufrière Guadeloupe in 1976 (Feuillard *et al.*, 1983; Fiske, 1984). Empirically, magma that is rising rapidly, with corresponding accelerating unrest, is more likely to erupt. Nishimura (2006) modeled accelerating ground deformation from undegassed magma, whereas degassed magma produced constant-rate deformation. There may be other differences between successful versus failed eruptions, also related to magma ascent rate and gas overpressures, but these are still in the research arena.

In 1999, precursors warned of an approaching eruption of Tungurahua volcano in Ecuador. The town of Baños, extremely vulnerable, was evacuated. The volcano did erupt and still erupts, but until 2006 the eruption was more spectacular than dangerous. The townspeople returned home after only a few months of evacuation and were loathe to ever evacuate again. It is very hard for a lay public to understand how degassing will increase magma viscosity and explosive potential, and that it will be very hard for the scientific monitoring team to know which of the daily explosions will be bigger than the rest and produce a pyroclastic flow. It is equally hard for the public to visualize that the deposits on which their town is built are from pyroclastic flows not much older than the town! Fortunately, stronger explosive activity in 2006 had clear precursors, clear warnings were issued, and those at highest risk (just outside Baños) did evacuate in time.

A stark illustration of another false alarm dilemma comes from Vesuvius and towns around its base. The population is so large (approximately 550 000 in the 'red' danger zone) that an evacuation will require at least 1 week. Yet, it is rarely possible to be sure that a volcano will erupt until just hours to days before an eruption. The earlier the evacuation, the better the chances that it will be successful, but also the greater the chances of false alarm.

4.12.7 Erosion of Volcanoes

Here is a brief introduction to hydrothermal alteration, sector collapse, and other processes that erode volcanoes.

4.12.7.1 Hydrothermal Alteration

Every volcano, to one degree or another, has a hydrothermal system – a system of groundwater heated by magma and circulating convectively through fractures and porous rock. Hydrothermal fluid that is heated directly by volcanic gases is acidic, and reacts with country rock to form clays and other alteration minerals plus a more neutral hydrothermal fluid. Largely hidden from view, hydrothermal alteration softens and weakens a volcanic edifice and makes it prone to large-scale collapse. Some volcanoes, for example, Yellowstone and Rainier have unusually extensive hydrothermal alteration. At Rainier, hydrothermal alteration weakens the steep slopes and has been partly responsible for some giant landslides (flank collapse) that evolved downslope into clay-rich debris flows.

4.12.7.2 Sector Collapses

Two distinctive geomorphic features – horseshoe-shaped craters and fields of small hills ('hummocks') at their feet – were recognized at a handful of volcanoes before 1980. Some were thought to be the result of large, possibly slow-moving landslides; others, including small hills north of Mount Shasta, were not understood at all. Immediately upon the giant landslide of Mount St Helens in 1980, hundreds of such pairs were found and understood at once to have formed in the same, catastrophic way (Ui, 1983; Siebert *et al.*, 1987; Siebert, 1996). Subsequent work at Mount St. Helens confirmed at least three previous collapses, two from the north flank about 3 ka BP and another of the south flank ~20 ka BP (Hausback, 2000).

Although the first parallels were noticed from steep-sided arc stratovolcanoes, workers in Hawaii soon realized that even gently sloping shield volcanoes have also collapsed, in landslides from hundreds and even up to 5000 km^3 (Moore *et al.*, 1994) Just one 10 min long event – the landslide of Mount St. Helens – changed the paradigm of thinking of volcanoes as fundamentally stable with rare exceptions to fundamentally unstable edifices that usually fail at least once in their lifetime.

4.12.7.3 Concentrated Sediment-Water Flows (Lahars)

Concentrated flows of volcanic sediment and water, called by the Indonesian term 'lahars', form by the following:

- eruption through crater lake (e.g., Kelut, Indonesia), followed by scouring of downstream deposits;
- sector collapse of water-saturated cone, followed by either downstream evolution of avalanche into

debris flow (e.g., Ontake, Japan; Mount Rainier, USA), or halting and compaction dewatering of the debris avalanche deposit (e.g., Mount St. Helens, USA);

- rapid melting of snow and ice by pyroclastic flows and incorporation of sediment from the pyroclastic flows and also from scoured downstream deposits (e.g., Nevado del Ruiz, Colombia);
- overtopping and rapid scouring failure of natural dams (e.g., Ruapehu, New Zealand; Pinatubo, Philippines); and
- heavy rainfall on fresh pyroclastic deposits (e.g., Pinatubo, Philippines). Fine ash on the surface reduces infiltration rates by an order of magnitude and watersheds become prone to sediment-laden flash floods, just as in arid climates or following forest fires.

The first three originate only during eruptions. The latter two can be syneruptive or posteruptive.

Lahars behave in ways quite unlike normal streamflow. Bulking up from clear water streamflow to >20 vol.% sediment creates hyperconcentrated flows, and further bulking up to >60 vol.% sediment creates debris flows, both termed lahars. Hyperconcentrated flow is dilute enough that it has no appreciable yield strength, yet it transports considerable volumes of sediment in short periods. Hyperconcentrated flow is turbulent and deposition from that flow is progressive (grain by grain) though fast enough that classical stream deposit features such as cross-bedding are absent. In contrast, a debris flow exhibits significant yield strength and laminar flow, the latter helping it to flow fast and far. Debris flows are analogous to fast-moving slurries of concrete, and the combination of their density and speed allows them to plow right through (or lift and float) buildings, bridges, and anything else in the way. Deposition from debris flows can range from progressive to sudden, *en masse* deposition, and resulting deposits are massive and without any internal stratification.

Lahars generate relatively high-frequency tremor and dense seismic networks can detect and track lahars. Inexpensive 10–300 Hz geophones installed on riverbanks (LaHusen, 2005) were used with good success to detect and quantify lahars at Pinatubo (Marcial *et al.*, 1996; Tungol and Regalado, 1996), and are presently deployed at a number of volcanoes including Rainier and Ruapehu to warn of approaching lahars.

4.12.7.4 Normal Streamflow, Sediment Issues, and Flooding

Lahars can transport very large amounts of sediment in the years following a large explosive eruption. In the case of Pinatubo, >3 km^3 was eroded from the roughly 5–6 km^3 of 1991 pyroclastic debris on the slopes of the volcano (Tungol, 2002). Most of this sediment was transported as lahars in the first 5 posteruptive years, but by 1997, 25% of declining annual sediment transport was by normal streamflow (Hayes *et al.*, 2002). Except for one large lake-breakout lahar in 2002, nearly all sediment transport since 2000 has been by normal streamflow. High sediment yields were also observed in the decades following the 1980 eruption of Mount St. Helens (Major, 2004), where torrential rains are rare and most sediment was carried by normal streamflow.

Copious loads of sediment are deposited in channels on the surrounding lowlands, sharply reducing their capacity to carry even modest normal runoff. Lowland flooding is a serious by-product of lahars and other heavy sediment transport after an explosive eruption.

4.12.8 People and Volcanoes

Demographers have long-recognized dense populations around tropical and temperate volcanoes. Reasons include fertile soils, healthy cooler climates above tropical lowlands, and esthetic and recreation appeal. The result is an unhealthy juxtaposition of large populations with dangerous volcanic phenomena, and an extra challenge for those who must forecast eruptions and avoid false alarms (Ewert and Harpel, 2004).

4.12.8.1 Risk Mitigation

The principal tools for long-range risk mitigation are hazard and risk maps (the latter, integrating hazard, vulnerability, and exposure), land use management, public education and dialog about volcanic risks, and early development and testing of emergency plans. The principal tools for immediate risk mitigation are short-range eruption forecasts, evacuations, aircraft reroutings, and other emergency measures such as installing ash filters and covering storm drains.

Can people, buildings, and infrastructure be protected against ashfall? In general, yes, though at some cost. Can flow hazards be stopped by engineering measures? The answer is a qualified ‘sometimes’. If topography is favorable, lava flows can be slowed or diverted away from critical structures. If topography is favorable and considerable time and money is invested in advance, lahars can be slowed and sediment can be sequestered by dams and related structures. Even bunker-like shelters from pyroclastic flows have been tried. However, the success rate of such efforts to control volcanic flows is low and costs are high. Usually, it makes more sense simply to avoid building in the path of volcanic flows and to move out of the way if threatened.

4.12.8.2 Probabilities and Acceptable Risk

Communities and individuals generally resist evacuation and other costly mitigation measures, especially if eruptions of that volcano are infrequent or if forecasts are conspicuously uncertain. In some cases, this stems from ignorance or denial in the face of clear and present danger, but in other cases it stems from a calculated decision to accept the risk from the volcano rather than take socially and economically costly mitigation measures. Quantification of volcanic risk, even with high uncertainties, usually helps citizens and officials to decide whether that risk is acceptable to them. Further discussion of this can be found in Newhall and Hoblitt (2002) and Aspinall *et al.* (2002), Hincks *et al.* (2005).

4.12.9 Future Directions

Brodsky (2004) offered interesting thoughts on outstanding challenges in volcanology. The author concurs, and adds these additional thoughts and questions.

What causes magmas to pond and accumulate in reservoirs with tops between 3 and 6 km?

This has been termed a region of 'neutral buoyancy' by Ryan (1987). Is this the level at which decompression degassing increases viscosity sharply and causes magmas to stall? Or is 3–6 km a region in which rising magma encounters significant groundwater and is thus quenched and slowed and/or can spread by hydrofracturing of country rock? Is this a level at which lithostatic pressure that is squeezing magma upward becomes too low to force open new fractures toward the surface? Or is it a level where exsolution of volatiles forms a barrier of low-density magma that blocks ascent of following magma? Understanding when and where each of these processes is dominant may help us to understand periodicity of shallow intrusions and eruptions, and to understand how long-dormant volcanoes reawaken.

Is magma supply from the mantle and lower crust continuous and are eruptions thus controlled by buffering in shallow reservoirs or control by near-surface valves? Or, alternatively, is supply into volcanic roots episodic and, itself, a major control on when and how volcanoes erupt?

One way to address this question is by InSAR monitoring of broad deformation, indicating resupply of magma into volcanic roots. With C-band radar, only a limited number of unvegetated high-latitude or high-altitude volcanoes could be monitored. However, now that L-band radar is once again available from Japan's ALOS/PALSAR satellite, we will see good data from many volcanoes, even in the tropics.

Supply of magma from depth is probably episodic if viewed to the resolution of hours to years. But as the viewing resolution blurs to decades or centuries, supply might be more or less steady. One hypothesis to be tested is that if supply is episodic and less frequent than the time required for magma in the conduit to degas, cool, and solidify after an eruption, then eruptions will occur only after fresh magma resupply and perhaps only after repeated episodes of resupply. If supply is more or less steady on the timescale of solidification of magma in conduits, then solidification will occur only when the rate of magma supply is relatively low. When supply rates are higher, conduits will stay open and filled with convecting magma, and eruptions can occur at anytime or with just the slightest fresh resupply.

How, in detail, does magma degas, and how does this control eruptive style and magnitude?

This topic was introduced earlier in two contexts: leakage during magma ascent from depth if that ascent is slow enough, and leakage during the moment of eruption (modulated expansion of magma foam) if the decompression is slow enough to allow development of shear permeability before the magma explodes. Both of these topics invite further laboratory experimentation and field confirmation. There are great opportunities for more interaction between those who measure gas emissions in the field, those who study degassing and fragmentation in the laboratory, and those who reconstruct it from the petrology of erupted products.

Specifically, field and lab investigators can test the working hypothesis that volcanoes which are plugged tight (i.e., with minimal inter-eruption passive degassing) accumulate gas in excess of saturation and have high explosive potential, especially if their eventual ascent to the surface is rapid. Did all magmas that fed plinian eruptions have an excess volatile phase? Data with which the author is familiar suggest yes, but any exceptions will be even more interesting! At volcanoes with open conduits, what causes the occasional explosions that are much larger than normal? Is it simply that the convection is speeded up, or is homogeneous nucleation also involved?

Do early bubbles in gas-charged magma, acting as potential sites for further heterogeneous nucleation, ever reduce explosivity? In relatively viscous magmas, what is the relative importance of foaming versus shear permeability as a means to efficiently degas magma? How was strong degassing like that of the past decade at Popocatépetl sustained even when little or no magma was erupting? How does Iwo-jima degas so efficiently and, in particular, might groundwater and hydrofracturing play a role? What independent evidence can be found to support or reject lava-lamp style conduit convection, and associated degassing at the top of an inferred 'convection cell?' Combined geologic, geophysical, and geochemical studies are needed to improve forecasts of explosive magnitude (VEI) so that evacuations and other measures are of appropriate scale. For example, at Tungurahua volcano in Ecuador in 1999, the geologic record warned of pyroclastic flows and the population of Baños was briefly evacuated, but no major pyroclastic flows occurred until 2006. What can be monitored in small, strombolian explosions that will foretell of whether and when the eruption will escalate into a more explosive one? The same general issue of forecasting VEI arises during unrest at any volcano that exhibits a range in the VEIs of its eruptions. Can estimation of a gas budget, as in **Figure 3**, help in estimating the VEI of an impending eruption?

At what timescales are eruptions stochastic versus predictable?

Eruptions may seem to occur randomly or predictably, depend on the timescale. Many variables influence whether a volcano will erupt in a given time period (e.g., history of magma supply, gas leakage, local stress–strain relations), so on longer timescales, eruptions may indeed be stochastic. At the same time, systematic buildup of precursors before eruptions argues for predictability. On timescales of weeks to months, volcanoes with open conduits, erupting frequently, may appear to be stochastic while those with closed conduits, erupting infrequently, appear predictable. On the other hand, if one looks closely enough at those with open vents, they, too, exhibit precursors and predictability on timescales of seconds to minutes. For forecast purposes, volcano monitoring must be scaled to the openness of the vent.

How can we best estimate critical intrinsic parameters (e.g., viscosity, water content, overpressure) from observable parameters (seismicity, deformation, gas leaks, etc.). Can active probes help?

A conundrum facing volcanologists is that the parameters which directly indicate explosive potential (volatile content of magma, gas overpressures, magma viscosity) cannot be measured before eruptions. What can be measured – seismicity, deformation of the ground surface, volumetric strain, gas emissions, ground or fumarole temperatures, and a few other parameters – are at best indirect indicators of the state of the magma.

Two general approaches might help. One is to track the volcano's changes (seismic, ground deformation, volumetric strain, gravity, gas) in response to known, repeating natural signals such as earth tides. In response to tidal decompression, we might expect greater expansion response if shallow magma is charged with a separate volatile phase (high explosive potential) than if it is gas poor. Less viscous magmas might respond more quickly than more viscous magmas. Microseismic signals and hour-to-hour gas emissions might vary with volatile concentration, state, and magma permeability.

As a second approach, we might try artificial sources. As a magma becomes charged with bubbles, and especially as those expand as gas, seismic velocities through the magma will slow down, but the general problem is how to focus a repeating source through relatively small target bodies of magma? Stacking of repeating artificial signals, such as in the Japanese ACROSS project (Ikuta *et al.*, 2002; Ikuta and Yamaoka, 2004), is promising but has not been proved at volcanoes yet. Other ideas are needed!

How can we know if an impending eruption from a large volcanic system will be just a small aliquot or a colossal eruption?

Large magnitude eruptions are thought to grow from small ones by runaway vesiculation (bubble nucleation and growth), just as large earthquakes may grow from small ones by cascading rupture. Volcanic gas is concentrated in the uppermost parts of magma reservoirs. An eruption might tap just a small volume of gas-charged magma from the top of a reservoir, if the gas content of underlying magma is too low and/or the vent and conduit get clogged quickly. On the other hand, if the gas content is high, and the bleed rate is too low, an explosive eruption may escalate. Runaway vesiculation may increase the volume and mass discharge rate of an eruption to exceptionally high values. Hoblitt *et al.* (1996) interpreted the pattern of more and more closely spaced, smaller and smaller eruptions of Pinatubo in 1991 leading up to the runaway climactic phase as a result of tapping magma with increasing gas concentrations, which systematically reduced the time between explosive pulses. At the same time, the vent was also getting reamed out, also decreasing resistance to eruption. Eventually, the pulse that began at 1342 h on June 15 tapped magma that was rich enough in gas to sustain and escalate the discharge. Harlow *et al.* (1996) reported a parallel escalation in the energy release of shallow long-period earthquakes. The runaway eruption continued until the eruption tapped into deeper, gas-depleted magma, and/or tapped so deep in the conduit that overlying foam and debris shut off vesiculation, or the conduit was squeezed shut with the same effect.

We do not know if the patterns observed by Harlow *et al.* (1996) and Hoblitt *et al.* (1996) are characteristic of the buildup to large (plinian) explosive eruptions. If they are, then we might have hours of forewarning of larger explosive events. From a practical point of view, populations must already be evacuated on the basis of any geologic evidence of past plinian eruptions, as the confirming indications from monitoring might come too late for evacuations.

4.12.10 Conclusion

This review began with analogies between volcanoes and fault zones. Both are subject to stress and strain, both produce distinctive patterns of seismicity, and both exhibit 'stick-slip' behavior with similar power-law magnitude–frequency relations. In the course of this review we have also seen a further analogy: that the variable degassing pattern of volcanoes, some open and freely degassing versus others plugged tight, is analogous to creeping versus locked segments of faults. Gas accumulation in magma is analogous to strain accumulation on faults, steadily accumulating unless released by passive leakage or creep. Estimation of explosive potential demands an accounting of gas accumulation versus passive degassing, over the timeframe since the previous explosive eruption (typically, years to decades) and also over the timeframe of current magma ascent (typically, hours to weeks).

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