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RESEARCH ARTICLE

Failed magmatic eruptions: late-stage cessation of magma ascent

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Abstract When a volcano becomes restless, a primary question is whether the unrest will lead to an eruption. Here we recognize four possible outcomes of a magmatic intrusion: "deep intrusion", "shallow intrusion", "sluggish/viscous magmatic eruption", and "rapid, often explosive magmatic eruption". We define "failed eruptions" as instances in which magma reaches but does not pass the "shallow intrusion" stage, i.e., when magma gets close to, but does not reach, the surface. Competing factors act to promote or hinder the eventual eruption of a magma intrusion. Fresh intrusion from depth, high magma gas content, rapid ascent rates that leave little time for enroute degassing, opening of pathways, and sudden decompression near the surface all act to promote eruption, whereas decreased magma supply from depth, slow ascent, significant enroute degassing and associated increases in viscosity, and impingement on structural barriers all act to hinder eruption. All of these factors interact in complex ways with variable results, but often cause magma to stall at some depth before reaching the surface. Although certain precursory phenomena, such as rapidly escalating seismic swarms or rates of degassing or deformation, are good indicators that an eruption is likely, such phenomena have also been observed in association with intrusions that have ultimately failed to erupt. A perpetual difficulty with quantifying the probability of eruption is a lack of data, particularly on instances of failed eruptions. This difficulty is being addressed in part through the WOVOdat database. Papers in this volume will be an additional resource for scientists grappling with the issue of whether or not an episode of unrest will lead to a magmatic eruption.

Keywords Magmatic intrusions · Volcanic unrest · Eruption forecasting · Factors influencing magmatic eruption

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Introduction

When a volcano becomes restless, one of the primary questions asked by and of observatory scientists is whether the unrest will lead to an eruption. To address this question, many factors are taken into account, including seismicity, deformation, gas emissions, the geologic and/or historic record of past eruptions, and analogues from the global record of volcanic unrest and eruptions. In several instances, two of which are detailed in this special volume (Crider et al. 2011; Nishimura and Ueki 2011), indicators of unrest have become significant enough to prompt scientists to issue public statements; to engage with the media, local land managers, public officials; and to engage in other activities in preparation for a potential eruption, only to have the unrest de-escalate without magma reaching the surface. Such instances can threaten the credibility of scientists in volcano observatories and public officials responsible for evacuations and access restrictions, and can also bring unwanted economic and social disruption.

Commonly after unrest begins, observatory scientists search for analogous instances of unrest observed at other volcanoes. However, one problem in the use of analogues to assess the likelihood of a future eruption is that the global volcanism record is biased towards incidents when unrest culminated in an eruption. When unrest leads to an eruption, scientific interest is piqued, monitoring capabilities are enhanced, and special collections of papers appear in scientific journals or monographs, e.g., Mount St. Helens, Washington, 1980 (Lipman and Mullineaux 1981), Unzen, Japan, 1990–1995 (Nakada et al. 1999), Soufrière Hills, Montserrat, 1995-present (Young et al. 1998; Druitt and Kokelaar 2002). When unrest instead dies away, so too does scientific interest. Few if any additional monitoring stations are installed, and few if any studies are published in scientific journals. Although the Smithsonian's Global Volcanism Program attempts to catalog all modern episodes of unrest, episodes not leading to eruption may be under-reported, or not reported at all, by local observatories. Such a bias in the record can result in incorrect estimation of probabilities that a given episode of unrest will lead to an eruption.

Over the last several decades unrest at stratovolcanoes that might have led to an eruption has included various combinations of increased degassing and thermal output, phreatic explosions, shallow earthquake swarms (some with felt and/or low-frequency events), and notable ground deformation. Examples include Soufrière Guadeloupe, 1975–1976 (Hirn and Michel 1979; Dorel and Feuillard 1980; Feuillard et al. 1983; Beauducel and Besson 2008), Akutan, Alaska, 1996 (Lu et al. 2000; Power et al. 2008), Mount Baker, Washington, 1975 (Crider et al. 2011), Iliamna, Alaska, 1996 (Roman et al. 2004; Roman and Power 2011), Iwate, Japan, 1998 (Nishimura and Ueki 2011), Parícutin, Mexico, 2006 (Gardine et al. 2011), Fourpeaked, Alaska, 2006 (Gardine et al. 2010), Mount Spurr, Alaska, 2004–2006 (Power 2004), and others. A few cases are well-studied, but many are poorly documented in the literature and so details of such events are often unavailable for comparison during an unrest episode. Thus, a principal goal of this special volume is to present retrospective studies of episodes of volcanic unrest at stratovolcanoes that ultimately failed to produce a magmatic eruption. In this introductory paper we define a failed eruption, discuss why some restless magmatic systems fail to erupt, and review possible discriminants that may allow future identification of episodes of magmatic unrest that are unlikely to culminate in eruption.

Failed eruptions: a definition, and general characteristics

A magmatic intrusion starting at depth can remain at depth; stall just before reaching the surface; erupt in a sluggish, viscous extrusion; or erupt rapidly, often explosively (Figs. 1 and 2). Here, we define a "failed eruption" as an instance in which magma has intruded to shallow depths (generally < 2–3 km below the surface), accompanied by anomalous seismicity, deformation, degassing, and in some cases even phreatic explosions, but ultimately fails to reach the surface. Because we emphasize magma, this definition classifies eruptions featuring only phreatic activity as failed eruptions.

Principally to narrow the scope of this special volume, we have elected to not include discussion of large caldera systems such as Yellowstone (Wyoming), Long Valley (California), and Campi Flegrei (Italy), which also experience episodes of unrest (Newhall and Dzurisin 1988) that would qualify as failed eruptions under our definition. Beyond calderas, most known failed eruptions occur at arc stratovolcanoes, although failed eruptions may also occur in monogenetic fields (e.g., Gardine et al. 2011), shield volcanoes (e.g., Cervelli et al. 2002), and rift settings (e.g., Calais et al. 2008; Pallister et al. 2010).

We recognize three main variants in the nature of unrest associated with failed eruptions: (1) instances in which volcanoes exhibit only strong steaming, changes in gas flux, and/or elevated fumarole temperatures, without other notable unrest e.g., Vulcano, 1977–2006 and beyond (Martini 1993; Granieri et al. 2006); (2) instances in which seismic swarms, inflation, and other evidence of pressure buildup simply stops, abruptly or slowly, e.g., Akutan, 1996 (Lu et al. 2000), Iliamna, 1996 (Roman et al. 2004; Roman and Power 2011), and Parícutin, 2006 (Gardine et al. 2011); and (3) instances in which unrest culminates in phreatic explosions, e.g., Soufrière Guadeloupe, 1976 (Feuillard et al. 1983; Beauducel and Besson 2008) and Fourpeaked, 2006 (Werner et al. 2011; Gardine et al. 2010). Barberi et al. (1992) reviewed documented instances of phreatic explosions and found that a majority of phreatic explosions were not followed by a magmatic eruption. Another common form of unrest involving deep long-period earthquakes and/or deep-focus inflation that stops after little or no shallow unrest, e.g., Three Sisters, Oregon, 1997–2006 (Wicks et al. 2002; Dzurisin et al. 2006), Mount Fuji,

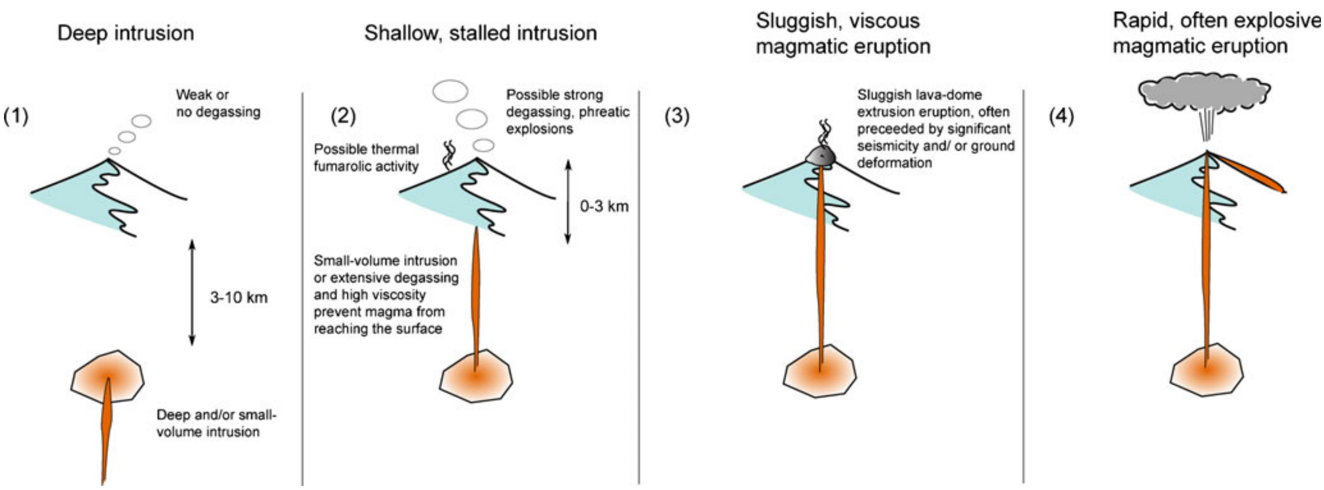

Fig. 1 Illustration of the four possible outcomes of unrest—deep intrusion; shallow, stalled intrusion; sluggish, viscous magma extrusion; and rapid, often explosive magmatic eruption

Japan, 2000–2001 (Ukawa 2005), is not considered here to represent a failed eruption because magma apparently did not reach the shallow intrusion stage.

As magma rises, a variety of processes act to either promote or hinder its reaching the surface (Fig. 3). Magmatic eruption is promoted by: (a) high gas content at depth and minimal degassing enroute to the surface (due, for example, to rapid ascent rate); (b) an increase in driving forces, such as increasing gas pressure resulting from gas exsolution as magma reaches shallower depths and/or a fresh supply of magma from depth, particularly high temperature or more gas-rich magma; (c) opening of pathways (e.g., through opening of pre-existing faults due to tectonic strain or magma intrusion); or (d) sudden decompression of magma at shallow levels (e.g., through rapid unloading from edifice/lava dome failure or a phreatic explosion at the surface). Magmatic eruption can be hindered by: (a) a loss of driving force (drop in gas pressure or magma supply rate); (b) an increase in viscosity and associated decrease in mobility due to cooling or degassing/loss of water from the melt and associated microlite crystallization (e.g., Hammer et al. 2000; Cashman and Blundy 2000); (c) impingement upon a physical barrier, such as an impermeable cap-rock (e.g., Taisne et al. 2011) or alternatively a “soft” rock layer (Gudmundsson and Philipp 2006); or (d) a change in fracture orientations and/or stress conditions in the host rock that allows for lateral diking instead of continued vertical migration (e.g., Gudmundsson and Brenner 2004; Taisne et al. 2011). All of these factors can interact with each other; for example, lateral redirection of magma caused by encountering a rift zone will allow further magma degassing and in turn reduce the vertical driving force AND increase magma viscosity. Opportunities abound for ascending magma to stall. Indeed, judging from the number of earthquake swarms (e.g., Benoit and McNutt 1996), phreatic explosions (e.g., Barberi et al. 1992), and other potential indicators of magma ascent versus the number of magmatic eruptions, it appears that the majority of intrusions stall at some depth without erupting.

Potential “will it erupt/won’t it erupt?” discriminants

Qualitative assessments of eruption likelihood

The most diagnostic geophysical and geochemical signs that a magmatic eruption is likely are:

(1) Swarms of low-frequency earthquakes, volcanic tremor, and/or deformation well above baseline, suggesting increased gas pressures in the magma. Examples include Mount St. Helens, 1980 (Christiansen and Peterson 1981), Unzen, 1991 (Nakada et al. 1999), Pinatubo, 1991 (Wolfe and Hoblitt 1996; Harlow et al. 1996), and Shishaldin, Alaska, 1998 (Nye et al. 2002). However, we note that swarms of shallow low-frequency events can also occur without culminating in eruption, e.g., Shishaldin, 2001–2004 (Petersen 2007). Low-frequency and hybrid seismicity can also increase greatly as viscosity increases due to degassing, in which case magma may erupt as a viscous dome, e.g., Pinatubo, 1992 (Ramos et al. 1996), and Kelut, 2007 (Hidayati et al. 2009), or it may stall just before extrusion, e.g., Usu Volcano, 1910 and 1943–45 (Minakami et al. 1951; Yokoyama 2002; Goto et al. 2004).

(2) Seismic or geodetic evidence of relatively rapid magma ascent (i.e., cm to dm/s). Rapid ascent minimizes the time available for magma to bleed off its gas and crystallize. Evidence of accelerating ascent,

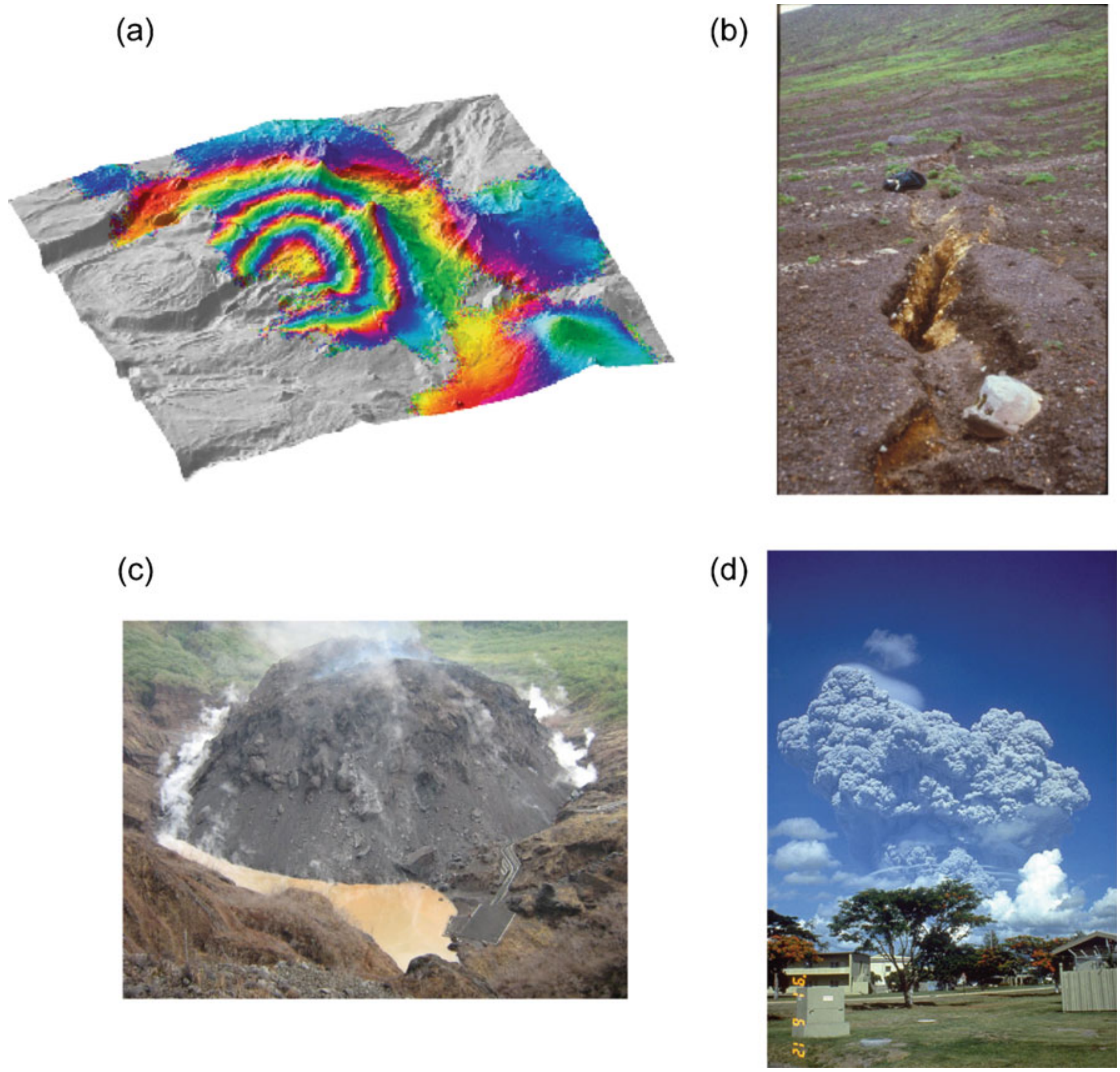

Fig. 2 Examples of the four outcomes of unrest: **a** "deep intrusion": Three Sisters, Oregon, 1997–2006—InSAR image showing deformation associated with intrusion of magma at ~8 km depth (modified with permission from Figure 1 of Wicks et al. 2002); **b** "shallow, stalled intrusion": Akutan, Alaska, 1996—picture showing ground cracks on NW flank of Akutan volcano in association with intrusion in 1996 (photo by R.G. McGimsey, U.S. Geological Survey); **c** "sluggish, viscous magma extrusion": Kelut, Indonesia, 2008—picture showing dome that emerged at Kelut following an energetic seismic swarm in 2007 (photo from CVGHM, Indonesia, reprinted with permission); **d** "rapid, often explosive magmatic eruption": Pinatubo, Philippines, 1991—the June 12, 1991, eruption column (photo by R. Hoblitt, U.S. Geological Survey)

largely in the form of accelerating deformation rates and/or seismicity rates, is a particularly important signal (e.g., Swanson et al. 1983; Malone et al. 1983).

(3) Sharp increases in SO_2 and CO_2 flux, to thousands or even tens of thousands of tonnes/day. However, some caution is needed when interpreting gas data. CO_2 often degasses early and can be in decline by the time magma nears the surface. SO_2 flux will increase whenever fresh gas-rich magma reaches shallow levels, either by intrusion or by acceleration of convection, but it can be masked by "scrubbing" through absorption into groundwater or crater lakes (e.g., Doukas and Gerlach 1995). For SO_2 to reach the surface and be measurable, hot gases must dry out a pathway. Thus, an increase in SO_2 flux can reflect either magma ascent or simply drying of a pathway through which it can now escape (see Werner et al. (2011) for further discussion). In the case of a conduit already filled with convecting magma and open to the surface, increases in SO_2 can be the most diagnostic change prior to a magmatic eruption (Newhall 2007). Such an increase may reflect an increased velocity (rate) of convection, with correspondingly faster supply of magma to the tip of the column where it can foam and degas. Seismicity and deformation in such cases are often remarkably minimal, e.g., Mayon, Philippines (Ramos-Villarta et al. 1985),

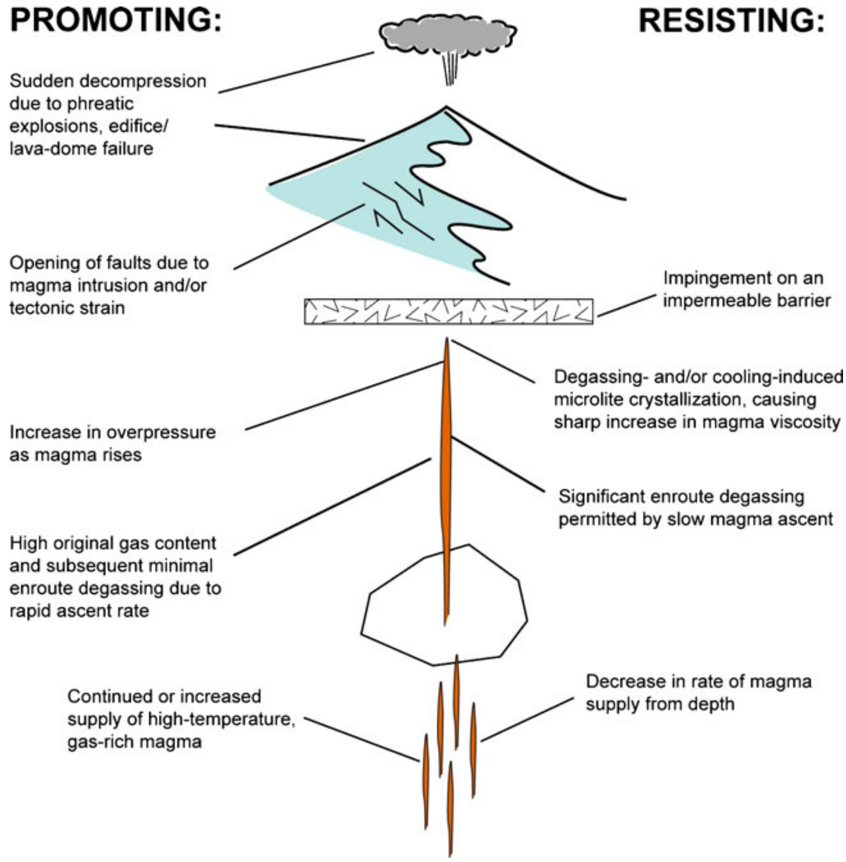

Fig. 3 Graphical representation of the principal processes that promote and resist magma ascent

Shishaldin (Moran et al. 2006), except if measured right at the summit/crater rim.

(4) Sudden (hours-days) declines in seismicity (e.g., Newhall and Endo 1987; Endo et al. 1996) and/or gas emissions (e.g., Stix et al. 1993). Seismicity declines may occur when magma is so close to the surface that no further fracturing is needed, when quenching of rising magma (in groundwater) forms a carapace around the tip of the magma column and briefly stops magma ascent, when rapid degassing "overshoots" and causes an increase in magma viscosity that similarly stops magma ascent, and/or when phreatic explosions temporarily reduce overpressure-driven seismicity, e.g., the vent-clearing phase of the Mount St. Helens 2004–2008 eruption (Moran et al. 2008). Gas emissions can decline suddenly just before an eruption if a quenching-induced carapace blocks further degassing or if flooding of a formerly dry open conduit suddenly scrubs magmatic gasses.

(5) A sudden reversal of deformation at summit stations, consistent with magma reaching a very shallow level. Models of deformation expected from shallow dikes show subsidence of the ground directly above the intrusion, after earlier uplift (e.g., Pollard et al. 1983; Lanzafame et al. 2003).

(6) Repeated, sometimes increasingly frequent and/or energetic phreatic eruptions, e.g., Augustine, 2006 (Power et al. 2006), suggesting that magma is pushing into and quickly heating groundwater. Each phreatic eruption causes decompression of the underlying magma, promoting gas exsolution. Such explosions also weaken the overlying rock, which can increase the likelihood of subsequent magmatic eruptions.

(7) Inclusion of small fractions of apparently fresh, juvenile glass shards in the ash of otherwise phreatic eruptions (e.g., Watanabe et al. 1999; Cashman and Hoblitt 2004).

There are also a few geophysical and geochemical signs that magma will probably not erupt. Decreasing rates of seismicity, deformation, and gas emission over timescales of weeks to months, as well as progressively longer pauses between phreatic explosions, are all signs indicating waning driving forces and/or increasingly sluggish magma ascent. Nishimura (2006) inferred that deformation will be constant (rather than accelerating) if magma nearing the surface has already been degassed to the point that it may fail to reach the surface. In the past, absence of significant SO_2 and juvenile glass shards has been interpreted to be evidence that magma will not erupt (e.g., Tazieff 1977). However, subsequent studies (e.g., Doukas and Gerlach 1995) have shown that magma can be close to eruption but SO_2 emissions still masked if groundwater is absorbing (or "scrubbing") all the SO_2 and other acid gases. Thus, absence of significant SO_2 is not a reliable indicator against eruption.

Quantitative assessments of eruption likelihood

The qualitative indicators noted in the preceding sections are useful, but still require quantification during episodes of unrest. Steps toward quantification are described in Newhall and Hoblitt (2002), Aspinall et al. (2002, 2003), Marzocchi et al. (2004) and Aspinall (2006). A perpetual difficulty with quantifying the probability of eruption is a lack of data—from the specific volcano in question as well as from analogous volcanoes. Many volcanoes erupt too infrequently for there to be good data on which intrusions will and will not lead to eruption. Analogues help, but are only as good as the inferred analogies. "WOVOdat", a project of the World Organization of Volcano Observatories, is currently compiling the collective experience of volcano monitoring agencies around the world into a searchable database (Venezky and Newhall 2007; Ratdomopurbo et al. 2009; http://www.wovodat.org). From WOVOdat and a working model of the volcanic system, scientists will be able to select appropriate analogues and thereby find enough data to use in decision frameworks such as that of Marzocchi and Woo (2007). In the interim, examples of failed eruptions highlighted in this volume should provide useful analogues for observatory scientists to consider during future episodes of unrest.

Conclusions

Examples cited above, in other papers of this volume, and elsewhere in the volcanology literature, show a complete spectrum through our classification of episodes of unrest into "deep intrusion", "shallow intrusion", "sluggish, viscous eruption", and "rapid, often explosive magmatic eruption." The outcome of any intrusion (and associated unrest) is a result of the competition between factors promoting and hindering magma ascent and eruption. Rapid and often explosive magmatic eruptions occur if magma intrusions accelerate into runaway ascent, i.e., when a feedback loop speeds up driving forces, such as gas exsolution, and overwhelms processes (degassing, crystallization) that would otherwise slow the process. For sluggish, viscous eruptions of lava domes, the processes which resist eruption (degassing, sharp increases in viscosity, encountering of physical barriers) almost stop the magma before it reaches the surface. For shallow intrusions that stall (our definition of a failed eruption), the balance is tipped even further in favor of resisting forces, resulting instead in plugs, dikes, and perhaps elevated geothermal activity. The papers in this volume provide a variety of case histories (Crider et al. 2011; Gardine et al. 2011; Nishimura and Ueki 2011; Roman and Power 2011; Werner et al. 2011) and perspectives from modeling (Taisne et al. 2010) that we hope will be a significant resource for scientists grappling with the issue of whether or not a future episode of unrest will lead to a magmatic eruption.

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