Volcanoes: From Mantle to Surface

Down the Crater: Magma Storage and Eruption
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Volatile Formation in Volcanic Systems
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Volcanic Tipping Points & Unknowns

KEITH D. PUTIRKA and KARI M. COOPER, Guest Editors
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INTRODUCTION

Whereas geological mapping allows scientists to see the inner plumbing of old volcanoes that have been exhumed by erosion, it is geophysical methods that allow us to study the internal structure of active volcanoes and the processes that may ultimately lead to eruption. Geophysical techniques make measurements near the surface to image the structure beneath a volcano, recording the stress and strain associated with magmatic processes (e.g. Pritchard and Gregg 2016). A common limitation is that multiple sensors must be placed around the volcano in order to make detailed inferences about the subsurface, and the majority of the world’s 1,400 subaerial volcanoes do not have such networks (e.g. Loughlin et al. 2015). Observations of surface deformation (‘geodesy’) can now be made using satellites and can be made globally (e.g. Fournier et al. 2010; Biggs et al. 2014). The number of known deforming volcanoes has quintupled over the last 20 years as traditional ground-based survey methods have been complemented with satellite technology. The ability to observe how different types of volcanoes deform offers a unique perspective on the behaviour of magmatic systems throughout the eruption cycle.

Despite the explosion of observations, our understanding of the significance of deformation (or the lack of deformation) at volcanoes is still in its infancy. The challenge for 21st century volcanologists is to link the new observations of surface deformation to volcanic processes, within the framework of other geological and geophysical studies of magmatic systems. In the coming decades, much more will be learned about complex magmatic systems, from the generation of melt to its interaction with the surface, atmosphere – or shallow crust – and there will be improved monitoring and forecasting of volcanic hazards.

This article will first review the geodetic methods that allow us to study subsurface processes in active volcanoes, and then summarize the broad range of volcanic behaviour that has been observed geodetically.

GEODETIC TECHNIQUES

In 1997, deformation had been reported at 44 volcanoes (Dvorak and Dzurisin 1997); by 2010, there were 118 (Fournier et al. 2010) and; at the writing of this article, there are over 220, which we document in the supplementary table. This rapid increase is not the result of a rise in volcanic activity but is a consequence of improved observation and reporting, particularly in the developing world. Technological advances have been vital: a fleet of international satellites (10 radar satellites in 2017 with more planned; e.g. Pinel et al. 2014) make routine, global observations, and global positioning system (GPS) networks make ground-based geodetic observations routine and affordable. However, the list of deforming volcanoes is still incomplete because some volcanoes have never been studied or have only been studied incompletely due to inadequate data.

This review focuses on interferometric synthetic aperture radar (InSAR), the main satellite-based tool that is used to measure surface deformation at volcanoes. Radar (a name itself that derives from ‘radio detection and ranging’) involves the transmission and reception of microwave electromagnetic radiation (roughly $10^8$–$10^{11}$ Hz, or wavelengths of 1 mm to 1 m) (e.g. Pinel et al. 2014). At these wavelengths, radar systems can see through clouds and most types of precipitation and do not rely on the sun’s illumination, giving them a unique all-weather, day–night capability. Each pixel in a synthetic aperture radar (SAR) image is represented by a complex number, with the amplitude corresponding to the intensity of the returned radar energy and the phase equalling a fraction of the complete wavelength (having a value between 0 and $2\pi$). When the phases from two images are combined to form an interferogram, the phase difference reveals variations in the distance between the ground and the satellite that appear as coloured fringes (Fig. 1). Each fringe corresponds
to half the radar wavelength. The rate of ground displacement can be measured by taking the observed change in distance divided by the time interval between SAR images—this interval usually varies between a day and a few weeks, with rates of deformation ranging from mm/y to several m per day (Fig. 2). There are two notable limitations to the InSAR technique. First, decorrelation, which occurs when characteristics of the ground and its ability to reflect radar waves change rapidly, for example in heavily vegetated or agricultural regions. Second, atmospheric delays caused by water vapour in the troposphere. Time-series methods that can combine hundreds or thousands of images are increasingly important in overcoming these limitations (e.g. Dzurisin et al. 2006; Pinel et al. 2014).

The revolution in volcano deformation studies as a result of InSAR comes from the ability to routinely image deformation at nearly all of the world’s volcanoes with an accuracy of a few millimetres to a few centimetres. Even so, volcanoes can deform very rapidly (e.g. Fig. 2) and the repeat time between satellite overpasses may be too long to capture these temporal changes, particularly as the amount of data...
A variety of processes can cause ground deformation at volcanoes: magma movements, landslides, faults, hydrothermal systems, and thermal or thermodynamic volume changes from heating, cooling, melting or crystallization (e.g. Dzurisin et al. 2006). Nevertheless, certain patterns of ground deformation can, to a limited extent, be diagnostic of specific physical process, e.g. a dyke intrusion, cooling lava/pyroclastic flow, or pressurizing ‘magma reservoir’, all of which have distinct patterns of ground displacement viewed with InSAR (Fig. 1A-E). Complexities arise depending on the direction the satellite is looking relative to the ground displacement (e.g. Dzurisin et al. 2006) and when multiple processes occur at nearly the same time (Figs. 1F, 1G). In this section, we describe some commonalities between deforming volcanoes and global patterns in the types of processes that produce ground deformation.

**The Classic Volcano Deformation Cycle**

Prior to an eruption, according to the classic model of the ‘volcano deformation cycle’, magma gradually inflates a magma chamber directly beneath the volcanic edifice until a threshold is reached at which point the chamber ruptures and an eruption rapidly empties and deflates the chamber (e.g. Dzurisin et al. 2006). The inflation phase causes uplift of the ground surface and large numbers of small earthquakes (volcano-tectonic seismicity), while eruption is accompanied by rapid subsidence (Fig. 3A). To a first approximation, this pattern of co-eruptive subsidence and inter-eruptive uplift has been observed at a number of volcanoes with different characteristic length scales and timescales and has been used to provide eruption forecasting in a number of cases (e.g. references in Dzurisin et al. 2006; Fournier et al. 2010). However, the recent explosion in volcano monitoring data has demonstrated that the classic model is an oversimplification. Many volcanoes just do not behave in this way. Furthermore, a simple, large, liquid-filled magma chamber is not compatible with modern geological or petrological observations. Nonetheless, the classic model of the volcano deformation cycle remains a useful concept from which to start.

Long-lived eruptions and intrusions provide an alternative opportunity to observe repeated cycles of behaviour. For example, the multi-decadal eruptions of Soufrière Hills Volcano (SHV) on Montserrat (an andesitic stratovolcano in the Caribbean) and Kilauea (Hawai‘i USA) (a basaltic shield volcano) have served as test-beds for new ideas and new monitoring systems. Global positioning system instruments at SHV observed a simple first-order pattern: when lava is erupting rapidly, the surface subsides; during periods of no extrusion, it inflates (Wadge et al. 2014). Similar patterns have been observed at volcanoes with multiple, distinct eruptions, such as Fernandina in the Galapagos (e.g. Pinel et al. 2014) and Okmok (Alaska, USA) (e.g. Lu and Dzurisin 2014). However, close inspection of the time series shows that inter-eruptive uplift is interrupted by short reversals that are not associated with any magma output at the surface. At Okmok, these reversals have been attributed to pulses of gas loss or re-absorption (e.g. Carinci et al. 2014).

**Intrusions**

Most magma does not reach the surface – but sometimes, the intrusion of dykes and sills and the growth of plutons are visible in the geodetic, as well as the geological, record. Earthquake swarms at Eyjafjallajökull (Iceland) in 1994 and 1999 were associated with tens of centimetres of surface uplift but no eruption (Sigmundsson et al. 2010 and references therein). Similar patterns of seismicity and deformation in 2010 culminated in the eruption that disrupted Europe’s air traffic, causing huge economic losses. Magma is supplied in batches, some of which ‘stall’ and form an intrusive complex at the roots of the volcano, while others, as happened in the 2010 eruption, reach the surface, triggering large eruptions which can tap multiple reservoirs (Sigmundsson et al. 2010). Intrusions also occur within volcanic edifices. The orientation of these intrusions is
Temporal patterns of volcano deformation

**Volcanic Conduits**

Volcanic conduits are thought to be extremely narrow (<10 m radius) and are, therefore, difficult to observe geophysically. Nonetheless, clues to their behaviour are given by sub-daily deformation cycles that are typically observed in association with Strombolian or Vulcanian eruptions.

**Restless Calderas**

Caldera systems have long repose periods between very large eruptions, but do not remain quiescent – many calderas have frequent, small eruptions, known as resurgent volcanism. In the case of Santorini (Greece) the volume of erupted lava is directly proportional to the time since the previous eruption, evidence that magma supply from depth is continuous (Parks et al. 2012). However, in 2010, a short period of rapid uplift occurred with an equivalent volume to that anticipated for the next eruption, indicating that the magma supply to the shallow reservoir is pulsed rather than continuous (Parks et al. 2012).

Some caldera systems have been known to deform for decades without erupting. At Campi Flegrei (Italy) gradual subsidence over centuries has caused the Roman market at Pozzuoli to become submerged. The subsidence is occasionally interrupted by pulses of uplift but it remains unclear whether the cause is magmatic or hydrothermal (e.g. Chiudini et al. 2010). At Yellowstone (USA), deformation has a spatially and temporally variable pattern and is attributed to hydrothermal fluids moving between reservoirs as well as magma intrusion (e.g. Chang et al. 2007).

More recently, satellite observations have shown that deformation is occurring at many caldera systems that are not erupting, that have no record of historical volcanism, and that have no real-time monitoring. This includes many volcanoes along the densely populated East African Rift (e.g. Biggs et al. 2014 and references therein) and raises the question of how to interpret caldera deformation in terms of volcanic hazards. The fastest uplift rate seems to be at Laguna del Maule (Chile), which has been uplifting at a rate of 280 mm/y since 2007 (e.g. Fournier et al. 2010). However, the implications for the shallow magma body underneath remain unclear.

Is the behaviour of restless calderas related to external influences? The passage of seismic waves associated with large earthquakes has been shown to influence hydrothermal systems, probably as a result of interactions with gas bubbles. For example, both the 2010 Maule (Chile) and 2011 Tohoku (Japan) earthquakes caused subsidence at several nearby volcanoes, but the subsidence mechanism could be due either to changes in the hydrothermal system or to stress changes acting on the weaker rheological properties of a pluton (e.g. Pritchard et al. 2013).

**During and After an Eruption**

While co-eruptive subsidence associated with subsurface magma withdrawal is a common feature of eruptions, in many cases it is masked by local surface changes, including both the destruction and creation of topography. Explosive eruptions typically form new craters and vents, they can

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**Figure 3** Temporal patterns of volcano deformation (schematic); stars represent eruptions. (A) Classic model of the eruption cycle where emptying of a magma reservoir during eruption causes subsidence, and refilling of the reservoir at a constant rate between eruptions causes uplift at a constant rate. (B) Modified eruption cycle, such that the rate of refilling between eruptions decays exponentially as magma flows along a pressure gradient from a deep reservoir. (C) Magma rises quickly and erupts immediately, such that deformation is rapid, recoverable and is likely to be undetected. (D) Pulsed magma supply, batches of magma are intruded causing uplift until a threshold is reached and an eruption is triggered. (E) Continuous unrest without eruption, caused by phase changes and mixing within shallower magma storage or by an overlying hydrothermal system. (F) Constant rate uplift or subsidence that may continue for several decades because of the growth or cooling of deep magma bodies.
alter the morphology of existing structures, and they may trigger large collapses. Eruptive products include lava flows, pyroclastic flows and lahars and these can fill low topography. A new dome may grow near the vent (e.g. Pinel et al. 2014). Ash fall, while rarely thick enough to dramatically alter topography, can alter the appearance of the ground surface making it incoherent to satellite radar and so reducing the efficiency of ground-based monitoring that relies on solar panels. After an eruption, the new topography is typically oversteep, and processes such as landsliding and gravitational spreading (e.g. Ebmeier et al. 2014) act to restabilise the landscape. Subsidence of cooling and compacting lava flows can last for decades: Paricutin (Mexico) erupted in 1952 and is still subsiding today (e.g. Fournier et al. 2010).

Magmatic systems can respond to eruption in multiple ways. In the classic model of the volcano deformation cycle, the magma chamber begins to refill within days of the eruption (e.g. Okmok volcano, described in Lu and Dzurisin (2014)). But sometimes the volcano just continues to subside (e.g. Kiska volcano in the Aleutian Islands of Alaska, described in Lu and Dzurisin (2014)). For decades after an intrusion, the subsurface magma body will continue to cool, crystallize and degas: at Medicine Lake Volcano (California, USA), modern geodetic observations were combined with levelling surveys from the 1950s to demonstrate that subsidence has continued at ~10 mm/y for at least the last 65 years (Parker et al. 2014).

**Limits of Detectability**

Several eruptions appear to have taken place at volcanoes with no known deformation (e.g. Fournier et al. 2010; Lu and Dzurisin 2014). Even when co-eruptive subsidence is observed, the volume is usually less than the dense rock equivalent of the erupted products. The classic model of the volcano deformation cycle assumes that there is a constant volume flux of magma being supplied to the magma chamber, producing a linear rate of uplift between eruptions. Altering the boundary conditions such that magma is supplied from a deeper chamber at constant pressure modifies this model such that eruption is followed by exponentially decaying uplift. This may explain observations of eruptions that occur with negligible uplift prior to eruption and very rapid uplift following eruption, such as was observed at Westdahl (Alaska) (e.g. Lu and Dzurisin 2014).

The simplest explanation for a lack of co-eruptive subsidence is that deformation is occurring at a magnitude or resolution beneath our current ability to observe: too fast or too slow, too shallow or too deep, or obscured by atmospheric effects. Already, new satellites with higher resolution and faster orbital repeat (e.g. TerraSAR-X and CosmoSkyMed) have been used to observe previously undetectable processes (e.g. Salzer et al. 2014). Long-term monitoring missions, such as the European Space Agency’s Sentinels programme, have the potential to revolutionise detection capabilities. An alternative explanation is that we ought to consider mass balance, rather than volume balance, because density is far from constant in a three-phase magmatic system undergoing changes in temperature and pressure. Gas bubbles, after all, do make magmas very compressible (e.g. Caricchi et al. 2014).

**GLOBAL SYNTHESIS**

The rapid increase in the number of geodetically studied volcanoes means that it is now feasible to treat the observations statistically. Biggs et al. (2014) showed that there is a strong link between deformation and eruption for over 500 systematically studied volcanoes, and also between non-deformation and non-eruption. Further, this relationship varies with volcanic parameters: shield volcanoes have a strong link between deformation and eruption and come closest to the classic volcano deformation cycle, while calderas frequently deform without erupting, suggesting that large volumes of magma are stored in the upper crust and that the deformation is caused by gas or hydrothermal fluids. Stratovolcanoes are the most likely to erupt without observable deformation, perhaps because the mass changes are accommodated in a way not detectable by current observation satellite systems, which have revisit times of weeks.

How do we know when a given deformation event may lead to eruption? One way to answer this question is to compare the duration of deformation events to their magnitude and see whether they led to an eruption (Fig. 2). As expected, deformation events that had large magnitudes usually lasted for a short amount of time and typically led to an eruption (e.g. Fournier et al. 2010). Yet there doesn’t seem to be a simple threshold at which deformation duration and/or magnitude should cause concern. On the other hand, there are many volcanoes that can deform at rates of 1–1,000 mm/y without causing eruption in the short term, and so these types of deformation events are not always hazardous. Figure 2 and the online supplementary table with this article are incomplete in several respects. Some volcanoes have never been studied. At others, the temporal sampling is inadequate to resolve pulses of rapid deformation, the deforming area being too small to resolve or had occurred within a data gap. In some cases, deformation was ongoing at the start or end of the available observations, making the specified duration a minimum estimate. There is a cluster of deformation events with a one month duration that likely correspond to even shorter time periods, but these short events could not be constrained because observations were not sufficiently frequent.

Although deformation has been reported at over 200 volcanoes, there remain several eruption styles that have been observed rarely, or not at all. The only geodetically observed rhyolite eruption occurred at Chaitén (Chile) in 2008, with only a few hours of pre-eruptive warning (e.g. Wicks et al. 2011). With a sample size of one, it is impossible to state whether the pattern observed is representative. The latest monogenetic eruption occurred at Paricutín during the period 1943–1953, which was before routine satellite observations. Perhaps most alarming is the lack of observation of the very largest eruptions. The most recent magnitude 7 eruption (i.e. erupting a volume in dense rock equivalent of $10^{12}$–$10^{13}$ m$^3$) occurred at Tambora (Indonesia) in 1815. Even recent magnitude 5–6 eruptions, with erupted volumes of $10^3$–$10^4$ m$^3$ (such as Mount St. Helens (Washington, USA), and Mount Pinatubo (Philippines)) have limited geodetic observations. The only available example is the relatively small eruption of Campi Flegrei (Italy) in 1538, where historical records suggest that several metres of deformation occurred in the years before the eruption (Guidoboni and Ciucarelli 2011). Calderas have the potential for extremely large eruptions, but could we distinguish between the semi-continuous unrest seen at so many calderas and the precursors to a major eruption?

**SUMMARY AND PERSPECTIVES**

Satellite techniques are rapidly improving our ability to monitor volcanoes on a global basis and have provided insight into the spatial and temporal changes in the subsurface stress fields around volcanoes and intrusions. The classic model of the volcano deformation cycle – co-eruptive deflation and inter-eruptive inflation – is seen at many volcanoes, but the rapid increase in geodetic...
monitoring has demonstrated that such a simple model is not always applicable. Deformation can be attributed to the movement and phase transitions of magma, volatiles and hydrothermal fluids and to the intrusions of dykes, sills and the growth of plutons, each of which have their own characteristic geodetic signatures. Surface processes during and after eruptions, such as the formation of volcanic flows and domes as well as edifice instability, may obscure subsurface processes. Alongside developments in observational ability, modelling capabilities have evolved from analytic solutions for point sources and other simplified geometries, to finite element models that can incorporate more complex rheological and structural information, to physics-based models capable of integrating geophysical monitoring with observations of degassing and petrology (e.g. Masterlark 2007; Anderson and Segall 2011). Satellite imagery has proved a remarkable reconnaissance tool to discover new phenomena. As new satellites and constellations of satellites are launched in the coming years (e.g. Pinel et al. 2014), even more new discoveries will be made. Yet, the biggest questions remain: “Which of the deforming volcanoes are a threat?” and, “Under what circumstances does deformation lead to eruption?”

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