CHAPTER 5.1 of PART 1: Tectonics and plumbing systems

Eruption triggering

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Abstract

Volcanic systems behave episodically and require mechanisms for magma segregation, instability and ascent. Here, we discuss the processes that promote the ascent of magma to the Earth's surface and prepare the onset of an eruption, thus acting as triggers, as well as the factors that prevent eruption. We describe the various petrological, geochemical and geophysical observations that reveal the triggering processes and events. Based on emerging concepts of magma plumbing systems, we consider the mechanisms involved, the buoyancy and overpressure of magmas and associated fluids, using modelling approaches. The transient increase in magma inflow at depth, leading to magma recharge at shallow levels, and the increase in volatile content during ascent are key ingredients of internal triggering. Events, such as earthquakes, changes in surface loading and interactions with external water and hydrothermal systems, perturb the stress field, change magma and fluid pressures to act as external triggers.

Keywords

Internal/External triggering, Critical state, Magma chamber failure, Recharge, Unrest

I Introduction

More than 500 million people now live less than 100 km from an active volcano and can be threatened by volcanic phenomena. In the context of increasing populations and anthropogenic environmental impacts, understanding the processes that lead to a volcanic eruption contributes to achieving the United Nations (UN) Sustainable Development Goal (SDG) 11 to “make cities and human settlements inclusive, safe, resilient and sustainable”. To reduce deaths and economic losses due to volcanic eruptions, volcanologists need to anticipate the location, timing and magnitude of eruptions by characterising the mechanisms acting on evolving volcanic systems. This knowledge will strengthen resilience and the ability to adapt to the potential feedback between environmental changes and volcanic activity, thereby making progress for achieving the UN SGD 13.

In this chapter, which is focused on eruption triggering, we refer to any potential phenomena that could significantly increase the activity level of a volcanic system. We consider events leading to unrest, the onset of a magmatic or phreatic eruption, and the transition to a paroxysmal phase during long-lasting eruptions. By ‘eruption triggering’, we mean phenomena or processes that enhance the eruption probability by increasing the amount of magma arriving at shallow depth (thus inducing unrest or phreatic activity) or at the surface (thus feeding an eruption). Triggering may involve one or more of the following processes: 1) magma generation at depth; 2) destabilisation of crustal magma storage zones; 3) magma ascent to the surface, by affecting the driving forces of rock failure, magma flow and; 4) changes in environmental conditions (such as the local stress field) in an active system initially at rest or at low levels of activity. These events may be internal (with a magmatic origin and driven from deep levels) or external factors or their combination (see Figure 1).

Many volcanic systems are characterised by episodic behaviour. Most volcanoes are dormant most of the time and eruptions are discrete events. The episodic character of volcanic activity requires mechanisms for magma instability and ascent. Magma propagation is necessary but not sufficient for magma to erupt. Many internal and external processes may affect whether magma reaches the Earth's surface or not. Most magma present at depth typically does not erupt, so a triggering mechanism is often necessary for an eruption to occur. However, not all triggering mechanisms necessarily lead to an eruption. Improved understanding of triggering mechanisms enables better characterisation of the evolution of a volcanic system and to better interpret the observational changes recorded prior to eruptions as a preliminary step towards improving eruption warning and reducing volcanic risks.

Triggering mechanisms can be identified from many different kinds of observations, including petrological observations of eruptive products and monitoring data from active volcanoes (section II, link to chapter 5.4), and can be studied by theoretical, numerical and experimental approaches (section III, link to chapter 5.3). Various triggers can be considered, with many potential interactions (section IV). The actual influence of the triggers depends on the current state of the system (section V, link to chapters 3.2 and 4.3).

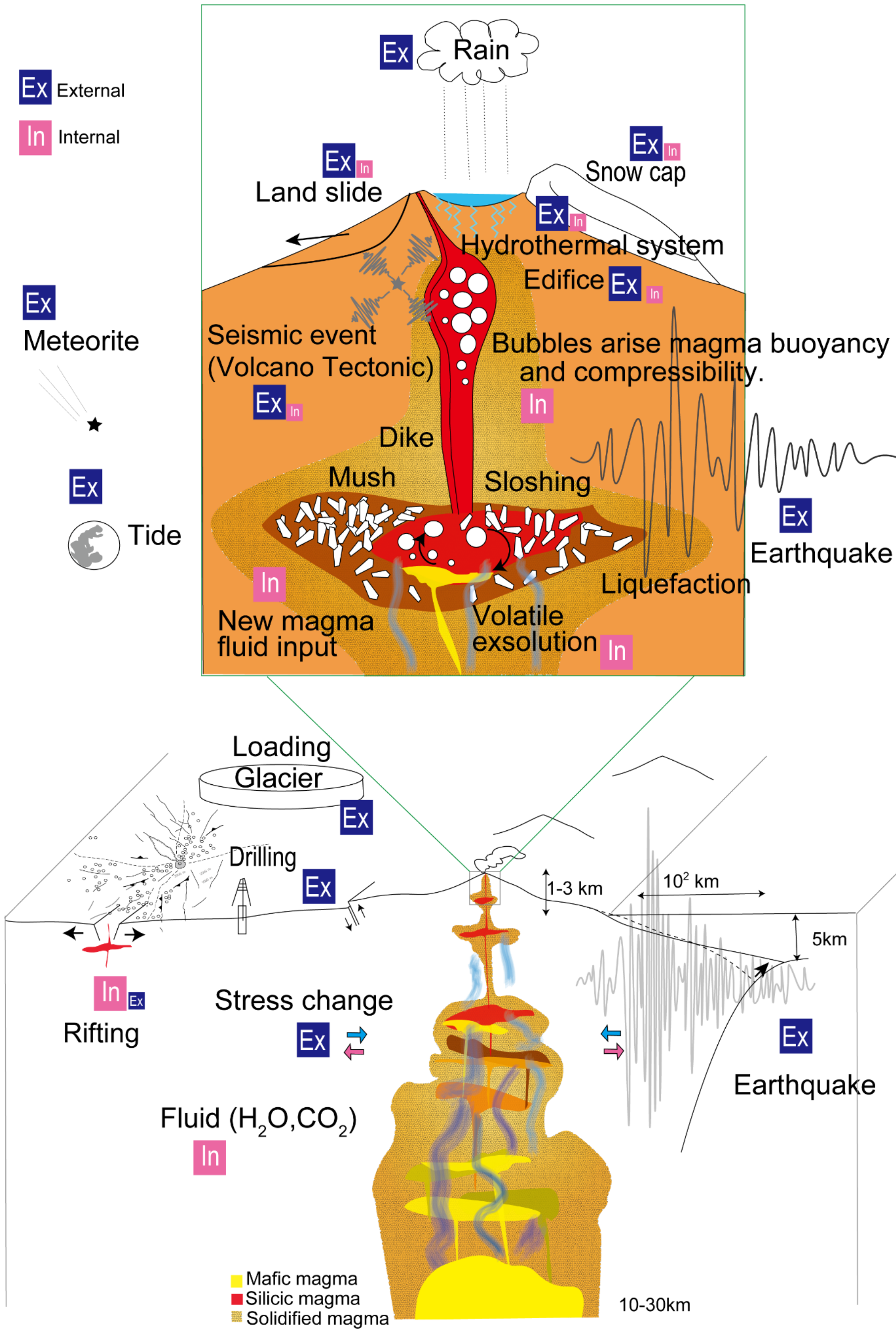


Figure 1. Summary of the various potential triggers for volcanic activity presented both on a large scale (the whole plumbing system in its geodynamic context) and on a local scale (volcanic edifice and shallow mature magma system). Most local external triggers, with the exception of precipitation and distant tectonic earthquakes, can be strongly influenced and possibly triggered by magma inflow at depth (see section IV, c). See the main text and Video 1 for detailed explanations of the phenomena. This illustration exaggerates some crucial phenomena, and the

proportions of the objects may not be realistic. [Link by QR code to the online animated version of figure 1 labelled Video1]

II Triggering processes

Development of ideas on the causes of volcanoes and volcanic processes extends back to the early 19th century, but specific mention of eruption triggering mechanisms surprisingly only developed in the 1970s. However, many of the key hypothesis were already established in previous decades, framed by the quest to understand volcanic processes and phenomena, but without discussing triggering mechanisms explicitly. Key ideas included: formation of magma pathways (e.g. dikes) to the Earth's surface due to magma overpressures and favourable stress fields; ascent of magma due to buoyancy; caldera collapse into large shallow magma chambers; and exsolution of volatiles (in particular water) to drive magma ascent and explosive volcanism. Understanding of phase equilibria and experimental measurements of key physical properties, such as pressure dependent water solubility were particularly influential in understanding the causes of explosive eruptions.
Field and petrological evidence of plutonic and volcanic rocks suggests a link between the start of eruptive activity and magma inflow (recharge) into upper crustal magma reservoirs. In plutonic systems field evidence includes occurrences of mafic synplutonic dikes and mafic enclaves, whereas in volcanic products, the occurrence of banded pumice fragments, complex crystal and melt inclusions populations, and mafic enclaves, indicate that mixing and mingling are ubiquitous processes in active magmatic systems. Sparks et al. [1] proposed triggering of explosive eruptions by recharge and magma mixing.

Detailed analysis of geophysical time-series datasets gathered during recent large eruptions (e.g. Pinatubo, 1991; Soufriere Hills, 1995-2010 (see Box1: Case Study box Soufrière Hills)) provides abundant evidence for recharge. During these eruptions, scarce, deep (medium-to-lower crust) long-period seismicity and/or CO_2 degassing has been recorded several months before the paroxysms. Abrupt changes in deformation, volcano-tectonic and long-period seismicity, and degassing patterns (changes in gas content and ratios) are commonly observed and interpreted as the arrival of magmas into the upper part of the magmatic system.

Petrological and geochemical studies of eruptive products sampled during short-lived large eruptions and long-lasting eruptive events also point to magma recharge as a common triggering mechanism. Among the conspicuous petrological features associated with recharge, magma mixing and mingling, we can mention the large heterogeneities of the whole-rock geochemical compositions in a single deposit which include compositionally banded rocks with dark (mafic) and light (felsic) bands, as well as hybridised (intermediate) compositions. These compositionally different samples typically define linear trends in binary plots and curved arrays in diagrams between incompatible trace elements ratios. Samples show complex petrography and mineralogy including disequilibrium mineral assemblages (such as the coexistence of Mg-rich olivine and quartz), and large variabilities in mineral chemistry, commonly showing reverse zoning patterns (e.g. anorthite-rich rims in plagioclase, forsterite-rich rims in olivines, Mg-rich rims in ortho- and clinopyroxenes). Disequilibrium textures (sieve and patchy phenocryst textures, dissolution surfaces and overgrowth rims) are commonplace. Large chemical diversities in melt inclusions are also consistent with magma mixing.

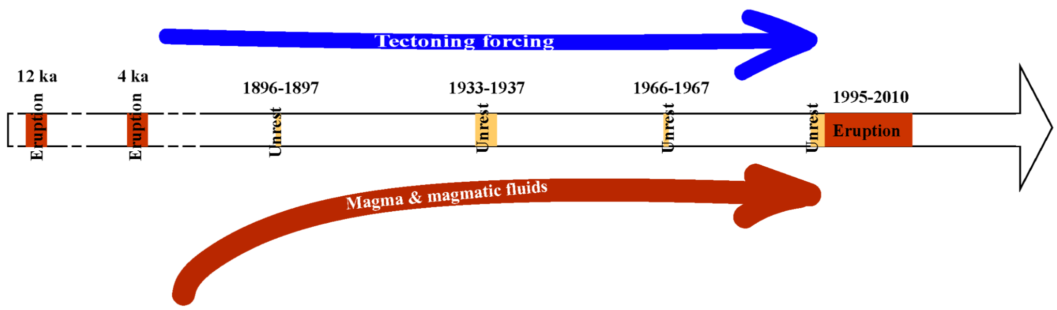
However, an alternative interpretation of most of the petrological and geochemical evidence is that recharge is not the cause of the eruption but rather its consequence. In this scenario, an eruption is triggered by a near-surface mechanism and, once initiated, the eruption destabilises the magmatic system, leading to recharge. It is therefore not always easy to distinguish between top-down and bottom-up eruptions from petrological and geochemical observations alone, but geophysical signals can provide useful constraints on timing.

**************BOX1 : CASE STUDY BOX : '**The 1995 – 2010 eruption of the Soufriere Hills Volcano, Montserrat.** ' ****************************

Box1Figure1: Photograph of the Soufriere Hills volcano, courtesy of S. Sparks.

The Soufrière Hills Volcano (SHV), Montserrat illustrates the interplay of external and internal factors in triggering a major eruption in an arc setting. The 1995 to 2010 eruption needs to be considered in the context of its volcanic history. The two previous major eruptions took place at about 4 and 12 ka. The eruptions of SHV conform to an episodic model of short intense periods of silicic andesite eruption and very long periods (typically several thousands of years or more) of dormancy. In addition, SHV has experienced periods of unrest characterised by seismic activity and enhanced fumarolic activity but without eruption as in *CE* 1896-97, 1933-37 and 1966-67. With approximately 30-year intervals of unrest it was not unexpected when seismic activity resumed in 1992, but this time a major eruption started in 1995. The earlier periods of unrest are interpreted as failed eruptions with ascent of magma and associated fluids into the shallow magma system from deeper regions. It seems then that deeper internal magmatic processes triggered both the periods of unrest and the 1995-2010 eruption. However, earlier attempts for magma to reach the Earth's surface were unsuccessful with the shallow cold and brittle crust plus the hydrothermal system providing physical barriers that required thresholds of physical conditions (such as magma pressure and heat transfer rates) to be exceeded for an eruption to be triggered.

While internal and deeply sourced magma and magmatic fluid instabilities seem to be the primary cause of triggering eruptions at SHV, the causes of these instabilities are not well understood. Some key observations have led to the hypothesis, widely applied to many arc volcanoes, that shallow magma chamber recharge by mafic magma is the triggering mechanism. Mafic inclusions are abundant in the silicic andesite lava and ejecta of the 1995-2010 eruption. Large fluxes of SO_2 during and following the eruption are attributed to degassing of unerupted basaltic magmas intruded at depth. Mafic magma inclusions and excess SO_2 fluxes are very common in andesite volcanoes worldwide. Despite its attractions, the mafic recharge model fails to fully explain key phenomena, such as the cause of the recharge itself and the formation of the shallow magma system which implies transfer of silicic andesite magma into the shallow crust. Also, alternatively magma mixing with formation of mafic inclusions and the excess SO_2 fluxes are a consequence rather than a cause of eruption. Cause and effect are far from clear. Indeed, as more detailed petrological and geophysical evidence accumulates, it seems more likely that instability develops across a considerable vertical depth range in a transcrustal magma system.



Box1Figure2

Tectonics is another major factor in triggering eruptions and moderating frequency of eruption. Again, like many arc volcanoes, SHV is located within an active transtensional fault system. The interactions of magma and fault systems are evident from: concurrent seismicity especially in the precursory period prior to eruption; from the location of vents and domes; and from cyclic deformation and eruptive patterns observed by geophysics during the eruption. Modelling of deformation patterns at SHV (tilt, strain meter, fault plane solutions and GPS patterns) indicate emplacement of dikes, with orientations controlled by local tectonics, that provide pathways for erupting magma. More generally, observations and theoretical concepts indicate that interaction of tectonic and magmatic stresses and associated deformation rates influence ascent of magma through the shallow brittle crust to enable eruptions.

[Reference for the case study box:

Wadge, G., Voight, B., Sparks, R.S.J., Cole, P., and Loughlin, S.C. An Overview of the Eruption of Soufriere Hills Volcano from 2000-2010. In "*The Eruption of the Soufriere Hills Volcano, Montserrat from 2000-2010*" edited by Wadge, G., Robertson, R.A.E. and Voight, B. Geological Society Memoir 39, 1-40, 2014.]

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The possibility of triggering by external events, such as earthquakes, intense rain, deglaciation, and large edifice sector collapse, has been recognized by the coincidence of external event occurrences with eruptions, increase in eruptive rates, or volcanic unrest episodes (e.g. Hill et al. [2]; Watts et al. [3]). Such events may change the stress field surrounding volcanoes and magma chambers, which can be a cause-effect link (see Section IV b for details). Statistical analysis indicates that the number of eruptions increases within several years following major earthquakes (e.g. Sawi and Manga [4]). The advent of satellite monitoring yielded more systematic and consistent detection of these events on a global scale (e.g. Delle Donne et al. [5]), revealing more subtle volcanic responses to external events, such as large-scale surface deformation, and increased heat flux or degassing. The increases in data quality and exponential increase in quantity enable verification of models, understanding of triggering mechanisms and more robust statistical testing.

III Prerequisites for triggering, modelling keys

The different triggering mechanisms can be described in the framework of the current knowledge of the magmatic plumbing systems. There is a consensus in the scientific community to picture magmatic systems as complex environments composed of melts, crystals and volatiles in different proportions. These regions can extend through the crust from the surface to the source regions in the lower crust and the mantle wedge (chapters 1.4.1 1.4.2 and 1.4.3). Magma reservoirs are composed of melt-rich regions (magma chambers), crystal-rich regions (mush), and exsolved volatiles pockets.

At depth, the underlying cause of volcanism is the production of magma in the Earth's interior by partial melting, segregation and subsequent ascent. In source regions magmas have a lower density than their surroundings so buoyancy forces play a critical role in melt segregation and the instabilities that cause magma ascent, thus causing episodic volcanism. Unfortunately, the deeper processes are more difficult to detect by near-surface geophysical and geochemical observations. Once segregated by compaction and shear from within regions of partial melting, buoyant melt layers grow and become unstable. Seropian et al. [6] show how episodic magma ascent is controlled by the time scales and volumes of melt layers. Coupling melt segregation, melt layer instability and a simple failure criterion for deep reservoirs in a model allows the study of episodic volcanic eruption initiation from the roots of the system [7].

In mature systems, melt-dominated magma chambers form at shallow levels. Once a shallow magma chamber is established, two conditions must be met for an eruption to occur. The first condition is that the wall of the chamber must rupture, allowing the magma to flow out and start its ascent towards the surface [Reference to Box2: chamber stability How-to Box]. The second condition is that the magma ascending can reach the surface. These two conditions both depend on the balance of forces at play in the thermomechanical system resulting from the coupling between a complex multi-phase fluid (the magma) and a heterogeneous solid (the surrounding crust, partly made up of mush). The driving forces for magma ascent are the magma internal overpressure and the magma buoyancy resulting from the relatively lower density of the magma compared to the mantle and crust. These driving forces typically increase with magma volume, must exceed the strength of the crust to form a fracture. Once a magma-filled fracture has formed its propagation towards the surface involves complex dynamics and mechanics so there are many factors that influence whether it will succeed in reaching the surface or not. Factors that may prevent reaching the surface include: decrease in overpressure resulting in the fracture closing; cooling and degassing with crystallisation which increases viscosity and ultimately freezes the magma; meeting variations in rock strength along the pathway resulting from crustal layering; and an unfavourable local stress field (e.g due to edifice load) that divert the magma propagation laterally. Meeting hydrothermal and ground water systems might enhance cooling. In contrast, factors that enhance the chances of reaching the surface include exsolution of volatiles which increase the buoyancy contribution of overpressure and in some instances may lead to explosive fragmentation that disrupt host rocks and, in some cases, lead to explosive break through.

It follows from the above that any process that pressurises the magma (e.g. recharge, volatile exsolution by crystallisation) or reduces its density (e.g. magma inflow, increase in gas content) will promote magma chamber failure and consequent magma ascent. In addition, any event that modifies the local stress field (e.g. fault movement or variations in surface loading) or the physical properties of the surrounding medium can also act as a potential trigger if it promotes magma ascent. The magma and the surrounding medium are coupled. For instance, changes in magma pressure affects the local stress field and any change in the stress field leads, in turn, to a change in magma pressure, the magnitude of which depends on the shape of the magma body and compressibility of the host rocks. The strength of the crust, the size of the magma chamber and magma recharge rate play important roles in whether conditions to rupture a magma chamber and thereby trigger an eruption or not [8].

**************BOX2: HOW TO BOX : '**How can mechanical modelling be used to characterise the stability and rupture of magma chambers? '**
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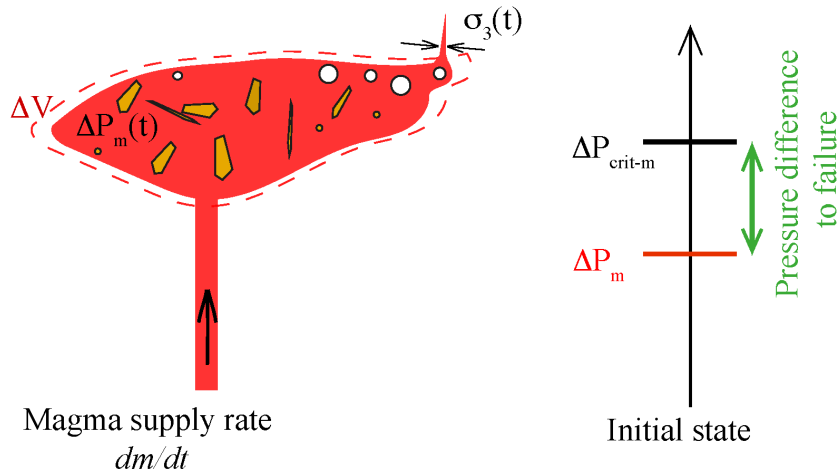
We here discuss in more detail criterion for magma intrusion and magma chamber failure.

Failure criterion
A rupture in opening mode is required to favour magma intrusion into the crust. It can be expressed as:
$\Delta P_m \geq \gamma T_s + \sigma_3$
with ΔP_m the magma overpressure (pressure in excess of the lithostatic pressure), σ_3 the minimum compressive (maximum tensile) deviatoric stress at the boundary of the magma body (here negative for tension, note that σ_3 depends on the magma overpressure), T_s being the effective tensile strength of the crust (on the order of few MPa, up to an order of magnitude lower than laboratory values for intact rocks) and γ a factor depending on the shape of the chamber wall. This rupture criterion is the same for hydraulic fracture or dike opening and thus ensuring that magma can propagate into the crust. It follows that for a given stress field and tensile strength, a **critical magma overpressure**, $\mathbf{\Delta P_{crit\_m}}$, can be defined for rupture of a chamber wall.

Changes in magma pressure ΔP_m (t)
Magma pressure changes over time and is directly linked to changes in the volume of the magma chamber through magma compressibility. Magma pressure is increased by the influx of magma or volatiles from deeper regions or by internal processes such as volatile exsolution and redistribution of exsolved volatiles within the chamber. In most cases these same processes increase buoyancy (e.g. by reducing the density of the magma or increasing its vertical extent). In addition, any slight displacement of the chamber wall due to a change in the crustal stress field acts on the chamber volume and therefore leads to a change in magma pressure.

Changes in crustal stress σ_3(t)
The crustal stress around the magma chamber changes over time in response to tectonic, topographic or magmatic forcing. Crustal strength is also strongly affected by fluid pressures and so may be locally affected by volatiles released from the chamber and changes in hydrothermal systems.

Rate of change and rheology
The crust behaves as a viscoelastic solid characterised by a relaxation time proportional to the ratio of its viscosity to its stiffness. Consequently, the rate of change of magmatic pressure and crustal stress controls the effective rheology and may also change the effective tensile strength of the crust; the crust behaves as a brittle elastic solid in response to sudden changes and as a viscous body when the timescale of change reaches values similar to or greater than the relaxation time. More complex rheologies, such as poro-elasticity, should also be considered.



Box2Figure1

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IV Various type of triggering: interplay between internal and external processes

The phenomena that trigger an eruption are distinguished as *internal* when they are directly linked to magmatic processes and associated volatiles and *external* when they are due to a perturbation initiated outside the magmatic system [9]. Figure 1 summarises the main internal (In) and external (Ex) triggers, which are described in more detail in the next two subsections. These triggers can occur at different distances from a volcanic system and act with varying temporal latency. In most cases, several triggering mechanisms may occur simultaneously, potentially influencing each other in complex feedback loops, including both internal and external processes (see below).

A complementary approach consists to distinguish bottom-up and top-down processes. In the bottom-up processes a cascading sequence of processes initiate at depth and progress towards the surface, while top-down sequence starts at or near the surface and propagate downwards. While most bottom-up processes are internal, magmatic processes may take place in the shallow parts of the magmatic system and trigger deeper events. At least in principle external triggers could be bottom-up, but the vast majority are likely to involve top-down processes.

IV a Internal triggering

When considering shallow reservoirs formed in mature volcanic systems, the most common magmatic process invoked for internally-triggered eruptions is open-system magma recharge from depth (Figure1(In)). Here deeper sourced magmas are postulated to be injected into a shallow magma chamber located in the upper crust increasing the volume and pressure [1,10]. In most studies the recharge magma is more mafic than the resident magma. However, observation of plutonic rocks indicates that silicic magma bodies are emplaced incrementally by successive injections of magma of the same composition. It is likely that some of these recharge events are associated with eruptions. Indeed petrological characteristics of some volcanic rocks suggest recharge by broadly the same magma composition.

Magma recharges advect heat and sometimes volatiles into the chamber and may induce convection. Recharged and the resident magmas may mingle and mix, producing a homogenised magma with intermediate physical and chemical properties. Recharging magma may not erupt but its existence may be indicated by other evidence, such as excess SO_2 fluxes. In addition, there is strong petrological evidence of phenocryst thermal cycling linked to successive recharges that did not lead to an eruption. In this case, each recharge seems to be part of a recurrent process that primed the reservoir for the next eruption.

Exsolution of volatiles in ascending decompressing magma enhances the prospects of magma reaching the surface. There may be multiple complementary mechanisms for volatile exsolution to make eruption favourable and thus be a trigger. Volatiles can separate such that deep fluid recharge might also trigger an eruption. Under closed-system conditions, a melt-dominated magma chamber should follow a temperature decrease, which induces magma crystallisation, producing an increase of exsolved volatiles that pressurises the magma chamber [11]. Clear examples are scarce but, probably because the increase in pressure induced by the volatiles exsolution remains relatively small compared with that induced by the influx of new magma [12]. Other mechanisms ascribed as internal triggers include the slow segregation of volatiles from saturated melts with the subsequent accumulation of exsolved volatiles on top of the upper crustal magma chambers, or the CO_2 flushing through the magmatic system [9].
Magma or fluid recharge between deeper and shallower portions seems to be a frequent phenomenon for mature systems, intrinsically related to the upper crustal magma chamber assembly. Eruptions should occur when an interconnected network of melt-dominated regions developed inside the transcrustal magmatic system, allowing transport of volatiles and melts from deeper to shallower zones at relatively short time-scales. In this context, magma recharge is a necessary condition (at least to shallow magma chamber assembly), but eruption triggering may require additional internal and external processes.

IV b External triggering

External mechanisms can modify the stress field around the magma reservoir and therefore affect the magma's ability to reach the surface (Figure 1(Ex)). Changes in stress can promote, arrest, or inhibit magmatic eruptions. The most commonly invoked external triggers are related to surface load variations or earthquakes. Changes in ground water and hydrothermal systems can lead to stress changes and fluid pressure which affect rock strength. Anthropogenic activities, meteorological events and climate change, or even asteroid impact, also are potential triggers. External triggers generally produce modest stress changes (10^3-10^7 Pa). These will only lead to eruptions in systems already close to failure.

Changes in surface loading

Many phenomena change surface loads on volcanoes. Eruptions cause major changes in the topography of active volcanoes either by accumulation of eruptive products (lava flows and domes, pyroclastic deposits, debris avalanches deposits, etc.) or by destructive processes as the formation of calderas (chapter 2.3.4) and flank or sector collapses (Figure 1). These changes in surface load are typically sudden and permanent. In addition, many volcanoes are covered by a permanent ice cap and/or seasonal snow, the thickness of which can change either seasonally or due to long-term trends (e.g. retreat of the ice during warmer periods). Variations in the water table induced by changes in precipitation can also lead to transient hydrological load effects. Some volcanoes have crater lakes of varying sizes and some are located near the sea. Tidal effects induce periodic changes in load when variations in sea level correspond to changes characterized by longer temporal trends.

Any change in the surface load modifies the state of stress in the underlying crust (see box3 'How to quantify the stress perturbation induced by variations in surface load'), which influences volcanic activity (e.g. Pinel and Jaupart [13], Sigmundsson et al. [14]). A surface load increases the compressive stress within the crust at shallow depth, acting against opening of magmatic intrusions and ascent of magma towards the surface, whereas a surface unloading, on the contrary, favours the ascent of magma at shallow depths. The local perturbation of loading stresses within the crust acts both on the magma pressure within the storage zone (a load applied to the surface induces an increase in magma pressure) and on the pressure threshold required to rupture the chamber wall and initiate magma propagation. The resulting effect is therefore highly dependent on the shape of the storage zone (which is often poorly known) and the compressibility of the magma. In addition, changes in the stress field have a strong influence on magma path towards the surface, which affects the location of eruptive vents.

Earthquake triggering

Tectonic earthquakes may significantly affect the stress field around a volcano, thereby triggering unrest phenomena (e.g. seismicity, surface deformation, degassing), changes in eruption style or possibly a new eruption (see Seropian et al. [15]). Stress transfer from an earthquake may be either static or dynamic.

Static stresses result from the permanent deformation of the rocks, after the seismic waves have passed. Given that static stresses decay rapidly with distance from the epicentre ($\propto r^{-3}$), they can only trigger an eruption at volcanoes within a few fault lengths of the epicentre. Static stresses may be either compressive or extensional. Compressive stresses may aid magma ascent by squeezing magma upwards, but can also inhibit dike opening. Extensional stresses facilitate dike propagation and also favours bubble nucleation and growth, therefore increasing overpressure in a magma body. Local permeability may also be increased, enhancing the upwards advection of volatiles and movement of hydrothermal fluids.

Dynamic stresses are produced by the passage of seismic waves. They are temporary (seconds to minutes) and oscillatory by nature. The amplitude of dynamic stresses falls off much more gradually with distance from the epicentre ($\propto r^{-1.66}$) than static stress and can therefore trigger eruptions at much greater distances. Dynamic mechanisms are varied, and depend on both the earthquake and volcano characteristics. There are three broad mechanisms: (1) magmatic volatiles nucleation, (2) resonance and (3) hydrothermal processes. Volatile-related processes include a series of mechanisms related to bubbles in magma (e.g. bubble nucleation). These are particularly important for low viscosity, supersaturated magmas. Resonance processes will occur when the seismic wave frequencies match the volcanic edifice's natural frequency. In this case, the edifice movement is greatly augmented, which could lead to increased melt and volatile migration in the plumbing system or sloshing in the conduit. Given the right conditions, these processes could trigger an eruption. Finally, volcanoes that host persistent hydrothermal systems are more prone to dynamic triggering. Hydrothermal systems are particularly sensitive to seismic waves, and the stress oscillations could induce rapid permeability, pressure and phase changes. Destabilisation of the hydrothermal systems could lead to a variety of unrest phenomena, and trigger top-down pressurisation of the whole magmatic system, potentially resulting in a magmatic eruption.

Anthropic activity and climate change

The impact of anthropogenic climate change on volcanic activity is receiving increased attention [16]. Some phenomena such as the retreat of ice sheets and the rise in sea level produce changes in surface loads that can be easily quantified. Other processes remain difficult to constrain, for instance the influence of local increases in extreme precipitation that modify pore pressure and the mechanical behaviour of rocks, thereby affecting the slope stability. Some human activities, such as drilling or damming lakes, may also have a direct impact on magmatic reservoirs. For example, magma erupted through a geothermal borehole in 1977 at Námafjall, Iceland during a rifting event in Iceland [17].

Potential impact of extraterrestrial events

Asteroids also have the potential to induce large dynamic stresses on volcanoes, although these events are extremely rare. The only known example is the Chicxulub impact, 66 Ma ago, which potentially favoured eruption of the Deccan Traps.

Proving causality between a given internal or external event and volcanic eruption remains a formidable challenge. Volcanic eruptions result from a long chain of complex, intertwined processes, and a single event cannot be the only culprit. It would therefore be more appropriate to speak of eruptions that are facilitated or promoted, rather than eruptions that are triggered [18]. All the more so as several factors are generally at play simultaneously.

IV c Multiple triggering

Several mechanisms can act together or in sequence to trigger an eruption. There are multiple examples of one trigger inducing another, with possible feedback. Tectonic earthquakes may affect the environment around a volcano by, for instance, rupturing gas reservoirs at depth or causing landslides. Such seismically-triggered non-volcanic events can in turn affect the magmatic system. A landslide can decompress the magmatic reservoir. Landslides can be induced by episodes of heavy rainfall or edifice deformation from very shallow magma intrusion as happened on 18th May

1980 at Mount St Helens. The latter case corresponds to a mixture of internal and external triggering.

**************BOX3 HOW TO BOX : '**How to quantify the stress perturbation induced by variations in surface load**' ******************************

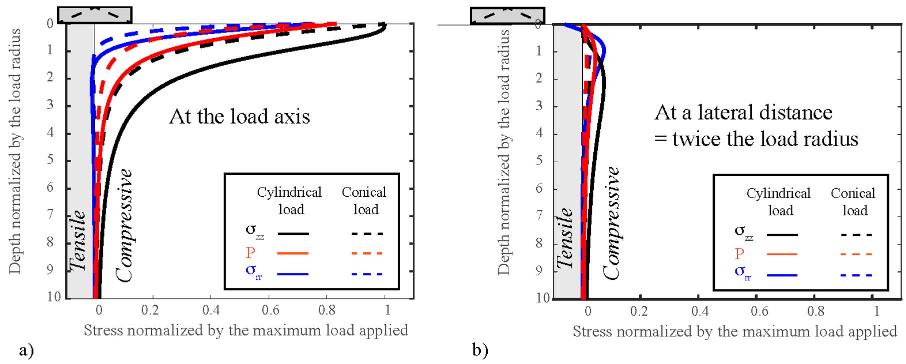
Stress perturbations induced by surface load variations applied on an elastic half-space can be inferred from the solution to a surface point load, known as a solution to the Boussinesq problem. In the case of a load of axisymmetric geometry applied at the surface of an elastic half-space, the induced stress field can be derived analytically considering various load distribution by numerical integration (see for instance Matlab codes provided on the repository https://zenodo.org/records/8348775). The pressure within the crust below the centre of a circular load of radius R, thickness H and density ρ is expressed as a function of the depth z by:

$$P(r=0,z) = \frac{2}{3}\rho g H(1+\nu)\left[1 - \frac{z}{(R^2+z^2)^{\frac{1}{2}}}\right]$$

with g the gravitational acceleration and ν the Poisson's ratio of the elastic medium. This equation can also be used with a negative thickness to calculate the depressurisation due to surface unloading like the formation of a caldera or a flank sector collapse.

The equation shows that the magnitude of the stress perturbation at depth depends directly on the magnitude of the load. The effect is greatest at a shallow level and decreases with depth at a rate depending on the lateral extent of the surface load; a larger load produces a deeper effect. This behaviour can be seen in figure Box3Figure1a, which displays the vertical (σ_{zz}) and horizontal (σ_{rr}) stress components (positive stress corresponding to compression as induced by surface loading) and the pressure (P) as a function of depth along the vertical axis at the centre of the applied load. The pressure perturbation becomes less than one tenth of the surface load at a depth greater than twice the radius of the load for a circular shape. This occurs at a shallower depth (once the depth reaches the radius of the load) when considering a conical load. In addition, the stress perturbation induced by the load decreases as one moves laterally away from the load, as shown in figure Box3Figure1b. Once again, the stress perturbation is already reduced by a factor of ten at a lateral distance equal to twice the load radius. Applying a surface discharge produces the same effect but with the opposite sign (decompression).

The elastic response provides the Earth's initial response to the load variation, but viscoelastic relaxation of the deviatoric stress is expected over time. The characteristic relaxation time, which depends on the viscosity and elastic stiffness of the crust, is longer at shallow depths within the upper brittle crust than within the lower crust, where it is expected to remain longer than a few years. Thus, stress perturbations induced by sudden or seasonal variations in surface loading occurring over a lateral extent of about 10 km can be reasonably well estimated using an elastic crust model.



Box3Figure1

V Importance of the current state and dynamic of the volcanic system on its sensitivity to triggers-Unrest, phreatic/magmatic eruptions

On many volcanoes, episodes of unrest are much more common than eruptions and are often interpreted as failed eruptions. There are also many examples of phreatic eruptions, which are sometimes viewed as a part of unrest, and attributed to the interactions of shallow magma with hydrothermal systems or groundwater. Very shallow stalled magma can heat groundwaters, re-invigorate hydrothermal systems and transfer magmatic gases to hydrothermal systems (e.g. Stix & de Moor [19]). Although phreatic activity is not always followed by a magmatic eruption, and is sometimes considered a failed eruption, if phreatic explosions are sufficiently vigorous, new pathways and conduits can be created that allow magma to erupt. In this case, the phreatic explosions may be viewed as the direct trigger of the magmatic eruption.

With the exception of a major magma recharge from depth, which can force a system far from the critical state to erupt, any event, internal or external, can only trigger a volcanic eruption if the magma system is already in a critical state, meaning that the volcanic system is already sufficiently close to eruption. The ability of a given event to actually trigger an eruption depends on the magnitude of the change in stress relative to how close to rupture the system already is, as illustrated in Figure 2. Since variations in static stresses decrease rapidly with distance from the source (see the detailed explanation in the case of surface load variations in the box3 'How to quantify the stress perturbation induced by surface load variations'), many external triggers, with the exception of those that cause transient dynamic stresses (earthquakes, meteorites, etc.), will only have an effect if they are located at a short distance from the magma plumbing system and mostly on the upper part of the plumbing system.

Schematically, a magma system can be considered to be in a given initial state (Figure 2a), either stable over time (solid red line), or evolving over time at different rates (dashed or dotted red lines) and moving towards or away from the critical state (blue lines), and thus towards or away from eruptive conditions. An eruption occurs when the magma system reaches a critical state, which can also evolve over time. If the system is evolving towards an eruption, the timing of the eruption depends on the rate of evolution. Any event that reduces the difference between the critical stress required for rupture and the stress in the magma system favours an eruption by reducing the time remaining before the eruptive event. A distinction can be made between:

- sudden events (Figure 2b), such as fault movements, edifice flank collapses or sudden magma recharge, which can directly trigger an eruption if the magnitude of the change in stress is sufficiently large, and can otherwise promote an eruption.
- seasonal events (Figure 2c), such as variations in snow thickness, which should only modify the timing of an eruption for systems characterised by slow evolution.
- long-term trends (Figure 2d), such as tectonic forcing, glacier retreat, the construction of volcanic edifices or magma/volatile accumulation at depth, which may lead to an earlier-than-expected eruption (Figure 2d). In this case, the key parameter is the rate of change being considered (for example, the rate of magma recharge).

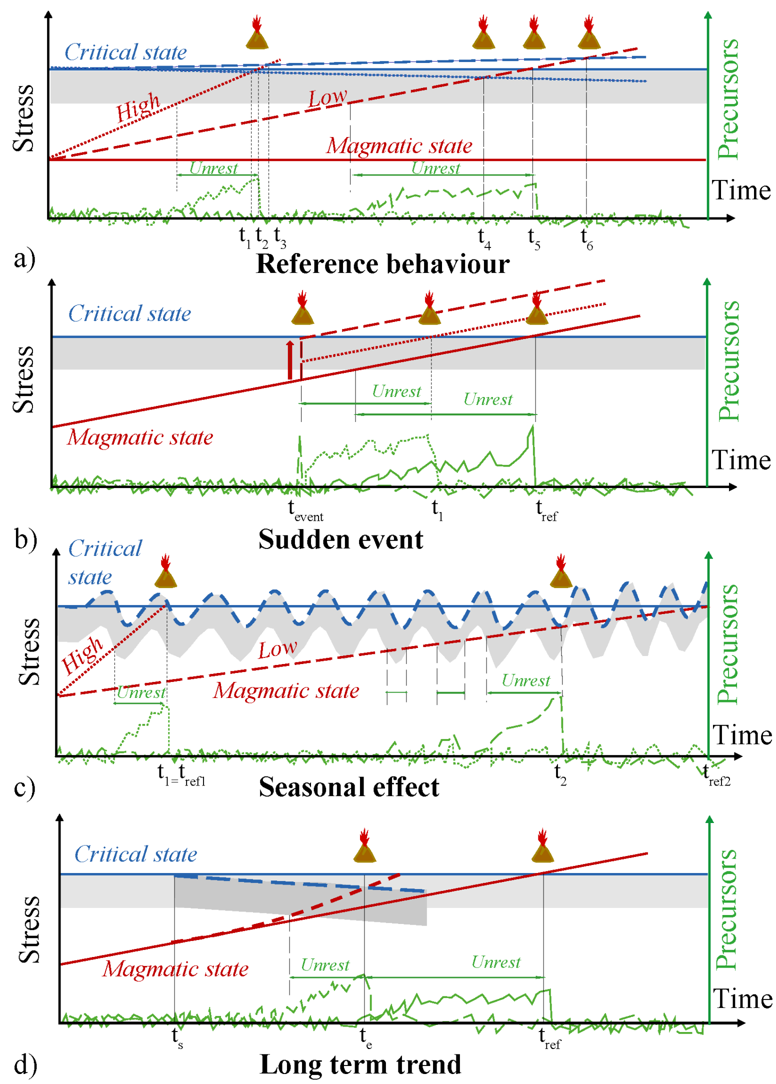


Figure 2. Sensitivity to various triggers of volcanic systems: a) Reference behaviour, Initial state and dynamics: An eruption occurs when the magma system reaches a critical state. If the critical state and the magma system do not change over time, no eruption is expected (solid line). The magma system and the critical state (depending on the surrounding medium) can evolve over time (towards or away from the eruption) at different rates (which should be lower for the critical state) leading to different timing of the eruption. As the state of the system approaches the critical state (grey area), it is expected that unrest will occur, leading to the appearance of precursors with certain recorded signals exceeding the background level (green lines). A higher rate of increase in the stress of the magma system will always result in an earlier event (red dotted line) and shorter unrest period. b) **Sudden event favouring an eruption, occurring at time t_{event} and affecting only the magmatic system** (e.g. internal trigger as sudden magma inflow): Depending on the amplitude of the change in stress, it can either induce an eruption without unrest period (dashed line), or reduce the time before the next eruption and shorten the period of unrest (dotted line, eruptive time $t_1<t_{ref}$, with t_{ref} the eruptive type in the absence of trigger). c) **Seasonal effect**

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modulating the critical state (e.g. snow load): The eruption timing is changed only for magmatic systems with a low evolution compared to the amplitude of the seasonal change ($t_2<t_{ref2}$), in this case eruption is more likely in the period favouring the eruption but the eruption timing is not affected when the magmatic system has a rapid evolution ($t_1=t_{ref1}$). Note that seasonal modulation can also affect the magmatic system evolution. d) **"Long term" trend affecting both the magmatic system and the surrounding medium** (e.g. glacier retreat) **and starting at time t_s**. If it favours an eruption, it may lead to an earlier eruption than expected ($t_e<t_{ref}$) and shorter period of unrest.

VI Recent advances and remaining challenges

Understanding how eruptions are triggered has made significant progress from improved volcano data recording, detailed petrological characterization of eruptive products, and advances in modelling framed by a new emerging paradigm of volcanic systems. The boom in remote sensing data has greatly increased the number of volcanoes for which geophysical information is available and has enabled systematic checking *a posteriori* for precursor signals and the identification of potential triggers by improved statistical analysis. Ongoing efforts to interpret geophysical and petrological monitoring to track fluids transfer at depth are also very promising. In seismology, new imaging tools (e.g. coda-wave interferometry) and improved detection of low-frequency deep earthquakes (e.g. Melnik et al. [20]) have advanced understanding of the deep roots of magma plumbing systems and the detection of subtle changes prior to eruptions. These precursors could be related to segregation of volatiles from melts. At depth, segregated volatiles can ascend independently of the melt from which it was extracted, driving volcanic phenomena. Evidence for such deep fluids comes from geochemistry such as the presence of Helium isotopes ratios indicative of a mantle source. This raises the possibility of deep fluid recharge rather than magma recharge as a triggering mechanism. The petrological characterization of volcanic products, based on high resolution images and microanalysis, allows us to constrain timing of magma recharge (by using diffusion chronometry) and to identify the cyclic behaviour of recharge events that prepare the reservoir for the next eruption. Numerical modelling of magmatic systems is also making significant progress, particularly with regard to the influence and transport of volatile substances and the inclusion of more realistic rheologies resulting from the new vision of the plumbing system. Laboratory fluid experiments are used to determine how the distribution of crystals and bubbles (by nucleation, growth and coalescence) are affected by changes in the stress field and have an impact on the physical properties of magma. This will undoubtedly improve our understanding of triggering mechanisms in the years to come. The rapid changes in the environment could also challenge perception of triggering.

Summary

The prerequisite for an eruption to occur is the transfer of a significant amount of magma from the melting zones, at the deep roots of the magmatic system, to the surface by buoyancy. All the processes that promote the ascent of magma from the reservoirs, leading either to the recharge of a shallower chamber, the destabilisation of the hydrothermal system in preparation for a phreatic eruption, or an eruption at the surface, can be considered as internal triggers. Volatiles play a key role in these internal triggers increasing both the buoyancy of the magma and its driving pressure, volatiles can also separate and move ahead of the magma, creating a path for the magma to follow. External triggering by any event/process acting on the magma's stress field and pressure can promote eruptions mostly acting on the upper part of the plumbing system. Some triggering mechanisms are fairly well understood (e.g. new magma inflow), while others remain less well identified. In any case, triggering is only possible for systems that are close to eruption, so a succession of several episodes of triggering events, sometimes spread out over time, may be necessary to move the systems towards eruption. Because of the interaction between several factors, both internal and external, it is often difficult to identify a dominant trigger and multiple triggers are likely to be the rule.

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[Additional references, further reading to be included in companion website:
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Additional material to be added to companion website (e.g. further reading, data, images, figures, videos):
Video1 : Animated version of figure 1
Machine narration in the animation is for submission purposes only; the final version will be in human voice.

Glossary:

Banded pumice: solid physical and chemical mixtures obtained by mixing two magmatic compositions, generally present in the form of discrete bands.
Brittle crust: upper part of the crust where rocks can fracture when stresses exceed the fracture toughness whereas in the ductile lower crust, rocks deform ductilely by creep.
Critical state: state of the volcanic system in which the conditions for an eruption are fulfilled.

Elastic behaviour: The rocks are elastic when their stress varies linearly with strain. Many models of how magma chamber walls react to overpressure and many models of dyke propagation assume that the crust is elastic.
Eruption trigger: phenomena or processes that increase the probability, or amount, of magma arriving at the surface or at shallow depth. Eruption triggers are distinguished as *internal* when they are directly linked to magmatic processes and associated volatiles and *external* when they are due to a perturbation initiated outside the magmatic system.
Liquefaction: Transformation of a magma mush into a liquid eruptible magma, for example due to an earthquake.
Magma buoyancy: upward force exerted on the magma body corresponding to the weight of the surrounding medium, crust or mantle, occupied by the magma according to Archimedes' principle. When the density of the magma is lower than that of the surrounding medium, this force exceeds the weight of the magma and the resulting force is directed upwards, promoting the vertical ascent of the magma.
Magma overpressure: magma pressure in excess of the lithostatic pressure.
Magma reservoir: any region in which melt and magmatic fluids are present potentially with a large proportion of crystals (mush). In the case of a volume containing eruptible magma, the term chamber is preferred.
Plutonic system: part of the magma system, which crystallised entirely at depth, without ever reaching the surface.
Precursors: change in the activity of a volcano (e.g. amount or composition of degassing) or geophysical signals (e.g. seismicity, ground deformation, etc) that occur before a volcanic eruption.
Recharge: addition of new magma to a volume of magma accumulated at depth in a storage zone.
Sloshing: resonant free-surface flows. Large earthquakes can potentially slosh a magma reservoir with a free surface or density-stratification, which can trigger an eruption.
Stress field: spatial distribution of internal forces in a body defined by the stress tensor as a function of position, knowledge of which makes it possible to quantify the surface force acting on a given surface.
Tectonic forcing: external stress induced by the movement of tectonic plates and the resulting stress field: e.g. extension in rift zones, compression in subduction zones.
Transcrustal magmatic system: magma system extending through the crust, comprising several reservoirs and their connections, made of melt, crystals, and exsolved volatiles that are heterogeneously distributed in space and time.
Viscoelastic behaviour: The rocks are viscoelastic when their response to stress is both viscous and elastic, resulting in a time-dependent relationship between stress and strain.
Volatile exsolution: the physical process by which dissolved gas within the melt forms gas bubbles. It occurs when the volatile concentration exceeds its solubility in the melt, through pressure decrease, temperature increase or a change in composition.
Volcanic unrest: significant deviation from the normal or reference state of a volcano; during an unrest, one or more signals (e.g seismicity, degassing) clearly deviate from the background level.