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Recognizing and tracking volcanic hazards related to non-magmatic unrest: a review

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Abstract

Eruption forecasting is a major goal in volcanology. Logically, but unfortunately, forecasting hazards related to non-magmatic unrest is too often overshadowed by eruption forecasting, although many volcanoes often pass through states of non-eruptive and non-magmatic unrest for various and prolonged periods of time. Volcanic hazards related to non-magmatic unrest can be highly violent and/or destructive (e.g., phreatic eruptions, secondary lahars), can lead into magmatic and eventually eruptive unrest, and can be more difficult to forecast than magmatic unrest, for various reasons. The duration of a state of non-magmatic unrest and the cause, type and locus of hazardous events can be highly variable. Moreover, non-magmatic hazards can be related to factors external to the volcano (e.g., climate, earthquake). So far, monitoring networks are often limited to the usual seismic-ground deformation-gas network, whereas recognizing indicators for non-magmatic unrest requires additional approaches. In this study we summarize non-magmatic unrest processes and potential indicators for related hazards. We propose an event-tree to classify non-magmatic unrest, which aims to cover all major hazardous outcomes. This structure could become useful for future probabilistic non-magmatic hazard assessments, and might reveal clues for future monitoring strategies.

Keywords: Non-magmatic unrest; Volcanic hazard; Forecasting; Volcanic surveillance; Event tree

Introduction

During long inter-eruptive periods of hundreds to thousands of years, a volcano passes through different stages of activity including periods of dormancy, quiescence, reawakening and unrest. Due to this very long-term behavior, hazard forecasting becomes challenging, in particular for short-term time frames (days to few months). So far, most volcanic hazard assessments have focused on magmatic unrest (e.g., Sparks 2003). Effective hazard assessment and risk mitigation during unrest depends on the early and reliable identification of changes in volcanic dynamics and their recognition as potential precursors to a hazardous event (Selva et al. 2012). Major uncertainties in the identification of the causative processes of unrest translate into uncertainties in short-term forecasting. The problem is made even more complicated by the intrinsic and almost inevitable subjectivity in the definition of unrest. Phillipson et al. (2013) defined volcanic unrest as the "deviation from the background behavior of a volcano towards a behavior, which is cause for concern in the short-term because it might be a prelude to an eruption", a concept that we adopt in this study. Unrest depends also on the volcano itself, reflected by its background activity, and should thus be defined for each volcano separately (e.g., Potter 2014; Sandri et al. 2014).

If the cause of concern during a stage of volcanic unrest is the recognition of the migration from a magma reservoir (Figure 1), the volcano is in a stage of magmatic unrest. Hence, this process does not attest to processes and characteristics that are intrinsic of stagnant cooling magma batches at given P-T conditions such as convection or crystallization, which may contribute to degassing (Figure 1). If no evidence for "magma-on-the-move" exists, but concern remains high for any other reason, the volcano is in a phase of non-magmatic unrest. This apparently banal but highly practical distinction does not imply that recognizing non-magmatic unrest is easy. One important task of this study is to shed light on the detectable indicators of non-magmatic unrest in its various ways.

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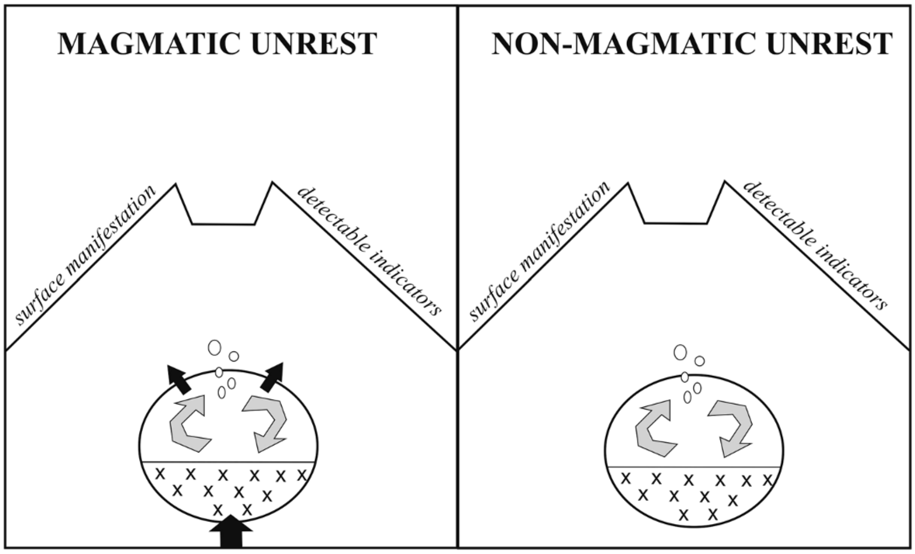

Figure 1 Magmatic vs non-magmatic unrest. A volcano is in a stage of magmatic unrest if volcanic unrest is caused by the migration from a magma reservoir (left), which we are able to recognize as such. If we cannot recognize magma migration (regardless whether it migrates or not), but concern remains high for any other reason, the volcano will be in a state of non-magmatic unrest (right). Magma convection and crystallization in a magma chamber are often not recognized as magma migrations, and thus would results in non-magmatic unrest.

Probabilistic methods can be applied to forecast volcanic activity across different timescales. Several tools have been developed and applied over the past decade including event trees (Newhall and Hoblitt 2002), some focusing only on magmatic eruptions (Bayesian Belief Network, BBN, Aspinall et al. 2003; Bayesian Event Tree for Eruption Forecasting, BET_EF, Marzocchi et al. 2004a, 2008; Sandri et al. 2009, 2012; Selva et al. 2012; Marzocchi and Bebbington 2012), others also focusing on long-term hazard assessments (Bayesian Event Tree for Volcanic Hazard, BET_VH, Marzocchi et al. 2010; Selva et al. 2010; Sandri et al. 2014; HASSET, Sobradelo et al. 2014). So far, the schemes for short-term hazard assessment have put only minor emphasis on evaluating phenomena associated to non-magmatic hazards, despite the fact that these have had significant economic and social impact in the past. Prominent examples of activity at volcanoes which did not probably involve magmatic eruptions include the phreatic activity at Soufrière de Guadeloupe in 1976 (Shepherd et al. 1979; Shepherd and Sigurdsson 1982), non-eruptive unrest at Campi Flegrei in the 1980's (Barberi et al. 1984; Dvorak and Mastrolorenzo 1991; Orsi et al. 1999; Chiodini et al. 2001, 2003), and the flank collapse of Casita volcano in 1998 (Kerle and van Wijck de Vries 2001), to name a few.

This paper reviews various non-magmatic unrest phenomena, in order to help recognize and track potentially hazardous outcomes. We will present an event tree to map the evolution of non-magmatic unrest along different stages, with an increasing level of detail. This event tree has not yet been applied to probabilistic hazard forecasting during stages of non-magmatic unrest, although it is intended to become the basis for a future BET_UNREST code. Adding this branch to the BET (Bayesian Even Tree) provides a means to assess all hazardous outcomes that should be considered in real-time updating of the BET model. The background idea is that "magma-on-the-move" is a necessity before a magmatic eruption. However, it is not a necessity for the evolution of different stages of non-magmatic unrest. Hence, we stress the importance of recognizing magma on the move to distinguish between magmatic and non-magmatic unrest. Here we present a concept for the recognition of precursory symptoms and resultant potential threats from non-magmatic unrest activity.

Non-magmatic hydrothermal unrest

Many volcanoes hosting hydrothermal systems are in a state of quiescence (i.e. background activity) during prolonged inter-eruptive phases, manifested as low-temperature fumarolic emissions (Giggenbach et al. 1990; Sturchio and Williams 1990; Giggenbach and Corrales Soto 1992; Sturchio et al. 1993; Fischer et al. 1997; Lewicki et al. 2000; Rouwet et al. 2009; Joseph et al. 2011, 2013; Chiodini et al. 2012), diffuse CO_2 soil degassing and steaming ground (Cardellini et al. 2003; Werner et al. 2003, 2008; Bergfeld et al. 2006, 2012; Werner and Cardellini 2006; Lewicki et al. 2007a, 2007b; Mazot et al. 2011; Inguaggiato et al. 2012; Lewicki and Hilley 2014), thermal spring discharges (Taran et al. 2008; Taran and Peiffer 2009), low-activity crater lakes (Pasternack and Varekamp 1997; Taran et al.

1998; Stimac et al. 2004; Rouwet et al. 2004, 2008, 2014; Taran and Rouwet 2008), and/or hydrothermal alteration (Getahun et al. 1996; Gehring et al. 1999; Salaün et al. 2011). Volcano-hydrothermal systems seem to develop more frequently in closed-conduit dome complexes and calderas (e.g. Yellowstone, Long Valley and Campi Flegrei), rather than cone-shaped central-conduit stratovolcanoes (Rouwet et al. 2015), and surface manifestations often reflect the local tectonic regime of the volcano (e.g., Mazot et al. 2011). Despite the relatively non-hazardous nature of such prolonged hydrothermal background activity, a profound knowledge of quiescent activity of individual volcanoes is necessary to understand the short- (daily) and long-term (seasonal, yearly) baseline behavior from which to deviate when evolving towards a state that causes concern.

Hydrothermal eruptive unrest

Eruptions during stages of non-magmatic unrest, by definition, do not deal with lava or juvenile tephra emission, as no magma migration is involved. However, besides lava and juvenile tephra, volcanoes are able to release fluids in an effusive or explosive manner. Below, we describe several observed mechanisms leading to such events.

Effusive eruptive unrest

Water effusion Volcano-hydrothermal systems can manifest non-explosive expulsion of water. A rise of the water level in volcano-hydrothermal systems can be caused by the interplay of (1) increased or prolonged infiltration of meteoric water into a hydrothermal system, (2) increased or prolonged condensation of steam into a near-surface aquifer, (3) variations in the vapor pressure regime in a hydrothermal system, or (4) the rise of perched aquifers due to buoyancy effects of a rising bubble-rich or vapor-rich fluid front.

Variations in the phreatic level of an aquifer often pass visually unobserved, as the effect does not necessarily reach the surface. Nevertheless, changes in the water saturation state in a volcanic edifice can drastically change its mechanical and hydraulic conditions (Reid 2004). These variations in hydraulic regime in a volcano can be detected through self potential, microgravity, resistivity or VLF surveys (Finizola et al. 2003; Révil et al. 2004; Zlotnicki et al. 2006; Gottsmann et al. 2007; Fournier et al. 2009; Villasante-Marcos et al. 2014). Water level gauging in wells or flow rate measurements from (thermal) springs at volcano flanks is the most direct way, although few examples of frequent monitoring exist (Ingebritsen et al. 2001; Hurwitz et al. 2002; Taran and Peiffer 2009), which inevitably leads to the need for numerical modeling procedures to increase theoretical insights (Hurwitz et al. 2003; Todesco and Berrino 2005).

Variations of the water level in a hydrothermal system is most obvious when aquifers intersect the surface, through direct outpouring of water from wells or springs, or variations in the water level of crater lakes. El Chichón volcano (Mexico), in a state of non-magmatic unrest since soon after the 1982 Plinian eruptions, hosts a boiling spring which periodically discharges water towards its crater lake (Rouwet et al. 2004, 2008) (Figure 2). This particular dynamics results from the prolonged infiltration of meteoric water into the volcano-hydrothermal system which, together with the local boiling regime, creates a steam cap that leads to non-explosive water expulsion.

An intrinsically gravitationally unstable system is manifested as a near-boiling lake that occasionally drains and refills its water, without necessarily a clear cycle (in contrast to geysers) (Inferno Lake, New Zealand, Vandemeulebrouck et al. 2005; Boiling Lake, Dominica, Fournier et al. 2009; Di Napoli et al. 2013). Such lakes can be sustained above the regional aquifer by the drag force caused by a gas phase flowing through a liquid-filled permeable conduit. At equilibrium, the hydrostatic pressure of the water raised above the water table counterbalances the drag stress for each bubble. In order words, bubbles "carry" the lake. Moreover, denser cold water near the lake will be lifted by underlying hotter water, which needs to boil and bubble to keep the unstable cooler lake suspended.

Poás' crater lake Laguna Caliente peculiarly overflowed its stable and steep-walled crater lake basin from January to April 2005 (despite the dry season), a year before the onset of the 2006-ongoing phreatic eruption cycle. This sudden rise was probably induced by an upwelling vapor front beneath the crater lake, or the injection of a liquid (direct or as condensed steam) into the cooler lake.

Sulfur volcanism Sulfur is a low viscosity liquid between ~116° and ~159°C (Oppenheimer 1992; Takano et al. 1994). If a fumarolic system is heated to >116°C preexisting sulfur deposits in vugs and vents can be remobilized, manifested as a sulfur flow at the surface (Naranjo 1985; Oppenheimer 1992; Harris et al. 2004). This feature is a clear sign of heating of a previously colder system, and thus often the onset of a state of non-magmatic unrest. Recently, sulfur flows were observed at White Island (New Zealand, 2013); Poás (Costa Rica, in May 2005, 10 months before the onset of the 2006-ongoing phreatic eruption cycle) (Figure 3), Turrialba (Costa Rica, the morning of the 11 January 2012 phreatic eruption, G. González pers. comm.), andat Sulphur Springs, Soufrière (Dominica) in 1994, four years before an intense seismic swarm (J. Lindsay pers. comm.). "Sulfur volcanism" is exhibited as spectacular sulfur volcanoes, boiling sulfur ponds, or liquid sulfur pools at the bottom of active crater lakes, manifested at the surface as floating sulfur spherules (e.g., Lake Yugama, Kusatsu-Shirane volcano, Japan, Takano et al. 1994; Poás,

Figure 2 Water expulsion from a boiling spring discharging towards the crater lake (top) inside the active crater of El Chichón volcano, Mexico. The image in the inset shows a detail of the spring (Picture by Yuri Taran, November 2009).

Oppenheimer and Stevenson 1989, Oppenheimer 1992; Keli Mutu, Pasternack and Varekamp 1994).

Explosive eruptive unrest

Phreatomagmatic eruptions A phreatomagmatic eruption implies the presence of magma, and thus seemingly a state of magmatic unrest. Unfortunately, phreatomagmatic eruptions can occur without a clear precursor during a phase of non-magmatic unrest, (i.e. no signal of magma migration is recognized) as a major trigger mechanism for phreatomagmatic eruptions is decompression. If a stagnant magma body is present near the surface (common in the most active hydrothermal systems) it can easily be triggered into phreatomagmatic activity.

There are examples of decompression triggered phreatomagmatic eruptions, which occurred during non-magmatic hydrothermal unrest. During the period before the 25 September 2007 phreatomagmatic eruption of Ruapehu (New Zealand) the volcano was seismically quiet and the crater lake water temperature was low (13°C, Christenson

Figure 3 Deposit of the May 2005 sulfur flow at Poás volcano (Picture by D.R.).

et al. 2010; Jolly et al. 2010), i.e. at that point, lacking any sign of magmatic unrest (Figure 4). The volcano experienced at that time a prolonged period of non-magmatic hydrothermal unrest. Volcano-tectonic earthquakes and volcanic tremor occurred only 10 minutes and 1 minute, respectively, before the eruption (Jolly et al. 2010). Within the practicality of volcano monitoring, precursory signals for such phreatomagmatic eruptions remain extremely difficult to recognize in time. The eruption itself was of short duration (~1 minute), but violent, and injured two climbers. The phreatomagmatic eruption partially expelled the crater lake, triggering two lahars down Ruapehu's flanks (see section Lahars, volcanic debris flows, floods and jökulhlaups) (Figure 4) (Kilgour et al. 2010).

Phreatic eruptions Several definitions are available for phreatic eruptions (e.g., Barberi et al. 1992; Mastin 1991; Browne and Lawless 2001; Rouwet and Morrissey 2015, and references therein) which are not necessarily mutually consistent. Phreatic eruptions are triggered by the input of fluids and heat of magmatic origin into a shallow aquifer (sometimes into a lake), followed by overpressurization of the hydrothermal system, but without the eruption of juvenile magmatic material (Figure 5). In many cases, phreatic eruptions are ubiquitous precursors to magmatic eruptions of both explosive or effusive nature, or could serve as the decompression mechanism prior to phreatomagmatic eruptions. Phreatic eruptions often occur during prolonged periods of non-magmatic hydrothermal unrest (e.g., Pisciarelli, Campi Flegrei), and can occur as a single, major event, or as minor events within a phreatic eruption cycle (Rouwet et al. 2014). Given the state of hydrothermal unrest and the high compressibility of fluid phases (gas, vapor and water), precursory signals are buffered by the hydrothermal system within the time-frame hydrothermal systems are often monitored (yearly, monthly or weekly in the best case). The input of mass and heat through a magmatic fluid pulse into a shallow aquifer, the instigator of a phreatic eruption, is often a short-term and too low-amplitude event which passes unobserved within the time-frame of monitoring set-ups, or discontinuous geophysical surveys will not be able to detect a minute to hour-scale sudden heat and fluid input. Continuous temperature monitoring of hydrothermal systems (e.g., real-time FLIR-imaging, T-gauging) could perhaps reveal a short-term precursor for phreatic eruptions (Ramírez et al. 2013).

Secondary mineral precipitation after prolonged alteration, or the presence of low-permeable elemental sulfur can seal hydrothermal systems by reducing permeability of country rock increasing the possibility of localized pressurization, one of the possible constraints to reach a pressure threshold prior to phreatic eruptive activity. Monitoring the evolution of alteration mineralogy, microgravity, hydrology and fluid geochemistry could lead to an indication of the most probable locus and timing of a future phreatic eruption.

Hydrothermal explosions Some volcano craters contain subaerial or sublacustrine geyser-like boiling springs. Geyser eruptions are boiling-point eruptions, which only expel water (Mastin 1995). These non-violent eruptions

Figure 4 View of the Ruapehu summit area after the September 2007 "surprise" phreatomagmatic eruption. The yellow arrow indicates the tephra fall; the open white arrow indicates the proximal deposit of the lahar, that was generated after partial crater lake expulsion due to the phreatomagmatic eruption (filled white arrow) (Picture by Karoly Németh).

Figure 5 Phreatic eruption at Poás volcano breaching the Laguna Caliente crater lake (Picture by A.B. Castro). Phreatic eruptions may result in pyroclastic density currents and non-juvenile tephra fall out.

occur when a buried fluid near boiling conditions is depressurized leading to the creation of a bi-phase liquid–vapor mixture, expansion and, finally, explosion (White 1967; Kieffer 1984). If we assume a constant water recharge and heating during intereruptive periods, what causes the cyclicity of geyser eruptions? A key property of the geyser plumbing system is its heterogeneous geometry and/or permeability (Kieffer 1984; Ingebritsen and Rojstaczer 1993, 1996). During reservoir refill, this leads to a discontinuous rise in water level, and thus a step-wise variation in hydrostatic pressure: boiling will eventually occur when higher permeable zones (e.g., fractures) are filled slowly, while boiling will be suppressed when less permeable parts (e.g., narrow conduits) are quickly filled (Kieffer 1984; Ingebritsen and Rojstaczer 1993). Brown et al. (1989) and Dowden et al. (1991) concluded that frequent (every 10 min) and short (10 sec) geyser-like eruptions offered an efficient means to dissipate energy at the 1985–1988 Laguna Caliente crater lake (Poás volcano, Costa Rica). Overlying liquid water will be disrupted when a submerged vent or fumarole filled with vapor under pressure is subjected to a pressure release paired with the water ejected upward (Dowden et al. 1991). It is noteworthy that Laguna Caliente passed a phase of lake level drop, due to enhanced lake evaporation that steadily decreased the hydrostatic pressure, teasing the underlying system with near-boiling conditions, and thus potential geysering. During complete dry-out of the lake in 1989, Poás exhibited nearly continuous geysering (Dowden et al. 1991). Since 2006, Laguna Caliente exhibits a similar behavior (Rymer et al. 2009): a phreatic eruption cycle and contemporaneous lake level decrease. It sounds reasonable that the ongoing periodical fluid injections in more peaceful manner into Laguna Caliente are controlled in some way by a geyser-like mechanism. Consequently, more powerful eruptions at a crater lake are rather phreatic, which can eventually destroy geyser plumbing systems.

Limnic Nyos-type gas release Deep lakes in maars, craters and calderas can become hazardous if they are fed by gas-rich (mainly CO_2) regional meteoric groundwater (Tassi and Rouwet 2014). Due to the hydrostatic pressure of the lake water column, the entering gas remains dissolved in the lake bottom waters (hypolimnion). The lethal gas (CO_2 is an asfixiating gas, denser than air) can be released from the lake when (1) the lake will periodically overturn during the cold season, when colder and thus denser surface waters sink towards the lake bottom, (2) the dissolved gas pressure at depth exceeds the hydrostatic pressure (spontaneous release by supersaturation), or (3) an external trigger disturbs the chemical and thermal lake stratification (e.g., an earthquake, rock fall into the lake, strong winds, internal waves, the sudden input of cold rain water during rainstorms, etc.) (see Kusakabe 2015, for a review). Explosive gas release occurred in 1984 and 1986 at Lake Monoun and Lake Nyos (Cameroon), respectively. The Lake Nyos event killed >1,800 people by CO_2-asfixiation (Kling et al. 1987).

At the lake surface, "Nyos-type" lakes appear peaceful, and the only way to recognize a potential CO_2-accumulation in bottom waters is by lowering a CTD-probe (conductivity-temperature-depth), dissolved gas-pressure probes, or by sampling the lake water at depth, followed by chemical analyses (Tassi and Rouwet 2014). Once a CO_2-accumulation is recognized and the CO_2 influx rate is known, it can be estimated when such lakes reach near-critical pressures of dissolved gases at depth

(Kusakabe 2015). The only way to mitigate a limnic gas burst is by artificially degassing the lake bottom waters, lowering pipes into the deep water layers, inducing degassing through the gas self-lift principle. Such pipes are efficiently degassing Lake Nyos since 2001, leading to safe gas contents in the near future (Kusakabe 2015). Many of the deep volcanic lakes world-wide remain unstudied, and it is thus unknown if these lakes are potentially hazardous or not.

Hydrothermal non-eruptive unrest

Gas emission

An increase in volcanic degassing can occur by (1) gas exsolution upon decompression, when a deeper magma rises towards the surface (Aiuppa et al. 2002, 2004), (2) dynamic magma convection inside the magmatic plumbing system, driven by the density difference between lower-density non-degassed and higher-density degassed magma (Kazahaya et al. 1994), or (3) crystallization of a stagnant, cooling magma batch and gas exsolution as the gas fraction in the magma reservoir increases with respect to the melt towards gas supersaturation (Oppenheimer 2011). The first process will generally be accompanied by seismic activity, deformation, variations in gas composition or even magmatic eruptions, and indicate a state of magmatic unrest, and is thus outside the scope of our review. The second process can be responsible for long-term degassing of large-volume magma bodies. Depending on the plumbing systems and magma depth, magma convection can be accompanied by signs of magmatic unrest (and eventually eruptions, e.g., Stromboli), or not (e.g., prolonged high-T fumarolic degassing). The third process is more complex in terms of time, space and its effect on surface manifestations. Magma crystallization can occur without any physical-chemical indicators measured at the surface, and the crystallization history will only be revealed when the magma is finally erupted. This type of degassing and related heat transfer, a necessary constraint to sustain a hydrothermal system, can explain prolonged periods of non-magmatic unrest, as there is no evidence of migration from a magma reservoir. Depending on the crystallization rate, depth and size of the magma batch, the period of "gas unrest" can cover entire inter-eruptive periods of volcanoes. Such unrest is often detected at passively degassing, closed-conduit volcanoes, often with long-lived active hydrothermal systems.

A few volcanoes are characterized by long-term, high-temperature (700-900°C) fumarolic degassing without further evidence of magma migration (e.g., Momotombo, Nicaragua, Menyailov et al. 1986; Satsuma-Iwojima, Japan, Shinohara et al. 1993, 2002; Kudryavy, Kuril Islands, Taran et al. 1995). Despite the fact that magmatic temperatures of the gases suggest the presence of a shallow magma body (hundreds of meters or less), no recent eruptions have occurred. The last major eruption at Satsuma-Iwojima occurred ~500 years ago, and high-T fumarolic degassing is reported for the past ~800 years (Shinohara et al. 2002). At Momotombo, the last magmatic eruption occurred in 1905 (lava flow, Menyailov et al. 1986). Slow crystallization and/or dynamic magma convection of a large stagnant magma body can explain long-term degassing. A major question remains: why do some high-temperature degassing volcanoes pass through decade-long phases of non-magmatic unrest, while other high-temperature degassing volcanoes frequently evolve into magmatic unrest, or eventually culminate into eruptive activity? The answer is probably found in the magma volume and magma recharge rate: large magma bodies with slow or absent magma refill will tend to degas without evolving towards eruptions.

The release of acidic gas species (SO_2, HCl, HF) during prolonged degassing in a phase of non-magmatic unrest can reduce the quality of human activities near volcanoes (e.g., agriculture, tourism) (van Manen 2014). The presence of SO_2 and HCl, clearly magmatic gases, does not necessarily imply the migration of a magma, as exsolution of both species can occur through the above processes (2) and (3). For open-conduit volcanoes, massive plume degassing can become very harmful (e.g., Masaya volcano, Nicaragua, Williams-Jones et al. 2003; Martin et al. 2010; Merapi, Java-Indonesia, Zimmer and Erzinger 2003), although in these cases magmatic unrest is ubiquitous. Besides the direct hazardous impact of acidic gas plumes, the absorption of acidic gases in humid air can cause acid rain and the formation of "dead zones" downwind volcano flanks (e.g., at Poás and Turrialba volcanoes, Costa Rica) (Rymer et al. 2009; van Manen 2014).

Acid contamination

The prolonged infiltration of acidic fluids into a volcanic edifice dissolves the host rock and can lead to (1) mechanical instability of volcano flanks, and thus a higher probability of flank failure (section Ground deformation), or (2) the dispersion of contaminants (e.g., heavy metals, fluorine, As, Hg, extreme acid waters) into the hydrologic network and regional aquifers around a volcano (Sriwana et al. 1998; Delmelle and Bernard 2000; Varekamp et al. 2001; van Rotterdam-Los et al. 2008; van Hinsberg et al. 2010). When such fluids are used for direct (drink water) or indirect (irrigation) human consumption, entering the food cycle, this long-term volcano-related process poses a health risk for the surrounding people (e.g., fluorisis, Löhr et al. 2005).

The most striking example of this situation is seepage from the hydrothermal system beneath the crater lake of Kawah Ijen volcano (Java, Indonesia), feeding the Banyupahit stream ("bitter river", Delmelle and Bernard 2000; van Hinsberg et al. 2010). Since the last magmatic eruption in 1817, Kawah Ijen has mainly been in a state of non-magmatic unrest, with the occurrence of only phreatic or geyser-like eruptions (Newhall and Dzurisin 1988). Nevertheless, the volcano hosts the largest reservoir of acidic surface water on earth, continuously fed by the input of magmatic gases and volatilized metals (Delmelle and Bernard 1994; Delmelle et al. 2000). Other examples, although with less hazardous impacts, are Río Agrio ("bitter river") at Copahue volcano (Argentina; Varekamp et al. 2001), Río Agrio at Poás volcano (Costa Rica; Rowe et al. 1995), and Ciwidey river at Patuha volcano (Java, Indonesia; Sriwana et al. 1998).

Ground deformation

Flank failure and sector collapse Acid fluids in volcanic-hydrothermal systems dissolve volcanic host rocks. Prolonged chemical leaching can finally result in physical rock removal, when such fluids exit at the volcano flank through acid saline thermal springs. Within the lifetime of hydrothermal systems, this rock mass removal can modify the morphology and weaken the mechanical stability of the volcanic edifice, increasing the probability of avalanches and sector collapses, even during periods of magmatic quiescence (Voight et al. 1983; López and Williams 1993; Kerle and van Wijck de Vries 2001; Reid 2004; Jolly et al. 2014; Fournier and Jolly 2014). Intense fumarolic activity at volcanic domes, not recently fed by a rising magma, can weaken the mechanical stability of the dome, eventually leading to collapse. Resulting pressure drop after unloading of the volcano flank or dome can lead to increased degassing or can trigger phreatic eruptions. The best physical evidence of acidic fluid dispersion and rock leaching within a volcanic edifice is given by seeping crater lakes (Rowe et al. 1995; Kempter and Rowe 2000; Varekamp et al. 2001; Taran et al. 2008; Delmelle et al. 2015).

Acid dispersion is an indication of eventual collapse-prone sectors of a volcanic edifice, but the presence of a less extreme hydrothermal system can already be sufficient to cause massive edifice collapse (López and Williams 1993; Reid 2004). Fluid-pressure evolution within a heterogeneous volcanic edifice is extremely complex, and depends on hydraulic (permeability and porosity) and thermal properties of the rock (Reid 2004; Fournier and Chardot 2012). To anticipate hazardous collapses intensive monitoring of pore-fluid pressure in the hydrothermal system and/or detailed deformation surveys (e.g., INSAR) of the edifice are key.

Ground deformation due to hydrothermal activity Long-term ground deformation often occurs at calderas without the occurrence of an eruption, or clear signals of magma migration (Rabaul, Papua New Guinea; Long Valley Caldera, Evans et al. 2002; Lewicki et al. 2007b; and "supervolcano" Yellowstone, USA, Dzurisin and Yamashita 1987; Werner and Brantley 2003; Lowenstern and Hurwitz 2008; Bergfeld et al. 2012; Chiodini et al. 2012; Lowenstern et al. 2014; Campi Flegrei, Italy, Amoruso et al. 2014). The discussion as to whether the unrest at Campi Flegrei since the 1980's is magmatic (Bianchi et al. 1987; Bonafede et al. 1986; Gottsmann et al. 2006), or non-magmatic (Casertano et al. 1976; Bonafede 1991; De Natale et al. 1991; Gaeta et al. 1998) is still ongoing. Recently, the tendency to explain the uplift by the expansion of the underlying hydrothermal system has become more plausible as it is supported by decade-long monitoring time series (geochemistry, geodesy, geophysical surveys) (Chiodini et al. 2001, 2003; Todesco et al. 2003), even though the deformation pattern may be explained by alternative models (e.g., Amoruso et al. 2014). The Campi Flegrei unrest, causing major concern also because the caldera is highly populated, is an example of prolonged hydrothermal non-eruptive unrest manifested as ground deformation paired with diffuse degassing.

The most convincing argument to explain the uplift phases at Campi Flegrei by hydrothermal circulation is an observed time-lapse of ~100 days of increased diffuse degassing following the uplift pulses (Chiodini et al. 2003). This implies that both processes have a common cause, which probably finds its origin in the input of a deep fluid, which may or may not be released by the underlying magma. The time-lapse between the "fast" deformation and "slow" degassing is due to the difference between the immediate elastic expansion of the highly altered rocks (De Natale et al. 1991), and the slower fluid rise (Chiodini et al. 2003). An additional argument in favour of the hydrothermal system being the cause of the observed bradyseismicity (ground deformation) is the thermal energy release by steam injection into the shallow system, an order of magnitude higher than the energy release by seismicity. Within the present view, as long as the Campi Flegrei caldera only deforms and degasses, without the occurrence of volcano-related seismicity (D'Auria et al. 2011), the current unrest is better defined as non-magmatic hydrothermal unrest. At Long Valley, following large seismic events and widespread passive CO_2 degassing there was debate as to whether the inflation was magmatic or hydrothermal in origin (Battaglia et al. 2003; Battaglia and Vasco 2006). A full range of methods and monitoring techniques (geophysics, seismology, geochemistry) is needed to better distinguish whether the cause of deformation is magmatic or non-magmatic.

Non-magmatic tectonic unrest

A volcano enters in a state of non-magmatic tectonic unrest when a seismic event (e.g. tectonic earthquake or seismic swarm) in the absence of any recognition of magma migration, causes concern. Despite the probable absence of indicators, other than the tectonic event itself, volcanoes in a delicate equilibrium (e.g., active hydrothermal systems) could enter a state of tectonic unrest because of the seismic event. Considering an already close-to-critical pressure state, the physical equilibrium beneath mineral caps or above chilling margins of magma bodies can be easily disturbed by major tectonic earthquakes, which can induce volcanic unrest (magmatic or not) or eventually trigger phreatic or phreatomagmatic eruptions (Brodsky et al. 1998; Ichihara and Brodsky 2006; Manga and Brodsky 2006).

Within the scope of the present study, the introduction of seismic events as a possible "eruption or unrest trigger" should be supported by (1) the probability of coincidence between an earthquake and an eruption, or an earthquake and an evolution from volcanic quiescence to unrest, and (2) the state of unrest of the volcano before the earthquake. Correlations between large tectonic earthquakes and major volcanic eruptions up to distances of 500–1000 km are found, with time lapses between the earthquake and eruption even up to 30–35 years (Marzocchi 2002; Marzocchi et al. 2002, 2004b). It is not excluded that lower magnitude earthquakes (especially if the epicentre is near the volcano), or major earthquakes at larger distances could possibly affect the state of unrest of a volcano (e.g., Walter and Amelung 2007). Detecting a change in the state of unrest caused by an earthquake of any kind is a lot more subtle than it is for extensively reported volcanic eruptions (e.g., increased heat flux, Delle Donne et al. 2010). A major problem in calculating the probability that the state of unrest is changed by the earthquake is not only to know how many times a volcano changed its state of unrest due to the earthquake, but also to know how many times the many other volcanoes did not change their state of unrest, despite the earthquake. For this reason, the pre-tectonic unrest state of the volcano should be known for as many volcanoes as possible, as seismic events are possible at any moment and place in a subduction tectonic setting.

Coinciding seismic and eruptive activity has been reported for hydrothermal systems and volcanoes in unrest (Healy et al. 1965; Hurst and McGinty 1999; Christenson 2000; Christenson et al. 2007, 2010; Ohba et al. 2008; Watt et al. 2009). The time lapse between the tectonic earthquake and the eventual variation in the behaviour of the volcano causing concern, can be short (hours to days) or long (weeks to months, to even several years, Marzocchi 2002; Eggert and Walter 2009). A short time lapse between tectonic earthquakes and volcanic events is observed for the November 1964 and 2006 Raoul crater lake breaching eruptions (7 days after, Healy et al. 1965; Christenson et al. 2007), for post-earthquake deformation and increased seismicity at Long Valley Caldera (within 5 days after, Linde et al. 1994), for increased heat flow from volcanoes worldwide after major tectonic earthquakes (>Mw 7.9) (3–5 days after, Delle Donne et al. 2010), and for a phreatic eruption at Laguna Caliente crater lake, Poás volcano (Costa Rica), four days after a Mw 6.2 earthquake at 6 km from the volcano in January 2009.

Not surprisingly, the dynamic stress created by the seismic surface wave is a major cause in disrupting the delicate equilibrium in magmatic-hydrothermal systems (Hill 2008; Delle Donne et al. 2010) switching from hydrothermal to tectonic unrest.

Event tree for non-magmatic unrest

Figure 6 proposes the structure of the non-magmatic unrest branch of the event tree that, for now, serves as a classification system for non-magmatic unrest. The branch first discriminates between a hydrothermal and tectonic state of unrest.

Many volcanoes experience a hydrothermal unrest phase that lasts for years, decades or even centuries, which can eventually evolve into eruptive or non-eruptive activity and related hazardous outcomes. This long-term constant behavior often makes it difficult to recognize how hydrothermal unrest can lead to related hazards in the short-term. Hydrothermal unrest can lead to non-magmatic eruptions, which can be explosive or effusive. Where the driving agent in magmatic eruptive unrest is magma, water (liquid or vapor, and occasionally liquid sulfur or gas) is the driving fluid and main eruptive product during non-magmatic eruptive unrest. This water is part of the volcanic edifice or its subsurface parts (i.e., hydrothermal aquifers, springs, rivers or crater lakes), and its expulsion can be effusive or explosive (phreatic, "surprise" phreatomagmatic, geyser-like water explosions and Nyos-type limnic gas burst). On the other hand, non-eruptive hydrothermal unrest can also lead to volcanic hazards after prolonged gas emission, acidic fluid infiltration into aquifers, soils and the hydrologic network, or deformation induced by a rising fluid front (i.e., bradyseismicity).

A state of non-magmatic unrest can be also characterized by the occurrence of earthquakes, unrelated to magma migration. Non-magmatic tectonic unrest can escalate into non-magma related seismic activity, or earthquake triggered tsunamis at subaqueous volcano flanks.

In the next sections, we follow that structure of the proposed event tree (Figure 6) to enumerate all potential hazards that such phenomena may cause in volcanic areas. In doing so we try to organize the potential hazards related to non-magmatic phenomena in order to provide the basis for a probabilistic quantification of the related hazards.

Major hazards related to non-magmatic unrest

Within the classification purpose of this review, the hazardous event can (1) only result from the previous causative

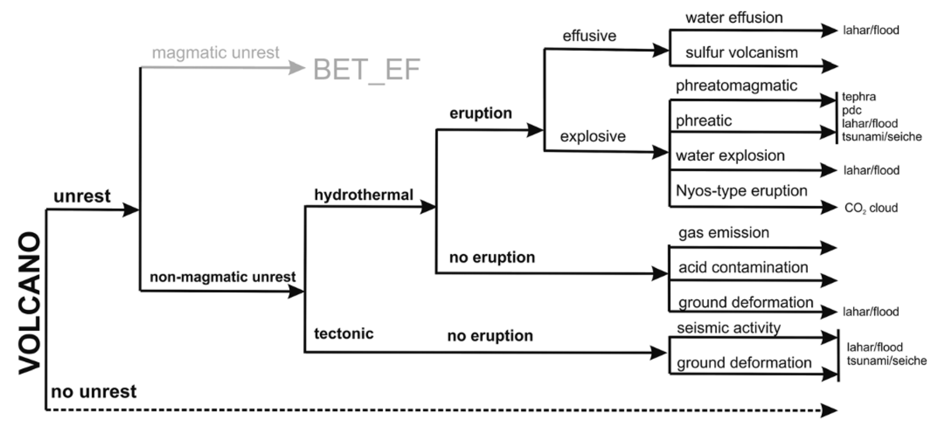

Figure 6 Proposed event tree for non-magmatic unrest. An event tree for magmatic unrest (BET_EF; in grey) is dealt with in previous studies (Marzocchi et al. 2004a, 2008; Sandri et al. 2009, 2012).

conditions (followed track) imposed by the event tree structure, and (2) not occur contemporaneously with another hazardous event, resulting from a different track. As the aim is to include all possible tracks that possibly lead to a hazard, all branches of the event tree have to exist and be possible at all time. A key point is that we do not know which track will be followed by the volcano and each new signal of the volcano challenges the user of the event tree to update and possibly even change track towards a different hazard, when describing non-magmatic unrest.

From the event tree structure it is clear that mass removal events (e.g. lahars, flood, tsunamis, ground deformation, jökulhlaups and flank collapses) can have several causes, depending on the state of unrest the volcano passes through. This principle gets even more complicated if external water (e.g., torrential rain) disturbs the state of unrest of a volcano. Besides having implications for future probabilistic hazard assessment, the potential for external water to influence the system highlights the need to a larger variety of monitoring methods, both internal (e.g., to better track hydrothermal activity) and external to the volcano (e.g., to track the weather). We now scan the event tree from right to left (Figure 6), to recognize indicators of non-magmatic unrest as we should aspire to track in monitoring networks.

Pyroclastic density currents and tephra fall out

Phreatomagmatic and phreatic eruptions may result in pyroclastic density currents (PDC) and tephra fall out. The 1963 Surtsey eruption off the coast of Iceland is the first well-documented phreatomagmatic eruption (Thorarinsson 1967) to show eruption dynamics. Phreatomagmatic and phreatic eruptions are short-lived explosions accompanied by an upward rush of black tephra (Kokelaar and Durant 1983) jets spread into cockscomb or cypress tree shapes (Kokelaar 1983) (Figure 5). Water-dominated hydrothermal systems (e.g., crater lakes, submarine settings) favor the generation of Surtseyan eruptions (Rouwet and Morrissey 2014a,b). During the phreatomagmatic eruptions at Lake Voui on Ambae Island in 2005, a tuff cone was constructed from material deposited from subaerial tephra jets leading into subaqueous PDCs after column collapse (Németh et al. 2006). Base surges are also common features that accompany tephra jets (Belousov and Belousova 2001; Németh et al. 2006). Phreatic eruptions appear extremely similar in morphology and dynamics (explosivity, Figure 5) but lack the injection of magma. The main distinction is the absence of juvenile material in the erupted products.

Lahars, volcanic debris flows, floods and jökulhlaups

A lahar consists of high-concentration sediment-charged flows that occur at volcanoes (Scott 1988; Manville et al. 2009; Pistolesi et al. 2014). Lahars are generated when three requisites are met: (1) a trigger mechanism that provides a sudden availability of sufficient water, (2) the presence of abundant loose volcanic debris along the flow path, and (3) steep slopes to increase gravitational flow (Vallance 2000; Pistolesi et al. 2014). Lahars can be primary or secondary. The former are instigated by eruptive activity (e.g., Nevado del Ruíz 1985, Pierson et al. 1990), and thus within a stage of magmatic unrest. The latter result from post-eruptive mobilization of unconsolidated volcanic debris (e.g., de Bélizal et al. 2013), sometimes even originating from a neighboring erupting volcano. The indefinite time between the eruption and the secondary lahar implies that the volcano can have re-entered into a state of non-magmatic unrest or even non-unrest, or that the occurrence of the lahar itself is independent of the actual state of unrest of the volcano.

Any process that suddenly liberates large amounts of water other than a magmatic eruption can trigger a lahar within a state of non-magmatic unrest. Rainfall-triggered lahars are arguably the most hazardous secondary events at volcanoes (e.g., Volcán de Fuego de Colima, Mexico, Capra et al. 2010, post-1991 Pinatubo eruption, Philippines, Pierson et al. 1992; Rodolfo et al. 1996; post-1965 Irazú eruptions, Costa Rica, Pavanelli 2006). Another frequent hazardous scenario to generate lahars are sudden expulsions of crater lake water after phreatic eruptions or crater failure (Mastin and Witter 2000; e.g., 1919, 1966, 1990 Kelud eruptions, Indonesia, Neumann van Padang 1960, Zen and Hadikusumo 1965; Thouret et al. 1998; 1965 Taal eruption, Philippines, Moore et al. 1966; Ruapehu 2007, Kilgour et al. 2010).

Increased heat transfer from a hydrothermal system into a glaciated volcanic edifice can lead to sudden ice melt. In general, any snow-capped volcano is lahar-prone even during stages of non-magmatic unrest, when sudden snow melt is induced (e.g., Cotopaxi, Ecuador, Hall et al. 2004; Mothes et al. 2004, Pistolesi et al. 2013, 2014). For snow-covered summits of crater lake bearing volcanoes entrainment of ice-slurry along the lahar flow path can also provide an additional water source (e.g., 1953, 1995 and 2007 Ruapehu eruptions, New Zealand; Nairn et al. 1979; Blong 1984; Cronin et al. 1997; Kilgour et al. 2010). A less hazardous scenario may include a newborn crater lake which fills a pre-existing previously snow-covered summit crater, with potential to eventually breach or overflow if melt water input continues (e.g., Chiginagak-2005, Alaska, Schaefer et al. 2008) (Figure 7). A more hazardous case is a subglacial lake with sudden and catastrophic melt water release which results in a jökulhlaup. Jökulhlaup is an Icelandic term and refers to a subglacial outburst flood (Björnsson 2002). The remaining ice mass can be entrained to add bulk mass to the flood, besides the entrainment of rock mass. Jökulhlaups can result from seasonal ice melting, heat input (non-magmatic unrest) or subglacial volcanic eruptions (magmatic unrest).

Figure 7 Picture of the natural "snow dam" blocking the newly formed Chiginagak crater lake, Alaska. Lake water drained at the base of the snow mass to generate a lahar/jökulhlaup downstream in early 2005 (Schaefer et al. 2008) (Picture by Game McGimsey, Alaska Volcano Observatory, US Geological Survey).

Jökulhlaups can probably be forecast by tracking volume decrease in ice mass, or increased temperatures.

Tsunamis and seiches

A tsunami occurs when a large body of water is suddenly displaced from its equilibrium position, generating long waves which propagate with a low energy loss from deep to shallow waters, where they rapidly decrease in velocity and reach high amplitudes in coastal areas. Volcano-related tsunamis can be associated with a variety of volcanic activities, such as submarine explosions in shallow waters, dense pyroclastic flows entering in the water and submarine mass movements. Within the context of the present study, these volcanic activities are often associated with magmatic unrest. Non-magmatic unrest phenomena are often slow and continuous, suggesting that non-magmatic unrest manifestations (apart from phreatomagmatic and phreatic eruptions) are not advantageous for tsunami triggering, however, tsunamis can also be generated as a secondary effect of non-magmatic outcomes (e.g., mass failures and PDC). In general, tsunamis due to volcanic activity remain a poorly investigated field, with only a few recent studies (e.g., Maeno and Imamura 2007, 2011; Paris et al. 2014).

Seiches are standing waves mostly due to meteorological effects (e.g., atmospheric pressure variations), but also due to earthquakes or tsunamis. They may occur in closed or partially closed basins, as such, lake seiches are most common. Seiches form as waves move back and forth hitting lake basin walls (Ichinose et al. 2000). When a phreatic eruption disrupts a crater lake basin, waves concentrically move outwards from the centre of initial water mass displacement (Figure 5). The wave height of the seiches at the lake shore depends on the violence of the trigger, lake water volume, and basin morphology. Similarly, seiches have also been observed also in open sea bays, harbors and gulfs, where natural resonant oscillations are permitted by the basin geometry. Tsunami and seiches related to volcanic activity are very rare events, nonetheless their impact could be extremely high, depending on specific conditions of the event (e.g., dominant wavelength, period).

Conclusions

In this review paper we have attempted to categorize non-magmatic unrest, and have sought to describe indicators how to recognize non-magmatic unrest. With the aim to classify non-magmatic unrest features, we presented an event tree structure with a progressive level of detail along various tracks towards hazardous outcomes. This implies that, beyond the current classification, the non-magmatic unrest branch in the event tree could become the base for future probabilistic hazard assessment at any type of volcano, regardless of its state of unrest.

The role of gas and water (liquid or vapor) instead of magma as the driving agent for unrest is stressed. Many volcanoes experience a prolonged stage of hydrothermal unrest, which can alter the hydraulic and rock mechanical properties leading to destabilization of a volcanic edifice. Recognizing an increase in vapor pressure in hydrothermal systems may provide warning of phreatic eruptions. A sudden release of water from an aquifer or lake, or intense rainfall, can trigger lahars when these mobilize poorly consolidated volcanic deposits or altered sections of a volcano.

Hydrothermal unrest is the most obvious expression of non-magmatic unrest, which could eventually lead to hazardous outcomes. Tracking temporal variations in fluid migration, hydraulic pressure regimes, as well as variations in temperature and chemical compositions of fluids (gas and water) could help to forecast hazardous outcomes. This means that traditional monitoring schemes (seismic-geodetic-geochemical) should be expanded to include other monitoring methods that might reveal precursory signals of non-magmatic volcanic hazards.

Moreover, we do not know if many of the ~1300 poorly studied Holocene active volcanoes are in a state of non-magmatic unrest, and thus, as stressed by this study, could be potentially hazardous. Knowing whether a volcano is dormant, in a state of quiescence, or in a state of non-magmatic unrest is a first requisite for hazard forecasting. This basic principle could guide future monitoring strategies for those volcanoes that are potentially more hazardous than currently thought. We invite the scientific community to define and track the background behavior of many poorly studied, but potentially hazardous volcanoes in order to recognize, in a timely manner, a state that raises concern.

Competing interests

The authors declare that they have no competing interests.

Authors' contributions

DR guided the review process of the research, and is the main author of the manuscript. LS participated in the scientific discussions and drafted the manuscript. WM participated in the scientific discussions and drafted the manuscript. JG participated in the scientific discussions and drafted the manuscript. JS participated in the scientific discussions and drafted the manuscript. RT participated in the scientific discussions and drafted the manuscript. PP participated in the scientific discussions. All authors read and approved the final manuscript.

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