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Volcanic precursors in light of eruption mechanisms at Vesuvius

Roberto Scandone*, Lisetta Giacomelli

Università di Roma Tre, Dipartimento di Matematica e Fisica, Rome, Italy

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ABSTRACT

Vesuvius entered a quiescent stage after the eruption of March-April 1944. The eruption was not much different or larger than other before, like for example the one of 1906, but it occurred at the end of a long period during which it was observed a decreasing trend of explosivity of eruptions [Scandone et al. 2008]. The parallel increase in the frequency of slow effusive eruptions, with respect to violent strombolian eruptions, point out to a process of average slower rate of magma ascent possibly due to a progressive sealing of the ascent path of magma to the surface. The small caldera collapse following the 1944 explosive phase effectively sealed the upper conduit, and since then the volcano entered a quiescence stage that was unusual with respect to the pattern of activity of the previous 300 years when the maximum repose time had been of 7 years (after the eruption of 1906). Most of the uncertainty on the duration of the present stage and character of a future renewal of activity derives by the basic questions regarding the nature of the current repose: due to a diminished supply of magma, related with structural condition or a sealing of the upper ascent path to the surface? Possibly the variation of structural conditions caused average slower ascent rates of magma favoring its cooling in the upper part of the crust, and effectively sealing the ascent path.

Introduction

Mount Vesuvius is presently in a quiescent stage that lasts since the eruption occurred 68 years ago, in March 1944. The current repose is the longest since 1631 when the volcano ended a centuries long quiescence, even if dubious reports of activity refer to an eruption in 1500 [Alfano and Friedlander 1929] or some kind of activity or unrest in 1571 [Guidoboni and Boschi 2006]. The majority of sources suggest that the repose may have begun as back as 1347 or 1139. During the period 1631-1944, the longest repose period (7 years) had occurred after the eruption of 1906. Significantly, the longest reposes occurred immediately after Violent Strombolian Eruptions as defined by Macdonald [1972] and classified by Walker [1973], which, accordingly, were termed "final" by Alfano and Friedlander [1929] and Carta et al. [1981] (Figure 1). The peculiarities of such eruptions were the rapid emission of a fast-moving lava flow, which eventually attained the nearby villages, a following explosive phase with high incandescent lava jets and fountains transitional to a sustained buoyant column, and the collapse of the summit cone with the formation of a small caldera of hundreds of meters depth [Scandone et al. 2008] (Figure 2). This occurrence was noted several times, and the formation of a small caldera was correctly interpreted already by Breislak and Winspeare [1794] during the lateral euption of 1794. They explained the decapitation of the summit of the volcano, following the explosive phase of the 1794 eruption, as caused by a collapse due to the fast emission of lava flows because in the proximity of the cone there were not enough detrital deposits justifying an erosion of the crater.

Such an explanation put forward much of the debate on caldera formation of the following two centuries. It is worth mentioning that during the lateral eruptions of 1760, 1794, and 1861, the explosive phases occurred at the summit cone, and were followed by the caldera collapse thus suggesting that, even in these cases, the eruptions were fed by a central feeder dyke. We emphasize that the caldera collapse is triggered by the rapid drainage of magma residing at very shallow depth (<2 km) [Marianelli et al. 2005] under the cone of Vesuvius. The magma resided at shallow depth for period ranging 0-10 years [Morgan et al. 2006], before being erupted either with slow effusion or during a violent strombolian event. The sequence of the violent strombolian eruptions are driven by the arrival of fast ascending, and gas-rich magma batches that act as a piston on the shallow reservoir, giving rise first to the fast lava flow, and the following explosive activity [Morgan et al. 2006, Scandone et al. 2008]. As mentioned before, after the most violent such episodes, the volcano entered a complete quiescence, eventually lasting for several years, and the renewal of activity, usually occurred with mild strombolian activity at the bot-

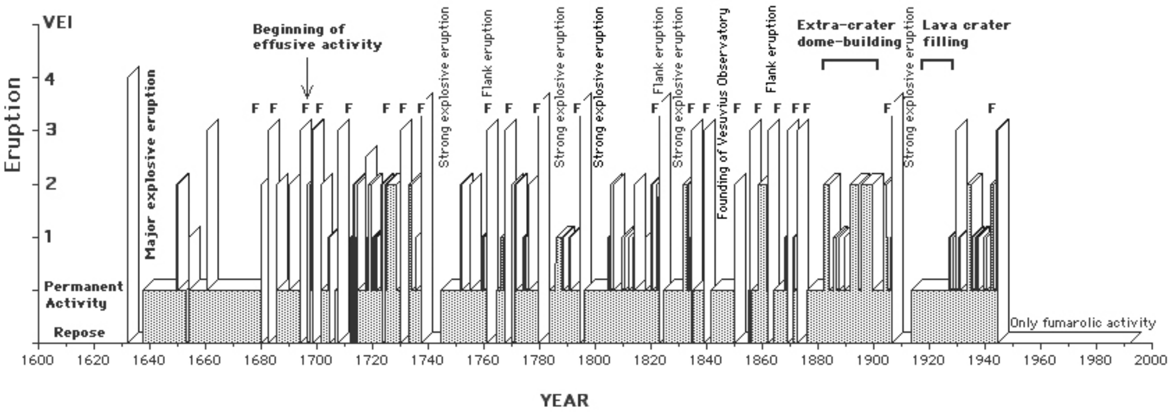

Figure 1. Scheme of the eruptive activity between 1631 and 1944. The horizontal bars indicate periods of permanent activity. The height of the vertical bars indicate the VEI of the eruptions; F indicates the major Violent Strombolian Eruptions eventually causing a small collapse of the "crater" and a subsequent short quiescence (modified after Scandone et al. [1993]).

tom of the caldera and slow lava effusion, that eventually filled the caldera. These mild renewals of activity were possibly due to an incomplete sealing of the ascent pathway following the "crater" collapses after the Violent Strombolian Eruptions.

The volcano structure and the current repose

The stratigraphy below Vesuvius is characterized by a sharp density transition at 2 km b.s.l. from limestone (with a density of 2500-2700 kg/m^2) to loose volcanic, alluvial and marine sediments (with an average density of 2300 kg/m^2) [Berrino et al. 1998, Bruno et al. 1998]. The sedimentary fill of the Campanian plain represents a low-density barrier to the ascent of dense magma and there are strong evidences that shallow magma accumulation below this barrier provides the source for medium-sized reservoirs, which fed the plinian and sub-plinian eruptions [Delibrias et al. 1979, Barberi and Leoni 1980, Belkin and De Vivo 1993]. Hydrothermal circulation, mineral deposition and repeated intrusions of dykes and their rapid cooling tend to reduce the density barrier and build up a structure favourable to magma ascent. Bouguer gravimetric anomalies of the volcano, nowadys, provide evidence of a shallow denser structure without very deep roots [Cassano and La Torre 1987]. A prominent high-density core has been also identified by seismic tomography [Zollo et al. 1996] concentric with the caldera structure. Seismic tomography evidenced also an extended low velocity layer at about 8-10 km depth, interpreted as the top of a magma reservoir, having a surface area of at least 400 km^2 [Auger et al. 2001].

Gravity inversion [Berrino and Camacho 2008] suggest the the upper part of Vesuvius is characterized by an high density body (density = 2450 Kg/m^3) similar to that of cooled lava. Such body has also a strong magnetization [Fedi et al. 1998], thus confirming the idea of a cooled magma intrusion.

A shallow reservoir residing near the surface fed the persistent activity between 1631 and 1944 [Marianelli et al. 2005]. During this period, violent strombolian eruption occasionally caused small caldera collapses, and a following short repose. The caldera, formed after the eruption of 1944, was smaller than that formed after the eruption of 1906 (25×10^6 m^3 *versus* 80×10^6 m^3, reduced to 50×10^6 m^3 after the landslides occurred before the renewal of activity in 1913). Also the erupted volume was smaller than that of 1906 [Imbò 1949], even if recent estimates suggest a much larger estimate [Cole and Scarpati 2009]. There is however a strong suggestion that in both cases, the shallow reservoir, that had fed the strombolian activity in the previous period, was almost completely evacuated [Scandone et al. 2008]. The reservoir was located above 3 km depth from the summit (between 10 and 100 Mpa pressure) [Fulignati et al. 2004] that means immediately above the limestone basement that floor the sedimentary cover of the campanian plain. It is tempting to suggest that the caldera collapses cause a density decrease in the central part of the volcano with respect to the surrounding rocks, thus preventing the buoyancy ascent of degassed magmas until the condition for magma ascent are re-established by repeated intrusions and the deposition of hydrothermal minerals.

After the 1944 eruption, signs of possible renewal of activity were observed in two occasions. The first significant anomalies were observed from 1952 to 1954 with an increase in the temperature of fumaroles within the crater from 350 to 600 °C [Imbò et al. 1964a] and on May 11, 1964, an earthquake swarm with a widely felt shock (I_{max} = 5-6 Mercalli Scale) and the formation of a funnel depression within the crater. [Imbò et al. 1964b] Since then, the temperature of the fumaroles has decreased [Nazzaro 1997], but occasional earthquake swarms have been observed in 1978-80, when there was a swarm of moderate earthquakes (M<3). The most significant earthquake since 1964 occurred in 1999 with $M_d = 3.6$, accompanied by a relatively small number of earthquakes with $M_d>1.8$.

Caliro et al. [2011] detected a significant variation in the chemical composition of fumaroles over the 1999-2002 period, which accompanied the seismic crisis of autumn 1999, with a continuous increase in the relative concentrations of CO_2 and He and a general decrease in the CH_4 concentrations which were interpreted as the consequence of an increment in the relative amount of magmatic fluids in the hydrothermal system.

Since 1985-86, there has been a progressive decrease of the b-value of the earthquakes of the Gutenberg-Richter law, an alternating sequence of seismic quiescence and activity, and an increase in the observed earthquake magnitude [De Natale et al. 2004]. The source mechanisms of the major earthquakes are compatible with isotropic components that indicate volumetric expansion [De Natale et al. 2004]. The average depth of the earthquakes is comprised between 2 and 6 kilometers [De Natale et al. 1998, Lomax et al. 2001, De Natale et al. 2004]. The recurrence rate of seismic swarms and other unrest phenomena in the past 50 years is = 0.14 y^{-1}, similar to the average rate of paroxistic eruptions evaluated until 1944 (0.13 y^{-1}) [Scandone et al. 2008].

Scandone et al. [2008] suggested that the violent strombolian eruptions were caused by fast ascending magma batches, which unplugged the upper conduit of the cooler and crystal-rich magma, and caused the following collapse of detrital material into the evacuated conduit. The number of violent strombolian eruption decreased with time and gradually made this process less effective. The volcano finally entered a quiescent stage because the slow ascending magma was not able to reach the surface and crystallized extensively before the arrival of the next energetic magma batch (1964). The crystallized magma and the mineral deposition caused by the hydrothermal circulation in the upper part of the volcano formed an extensive barrier [Bons et al. 2001] which further grew because of subsequent intrusions (e.g. 1999-2002).

Collectively these data indicate the lack of any relevant shallow magma chamber in the first two kilometers depth, like the ones feeding the persistent activity in the period between 1631 and 1944, and at the same times suggest the existence of a rigid structure with density higher than the surrounding rocks, that represent a physical, unfractured barrier to the ascent of magma.

We suggest that below such structure, magma has continued to intrude, occasionally with batches rising with higher velocity thus causing episodic seismic swarms. The eruptable magma depends on the ratio between the intrusion rate and the cooling rate which varies with depth and the surrounding temperature.

Figure 2. Oblique view of the Vesuvius crater (Gran Cono) formed during the eruption of 1944. The steep walls of the crater suggest that it was formed by a collapse inside the evacuated portion of the shallow magma reservoir (<2 km) feeding the permanent activity between 1913 and 1944. On the background, the border of the Somma caldera, formed as a result of the collapses following the main plinian eruptions of the last 17,000 years. (Photo Scandone).

Precursory signals for an eruption at Vesuvius

Eruptions are commonly preceded by a number of precursors signalling an "anomalous" state of the volcano. Precursors are generally observed in the last stages of processes that lead to an eruption when magma finally approaches the surface. According to Dsurisin [2003], the absence of measurable precursors until magma begins its final ascent, is mainly a consequence of:

(1) the brittle-ductile transition relatively shallow beneath most volcanoes;

(2) the lack of monitoring in the types and amounts of volcanic gases emitted over time;

(3) the most classical geodetic techniques are not sensitive enough to detect subtle ground deformation that might occur when magma accumulates near the brittle-ductile transition, especially if the intrusion occurs gradually or episodically in a series of small events.

In particular, it is highly probable that the processes leading to the occurrence of an eruption of a quiescent volcano, may occur on time scales of the order of hundred of years, as has been proved, for example, for the Monte Nuovo eruption of Campi Flegrei in 1538, where the early signs of unrest dated back more than a century [Giacomelli and Scandone 2012]. It is otherways true, that we are more capable of detecting signs of unrest in calderas than in central volcanoes.

Based on this assumptions it is likely that we are: either uncapable of detecting slow magma accumulation in a crustal reservoir, or we are overlooking the weak signals that the volcano may display.

As mentioned in the previous paragraph, the occasional seismic swarms observed at Vesuvius in the last 50 years may represent a weak signal of magma accumulation. Within this frame, the renewal of volcanic activtiy may only be signalled by the fracturing of the rocks above an inferred magma accumulation zone 3-5 km deep, similar to that which fed the 1631 eruption after the last prolongued period of repose [Scaillet et al. 2008].

In this respect it is unlikely that we may observe significant changes of ground deformation or seismicity until the process is well underway and unlikely to stop (like for example in Campi Flegrei).

In some respect, the start of precursory earthquakes may be not much different from a "normal" seismic crisis like that observed, for example in 1999, with the main difference that it may continue with a swarm of magnitude 4 earthquakes.

Roman and Kashmann [2006] suggests that two main processes dominate the seismic swarms preceding an eruption:

– failure within the walls of inflating dikes with a local 90° stress field rotation with respect to to the axis of maximum compressive stress,

– failure at the dike tip with hypocenter propagation, although in some cases dike propagation may be effectively aseismic and in other cases dike inflation may be effectively aseismic.

In the case of Vesuvius, Bianco et al. [1998] suggest that there are two main NNE–SSW and ESE–WNW-trending extension directions with the first one related with the regional stress field and the second, local, acting inside the Somma caldera.

Basing also on the last renewal of activity of Vesuvius in 1631, when the early phase of the eruption was fed by a lateral vent we may envisage a precursor scenario possibly into two stages:

1) Magma ascent filling a dike along the central axis of the volcano perpendicular to the local ESE–WNW-trending extension direction

2) Possible migration of hypocenters toward the SSW flank of the volcano with a possible opening of a lateral vent.

Within this frame, we may expect an early swarm of VT earthquakes with mostly strike-slip mechanism oriented along an ESE–WNW trending direction, followed by a possible migration of hypocenters toward the SSW flank of Vesuvius. The depth of the earthquakes should be in the uppermost 3 km of crust. The duration of such swarm may be as short as a few hours, as the one occurred in the night of December 16, 1631, and with an high seismic destructive potential, expecially in the environment of Torre del Greco as this city lies along one of the main fault systems feeding the activity of Vesuvius

A short duration of the swarm and lack of LP events are likely in the case of a sudden rupture of a shallow magma reservoir and rapid magma ascent. The lack of LP earthquakes may be due to an insufficient time to allow efficient degassing from the ascending magma batch [Scandone et al. 2007]. In such case it is likely an explosive eruption with kilometers high lava-fountain and the formation of a plinian eruption cloud. Rapid rupturing of a crustal reservoir and the lack of efficient degassing make unlikely early detection of extensive geochemical anomalies within fumaroles and water wells.

A longer duration of the swarm and the occurrence of LP earthquakes and possible lateral migration of hypocenters could be suggestive of very shallow residence of the ascending magma and of extensive degassing. This could possibly result in an early lateral effusive eruption, eventually followed by an explosive phase from the central vent, as soon as conditions for fast magma ascent are attained (e.g. 1631 or 1794 eruptions). Extensive degassing may allow detection of geochemical anomalies in fumaroles. Shallow intrusion may be also evidenced by deformation of the flank of the volcano. This picture however, does not ensure that the outbreak may be only effusive, as the opening of the ascent path may be operated by several ascending magma batches each one utilizing the path of the previous one and displacing or intruding the stationary early degassed magma (like in the case of Pinatubo eruption where the early degassed dome was rapidly diplaced by faster ascending batches) [Hattori and Sato 1996].

Conclusions

The analysis of the past pattern of activity, as well as recent unrest episodes at Vesuvius, suggest that the volcano is in a sort of dynamical repose where episodic input of new magma in the shallow crust (3-5 km) are not allowed to reach the surface because of a solidified plug. The lack of evidence of an extensive magma chamber in the shallow crust, suggests that most of the new magma arrivals cool rapidly. Relevant accumulation of enough hot magma, capable of eruption may eventually occur with an increase of the episodic supply rate. Such occurrence should be eventually detected by a higher frequency of seismic crises and possible increase of magmatic gas emission.

We do have a relative good record of the phenomena preceding the last renewal of activity of 1631, after a long repose period [Alfano and Friedlander 1929, Bertagnini et al. 2006, Guidoboni 2008, Scandone and Giacomelli 2008, Guidoboni and Mariotti 2011].

The suggestion that a two-step scenario occurred at Vesuvius, in 1631, is suggested by the sequence of observed precursors, which records a months-week long, succession of "low intensity precursors", not raising particular alarm, and then a climax of seismic activity in the day before the outbreak of the eruption [see for example: Braccini 1632, Guidoboni and Mariotti 2011].

The idea that high intensity, short-term precursors may be followed by an explosive eruption is based on the suggestion that short ascent times are required to prevent extensive degassing from the ascending magma batches [Scandone et al. 2007]. This requirement may be obtained also late, in the development of the eruption process, if precursory magma batches prepare the ascent path for later fast ascent and eventually full connection between the magma chamber and the surface. In this case, it is likely that a sustained plinian eruption may develop like the one of 1631 or even 79 A.D., if a large enough magma chamber formed at depth below the volcano and became connected to the surface. Such case has been observed for the eruption Pinatubo, in 1991. In that case, a slow ascending magma produced months-long precursors to the extrusion of a dome. The following explosive climax was preceded by a sequence of of subplinian eruptions, which started three days before the cataclysmic event.

The short term precursors to the climactic phase, as often occurs in many explosive eruptions, generally causes a self-evacuation of the area, and reduces the impact of the eruption. A self-evacuation of the area occurred during the 1631 eruption, but the relative, high number of casualties was due to the prohibition to the evacuees to enter Naples because the fear of plague [Giuliani 1632].

We are conscious of the multiple possibilities that may develop when a volcano come backs into activity, so that, risk scenarios cannot be based only on a single case, but must take into account also an unlikely succession of events. We suggest that the population should be aware of this potential, and be made capable of reacting to unforeseen events.

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*Corresponding author: Roberto Scandone,
Università di Roma Tre, Dipartimento di Matematica e Fisica,
Rome, Italy; email: scandone@uniroma3.it.