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RESEARCH ARTICLE

Volatile-rich magma injection into the feeding system during the 2001 eruption of Mt. Etna (Italy): its role on explosive activity and change in rheology of lavas

Carmelo Ferlito · Marco Viccaro · Renato Cristofolini

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Abstract The explosive behavior and the rheology of lavas in basaltic volcanoes, usually driven by differentiation, can also be significantly affected by the kinetics of magma degassing in the upper portion of the feeding system. The complex eruption of 2001 at Mt. Etna, Italy, was marked by two crucial phenomena that occurred at the Laghetto vent on the southern flank of the volcano: 1) intense explosive activity and 2) at the end of the eruption, emission of a lava flow with higher viscosity than flows previously emitted from the same vent. Here, we investigate the hypothesis that these events were driven by the injection of volatile-rich magma into the feeding system. The input and mixing of this magma into a reservoir containing more evolved magma had the twofold effect of increasing 1) the overall concentration of volatiles and 2) their exsolution with consequent efficient vesiculation and degassing. This led to an explosive stage of the eruption, which produced a ~75-m-high cinder cone. Efficient volatile loss and the consequent increase of the liquidus temperature brought about the nucleation of Fe-oxides and other anhydrous crystalline phases, which significantly increased the magma viscosity in the upper part of the conduit, leading to the emission of a high viscosity lava flow that ended the eruption. The 2001 eruption has offered the opportunity to investigate the important role that input of volatile-rich magma may exert in controlling not only the geochemical features of erupted lavas but also the eruption dynamics. These results present a new idea for interpreting similar eruptions in other basaltic volcanoes and explaining eruptions with uncommonly high explosivity when only basic magmas are involved.

Keywords Etna · Exsolution · Degassing · Viscosity · Mixing · Eruptive behavior

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C. Ferlito · M. Viccaro (✉) · R. Cristofolini
Dipartimento di Scienze Geologiche, Università di Catania,
Corso Italia 57,
I-95129 Catania, Italy
e-mail: m.viccaro@unict.it

C. Ferlito
e-mail: cferlito@unict.it

R. Cristofolini
e-mail: rcristof@unict.it

Introduction

Volcanic activity is a hazard for land, property and population. In many areas of the world mitigation of potentially dangerous eruptions is based on studies aimed at recognizing eruption precursors. Particular attention has been devoted to stratovolcanoes characterized by long periods of quiescence followed by sudden awakening and strongly explosive (Plinian or Ultra-Plinian) eruptions (e.g. Mount St. Helens, Pinatubo; cf. Schminke 2004 and references therein). Less attention has been addressed to persistently active basaltic volcanoes, such as Mt. Etna, which are characterized by prevailing Strombolian explosions and low rates of lava emission, and generally do not represent as serious a hazard. However, these volcanoes occasionally generate strongly explosive eruptions which can greatly increase the associated hazard. A recent example is the episode of April 5, 2003 at Stromboli volcano (Italy), where a 5-km-high eruptive column resulted from a sudden input of fresh and undegassed magma that mingled with more degassed magma already occupying the open conduit (Francalanci et al. 2005). Recently, Mt. Etna volcano experienced a comparable event, when during the complex eruption of July–August, 2001, eruptive activity was fed by two distinct batches of magma through different fracture systems (Behncke and Neri 2003; Clocchiatti et al. 2004; Métrich et al. 2004; Corsaro et al. 2007). One system, NNW-SSE-oriented, erupted products related to the open-conduit connected to summit craters, whereas the other, N-S-oriented, erupted lavas mildly evolved in a shallow closed reservoir (ABT in Ferlito et al. 2008; cf. also Clocchiatti et al. 2004; Métrich et al. 2004; Corsaro et al. 2007; Viccaro et al. 2007). Some authors working on the eruption have suggested that the two fracture systems intersected the surface in the Laghetto area (Fig. 1), doubling the extensional offset and allowing the ascent of a magma (LAG in Ferlito et al. 2008) resulting from mixing between the resident and more evolved magma (ABT) and a more basic and undegassed one (DBM in Ferlito et al. 2008) ascending through the NNW-SSE fractures (Monaco et al. 2005; Viccaro et al. 2006). From July 25 to 30, short and thin flows of lava, derived from mixing, were erupted and accompanied by intense Strombolian activity. On the evening of July 30, the emission of a high viscosity lava flow concluded the lava effusion at the Laghetto vent.

Focusing on the exceptional explosive activity and the abrupt transition from low- to high-viscosity lavas at the Laghetto vent, this work is aimed at clarifying the role played by the input of volatile-rich magma in modifying the intensity of the eruption and the rheology of lavas. Since the processes we are dealing with occurred within the subsurface feeding system, we consider seismological and geochemical data in order to constrain the parameters for mixing and degassing. Modeling of the quantity of gas able to exsolve from the ascending magma was performed using VOLATILECALC (Newman and Lowenstern 2002), and was constrained by data on melt inclusions in olivine (Métrich et al. 2004). Modeling of the phase assemblage and proportions of phenocryst minerals in the system has been performed using MELTS (Ghiorso and Sack 1995; Asimow and Ghiorso 1998). Physical-chemical parameters were constrained by geochemical and geophysical data from literature (Bonaccorso et al. 2002, Patanè et al. 2002; Monaco et al. 2005; Viccaro et al. 2006; 2007; Ferlito et al. 2008). The reliability of the modeling was constrained by comparing the results with volcanological observations. The results obtained have allowed us to quantify the role played by the degassing dynamics in increasing explosivity and modifying lava rheology.

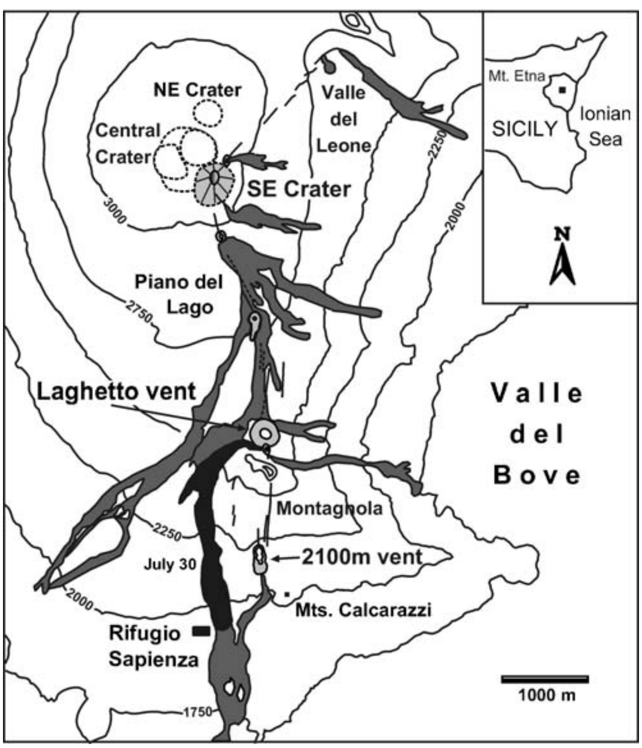

Fig. 1 Sketch map of lava flows and cinder cones of the 2001 eruption. Lava flow emitted by the Laghetto vent in the evening of July 30 is highlighted in black

Volcanological background

The eruptive event occurred between July 13 and August 9, 2001 on the upper southern slopes of Mt. Etna (Fig. 1). The eruption was preceded by shallow seismic swarms totaling ~2,600 earthquakes (Bonaccorso et al. 2002; Patanè et al. 2002; Monaco et al. 2005). On July 12, a seismic swarm preceded the opening of fissures at the surface between 3,100 and 2,100 m (a.s.l.). Hypocenters were distributed within a crustal volume elongated NNW-SSE at depths between 1.5 and 4 km (b.s.l.). On July 13, the first lava fountains occurred at the South-East (SE) crater, and emission of lava flows began from the base of the cone. From July 15 to 18 the number of deeper events gradually decreased with focal depths concentrated in the upper 2 km of the edifice (Monaco et al. 2005). In the early morning of July 18, a new N-S-oriented fissure developed near the Rifugio Sapienza at 2,100 m (a.s.l.). Eruption at this fissure was characterized by Strombolian activity and lava effusion at low emission rates (6 m^3/sec; Behncke and Neri 2003). On July 20, fracturing progressed northwards for about 1.5 km and reached the Laghetto area at 2,550 m. Here, the NNW-SSE and N-S fracture systems intersected, doubling the extensional offset which produced ground collapse and the formation of a pit-crater (Fig. 2; Monaco et al. 2005). At first the activity here was phreatomagmatic (Behncke and Neri 2003), but in a few days evolved to spectacular Strombolian activity, which produced a 75-m-high cinder cone (Calvari and Pinkerton 2004).

On July 25, at the Laghetto cone, the transition to a purely magmatic activity was marked by 500-m-high lava fountains accompanied by lava flow emission with rather low output rate (<10 m^3/sec) from the southern base of the

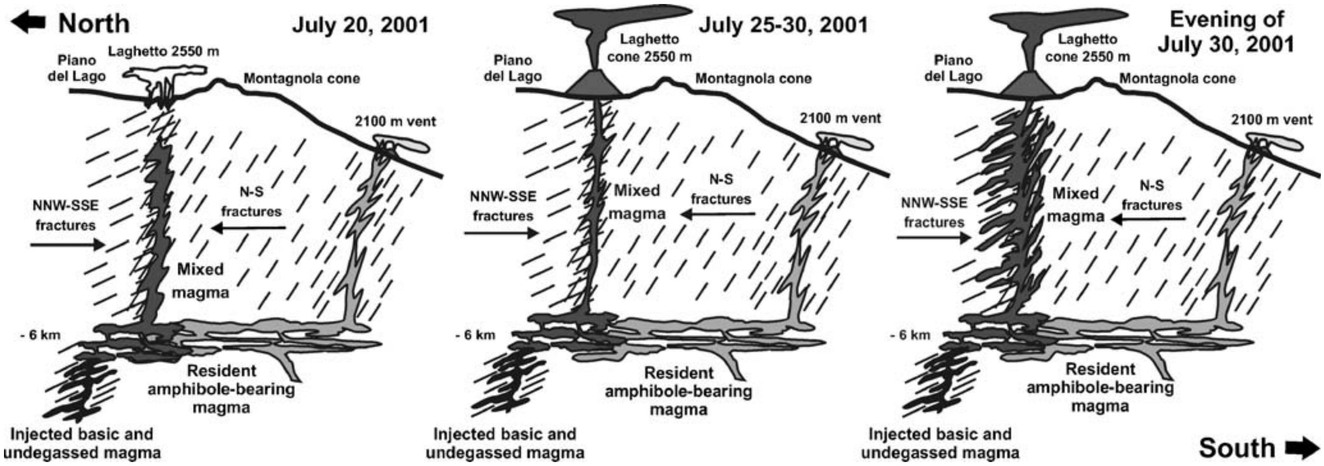

Fig. 2 Schematic cartoon representing the evolution of the 2001 eruption at Mt. Etna volcano. During the first days of eruption, until July 20, the activity took place at the 2,100 m vent and at the South-East crater (off figure to the north, see Fig. 1). On July 20, NNW-SSE and N-S fracturing met in the Laghetto area at ~2,550 m of altitude and here phreatomagmatic activity began, associated with the injection of magma resulting from mixing (*left panel*). The pure juvenile component of the mixed magma started to be emitted from the Laghetto crater on July 25, lasting until the afternoon of July 30, as thin lava flows and pulsing lava fountains up to 500-m-high (*central panel*). On the evening of July 30, an increase in the emission rate occurred, suggesting that the conduit was able to supply magma at higher rate than on the previous days (*right panel*). The extrusion of magma continued to be very violent at the Laghetto vent, but it was also associated with the emission of a thick and highly viscous lava flow that moved like a Bingham body

cone. Lava flows were directed east into the Valle del Bove, where they extended to a length of ~1.5 km, and west, where they extended for ~400 m (Fig. 2; Calvari and Pinkerton 2004; Ferlito and Siewert 2006; Siewert and Ferlito 2008). On July 30, at around 8.30 pm (Local Time), the lava output from the southern base of the Laghetto vent increased dramatically and gave rise to a large high viscosity flow that reached the Rifugio Sapienza area in less than 12 h (Fig. 2; Ferlito and Siewert 2006). This was the peak of the eruption as far as emission rate is concerned, and led to the end of lava outpouring at the Laghetto vent; explosive activity continued, decreasing slowly until August 9.

The 2001 eruption produced an estimated volume of about 25×10^6 m^3 of lava and 7×10^6 m^3 of tephra (Behncke and Neri 2003; Calvari and Pinkerton 2004; Lautze et al. 2004; Ferlito and Siewert 2006), with an Explosivity Index = 63, testifying that it can be regarded among the most explosive in the historical record of Mt. Etna (Fig. 3; cf. Romano and Sturiale 1982).

Petrochemical background

Petrography and geochemistry of lavas and tephra of the 2001 eruptive episode indicate that this event was characterized by the simultaneous emission of distinct magmas (Clocchiatti et al. 2004; Monaco et al. 2005; Métrich et al. 2004; Corsaro et al. 2007). In the upper part of the volcano (at the base of the SE crater and at the Piano del Lago area), lavas emitted from the NNW-SSE fracture system were trachybasalts with a Porphyritic Index ~30–40 (PI in vol%), characterized by abundant phenocrysts of plagioclase and subordinate augite, olivine, and titanomagnetite. Texture and mineral assemblage are similar to those of magmas associated to the open conduit system (Clocchiatti et al. 2004; Métrich et al. 2004; Corsaro et al. 2007; Viccaro et al. 2006; Viccaro and Cristofolini 2008). In the Calcarazzi (2,100 m a.s.l.) and Laghetto (2,550 m a.s.l.) areas, lavas emitted from the N-S fissures were amphibole-bearing trachybasalts (PI ~10–15) with phenocrysts, in order of decreasing abundance, of augite, plagioclase, titanomagnetite,

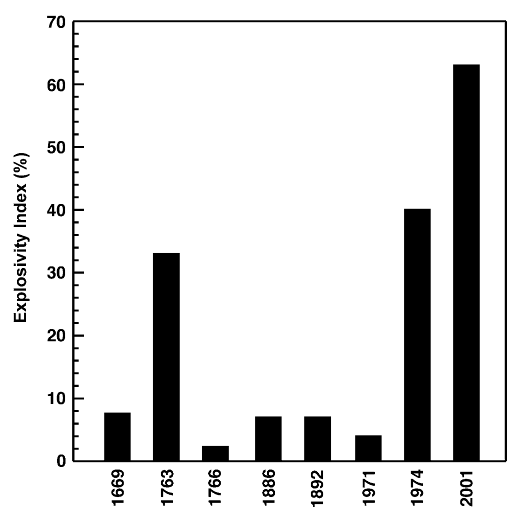

Fig. 3 Histogram showing some historical explosive eruptions vs. their Explosivity Index calculated as $EI = V_t/(V_1 + V_t)$, where V_t is the volume of tephra and V_1 that of lava flows (cf. Romano and Sturiale 1982). Tephra and lava volume for the 2001 eruption includes only tephra and lava erupted from the Laghetto vent

olivine and with megacrystic amphibole. Their texture and mineral assemblage are consistent with differentiation of a magma residing in a closed reservoir (Clocchiatti et al. 2004; Métrich et al. 2004; Corsaro et al. 2007; Viccaro et al. 2007; Viccaro and Cristofolini 2008). A thorough examination of the geochemical features of lavas and tephra erupted at the Laghetto vent, compared with products emitted at the Calcarazzi area, led to the hypothesis that a third magma, more primitive and volatile-rich than the resident magma, intruded the closed reservoir (Viccaro et al. 2006; Ferlito et al. 2008). This new magma mixed with the amphibole-bearing magma, as revealed by variations of the concentrations of some major and trace elements of minerals and whole rock (cf. Viccaro et al. 2006), but did not erupt independently. Mixing occurred within about 5 days, as suggested by the evolution of volcanic phenomena, and was preceded by an influx of volatiles which modified the composition of the melt fraction, by increasing its concentrations in TiO_2, FeO, P_2O_5 and particularly in K_2O and chlorine (Ferlito et al. 2008).

Volatile-rich magma injection and explosive activity

The injection of hot, primitive magma into reservoirs of more evolved magma can produce at depth two fundamental linked processes: 1) the exsolution of volatiles in the cooler resident magma, due to temperature rise (Sparks et al. 1977; Snyder 2000 and references therein); 2) nucleation and growth of anhydrous crystalline phases due to the lowered temperature in the hotter injected magma, which increases volatile concentration, possibly to saturation and exsolution (Blake 1984; Huppert et al. 1982a; Tait et al. 1989; Folch and Martì 1998).

In order to describe the above mentioned scenarios, the relative amounts of melt and crystals have been constrained with thermodynamic simulations performed with MELTS (release v5.0; Ghiorso and Sack 1995; Asimow and Ghiorso 1998), whereas conditions of volatile saturation and overpressure of magmas have been estimated by means of the VOLATILECALC software (Newman and Lowenstern 2002).

Compositions of the two end-members involved in the mixing process have been chosen from Ferlito et al. (2008) for MELTS simulations (Table 1). The stable assemblage and thermodynamic properties of the resident magma were calculated at the following conditions: 1) T=990°C, according to phase equilibria between Mg-hastingsite amphibole and other phases in the system (Pompilio and Rutherford 2002; Viccaro et al. 2006, 2007); 2) P=200 MPa, based on aforementioned phase equilibria and the hypocentral depths (6 km b.s.l.) of the earthquake swarm that preceded the eruption of this magma (Bonaccorso et al. 2002; Patanè

Table 1 Average chemical compositions (n = number of analyses) and physical conditions for end-members of the mixing process and the mixed magma used for thermodynamic simulations (data from Ferlito et al. 2008 and Viccaro and Cristofolini 2008)

	n	SiO_2	TiO_2	Al_2O_3	FeO_{tot}	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Fe_2O_3	FeO	T°C	P MPa	fO_2
Resident Magma	5	48.65	1.40	18.21	8.84	0.16	5.67	10.29	4.19	2.06	0.53	3.74	5.44	990	200	QFM
Injected Magma	10	47.06	1.80	17.41	10.59	0.18	5.87	11.19	3.42	2.05	0.43	5.28	5.84	1,100	200	QFM
Mixed Magma	12	47.66	1.60	17.58	10.07	0.18	5.93	11.04	3.53	1.95	0.46	4.86	5.65	1,050	200	QFM

See text for details on the choice of T, P and fO_2

Simulations run at H_2O=2.5 wt% and CO_2=500 ppm for the resident magma (cf. Armienti et al. 2002) and H_2O=3.4 wt% and CO_2=1,100 ppm for the injected one (cf. Métrich et al. 2004). H_2O and CO_2 concentrations for the mixed magma were approximated to 2.8 wt% and 680 ppm, respectively, considering the addition of ~30% by volume of the initial volatile content of the injected magma (volatile saturated at 200 MPa) to the resident one et al. 2002; Monaco et al. 2005); 3) fO_2 at QFM buffer, in agreement with the literature data for most Etnean magmas (cf. Pompilio et al. 1998). H_2O and CO_2 contents of 2.5 wt% and 500 ppm respectively, have been assumed for this magma at liquidus conditions, on the basis of values generally accepted for Etnean magmas of similar compositions (Armienti et al. 2002). Run simulations led to a stable assemblage composed of ~58% liquid, ~25% augite, ~10% plagioclase (An_{71}), ~4% magnetite, ~3% olivine (Fo_{72}) and traces of apatite. This assemblage matches fairly well the one in the actual products as described in the literature (Clocchiatti et al. 2004; Métrich et al. 2004; Viccaro et al. 2006, 2007; Ferlito et al. 2008). However, due to MELTS limits for hydrous phases, amphibole was not considered in the simulation.

Conditions were set for the basic undegassed magma as following: 1) T=1,100°C corresponding to a near-liquidus T for trachybasaltic compositions (as inferred in Ferlito et al. 2008) and close to T values deduced by Métrich et al. (2004) for H_2O- and CO_2-rich primitive melt inclusions in olivine erupted at the Laghetto vent; 2) P=200 MPa and 3) fO_2=QFM buffer, as in the resident magma. Volatile content for this magma was assumed to be 3.4 wt% H_2O and 1,100 ppm CO_2, the volatile concentrations of olivine-hosted melt inclusions found in this magma (cf. Métrich et al. 2004). Under these conditions the computed stable assemblage is composed of ~97% of liquid and ~3% of olivine and augite.

In order to assess how the system equilibrated after mixing, the process was simulated as an isenthalpic assimilation between the two magmas at thermodynamic equilibrium conditions. The possibility of exsolving volatiles from the resident magma by transferring heat from the hotter injected one was considered first. Heat will be transferred from the high-H to low-H system (where H is enthalpy) until the system attains the new thermal equilibrium. The simulation was run so that, at each step of mixing, temperature variation is computed to keep constant the total enthalpy of the system (cf. Ghiorso and Sack 1995; Asimow and Ghiorso 1998). The resulting temperature is strictly dependent on mixing proportions and heat capacity (C_p) of the two magmas. As suggested by Viccaro et al. (2006), on the basis of geochemical evidence, the proportion of basic undegassed magma in the mixing was ~30 vol%. The final T of the system after mixing should be ~1,050°C. This change in temperature, however, can cause volatile exsolution only if magma reaches gas saturation. Calculations performed with VOLATILECALC (Newman and Lowenstern 2002), adopting the same parameters of MELTS simulations, give a minimum pressure of saturation for a multi-component (H_2O+CO_2) gas phase at ~160 MPa for the resident magma and at ~250 MPa for the primitive undegassed magma. This suggests that at 200 MPa (the pressure at which mixing occurred; cf. Viccaro et al. 2006), the resident magma was volatile undersaturated. At a constant confining pressure of 200 MPa, the resident magma could reach gas saturation only by increasing its volatile concentration (from 2.5 to 3 wt% H_2O and from 500 to 800 ppm CO_2; Fig. 4a). The injected volatile-rich magma reached saturation at ~250 MPa and was therefore able to provide the necessary gas phase. However, even considering that the resident magma might have reached saturation through gas addition, the computed ΔT_{max} ~60°C due to mixing is not capable in further increasing the exsolution (Fig. 4a).

In the injected magma, mixing with the cooler resident magma induces temperature drop and crystallization, which enhances exsolution of volatiles and generates overpressure. Many authors (Folch and Martì 1998; Huppert et al. 1982a; Huppert et al. 1982b) have suggested that overpressure can be generated relatively quickly when the initial T contrast between the two magmas is high (ΔT of 200–300°C). In the present instance, the initial ΔT_{max} between the two magmas is only ~110°C. The simulated T decrease (around 50°C), due to mixing with proportions of 70% resident and 30% injected magma (see above), has produced the additional precipitation of ~10% of augite and Fe-oxide during thermal re-equilibration of the system. The effect on exsolution-driven volatile overpressure due to crystallization can be quantified by using the equation formulated by Tait et al. (1989, eq. 20) in the form:

$$1-(P_i/P)^n+(P/P_i-1)\times(K_1+K_3)=m_c/M \quad (1)$$

where P_i is the initial pressure at which crystallization occurs; P the overpressure generated by crystallization; n is a dimensionless exponent depending on the liquid composition; K_1 is a dimensionless number depending on the elastic deformation of the reservoir wall-rock, the bulk modulus of the liquid, and H_2O solubility; K_3 depends on the mass fraction of gas exsolved in the system; m_c/M is the mass fraction of the crystallized magma (see Tait et al. 1989 for further details on the mathematic derivation of K_1 and K_3).

We chose P_i=200 MPa (the pressure at which mixing occurred); n is 0.7 for H_2O solubility in basaltic melts. Tait et al. (1989) calculated a K_1=9.8×10^{-2}, which considers deformation of the surroundings induced by basaltic magma on sedimentary wall-rocks (conditions comparable with those of Mt. Etna) at initial pressure (P_i) = 200 MPa and T=1,100°C. For the modeled case, since the injected magma had already reached saturation at ~250 MPa, a certain amount of gas was in the system when mixing occurred, and K_3 was therefore ≠ 0. For 0.5 wt% and 1 wt% of exsolved water, values of K_3 are 0.23 and 0.11 respectively. Solving Eq. (1), an overpressure ranging between 24 and 27 MPa is obtained, depending on the existing mass of exsolved gas (i.e., $0.11<K_3<0.23$; Fig. 4b). The computed overpressure reached by the magma as a

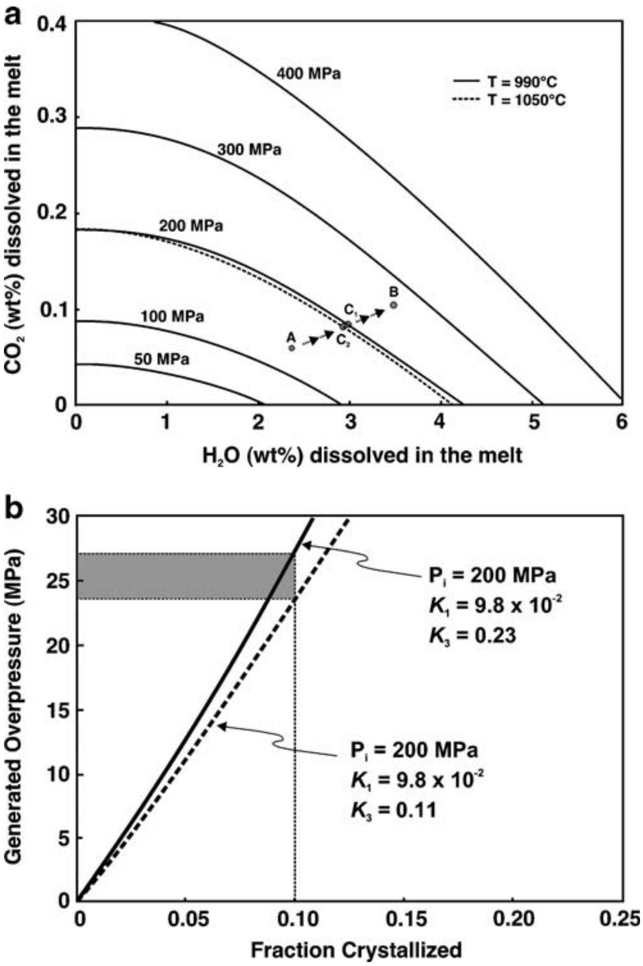

Fig. 4 **a** H_2O and CO_2 (wt%) dissolved in the melt following the solubility model of VOLATILECALC (Newman and Lowenstern 2002). Point A indicates the saturation of a multi-component gas phase for the resident magma at P ~160 MPa with H_2O=2.5 wt%, CO_2=500 ppm and T=990°C. Point B indicates the saturation of a multi-component gas phase for the injected volatile-rich magma at P ~250 MPa with H_2O=3.4 wt%, CO_2=1,100 ppm and T=1,100°C. Arrow-headed line models the path that the resident magma should follow after mixing in order to reach the saturation (Point C_1) of a multi-component gas phase at P=200 MPa, considering that the system partly undergoes thermal equilibration following mixing (T increases to ~1,050°C at the end of the process). The model assumes the injected volatile-rich magma as the volatile source, since it was already saturated in a multi-component H_2O-CO_2 gas phase at P= 200 MPa. The figure shows that H_2O and CO_2 contents of the resident magma at P=200 MPa must increase by ~0.5 wt% and ~300 ppm respectively in order to reach saturation, and therefore exsolution. The isobaric T increase of 60°C (from 990° to 1,050°C) had no measurable effect on additional exsolution of the gas phase at P=200 MPa (Point C_2). **b** Overpressure of the system as a function of T decrease and consequent nucleation and growth of anhydrous mineral phases in the injected volatile-rich magma following the model of Tait et al. (1989). See text for details on the choice of parameters. Assuming that ~10% of solid phases crystallized from magma, a significant overpressure ranging between 24 and 27 MPa can be generated as a function of the gas already exsolved in the system (0.5 to 1 wt%)

result of the crystallization pursuant to mixing was more than 10% of the lithostatic pressure (~200 MPa). This pressure increase was certainly not sufficient to overcome the tensile strength of rock and trigger an eruption. However, it might have facilitated the ascent of magma after fracture opening and very likely it increased the intensity of explosive activity at the surface.

Magma degassing and change in rheology of lavas

In this section the effect on physical properties of lavas erupted at the Laghetto vent resulting from the massive magma degassing that occurred in the upper section of the plumbing system is evaluated.

Low viscosity lava flows

The effusion of lava, from the southern vent at the base of the Laghetto cone, began on July 25, and continued undisturbed for about 5 days at a low output rate (<10 m^3/s), whereas lava fountaining changed to Strombolian activity on July 30. It produced several thin (0.5 m thick) and short (~400 m) flow units, characterized by a quasi-Newtonian rheology, directed westerly on a gentle slope (less than 10°; Calvari and Pinkerton 2004; Ferlito and Siewert 2006; Siewert and Ferlito 2008).

The viscosity η of these flow units was estimated based on their flow rate and thickness, and assuming a Newtonian behavior (Jeffrey 1925):

$$\eta = \rho\ g\ \sin\alpha\ d_f^2 \Big/ 3V \tag{2}$$

where ρ is the specific gravity of the lava, g the gravitational acceleration, α the slope angle, d_f the thickness of the flowing lava unit (0.5 m), and V the average flow rate. ρ was assumed as 2,800 kg/m^3, in agreement with values calculated by MELTS for this trachybasalt lava composition; α=10°; V, measured next to the vent, was around 1 m s^{-1}. The empirical value of Newtonian viscosity from (2) is on the order of 10^3 Pa s. By the end of the eruption, 11 flow units with average thickness of 0.3 m (d_s) had been emplaced (Ferlito and Siewert 2006). This was used to calculate the minimum pressure exerted by the lava load needed to flow on a slope (**τ_0**, **yield strength)**, that is according to Robson (1967) and Hulme (1974):

$$\tau_0 = d_s\,\rho\ g\ \sin\alpha \tag{3}$$

τ_0 is ~1.4×10^3 Pa for the thin lava flows of the Laghetto, well in the range measured and calculated for other Etnean lavas (Walker 1967; Kilburn 1990; Pinkerton and Norton 1995).

High viscosity lava flow

The output of short, low viscosity lava flows from the southern vent of the Laghetto cone lasted 5 days. On the evening of July 30 (8.30 pm Local Time), an abrupt change of the lava output rate occurred, and a huge flow was emitted. Direct observations allowed an estimate of its greatest thickness of ~20 m while moving on a gentle slope (<10°), reduced to 3–4 m when the flow reached a steeper slope (>20°). The lava advanced westward, and extended between the Laghetto vent and the northern flank of the Montagnola cone. Due to a change in the slope direction 300 m from the vent, the flow turned to the south, passed by and set fire to the upper cable-car station (2,500 m a.s.l.), and advanced ~2 km in nearly 12 h. On the early morning of July 31, lava emission at the Laghetto vent ended, when the flow had almost reached Rifugio Sapienza. This large lava flow, ~60 times thicker than the flow units previously emitted from the Laghetto vent, exerted a mechanical erosion at its base and produced a large amount of debris pushed ahead of its front. The overall flow thickness can be related to its Bingham rheology or the resistance to flow at its front (Ferlito and Siewert 2006; Siewert and Ferlito 2008).

To estimate the viscosity of this flow we must consider that the lava moves in a Bingham fashion and a minimum yield strength must be applied to make it flow. From the general relation for Bingham fluids, the following equation is derived:

$$\eta = \tau/(\partial U/\partial y) \qquad (4)$$

where τ is the stress applied to the flowing lava:

$$\tau = d_f \rho \ g \ \sin\alpha \qquad (5)$$

that, for an assumed conservative thickness d_f of 10 m on the 10° slope, is ~4.6×10^4 Pa; $\partial U/\partial y$ is the variation of flow rate perpendicular to flow direction: the flow velocity (U) at the vent was ~0.5 m s^{-1}, the width of the flowing lava (y) at the vent was assumed to be few meters (5–10). The estimate of viscosity so obtained from Eq. (4) is on the order of 10^5 Pa s, which is two orders of magnitude higher than viscosity of the thin flows erupted at the Laghetto vent from July 25 to July 30.

Transition from quasi-Newtonian to Binghamian rheology

Viscosity is among the most important of the physical properties of magmas that controls phenomena such as their flow, crystallization rate and style of volcanic activity. Lava flows emitted from the Laghetto vent displayed two clearly distinct and contrasting rheological behaviors: many thin low viscosity flow units characterized the first phase of activity, whereas one thick and very high viscosity flow unit was emitted in the final phase. This later flow has a trachybasaltic composition like the short low viscosity flow units erupted in the previous days. This means that the abrupt increase of viscosity and therefore the transition from quasi-Newtonian to Binghamian behavior, did not result from change in magma composition, and must be accounted for by other factors.

In addition to chemical composition, the transition from Newtonian to Binghamian rheology is dependent on crystal content (Barmin et al. 2002 and references therein). Here, an effort is made to correlate the increased viscosity of the final lava flow erupted at the Laghetto vent with massive nucleation of crystallites, which is in turn related to the extensive magma degassing that occurred in the upper portion of the Laghetto conduit, and revealed by the exceptionally explosive activity that preceded the emission of the final high viscosity lava flow.

Water dissolved in silicate melts dissociates producing hydroxyls, which reduce polymerization by saturating free valences and lowering O bridging (Burnham 1975; Stolper 1982; McMillan 1994 and references therein). In magmatic systems a pressure decrease lowers H_2O dissociation and hence increases viscosity. If the pressure decrease continues up to H_2O saturation, the ensuing bubble growth will further contribute to raise magma viscosity, due to the enthalpy variation of the magma.

At the Laghetto vent, the magma that fed the high viscosity flow should have lost a large part of its original H_2O content during the preceding explosive activity. The effect of such intense and rapid degassing on magma viscosity has been computed by using MELTS (Ghiorso and Sack 1995; Asimow and Ghiorso 1998). Physical and chemical conditions of magma were set as: T=1,050°C, after thermal equilibration of the system; $P_{initial}$ = 200 MPa and P_{final} 0.1 MPa, to simulate magma ascent from the chamber up to the surface; fO_2=QFM buffer (See also Section 4 for the choice of parameters). The molten fraction was ≥80%, as indicated by the presence of 15–20% of phenocrysts. Viscosity evaluations were performed with decreasing values of P to simulate the effect of decompression and the related volatile loss during ascent (H_2O ranging from 2.8 wt% to less than 1.0 wt%; Fig. 5). A final viscosity of ~10^3 Pa s was obtained by MELTS at the lowest H_2O content, well comparable with the estimate derived by using Eq. (2) for flows of the first phase of emission at the Laghetto vent. However, the viscosity estimated for the lava flow of July 30 is on the order of 10^5 Pa s, which cannot be fully justified as the effect of degassing.

A significant contribution in increasing the viscosity can be provided by crystal nucleation and bubble growth as a consequence of undercooling associated with degassing. In detail, the phase that crystallizes under the lowest degree of

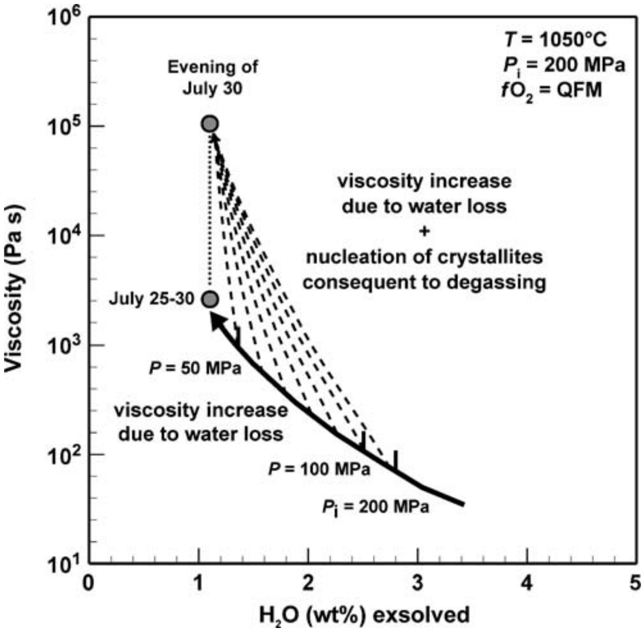

Fig. 5 Viscosity of the system as a function of wt% H_2O exsolved from magma. Solid line indicates the path of viscosity increase calculated by MELTS for an average composition of the mixed magma (Table 1) at the initial conditions of T=1,050°C, P_i=200 MPa, fO_2=QFM buffer and H_2O=2.8 wt%. We have assumed that magma was gradually decompressed from 200 MPa to the surface, loosing its water content. A viscosity on the order of 10^3 Pa s is obtained. The lava of July 30 had a viscosity of ~10^5 Pa s; dashed lines indicate possible paths due to water loss and nucleation of crystallites consequent to degassing that lead to the viscosity increase of about 2 orders of magnitude. Dashed lines depart from the solid one probably when the critical threshold of magma undercooling and nucleation of Fe-oxides and bubbles is initiated

undercooling in mafic magmas is magnetite (Lofgren 1980 and references therein). The presence of crystal nuclei, especially of Fe-Ti oxides, directly controls viscosity and nucleation of the gas phase. Homogeneous gas bubble nucleation from magma can only be initiated under oversaturation conditions, so that gas nuclei can overcome their critical size threshold and grow. The necessary oversaturation can be strongly decreased in presence of crystals which can operate as bubble nucleation germs, whose efficiency will depend on the wetting angle between liquid and bubble, which is higher for Fe-Ti oxides than for other minerals (cf. Burkhard 2002, 2005; Holness 2006; Gualda and Ghiorso 2007).

Once nucleated, bubbles owe their growth to the viscous deformation of the melt and they are therefore favoured in basaltic magmas (Sparks 1978; Toramaru 1995; Gardner et al. 1999; Rust et al. 2003). In the case under study, bubbles nucleating from a trachybasaltic magma within the Laghetto conduit, must have had a very efficient growth and upward movement. The intense degassing rapidly depleted the magma of its H_2O content, and increased crystal nucleation and bubble growth. The final outcome of this process was a dramatic increase of magma viscosity. A quantification of viscosity variation due to H_2O loss in the conduit may be done using thermodynamic simulations (see above), while quantifying the contribution of crystal and bubble growth in increasing viscosity is more difficult.

An attempt to calculate effects of microlites and bubbles on viscosity can be done using the phenomenological Einstein-Roscoe relation (Roscoe 1952) for systems constituted by crystals + melt, where the relative viscosity η_r is defined as:

$$\eta_r = \eta \Big/ \eta_m = (1 - \phi/\sigma)^{-2.5} \tag{6}$$

where η is the effective overall viscosity of the heterogeneous system; η_m is the melt viscosity, ϕ the crystal content and σ is a value that depends on the amount of crystallites and their packing in the melt (cf. Hoover et al. 2001; Saar et al. 2001). Recently Sato (2005), on the grounds of viscosity measurements on basaltic melts of Fuji volcano, proposed a σ value of 0.3 in relation to crystal and distribution morphology. Applying Eq. (6) to the high viscosity lava flow of July 30, a η_r of 10^2 is obtained using for η an estimated viscosity (10^5 Pa s) and for η_m the viscosity calculated by MELTS (10^3 Pa s). The assumed crystal content, on the basis of petrographic evidence, is 0.15. The calculated σ is 0.2, lower than that found by Sato (2005). The difference between σ and ϕ can represent the effective contribution to viscosity due to the fast nucleation of crystals and bubbles following gas loss.

On the evening of July 30, the rheological behavior of the system changed from quasi-Newtonian into Binghamian in a few hours. This implies that, although gas loss was gradual and went on for several days (See Section 2), its effects on the physical properties of the ascending magma were not gradual and became effective only when a critical threshold was overcome and a high yield-strength body was produced. The magma became too rigid to ascend, which led to the end of lava emission and formation of a viscous plug that can still be seen in the Laghetto crater.

Conclusions

The 2001 eruption at Mt. Etna was one of the most interesting episodes of the recent activity, which drew the interest of the scientific community on this basaltic volcano. Here, two particular episodes, reckoned as critical, followed each other at the Laghetto eruptive vent from July 20 to 30. They consisted of 10-day-long period of intense explosive activity, phreatomagmatic to purely magmatic with effusion of low viscosity lava flows, followed by extrusion of a high viscosity lava flow on July 30. These episodes have been investigated and related to input of more primitive and volatile-rich magma into the feeding system of the volcano. Prior to magma ascent, mixing with resident magma caused an increase in exsolution of gas which generated significant magma overpressure and the high explosivity of the eruption at the Laghetto. Later, the emission of a high viscosity flow and, therefore, the transition from quasi-Newtonian to Binghamian rheology of emitted lavas, highlights the role of undercooling related to the intense degassing. In calcalkaline volcanoes, the volatile-dependent viscosity of the ascending magma is viewed as a decisive factor in controlling the fluctuating eruptive regime (cf. Wylie et al. 1999). This is the first time that a volatile dependent viscosity has been investigated during an eruption of Mt. Etna and an attempt made to quantify the relative contributions of different factors.

Explosivity of eruptions and rheology of lava flows are important factors in assessing volcanic hazards. The example discussed here, at the scale of a single eruption, provides new clues for considering additional sources of risk induced by inputs of basic and volatile-rich magmas into the feeding systems of basaltic volcanoes.

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