Publisher: GSA
Journal: GEOL: Geology
DOI:10.1130/G37807.1

Temporal variations of helium isotopes in volcanic gases quantify pre-eruptive refill and pressurization in magma reservoirs: The Mount Etna case

Antonio Paonita*, Antonio Caracausi, Mauro Martelli, and Andrea L. Rizzo

Istituto Nazionale di Geofisica e Vulcanologia–Sezione di Palermo, Via Ugo La Malfa 153, 90146 Palermo, Italy

*E-mail: antonio.paonita@ingv.it

ABSTRACT

Two approaches to the challenging aim of forecasting impending eruptions are searching for empirical precursors and developing suitable interpretative models. Here we present high-resolution time series of $^3He/^4He$ ratios measured in gases emitted from peripheral vents around Mount Etna volcano (Italy), which revealed variations with strong correlations over both time and a broad spatial scale. The main eruptive episodes are preceded by increases in $^3He/^4He$, making this ratio a unique tracer for monitoring volcanic activity. These features strongly reflect pressurization beneath the volcano due to deep magma influx. We propose a pioneering model that relates the changes in $^3He/^4He$ to the time-dependent outflow of volatiles from a magmatic chamber subjected to evolution of its internal pressure due to magma injection. At Mount Etna, the model makes it possible to estimate in near real time key parameters such as the rate of magma input and volume change in deep chamber preceding eruptions, and to compare them with geodetic estimations. This represents an unprecedented use of $^3He/^4He$ to obtain quantitative information on the physics of magmatic systems. Volcanoes showing

changes of $^3He/^4He$ ratio in discharged gases due to unrest episodes are widespread in the world, and therefore we envisage extensive future applications of this approach.

INTRODUCTION

Episodes of magma injection into subvolcanic reservoirs are recognized as the main trigger of the eruptive activity of volcanoes (Caricchi et al., 2014). Along with other signals, the arrival of the deep magma can be detected by means of increases in the He isotope ratio ($^3He/^4He$) measured in outgassed fluids (Sano and Fischer, 2013, and references therein). This is possible because the ascending primitive magmas outgas volatiles having $^3He/^4He$ ratios that are higher than those of fluids from more evolved melts and/or crust. In the recent case of the Mount Ontake eruption in Japan (Sano et al., 2015), this tracer was the only one capable of providing clues about increasing activity over a time scale of years. In spite of this, the $^3He/^4He$ ratio has never been used to achieve physical information of magmatic systems.

Here we describe a 12-yr-long time series of the $^3He/^4He$ ratio in volcanic gases emitted from Mount Etna volcano (Italy) that clearly shows that the main eruptive phases are preceded by increases in this ratio. An unprecedented use of this tracer allows us to quantify, in near real time, the flow of magma entering a reservoir beneath the volcano and the related pressure buildup at depth.

DATA AND OBSERVATIONS AT MOUNT ETNA

The $^3He/^4He$ ratios of the gases emitted from 4 vents located along the flanks of Mount Etna (Fig. 1), 15–20 km from the summit, have been monitored for two decades at a nearly bimonthly sampling rate, yielding a matchless time series. The $^3He/^4He$ ratio of each vent site fluctuates around its specific average value (Fig. 1), and the decreases in

DOI:10.1130/G37807.1

the average $^3He/^4He$ values from the P39 to Fd vents cannot be caused by an increasing contribution of crustal/hydrothermal fluids, as clearly indicated by the geochemical features of the gases (Caracausi et al., 2003). Instead, the $^3He/^4He$ ratios of the sites indicate that magmas with isotope ratios lower than that of pristine Etnean magma (~7.6 Ra, where Ra = $^3He/^4He_{air}$; Correale et al., 2014) are stored in the plumbing system and feed the vents. En route processes (chiefly, ^4He production by U and Th radioactive decay) reduce the isotope ratios of Mount Etna resident magmas (Correale et al., 2014). Three of the monitored sites are separated by only 2–3 km (Fig. 1), making it highly unlikely that each of them is fed by a distinct magmatic body with its own $^3He/^4He$ marker. More probably, the gas end member from the high $^3He/^4He$ pristine magmas (7.6 Ra) mixes with a low $^3He/^4He$ end member from resident magmas before feeding the vents (Fig. 2). The differences in average values among the vents hence reflect different mixing proportions of these two end members.

Figure 1 indisputably shows that increases of 0.2–0.7 Ra occur in all the vents simultaneously, followed by similar decreases. Considering that one vent is up to 35 km from the others, the synchronous changes on such a wide scale testify to a common deep process occurring beneath the volcano (Fig. 2). A relative increase in the contribution of the deep-origin **[[SU: correct meaning, vs. "coming"?]]** high $^3He/^4He$ end member with respect to the low $^3He/^4He$ one explains the growth of isotope ratio at all sites very well.

The investigated period represents a phase of considerable activity of Mount Etna during which the volcano exhibited at least six major eruptions that occurred in 2001, 2002–2003, 2004–2005, 2006, 2008, and 2011–2012. All of these eruptions were characterized by considerable volumes of emitted magma (>10^7 m^3), and most were gas rich and highly explosive, with remarkable fire-fountaining activity (Bonaccorso and Calvari, 2013). The 2004–2005 eruption was the only one triggered by a passive response to eastward flank movements, with magmatic overpressure playing a minor role (Burton et al., 2005). An impressive feature in Figure 1 is that the main eruptive phases are systematically preceded by increases in $^{3}He/^{4}He$ lasting from a few weeks to several months, with the sole exception of the 2004–2005 event. This emphasizes the potential of this tracer in forecasting eruptive periods, as already highlighted at Mount Etna (Paonita et al., 2012, and references therein). Moreover, it strongly suggests that the increases in $^{3}He/^{4}He$ (and consequently the increase in the contribution of the high $^{3}He/^{4}He$ end member) are associated with input of high $^{3}He/^{4}He$, gas-rich pristine magma into the system and consequent pressurization that may lead to eruptions.

PHYSICAL MODEL

Each recharge episode thus introduces a volume of the high $^{3}He/^{4}He$ magma into the reservoir that is connected more directly to the source (i.e., the deep reservoir in Fig. 2). Due to this direct connection, such an entering magma will have a very similar He isotope signature with respect to the magma already stored in the deep reservoir (i.e., 7.6 Ra). The massive entrainment of melts in this confined zone increases the internal pressure and hence the amount of outgassed fluids, namely an increase in the contribution of the high $^{3}He/^{4}He$ gas end member. Conversely, magma batches stored more marginally in the system are poorly hydraulically connected, if at all, to the deep reservoir, as supported by models of plumbing systems operating on time scales of months to years that include magma compartments (Gudmundsson, 2012). Both the internal pressure and the low $^{3}He/^{4}He$ signature of these distal batches of magmas are not therefore affected by the refilling event and, for the same reason, the output of fluids from these magmas does not undergo any change. After the end of the magma input, when the overpressure exhausts, the excess contribution of the high $^3He/^4He$ end member terminates, and the $^3He/^4He$ ratios of the sites return to their background values.

We modeled this complex process by coupling a mass balance between the two gas end members to a physical model of a magma chamber, which is **[[SU: correct?]]** the part of the deep reservoir that pressurizes (Fig. 2; see the Appendix; details are provided in the GSA Data Repository[1]). During periods without input of new magmas (background), a site S **[[SU: italicize variables only; and not subscripts]]** receives magmatic gases with a measured $^3He/^4He$ ratio ($R_{S,B}$) and He concentration ($He_{S,B}$). This already constitutes a mixture of high and low $^3He/^4He$ end members, with the total flow (Q_B) of this mixture being constrained by measurements of the total output of volatiles from the volcano during quiescent degassing. When a deep input pressurizes the chamber, the latter releases large amounts of gases with flow rate $Q_{g,o}(t)$, **[[SU: define *t* (time?) here]]** known He concentration He_o (Paonita et al., 2012), and having the marker of the high $^3He/^4He$ end member (R_o). The gases combine with the previous mixture according to a mixing fraction x_S, so as to change $^3He/^4He$ at site S [$R_S(t)$]. $Q_{g,o}(t)$ is fed by a foam layer that grows in the chamber, and depends on the rock permeability of the chamber roof, which in turn is controlled by overpressure (Zencher et al., 2006). The chamber overpressure (ΔP) changes over time as a function of (1) parameters constrainable by data in the literature or those having only a minor effect (e.g., chamber and rock properties; for explanation, see the Data Repository) and (2) mix fraction x_S (not constrainable from external information) and melt input $Q_{m,i}$ in the chamber (to be estimated for each recharge event), which are discussed in the following.

RESULTS AND DISCUSSION

We apply the model to two particular phases of Mount Etna activity: the 2010–2012 and 2001 events. During the 2010–2012 period, the eruption started while $^3He/^4He$ was increasing, and ended in early 2012 when the ratios were at the minimum values (Fig. 1). This suggests that the entire erupted volume was probably supplied by a single recharge event. Because Mount Etna has displayed a tendency to maintain an equilibrium between magma input and output over the past 40 years (Bonaccorso and Calvari, 2013), the volume that erupted during 2010–2012 ($50 \pm 10 \times 10^6$ m^3; mean from Bonaccorso and Calvari, 2013; Behncke et al., 2014) is a good proxy of the recharge volume from depth. Divided by the time span of the $^3He/^4He$ variation, we achieve a deep input of melt of ~0.95 m^3/s (plus the gas input computed from the gas/melt ratio; Table 1; see the Data Repository). This input in the chamber results in the overpressure increasing toward an asymptotic value (Fig. 3A). Due to the overpressure-dependent rock permeability, the increase in the gas flow rate from the chamber [$Q_{g,o}(t)$] lags both the overpressure increase and the gas flow from melt into the foam (F_g), thereby producing a sigmoid curve. As soon as the deep input ends, the chamber elastically returns to its original volume while the overpressure decays exponentially (Woods and Huppert, 2003) (Fig. 3A). $Q_{g,o}$ decays even more quickly due to the overpressure-driven decrease in permeability. Figure 3B shows the calculated $^3He/^4He$ ratio for two sites [$R_S(t)$]. The simulated and observed signals are surprisingly similar in both shape and time scale. The model predicts the sharp increase in $^3He/^4He$, followed by a smooth trend toward an asymptote. When the deep input of magma terminates, the model correctly predicts the $^{3}He/^{4}He$ decrease and the restoration of background values. We note that the selected mixing fraction x_S (the only parameter we are constraining based on $^{3}He/^{4}He$ data) affects the highest $^{3}He/^{4}He$ value reached during the simulation, but not the time scale over which this is reached. Instead, the time scale is affected by the deep input of magma entering the chamber. Curves computed with markedly differing values of deep input are unable to reproduce the observables irrespective of x_S (see the Data Repository).

The model suggests a striking sequence of events during the 2010–2012 period (Fig. 3B). First, a melt plus gas input started to feed the deep plumbing system in February 2010, causing the observed increase in $^{3}He/^{4}He$ that peaked ~1 yr later close to the maximum possible value under those conditions. This suggests that the overpressure, and then the ascent of melt toward shallow levels, reached the highest values, and accordingly the eruptive activity (fire fountaining) started. At this stage the model predicts a volume increase of ~2×10^{7} m^3 for the chamber. At the end of 2011, while Mount Etna was still erupting, the $^{3}He/^{4}He$ ratio started to decrease dramatically. Our model suggests that the deep input ceased and, at depth, both the overpressure and magma ascent to shallow levels were decreasing rapidly. This resulted in the supply of eruption being extinguished, and accordingly the fire fountaining ended in April 2012. Supporting evidence comes from areal dilatation signals calculated from data collected at three summit GPS stations and three intermediate altitude ones, which are considered as representative of shallow and deep volume changes in Mount Etna plumbing system, respectively (Aloisi et al., 2011). The input that occurred during February 2010 matches a change in the rate of dilatation at intermediate altitude, whereas there was negligible deformation of the summit area (Patanè et al., 2013). This pattern of observations was attributed to an active inflation source located 8 km below sea level (bsl) under the summit craters (Patanè et al., 2013), in agreement with the depth of our modeled chamber (see Table 1). The observed deformation was generated by a volume change of ~5×10^7 m^3 (Patanè et al., 2013), which is amazingly close to the volume increase estimated by our model. The input started some weeks after the deep (20–30 km) earthquake swarm affecting the northwestern sector of the volcano on 19 December 2009, which was the most intense event during that period (Patanè et al., 2013). Volcano-tectonic swarms that occur at these levels are thought to be predictive of magmatic replenishments feeding future volcanic activities (Patanè et al., 2013).

The rate of $^3He/^4He$ increase and its absolute value were highest during a violent eruption that occurred in July–August 2001 (Fig. 1). This sharp peak followed a smoother increase that started several months earlier, suggesting that the paroxysmal recharge occurred during a longer period of slower injection of magma. This means that we cannot use the erupted volume as a proxy of the recharge volume to constrain the deep input of magma. Instead, we utilize the constrained x_S value during 2010–2012, and adjust the deep melt input $Q_{m,i}$ to fit the $^3He/^4He$ increase that occurred in July–August 2001. As a result, the $^3He/^4He$ increase was due to a deep magmatic input much larger (~2.7 m^3/s) than that predicted for 2010–2012 (Fig. 3D). It further agrees with the magma emission rates observed during the 2001 eruption, which were the highest of the entire investigated period (Behncke and Neri, 2003). The input started in May (Figs. 3C and 3D), ~2 weeks after the seismic swarm of 22 April 2001, interpreted as an intrusive process occurring at ~8 km bsl (Bonaccorso et al., 2004). The onset of the 2001 eruption on 17 July was

DOI:10.1130/G37807.1

anticipated by an impressive seismic swarm (12–17 July; Figs. 3C and 3D) and abrupt ground deformation that indicated the propagation of a dike from the shallow reservoir at 3 km bsl (Neri et al., 2005), which overtops our modeled chamber (Fig. 2). Our simulation predicts that, at the moment of the seismic swarm, the overpressure in the deep chamber exceeded the tensile strength of the local wall rocks (Fig. 3C; see the Appendix). The connection between the deep and shallow reservoirs also suggests that the latter was seriously overpressurized.

Modeling the effect of dike opening on overpressure and gas release from the deep chamber is outside the aim of this work. The high $^{3}He/^{4}He$ of olivine-hosted fluid inclusions that erupted in 2001 suggests that the new dike carried high $^{3}He/^{4}He$ magmas directly to the surface (Correale et al., 2014), explaining the amazingly high $^{3}He/^{4}He$ ratio at vent P39. In addition, the abrupt outpouring of magma after dike opening guaranteed a drop in the overpressure. The consequent decrease in gas output $Q_{g,o}(t)$ from the chamber could explain the very high rate of $^{3}He/^{4}He$ decrease (higher than model prediction; Fig. 3D) observed between 7 and 10 August, when the eruption was ending.

CONCLUSION

This work highlights that temporal variation of $^{3}He/^{4}He$ is a valuable indicator for assessing the level of volcanic activity and can provide key physical parameters such as the magma input rate and overpressure in the chamber, which until now have only been available from geodetic approaches. The conditions for chamber stability and wall-rock failure can be determined, and will improve the accuracy of eruption forecasting. Our approach is predicted to be widely applicable, because time-dependent He isotope mixing between primitive and more radiogenic end members appears to be common in many active volcanoes (Sano and Fischer, 2013). Long-lasting time series with sufficiently frequent samplings and high-precision $^{3}He/^{4}He$ measurements in air-free volcanic gases are required, since even small isotope variations (fractions of 1 Ra unit) can reflect important volcanic processes.

APPENDIX: MODEL AND PARAMETERS

Here we recap the main equations of the model (for full details, including parameterization and sensitivity study, see the Data Repository). The $^{3}He/^{4}He$ ratio at site S, after mathematical manipulation, can be computed using**[[SU: Note, the only characters that should be italicized in Equation 1 are variables R, Q, t, and x; subscripts and He (helium) are not italicized]]**

$$R_{\mathrm{S}}(t)=\frac{0.5Q_{\mathrm{B}}\mathrm{He}_{\mathrm{S,B}}R_{\mathrm{S,B}}+Q_{\mathrm{g,o}}(t)\mathrm{He}_{\mathrm{o}}R_{\mathrm{o}}x_{\mathrm{S}}}{0.5Q_{\mathrm{B}}\mathrm{He}_{\mathrm{S,B}}+Q_{\mathrm{g,o}}(t)\mathrm{He}_{\mathrm{o}}x_{\mathrm{S}}} \quad (1)$$

(all symbols are defined in the text). Calculation of $Q_{\mathrm{g,o}}$ (t), i.e., the gas outflow from the chamber, requires a physical model of the reservoir. We modified the approach of Woods and Huppert (2003) to account for the input and output of gases, as well the formation of a foam layer. This approach has no pretentions of being complete, but it does capture the key processes governing chamber overpressure (Woods and Huppert, 2003). Pressure and volume changes in the chamber can be described by:

$$f\,\frac{d\Delta P}{dt}=\frac{d\Delta V}{dt}=Q_{\mathrm{g,i}}+Q_{\mathrm{m,i}}-Q_{\mathrm{g,o}}-Q_{\mathrm{m,o}}, \quad (2)$$

where subscripts m, g, i, and o refer to melt, gas, input, and output volumetric flows into and out from the chamber, respectively, and f is the chamber volume divided by the effective compressibility of the melt-gas-rock system, which depends on gas fraction n_{g}

in the chamber (Woods and Huppert, 2003). The value of n_g depends on the difference between the input and output of gas:

$$\frac{dn_g}{dt} = \frac{\rho_g}{M_m}(Q_{g,i} - Q_{g,o}), \tag{3}$$

where M_m is the mass of magma (with density ρ_m) in chamber volume V_{ch}, and ρ_g is the gas density. At any time n_g will be shared between dispersed bubbles in the melt plus a fraction in the foam. If n_b is the amount of gas in the dispersed bubbles, it evolves as (Woods and Cardoso, 1997):

$$\frac{dn_b}{dt} = \frac{Q_{g,i}\rho_g}{M_m} - \frac{v_b}{H} n_b, \tag{4}$$

where v_b is speed of bubble ascent (from Stokes's law). Output flows $Q_{g,o}$ and $Q_{m,o}$ are functions of ΔP:

$$Q_{g,o} = \frac{k(\Delta P) \cdot A_{ch}}{\mu_g}\left[\frac{\Delta P}{h} + (\rho_m - \rho_g)g\right], \tag{5}$$

$$Q_{m,o} = \frac{\pi r^4}{8\mu_m Z}\Delta P, \tag{6}$$

where $k(\Delta P)$ is the overpressure-dependent permeability (Zencher et al., 2006), h is a characteristic vertical distance from the chamber roof at which the effect of ΔP disappears, A_{ch} is the roof area, μ_m and μ_g are the melt and gas viscosities, respectively, r and Z are the radius and height of the output conduit of melt, respectively, and g is gravity.

All of the required parameter values are listed in Table 1.

ACKNOWLEDGMENTS

We thank J. Lowenstern, R. Herd, and an anonymous reviewer for their helpful comments.

REFERENCES CITED

Aiuppa, A., et al., 2008, Total volatile flux from Mount Etna: Geophysical Research Letters, v. 35, L24302, doi:10.1029/2008GL035871.

Aloisi, M., Mattia, M., Monaco, C., and Pulvirenti, F., 2011, Magma, faults, and gravitational loading at Mount Etna: The 2002–2003 eruptive period: Journal of Geophysical Research, v. 116, B05203, doi:10.1029/2010JB007909.

Behncke, B., and Neri, M., 2003, The July–August 2001 eruption of Mt. Etna (Sicily): Bulletin of Volcanology, v. 65, p. 461–476, doi:10.1007/s00445-003-0274-1.

Behncke, B., Branca, S., Corsaro, R.A., De Beni, E., Miraglia, L., and Proietti, C., 2014, The 2011–2012 summit activity of Mount Etna: Birth, growth and products of the new SE crater: Journal of Volcanology and Geothermal Research, v. 270, p. 10–21, doi:10.1016/j.jvolgeores.2013.11.012.

Bonaccorso, A., and Calvari, S., 2013, Major effusive eruptions and recent lava fountains: Balance between expected and erupted magma volumes at Etna volcano: Geophysical Research Letters, v. 40, p. 6069–6073, doi:10.1002/2013GL058291.

Bonaccorso, A., D'Amico, S., Mattia, M., and Patanè, D., 2004, Intrusive mechanisms at Mt. Etna forerunning the July–August 2001 eruption from seismic and ground deformation data: Pure and Applied Geophysics, v. 161, p. 1469–1487, doi:10.1007/s00024-004-2515-4.

Bonaccorso, A., Cianetti, S., Giunchi, C., Trasatti, E., Bonafede, M., and Boschi, E., 2005, Analytical and 3-D numerical modelling of Mt. Etna (Italy) volcano inflation: Geophysical Journal International, v. 163, p. 852–862, doi:10.1111/j.1365-246X.2005.02777.x.

Burton, M., et al., 2005, Etna 2004–2005: An archetype for geodynamically-controlled effusive eruptions: Geophysical Research Letters, v. 32, L09303, doi:10.1029/2005GL022527.

Caracausi, A., Italiano, F., Nuccio, P.M., Paonita, A., and Rizzo, A., 2003, Evidence of deep magma degassing and ascent by geochemistry of peripheral gas emissions at Mt. Etna (Italy): Assessment of the magmatic reservoir pressure: Journal of Geophysical Research, v. 108, no. B10, 2463, doi:10.1029/2002JB002095.

Caricchi, L., Annen, C., Blundy, J., Simpson, G., and Pinel, V., 2014, Frequency and magnitude of volcanic eruptions controlled by magma injection and buoyancy: Nature Geoscience, v. 7, p. 126–130, doi:10.1038/ngeo2041.

Correale, A., Paonita, A., Martelli, M., Rizzo, A., Rotolo, S.G., Corsaro, R.A., and Di Renzo, V., 2014, A two-component mantle source feeding Mt. Etna magmatism: Insights from the geochemistry of primitive magmas: Lithos, v. 184, p. 243–258, doi:10.1016/j.lithos.2013.10.038.

Gudmundsson, A., 2012, Magma chambers: Formation, local stresses, excess pressures, and compartments: Journal of Volcanology and Geothermal Research, v. 237–238, p. 19–41, doi:10.1016/j.jvolgeores.2012.05.015.

Jamtveit, B., and Yardley, B., 1997, Fluid flow and transport in rocks: Mechanisms and effects: London, UK, Chapman & Hall, 315 p.

Johnson, S.E., Jin, Z.-H., Naus-Thijssen, F.M.J., and Koons, P.O., 2011, Coupled deformation and metamorphism in the roof of a tabular midcrustal igneous complex: Geological Society of America Bulletin, v. 123, p. 1016–1032, doi:10.1130/B30269.1.

Neri, M., Acocella, V., Behncke, B., Maiolino, V., Ursino, A., and Velardita, R., 2005, Contrasting triggering mechanisms of the 2001 and 2002–2003 eruptions of Mount Etna (Italy): Journal of Volcanology and Geothermal Research, v. 144, p. 235–255, doi:10.1016/j.jvolgeores.2004.11.025.

Paonita, A., Caracausi, A., Iacono-Marziano, G., Martelli, M., and Rizzo, A., 2012, Geochemical evidence for mixing between fluids exsolved at different depths in the magmatic system of Mt Etna (Italy): Geochimica et Cosmochimica Acta, v. 84, p. 380–394, doi:10.1016/j.gca.2012.01.028.

Patanè, D., et al., 2013, Insights into magma and fluid transfer at Mount Etna by a multiparametric approach: A model of the events leading to the 2011 eruptive cycle: Journal of Geophysical Research, v. 118, p. 1–21, doi:10.1002/jgrb.50248.

Sano, Y., and Fischer, P.T., 2013, The analysis and interpretation of noble gases in modern hydrothermal systems, *in* Burnard, P., ed., The noble gases as geochemical tracers: Berlin, Springer-Verlag, p. 249–317, doi:10.1007/978-3-642-28836-4_10.

Sano, Y., Kagoshima, T., Takahata, N., Nishio, Y., Roulleau, E., Pinti, D.L., and Fischer, P., 2015, Ten-year helium anomaly prior to the 2014 Mt Ontake eruption: Scientific Reports, v. 5, 13069, doi:10.1038/srep13069.

Woods, A.W., and Cardoso, S.S.S., 1997, Bubble magma separation as a trigger of basaltic volcanic eruptions: Nature, v. 385, p. 518–520, doi:10.1038/385518a0.

Woods, A.W., and Huppert, H.E., 2003, On magma chamber evolution during slow effusive eruptions: Journal of Geophysical Research, v. 108, 2403, doi:10.1029/2002JB002019.

Zencher, F., Bonafede, M., and Stefansson, R., 2006, Near-lithostatic pore pressure at seismogenic depths: A thermoporoelastic model: Geophysical Journal International, v. 166, p. 1318–1334, doi:10.1111/j.1365-246X.2006.03069.x.

FIGURE CAPTIONS

Figure 1. Time variations of $^{3}He/^{4}He$ measured at P39, St, VS, and Fd gas vents of Mount Etna (Italy) during the period 2001–2012 (locations shown in inset; see Paonita et al., 2012, for sampling and analytical methods). Sampling times are the same for all sites and are shown only for VS. Isotope values are in units of R/Ra = $(^{3}He/^{4}He_{sample})/(^{3}He/^{4}He_{air})$, and are corrected for (very small) air contamination (Caracausi et al., 2003). The values of mean R/Ra ± 1σ at P39, VS, St, and Fd are 7.05 ± 0.17, 6.57 ± 0.14, 6.21 ± 0.15, and 6.17 ± 0.14, respectively. Red lines highlight the eruptive phases; gray areas indicate the modeled periods.

Figure 2. Sketch of the plumbing system, with magma-plus-fluids storage zones (yellow) and the crystalline body (blue; modified from Patanè et al., 2013). The shallow (3 km below sea level, bsl) and deep (5–12 km bsl) reservoirs (res.) are from Paonita et al. (2012); dark red to orange indicate magmas with high to low $^{3}He/^{4}He$. Dotted arrows mark the magma ascent pathway. Brown and orange arrows indicate pathways of high and low $^{3}He/^{4}He$ gaseous end members, respectively. The high $^{3}He/^{4}He$ magma chamber is shown (ellipse). The right diagram provides details of the magma chamber (symbols as in the text; also see Table 1). Fumar.—fumarole. **[[SU: symbols and variables should be as in text (subscripts and Greek characters should not be italicized, but P and t should be).]]**

Figure 3. A: Time variation of overpressure Δ*P*, gas flow into foam F_g, and gas output from chamber $Q_{g,o}$. B: Time variation of $^3He/^4He$ of gases from vents St and P39, computed for the 2010–2012 input. C, D: Same as A and B, but computed for the 2001 input (P39 site only). B and D also show the measured $^3He/^4He$ signals. The start and stop of magma input are indicated. Thick curves for vents St and P39 in B differ by the selected background and x_S values (see Table 1). Background for each site is the minimum $^3He/^4He$ ratio of the time series. Thin dotted curves for St show the effect of changing x_S: 0.25 (higher) and 0.15 (lower) versus 0.18 (best fit; see Table 1). Eruptions are shown as red bars. **[[SU: Need uppercase A–D in figures; "g,o" after Q should be subscripts, Q should be italicized, g after F should be subscript, F should be italicized. On left axes and in legends of A and C, "P" should be italicized; "mar" at bottom of D should be spelled out (March).]]**

[1]GSA Data Repository item 2016xxx, **[[SU: Need DR item names and brief descriptions here]]**, is available online at www.geosociety.org/pubs/ft2016.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

Publisher: GSA
Journal: GEOL: Geology
DOI:10.1130/G37807.1

TABLE 1. MODEL PARAMETERS

Chamber:	
height (H), volume (V_{ch})	1 km†, 4.0 km^3 [1]
temperature (T_m)	1150 °C[2]
pressure (P_m), depth (d)	300 MPa*[2], 8.0 km bsl*
dissolved H_2O, CO_2	3.5 wt%, 1800 ppm [2]
H_2O, CO_2 in vapor	35 mol%, 65 mol% [2]
radius of bubbles	200 μm [S]
Conduit:	
radius (r), height (Z)	0.2 m*, 3.0 km*
Density:	
melt (ρ_m), gas(ρ_g)	2600 kg/m^3*, 870 kg/m^3*
Viscosity:	
melt (μ_m), gas (μ_g)	3.0 Pa·s[S], $5 \times 10^{\square 5}$ Pa·s[S]
Young's modulus (G)	15 GPa [3]
Poisson's ratio (ν)	0.25 [3]
Melt bulk modulus (β_m)	10 GPa*
Distance (h)	200 m [4]
Deep input:	
melt ($Q_{m,i}$) in 2010–2012	0.95 m^3/s*[5]
melt ($Q_{m,i}$) in 2001	2.7 m^3/s*
gas/melt ratio ($Q_{g,i}/Q_{m,i}$)	0.5*[5,6]
Gas output from chamber:	
He concentration (He_o)	6.0 ppm [2]
$^3He/^4He$ (R_o)	7.6 R/Ra [7]
Rock permeability:	
constant permeability (k^o)	$5 \times 10^{\square 18}$ m^2 [8]
fracture half-length (l)	2.0 mm*[9]
fracture distance (D)	0.5 m*[9]
Background gas flow:	
Gas flow rate (Q_B)	0.14 m^3/s*[6]
He concentration ($He_{S,B}$)	6.0 ppm [2]
$^3He/^4He$ at vent St ($R_{S,B}$)	5.96 R/Ra*
$^3He/^4He$ at vent P39 ($R_{S,B}$)	6.76 R/Ra*
mix fraction at St ($x_{S,}$)	0.18*
mix fraction at P39 (x_S)	0.88*

Note: bsl—below sea level. References in superscript brackets: 1—Bonaccorso et al., 2005; 2—Paonita et al., 2012; 3—Aloisi et al., 2011; 4—Johnson et al., 2011; 5—Bonaccorso and Calvari, 2013; 6—Aiuppa et al., 2008; 7—Correale et al., 2014; 8—Jamtveit and Yardley, 1997; 9—Zencher et al. 2006; S—Other sources (see text footnote 1**[[SU: which DR item?]]**).

*Computed.

†Assumed in this work.

[[SU: what should highlighted superscript be before 18? What is the superscript o in permeability k^o?]]