Stability and instability of quiescently active volcanoes: The case of Masaya, Nicaragua

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ABSTRACT

Quiescently active volcanoes are enigmatic due to their restlessness but lack of eruptive activity. I present a model of coupled conduit convection and foam accumulation to explain degassing behavior of Masaya, an active volcano in Nicaragua that is currently emitting large amounts of gas but not erupting. Gas-rich magma is transported through a conduit 2–6 m in radius and then released into a shallow reservoir. The magma is degassed in the reservoir and forms a foam 1–3 m thick at the top with bubbles 20–60 μm in diameter. The foam layer is stable because the input of gas into the foam is balanced by gas release through a vent to the surface. If the foam layer is destabilized, the volcano can erupt explosively. The most likely cause of foam destabilization is a large injection of volatile-rich magma from deeper levels into the shallow conduit-reservoir system, thereby increasing magma fluxes and gas fluxes, exsolving large gas bubbles, and reducing surface tension of the magma.

Keywords: quiescent volcanoes, conduit convection, foam accumulation, volcanoes, Masaya.

INTRODUCTION

Volcanic activity is manifested principally by eruptions of lava and pyroclastic debris. Yet some volcanoes, such as Masaya (Nicaragua) and Miyakejima (Japan), exhibit little or no eruptive activity, but are still dynamic systems. Their activity is commonly characterized by high levels of geophysical unrest, such as degassing and seismicity, but these volcanoes erupt little or no solid material. Because of this behavior, the volcanic systems may be considered as stable or metastable, because they erupt infrequently. Such volcanoes pose a fundamental problem for volcano hazards and forecasting. Although we can observe that these systems are active, we generally do not know the parameters that are important for destabilizing such systems and generating eruptions, which can threaten lives and livelihoods.

This paper addresses the problem of persistent quiescent volcanism and the mechanisms that control it. The goal of the paper is to apply some well-established models of magma dynamics in reservoirs and conduits to examine if they can adequately explain the behavior of natural systems, in this case Masaya volcano in Nicaragua. I first examine magma behavior and degassing by means of these models, determining gas fluxes, conduit dimensions, and the characteristics of a foam layer at the top of a magma reservoir. I then discuss the strengths and weaknesses of each, and determine if the different models can be coupled. I conclude by proposing a general view of the conditions under which quiescent volcanism occurs, as well as the parameters that potentially can destabilize such systems and produce eruptions.

MASAYA VOLCANO

Masaya volcano in Nicaragua is a basaltic caldera that forms part of the Central American volcanic arc system. Masaya has undergone several episodes of caldera subsidence associated with Plinian pyroclastic surge and flow eruptions (Williams, 1983). The youngest part of the volcano consists of a series of cones and pit craters that have built an edifice within the most recent caldera. Magma has erupted from this edifice in historic time in the form of lava flows that have accumulated within the caldera depression. The current activity is centered on Santiago crater, a pit crater ~1 km in diameter that formed in 1859 after a series of explosions (McBirney, 1956). Since then, Santiago has had episodes of intense degassing that last years to decades. Intermittent lava lakes at the bottom of the crater are also observed, but little or no magma is erupted out of the crater. The volcano is characterized by persistent tremor (Metaxian et al., 1997), most likely the result of magma movements and/or degassing. The most recent degassing crisis began in 1993 and continues as of this writing (January 2007).

In terms of quiescent volcanism, Masaya is important for two reasons. First, it shows extreme behavior, in the sense that strong magmatic degassing occurs with nearly no magma erupted. Second, the volcano has been studied intensively since 1992 by a series of integrated microgravity and gas investigations (e.g., Delmelle et al., 1999), which provide high-quality data essential for modeling purposes.

MODELS OF QUIESCENT VOLCANISM

Two models have been proposed to explain the behavior of quiescent volcanoes. The first considers magma convection in a conduit (Kazahaya et al., 1994; Stevenson and Blake, 1998), in which buoyant gas-rich magma is transported upward in the central portions of the conduit, releasing its gas by decompression as it rises. The degassed magma is denser and descends along the margins of the conduit (Fig. 1A). The second model addresses the development of a foam layer (Jaupart and Vergniolle, 1989), whereby bubbles rising through magma accumulate at the roof of a reservoir. The rise and accumulation of bubbles is balanced by gas loss through a conduit connecting the top of the reservoir to the surface (Fig. 1B). I apply the two models to Masaya.

Conduit Convection

Conduit convection and degassing comprise a central plug of rising magma that is buoyant because of volatiles dissolved in the magma. As the magma rises, it loses gas by decompression. The gas loss increases the density and viscosity of the magma, causing it to become gravitationally unstable and descend along the walls of the conduit. In this manner, a convective circulation system is established within the conduit system, promoting efficient magma degassing. The mass upflow rate of the magma was defined by Stevenson and Blake (1998) as follows:

$$Q_m = \frac{\rho_a \pi R^{*2} P_s g \Delta\rho R^4}{\mu_d}, \quad (1)$$

where ρ_a is the density of gas-rich upflowing magma, R^* is the ratio of the radius of upwelling magma to that of the conduit R, g is gravitational acceleration, $\Delta\rho$ is the density difference between ascending and descending magma, and μ_d is the viscosity of descending magma. The parameter P_s is a dimensionless Poiseuille number, $v\mu_d/g\Delta\rho R^2$, where v is the velocity of rising magma at a low Reynolds number.

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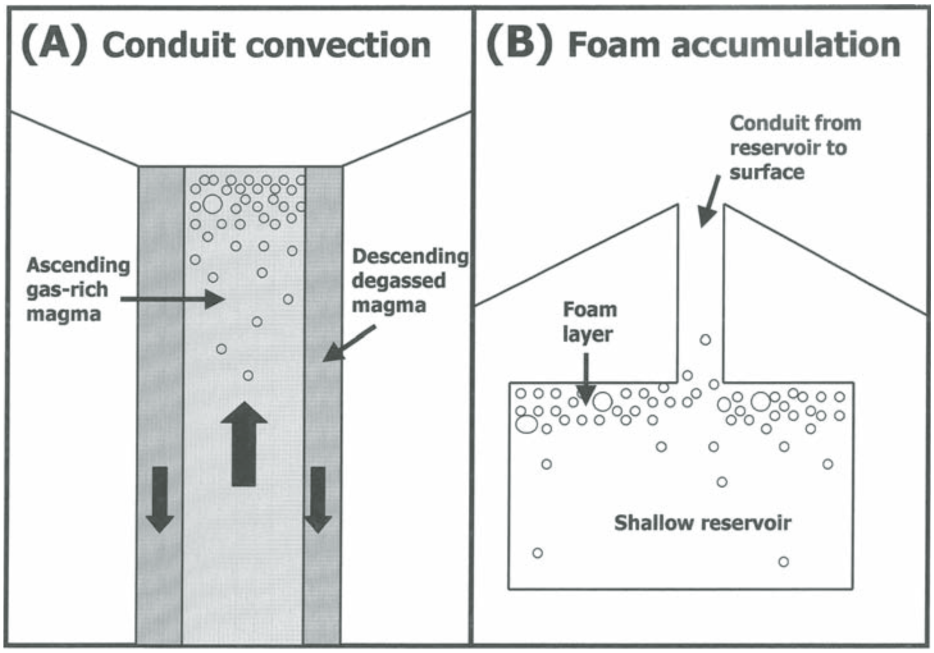

Figure 1. Two models of quiescent degassing. A: Convection in conduit. B: Foam accumulation.

Using this approach, I analyze the magmatic system beneath Masaya volcano. At Masaya, we have abundant measurements and data on gas fluxes but few data on the dimensions of the conduit or on the flux of magma to the shallow reservoir, because no magma is erupted. To simulate relatively dry and wet magmas, the ascending magma in the conduit contains 0.5 and 3 wt% dissolved water, with corresponding densities of 2679 and 2574 kg m^{-3} and Poiseuille numbers of 0.04 and 0.064, respectively, as shown by Stevenson and Blake (1998). The descending magma is fully degassed with a density of 2700 kg m^{-3} and a viscosity of 630 Pa s estimated from Khitarov et al. (1976). The R^* value is 0.6 based on experiments by Stevenson and Blake (1998), and I neglect crystals and bubbles in the magmas for simplicity. Figure 2 shows the magma flow rate as a function of the conduit radius for ascending magma with 0.5 and 3 wt% H_2O. The magma flow rate is extremely sensitive to small changes in conduit radius, varying by more than seven orders of magnitude for conduit radii of only 1–50 m. This large variation poses a problem, because there are few independent constraints on either parameter.

To resolve this problem, I estimate the magma flux rate using the petrologic approach of sulfur degassing from Williams-Jones et al. (2001):

$$Q_m = 500{,}000\, Q_{SO_2}/\Delta S, \qquad (2)$$

where Q_m is the magma flux in kg s^{-1}, Q_{SO_2} is the sulfur dioxide flux in kg s^{-1}, ΔS is the amount of sulfur degassed from the melt (in ppm), and the constant of 500,000 is a conversion factor. To calculate magma fluxes, I vary both the SO_2 flux and the amount of sulfur degassed from the melt. The SO_2 flux at Masaya is well delimited, varying from 5 kg s^{-1} to 21 kg s^{-1} between 1993 and today (Duffell et al., 2003). The amount of sulfur degassed from the melt is less well defined, but values range from 240 ppm to 1000 ppm, which is a reasonable upper limit (Stoiber and Williams, 1986). Using these values, magma fluxes range from 2500–44,000 kg s^{-1} (Fig. 3), with most values clustering between 10,000–44,000 kg s^{-1}. Conduit dimensions (Fig. 2) beneath Masaya can be estimated by using these relatively well defined magma fluxes obtained by the sulfur degassing calculations. For initial magma water contents of 0.5 wt%, the conduit radius ranges from 2.8 m to 5.8 m, averaging 4.3 m, while for 3 wt% H_2O, the radius is 1.6–3.3 m, averaging 2.5 m. Thus, a reasonable conduit radius under these conditions is 2–6 m.

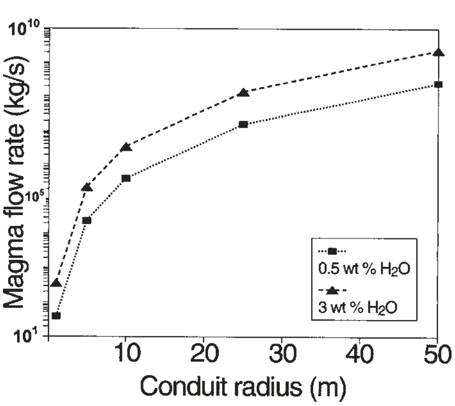

Figure 2. Plot of magma flux rate as function of conduit radius for magmas with initial 0.5 and 3.0 wt% dissolved H_2O.

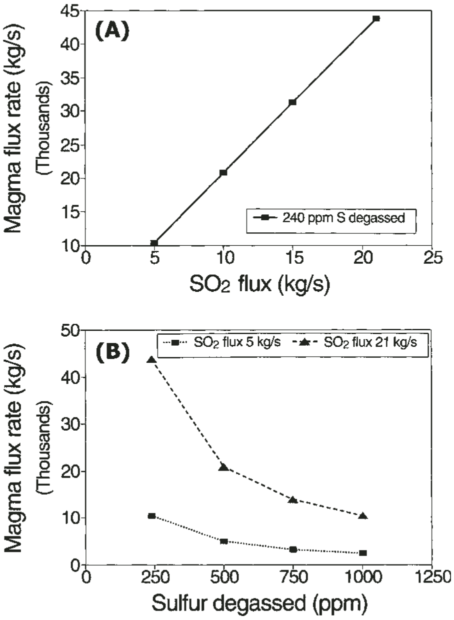

Figure 3. Magma flux rates calculated using petrologic and correlation spectrometer (COSPEC) data. A: Magma fluxes based on observed SO_2 fluxes during 1993–2006 and degassing of 240 ppm of sulfur from melt. B: Magma fluxes as function of different amounts of sulfur degassed from melt, using minimum and maximum SO_2 fluxes during 1993–2006.

These values can be verified at Masaya. Using the magma fluxes and conduit dimensions calculated here, a theoretical value for water degassing is ~10^1–10^3 kg s^{-1}. These amounts compare favorably with field measurements and estimates ranging from a low of 2.3 × 10^1 kg s^{-1} to a high of 4.2 × 10^2 kg s^{-1} (Rymer et al., 1998; Burton et al., 2000). In summary, degassing rates based on the conduit convection model are consistent with those observed at Masaya.

Foam Accumulation

Recent microgravity work at Masaya has shown that the summit region has undergone overall declines in microgravity since 1993, interpreted as increasingly vesicular magma present at shallow depths beneath the volcano as a result of degassing processes (Rymer et al., 1998; Williams-Jones et al., 2003). These microgravity decreases cover much of the summit region, implying that the degassing is not simply taking place through a narrow conduit. Instead, the development and accumulation of vesicular magma appear to be occurring at the top of a shallow magma reservoir beneath the summit. It is thus appropriate to examine degassing behavior at Masaya by applying the foam model of Jaupart and Vergniolle (1989), which defines a dimensionless measure of foam stability, N_1:

$$N_1 = \frac{\left(3Q_g\mu_f/\pi\varepsilon^2\rho_1 g\right)^{1/4}}{\left(4\sigma/\varepsilon\rho_1 g d\right)}, \qquad (3)$$

where Q_g is the gas flux into the foam, μ_f is the viscosity of the foam, ε is the volume fraction of gas in the foam, ρ_1 is the density of degassed liquid, σ is the coefficient of surface tension, and d is the bubble diameter. If $N_1 > 1$, the foam is unstable, leading to periodic eruptive activity such as observed at Stromboli and Pu'u O'o in Hawaii (Jaupart and Vergniolle, 1988). If $N_1 < 1$, the foam at the top of the magma reservoir is stable, the gas input from below being balanced by gas flowing out of the foam layer into a conduit connected to the surface. This is the case at Masaya. In general, foams are stabilized by lower gas fluxes, smaller gas bubbles, reduced magma viscosity, and higher magma surface tension.

Using a range of bubble sizes, I adjust gas fluxes into the foam to maintain $N_1 \leq 1$. If gas flux into the foam equals that at the surface, the thickness h_c of the stable foam and associated bubble sizes can be calculated (Fig. 4), where $h_c = 2\sigma/\varepsilon\rho_1 gr$ (Jaupart and Vergniolle, 1988), r is the bubble radius, $\rho_1 = 2700$ kg m^{-3}, $\sigma = 0.35$ N m^{-1}, $\mu_f = \mu_1(1-\varepsilon)^{-2.5}$, and μ_1 is the liquid viscosity (630 Pa s). The critical foam thickness is the thickness to which the foam at the top of the reservoir can grow without becoming unstable, i.e., causing the volcano to erupt. At degassing rates of 10^1–10^3 kg s^{-1}, the stable foam can grow to ~1–3 m thick, consisting of small bubbles whose average diameters range between 20 and 60 μm.

Bubbly Magma

The presence of bubbles in a magma modifies its rheology and dynamics. The conduit modeling assumes that bubbles are released principally at the top of the conduit. If bubbles are released and retained at an earlier stage within the conduit by rising decompressing magma, the bubbles will increase the buoyancy of the magma and the density difference between rising and sinking magma, resulting in enhanced magma upflow rates. To maintain volumetric flow rates as calculated here necessitates a slightly smaller conduit radius.

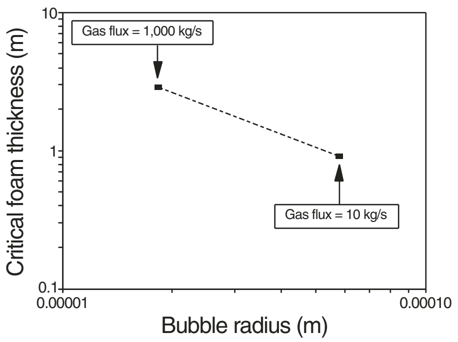

Figure 4. Critical foam thickness and mass flux of gas as function of average bubble size in upper foam layer.

Phillips and Woods (2001) showed that the accumulation of foam within a reservoir may occur due to the rise of isolated bubbles, as originally envisioned by Jaupart and Vergniolle (1989), or by a plume of bubbles or bubbly magma that is discharged through a crack or dike at the base of the reservoir. In this case, the bubbly plume mixes with magma resident in the reservoir. This entrainment process can slow the development of the foam layer at the top of the reservoir.

STRUCTURE AND MAGMA DYNAMICS BENEATH MASAYA

Conduit and Reservoir Connection

The conduit convection and foam models alone are insufficient mechanisms to explain gas release at Masaya. A degassing pipe is not consistent with the microgravity decreases observed for the summit region. A shallow reservoir containing an upper foam layer does not explain the source of gas from deeper levels. I propose that the conduit system and the shallow reservoir are coupled, providing a viable explanation for unrest and degassing at Masaya (Fig. 5). The conduit supplies fresh, low-density, gas-rich magma to the shallow reservoir. As the gas-rich magma exits the conduit and enters the reservoir, the magma exsolves bubbles, which rise and accumulate in a stable foam layer at the top of the reservoir. The rise of bubbles into the foam layer is balanced by gas that is released from the top of the foam layer and transferred through the open vent at the bottom of Santiago crater. In this manner, the magmatic system of the volcano is kept in a steady-state configuration.

The presence of a shallow magma reservoir system beneath Masaya may be structurally controlled. The pit craters at Masaya and many other volcanoes are the result of collapse processes akin to caldera formation, where support is removed from the roof of a magma chamber by either eruption or drainage of magma, causing collapse of the roof (Roche et al., 2001; Geshi et al., 2002). While collapse commonly takes place at the surface in the form of pit craters, subterranean collapse also may occur. In this case, subsurface cavities develop, which can then be used to store magma at shallow levels. Such voids, therefore, are logical candidates for shallow subsurface magma reservoirs. At Masaya, the extent of this shallow reservoir is uncertain, but microgravity data collected during the past decade suggest that it could be of larger extent than the Santiago pit crater, which has a diameter of ~1 km. Rymer et al. (1998) and Williams-Jones et al. (2003) showed that the gravity decreases are focused around the pit crater of Nindirí, suggesting that the shallow magma reservoir is centered under Nindirí and extends beneath the Santiago and possibly San Pedro pit craters. Alternatively, the gravity anomalies may simply reflect air-filled cavities and caves, without a more extensive shallow magma system.

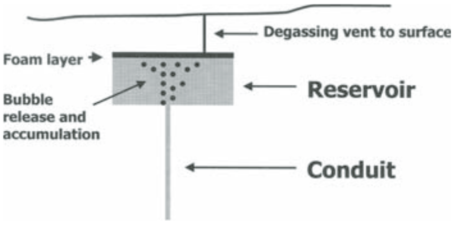

Figure 5. Model of gas flow and accumulation beneath Masaya. Steady-state situation exists between gas flow from conduit into foam, and from foam upward into degassing vent at surface.

The configuration of the reservoir's roof is also an important consideration. In Jaupart and Vergniolle's (1989) model, the foam layer accumulates beneath a flat roof. This may be the case at Masaya, because many of the lavas within the pit craters are flat lying due to ponding in lava lakes. If subsurface collapse that created the cavity was symmetric and piston style, the roof of the reservoir may be flat. If subsidence occurred in a piecemeal or asymmetric fashion, the roof may be irregular, allowing foam to accumulate to different thicknesses depending on the internal topography of the roof. Such a configuration is consistent with small differences in microgravity decreases observed among individual localities around the various pit craters (Rymer et al., 1998; Williams-Jones et al., 2003).

Toward More Complex Magma Plumbing

The model proposed here is simple, whereby the overlying reservoir and underlying conduit are coupled, with upflow of gas-rich magma and return flow of degassed magma occurring within the same conduit. Other configurations can be envisaged. Beneath the shallow reservoir, the rocks through which magma flows may have variable permeability. These rocks may be fractured and broken into blocks as a result of subsidence events (Roche et al., 2001; Geshi et al., 2002). Individual blocks consisting of lavas and pyroclastics have comparatively low permeabilities, while fault and fracture surfaces between blocks will have high permeabilities. Such structural pathways promote magma transport and degassing, as observed at Miyakejima volcano in 2000 (Geshi et al., 2002). The rate of upward magma flow will be influenced by this permeability and the magma's buoyancy, while return flow of magma will be determined by the permeability and the magma's density. This structural configuration raises the interesting possibility that the locus of magma replenishment on the floor of the shallow reservoir is distinct from the locus of return flow of magma. Further work is needed to elucidate this type of magma plumbing.

DESTABILIZATION OF A QUIESCENT VOLCANO

What factors might destabilize the magmatic system beneath Masaya, potentially leading to eruptive activity? If the foam layer is destabilized, it can erupt. Since 1993, small accumulations of Pele's hair have been noted at times around Santiago, suggesting that trace amounts of the upper foam layer are ejected when small instabilities in the system occur. In terms of the dimensionless measure of foam stability N_1, three parameters are critical: gas flux, bubble size, and surface tension.

The link between the conduit and the reservoir is important. If the reservoir overlies the conduit, then changes to the conduit's dimensions or magma flow therein may influence and potentially destabilize the foam layer at the top of the reservoir, causing it to be erupted. Deeper magma movements or structural adjustments due to earthquakes could cause the conduit to widen, resulting in enhanced conduit convection and increased magma and gas fluxes. An increased flux of gas into the foam layer would increase the N_1 parameter, potentially destabilizing the foam layer. Enhanced degassing could be accompanied by exsolution and rise of comparatively large bubbles, which would contribute to destabilizing the small bubble sizes present in the foam layer. Ground shaking by earthquakes also could release bubbles from the walls of the magma reservoir (Linde et al., 1994), causing them to rise and destabilize the foam. Volatile-rich magma will have reduced surface tension (Khitarov et al., 1979; Bagdassarov et al., 2000). If the surface tension contrast ($\Delta\sigma$) is large between replenishing volatile-rich magma and resident volatile-poor magma, the foam can be readily destabilized by this mechanism as well.

In summary, a large-scale replenishment of the upper conduit-reservoir system by volatile-rich magma provides the means to destabilize foam and trigger an eruption. The enhanced flux and gas-rich nature of the replenishment together provide a viable mechanism to disrupt quiescently degassing volcanoes such as Masaya. It is difficult to judge how fast such a process can occur. Basaltic magma at volcanic arcs is hot, gas rich, and mobile, so the time scales for destabilization can be short in many situations. Other mechanisms such as groundwater interaction also could trigger explosive eruptions.

CONCLUSIONS

What causes a degassing crisis at Masaya to end? The answer may be both in structural adjustments to the system and changes to the deep magma supply. Partial closing of the conduit by crystallization will limit the supply of magma to the overlying reservoir. The supply of magma from the deep system also may change, either by less magma fluxing through the system and/or lower volatile contents. Ultimately, these factors will reduce the foam layer and cause reservoir magma to stagnate.

Quiescent volcanoes are notable for their extended periods of restlessness and lack of eruptive activity. The ideas presented in this paper may aid volcanologists to develop a more sophisticated appreciation of these enigmatic volcanoes. The open nature of the systems, both in terms of magma dynamics and the pathways through which magma and gas flow, appears to be an essential component to their behavior. Understanding the factors that stabilize and destabilize these volcanic systems will allow us to forecast and predict their activity with greater confidence.

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REFERENCES CITED

Bagdassarov, N., Dorfman, A., and Dingwell, D.B., 2000, Effect of alkalis, phosphorus, and water on the surface tension of haplogranite melt: American Mineralogist, v. 85, p. 33–40.

Burton, M.R., Oppenheimer, O., Horrocks, L.A., and Francis, P.W., 2000, Remote sensing of CO_2 and H_2O emission rates from Masaya volcano: Geology, v. 28, p. 915–918, doi: 10.1130/0091–7613(2000)028<0915:RSOCAH>2.3.CO;2.

Delmelle, P., Baxter, P., Beaulieu, A., Burton, M., Francis, P., Garcia-Alvarez, J., Horrocks, L., Navarro, M., Oppenheimer, C., Rothery, D., Rymer, H., St-Amand, K., Stix, J., Strauch, W., and Williams-Jones, G., 1999, Origin, effects of Masaya volcano's continued unrest probed in Nicaragua: Eos (Transactions, American Geophysical Union), v. 80, p. 575, 579, 581.

Duffell, H.J., McGonigle, A.J.S., Burton, M.R., Oppenheimer, C., Pyle, D.M., and Galle, B., 2003, Changes in gas composition prior to a minor explosive eruption at Masaya volcano, Nicaragua: Journal of Volcanology and Geothermal Research, v. 126, p. 327–339, doi: 10.1016/S0377–0273(03)00156–2.

Geshi, N., Shimano, T., Chiba, T., and Nakada, S., 2002, Caldera collapse during the 2000 eruption of Miyakejima Volcano, Japan: Bulletin of Volcanology, v. 64, p. 55–68, doi: 10.1007/s00445–001–0184-z.

Jaupart, C., and Vergniolle, S., 1988, Laboratory models of Hawaiian and Strombolian eruptions: Nature, v. 331, p. 58–60, doi: 10.1038/331058a0.

Jaupart, C., and Vergniolle, S., 1989, The generation and collapse of a foam layer at the roof of a basaltic magma chamber: Journal of Fluid Mechanics, v. 203, p. 347–380, doi: 10.1017/S0022112089001497.

Kazahaya, K., Shinohara, H., and Saito, G., 1994, Excess degassing of Izu-Oshima volcano: Magma convection in a conduit: Bulletin of Volcanology, v. 56, p. 207–216, doi: 10.1007/s004450050029.

Khitarov, N.I., Lebedev, Y.B., Slutskin, A.B., Dorfman, A.M., Soldatov, I.A., and Revin, N.I., 1976, The pressure dependence of the viscosity of basalt melts: Geochemistry International, v. 13, no. 5, p. 126–133.

Khitarov, N.I., Lebedev, Y.B., Dorfman, A.M., and Bagdasarov, N.S., 1979, Effects of temperature, pressure, and volatiles on the surface tension of molten basalt: Geochemistry International, v. 16, no. 5, p. 78–86.

Linde, A.T., Bilham, R.G., Sacks, I.S., Johnston, M.J.S., and Hill, D.P., 1994, Increased pressure from rising bubbles as a mechanism for remotely triggered seismicity: Nature, v. 371, p. 408–410, doi: 10.1038/371408a0.

McBirney, A.R., 1956, The Nicaraguan volcano Masaya and its caldera: Eos (Transactions, American Geophysical Union), v. 37, p. 83–96.

Metaxian, J.-P., Lesage, P., and Dorel, J., 1997, Permanent tremor of Masaya volcano, Nicaragua: Wave field analysis and source location: Journal of Geophysical Research, v. 102, p. 22,529–22,545, doi: 10.1029/97JB01141.

Phillips, J.C., and Woods, A.W., 2001, Bubble plumes generated during recharge of basaltic magma reservoirs: Earth and Planetary Science Letters, v. 186, p. 297–309, doi: 10.1016/S0012–821X(01)00221–7.

Roche, O., Van Wyk de Vries, B., and Druitt, T.H., 2001, Sub-surface structures and collapse mechanisms of summit pit craters: Journal of Volcanology and Geothermal Research, v. 105, p. 1–18, doi: 10.1016/S0377–0273(00)00248–1.

Rymer, H., van Wyk de Vries, B., Stix, J., and Williams-Jones, G., 1998, Pit crater structure and processes governing persistent activity at Masaya volcano, Nicaragua: Bulletin of Volcanology, v. 59, p. 345–355, doi: 10.1007/s004450050196.

Stevenson, D.S., and Blake, S., 1998, Modelling the dynamics and thermodynamics of volcanic degassing: Bulletin of Volcanology, v. 60, p. 307–317, doi: 10.1007/s004450050234.

Stoiber, R.E., and Williams, S.N., 1986, Sulfur and halogen gases at Masaya caldera complex, Nicaragua: Total flux and variations with time: Journal of Geophysical Research, v. 91, p. 12,215–12,231.

Williams, S.N., 1983, Plinian airfall deposits of basaltic composition: Geology, v. 11, p. 211–214, doi: 10.1130/0091–7613(1983)11<211:PADOBC>2.0.CO;2.

Williams-Jones, G., Stix, J., Heiligmann, M., Barquero, J., Fernandez, E., and Gonzales, E.D., 2001, A model of degassing and seismicity at Arenal volcano, Costa Rica: Journal of Volcanology and Geothermal Research, v. 108, p. 121–139.

Williams-Jones, G., Rymer, H., and Rothery, D.A., 2003, Gravity changes and passive SO_2 degassing at the Masaya caldera complex, Nicaragua: Journal of Volcanology and Geothermal Research, v. 123, p. 137–160, doi: 10.1016/S0377–0273(03)00033–7.

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