Unique strainmeter observations of Vulcanian explosions, Soufrière Hills Volcano, Montserrat, July 2003

B. Voight,[1] D. Hidayat,[1] S. Sacks,[2] A. Linde,[2] L. Chardot,[3] A. Clarke,[4] D. Elsworth,[1] R. Foroozan,[1] P. Malin,[5] G. Mattioli,[6] N. McWhorter,[2] E. Shalev,[5] R. S. J. Sparks,[7] C. Widiwijayanti,[1] and S. R. Young[1]

Received 15 January 2010; revised 14 April 2010; accepted 19 April 2010; published 2 June 2010.

[1] Five Vulcanian explosions were triggered by collapse of the Soufrière Hills Volcano lava dome in 2003. We report strainmeter data for three explosions, characterized by four stages: a short transition between the onset of disturbance and a pronounced change in strain; a quasi-linear ramp accounting for the majority of strain change; a more gradual continued decline of strain to a minimum value; and a strain recovery phase lasting hours. Remarkable ~800 s barometric gravity waves propagated at ~30 m s^{-1}. Eruption volumes estimated from plume height and strain data are 0.32–0.42 × 10^6, 0.26–0.49 × 10^6, and 0.81–0.84 × 10^6 m^3, for Explosions 3–5 respectively, consistent with quasi-cylindrical conduit drawdown <2 km. The duration of vigorous explosion is given by the strain signature, indicating mass fluxes of order 10^7 kg s^{-1}. Conduit pressures released reflect static weight of porous gas-charged magma, and exsolution-generated overpressures of order 10 MPa.
Citation: Voight, B., et al. (2010), Unique strainmeter observations of Vulcanian explosions, Soufrière Hills Volcano, Montserrat, July 2003, *Geophys. Res. Lett.*, *37*, L00E18, doi:10.1029/2010GL042551.

1. Introduction

[2] Most analytical and numerical eruption models address Plinian-type eruptions where vent conditions are quasi-steady. Only a relatively few studies have explored the physics of short-duration Vulcanian explosions with unsteady conditions, although such events are common at andesitic volcanoes such as Sakajurima and Soufrière Hills Volcano [*Ishihara*, 1985; *Clarke et al.*, 2002a, 2002b; *Druitt et al.*, 2002; *Mason et al.*, 2006] (see Text S1 in the auxiliary material).[8] The interpretations of such events have been diverse, and there is a need for observations that constrain alternative physical models. Here we use high-precision volumetric strain data and broadband seismic data to constrain the source mechanism, explosion physics and transport dynamics of three well-documented Vulcanian eruptions in 2003 at Soufrière Hills Volcano (SHV), Montserrat, in 2003 [*Herd et al.*, 2005].

2. July 2003 Explosions on Montserrat

[3] On 12–13 July 2003, after eight years of eruption at SHV, a major collapse of the lava dome occurred. The collapse followed two years of sustained dome growth and lasted about 18 h. The loss of dome height over the vent was ~500 m, and after collapse the vent was surrounded by crater walls 140–200 m high [*Herd et al.*, 2005]. Decompression of magma in the conduit of ~12 MPa caused by loss of overlying dome material triggered a sequence of several Vulcanian explosions. Explosion 1 occurred during the peak of collapse at 03:35 13 July (all times cited are UT), and caused an island-wide fallout of dense juvenile clasts and release of a SO_2-rich gas plume [*Herd et al.*, 2005]. Explosion 2 occurred at 05:08 during waning stages of the collapse, when the level of volcano seismicity remained high and was dominated by multi-frequency signals from collapse and pyroclastic flow activity [*Herd et al.*, 2005].

[4] Three further Vulcanian explosions followed at progressively increasing intervals, after seismicity had returned to background. Explosion 3 occurred at 13:09 13 July, Explosion 4 at 05:15 on 14 July, and Explosion 5 at 05:29 on 15 July [*Herd et al.*, 2005]. The explosions were broadly similar in scale, generated lithic and pumice clasts, and were each followed by 2–4 h of ash venting and associated low amplitude tremor. No pyroclastic density currents were detected. Shock waves were reported for several explosions. Washington VAAC reported plumes for Explosion 3 at about 12.2 km altitude, for Explosion 4 about 11.0 km, and for Explosion 5 about 14.3 km high. Strong gas emissions occurred on 16 and 17 July [*Edmonds and Herd*, 2007], and a small lava dome was extruded 21–28 July, terminating the eruptive episode.

[5] As in 1997 [*Druitt et al.*, 2002], each explosion had an initial, high intensity phase that included the main explosion and peak magma discharge rates, formation of a buoyant plume, and plume ascent to a level of neutral buoyancy in the atmosphere, and a second, waning phase of comparatively negligible discharge, lasting a few hours and characterized by weak venting of ash and gas that generated a low, gradually diminishing bent-over plume. Given the short period of intense discharge compared to explosion plume rise times,

[1] Earth and Mineral Sciences, Pennsylvania State University, University Park, Pennsylvania, USA.
[2] Department of Terrestrial Magnetism, Carnegie Institute of Washington, Washington, D.C., USA.
[3] Montserrat Volcano Observatory, Flemming, Montserrat, West Indies.
[4] School of Earth and Space Exploration, Arizona State University, Tempe, Arizona, USA.
[5] Institute of Earth Science and Engineering, University of Auckland, Auckland, New Zealand.
[6] Department of Geosciences, University of Arkansas, Fayetteville, Arkansas, USA.
[7] Department of Earth Sciences, Bristol University, Bristol, UK.

0094-8276/10/2010GL042551

[8] Auxiliary materials are available in the HTML. doi:10:1029/2010GL042551.

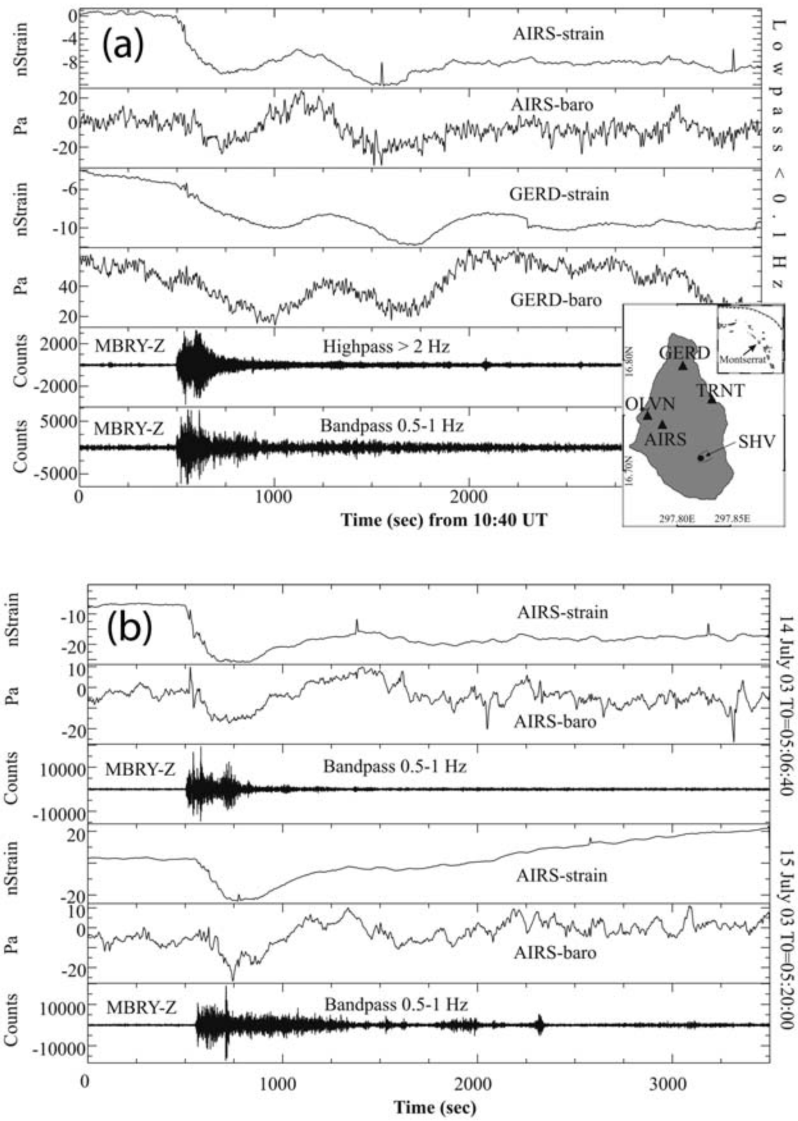

Figure 1. (a) Strain and barometer records for 13 July 2003 explosion, at AIRS and GERD strainmeter sites. Inset map shows site locations. Data are lowpass filtered at 0.1 Hz. Note strain ramps corresponding to explosion, followed by very-long-period waves shown on both barometer and strain records. Lower panels show highpass and bandpass filtered seismic data recorded at station MBRY, 2 km ESE of the vent. (b) Strain and barometer records for 14 and 15 July 2003 explosions at AIRS strainmeter site. Data are lowpass filtered at 0.1 Hz. Note strain ramps corresponding to explosions, followed by very-long-period waves shown on both barometer and strain records. Data are compared to bandpass filtered seismic data from station MBRY.

the plumes are modeled as the instantaneous injection of a buoyant thermal of a given initial mass [*Morton et al.*, 1956; *Woods and Kienle*, 1994; *Druitt et al.*, 2002]. The height H_t (m) of a thermal is given by $H_t = 1.89\ Q^{0.25} + H_v$, where H_v is vent height and Q (J) is the excess thermal mass of injection, taken as the product of $\Phi\ M\ C\ T$, where Φ is fraction of solids releasing heat to the plume, M (kg) is mass of solids in plume, C (J kg^{-1} K^{-1}) is specific heat of solids, and T (K) is initial temperature difference between eruption mixture and ambient air. We assume $\Phi = 0.8$, $C = 1100$ J kg^{-1} K^{-1}, $T = 1123$ K, and $H_v = 1000$ m. We note further that in a humid tropical environment, entrainment of humid air and tropospheric instability can make plume rise exceed the height calculated from the general formula for a given erupted mass, typically by 5–20% [*Sparks et al.*, 1997; *Tupper et al.*, 2009]. To adjust for this observation, we follow the approach of *Druitt et al.* [2002] and reduce H_t by 10% prior to calculation.

[6] Given the adjusted plume heights cited previously, the equation yields erupted volumes of 0.42, 0.26 and 0.84 Mm^3 dense rock equivalent (DRE) for Explosions 3–5, respectively, using a reference density of 2600 kg m^{-3}. These volumes are approximate, because in addition to modeling issues, the VAAC height estimates for SHV have an uncertainty of about 1 km that generates a variation range of about 0.62–1.5 times the central volume estimates.

3. Instrumentation Network and Data Description

[7] We present observations of strain and air pressure associated with these explosions, and compare these data with broadband seismic observations (Figures 1 and 2). Data were

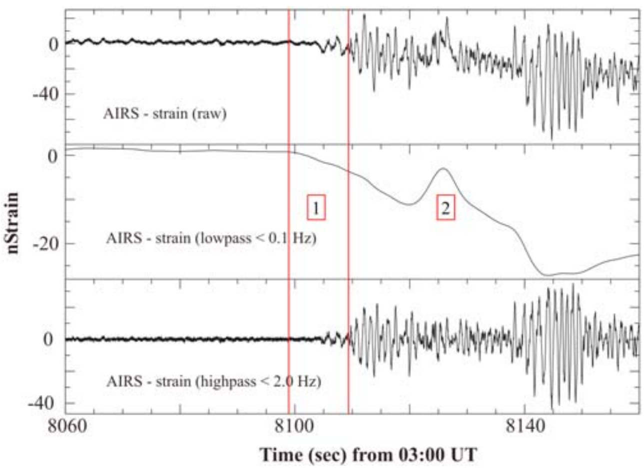

Figure 2. Detail of strain records at AIRS site at onset of 14 July explosion. (top) Raw strain record; (middle) 0.1 Hz lowpass filtered strain; and (bottom) 2.0 Hz highpass filtered strain. The data suggest strain onset about 10 s prior to strong explosion-induced seismicity marking Stage 2 onset.

obtained from the *CALIPSO* network [*Mattioli et al.*, 2004], which investigates the dynamics of the Soufrière Hills Volcano (SHV) magmatic system by specialized instruments in four ~200-m-deep boreholes (Figure 1, inset). Three stations (AIRS, TRNT, GERD) were operating for the 13 July explosion, and two (AIRS, TRNT) for 14 and 15 July. Each borehole site, at distances of 5.1 to 9.6 km from the crater (Figure 1a), includes a Sacks-Evertson volumetric strainmeter, capable of recording strain changes of about 0.1 nanostrain in periods from DC to about 50 Hz [*Linde and Sacks,* 1995]. Each strainmeter recorded data at 50 samples per second (sps) with 24 bit ADC's, and a barograph recorded data at 50 sps with 20 bit resolution. We compare these data with vertical-component broadband seismic data from the surface station MBRY (Figure 1).

[8] The strainmeter responds to ground deformation due to changes in atmospheric air pressure, in addition to underground sources, and these effects were removed in order to determine the strain changes from the explosions [*Nakao et al.*, 1989; *Linde et al.*, 2010]. At SHV the explosions typically generated a complicated but intriguing airwave on barographs. The strain shows a rapid decrease in conduit pressure, and the barograph indicates a decrease in air pressure after onset of the eruptive process (Figure 1). For July 2003 events there is no sharp initial positive pressure pulse as observed in 2008–09 explosions, probably because of blocking of the acoustic pulse by crater walls. The remarkable long period ~800 s wave train from these early explosions propagated at about 30 m s^{-1} as gravity waves for distances >5 km, possibly generated by the rising plume [*Kanamori et al.*, 1994; *Lighthill*, 1967; *Voisin*, 1994]. Some shock wave spikes appear in both barometer and strain records during the evacuation process.

[9] The data from AIRS and GERD display good signal/noise ratio (Figure 1). TRNT is located near the coastline and data (not shown) are relatively noisy, but the site is nevertheless important to constrain polarity. For these events the strainmeters at all operating sites expanded (millivolts go negative). AIRS is well beyond the nodal distance for a relatively shallow axisymmetrical source [*Linde and Sacks*, 1995] (see Text S1), so the source itself has to contract which indicates a decrease in source pressure. The data at AIRS, TRNT and GERD have the same sign and can be satisfied by a relatively shallow (cylindrical) conduit source. The explosion patterns (Figure 1) comprise several stages: 1) a short transition between the onset of disturbance and a pronounced change in strain; 2) a quasi-linear ramp accounting for the majority of strain change; 3) a continued and more gradual decline of strain to a minimum value; and 4) a strain recovery phase lasting hours.

[10] The first of these, Stage 1, is a brief precursor period involving small strain and small seismicity that gradationally precedes the strong explosive release recorded by the onset of large amplitude seismic energy (Figure 2). This precursor period is interpreted to represent initial progressive fracturing as the pressure beneath the conduit plug reaches a failure threshold. In laboratory experiments on samples subjected to a sudden pressure drop [*Spieler et al.*, 2004] tensile failure initiates a fragmentation wave that propagates through the sample. We suggest that Stage 1 occurs when the pressurized gas, accumulating in glassy foam or as a discrete pocket under a strong plug, reaches the tensile strength of the cap [*Alidibirov and Dingwell*, 1996; *Spieler et al.*, 2004].

[11] Stage 2 denotes a rapid source contraction that we correlate with fragmentation and vigorous explosive evacuation of magma in the conduit. The stage 2 pattern in the raw time series is interrupted by the superposition of very-long-period waves in both barometer and strainmeter signals, but we minimize this influence by applying corrections for air pressure. We propose that most of the ejecta is expelled during the nearly linear strain ramp, and that the subsequent gradual further decline of strain to a minimum value in Stage 3 reflects superposed effects of discharge of ash and degassing at the top of the magma column, and rise of viscous magma in the conduit. Stage 4 continues beyond the strain-minimum value and represents a gradual several hour strain recovery that reflects magma ascent in the conduit and exsolution.

4. Strain Analysis and Other Results

[12] The conduit dimensions are constrained by spine dimensions, magma ascent rates and volume extrusion rates [*Voight et al.*, 1999; *Melnik and Sparks*, 2002], and we assume for analysis a conduit diameter of 30 m. A transition from a cylindrical conduit to an underlying dike has been proposed for SHV [*Mattioli et al.*, 1998; *Costa et al.*, 2007; *Hautmann et al.*, 2009; *Linde et al.*, 2010], and for analysis we assume a nominal transition depth roughly 2 km beneath the vent at ~700 m a.s.l. There is no suggestion in our strain data that a dike was depressurized, but this fact provides a useful constraint.

[13] The conduit drawdown depth is estimated from erupted DRE volume adjusted for porosity, divided by conduit area. For simplicity we assume the porosity is constant with depth, and we use a average pumice density of ~1500 kg m^{-3}, based on data from both the 1997 and 2003 explosions [*Clarke et al.*, 2002b; *Edmonds et al.*, 2006]. We first reconsider eruption volume. The plume heights suggest erupted volumes of 0.42, 0.26 and 0.84 Mm^3 DRE for Explosions 3–5, implying for a 30-m conduit, drawdowns of 1.0, 0.65 and 2.1 km, respectively. The 2.1 km estimate matches within standard error our proposed nominal depth to top of dike.

[14] We also propose another volume-estimation method, based on strain data. We assume eruption volume is proportional to the relative sizes of strain-ramps, with strain-ramp ratios of 0.4, 0.6, and 1.0 for Explosions 3–5, respectively. The procedure is based on the interpretation that the strain-ramps correspond to vigorous evacuations of the conduit, and that other variables such as conduit overpressure are similar. The maximum erupted volume, conditional on a dike top at 2.0 km and 30-m conduit, is 0.81 Mm^3 DRE, applied to Explosion 5. Multiplying by strain ratio yields 0.32 and 0.49 Mm^3 DRE, for Explosions 3 and 4. Thus the corresponding drawdowns for the three explosions are 0.79, 1.2, and 2.0 km, respectively. We accept the fact of volume uncertainty as previously expressed, and include its consideration in analysis.

[15] Next, we use our strain-ramp amplitudes to estimate average conduit wall pressures released in the explosions. We use an axisymmetrical elastic finite element model (13,000 elements with quadratic shape functions) (see auxiliary material) and a cylindrical conduit of 30 m diameter, applying a magmastatic load for a nominal average density of 1500 kg m^{-3} and a superposed uniform overpressure (pressure exceeding magmastatic), over the lengths of conduit specified above. We have developed several sub-models, e.g. using homogeneous media for one case, and using inhomogeneous media in another, based on the average one-dimensional seismic velocity model (see auxiliary material), with velocity/modulus increasing with depth, obtained in the *SEA-CALIPSO* tomography experiment [*Shalev et al.*, 2010]. We assume Young's modulus E = 5 GPa for the homogeneous case; this is the average local modulus value calculated for the strainmeter sites. Similar values have been justified for modeling work at other volcano sites [e.g., *Houlié and Montagner*, 2007; *Cayol and Cornet*, 1998], implying a fractured and/or hydrothermally altered rock mass. In all cases we assume a Poisson's ratio of 0.25. For inhomogeneous media, plume height-based volumes, and Explosions 3, 4, 5, the air pressure-corrected strain-ramps of Stage 2 (8, 12, 20 nanostrain) yield average overpressures of about 15, 50, and 25 MPa, respectively (rounded-off values). For the strain ratio-based volumes, the overpressures are about 25 MPa for the three cases.

[16] The general consistency of average pressures is probably more significant than results for any specific explosion. Further, this result is dependent on modulus: assuming E = 4 GPa reduces overpressures ~30%, and E = 3 GPa yields overpressures of order 10 MPa. The latter modulus value is considered typical for volcanic areas by *Lisowski* [2007] (see auxiliary material). For Explosion 3, we can also calculate the ratio of measured strain changes at AIRS and GERD sites (Figure 1) as about 4 (GERD data are not available for the other explosions). The corresponding modeled values are >5. This result might be explained by modulus anomalies recognized in the 3D tomography [*Shalev et al.*, 2010], not yet considered in our modeling.

[17] To estimate mass evacuation rates, we divide the erupted mass by the corresponding durations of the strain-ramp in stage 2, giving 1.2×10^7, 0.78×10^7, and 2.1×10^7 kg s^{-1} for Explosions 3–5, respectively, using plume height-based volumes. For the strain ratio-based volumes, the estimated average rates are 0.57×10^7, 0.91×10^7, 1.3×10^7 kg s^{-1}. The order of magnitude of average mass flux is 10^7 kg s^{-1}, similar to the estimate of *Druitt et al.* [2002] for the 1997 explosions. In comparison, model-based estimates of *Clarke et al.* [2002a, 2002b], *Melnik and Sparks* [2002] yielded peak discharges of 3–7 × 10^7 kg s^{-1}.

[18] The volumes erupted in these explosions are necessarily uncertain, with our estimated ranges for Explosions 3–5 being 0.32–0.42 × 10^6, 0.26–0.49 × 10^6, and 0.81–0.84 × 10^6 m^3, respectively. These values are similar to the average discharge of 0.3×10^6 m^3 for 88 explosions at SHV in 1997 [*Druitt et al.*, 2002], which occurred at intervals of 2.5–63 h, with a mean of 10 h. However in 2003, the intervals between successive explosions increased: 90 min, to 480 min, to 960 min, to 1450 min, between Explosions 1 and 5 [*Herd et al.*, 2005]. The volumes erupted in the first two explosions are unknown because they occurred during the dome collapse. For refilling the conduit after Explosions 3 and 4, the data suggest average conduit recharge fluxes of 5.6–7.3 m^3 s^{-1}, and 3.0–5.6 m^3 s^{-1}, respectively. These values are roughly consistent with the enhanced explosion potential noted for fluxes exceeding ~5 m^3 s^{-1} [*Sparks et al.*, 1998]. The 2003 explosions appear to be close to the end-member case of zero volatile mass transfer, where magma rising in a conduit acquires a porosity and overpressure structure that evolves to critical conditions to trigger an explosion [*Melnik and Sparks*, 2002; *de' Michieli Vituri et al.*, 2008]. But the decrease noted in recharge flux rates suggests a trend toward more effective magma degassing from the conduit that could explain the delay in meeting the explosion criterion, as well as the lack of significant explosions after 15 July.

5. Discussion

[19] Our calculations underscore the importance of overpressures in Vulcanian eruptions, and our modeling suggests average overpressures of about ten to possibly a few tens of MPa. These preliminary results are approximate because of (a) unavoidable eruptive volume uncertainty; (b) uncertainty regarding precise conduit geometry and possible variations with depth; (c) uncertain magma bulk density with depth [*de' Michieli Vitturi et al.*, 2008]; (d) a simplified pressure distribution in comparison to conduit flow models [*Melnik and Sparks*, 2002; *de' Michieli Vitturi et al.*, 2008], and (e) the fact that rock modulus and its distribution is only crudely understood and modeled. Nevertheless, the suggestion of pressurized gas-charged magma at high levels in the conduit immediately prior to the explosion supports the observations of exit velocities >100 m/s only a few seconds after Vulcanian explosions began [*Druitt et al.*, 2002; *Clarke et al.*, 2002a, 2002b]. In comparison, *Robertson et al.* [1998] reported a pressure of 27.5 MPa, based on modeling of ballistic clasts in the 17 September 1996 explosion. This result is similar to some of ours, although we consider our larger estimates as uncomfortably high and emphasize that a reduced modulus can yield overpressures of order 10 MPa Overpressures estimated by conduit models for magma properties equivalent to those for SHV do not typically exceed 15−20 MPa [*de' Michieli Vitturi et al.*, 2008; *Melnik and Sparks*, 2002].

[20] If the explosion threshold is conduit overpressure, then this should in principle be detectable by dilatometer strain. Inspection of the strain records over a three-day period encompassing the explosions shows no clear indications of overpressure buildup. The multiple-day records are, however, uncorrected for tides, ocean loading, or meteorological effects. On timescales of about 1 h we see no evident pressure

rises preceding the explosions. Over a period of about 80 min following the explosions, we note that the strain levels for 13 and 14 July had built to about two-thirds of the pre-explosion values, whereas for 15 July, the recorded strain had exceeded the pre-explosion value.

[21] **Acknowledgments.** This research was supported by NSF, CD and I&F Programs, and by NERC (UK). We thank particularly L. Johnson and R. Kelz at NSF. RSJS is supported by an ERC Advanced Grant. Additional support was provided by CALIPSO Consortium and Montserrat Volcano Observatory. B.V. is grateful to G. Swanson at NOAA, and we thank A. Neri and an anonymous reviewer.

References

Alidibirov, M. A., and D. B. Dingwell (1996), Magma fragmentation by rapid decompression, *Nature*, *380*, 146–148, doi:10.1038/380146a0.

Cayol, V., and F. H. Cornet (1998), Effects of topography on the deformation field of prominent volcanoes: application to Etna, *Geophys. Res. Lett.*, *25*, 1979–1982, doi:10.1029/98GL51512.

Clarke, A. B., et al. (2002a), Transient dynamics of vulcanian explosions and column collapse, *Nature*, *415*, 897–901, doi:10.1038/415897a.

Clarke, A. B., et al. (2002b), Computational modeling of the transient dynamics of the August 1997 Vulcanian explosions at Soufrière Hills Volcano, Montserrat, in The *Eruption of Soufrière Hills Volcano, from 1995 to 1999*, edited by T. H. Druitt and B. P. Kokelaar, *Mem. Geol. Soc. London*, 21, 319–348.

Costa, A., et al. (2007), Control of magma flow in dykes on cyclic lava dome extrusion, *Geophys. Res. Lett.*, *34*, L02303, doi:10.1029/2006GL027466.

de' Michieli Vitturi, M., A. B. Clarke, A. Neri, and B. Voight (2008), Effects of conduit geometry on magma ascent dynamics in dome-forming eruptions, *Earth Planet. Sci. Lett.*, *272*, 567–578, doi:10.1016/j.epsl.2008.05.025.

Druitt, T. H., et al. (2002), Episodes of cyclic Vulcanian explosive activity with fountain collapse at Soufrière Hills Volcano, Montserrat, in *The Eruption of Soufrière Hills Volcano, from 1995 to 1999*, edited by T. H. Druitt and B. P. Kokelaar, *Mem. Geol. Soc. London*, 21, 281–306.

Edmonds, M., and R. A. Herd (2007), A volcanic degassing event at the explosive-effusive transition, *Geophys. Res. Lett.*, *34*, L21310, doi:10.1029/2007GL031379.

Edmonds, M., R. A. Herd, and M. H. Strutt (2006), Tephra deposits associated with a large lava dome collapse, Soufrière Hills Volcano, Montserrat, 12–15 July 2003, *J. Volcanol. Geotherm. Res.*, *153*, 313–330, doi:10.1016/j.jvolgeores.2005.12.008.

Hautmann, S. J., et al. (2009), Modelling ground deformation caused by an oscillating overpressure in a dyke conduit at Soufrière Hills Volcano, Montserrat, *Tectonophysics*, *471*, 87–95, doi:10.1016/j.tecto.2008.10.021.

Herd, R., M. Edmonds, and V. Bass (2005), Catastrophic lava dome failure at Soufrière Hills Volcano, Montserrat, 12–13 July 2003, *J. Volcanol. Geotherm. Res.*, *148*(3–4), 234–252, doi:10.1016/j.jvolgeores.2005.05.003.

Houlié, N., and J. -P. Montagner (2007), Hidden dykes detected on ultra long period seismic signals at Piton de Fournaise volcano?, *Earth Planet. Sci. Lett.*, *261*, 1–8, doi:10.1016/j.epsl.2007.04.018.

Ishihara, K. (1985), Dynamical analysis of volcanic explosion, *J. Geodyn.*, *3*, 327–349, doi:10.1016/0264-3707(85)90041-9.

Kanamori, H., et al. (1994), Excitation of atmospheric oscillations by volcanic eruptions, *J. Geophys. Res.*, *99*(B11), 21,947–21,961, doi:10.1029/94JB01475.

Lighthill, M. J. (1967), On waves generated in dispersive systems by traveling forcing effects, with applications to the dynamics of rotating fluids, *J. Fluid Mech.*, *27*, 725–752, doi:10.1017/S0022112067002563.

Linde, A. T., and L. S. Sacks (1995), Continuous monitoring of volcanoes with borehole strainmeters, in *Mauna Loa Revealed: Structure, Composition, History, and Hazards*, *Geophys. Monogr. Ser.*, vol. 92, edited by J. M. Rhodes and John P. Lockwood, pp. 171–185, AGU, Washington, D. C.

Linde, A. T., et al. (2010), Vulcanian explosion at Soufrière Hills Volcano, Montserrat on March 2004 as revealed by strain data, *Geophys. Res. Lett.*, *37*, L00E07, doi:10.1029/2009GL041988.

Lisowski, M. (2007), Analytical volcano deformation source models, in *Volcano Deformation*, edited by D. Dzurisin, pp. 279–304, Springer, Berlin.

Mason, R. M., A. B. Starostin, O. Melnik, and R. Sparks (2006), From Vulcanian explosions to sustained explosive eruptions: The role of diffusive mass transfer in conduit flow dynamics, *J. Volcanol. Geotherm. Res.*, *153*, 148–165, doi:10.1016/j.jvolgeores.2005.08.011.

Mattioli, G., et al. (1998), GPS measurement of surface deformation around Soufrière Hills Volcano, Montserrat, from October 1995 to July 1996, *Geophys. Res. Lett.*, *25*(18), 3417–3420, doi:10.1029/98GL00931.

Mattioli, G., et al. (2004), Prototype PBO instrumentation of CALIPSO Project captures world-record lava dome collapse on Montserrat, *EOS Trans. AGU*, *85*(34), 317–325.

Melnik, O. E., and R. S. J. Sparks (2002), Modelling of conduit flow dynamics during explosive activity at Soufrière Hills Volcano, Montserrat, in *The Eruption of Soufrière Hills Volcano, from 1995 to 1999*, edited by T. H. Druitt and B. P. Kokelaar, *Mem. Geol. Soc. London*, 21, 307–317.

Morton, B., G. I. Taylor, and J. S. Turner (1956), Turbulent gravitational convection from maintained and instantaneous sources. *Proc. R. Soc. A*, *234*, 1–23.

Nakao, S., A. Linde, S. Miura, M. Mishina, and A. Takagi (1989), Nonlinear barometric response of borehole strainmeters and its interpretation, *J. Phys. Earth*, *37*, 357–383.

Robertson, R. E. A., et al. (1998), The explosive eruption of Soufrière Hills Volcano, Montserrat, September 17, 1996, *Geophys. Res. Lett.*, *25*, 3429–3432, doi:10.1029/98GL01442.

Shalev, E., et al. (2010), Three-dimensional seismic velocity tomography of Montserrat from the SEA-CALIPSO offshore/onshore experiment, *Geophys. Res. Lett.*, *37*, L00E17, doi:10.1029/2010GL042498.

Sparks, R. S. J., M. L. Bursik, S. N. Carey, J. S. Gilbert, L. S. Glaze, H. Sigurdsson, and A. W. Woods (1997), *Volcanic Plumes*, 574 pp., J. Wiley, New York.

Sparks, R. S. J., et al. (1998), Magma production and growth of the lava dome of the Soufrière Hills Volcano, Montserrat, West Indies, *Geophys. Res. Lett.*, *25*, 3421–3424, doi:10.1029/98GL00639.

Spieler, O., D. B. Dingwell, and M. Alidibirov (2004), Magma fragmentation speed: an experimental determination, *J. Volcanol. Geotherm. Res.*, *129*, 109–123, doi:10.1016/S0377-0273(03)00235-X.

Tupper, A., et al. (2009), Tall clouds from small eruptions: the sensitivity of eruption height and fine ash content to tropospheric instability, *Nat. Hazards*, *51*, 375–401, doi:10.1007/s11069-009-9433-9.

Voight, B., et al. (1999), Magma flow instability and cyclic activity at Soufrière Hills volcano, Montserrat, British West Indies, *Science*, *283*, 1138–1142, doi:10.1126/science.283.5405.1138.

Voisin, B. (1994), Internal wave generation in uniformly stratified fluids. Part 2. Moving point sources, *J. Fluid Mech.*, *261*, 333–374, doi:10.1017/S0022112094000364.

Woods, A. W., and J. Kienle (1994), The dynamics and thermodynamics of volcanic clouds: Theory and observations from the April 15 and April 21 1990 eruptions of Redoubt Volcano, Alaska, *J. Volcanol. Geotherm. Res.*, *62*, 273–299, doi:10.1016/0377-0273(94)90037-X.

L. Chardot, Montserrat Volcano Observatory, Flemming, Salem Montserrat, West Indies.

A. Clarke, School of Earth and Space Exploration, Arizona State University, Tempe, AZ, 85287 USA.

D. Elsworth, R. Foroozan, D. Hidayat, B. Voight, C. Widiwijayanti, and S. R. Young, Earth and Mineral Sciences, Pennsylvania State University, University Park, PA 16802, USA. (voight@ems.psu.edu)

P. Malin and E. Shalev, Institute of Earth Science and Engineering, University of Auckland, Auckland 1142, New Zealand. (p.malin@auckland.ac.nz; e.shalev@auckland.ac.nz)

G. Mattioli, Department of Geosciences, University of Arkansas, Fayetteville, AR 72701, USA.

N. McWhorter, S. Sacks, and A. Linde, Department of Terrestrial Magnetism, Carnegie Institute of Washington, Washington, DC 20015, USA.

R. S. J. Sparks, Department of Earth Sciences, Bristol University, Bristol, BS8 1RJ, UK.