Explosion dynamics from strainmeter and microbarometer observations, Soufrière Hills Volcano, Montserrat: 2008–2009

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[1] Vulcanian explosions with plumes to 12 km occurred at Soufrière Hills volcano (SHV) between July 2008 and January 2009. We report strainmeter and barometric data, featuring quasi-linear strain changes that correlate with explosive evacuation of the conduit at rates of ~0.9−2 × 10^7 kg s^{-1}. July and January explosion-generated strains were similar, ~20 nanostrain at ~5 km, and interpreted as contractions of a quasi-cylindrical conduit, with release of magmastatic pressure, and exsolution-generated overpressure of order 10 MPa. The 3 December 2008 event was distinctive with larger signals (~140–200 nanostrain at 5–6 km) indicating that a rapid pressurization preceded and triggered the explosion. Modeling suggests a dike with ENE trend, implying that feeder dikes at SHV had diverse attitudes at different times during the eruption. All explosions were associated with acoustic pulses and remarkable atmospheric gravity waves. **Citation:** Chardot, L., et al. (2010), Explosion dynamics from strainmeter and microbarometer observations, Soufrière Hills Volcano, Montserrat: 2008–2009, *Geophys. Res. Lett.*, *37*, L00E24, doi:10.1029/2010GL044661.

1. Introduction

[2] The fourth episode of eruptive activity at Soufrière Hills volcano (SHV) between 29 July 2008 and 3 Jan 2009 began and terminated with pumice-bearing explosions. The Vulcanian explosions in this episode were captured by the CALIPSO strainmeter/microbarometer array installed in 2002–03 to investigate the dynamics of the SHV magmatic system [*Mattioli et al.*, 2004]. We use volumetric strain, barometric, and broadband seismic data to constrain explosion source mechanisms and dynamic parameters. Such high-precision strain measurements of explosions are valuable but scarce. The innovation of using strain data acquired in 2008–09 to explore explosion dynamics is shared with data procured in 2003 [*Voight et al.*, 2010], but significant new features are also recognized here, including explosion precursors, pre-explosion pressurization for one case that is used to deduce feeder-dike geometry, and acoustic waves that define explosion onsets. We present unique pressure-corrected-strain and microbarometric gravity wave records for all events and all operating stations.

2. Background on Explosions of July 2008 to January 2009

[3] Prior to the explosion on 29 July 2008, a 38-h swarm of intense seismicity occurred with hypocenters at 0–3.4 km depth [*Stewart et al.*, 2008]. Then, following a quiet interval of a few hours, intense seismicity restarted at 03:32 29 July 2008 (all times are UTC), probably indicating a minor partial collapse of the dome carapace . The explosion followed at 03:38, generated a plume to 12 km a.s.l. (Washington VAAC), and produced pyroclastic flows in several drainages [*Stewart et al.*, 2008; *Komorowski et al.*, 2010]. The pyroclastic flows contained vesicular pumice and were generated by column collapse. Disregarding deposits in the Tar River valley due to a partial dome collapse [*Stewart et al.*, 2008], pumice deposits on the west flank covered 1.9 km^2, and assuming a thickness of 0.5–1 m, comprised a volume of 0.5–1.0 Mm^3 DRE, plus co-pyroclastic flow (co-PF) ash [*Komorowski et al.*, 2010; G. Wadge, written communication, 2010].

[4] The second extrusive phase a few months later was more complex [*Stewart et al.*, 2009]. Following volcano-tectonic seismicity starting 28 Nov, an explosion occurred at 01: 37 on 3 Dec 2008. It was preceded by 2 min of seismicity with rockfall signatures. Dome-rock ballistics were propelled 2 km from the vent, a plume rose 12 km high, and pyroclastic currents split around St. Georges Hill west of the volcano [*Komorowski et al.*, 2010]. Of the three events considered here, the Dec deposits are least well understood, because of burial during further events in Dec and Jan, and rainfall erosion. A surge possibly caused by a small directed blast from the dome base occurred in the Gages area. The deposit limits were carefully mapped [*Komorowski et al.*, 2010], yielding a volume of 1.34 Mm^3 DRE including co-PF ash, or 1.17 Mm^3 excluding co-PF ash. However, the proportions of lithics and pumice are not well defined. *Komorowski et al.* [2010] presumed most ejecta were from hot pre-Dec 2008 dome rock, in essence not fresh juvenile products; but *Wadge et al*. [2010] included the DRE volume as fresh lava.

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[5] Three subsequent smaller explosions (4 Dec, 0:43; 4 Dec, 20:22; 5 Dec, 9:41) caused plumes of limited height, but only the last generated a pyroclastic flow. The dome grew rapidly after 8 Dec, and after 21 Dec cyclic tremor was observed [*Stewart et al.*, 2009], similar to tilt/seismic cycles in 1997 and 1999–2000 [*Voight et al.*, 1999; *Young et al.*, 2003]. The cyclic tremor amplitude increased after 31 Dec, suggesting increased pressure [*Voight et al.*, 1999; *Young et al.*, 2003]. On 2 Jan 2009 a small pyroclastic flow developed, and on 3 Jan explosions occurred from the dome top at 6:55, 8:13, 8:47 and 11:07 [*Komorowski et al.*, 2010]. The last two began with jets to ~500 m, and generated pumice flows by column collapse. The last event produced a plume to 11 km. The pumice deposit volume is not so well defined as for the July explosion. For all 3 Jan events, *Komorowski et al.* [2010] reckoned a total volume of 0.95 Mm^3 DRE, but constraints are poor and partitioning between events is uncertain. For the largest event at 11:07, we take 0.5 Mm^3 DRE as a rough estimate of pumice deposits.

[6] Complete computations of eruption volumes including plume-height based estimates are given in the auxiliary material.[1] The plume is modeled as a buoyant thermal in a humid tropical environment [*Woods and Kienle*, 1994; *Sparks et al.*, 1997; *Druitt et al.*, 2002; *Tupper et al.*, 2009]. The above volumes are approximate, but they provide useful quantitative constraints for analyses based on strain observations.

3. Instrumentation and Observations

[7] Data were obtained from the CALIPSO array with instruments in ~200-m-deep boreholes and surface sites, at distances of about 5.1 to 9.6 km from the vent [*Mattioli et al.*, 2004] (Figure 1). We furnish observations of strain and air pressure linked to the explosions, and compare these data with broadband seismic records from surface station MBGH. Three sites (AIRS, TRNT, GERD) were operating for the 3 Jan 2009 explosion, and two (AIRS, TRNT) for 29 July and 3 Dec 2008. Each borehole contains a Sacks-Evertson high-dynamic-range volumetric strainmeter recording data at 50 samples per second (sps) with a precision of ~0.1 nanostrain (nS) in periods from 0–50 Hz [*Linde and Sacks*, 1995]. In this paper we round-off values to the nearest nanostrain.

[8] The strainmeters are affected by changes in atmospheric pressure, requiring corrections based on microbarometric data [e.g., *Nakao et al.*, 1989] (see auxiliary material). The explosions generated complicated airwaves recorded at 1 sps on barographs, with a shock wave spike at explosion onset, followed by a strong pressure decrease that evolved into long-period oscillations (Figures 2 and S1–S5). Similar distinctive ~800 s oscillations were observed with SHV explosions in 2003 [*Voight et al.*, 2010]. They propagated at about 30 m s^{-1} beyond 5 km as gravity waves, i.e., atmospheric oscillations generated by the rising plume, and were recorded at various azimuths and distances. Such observations of gravity waves in the near-field are unprecedented for explosions at this scale [cf. *Ripepe et al.*, 2010; *Kanamori et al.*, 1994].

[1]Auxiliary materials are available in the HTML. doi:10.1029/2010GL044661.

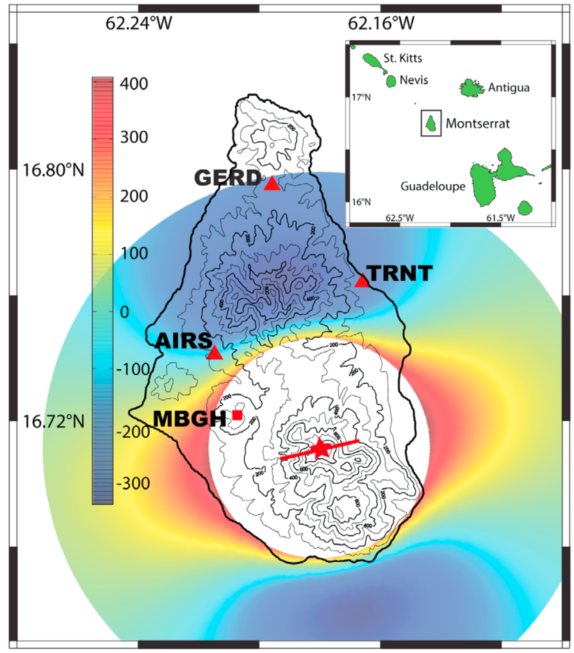

Figure 1. Montserrat map with location of strainmeter sites (red triangles), the broadband seismic station MBGH (red dot) and the approximate location of the active vent of Soufrière Hills Volcano (red star). Colored pattern is surface volumetric strain for modeled dike with N77°E trend, 600 m width, −2 to −3 km depth, 10 MPa pressure, 5 GPa rock modulus (see text for details). Blue is field of volumetric contraction, orange is field of expansion. Scale is nanostrains.

[9] The data from AIRS and GERD show high signal/noise ratios (Figure 2). TRNT, located near the shoreline, yielded noisy data for July 2008 and Jan 2009 explosions (amplitudes ~10 nS), but was crucial for identifying strain polarity (Figures S1–S3). For the July and Jan events the strainmeters at all operating sites expanded. Because AIRS and the other sites are far beyond the nodal distance for a shallow axisymmetrical source [*Linde & Sacks*, 1995], the source had to contract, and this indicates a source pressure reduction. The data at AIRS, TRNT and GERD have identical polarity and can be satisfied by a shallow cylindrical conduit source. Thus the July and Jan explosion patterns (Figures 2, S1, and S3–S5) confirm the same four strain stages recognized in 2003 explosion data [*Voight et al.*, 2010]: 1) a brief transition between disturbance onset and a profound change in strain; 2) a quasi-linear ramp producing most of the strain change; 3) a more gentle decrease of strain to a minimum value; and 4) a recovery lasting hours to approach a background magnitude. In several 2008 cases we recognize also a precursor stage (p). No precursor stage is recognized for 3 Jan, but for the 29 July explosion, seismic activity initiated 6 min prior to the explosion, interpreted as a minor rockfall.

[10] A different pattern developed on 3 Dec 2008, quite unlike any of the explosions of 2003 [*Voight et al.*, 2010]. In this case a strong and rapid (~100 s) strain pulse preceded the explosion, accompanied by bursts of seismicity (Stage p (Figure 2)). AIRS recorded peak values of ~143 nS, and the

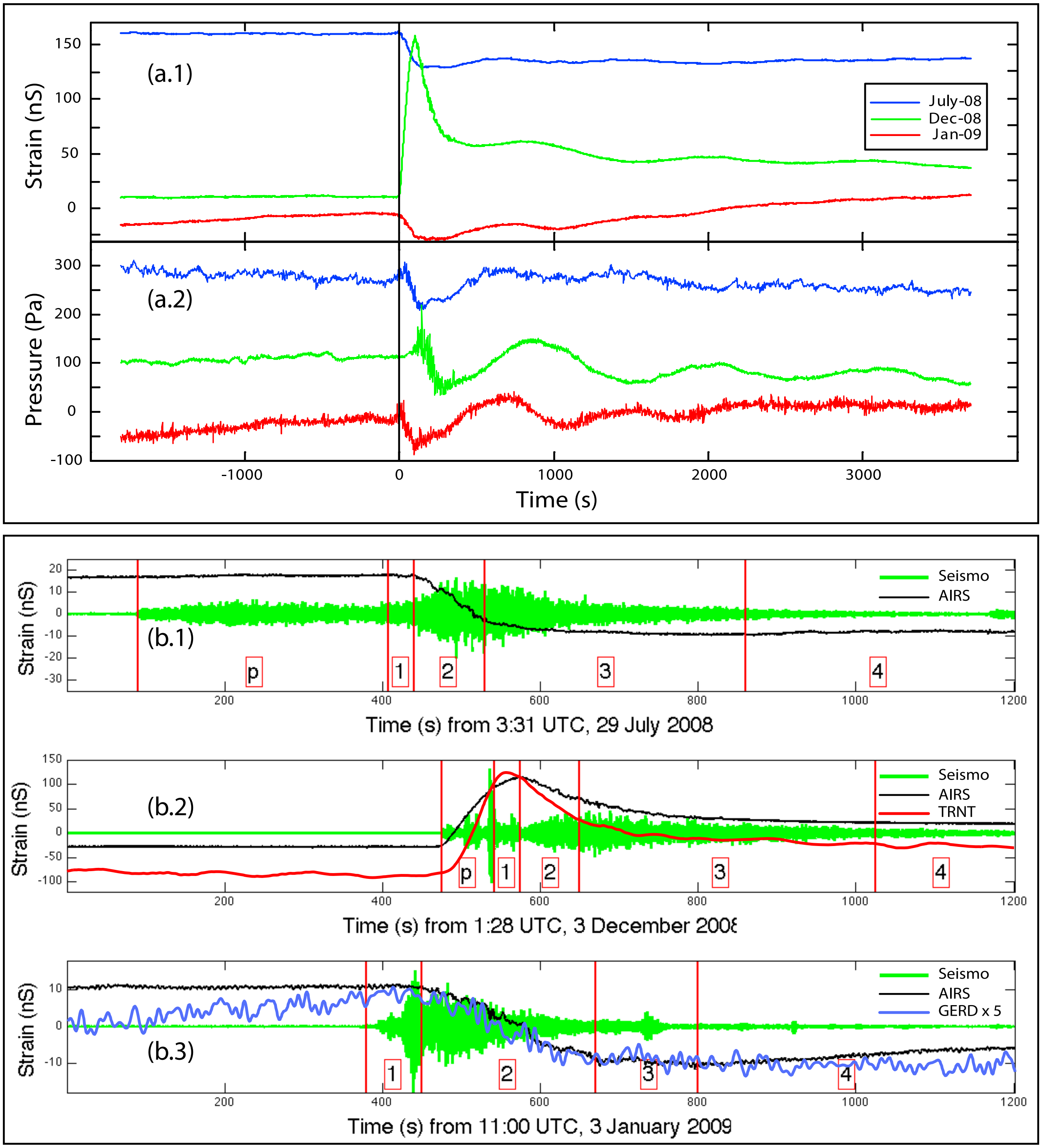

Figure 2. (a) Raw strain data and raw barometer data for three explosive events (from top to bottom: 29 July 2008, 3 Dec 2008, 3 Jan 2009) recorded at AIRS. Traces have been aligned with the approximate onset of the strain signal (29 July 2008 at 3:38:30; 3 Dec 2008 at 1:35:50; 3 Jan 2009 at 11:07:20; all times UTC), with the onsets marked by a vertical line on the two panels. In the Dec event, the explosion followed a rapid large strain buildup. (b) Strain signals corrected for atmospheric pressure for the same three explosions recorded at AIRS (black), TRNT (red) and GERD (blue). Strain signals are superposed on the respective seismic signals recorded at MGBH (green), with normalized amplitudes. Stages as explained in text are separated by red lines, with [*p*] [1] [2] [3] [4] the successive stages.

more distant TRNT ~202 nS (Figure S2), both with high signal/noise ratios and the same polarity (contraction). During the following ~10 minutes both sites showed reduction in strain (expansion), accompanied by explosive surface manifestations (acoustic and gravity waves, and plume generation). Seismic activity virtually coincided with the initial strain pulse (Figure 2), but before onset of the rapid pulse we detected no strain or seismic changes that could be related to either the strain pulse or explosion.

[11] The Stage *p* changes for the Dec event are inconsistent with solely a spherical pressure source at any depth under SHV, or a shallow cylindrical source, or a single-sensed combination of the two [*Lisowski*, 2007]. The observed strain polarities can be produced by such sources, but not strain magnitudes at TRNT that exceed those at the nearer station AIRS; also for a cylindrical source the implied pressures exceed conduit strength (see auxiliary material). Thus we sought to see if a dike source could help to satisfy the data. As developed in Section 4, our analyses strongly support inclusion of a dike source, with Stage *p* comprising mainly an intense, rapid dike pressurization. This pattern is comparable with that observed during the 3 March 2004 event, also modeled using a dominant dike source [*Linde et al.*, 2010].

[12] We judge Stage 1 of the Dec explosion as having started shortly before the AIRS strain peak was reached (the TRNT peak occurred slightly earlier), no doubt triggered by the pressure increase and indicated by a reduction in strain rate (slight break-in-slope) and generation of an acoustic wave that originated at the vent ~15 s before its arrival at AIRS (5.1 km/340 m s^{-1}). The peak of the AIRS strain curve (~143 nS) marks the point where incremental strain released by the explosion exceeds the dike source strain build-up. This defines the Stage 2 onset, and thereafter the stages appear similar to those of the other explosions. The strain recoveries at AIRS and TRNT are −88 and −135 nS, respectively, approximately 0.6–0.7 of the peak values.

[13] For all explosions, Stage 1 is a brief period with small strain and small seismicity that occurs when the pressurized gas under a strong plug reaches the strength of the cap [*Voight et al.*, 2010; *Alidibirov and Dingwell*, 1996]. The acoustic pulse is generated within Stage 1 and indicates that fragmentation and material expulsion is initiated in this stage. Stage 2 is a swift contraction of the pressure source that correlates with energetic expulsion of the pyroclast/gas mixture in the conduit, with durations of 105 and 230 s for the July and Jan events, respectively. *Voight et al.* [2010] propose that most of the ejecta are expelled during the quasi-linear strain-ramp, and that the further gradual reduction of strain to a minimum value in Stage 3 reflects combined effects of discharge of ash and gas from the top of the magma column, pressure adjustments within the deeper conduit system, and rise of magma. Stage 4 then represents a gradual strain recovery over several hours due to magma ascent and exsolution/gas pressure effects.

4. Analyses: Procedures, Results, and Discussion

[14] For the July and Jan events the strain-ramp is interpreted as a rapid reduction in internal pressure about a quasi-cylindrical conduit wall, associated with pyroclast/gas evacuation. We assume a conduit with diameter of 30 m [*Voight et al.*, 1999; *Melnik and Sparks*, 2002], positioned over a planar feeder dike that rises from the magma chamber at ~5 km depth [cf. *Mattioli et al.*, 1998; *Costa et al.*, 2007; *Hautmann et al.*, 2009; *Linde et al.*, 2010].We assume the cylinder/dike transition occurs roughly at 2 km beneath the vent at ~1000 m a.s.l. Our data suggest strong involvement of a dike in the Dec event.

[15] Following *Voight et al.* [2010] we estimate conduit drawdown from erupted DRE volume adjusted for porosity, divided by conduit area. We assume an average density of ~1500 kg m^{-3}, based on clast data from 1997 and 2003 explosions [*Clarke et al.*, 2002; *Edmonds et al.*, 2006]. For a 30-m conduit the total volume estimates of 0.92–1.42 Mm^3 DRE for July, and 0.76 Mm^3 DRE for Jan, suggest drawdowns of 2.3–3.5 km (July), and 1.9 km (Jan). The 2.3 km estimate for July is roughly compatible with our proposed 2 km nominal depth to the top of the dike, but the larger volume implies a 3.5 km drawdown. Because strain data for July suggest no dike signature, we prefer the lower volume and drawdown estimates. For the Dec explosion, discussed below, the issue is more complex.

[16] We then use our strain-ramp amplitudes to estimate average conduit pressures released in the explosions [*Voight et al.*, 2010]. Using an axisymmetric elastic finite element model (see auxiliary material), we apply a magmastatic load using an average density of 1500 kg m^{-3} and a superposed uniform overpressure (pressure exceeding magmastatic) over the drawdown lengths of conduit (see auxiliary material). We developed sub-models for homogeneous and inhomogeneous cases, in the latter instance applying a velocity/modulus increase with depth as based on the SEA-CALIPSO experiment [*Shalev et al.*, 2010]. For the homogeneous case we used Young's modulus $E = 5$ GPa, the average local modulus at the strainmeter sites, and a Poisson ratio of 0.25. Comparable moduli have been used in modeling other volcanoes [*Lisowski*, 2007; *Houlie and Montagner*, 2007] (and others cited in auxiliary material) and suggest a compliant, fractured and/or hydrothermally altered rock mass. For inhomogeneous media and AIRS data (best signal/noise ratio), the Stage 2 strain-ramps for July and Jan explosions (24 and 18 nS) suggest preliminary average overpressures of ~30 and ~25 MPa; homogeneneous-case values are slightly larger. These results are dependent on magma density and modulus. Increasing magma density can reduce overpressures <20%, whereas assuming $E = 3$ GPa (arguably typical for volcanic areas [see *Lisowski*, 2007]) yields overpressures of order 10 MPa. Although *Robertson et al.* [1998] proposed an overpressure of 27.5 MPa for a 1996 SHV explosion, we nevertheless regard our larger estimates with discomfort. Typically, overpressures <15–20 MPa are given by conduit models for magma attributes appropriate for SHV, consistent with our reduced-modulus results [*de' Michieli Vitturi et al.*, 2010; *Melnik and Sparks*, 2002; *Voight et al.*, 2010]. For Jan 2009 we can also calculate the ratio of measured strain changes at AIRS and GERD sites (Figure 2) as 18/4.5 = 4. The corresponding modeled values are >5; this discrepancy, also recognized for 2003 explosion data [*Voight et al.*, 2010], might reflect modulus heterogeneity [cf. *Shalev et al.*, 2010]. Associated mass evacuation rates are given by erupted mass divided by Stage 2 strain-ramp durations, yielding 2.3×10^7, and 0.86×10^7 kg s^{-1} for the July and Jan explosions, respectively. These results are similar to those reported for explosions at SHV in 1997 [*Druitt et al.*, 2002] and 2003 [*Voight et al.*, 2010].

[17] For the Dec explosion we propose that a dominant dike source (with lesser contributions from conduit and chamber sources) can explain the observations. Stage *p* and Stage 1 strains at AIRS, TRNT (c.143,202 nS) correspond mainly to pressurization of a dike; the contribution from the

upper cylindrical conduit is only of the order 20 or 10 nS at AIRS, TRNT (e.g., July and Jan data), with chamber contributions uncertain but likely relatively small. During Stage 2, with ejection of the gas/pyroclast mixture, the cylindrical conduit contracts and the dike partially closes. To test this hypothesis we calculated strain changes using several finite-element models (Figure 1 and auxiliary material). The top of the vertical dike was held to the depth range 1 to 2 km. The modeled strains at AIRS and TRNT vary rapidly with dike azimuth and their polarity requires an ENE dike orientation (best fit, ~N77°E). For the March 2004 event, *Linde et al.* [2010] found evidence to constrain a dike of WNW orientation. These and other results [e.g., *Hautmann et al.*, 2009] suggest that feeder dikes at SHV had diverse attitudes at different times during the eruption.

[18] We use a model dike width of 600 m but our conclusions are independent of this choice, as model strain values depend on the product of dike area and opening (or pressure). With $E = 5$ GPa, and pressure 10 MPa, Figure 1 shows model contractive strains at AIRS and TRNT as 125 and 185 nS, respectively. With addition of the upper conduit contributions at these sites as ~20 and 10 nS (contractive during Stage *p* pressurization), a nearly exact match is produced of the observed strains. This result demonstrates the plausibility of a dike source although the solution is not unique, and no contribution from a magma chamber is included in the calculation. For the March 2004 event, *Linde et al.* [2010] modeled a dike and a deep spherical pressure source, and neglected a shallow source. The pressure loss assumed for the deep source contributed contractions of 50 and 20 nS at AIRS and TRNT, respectively. However, we think it is not certain that a deep source involved a pressure loss. An increase in chamber pressure applied instantaneously to the bottom of a (cylindrical) conduit has been modeled by *de' Michieli Vitturi et al.* [2010]. For magma properties based on SHV, a pressure increase is transmitted though the lower 3/5 of the conduit in only about 1 min; ~10 min are required to pressurize the upper 4/5, and ~30 min to affect also the uppermost conduit. These results are germane to the Dec 2008 and March 2004 events, as they show that deep pressure increases can in principle be transmitted in very short times (few minutes) to mid-level portions of the transport system, and thereby be detected by our strainmeter array.

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