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Stress Triggering of the 2005 Eruption of Sierra Negra Volcano, Galápagos

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Key Points:

- New thermomechanical models provide an estimation of magma system stability in the lead up to the 2005 eruption of Sierra Negra
- Models suggest that Sierra Negra's magma system was in stable storage prior to eruption with minimal overpressure and no tensile failure
- Coulomb static stress calculations indicate that a M_w 5.4 earthquake likely triggered the 2005 eruption

Supporting Information:

- Supporting Information S1

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Abstract Extensive vertical deformation (>4.5 m) observed at Sierra Negra volcano Galápagos, Ecuador, between 1992 and the 2005 eruption led scientists to hypothesize that repeated faulting events relieved magma chamber overpressure and prevented eruption. To better understand the catalyst of the 2005 eruption, thermomechanical models are used to track the stress state and stability of the magma storage system during the 1992–2005 inflation events. Numerical experiments indicate that the host rock surrounding the Sierra Negra reservoir remained in compression with minimal changes in overpressure (~10 MPa) leading up to the 2005 eruption. The lack of tensile failure and minimal overpressure accumulation likely inhibited dike initiation and accommodated the significant inflation without the need for pressure relief through shallow trapdoor faulting events. The models indicate that static stress transfer due to the M_w 5.4 earthquake 3 hr prior to the eruption most likely triggered tensile failure and catalyzed the 2005 eruption.

Plain Language Summary Tracking the stability of a magma system in the lead up to a volcanic eruption requires investigating both the pressure state of the magma reservoir and stress accumulation in the host rock. New coupled conduit flow-magma reservoir pressurization models are used to evaluate the evolution of the magma reservoir of Sierra Negra volcano, Galápagos, in the lead up to its 2005 eruption. Stress calculations indicate that the magma reservoir was stable prior to the 2005 eruption and that the eruption was likely triggered by a M_w 5.4 earthquake that occurred 3 hours prior to the event. The new modeling approach has important implications for tracking the stress evolution of magma systems to evaluate future unrest and eruption triggering mechanisms at volcanoes worldwide.

1. Introduction

A classic paradigm in volcanology is that eruption occurs when the pressure within a magma reservoir exceeds the strength of the host rock surrounding it (Fowler & Spera, 2008; Huppert & Woods, 2002; Jaupart & Vergniolle, 1989; Tait et al., 1989; Wilson, 1980). Referred to as "overpressure," magma chamber pressurization is often cited as the primary mechanism for triggering volcanic eruption. Overpressure is hypothesized to be generated by the rapid change in reservoir volume resulting from mechanisms such as the injection of new material, phase change, the exsolution of volatiles, or a combination of these processes (Blake, 1981; Stock et al., 2016). In this paradigm, eruption is generally preceded by the swelling of the ground surface as the magma chamber expands. Magma reservoir inflation and subsequent volcano deformation is observable using geodetic methods such as tiltmeters, Interferometric Synthetic Aperture Radar (InSAR), and Global Positioning System (GPS; Biggs & Pritchard, 2017; Bjornsson et al., 1977; Dvorak & Dzurisin, 1997; Dzurisin, 2000; Lu & Zhang, 2014; Massonnet & Sigmundsson, 2000; Segall, 2013; Sparks, 2003). Overpressure can be estimated from surface inflation using classic analytical models (e.g., McTigue, 1987; Mogi, 1958), providing a first-order assessment of the system's evolution. However, the links between surface deformation, magma chamber pressurization, and eruption potential remain elusive (Gregg et al., 2012, 2013; Masterlark, 2007). While overpressure facilitates magma evacuation, failure in the host rock surrounding a magma system, and in particular tensile failure at the magma-rock interface, is critical for catalyzing an eruption. As such, determining the stability of magma in storage by estimating host rock stress evolution is necessary for assessing eruption potential (Acocella, 2007, 2010; Gerbault et al., 2012; Gregg et al., 2012; Grosfils, 2007; Grosfils et al., 2015; Marti et al., 2008).

While the impact of stress change on eruption triggering has previously been recognized as important for assessing unrest and forecasting eruption potential (Chadwick et al., 2006; Manga & Brodsky, 2006; Sulpizio et al., 2017; Sulpizio & Massaro, 2017), overpressure is more often used when discussing the threshold for magma reservoir stability. In the case of the prolonged unrest and uplift period prior to the 2005 eruption of Sierra Negra Volcano in the Galápagos, trapdoor-faulting events were postulated to have relieved overpressure accumulation in the rapidly expanding magma reservoir (Amelung et al., 2000; Chadwick et al., 2006). Unfortunately, estimating the overpressurization of a magma body is challenging, especially when observations of unrest begin in the middle of an eruption cycle. It is exceedingly difficult to back out how much deformation has occurred prior to detailed geodetic measurements to determine the baseline pressure and stress state (Del Negro et al., 2009; Gregg et al., 2012; Masterlark, 2007; McTigue, 1987; Mogi, 1958). Recent fluid-structure finite element method modeling approaches provide a means for producing model predictions of magma chamber stability by tracking stress and strain in the host rock while calculating changes in overpressure associated with deformation (Gregg et al., 2015; Le Mével et al., 2016).

In this investigation, we take advantage of recent advancements in volcano, fluid-structure interaction modeling to calculate the stress evolution of the Sierra Negra magma system in the lead up to its 2005 eruption. Overpressure, Mohr-Coulomb failure, and tensile stress are estimated to determine the stability of the magma system as it evolves. Of particular focus in this investigation is the interplay between overpressure and stress accumulation in the host rock prior to the onset of eruption. Additionally, we investigate stress evolution due to the thermal effect of prolonged magma flux, thermal contraction, and crystallization and discuss distinguishing volume change due to crystallization and volatile exsolution from volume change due to magma injection.

2. Unrest and the 2005 Eruption of Sierra Negra

Sierra Negra is a 60 × 40-km basaltic shield volcano that comprises most of the southern portion of Isabela Island and is the largest of the Galápagos volcanoes (Figure 1; Reynolds & Geist, 1995). Significant data collection efforts preceded the 2005 eruption of Sierra Negra, including deformation observations from InSAR, campaign GPS (2001–2003), the installation of continuous GPS stations in 2002, and campaign microgravity studies conducted in 2001–2002, 2005, 2006, and 2007 (Amelung et al., 2000; Chadwick et al., 2006; Geist et al., 2006, 2008; Vigouroux et al., 2008). In the lead up to the 2005 eruption, Sierra Negra experienced two episodes of extraordinary uplift: ~38 cm/a from 1992 to 1999 and ~120 cm/a in 2004 to 2005 when it reached a rate of ~1 cm/day prior to the eruption. The observed preeruption inflation culminated in nearly 5 m of cumulative vertical displacement (Chadwick et al., 2006).

Analytical models of the 1992–1999 InSAR observations indicate a rapidly inflating sill located ~2.1 km beneath the Sierra Negra Caldera (Amelung et al., 2000). Significant uplift during this time period, > 2 m, is thought to have triggered trapdoor faulting and a M_w 5.0 earthquake on 11 January 1998 (Amelung et al., 2000; Chadwick et al., 2006; Jonsson, 2009). From 2000 to 2002, deflation of 9 cm/a was observed and modeled as a contracting sill of similar depth and geometry as the inflation source modeled by Amelung et al. (2000; Geist et al., 2006). The subsidence might be attributed to degassing, magma withdrawal back into the plumbing system, and/or viscoelastic relaxation (Geist et al., 2006, 2008). In 2004, a new period of inflation commenced at a higher rate than observed in the 1990s (Chadwick et al., 2006) and with exponentially increasing rates (Figure 2a). At its peak, Sierra Negra experienced vertical uplift rates >1 cm/day, triggering additional moderate-sized earthquakes including a M_b 4.0 on 23 February 2005, a M_b 4.6 earthquake on 16 April 2005, and a M_w 5.4 earthquake ~3 hr prior to the 22 October 2005 eruption. The 2005 eruption initiated as a 13–14-km-high volcanic plume and included lava fountains originating from a 2-km-long fissure and continued focused lava fountains that fed lava flowing into the caldera (Geist et al., 2008). The eruption ceased on 30 October 2005 after ~8 days of activity.

3. Model Formulation, Results, and Discussion

Our numerical approach, described in detail in supporting information S1, utilizes previously developed and benchmarked, thermomechanical and fluid-structure interaction finite element method models (Anderson, 1936, 1951; de Silva & Gregg, 2014; Del Negro et al., 2009; Gregg et al., 2012, 2013; Grosfils, 2007; Hickey & Gottsmann, 2014; Le Mével et al., 2016; Smith et al., 2009). COMSOL Multiphysics 5.3a is used to calculate

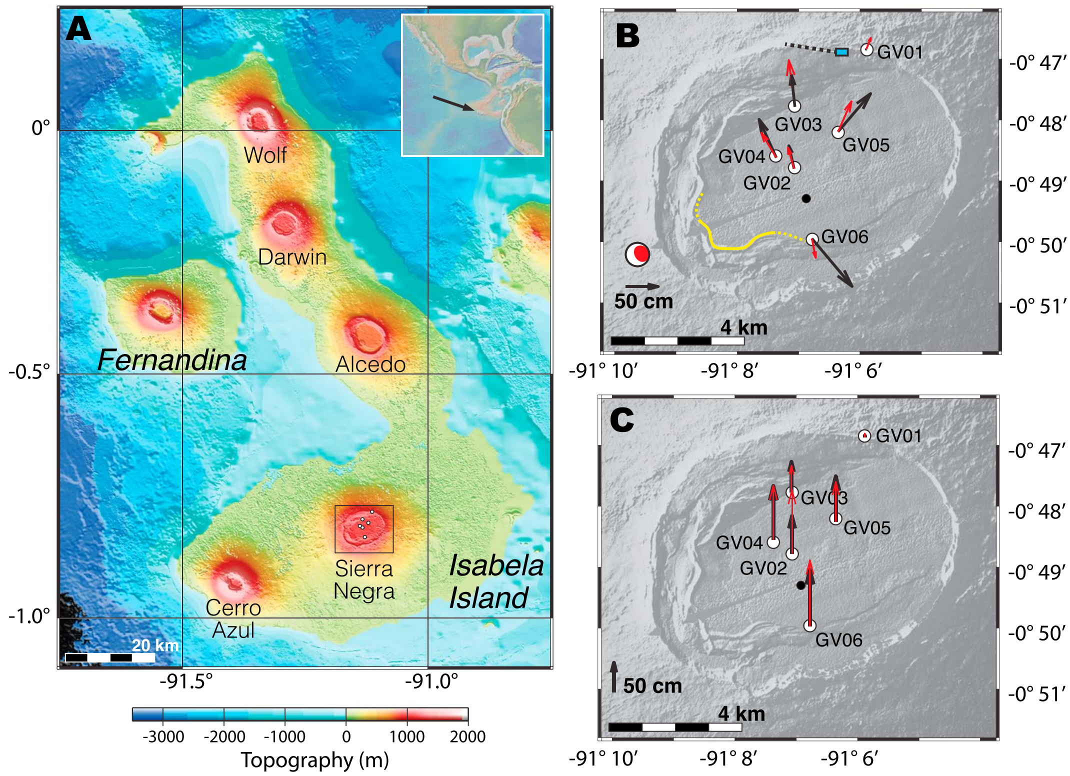

Figure 1. Best-fit thermomechanical model results compared to continuous GPS observations (Chadwick et al., 2006). (a) Location map. Data from GeoMapApp plotted with GMT. (b) Digital Elevation Model (DEM) of the summit of Sierra Negra (Yun et al., 2006) with the observed GPS horizontal displacements (black arrows) from 1 April 2003 to 21 October 2005 and modeled horizontal displacements (red arrows) for the entire time series. The center of the model source is depicted by the small black circle southeast of GV02. The precise location of the 22 October 2005 M_W 5.4 earthquake was not measured but is attributed to a fresh fault scarp (yellow and yellow dashed line) observed by Geist et al. (2008); the Global Central Moment Tensor (CMT) focal mechanism is provided (Dziewonski et al., 1981; Ekström et al., 2012). The initial eruption fissure is indicated by a black-dashed line on the north caldera scarp with the location of vents that remained active throughout the eruption depicted by a blue box. (c) The observed GPS vertical displacements (black arrows) from 1 April 2003 to 21 October 2005 and modeled vertical displacements (red arrows) for the entire time series from 1 April 2003 to 21 October 2005 (the misfit with GV02 reflects that the instrument went offline 10 June 2005, and GV01 was only operational until 3 December 2004). GPS = Global Positioning System.

the stress, strain, and temperature variations due to a viscous magma flowing from a deeper source into an existing, magma-filled reservoir (see supporting information S1 for full model details). A series of viscoelastic, fluid-structure interaction models was run to investigate the predicted stress evolution resulting from the observed surface deformation from 1992 to 2005. As the reservoir geometry and location (an ellipsoid 0.4 km tall × 7 km wide centered at 3-km depth) were constrained by previous geodetic studies (Amelung et al., 2000; Chadwick et al., 2006; Jonsson et al., 2005; Yun et al., 2006), we focus on the mass flux necessary to reproduce the uplift. Stability of the system is assessed by investigating the magnitude of the change in overpressure, the extent of Mohr-Coulomb failure in the host rock, and the presence of tensile failure along the magma reservoir boundary (Gregg et al., 2012, 2013, 2015; Grosfils, 2007; Grosfils et al., 2015). Overpressure is calculated as the local force per area along the reservoir-host rock interface with variable states of stress experienced for different regions of the interface. While the mere presence of significant pressurization, tensile failure, and/or through-going Mohr-Coulomb failure may not be sufficient to drive eruption, they provide a first-order approximation of stability. Although our discussions below focus on the fully coupled temperature-dependent viscoelastic calculations, a fully nontemperature-dependent, viscoelastic model (nTd1) and a model with nontemperature-dependent elastic moduli and a temperature-dependent viscosity (nTd2) are provided for comparison purposes (see Table S1 for model parameters and Table S2 for model variables). COMSOL's parameter sweep optimization was used to minimize the error between model predicted and observed vertical displacement.

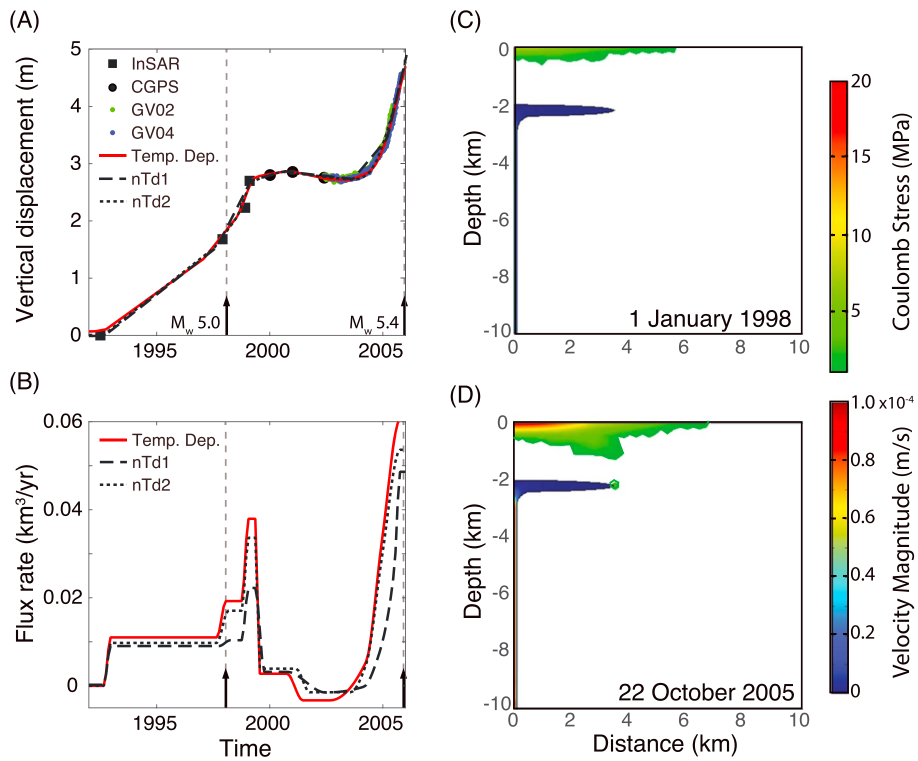

Figure 2. Model predicted flux variations leading up to the 2005 eruption. (a) Model predictions of vertical displacement from the best-fit temperature-dependent case (red solid line) and nontemperature-dependent models (black dashed and dotted lines) compared to InSAR (black squares), campaign Global Positioning System (black circles), and continuous Global Positioning System (colored circles) observations. (b) Flux boundary condition variations for the best-fit models. Data plotted from Chadwick et al. (2006). (c) Calculated Coulomb stress in the temperature-dependent model at 1 January 1998 just prior to the M_W 5.0 earthquake on 11 January 1998 (Amelung et al., 2000). White line is provided to highlight the 0-MPa Coulomb stress contour. (d) Calculated Coulomb stress in the temperature-dependent model on 22 October 2005 just prior to the recorded M_W 5.4 at 2034 UTC and eruption at 2330 UTC. Magma velocity magnitude is plotted within the reservoir and conduit. InSAR = Interferometric Synthetic Aperture Radar.

3.1. Preeruption Stress Evolution

The preeruption deformation of Sierra Negra is divided into three periods for model assessment: (1) the 1992–1999 inflation period; (2) the 2000–2003 deflation period; and (3) the preeruption 2004–2005 inflation period. To reproduce the 1992–1999 pattern of inflation, magma flux rates must average from 0.011 km^3/yr (Figure 2 and Tables S3–S5) through 1998 and increase to 0.03795 km^3/yr in 1998 (Figure 2b). In 1998, the best-fit model predicts Mohr-Coulomb failure and faulting in the upper 500 m of the model space (Figure 2c). The failure prediction coincides with the timing of observed seismicity including a M_W 5.0 earthquake on 11 January 1998 (Amelung et al., 2000). On the other hand, during the 1992–1999 inflation period, our models predict that the magma chamber should remain stable. There are no regions along the chamber boundary where tensile failure is observed and the predicted maximum change in reservoir overpressure remains modest (~6 MPa; Figure 3b and Table S4).

In 2000, there was deceleration and roll off in the uplift signal that is reproduced in the model by ceasing magma injection and allowing the host rock around the magma chamber to relax viscoelastically. However, reproducing the observed subsidence in 2001–2003 requires a decrease in the magma chamber volume. This can be achieved by allowing back flow of magma into the pipe at a rate of −0.0033 km^3/yr. Contraction due to conductive cooling during this time period does not produce an observable deformation signal due to its modest nature. In particular, the majority of the cooling occurs in a thin region along the flanks of the magma reservoir, while the central portion of the reservoir retains its temperature (within ~5 °C; Figure 3d). In response to the volume loss, the overpressure in the magma chamber is reduced by ~0.2 MPa (Figure 3b and Table S4). Similarly, tensile stress is reduced slightly, 0.2 MPa, along the chamber boundary

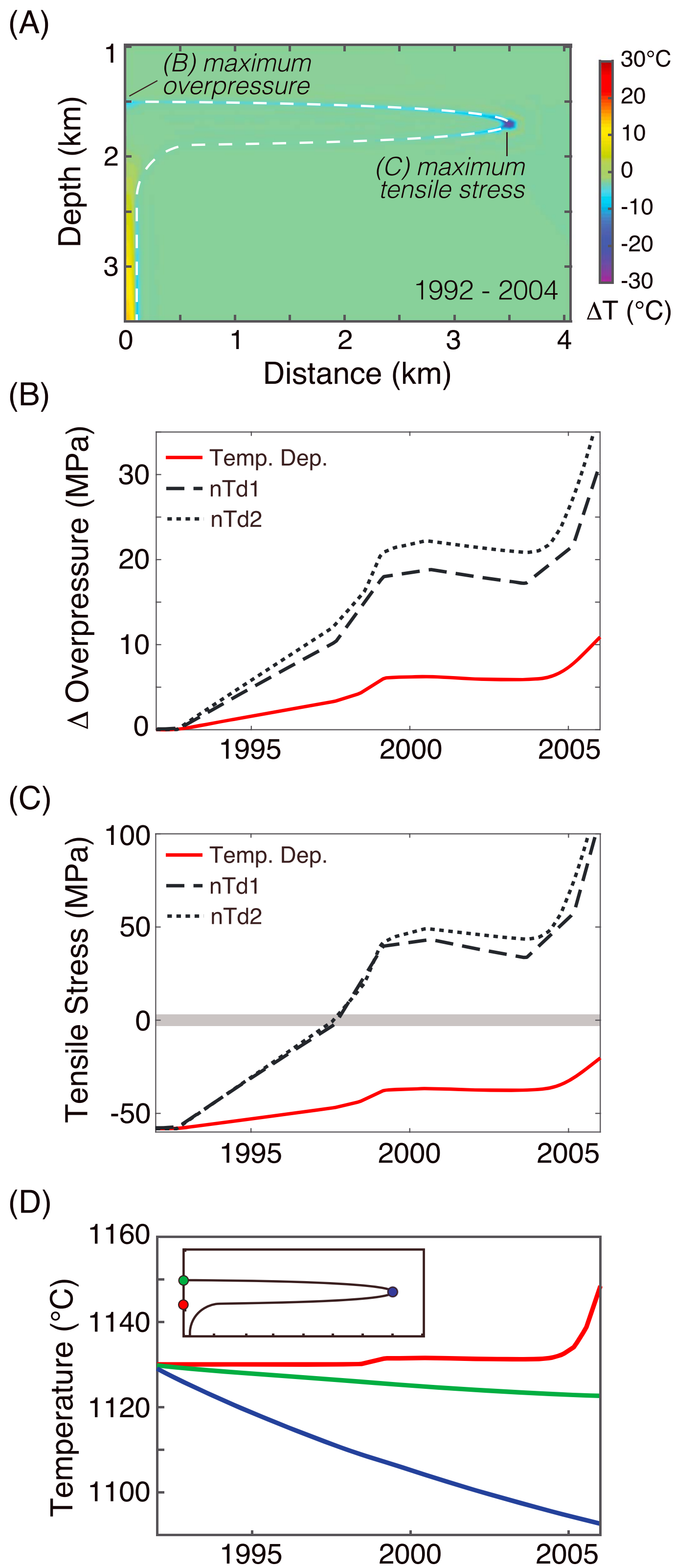

Figure 3. Model predictions of tensile stress and overpressure. (a) The change in model temperature observed from 1992 to 2005. The location of the maximum predicted tensile stress and maximum change in overpressure are indicated for (b) and (c). (b) Change in overpressure. (c) Maximum tensile stress. Negative tensile stress indicates compression, and positive tensile stress indicates tension. (d) Three points are investigated to compare the thermal evolution: above the central feeder conduit (red); the top, center of the reservoir (green); and the most distal point (blue).

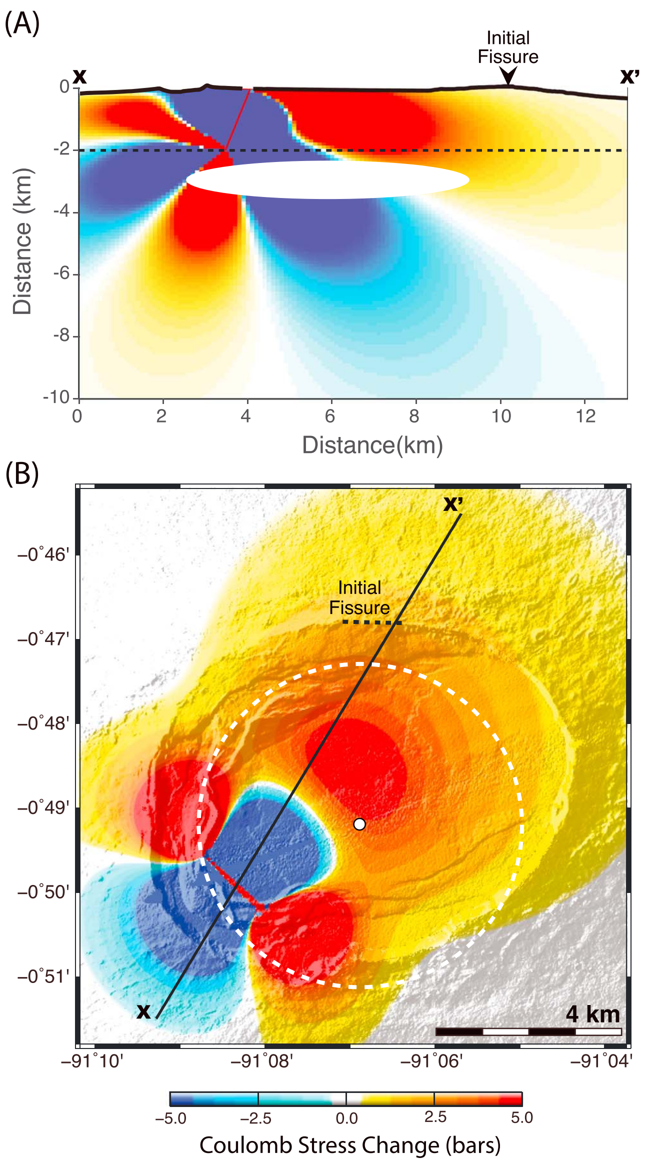

Figure 4. Calculated Coulomb static stress transfer on normal receiver faults due to the 22 October 2005 M_w 5.4 earthquake, which occurred 3 hr prior to Sierra Negra's eruption. Source parameters: seismic moment = 1.83e24 dyne * cm, strike = 146, dip = 74, Young's modulus = 50 GPa. (a) Cross section along X–X′ indicated on (b) through the source fault and the location of the initial eruption fissure (block arrow). The red line indicates the location of the source fault, and the magma reservoir location is indicated by the white ellipse. Please note that the magma reservoir is not part of the Coulomb 3.4 calculation. The dashed line indicates the location of the map view stress plot. (b) A map view stress section draped over Sierra Negra topography (shaded relief) taken at 2-km depth showing the reservoir center (white circle) and greatest extent (white-dashed line).

(Figure 3c and Table S5). Overall, the time period of quiescence appears to be too brief and the rate of volume loss too modest to significantly impact the stress state of the magma system.

At the end of 2003, magma injection commences again at a higher rate of 0.033 km^3/yr. As the ground surface inflates, significantly larger regions of the shallow crust undergo Mohr-Coulomb failure. By 2005, the regions of the model space exhibiting Mohr-Coulomb failure is extensive, impacting the overlying roof down to ~1.5-km depth (Figure 2d). Tensile stress along the magma reservoir boundary also increases, but the entire reservoir boundary remains in a compressional regime (Figure 3c and Table S5). Overpressure also significantly increases reaching its highest point at the center of the reservoir of 10.7 MPa just prior to the 2005 eruption (Figure 4b and Table S4). However, at the time step prior to the 2005 eruption, the modeled reservoir remains have no tensile failure and only moderate overpressure suggesting that it should be stable. Nevertheless, the widespread nature of Mohr-Coulomb failure and potential for triggering seismicity raises important considerations for the role of earthquakes leading up to the 2005 eruption.

3.2. Earthquake Catalyst for the 2005 Eruption

In the lead up to the 2005 eruption, the models predict that the magma reservoir remains in a stable stress state with minimal overpressure and no tensile stress above 0 MPa on or near the reservoir boundary. Although a predicted magnitude of overpressure $>$10 MPa is generally cited as a minimum level for dike propagation (Rubin, 1995), the greatest overpressures are observed in the central portion of the magma chamber where compressive stresses are greatest and dike initiation is unlikely. Additionally, the eruption appears to have initiated from the flank of the magma system (Geist et al., 2008). However, while the magma system appears to be stable in the model, large regions of the host rock are predicted to be in Mohr-Coulomb failure in the time steps leading up to the 2005 eruption (Figure 3c). Moreover, 3 hr prior to the 22 October 2005 eruption, a M_w 5.4 earthquake occurred, likely rupturing along the sinuous ridge on the southwestern base of the caldera where ~150 cm of dip-slip was observed (Geist et al., 2008).

The Coulomb static stress transfer due to a M_w 5.4 earthquake is investigated utilizing the U.S. Geological Survey (USGS) Coulomb 3.4 software (Lin & Stein, 2004; Toda et al., 2005). As the exact rupture geometry is not constrained, we have chosen a central segment to represent the multi-segment fault rupture (Figure 4b). Model results indicate that the 22 October 2005, 2034 UTC event likely relieved stress along the southwest portion of the caldera while increasing tensile stress, along the opposite, northeast region of the caldera and magma system (Figure 4). Three hours after the earthquake, the eruption initiated through a fissure on the north caldera rim. The triggering of volcanic events after has been previously observed (Chesley et al., 2012; Diez et al., 2005; Gregg et al., 2006; La Femina et al., 2004). Given the timing and location of the eruption, it is likely that the M_w 5.4 earthquake catalyzed the eruption on the opposite side of the magma reservoir where overpressure and tensile stresses were increasing but had not quite reached a critical threshold. Without the M_w 5.4 earthquake, the models predict that the magma reservoir would have continued to remain stable. The lack of eruption following the previous moderate earthquake events, M_w 5.0 on 11 January 1998 and M_b 4.6 on 16 April 2005, may indicate that the stress change due to the combination of reservoir inflation and the moderate-sized earthquake was not enough to trigger eruption.

An important caveat is that there is no deformation information prior to 1992 to constrain the stress state going into the inflation event. Two outcomes from the numerical experiments suggest that the approach is capturing some aspects of the preeruption stress evolution. First, the model-predicted increase in Coulomb failure (Figure 2c) is in agreement with the increase in seismicity and faulting observed at Sierra Negra in 1998 (e.g., Amelung et al., 2000). Were there to have been significant uplift prior to 1992, one would expect to see coincident seismicity, which is absent in the record. Second, the model indicates that the reservoir is priming for eruption but yet not failing in October 2008. Given the timing of the eruption, immediately following the M_w 5.4 earthquake, there is a strong suggestion that stress change induced by the earthquake pushed the reservoir, which was already primed, into eruption.

3.3. Thermal Impacts on Reservoir Stability

In long-lived magma systems with significant influx of heat and material, it is important to consider the thermal state of the host rock when assessing reservoir stability (Gregg et al., 2012, 2013; Jellinek & DePaolo, 2003). In the case of Sierra Negra, prolonged intrusion of material is likely to have increased the temperature of the host rock surrounding the reservoir, the effect of which is to lower both the viscosity and the elastic moduli of the host rock. The outcome of this reduction is to buffer stress accumulation, prevent failure, and increase stability (Gregg et al., 2012; Jellinek & DePaolo, 2003). By comparing the temperature-dependent model results to a nontemperature dependent, isoviscous (nTd1) case and a quasi-temperature-dependent case that includes constant elastic moduli but a temperature-dependent viscosity (nTd2), it appears that temperature plays a significant role in reducing the potential for failure when the impact on the elastic moduli is included. In the nTd model runs, the magma system is estimated to become unstable in 1998 after the first period of recorded inflation. At this time, the nTd models predict that the magma reservoir is in tensile failure and that the change in overpressure has surpassed 10 MPa. As the nTd models progress, both the tensile stress and change in overpressure are significant, reaching maximum values $>$30 and ~100 MPa, respectively. These values are likely untenable and indicate that models that do not include temperature-dependent elastic moduli are unable to provide a realistic estimation of the stress evolution of the Sierra Negra system given the lack of eruption until 2005.

3.4. Magma Injection Versus Volatile Exsolution

Microgravity observations led Vigouroux et al. (2008) to conclude that the rapid inflation period recorded in 2004–2005 leading up to the 2005 eruption may have been the result of second boiling (Blake, 1984; Tait et al., 1989) rather than the renewed injection of magma. The explosive nature of the opening phase of the 2005 eruption lends support to a volatile-driven pressure state. Furthermore, major element analysis using MELTS software indicates a 20 °C variation between the Early Phase and Main Phase erupted products of the 2005 event (1,128 to 1,108 °C, respectively), and ~13% crystallization is necessary to explain the trace element composition variation between the early and main phases (Geist et al., 2008). If the entire magma system were to have undergone such significant cooling and crystallization, crystallization-induced volatile exsolution is a very plausible eruption catalyst. As such, we investigate the extent of cooling during the 1992–2005 period of unrest.

When inflation paused between 1999 and 2003, the magma system should have experienced some cooling, but since information regarding the pre-1992 state of the magma system is unavailable, we cannot constrain whether the system had already undergone substantial cooling prior to the 1990s inflation event. The temperature-dependent model predicts a thin region of cooling, ~20°, at the flank of the reservoir between 1999 and 2003 but more modest cooling in the central portions of the reservoir near the conduit ($<$5°; Figure 3d). Although a volatile exsolution trigger cannot be fully ruled out, the subtlety and localized nature of the cooling observed in the model, which impacts only a small volume of the reservoir (Figure 3A), suggests that volatiles are an unlikely cause of the significant uplift observed prior to the 2005 eruption. Rather, a rejuvenation of magma injection is a more likely mechanism for the increased inflation in 2004–2005. Furthermore, the absence of crystals in the early phase of eruption (Geist et al., 2008) may indicate that the triggered dike that fed the initial eruption tapped a melt rich portion of the reservoir rather than a chilled and crystallized margin. An important caveat is that the true geometry and heterogeneity of the magma system may be more likened to an interconnected network of sills rather than a reservoir. Future efforts utilizing multiphase and perhaps 3-D modeling may be necessary to examine this problem in more detail.

Although magma compressibility is not included in the models, significant volatile content will impact the model results by increasing reservoir stability due to the inverse relationship between magma reservoir pressure and magma compressibility (see Section S1.1; e.g., Blake, 1981; Le Mével et al., 2016).

4. Conclusions

New numerical models focused on thermomechanical evolution indicate that the 2005 eruption of Sierra Negra volcano was likely triggered by a M_w 5.4 earthquake that preceded the event. Models of the stress evolution during the observed 18-year inflation event of >4.5 m of inflation suggest that Sierra Negra's magma system was in a stable storage configuration prior to its eruption in 2005 with ~11 MPa of accumulated overpressure and no indication of tensile failure along the reservoir boundary. However, the host rock accommodating the inflation had accumulated significant strain resulting in several earthquake/faulting events culminating in the M_w 5.4 earthquake, which triggered the eruption 3 hr later. Furthermore, estimates of temperature variation, which are modest and localized, suggest that a volatile overpressurization trigger is unlikely. The findings at Sierra Negra provide critical advancements for models of magma system stress state and system behavior in the lead up to eruptive events. This work has significant implications for the assessment of future unrest episodes, including the Sierra Negra's 2018 eruption, as well as forecasting magma system stability at other active volcanic systems.

Acknowledgments
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