Improved constraints on the estimated size and volatile content of the Mount St. Helens magma system from the 2004–2008 history of dome growth and deformation

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[1] The history of dome growth and geodetic deflation during the 2004–2008 Mount St. Helens eruption can be fit to theoretical curves with parameters such as reservoir volume, bubble content, initial overpressure, and magma rheology, here assumed to be Newtonian viscous, with or without a solid plug in the conduit center. Data from 2004–2008 are consistent with eruption from a 10–25 km^3 reservoir containing 0.5–2% bubbles, an initial overpressure of 10–20 MPa, and no significant, sustained recharge. During the eruption we used curve fits to project the eruption's final duration and volume. Early projections predicted a final volume only about half of the actual value; but projections increased with each measurement, implying a temporal increase in reservoir volume or compressibility. A simple interpretation is that early effusion was driven by a 5–10 km^3, integrated core of fluid magma. This core expanded with time through creep of semi-solid magma and host rock. **Citation:** Mastin, L. G., M. Lisowski, E. Roeloffs, and N. Beeler (2009), Improved constraints on the estimated size and volatile content of the Mount St. Helens magma system from the 2004–2008 history of dome growth and deformation, *Geophys. Res. Lett.*, *36*, L20304, doi:10.1029/2009GL039863.

1. Introduction

[2] Idealized models of reservoir-conduit systems predict effusive eruptions to be characterized by a generally monotonic decline in eruption rate [*Woods and Huppert*, 2003]. In these idealized systems, the duration and final volume reflect the size, initial overpressure, and compressibility of the magma-conduit system. Thus, measuring the duration and final volume should allow inferences to be made about these properties. But in general, dome growth is more complex than these models predict. Many domes grow in spurts ranging from months to years, with explosive activity that can vary over the course of an eruption. Even eruptions with simple monotonic effusion rates frequently deviate from theoretical predictions [*Stasiuk et al.*, 1993]. By contrast, we show in this paper that the 2004–2008 Mount St. Helens eruption history fits reasonably well with simple theoretical predictions.

[3] During this eruption we derived physically based models that predict histories of dome growth and of reservoir deflation with time, and compared their predictions with observed developments [*Mastin et al.*, 2008]. Now that the eruption has ended, we can test some of our predictions and derive new constraints on the size, volatile content, and other properties of the Mount St. Helens magma system.

2. Mount St. Helens Eruption and Measurements

[4] Mount St. Helens is the most active volcano in the Cascade Range of North America. Following eruptions in the late 1400s, mid 1600s and early 1800s, the volcano was dormant from 1857 until the spring of 1980. The 18 May 1980 eruption was followed by six years of dome growth, then eighteen years of quiescence, and four more years of dome growth from 2004–2008. Successive photogrammetric surveys documented growth of the recent dome to a final volume of 93±4 M m^3; nearly equal to that of the 1980–86 dome [*Schilling et al.*, 2008]. An additional 10±0.4 M m^3 of cold crater-floor material was pushed out ahead of the lava, making a final extruded volume V_e^f of about 103±4 M m^3. The intact lava contains only a few volume percent vesicles [*Pallister et al.*, 2008]. Assuming that <30% of the dome consisted of talus with a porosity of 30%, the dense-rock-equivalent (DRE) erupted volume would be within 10% of the dome volume. While the 1980–86 lava dome grew mostly in spurts lasting weeks, the 2004–2008 growth was continuous, decreasing monotonically from >6 m^3/s in October of 2004 to zero in 2008. The lava emerged as a nearly holocrystalline solid mantled by fault gouge, building spines and whalebacks up to 250 m high by early 2005. Subsequent growth and collapse kept the dome height between 150 and 250 m through the end of the eruption.

[5] Thirteen continuous GPS stations recorded surface displacements associated with deflation of the magmatic system [*Lisowski et al.*, 2008]. One continuous GPS instrument (JRO1, Figure 1) was operating at the start of the eruption; others were installed within a month thereafter. Their results suggest a vertical cigar-shaped source reservoir with a top ~4–7 km below the crater floor, a base deeper than 10 km, and a final volume loss in the reservoir ΔV_C^f of −19 to −32 M m^3—less than one third of the extruded volume. Mineral equilibrium studies [*Rutherford and Devine*, 2008] and seismicity [*Moran*, 1994] also suggest a reservoir top near 5 km, while the petrology of the 1980 products suggests a source region at 8–10 km depth [*Rutherford et al.*, 1985]. *Gerlach et al.* [2008] estimate the pre-eruptive volatile content and use the solubility model of *Newman and Lowenstern* [2002] to infer that the 2004–2008 magma contained about 1.2 volume percent bubbles in the 8–10-km-deep source region. If the gas released during this eruption emanated from a volatile-rich

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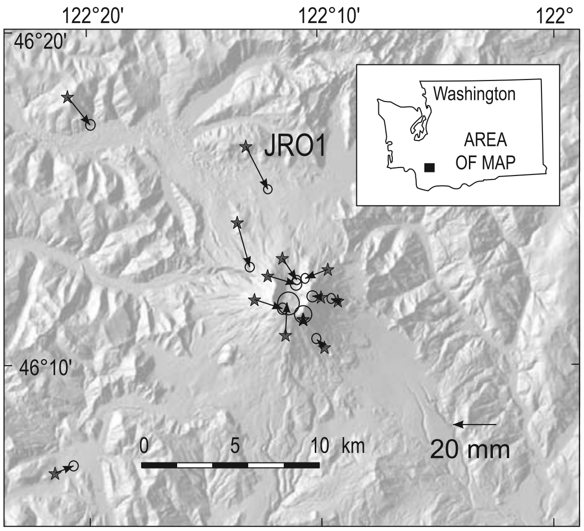

Figure 1. Map of the location of the continuous GPS stations around Mount St. Helens. Arrows indicate the horizontal displacement during the 2004–2008 eruption. The ellipses around each arrowhead indicate one standard error in measurement.

cap, the average reservoir bubble content below this cap could be less than the estimated 1.2%.

2.1. Question of Recharge

[6] The large discrepancy between final erupted volume V_e^f and reservoir deflation volume ΔV_C^f was initially thought to suggest that the reservoir was being recharged from depth at a low but sustained rate during extrusion. But extensive sampling uncovered no petrologic evidence for new or mingled magmas [*Pallister et al.*, 2008]. Two other considerations now lead us to think that this contribution was small. First, much of the discrepancy can be accounted for by expansion of bubbly magma remaining in the reservoir. Assuming no recharge, a magma of compressibility κ_m, and a reservoir consisting of a spheroidal void space with a linear (e.g., elastic) relationship between internal pressure and volume characterized by compressibility $\kappa_C \equiv (1/V_C)(\partial V_C/\partial p)$, the erupted volume V_e (DRE) at any given time is related to deflation volume ΔV_C (DRE) by [*Mastin et al.*, 2008]:

$$\frac{V_e}{\Delta V_C} = -\left(1 + \frac{\kappa_m}{\kappa_C}\right) \tag{1}$$

Using $\kappa_m = 2.5\text{–}5\times10^{-10}$ Pa^{-1} based on inferred bubble content, and $\kappa_C \approx 2\times10^{-11}$ Pa^{-1} based on seismic p-wave velocity (Table 1), we would calculate a volume ratio of -13.5 to -26–even greater than the inferred -3 to -5. One or both compressibilities must be adjusted, but it is clear that their difference can account for the disparate volumes.

[7] Secondly, sustained recharge seems unlikely based on the similarity of the dome growth and reservoir deflation curves (Figure 2). Assuming (1) a linear relationship between reservoir overpressure and extrusion rate, characterized by Poiseuille flow and flow of certain plugs with frictional or Newtonian margins [*Mastin et al.*, 2008], and

Table 1. Values of Some Variables, With Explanations

Variable	Value, Pa^{-1}	Explanation
κ_m	$2.5\text{–}5\times10^{-10}$	Value of bulk modulus [*Mastin et al.*, 2008] for a magma containing 0.5–2% bubbles at p = 220 MPa, the pressure of the magma source based on phase equilibrium studies of the 18 May 1980 magma [*Rutherford et al.*, 1985].
κ_C	2×10^{-11}	Reservoir compressibility calculated from the formula $\kappa_C = 3/4\ \mu$ for a spheroidal source [*McTigue*, 1987], using shear modulus μ = 40 GPa, which is calculated from the seismic p-wave velocity of 6.7 km/s at that depth [*Musumeci et al.*, 2002], density of 2700 kg m^{-3} [*Williams et al.*, 1987] and a Poisson's ratio of 0.25, using the formula [e.g., *Rubin*, 1990] $\mu = \rho_r v_p^2 (1 - 2\nu)/(2(1 - \nu))$.

(2) a constant rate of recharge c into the reservoir, we would predict the following volume changes with time t for the lava dome and reservoir:

$$V_e = a\left(1 - e^{-bt}\right) + ct, \tag{2}$$

$$\Delta V_C = -\frac{\kappa_C}{(\kappa_m + \kappa_C)} a\left(1 - e^{-bt}\right), \tag{3}$$

where a is the erupted volume not fed by recharged magma; b is the reciprocal time required for the dome to reach $(e-1)/e$ of its final volume; and c is the recharge rate. For Poiseuille flow, $a = V_C\ (\kappa_m + \kappa_C)\ p_{ex}^0$ and $b = \pi R^4/(8\eta\ HV_C\ (\kappa_m + \kappa_C))$, where p_{ex}^0 is initial reservoir pressure in excess of $\rho_m\ gH$, η is magma viscosity, H and R are conduit length and radius, respectively, g is gravitational acceleration, and ρ_m is mean magma density in the conduit. For flow of a

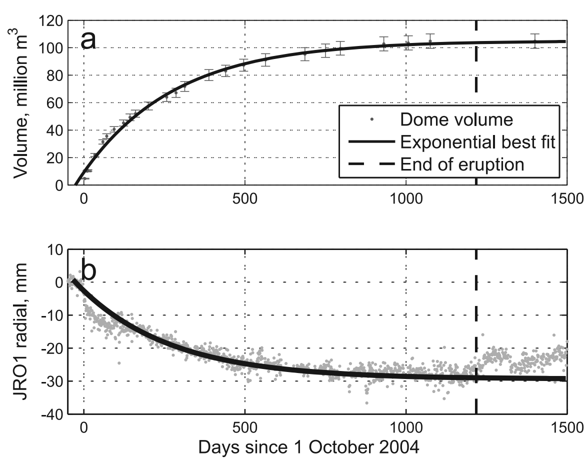

Figure 2. (a) Volume of dome lava plus cold rock extruded before 13 October 2004 as a function of time since 1 October 2004 and (b) outward displacement radial to the volcano measured at GPS station JRO1 (Figure 1). Error bars in dome volume are ±5%. To convert reservoir volume change ΔV_C in equation (3) to radial displacement at JRO1, we multiplied ΔV_C by 1.379×10^{-6} mm/m^3, the displacement per cubic meter volume loss at this location for an ellipsoidal source extending from 4 to 15 km depth [*Mastin et al.*, 2008].

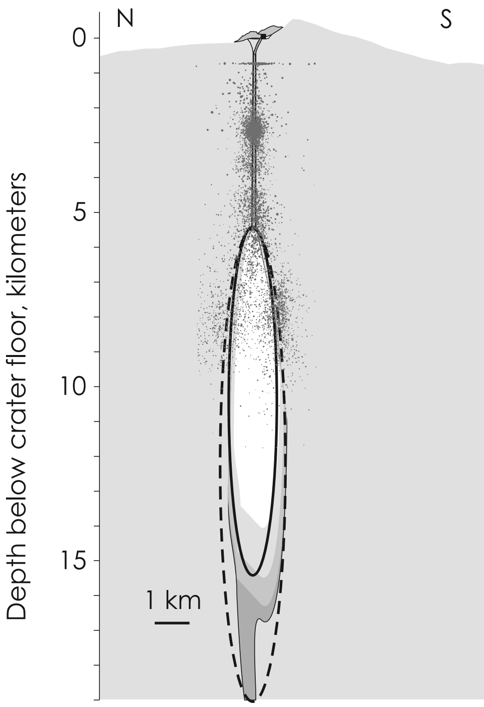

Figure 3. Cross section of the magma system and prolate ellipsoids of 10 km^3 (solid) and 25 km^3 (dashed) volume. Also shown are earthquake hypocenters from the University of Washington catalog between 1 January 1987 and 22 September 2004 within 2 km in plan view of the Mount St. Helens lava dome, which were detected by 7 or more stations with an azimuthal gap less than 135°. No vertical exaggeration.

solid plug through the conduit surrounded by a Newtonian annulus, the formula for a is the same, but that for b is slightly different [*Mastin et al.*, 2008].

[8] Equation (3) implies that the reservoir volume or its proxy, a geodetic displacement measurement, should exponentially approach a final, constant value; while (2) suggests that the dome volume should exponentially approach a constant rate of growth. If c>0, dome growth would have continued long after deflation stopped, which does not appear to be the case. Setting c=0, $(\kappa_C + \kappa_m)/\kappa_C$=4 (from (1)), $a = V_e^f$ = 103 M m^3, and a best-fit value of b=4.05×10^{-8} s^{-1}, the resulting curves (Figure 2) provide a good fit to the data.

2.2. Constraints on Estimates of Size, Overpressure, and Gas Content

[9] Equations (1), (2), and (3) provide the following constraints on the estimated size, initial overpressure, and volatile content of the reservoir:

2.2.1. Gas Content

[10] From equation (1), the ratio $V_e^f/\Delta V_C^f \approx 4$ implies $\kappa_m/\kappa_C \approx 3$, yet we would estimate a much larger ratio using κ_m =2.5–5×10^{-10} Pa^{-1} based on an estimated average 0.5–2% magma bubble content and κ_C =2×10^{-11} Pa^{-1} based on p-wave velocities at that depth (Table 1). In order to bring κ_m/κ_C in line with the observed volume ratio while holding κ_C constant, we would have to reduce κ_m to ∼6×10^{-11} Pa^{-1}, which is lower than that for this bubble-free magma (∼2×10^{-10} Pa^{-1} [*Mastin et al.*, 2008]). Such a low compressibility could be achieved only if more than 70% of the deforming volume consisted of holocrystalline host rock with negligible compressibility, which seems unrealistic. Moreover, field studies of intrusions [*Rubin*, 1995] find that host-rock stiffness of large, shallow rock masses is commonly 5 to 10 times less than predicted from seismic p-wave velocities, suggesting that an increase in κ_C is in order. Maintaining κ_m = 2.5–5×10^{-10} Pa^{-1} while raising κ_C to 0.8–1.5×10^{-10} Pa^{-1} would be consistent with this observation.

2.2.2. Reservoir Overpressure

[11] The growth of the lava dome to a maximum height of about 250 m would have imposed a back-pressure on the vent of several (6–7) megapascals. The fact that the eruption continued after the dome reached this height suggests that p_{ex}^0 exceeded this value. The maximum overpressure was likely constrained by hydrofracturing to be less than about 20 MPa, i.e. 5–10 MPa above lithostatic pressure at 5 km depth, assuming ρ_m =2400–2500 kg m^{-3} and host-rock density=2700 kg m^{-3} [*Williams et al.*, 1987].

2.2.3. Reservoir Volume

[12] The value $a = V_C\, p_{ex}^0\, (\kappa_m + \kappa_C)$ = 103 M m^3, combined with κ_m =2.5–5×10^{-10} Pa^{-1}, κ_C = 0.8–1.5×10^{-10} Pa^{-1}, and p_{ex}^0 =10–20 MPa, implies a reservoir volume V_C of about 8 to 31 cubic kilometers, with values in the middle of this range, around 10–25 km^3, most likely. This volume is a few to several times larger than the ∼4 km^3 DRE of the largest Holocene Mount St. Helens eruption [*Carey et al.*, 1995]. A 10 km^3 prolate ellipsoidal reservoir (solid ellipse, Figure 3) whose top is at 5.5 km

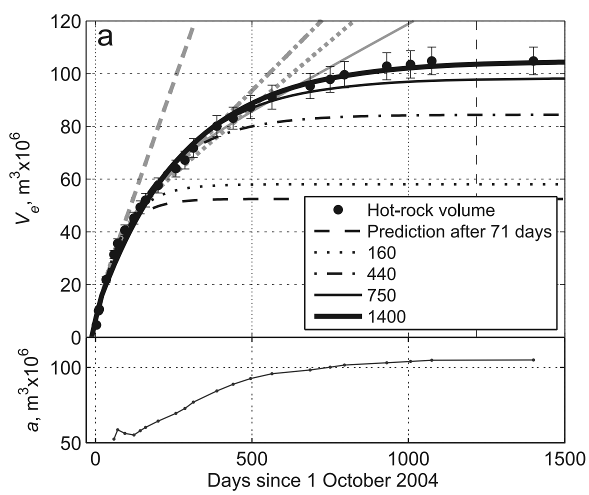

Figure 4. (a) Extrapolations of best-fit curves based on volume data available at different times since the eruption onset, as given in the legend. Gray lines are from equation (2) with best-fit values of a, b, and c. Black lines are also from equation (2) but with c=0. Black data points and error bars are dome volume as given in Figure 2a. (b) Best-fit values of a versus time for the case where c=0. For the recharge-free curves, root-mean-square differences between the volumes measured (V_e^m) and predicted (V_e^p), i.e., $\sqrt{\Sigma\left(V_e^m - V_e^p\right)^2/N}$ (where N is the number of measurements), are 0.33 M, 0.56 M, 2.24 M, 2.47 M, and 2.45 M m^3 for curves derived 71, 160, 440, 750, and 1400 days after 1 October 2004, respectively.

depth and center about 10 km fits roughly within the cloud of seismicity thought to define the top of the reservoir [*Moran*, 1994]. A 25 km^3 reservoir with similar top depth and aspect ratio (dashed ellipse, Figure 3), also fits within this seismic cloud but contains a center a few kilometers deeper than the 8–10 km inferred source depth of the 1980 magma.

2.3. Misfits, and What They Tell Us About the Magma System

[13] During the eruption, dome volumes were periodically measured and growth curves extrapolated to forecast the eruption's final volume and time of cessation. Using (2) with recharge (c) set to 0, each best-fit curve that we acquired suggested a greater final eruptive volume a than the previous one (Figure 4). Random measurement errors would not have produced this systematic trend. Inaccurate, low volume estimates in the first month of the eruption when part of the lava dome was buried in the glacier may have contributed to this trend; but changes in a continued for more than two years. Moreover, analyses using (2) with an adjustable recharge rate c did not improve the prediction, but found the apparent recharge rate to decrease to zero with time.

[14] Because $a = V_C\, p_{ex}^0\, (\kappa_m + \kappa_C)$, these trends suggest that the magma system volume V_C or compressibility ($\kappa_m + \kappa_C$) was increasing with time, roughly doubling over a 40-month period. Numerical solutions to the growth curve using time-varying magma compressibility can account for only a small part of the increase in volume [*Mastin et al.*, 2008]. Another possible interpretation is that the early eruption was fed by a ~5–10 km^3 volume near the top of the reservoir that was hot and crystal-poor enough to be hydraulically integrated. More viscous, crystal-rich regions may have deformed slowly over subsequent years, enlarging the mobile portion of the system downward and outward (as illustrated by the shaded regions near the base of Figure 3). Continued adjustments are suggested by a hint of re-inflation since the end of the eruption (Figure 2b).

[15] A recharge pulse might also increase a with time. We would expect such a pulse to produce a geodetic re-inflation signal, whereas downward-propagating relaxation would produce continued deflation. JRO1 did record a decrease in deflation rate about 50–100 days after 1 October 2004 (Figure 2b); but the most rapid increase in a (Figure 4b) occurred at ~200–500 days, a time when JRO1 recorded inward displacement that matched its long-term deflationary trend (Figure 2b). These results would seem to argue against a recharge pulse as a cause of the prolonged eruption.

3. Discussion and Conclusions

[16] Our model assumes a value of κ_C consistent with elastic deformation over a length scale of kilometers. If wallrock deformation were significantly inelastic, κ_C could be higher, implying a smaller V_C or p_{ex}^0. Smaller V_C or p_{ex}^0 would also be implied if reservoir gas content were higher than estimated. But the maximum volume of large recent eruptions (4 km^3) still implies that V_C exceeds several cubic kilometers; and dome growth to 250 m height despite backpressure on the vent suggests that p_{ex}^0 exceeded several megapascals.

[17] In recent centuries, major eruptions such as the Wn (2 km^3) and We (0.4 km^3) have been separated in time by as little as two years [*Yamaguchi*, 1985]. The 93 M m^3 1980–86 lava dome started its growth shortly after the ~0.5 km^3 eruption of 18 May 1980. These events reflect a resilient, dynamic system whose size and properties may vary over decades. Hints of these properties provided by the 2004–2008 eruption represent only the current state of the system.

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