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The effect of pressurized magma chamber growth on melt migration and pre-caldera vent locations through time at Mount Mazama, Crater Lake, Oregon

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ABSTRACT

The pattern of eruptions at long-lived volcanic centers provides a window into the co-evolution of crustal magma transport, tectonic stresses, and unsteady magma generation at depth. Mount Mazama in the Oregon Cascades has seen variable activity over the last 400 ky, including the 50 km^3 climactic eruption at ca. 7.7 ka that produced Crater Lake caldera. The physical mechanisms responsible for the assembly of silicic magma reservoirs that are the precursors to caldera-forming eruptions are poorly understood. Here we argue that the spatial and temporal distribution of geographically clustered volcanic vents near Mazama reflects the development of a centralized magma chamber that fed the climactic eruption. Time-averaged eruption rates at Mount Mazama imply an order of magnitude increase in deep magma influx prior to the caldera-forming event, suggesting that unsteady mantle melting triggered a chamber growth episode that culminated in caldera formation. We model magma chamber–dike interactions over ~50 ky preceding the climactic eruption to fit the observed distribution of surface eruptive vents in space and time, as well as petrologically estimated deep influx rates. Best fitting models predict an expanding zone of dike capture caused by a growing, oblate spheroidal magma chamber with 10–30 MPa of overpressure. This growing zone of chamber influence causes closest approaching regional mafic vent locations as well as more compositionally evolved Mazama eruptions to migrate away from the climactic eruptive center, returning as observed to the center after the chamber drains during the caldera-forming eruption.

1. Introduction

Long-lived volcanic centers commonly exhibit a wide variety of eruptions, including those that emanate from a centralized volcanic edifice as well as from regionally scattered monogenetic vents and shield volcanoes. The pattern of eruptions is generally highly variable in space and episodic in time: even from a single volcanic edifice, eruption style may vary widely in magnitude and intensity. Physical controls on the spatial and temporal organization of eruptions are poorly constrained. At some volcanic centers, a long history of repeated activity includes explosive eruptions of sufficient size to cause collapse of the central edifice, producing a caldera. Evaluation of which volcanoes are capable of caldera-forming eruptions (thereby defining the maximum potential hazard at a given center; Marzocchi and Bebbington, 2012) requires finding evidence for the presence of a large volume of eruptible magma at depth. Such evidence may be found in long-term patterns of eruptive activity that precede caldera collapse, outputs of the subsurface crustal magma transport network.

Volcanically active regions also exhibit spatially variable eruption patterns, such as the common along-strike volcanic vent density variation observed in active volcanic arcs (Siebert et al., 2011). This spatial localization is inexorably tied to temporal variability in eruption style and composition. A variety of mechanisms have been proposed for focusing of melt towards volcanic centers, all of which fall into three general classes of models for volcano spacing: "bottom up" models propose localization in the melt source region through buoyancy instabilities (e.g., Marsh and Carmichael, 1974; Olson and Singer, 1985), "top down" models propose dike focusing through stress interactions with surface edifice loads (e.g., Muller et al., 2001), while "internal" models rely on stress interactions between magma chambers – storage zones in which melt fractions are high enough that pressure gradients are homogenized by flow – and dikes to generate spatially discrete volcanic centers (e.g., Karlstrom et al., 2009).

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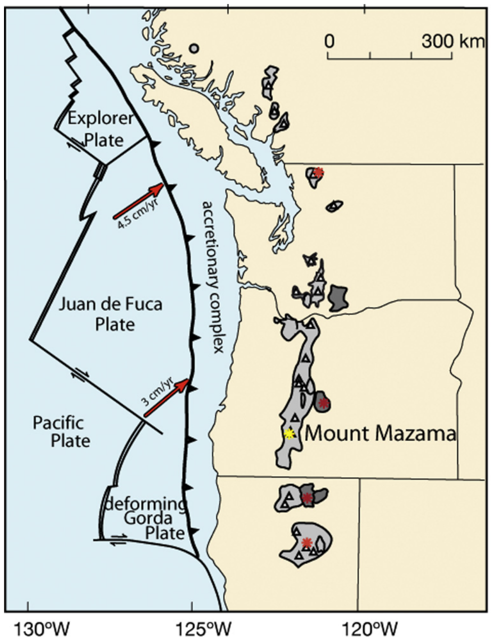

Fig. 1. Location of Mount Mazama and Crater Lake caldera in the Cascades chain (yellow asterisk). Other identified Quaternary calderas are shown with red asterisks. Modified after Hildreth (2007). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Here we use the eruptive history at Mount Mazama, Oregon (Fig. 1; Bacon and Lanphere, 2006; Wright et al., 2012), to probe evolving subsurface architecture of magma transport leading to the Mazama climactic caldera-forming eruption at ca. 7.7 ka (CFE hereafter) through application of a mechanical model for the plumbing system. Trends in erupted compositions and volumes, along with the caldera structure itself, indicate that an evolving magma chamber played a key role in the style and distribution of surface volcanism. Therefore we focus efforts on the "internal" class of models described above, using known locations and timing of eruptive episodes over 55 ky along with petrologically based estimates of melt influx to constrain models for silicic chamber growth and dike capture. Although the idea that mafic magma may be trapped or stalled by a shallow, less dense magma reservoir is not new (e.g., Walker, 1974; Smith and Shaw, 1975; Hildreth, 1981; de Silva, 1989), here we hypothesize that overpressure within the growing climactic chamber generates a deviatoric stress field that reorients more distant rising dikes (Karlstrom et al., 2009, 2010) to control the rate of chamber growth and surface eruption locations.

2. Mount Mazama eruptive history

The geologic record of Mount Mazama includes remarkable exposures of volcanic sequences in the caldera walls at Crater Lake that offer a rare opportunity to see otherwise buried deposits. Volcanic activity at Mount Mazama has persisted for >400 ky, with andesite and low-silica dacite lava flows dominating the eruptive products from the central volcano (Bacon and Lanphere, 2006; Bacon, 2008). The appearance of silicic dacites and rhyodacites occurs late in the eruptive sequence (71 ka and 27 ka, respectively), culminating in the 50 km^3 zoned eruption of Mount Mazama (CFE) that led to caldera collapse ca. 7.7 ka (Bacon, 1983; cf. 7627 ± 150 cal. yr. BP age of Zdanowicz et al., 1999).

For the present study we have compiled a database of Mazama volcanism that includes compositions, ages, and volumes of eruptive units along with vent locations (additional discussion of mapping and data selection in the Supplementary Material). Fig. 2 shows the distribution of eruptive vent locations for the last 400 ky, as a function of map distance (in km) from the caldera center for the Mazama climactic eruption, along with corresponding volumes (symbol size) and *maximum* SiO$_2$ contents (symbol color). Eruptions with increasing maximum silica content through time (Bacon and Lanphere, 2006; their Fig. 7) and a weak correlation between erupted volume and maximum silica content for the entire dataset (correlation coefficient 0.229, p-value 0.029) imply increased crustal storage in time (e.g., Bacon and Druitt, 1988).

This work focuses on the last 55 ky of Mazama eruptive history leading up to CFE, in which vents appear progressively farther from the climactic caldera center (Fig. 2.B). Starting with near-summit domes of the dacite of Munson Valley and the mingled lava of Williams Crater (both ∼35 ka), evolved silicic eruptions migrate steadily away from the climactic center, with some radial dike controlled patterns that reflect shallow fault control (Bacon, 2008), until two substantial rhyodacitic units (Llao Rock preclimactic rhyodacite and Cleetwood preclimactic rhyodacite) erupted near the climactic center decades before CFE. These eruptions are interpreted by Bacon (1983) as failed attempts or precursors to the climactic eruption. Nowhere prior in the Mazama history do regional eruptions and evolved silicic eruptions spatially migrate away from a large volume eruption (and CFE is the largest eruption recorded). Furthermore, silicic vents appear ∼6 km farther away from the climactic center during this period than elsewhere in the eruptive history (Fig. 2.A). Closest approaching regional mafic eruptions occur farther from the center than the evolved Mazama eruptions. After the caldera-forming event, eruptive activity again appears close to the climactic vent location.

2.1. Mazama melt influx estimates

We model melt evolution at Mazama using the Rhyolite MELTS program (RMELTS; Gualda et al., 2012). The initial magma composition in these petrologic simulations is equivalent to the whole rock composition of basaltic andesite of Red Cone (unit abbreviations br and brp; Bacon, 2008), a monogenetic volcano northwest of Crater Lake taken as a proxy for primary compositions of magmas entering the Mazama plumbing system. Additional discussion of these samples and RMELTS models may be found in the Supplementary Material.

The range of erupted magma compositions at Mount Mazama reflects a combination of fractional crystallization, assimilation and recharge (Bacon and Druitt, 1988; Bacon et al., 1994; Bacon and Lanphere, 2006; Wright et al., 2012). However much of this compositional variation can be explained as due to crystallization from parental melt (Bacon and Druitt, 1988). Therefore, we do not include magma mixing, recharge, or assimilation in RMELTS models. We recognize that recharge likely has important effects on magma chamber compositions in long-lived reservoirs such as at Mazama (e.g., Eichelberger, 1974). Our goal is not to model the details of Mazama compositional evolution but rather general trends, for which a melt mass fraction-composition parameterization based on fractional crystallization is sufficient.

We assume isobaric crystallization at 3 kbar with initial H$_2$O = 1 wt%, fitting compositional data to find a polynomial relation between melt mass fraction and wt% SiO$_2$ (Supplementary Material Fig. S1)

$$\begin{aligned}\%\,\text{melt} = {} & 48809.2 - 2973.58X_{\text{SiO}_2} + 67.932X_{\text{SiO}_2}^2 \\ & - 0.6888X_{\text{SiO}_2}^3 + 0.0026X_{\text{SiO}_2}^4, \qquad (1)\end{aligned}$$

where X_{SiO_2} is the weight percent SiO$_2$ in the melt.

Eruptive volumes and ages are then corrected via Eq. (1) to infer primary melt influx (Fig. 3) using a moving average over a