Did magma ascent rate control the explosive-effusive transition at the Inyo volcanic chain, California?

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ABSTRACT

Rhyolitic eruptions often begin explosively and then shift to extrusions of lava. It is widely believed that the style of eruption is dependent on magma flow rate within the conduit, with fast ascent leading to explosive eruption and slow ascent favoring effusive behavior. Currently, the velocity of rhyolite magma flowing in a conduit is unknown, and therefore eruption models remain untested against geologic evidence of magma ascent. We show that explosively erupted rhyolite magma from the Inyo volcanic chain ascended slowly (centimeters per second), at the same rates as the dome-building phases. We propose that the explosive-effusive transition in sub-Plinian rhyolite eruptions is governed by the alleviation of exsolved volatile pressure in the shallow conduit rather than the magma flow rate.

Keywords: microlites, pumice, magma ascent, decompression experiments.

INTRODUCTION

Volcanoes often display a pattern of activity beginning with column-forming explosive eruptions followed by lava extrusions. Understanding these eruptive transitions is important for volcanic hazards assessment. Eichelberger and Westrich (1981) suggested that the explosive-effusive pattern reflects stratification of dissolved H_2O within the magma source; hydrous pyroclasts were believed to represent the volatile-rich top of the magma body, whereas degassed lavas originated from deeper, drier magmas. It turns out that many silicic magmas that erupted explosively and effusively had the same initial volatile contents, and therefore must have experienced different degassing histories (e.g., Eichelberger et al., 1986). To explain the explosive-effusive transition in the absence of volatile gradients, many propose that explosive magmas degas in a closed-system manner (e.g., Newman et al., 1988), in that gas bubbles remain in contact with the melt until it fragments. By contrast, lavas undergo open-system degassing, in that exsolved volatiles separate from the melt through permeable magmatic foam (Eichelberger et al., 1986) or brecciated magma (Gonnermann and Manga, 2003).

The interdependence of eruption style and degassing mode is thought to be governed by magma ascent rate and its influence on degassing efficiency (e.g., Gonnermann and Manga, 2007). In Plinian eruptions, the explosive phase is caused by fast (meters per second) magma ascent in which bubbles remain coupled to the melt; this prevents degassing on the ascent time scale, and the magma fragments due to volatile overpressure, bubble crowding, and rapid acceleration (e.g., Sparks, 1978; Dingwell, 1996). Lava extrusion follows when the magma slows within the conduit as magma chamber overpressure declines (e.g., Woods and Koyaguchi, 1994) and conditions for fragmentation subside (Papale, 1999).

The extent to which magma ascent rate dictates eruptive transitions in smaller sub-Plinian rhyolite eruptions is unknown. In this study, we evaluate whether magma ascent controlled the explosive-effusive transition in a sub-Plinian rhyolite eruption along the Inyo volcanic chain, California (Miller, 1985). We do this by experimentally reproducing tiny (<100 μm) feldspar and clinopyroxene crystals found in Inyo pumice pyroclasts (Fig. 1). Because these microlites grew in response to degassing during magma ascent (Swanson et al., 1989), they record the magma ascent history in their textures and compositions. We have conducted decompression experiments (e.g., Hammer and Rutherford, 2002) in order to reproduce these features and thus quantify the rise rate of explosively erupted magma.

Our results indicate that despite erupting explosively, the Inyo magma ascended slowly, at rates similar to later effusive eruptions. We propose that initial explosivity was caused by closed-system accumulation of bubbly rhyolite in the conduit, followed by rapid pressure release. Continued slow magma ascent after rapid decompression fostered the explosive-effusive transition.

ERUPTIONS OF THE INYO VOLCANIC CHAIN, CALIFORNIA

The Inyo volcanic chain was active ca. 550 yr B.P. when four sub-Plinian eruptions occurred from three aligned vents, reflecting an underlying, segmented feeder dike (Fink, 1985). These events produced fall deposits composed of white pumice lapilli (~95%) and juvenile pyroclastic obsidian and lithic fragments (~2%–5%). Pyroclastic flows occurred after column-producing eruptions from the southernmost Deadman vent (Miller, 1985), and small surge and block-and-ash flow deposits near the northernmost Obsidian Dome vent may mark a brief period of dome collapse or phreatic activity soon after the explosive phase. Three obsidian domes extruded from the same vents that fed the pyroclastic eruptions (e.g., Miller, 1985) at the conclusion of the eruption sequence.

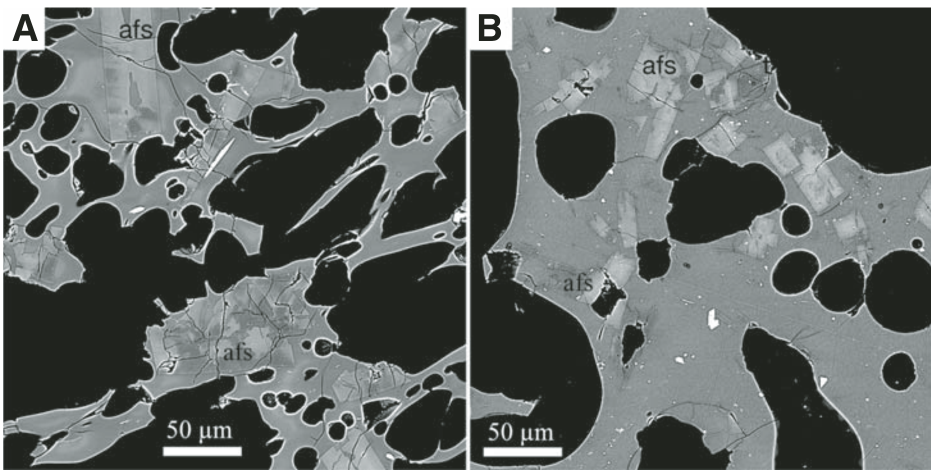

Figure 1. Backscattered electron (BSE) photomicrographs of an Inyo pumice pyroclast (A) and an experimental pumice produced by decompression of H_2O-saturated Inyo rhyolite from 200 to 4.4 MPa at 0.68 MPa h^{-1} and a temperature of 750 °C (B). Blocky phases are alkali feldspar (afs); bright phases are Fe-oxide and clinopyroxene.

DECOMPRESSION-INDUCED CRYSTALLIZATION

Inyo pyroclasts contain variable quantities (~1–37 vol%) of alkali feldspar microlites and uniformly dilute amounts (<3 vol%) of clinopyroxene, biotite, and Fe-oxide microlites (Castro and Mercer, 2004) (Fig. 1). Castro and Mercer (2004) quantified feldspar and clinopyroxene microlite characteristics in obsidian pyroclasts and identified different textural types based on the relative proportions of feldspar and clinopyroxene. The microlite textures of pumice pyroclasts were not examined, and because these are the volumetrically dominant component of the pyroclastic phase, we have performed additional textural measurements on pumice in order to establish a representative natural baseline for comparison with experiments. Measurements and experiments were conducted on pyroclasts collected from a fall deposit that originated from the Obsidian Dome vent, which is exposed ~1 km east of Obsidian Dome. The feldspar mode, number density, and compositions were measured on five pumice and four obsidian clasts; these samples are representative of the wide range of microlite textures examined in ~30 pumice and 50 obsidian clasts from the sample site and in fall deposits from the other vents (Castro, unpublished data). We also measured the size distributions of alkali feldspar and clinopyroxene microlites on a subset of these clasts. Detailed methods are described in the Appendix.[1]

In order to quantify the ascent rate of the Inyo magma, we conducted isothermal, H_2O-saturated ($P = P_{H_2O}$) decompression experiments on the Inyo rhyolite (Table DR1 in the GSA Data Repository[1]). The starting conditions for decompression experiments (200 MPa, 750 °C) were guided by feldspar geothermometry, having a precision of ±30 °C (Gibson and Naney, 1992), the upper stability limit of biotite, and the H_2O contents of glass inclusions (Hervig et al., 1989). At these conditions, the Inyo melt is saturated in clinopyroxene, biotite, and Fe-oxide. The saturation pressure, composition, and mode of alkali feldspar as a function of P, in addition to the stability field of biotite, were determined in several H_2O-saturated phase equilibrium experiments (Table DR1). Experiments above the clinopyroxene liquidus (~300 MPa at 750 °C) were not possible due to the mechanical limitations of the pressure vessels. Detailed experimental and analytical methods are described in the appendix.

All experiments were held at the starting conditions for four to six days and then decompressed at different rates (2.7–0.33 MPa h^{-1}) to the final pressure (P_f = 4.4 MPa). We also quenched several experiments at P_f > 4.4 MPa, in order to track changes in feldspar texture throughout the decompression interval.

EXPERIMENTAL RESULTS

Tabular alkali feldspar microlites were the primary crystallization product in decompression experiments (Fig. 1B; Table DR1). Alkali feldspar also crystallized on resorbed plagioclase grains, which represent a small fraction (<5 vol%) of phenocryst fragments in the starting powders. This overgrowth texture mimics ubiquitous anti-rapakivi textures in the natural pumice (Gibson and Naney, 1992). Experiments also produced clinopyroxene, Fe-oxide, and biotite microlites in low abundances (Table DR1).

The feldspar mode depended on the decompression rate, with more feldspar being produced at progressively slower rates (Fig. 2A). The range of feldspar modes produced at rates from 1.36 to 0.33 MPa h^{-1} overlaps the variability in Inyo pyroclasts. Experiments at 2.7 and 2.0 MPa h^{-1} produced very little feldspar (<1 vol%). Feldspar number density also varied with decompression rate (Fig. 2B); these values plot at the low end of what is observed in pyroclasts.

Feldspar crystal size distributions (CSDs) of the three slowest decompression experiments are roughly linear, and their extrapolated intercepts increase with decreasing decompression rate, reflecting an overall increase in number density and crystallinity with time (Table DR2; Fig. 2C). There is very good agreement between two pumice CSDs and the experiments at 0.68 and 0.33 MPa h^{-1}. The CSD of the third pumice sample overlaps a majority of the size scale of the experimental CSDs but departs in the smallest two size classes. There is no match between experiments and the obsidian pyroclast CSDs.

Experimental feldspar compositions match those in natural pyroclasts (Fig. 3) and reflect the equilibrium compositions. There is apparently no decompression rate dependence on feldspar composition between 0.33 and 0.68 MPa h^{-1}; the compositional ranges in these experiments are identical.

Clinopyroxene microlite textures produced in decompression experiments are broadly similar to those measured on pumice pyroclasts (Table DR3). The best textural match is produced by the 0.33 MPa h^{-1} experiment; it repli-

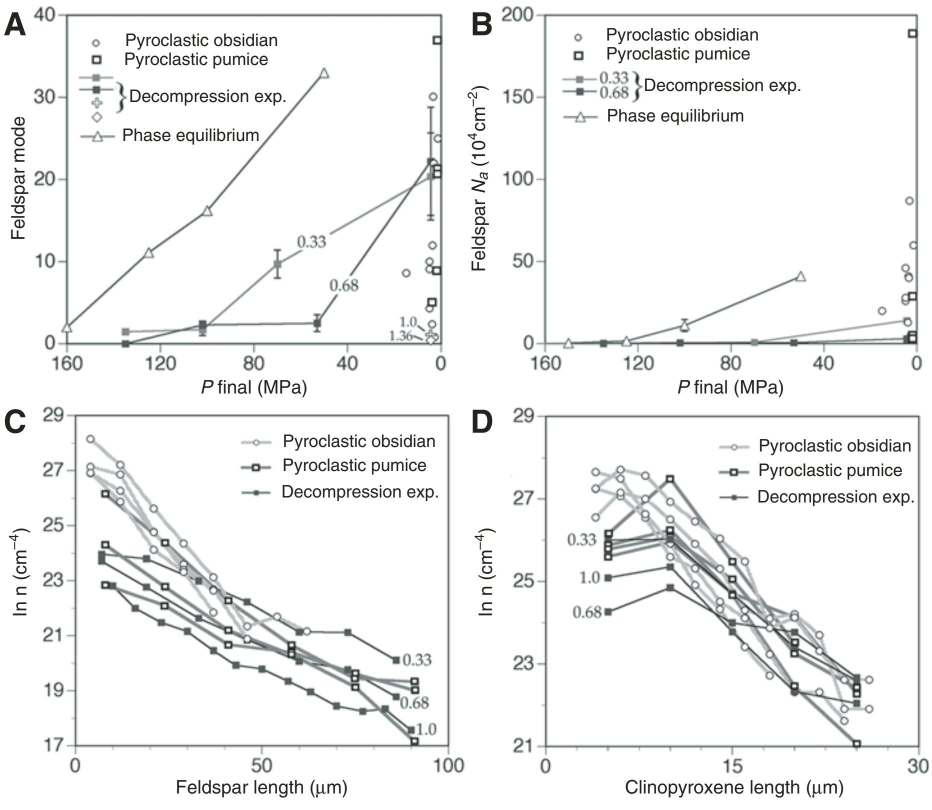

Figure 2. Microlite textures in experiments and Inyo pyroclasts. Numbers on experiment curves are the decompression rate (MPa h^{-1}). A: Feldspar mode in pyroclastic obsidians is from this study and Castro and Mercer (2004); pressure values are inferred from Castro and Mercer (2004). B: Feldspar area number density (N_a). C and D: Feldspar and pyroxene crystal size distributions (CSDs) in experiments and in pyroclasts. The ordinate is the natural log of the number of crystals per unit volume normalized to crystal size.

[1]GSA Data Repository item 2008068, Appendix (experimental and analytical methods), Table DR1 (summary of experiments), Table DR2 (feldspar CSD analysis), and Table DR3 (clinopyroxene textures and CSD data), is available online at www.geosociety.org/pubs/ft2008.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

cates both the number density and CSD found in three pumice samples (Fig. 2D). Experimental clinopyroxene CSDs do not fit the obsidian pyroclast CSDs, as they reflect much lower microlite number densities in the small size classes.

FORMATION OF OBSIDIAN AND PUMICE PYROCLASTS

The Inyo experiments simulate simple magma rise characterized by constant temperature and linear decompression. Despite these simplifications, the experiments reproduced many of the textural and compositional characteristics of microlites found in the natural pumice pyroclasts.

The departure between experiment and nature is evident in the CSDs of obsidian pyroclasts, which reflect higher numbers of small microlites than observed in the pumice and experiments. Because CSDs are the time-integrated product of microlite nucleation and growth, and may be modified after the initial nucleation event (e.g., Marsh, 1998), the disparate CSDs probably reflect processes operating during the formation of obsidian that did not hold for the pumice. For example, cooling and shearing at the conduit margins, where pyroclastic obsidian is believed to form (Rust et al., 2004), would result in a higher number density of small crystals and consequently a steep CSD (e.g., Bindeman, 2005). Because these factors were not modeled by decompression experiments, it is impossible to pinpoint the origin of the disparate CSDs. However, the parcels of magma that became obsidian and pumice must have experienced different ascent, degassing (Rust et al., 2004), and possibly cooling histories. Our experiments, therefore, are best used to interpret the rise conditions of the pumice pyroclasts.

MAGMA ASCENT RATE

Comparisons of experimental and natural feldspar textures can be used to estimate magma ascent rate. There is good agreement between textures and compositions of feldspar produced at decompression rates ranging from 0.33 to 1.36 MPa h^{-1} and those in pumice pyroclasts. Assuming an overburden density of 2.3 g cm^{-3}, the ascent rates corresponding to these decompression rates would range from ~0.4 to 1.7 cm s^{-1}. The 2.7 and 2.0 MPa h^{-1} experiments, by contrast, were too rapid to generate much feldspar and reproduce what is found in nature. The experiments at slower rates, therefore, establish a maximum ascent rate (i.e., ~1.7 cm s^{-1}) for the development of the feldspar textures in the Inyo rhyolite.

The experimental clinopyroxene textures, while similar to those in the natural pumice, cannot be used to estimate a threshold ascent rate, because all experiments contained clinopyroxene microlites at the onset of decompression; initial differences in these populations may have imparted some of the variability in experiments. These clinopyroxene data do, however, support the conclusions based on feldspar, given the relatively good overlap between experimental and natural clinopyroxene at slow decompression rates.

ERUPTION DYNAMICS AT INYO

The upper-bound ascent rate of explosive rhyolite (~2 cm s^{-1}) is slow and comparable to the average effusion rate (~1.6 cm s^{-1}) of the dome lavas (Castro et al., 2002). This result contradicts the view that "the style of eruption—explosive or effusive—is dependent on magma flow rate" (Woods and Koyaguchi, 1994), in that the magma supply rate apparently did not vary significantly between explosive and effusive stages. Thus, factors other than a fluctuation in ascent rate must have moderated the eruption style.

Presumably, a transition from an explosive to an effusive eruption occurs when conditions favoring fragmentation change or cease to exist. Two end-member fragmentation mechanisms have been proposed to operate during explosive eruptions (Cashman et al., 2000). One mechanism results from rapid ascent (meters per second) and high magma strain rates caused by acceleration and expansion of bubbly magma (Papale, 1999). Such conditions may cause brittle deformation (Dingwell, 1996) and late-stage disequilibrium exsolution of volatiles (e.g., Gonnermann and Manga, 2007).

A second fragmentation mechanism, typically ascribed to Vulcanian eruptions and dome-collapse events, involves sudden decompression of static, gas-pressurized bubbly magma. Rapid decompression of bubbly magma induces an unloading shock that causes brittle failure of bubble walls and gas expansion from those newly vented bubbles. The result is a fragmentation wave that propagates downward at several meters per second (e.g., Alidibirov and Dingwell, 1996).

As demonstrated by our experiments, the formation of feldspar microlites requires slow decompression and correspondingly low ascent rates from storage to surface. These facts argue against fragmentation by a rapid ascent mechanism, as fast ascent would not provide the requisite time for equilibrium degassing and undercooling required for microlite growth.

Fragmentation may have instead resulted from the sudden decompression of magma resident in the conduit prior to eruption. In this case, the slow decompression rates recorded in pyroclasts represent the slow supply of magma, possibly the intrusion of the Inyo dike, prior to the rapid evacuation of the Inyo feeder system. The physical characteristics of the shallow conduit, particularly a lack of permeability, may have hindered outgassing and allowed vapor pressure to increase. When the system vented, quite possibly by the propagation of a crack tip at the head of the rising dike (Reches and Fink, 1988), an unloading wave traveled downward through the system, fragmenting the vesicular magma, which would have then allowed trapped gasses to expand, accelerate, and exit the conduit (Alidibirov and Dingwell, 1996). Rhyolite melt will indeed fragment in response to sudden decompressions as small as 5 MPa if it contains a significant complement of bubbles (~60 vol%; Martel et al., 2001); such porosities are well within the range expected for H_2O-saturated magma residing in the upper 1 km of the conduit.

We postulate that fragmentation and degassing during sub-Plinian eruptions at Inyo were driven by unloading of a static gas-charged magma body. Viewed in this manner, explosive events at Inyo are mechanistically more akin to Vulcanian eruptions than Plinian eruptions, in that they are initiated from the top down and are fueled by shallow degassing and pressurization (Morrissey and Mastin, 2000).

The fact that the explosive phase drew on only a small volume of magma available in the Inyo

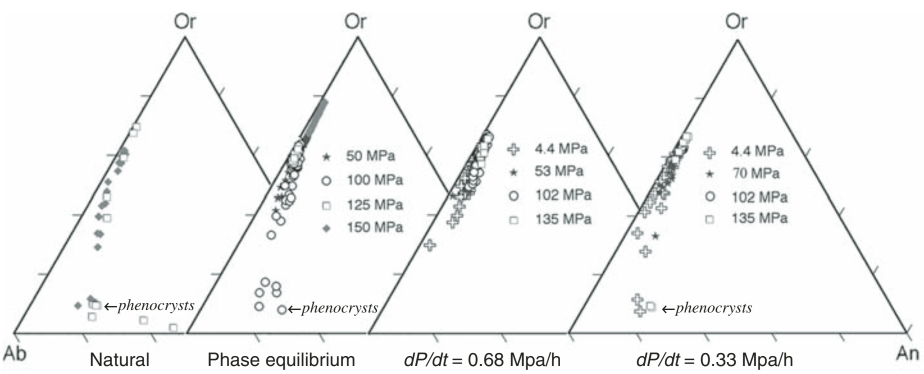

Figure 3. Feldspar compositions depicted as mole percent of orthoclase (Or), albite (Ab), and anorthite (An) components. Natural data are from this study (black diamonds) and from Gibson and Naney (1992) (open squares). Points near the Ab-An join are natural phenocrystic plagioclase fragments included in the experiment starting powders.

feeder dike supports our hypothesis that shallow processes governed explosivity. At the Obsidian Dome vent, for example, the explosively erupted magma would have filled the dike to a depth of ~1 km (Reches and Fink, 1988). This shallow depth implies, in turn, that the fragmentation depth was also limited to ~1 km, considering that fragmentation is a requirement for an explosive eruption. The maximum H_2O content (~1.8 wt%) of obsidian pyroclasts (Castro and Mercer, 2004) is evidence that fragmentation was limited to a shallow depth, as this H_2O content reflects a solubility pressure of ~28 MPa, and a corresponding depth of ~1.2 km.

If the explosive events at Inyo were driven by rapid decompression, then the explosive-effusive transition reflects the wholesale relief of *exsolved* volatile pressure and an inability for that pressure to build back up in the magma. This, in turn, implies that the pre-eruptive magma body was zoned with respect to bubble content, as this structure would provide not only the initial driving force for expansion, but also an ending point for fragmentation.

DEGASSING OF RHYOLITE MAGMA

A remarkable fact about many sub-Plinian rhyolite eruptions is that only ~10 vol% (dense rock basis) of the magma erupts explosively (Heiken, 1978); the rest is effusive. How then, does rhyolite magma degas? Recent studies suggest that degassing may occur by shear-induced fracturing of magma near conduit walls (e.g., Tuffen et al., 2003). Gonnermann and Manga (2003) show that fragmentation will occur over a broad range of conduit strain rates, provided that the melt viscosity (μ_{melt}) is high enough. They define strain rate as $Q/\pi R^3$, where Q is the volumetric flux (i.e., cross-sectional area × ascent rate) and R is the cylindrical conduit radius. The Obsidian Dome conduit radius varies from ~5 to 15 m between depths of ~600 and 400 m (Heiken et al., 1988). Considering an ascent rate of ~0.02 m s^{-1}, and an average conduit radius of 10 m, the conduit strain rate would be of order 10^{-3} s^{-1}, and the corresponding fragmentation threshold would be met when $\mu_{melt} \approx 10^{10}$ Pa s. That viscosity is about two orders of magnitude greater than the viscosity expected for the Inyo magma at 750 °C and having an H_2O content of ~1.2 wt% (Hess and Dingwell, 1996), which is commensurate with the depth (~600 m) at which the Inyo conduit was intersected during research drilling (Eichelberger et al., 1985). At the implied ascent rates, the Inyo magma would have to degas to near-atmospheric levels in order for its viscosity to rise to the critical values for shear fragmentation (Hess and Dingwell, 1996). It appears that over much of its ascent, the conditions for shear fragmentation, and thus degassing by that mechanism, were not met. Only during the latest stages of dome extrusion would this mechanism be important, and by that stage, most of the magma had already degassed.

ACKNOWLEDGMENTS

Partial support was provided by the National Science Foundation (EAR-0106658). We thank C. Bacon, M. Coombs, J. Hammer, M. Manga, A. Rust, and T. Vogel for helpful comments.

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Manuscript received 24 September 2007
Revised manuscript received 20 November 2007
Manuscript accepted 4 December 2007

Printed in USA