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Textural and mineral chemistry constraints on evolution of Merapi Volcano, Indonesia

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ABSTRACT

We analyze and compare the textures of Merapi lavas (basalts and basaltic andesites) ranging in age from Proto-Merapi through modern activity, with the goal of gaining insights on the temporal evolution of Merapi's magmatic system. Analysis of textural parameters, such as phenocryst and microphenocryst crystallinity, coupled with crystal size distribution theory, provides information about the storage and transport of magmas. We combine textural analyses with geochemical investigations for a comprehensive comparison of erupted lavas over time. The chemical analyses identify crystal growth processes in magma chambers and underline differences between sample groups. Our work suggests the occurrence of two distinct histories, presumably associated with (at least) two generally distinct types of rheological behaviors and storage/transport systems. These behaviors are associated with different plagioclase growth patterns, with both groups influenced by late-stage shallow decompression degassing-induced microlite crystallization. Both groups contain amphibole crystals that indicate an early period of mid-crustal to deep-crustal storage of water-rich magmas. Dome lavas from the 20th century eruptive activity indicate quasi-steady-state nucleation-and-growth evolution interspersed with episodes of reheating and textural coarsening, suggesting residence in magma storage at multiple depths, both >10 km, and <10 km, while samples from the older stratigraphic history of Merapi record both repeated attainment and loss of quasi-steady-state conditions. These observations, coupled with our companion study of Merapi tephra samples, suggest that the relatively benign type of activity observed in the 20th century will be interrupted from time to time in the future by more explosive eruptions, such as that of 2010.

1. Introduction

Igneous rocks preserve the history of magmatic crystallization, and for this reason there is a high degree of interest in understanding the kinetic processes and intensive parameters that can be determined by the quantitative study of their textures. Quantitative textural analyses have proven useful in identifying magma chamber processes such as accumulation, magma mixing and fractionation, and in estimating residence times in magma reservoirs (Marsh, 1988, 1998; Higgins, 2011). Cooling-induced crystallization has been studied extensively for Hawaiian lava flows (Cashman and Marsh, 1988; Cashman, 1993; Cashman et al., 1999), and decompression or degassing-induced crystallization has been studied in both natural systems (Hammer et al., 1999, 2000; Couch et al., 2003a) and laboratory experiments (Hammer and Rutherford, 2002; Couch, 2003; Couch et al., 2003a,b; Hammer and Rutherford, 2003). Both crystallization and degassing change the melt chemistry (Cashman and Blundy, 2000; Hammer et al., 2000; Harford et al., 2003) and magma rheology (Dingwell, 1998), and therefore can affect eruption dynamics and style (Melnik and Sparks, 2002). However, the interactions are intricate and the processes remain incompletely understood, despite significant recent modeling advancements (Melnik and Sparks, 1999; Barmin et al., 2002; Melnik and Sparks, 2002; Diller et al., 2006; de' Michieli Vitturi et al., 2010).

Merapi volcano is an example of a relatively well documented system with a complex variety of observed behaviors (Voight et al., 2000a,b), and therefore provides an ideal opportunity to investigate the relationship between petrographic textures and mineral chemistry in lavas produced by eruptions of different styles. The eruptions at Merapi have been well documented, and recent events have been monitored closely (Andreastuti et al., 2000; Newhall et al., 2000; Voight et al., 2000a; Gertisser and Keller, 2003a, b; Chadwick et al., 2007; Charbonnier and Gertisser, 2008; Gertisser et al., 2012; Surono et al., 2012; Chadwick et al., 2012; Costa et al., 2013).

Here we quantify textures and bulk, mineral, and glass chemistries from a representative sampling of Merapi lavas of different vintage, and correlate these data to the eruptive processes. We also explore the mutual influence of geochemistry and rheological parameters in affecting the observed crystallization patterns and textures. To achieve these goals, we focus on occurrence, textures, and mineral chemistry

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of plagioclase feldspar and amphibole crystals in lava domes and lava flows sampled from throughout the eruptive history of Merapi volcano. The crystal zoning patterns and crystal size distribution curves appear to indicate the development of two distinct rheological behaviors, and these are interpreted to reflect, in general, two basically distinct types of magmatic reservoir storage histories.

2. Geological background

Merapi has experienced quasi-continuous extrusion of lava throughout the late-19th, 20th, and early-21st centuries (Siswowidjoyo et al., 1995; Voight et al., 2000a; Voight and Davis, 2000; Charbonnier and Gertisser, 2008; Surono et al., 2012), and a growth history that extends back at least to 40 ka (Berthommier et al., 1990; Camus et al., 2000; Newhall et al., 2000), and perhaps to >100 ka if we include precursor edifices (Gertisser et al., 2012). Its volcanism is related to the subduction of the Indo-Australian plate beneath the Eurasian continental plate.

Several authors (del Marmol, 1989; Berthommier et al., 1992; Andreastuti et al., 2000; Camus et al., 2000; Newhall et al., 2000; Gertisser et al., 2012) have summarized Merapi's activity, and have used stratigraphic and geochronologic evidence to interpret its history and to distinguish several eruptive stages. Here we follow Newhall et al. (2000) who subdivide the volcanic activity as follows: Proto Merapi (~40,000 ybp to ~5000 B.C.), Old Merapi (~5000 B.C. to ~100 A.D.), and New Merapi (~100 A.D. to present). The boundary between Old Merapi and New Merapi is set at the somma-forming edifice collapse of ~100 A.D. (Newhall et al., 2000) and is marked by the Tegalsruni tephra deposit (Andreastuti et al., 2000).

Detailed tephrostratigraphic work (Andreastuti, 1999; Andreastuti et al., 2000; Gertisser and Keller, 2003a) demonstrated that Old and New Merapi exhibited intermittent small to large-scale explosive activity, ranging from Vulcanian to sub-Plinian and Plinian events. Modern activity (19th–21th century) has been characterized by frequent (every 3–5 years) small-scale lava effusions and dome collapses, associated with small to moderate Vulcanian events with pyroclastic density currents that traveled through channeled topography as much as 10–15 km from the summit (Andreastuti et al., 2000; Voight et al., 2000a; Charbonnier and Gertisser, 2008). Although this activity has taken human lives (Abdurachman et al., 2000; Voight et al., 2000a; Voight and Davis, 2000) and damaged homes, it represents relatively mild behavior compared to several pre-historical devastating eruptions (Andreastuti et al., 2000). It had been recognized before 2010 that Merapi would re-develop larger eruptions than those of the 20th century (Newhall et al., 2000; Voight et al., 2000a), and this remains a concern because over 400,000 people live within the area of high risk now covered by deposits of highly destructive older explosive eruptions (Hartmann, 1934; Newhall et al., 2000; Thouret et al., 2000; Voight et al., 2000a; Lavigne et al., 2008). The eruption in 2010 represents a larger than normal but certainly not a worst-case eruption (Surono et al., 2012; cf. Andreastuti et al., 2000).

2.1. Geophysical and petrological evidence for magma storage locations

Here we review some evidence on the location of magma storage systems at Merapi to provide background for interpretations of our textural data. First, Merapi eruptions are usually associated with both relatively deep and shallow earthquakes of volcano-tectonic (VT) and multi-phase and low-frequency types (Ratdomopurbo and Poupinet, 2000; Hidayat et al., 2000, 2002, 2003). The latter types are of shallow origin (<1.5 km), but so-called 'deep' VTs of 'A' type, are detected between 2.5 and about 5 km, and seem to correspond to injection of gas-enriched magma into the shallow part of the system (Ratdomopurbo and Poupinet, 2000). On the basis of the aseismic gap a small shallow magma chamber has been proposed at a depth of approximately 1.5–2.5 km beneath the summit, near the base of the edifice, where magma injected from below can be stored temporarily (Ratdomopurbo and Poupinet, 1995, 2000). Such a chamber is probably ephemeral, perhaps contains a highly-crystalline mush, and is small, inasmuch as some additional investigations (with fairly crude detection precision) have failed to detect it (*see e.g.*, Chadwick et al., 2012 and cited references). An additional shallow chamber below 5 km depth is suggested by the distribution of hypocentres, with pressurized magma breakout from this chamber, and conduit transport to the shallow chamber at ~2 km depth, causing the deeper set of VT events (Ratdomopurbo and Poupinet, 2000). A similar VT-defined chamber top was proposed at Montserrat (Aspinall et al., 1998) and subsequently proven by active-source tomography (Paulatto et al., 2012). A chamber at this level at Merapi is consistent with the idea that most evolved, wet magmas ascending and crystallizing as a result of decompression degassing, with consequently changed melt composition and greatly enhanced viscosity, will stall at roughly 4–10 km (Annen et al., 2006). Also consistent with a >5 km chamber is a pressurization center around 8 km beneath the summit, suggested by analysis of tilt and GPS data (Beauducel and Cornet, 1999). Their results are suggestive rather than conclusive, because wider-aperture deformation data (say at radii >10 km) are required to examine deep crustal structure and to discriminate between competing geometric models of pressurized chambers. On Montserrat, for example, wide-aperature GPS deformation data are compatible with vertically-elongate idealized source models that could represent a stack of discrete mid-crust chambers (Voight et al., 2010; Mattioli et al., 2010), and an interplay between deep and shallow chambers is shown to control the eruption dynamics (Foroozan et al., 2011). Merapi has lacked a wide-aperature GPS array, but on balance the data suggest magma is stored at several locations <10 km deep.

Second, recent tomographic investigations have recognized a large velocity anomaly in the crust and upper mantle beneath Merapi. The anomaly seems to extend all the way through the crust, and also >25 km into mantle toward the subducting slab (Wagner et al., 2007; Koulakov et al., 2007, 2009).

Third, petrology and geochemistry complement geophysics. Chadwick et al. (2012) applied the Nimis (1999) barometry method to pyroxenes from basaltic-andesites and inclusions in Merapi lavas. Pyroxene crystallization pressures ranged from ~100 to 1300 MPa, with a concentration of values between 400 and 700 MPa. Results for host basaltic-andesites range from 200 to 900 MPa, with a concentration between 400 and 500 MPa, in agreement with Gertisser (2001). The data suggest that the bulk of pyroxenes in the basaltic-andesites crystallized in mid- to deep-crustal reservoirs (roughly 12–18 km). Pyroxenes in felsic plutonic inclusions suggest crystallization pressures of 65–720 MPa (roughly 2–25 km) (Chadwick et al., 2012). The lower pressure results based on crystal rim compositions suggest some storage in very shallow parts of the plumbing system, with most crystallization at 300–400 MPa (approximately 10–15 km). Pyroxenes in mafic plutonic inclusions indicate 300–720 MPa, and co-magmatic basaltic enclave pyroxenes indicate the deepest pressures, 300–1300 MPa, consistent with magma storage extending into the upper mantle (Chadwick et al., 2012). Associated temperature estimates of Merapi magmas are inferred to be between 900 and 1100 °C, with most estimates around 1000 °C (Gertisser, 2001; Chadwick et al., 2012; Costa et al., 2013).

Similar indications derive from study of amphiboles. The common presence of amphibole within phenocryst assemblages in our study indicates that the early stages of crystallization in nearly all samples occurred under water-saturated conditions at considerable depth (Moore and Carmichael, 1998; Innocenti et al., 2013). Pressure and temperature estimates by Costa et al. (2013), based on the geothermobarometer of Ridolfi and Renzulli (2012), show for Al-rich amphiboles in 2006 and 2010 lava samples a large range of pressures from about 300 to 900 MPa, with most determinations around 300–450 MPa and 800 MPa (Costa et al., 2013). These results with amphibole are consistent with the pyroxene-based interpretations of Gertisser (2001) and Chadwick et al.