Journal of Volcanology and Geothermal Research 261 (2013) 69–86

Contents lists available at SciVerse ScienceDirect

Journal of Volcanology and Geothermal Research

journal homepage: www.elsevier.com/locate/jvolgeores

The pre-eruption conditions for explosive eruptions at Merapi volcano as revealed by crystal texture and mineralogy

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ARTICLE INFO

Article history:
Received 29 May 2012
Accepted 3 December 2012
Available online 24 March 2013

Keywords:
Merapi
Textural analyses
Tephra
Kepuharjo
Dome lava
Lava flow
CSD
Eruptive style
Explosive
Effusive

ABSTRACT

We analyze the textures and mineralogy of Merapi tephra generated during explosive VEI 3–4 eruptions over the past 2000 years, and compare these data with those observed for Merapi dome and flow lavas. We find that the Merapi pumiceous tephra and lava textures differ significantly with respect to small-size crystal populations, but that phenocryst textures are generally similar. A similar initial phase of crystallization is indicated for tephras and lavas in mid-crustal (>10 km depth) reservoirs. Subsequent textural differences are mainly affected by ascent rate and degassing during ascent, and, for dome lavas, with temporary storage in shallower reservoirs. These differences also correspond to different eruptive styles. Our analyses include study of pumices, lava, and breadcrust-bomb samples, including samples from some of the most recent explosive episodes prior to the 2010 eruption. Textural analyses of youthful breadcrust bomb samples yield insights on one type of transition between effusive and explosive eruptive styles, involving pressure build-up under a degassed crystalline shallow-conduit plug. In general, comparison of the crystal size distributions and calculated residence times among the effusive and explosive eruptive styles suggests that the two main magma-product types resided for similar lengths of time in a mid-crustal reservoir, before ascending toward the surface and either erupting explosively (tephra), or stagnating in a shallow magma chamber prior to extrusion (lava). The interpretation is supported by the occurrences of amphibole in pristine condition in tephra and in altered state in lava. Finally the 1872 and 2010 explosive eruptions are examined and compared with others over the past three millennia. The 2010 bulk-rock compositions overlap with products of other major explosive Merapi eruptions, such as the Ngrangkah, Tegalsruni, Temusari, and Kepuharjo tephras. The 2010 products show a gradational late-stage mafic enrichment, dissimilar to 1872 which showed little (<1%) compositional variation, but similar to the Ngrangkah, Tegalsruni, and Temusari tephras. Such variation implies conventional withdrawal from a zoned magma chamber, with the more-evolved products being erupted first.

1. Introduction

Despite the recent dominance of activity at Merapi by dome-forming eruptions and lava flows, tephra production has constituted an important role in the formation of Merapi's edifice (Andreastuti et al., 2000; Newhall et al., 2000; Voight et al., 2000a). The larger explosive eruptions responsible for significant tephra falls and fountain-collapse pyroclastic flows are not as frequent as effusive lava-dome dominated eruptions, and their recurrence interval is on the order of centuries (Andreastuti et al., 2000). The stratigraphic record for the past several millennia demonstrates that Merapi is capable of switching between explosive and effusive-dominated eruptive styles (Newhall et al., 2000; Voight et al., 2000a,b). The explosive Merapi eruption of 2010 (VEI ~ 4) serves as a vivid reminder of this potential (Surono et al., 2012). The tephra-generating eruptions usually generate a series of hazards involving pyroclastic currents and lahars as well as tephra falls, and hazards evaluations should directly benefit from an improved understanding of the mechanisms that lead to changes in eruption style.

In this paper we study this issue by examining the crystal textures and mineralogy of Merapi tephras produced in explosive eruptions over the past 2000 years (integrating our data on the explosions) and also compare them with similar observations for Merapi dome and flow lavas (cf. Innocenti et al., 2013). We find that the Merapi tephra and lava textures differ significantly, particularly with respect to small-size crystal populations. We interpret these differences in terms of magma evolution and residence times in crustal storage reservoirs, as well as transport through conduits and filters to the surface.

1.1. Geological background

Merapi has experienced a volcanic history that extends back at least to 40 ka and perhaps >100 ka (Berthommier et al., 1990; Camus et al., 2000; Newhall et al., 2000; Gertisser et al., 2012). Effusion of lava has

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http://dx.doi.org/10.1016/j.jvolgeores.2012.12.028

been important throughout this history, and the quasi-continuous extrusion of lava, interspersed with mild explosions, characterized the late-19th, 20th, and early-21st century eruptions (Siswowidjoyo et al., 1995; Ratdomopurbo and Poupinet, 2000; Voight et al., 2000a,b; Charbonnier and Gertisser, 2008).

The stratigraphic record also indicates that numerous relatively-large explosive eruptions have been generated at Merapi, with volume as much as 3.7×10^8 m^3 (VEI 4; Andreastuti et al., 2000; Newhall et al., 2000). The eruptive deposits are typically produced from pyroclastic fall and pyroclastic flow sequences. Most of the fall deposits are characterized by single to multiple, generally upward-fining, pumiceous or scoriceous units, sometimes overlain by ash-cloud surge beds (Andreastuti, 1999; Andreastuti et al., 2000; Newhall et al., 2000). Isopach maps have been produced for a number of these beds, indicating dispersion directions and establishing fallout volumes (Andreastuti et al., 2000). Sub-plinian or plinian to vulcanian open-vent styles of activity have been common, and contrast with the effusive style of eruption that was most pronounced during the 19th and 20th centuries (Newhall et al., 2000; Voight et al., 2000a; Innocenti et al., 2013). The tephras have been basalts and basaltic andesites with compositions that ranged from 48.9 to 55.8 wt.%.

1.2. Magma generation and storage

In this section we review some evidence on the location of magma storage systems at Merapi to provide background for interpretations of our tephra and lava mineralogical data. First, recent tomographic investigations have recognized distinctive velocity anomalies in the crust and upper mantle beneath Merapi. The anomalies extend through the crust, and also tens of km into the mantle, toward the subducting slab (Koulakov et al., 2007; Wagner et al., 2007; Koulakov et al., 2009).

Second, petrology and geochemistry constrain magma storage locations. Using the Nimis (1999) barometry method, Chadwick et al. (2012) used pyroxenes from basaltic andesite and plutonic inclusions in Merapi lavas to establish crystallization pressures of ~100 to 1300 MPa, with a cluster of data between 400 and 700 MPa. Results for pyroxenes from host basaltic-andesite range from 200 to 900 MPa, with a concentration between 400 and 500 MPa, in agreement with Gertisser (2001). These data suggest that the bulk of pyroxenes in the basaltic andesites crystallized in mid-crustal reservoirs, approximately 11–18 km deep, with some crystals also suggesting a deep crust origin, near 30 km, where evolved magmas are generated by fractionation of basalts and assimilation of crustal rocks in a deep crustal hot zone (Annen et al., 2006; cf. Hildreth, 2007).

Pyroxenes in felsic plutonic inclusions, based on crystal rim compositions, suggest storage can also occur in shallower parts of the plumbing system, approximately 10–15 km. Associated temperature estimates 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). Study of amphibole has yielded similar estimates. 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, 300–900 MPa, with data mainly clustered at 300–450 MPa, and some at 800 MPa (Costa et al., 2013).

Lower-pressure (<10 km depth) storage and crystallization of some magma is likely (Innocenti et al., 2013). Seismic data support the existence of a small, probably ephemeral, magma chamber at a depth of approximately 1.5–2.5 km beneath the summit, near the base of the edifice, and an additional larger chamber below 5 km depth, beyond the VT-seismicity limit (Ratdomopurbo and Poupinet, 2000). The latter is consistent with the idea that evolved, wet magmas ascending and crystallizing as a result of decompression degassing, with consequently changed melt composition and greatly enhanced viscosity, will stall at ~4–10 km depth (Annen et al., 2006), and is further supported by data from felsic inclusions (Chadwick et al., 2012), and deformation modeling (Beauducel and Cornet, 1999). Magma degassing can occur at depths of <8–10 km, as magmas with 4–6 wt.% H_2O reach volatile saturation around 200–300 MPa, or deeper with CO_2 considered (Costa et al., 2013; cf. Moore and Carmichael, 1998; Annen et al., 2006).

2. Tephra and lava samples

The samples examined in this study belong to both explosive and effusive phases of Merapi volcanism. Tephrostratigraphic reconstructions and radiocarbon dating place the tephra samples in a period of time that spans approximately 2000 y (Andreastuti et al., 2000), including both the Old Merapi and New Merapi periods of activity as defined by Newhall et al. (2000).

The tephra samples were collected by S. Andreastuti in field campaigns carried out during 1996–2000, and consist of pumice clasts and breadcrust bombs (for information on sample locations, see Andreastuti, 1999; Andreastuti et al., 2000). The tephra specimens belong to the following stratigraphic units, ranked from oldest to youngest: Tegalsruni, Temusari, Plalangan, Jrakah, and Kepuharjo (VEI = 3–4) (Fig. 1; Table 1), formed during vulcanian to sub-plinian eruptions (see Andreastuti et al., 2000, Table 2).

The breadcrust bombs from the Kepuharjo (sometimes spelled Kepuhharjo) tephra (~250 ± 50 y bp), and so-called ‘1822/72’ bomb deposits (cf. Andreastuti, 1999; Andreastuti et al., 2000), are also classified as tephras, but result from eruption mechanisms of a different style than those producing pumices. The ‘1822/72’ bomb sampled from near the Merapi Golf Course (see Newhall et al., 2000, Section D) is of historical age, with a radiocarbon age of 160 ± 30 y bp, and it is interpreted to represent either the 1822 (VEI 3 or 4) explosive eruption or, perhaps less likely, the 1872 (VEI 4) explosive eruptions (Newhall et al., 2000), both of which destroyed the lava dome and many villages (Kemmerling, 1921; Hartmann, 1934; Voight et al., 2000a).

The dome lavas used in this study include samples from lava domes or lobes for 1888–1984 lava eruptions (del Marmol, 1989; Innocenti et al., 2013), and from pyroclastic flow deposits representing the 1986–1988 and 1992–2002 periods of activity (noted by ‘1990’ (NM9) and ‘1993’ (KMC-93) dome lava samples, respectively). Dome eruptive volumes were estimated to be about 10^6 m^3 for each of the dome-forming episodes, apart from the 1984–1988 and the 1992–2002 periods of activity, which emplaced ~10^7 m^3 lava (Voight et al., 2000a,b). Other lava samples (see Tables 1–3) refer to the Proto Merapi, Old Merapi, and New Merapi periods (Newhall et al., 2000).

3. Methods

Samples were embedded in epoxy and then sliced to make polished thin sections. Petrographic work and modal estimates were performed using a Nikon Labophot petrographic microscope, a mechanical sample holder, and a laboratory counter at The Pennsylvania State University. Phenocrysts and microphenocrysts (crystal lengths >100 μm) were counted under the 4× objective to determine phenocryst crystallinity (Table 2). Modal analyses of microlites (defined as having crystal lengths <100 μm) were performed on tephra samples under the 40× objective (Table 3). The volumetric percentage of microlites (almost exclusively plagioclase with minor amounts of pyroxene) was determined, and converted to groundmass crystallinity by adjusting for vesicularity.

Microphotograph acquisition for amphibole crystals and groundmass was done with a Nikon Eclipse LV100 petrographic microscope at The Pennsylvania State University. Back-scattered electron images were subsequently acquired on C-coated thin sections using the FEI Environmental Scattered Electron Microscope at the Materials Characterization Laboratory at The Pennsylvania State University (www.mri.psu.edu/directory/mcl.asp). Image frames at 100× magnification were acquired continuously at horizontal intervals of 2.3 mm and