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Textural and micro-petrological variations in the eruptive products of the 2006 dome-forming eruption of Merapi volcano, Indonesia: Implications for sub-surface processes

Katie Preece [a,*], Jenni Barclay [a], Ralf Gertisser [b], Richard A. Herd [a]

[a] School of Environmental Sciences, University of East Anglia, Norwich, NR4 7TJ, UK
[b] School of Physical and Geographical Sciences, Keele University, Keele, Staffordshire, ST5 5BG, UK

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ABSTRACT

The interplay between magma ascent, degassing and changing magmatic properties are widely recognized as critical factors controlling the style of silicic volcanic eruptions. Microlite textures in samples from the prolonged dome-forming eruption of Merapi in 2006 provide a record of changing magmatic ascent conditions and shallow conduit processes throughout the eruption. Analysis of microlite textural parameters, including measurements of areal number density (N_A), mean microlite size, crystal aspect ratio and groundmass crystallinity (φ), combined with the monitoring record and field observations, indicate that magma ascent paths change between continuous ascent at varying rates from a deeper magma storage region, to ascent being temporarily stalled at shallow depths in the latter stages of the eruption, supporting the idea of an ephemeral shallow magma storage region at Merapi. Plagioclase microlite compositions show evidence of decompression-induced degassing, often displaying rims of anorthoclase and more K-rich alkali feldspar (sanidine). Anorthite contents also support the textural data of later erupted magma being temporarily stalled at shallow depths. Crystal size distributions (CSDs) are interpreted to show that both growth-dominated and nucleation-dominated crystallisation regimes existed during the 2006 eruption, resulting from changing conditions of undercooling (ΔT) during variable magma ascent paths. By contrast, microlite textural analysis and feldspar microlite compositions of samples from the fast-growing lava dome of the second phase of the 2010 eruption prior to the cataclysmic events on 5 November indicate faster ascent rates, a crystallisation regime more strongly dominated by nucleation due to high ΔT and interaction of the 2010 magma with more hotter magma from depth.

1. Introduction

Magma ascent and conduit dynamics exert a profound control over eruptive behaviour, influencing whether an eruption will be effusive or explosive (e.g. Eichelberger et al., 1986; Jaupart and Allègre, 1991; Woods and Koyaguchi, 1994). The style of magma ascent and syn-eruptive degassing during ascent plays a large role in determining the style of eruption of compositionally similar magmas (e.g. Hildreth and Drake, 1992; Platz et al., 2007; Constantini et al., 2011). For example, open-system degassing, where magma degasses freely during ascent, typically results in a less explosive eruption compared to a partly closed system, where gas is prevented from escaping, more often resulting in explosive activity (e.g. Westrich et al., 1988; Stix et al., 1993). Degassing also influences the physical and rheological properties of magma by affecting the permeability, viscosity and crystallinity, which in turn affect the style of eruption (e.g. Eichelberger et al., 1986; Mastin, 2005; Edmonds and Herd, 2007; Hale et al., 2007). It is therefore crucial to better understand magma ascent processes and syn-eruptive degassing and their respective influence upon eruptive behaviour, in order to aid in future hazard assessment of volcanic eruptions.

During magma ascent, crystallisation of groundmass microlites may occur due to adiabatic decompression and accompanying devolatilisation (e.g. Cashman, 1992; Geschwind and Rutherford, 1995; Hammer et al., 1999, 2000). As magma ascends prior to eruption, decompression leads to volatile exsolution and degassing of H_2O. The resulting H_2O loss causes an increase in the stability of anhydrous minerals, particularly plagioclase, and an increase in the liquidus temperature of the anhydrous minerals. As a corollary of the increase in liquidus temperature, there is an increase in the relative undercooling (ΔT), defined as the difference between the temperature of the liquidus and that of the magma, causing the melt to crystallise. The style of crystallisation depends upon the kinetics and relative importance of crystal nucleation and crystal growth, as a consequence of ΔT (Lofgren, 1980; Kirkpatrick, 1981; Swanson et al., 1989). For example, a large ΔT favours nucleation of new crystals, leading to the formation of many smaller crystals, whereas smaller ΔT leads to a growth dominated regime, resulting in fewer but larger crystals. At very high undercooling however, low rates of diffusion, due to high melt viscosity at low H_2O content (Hess and Dingwell, 1996), mean that the magma does not crystallise, but is quenched to glass upon

* Corresponding author. Tel.: +44 1603 591338.
E-mail address: K.Preece@uea.ac.uk (K. Preece).

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eruption. Microlite textures are therefore an indication of the relative undercooling of the magma, determined by its decompression path. In the case of dome-forming eruptions, magma may also reside for prolonged periods of time near the surface, at elevated temperatures within the dome, allowing for the possibility of further groundmass crystallisation to occur (Sparks et al., 2000). The final groundmass texture is a result of the often complex ascent path of a magma, i.e. ascent rate, depth and style, which may be continuous or intermittent, with temporary stalling at one or several levels in the crust, as well as any post-extrusion, near-surface crystallisation that may take place during dome residence.

Previous investigations into groundmass crystallisation have utilized natural volcanic samples (e.g. Cashman, 1992; Wolf and Eichelberger, 1997; Hammer et al., 1999; Nakada and Motomura, 1999; Hammer et al., 2000; Noguchi et al., 2006, 2008; Suzuki and Fujii, 2010), decompression experiments (e.g. Geschwind and Rutherford, 1995; Hammer and Rutherford, 2002; Couch et al., 2003a; Brugger and Hammer, 2010a; Martel, 2012), as well as numerical modelling of crystallisation and magma ascent (e.g. Brandeis and Jaupart, 1987; Melnik and Sparks, 1999, 2005; Clarke et al., 2007; Melnik et al., 2011).

Previous work (Hammer et al., 2000) carried out on the feldspar microlite textures of recent dome-forming eruptions at Merapi has investigated processes of magma ascent and the resultant degassing-induced crystallisation over timescales of several years, using samples produced in distinct episodes of dome extrusion between 1986 and 1995. The study found a correlation between microlite number density and effusive flux, indicating that crystallisation conditions (ΔT) and resultant microlite textures were determined by magma ascent rate that was cyclic over the time period studied.

This paper presents a detailed textural and petrological case study of the exceptionally well-documented 2006 eruptive episode, with the aim of elucidating the short timescale variations (>3 month duration) of shallow magma ascent and conduit processes during a single dome-forming (effusive) eruption cycle at Merapi (Charbonnier and Gertisser, 2008; Ratdomopurbo et al., 2013). Chronologically controlled sampling and textural analysis of juvenile dome material, incorporated into several block-and-ash flow (BAF) deposits during successive dome collapse events during the course of the eruption, provide a detailed insight into sub-surface magmatic processes. When compared to the products of the more explosive 2010 eruption (e.g. Pallister et al., 2013; Surono et al., 2012), these data allow the relation of those textural features to differences in magma ascent processes and their control on eruptive mechanisms and behaviour.

2. Background

2.1. Recent eruptive history of Merapi

Throughout the last two centuries, activity at Merapi has been almost continuous, with eruptions occurring every few years, typically consisting of prolonged periods of dome growth followed by multiple gravitational dome collapses (VEI 1-2), to produce BAFs [see Voight et al. (2000) for a detailed summary]. This type of activity has become so synonymous with Merapi that small volume pyroclastic flows from gravitational dome collapse are often termed "Merapi-type" nuées ardentes. However, more explosive eruptions (up to VEI 4; Voight et al., 2000) have occurred during this time, notably in 1872 and, possibly, in 2010, and were more common in pre-historical time (Andreastuti et al., 2000; Camus et al., 2000; Newhall et al., 2000; Gertisser and Keller, 2003a; Gertisser et al., 2012a). Variations in magma supply from depth, magma ascent rate, the degassing behaviour during ascent and the assimilation of crustal carbonates are thought to be important factors that control whether Merapi erupts effusively or explosively (Newhall et al., 2000; Chadwick et al., 2007; Gertisser and Keller, 2003a; Deegan et al., 2010; Gertisser et al., 2011; Troll et al., 2012).

2.2. Merapi magma plumbing system

There is both petrological and geophysical evidence for the structure of the magma plumbing system at Merapi. Gertisser (2001) found evidence, based on thermobarometric calculations, of a major magma storage region at mid to lower crustal levels. There is also petrological evidence of a storage region at mid- upper crustal levels, where crustal carbonate contamination contributes to the volatile budget of the system, intensifying and sustaining eruptions (Deegan et al., 2010; Troll et al., 2012, 2013). Costa et al. (2013) present further petrological data to elucidate the magma storage conditions before the 2006 and 2010 eruptions, concluding that there are three main zones of crystallisation, consisting of a deep reservoir (~30 km depth), a second zone at ~11 km and a shallow, crystal-rich region at <10 km depth. Petrological evidence, based on magmatic inclusions, also reveals that crystallisation at Merapi occurs at multiple depths (~2–45 km depth), with the bulk of it occurring at mid-crustal levels (Chadwick et al., 2013). Geophysical studies also indicate the presence of multiple magma storage regions at Merapi. Hypocentre distributions and seismic tomography confirm the presence of a low velocity and presumably melt and fluid bearing region extending from the shallow crust near Merapi to the subducting slab at depths of ~100 km (Koulakov et al., 2007; Lühr et al., 2013). GPS and tilt data suggest an average source depth for magma storage at 8.5 ± 0.4 km below the summit (Beauducel and Cornet, 1999), consistent with an aseismic zone at >5 km depth below the summit (Ratdomopurbo and Poupinet, 2000). As well as a deeper source, one or potentially two shallower storage regions have been proposed, linked to the deeper storage region(s) via an inclined conduit, based upon Bouguer gravity anomaly data (Saepuloh et al., 2010). In addition, a shallow aseismic zone located at 1.5–2.5 km depth below the summit has been proposed and is interpreted to be an ephemeral storage region, where magma is temporarily stored as it migrates from the deeper reservoir(s) before eruption (Ratdomopurbo and Poupinet, 2000). A low density region between 0.8 and 1.8 km below the summit has also been postulated from gravity data, although it is not clear if this is a zone of magma storage (Tiede et al., 2005). The plumbing system at Merapi is therefore complex and believed to consist of multiple regions of magma storage and crystallisation. These range over almost the entire thickness of the crust, with a main magma storage zone within the lower crust, linked via a network of other regions of crystallisation in the mid- to shallow-crust, to a shallow magma storage region within the edifice, which is probably small and ephemeral.

2.3. The 2006 eruption of Merapi volcano

The 2006 eruption is a well characterized example of extrusive, dome-forming activity at Merapi (Charbonnier and Gertisser, 2008, 2009, 2011; Lube et al., 2011; Gertisser et al., 2012b). The onset of the eruption (VEI 1) is estimated from monitoring data to be 26 April 2006 (Ratdomopurbo, 2011). Lava dome extrusion probably began on 1 May, although it was not directly observed until 5 May (Ratdomopurbo et al., 2013), after approximately 10 months of increased seismicity and summit deformation (Fig. 1). The first dome collapses occurred on 11 May 2006, and continued during May, with BAFs channelled and emplaced in river valleys, herein referred to as "Kali", (Indo. = river), abbreviated as "K.", on the south-west flank of the volcano, including Kali Bebeng, Kali Krasak, Kali Boyong and Kali Bedok, with run-out distances of less than 4 km. On 22 May, the dome volume was estimated to be $\sim 2.3 \times 10^6$ m^3. A M_w6.3 tectonic earthquake occurred ~40 km south of Merapi on 27 May, after which, a considerable increase in dome collapse events was noted (Walter et al., 2007; Charbonnier and Gertisser, 2008; Troll et al., 2012). On 4 June, the dome was estimated to have reached a height of 100 m above the former summit and a volume of $\sim 4.0 \times 10^6$ m^3 (Smithsonian Institution, 2007) and between 1 May and 8 June, extrusion rates were estimated to increase from 1.0 m^3 s^{-1} to 3.3 m^3 s^{-1} (Ratdomopurbo et al., 2013). Throughout the beginning of