1 **Self-mixing magma in the Ruiz Peak rhyodacite (New Mexico, USA): a**
2 **mechanism explaining the formation of long period polytypes of mica**

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17 **ABSTRACT**

18 The rhyodacite of Ruiz Peak Volcano (New Mexico, USA) is an exceptional rock because it
19 contains both long period and short period polytypes of mica. Our petrographic study shows
20 that this rhyodacite is characterized by numerous disequilibrium textures of phenocrysts
21 (mica, amphibole, clinopyroxene, olivine and plagioclase) contained within both dark-grey
22 and reddish coloured groundmass. The presence of two groundmasses, as well as of
23 disequilibrium textures (reaction rims, resorption, dendritic, skeletal morphologies, etc.)
24 suggests a complex magmatic history. These two types of groundmass are not due to a mixing
25 of magmas but result from a degassing process during the magma ascent in the conduit. The
26 disequilibrium textures are interpreted to be the result of small, short-lived convection cells in

the magmatic chamber, which may allow crystal-crystal, crystal-spiral and spiral-spiral interactions to occur, leading to the formation of long period polytypes of mica. For the first time, the relationships between the crystallographic features of mica and the host-rock formation are underlined in this study. It follows that long period polytypes of mica can be considered markers of the complex history of magmas.

Keywords: long period polytypes; mica; self-mixing magma; convective cells; disequilibrium textures.

1. Introduction

Mica structure is based on an octahedral sheet (O) between two tetrahedral sheets (T), forming a T-O-T or 2:1 layer, also known as the M layer (Takéuchi, 1971; Takéuchi and Haga, 1971). The M layers can take six orientations, rotated by angles multiples of 60° and they are expressed by OD symbols (Dornberger-Schiff et al., 1982) (Fig. 1a). The stacking of the M layer along the *c* axis gives rise to different mica polytypes. Polytypes with stacking sequences of several tens to hundreds of Å are not common and are termed ‘long period’ to distinguish them from shorter (10 to 30 Å) and more frequent polytypes (Fig. 1). Long period polytypes do not have a defined thermodynamic field of stability (Bozhilov et al., 2009).

The coexistence of short and long period polytypes in the same rock is extremely rare, but occurs in the Ruiz Peak Volcano rhyodacite (New Mexico, USA), where one third of micas are long period polytypes (Ross et al., 1966). This explains why crystallographers and mineralogists have been particularly interested in this rock since the 1960s. In fact, most of long period polytypes reported in literature come from this rock: $3A_1$, $4M_2$, $8M_8$, $8A_{12}$, $8A_1$, $10A_3$, $11M_1$, $14A_1$, $23A_1$ (Ross et al., 1966), $8A_2$ (Nespolo and Takeda, 1999) $32A_1$, $36A_2$ (Kogure and Nespolo, 1999), $4M_3$ (Pignatelli et al., 2011), and $5M_3$ (Pignatelli and Nespolo,

2011). Only few examples of long-period polytypes found in other rocks are known: $4M_3$ polytype of titanium phlogopite from the kamptonite dike of Vostochnyi Priazov - Azov Sea region, Russia- (Zhukhlistov et al., 1988), $3A_1$ polytype of annite from the alkalic trachyte of the Khibiny pluton, Russia (Borutskiy et al., 1987), the 94-layer polytype of Mg-rich annite from dacite rocks of Džep, Serbia (Fregola and Scandale, 2011), $6M_1$ and $3M_1$ polytypes found in a lepidolite from Tørdal, Norway (Rule et al., 1987), and $3A_1$, $5A_2$ muscovite polytypes synthesized by Baronnet et al. (1976).

Several hypotheses have been proposed to explain the formation of long period polytypes: i) kinetic factors (Sunagawa et al., 1968; Baronnet et al., 1981; Baronnet, 1992; Kogure and Nespolo, 1999); ii) structural controls, where strength varies as a function of the thermodynamic conditions during growth (Smith and Yoder, 1956; Güven, 1971); iii) non-equilibrium conditions related to compositional fluctuation of fluid during crystallisation (Xu and Veblen, 1995; Fregola et al. 2009); iv) the formation of a screw dislocation with an unfaulted (“*Perfect Matrix Model*”, Baronnet, 1975) or faulted exposed ledge (“*Faulted Matrix Model*”, Pandey et al., 1982); and finally, v) perturbations occurring during spiral growth (“*perturbative theory*”, Nespolo, 2001) i.e., spiral-spiral interactions, spiral-crystal interactions (“*platelet mechanism*”, Takeda and Ross, 1995) and crystal-crystal interactions.

Despite these diverse explanations, the formation of long period polytypes, as well as their coexistence with short polytypes, are not yet well-understood due to the scarce information available on both polytypism and host-rock relationships. The Ruiz Peak rhyodacite is therefore an important example that may shed new light on these relationships. Reconstruction of the formation and geological history of the Ruiz Peak rhyodacite is therefore essential to better understand the growth conditions of long period polytypes of mica. In this paper, we present a detailed petrographic and mineralogical description of Ruiz Peak rhyodacite and we propose a new hypothesis for its formation, taking into account the

disequilibrium textures of minerals. Our hypothesis is based on a self- mixing magma process that enables the formation of long period polytypes of mica. This hypothesis also describes how the formation of two groundmasses with different degrees of vesiculation is due to the degassing during the magmatic ascent and not to the mixing between mafic (basaltic andesite) and felsic (rhyolite) magmas as previously put forward by Guilbeau (1982) and Guilbeau and Kudo (1985).

2. Geological setting

Ruiz Peak Volcano is located in the Jemez Mountains (Fig. 2), a region that covers nearly 4000 km² in the north of New Mexico (Guilbeau, 1982). The Jemez Mountains were formed by intense volcanic activity that started in early Pliocene and culminated with an explosive caldera-forming eruption during the Pleistocene (Kudo, 1974). Pre-, post- and caldera formation rocks were studied by several authors in the 1960s and 1970s (Ross et al., 1961; Smith and Bailey, 1966, 1968; Bailey et al., 1969; Smith et al., 1970). Bailey et al. (1969) divided the oldest pre-caldera stratigraphic group, the “Keres Group”, into four geological formations (Table 1). The sample studied here belongs to an additional formation, the “rhyodacite of La Jara Canyon”, which is missing in the classification of Bailey et al. (1969) but was later identified by Guilbeau (1982) and Guilbeau and Kudo (1985). This formation corresponds to a dome and flow sequence located in the southern part of Ruiz Peak area (Guilbeau, 1982) (Fig. 2). According to Takeda and Ross (1975), the sample is collected from the top of the flow that is generated by the collapse and fragmentation of the dome.

It should be noted that we didn’t collect the sample, as it was kindly provided by Malcom Ross. The sample is the same one used for all previous studies about long period polytypes from Ruiz Peak (Ross et al., 1966; Nespolo and Takeda, 1999; Kogure and Nespolo, 1999; Pignatelli et al., 2011; Pignatelli and Nespolo, 2011). It is the reference sample with the

highest content of long period polytypes found on the Earth so far and it is, thus, the best candidate to understand the relationships between the polytypism and the magmatic processes.

3. Analytical techniques

Polished thin sections were cut parallel to the polished surface of the rhyodacite sample (Fig. 1). Petrographic and textural studies were undertaken using several techniques.

Images of mineral textures were obtained by back-scattered electron imagery (BSE), using a scanning electron microscope (SEM) JEOL J7600F at SCMEM laboratory (University of Lorraine). The images were acquired using an accelerating voltage of 15 kV and a beam current of 1.2 nA. The same acquisition conditions were used to determine the compositions of the different types of groundmass, but data were instead collected in raster mode with probe diameters of ~ 60, 30 and 15 μm (rastered areas of 3600 μm^2 ; 900 μm^2 and 225 μm^2 respectively). Three probe diameters were used in order to consider the different sizes of crystals and/or vesicles. The used standard were: albite for Na, forsterite for Mg, corundum for Al, orthoclase for K, andradite for Ca, pyrophanite for Ti and Mn, hematite for Fe. Mineral chemical compositions were determined by electron microprobe analysis (EMPA) at SCMEM laboratory (University of Lorraine) using a CAMECA SX100 microprobe. Minerals were analysed with an accelerating voltage of 15 kV, a beam current of 12 nA and a minimum probe diameter of 1 μm . The total iron content was reported as FeO. The microprobe was calibrated using the following standards: Amelia albite for Na and Si, San Carlos forsterite for Mg, corundum for Al, orthoclase for K, andradite from Corsica (BRGM) for Ca, pyrophanite for Ti and Mn, hematite for Fe, Cr_2O_3 for Cr, baryte for Ba and topaz for F.

Raman spectra were recorded using a LabRAM HR spectrometer (Horiba Jobin Yvon) equipped with a 600 gr.mm^{-1} grating and an Edge filter at GeoRessources laboratory

(University of Lorraine). The confocal hole aperture was set to 500 μm and the slit aperture to 100 μm . The excitation beam was provided by a Stabilite 2017 Ar^+ laser (Spectra Physics, Newport Corporation) at 514.53 nm and 50 mW power, using a $\times 100$ objective (Olympus) to focus on the sample. Acquisition time accumulations of 5 and 10s were selected to optimize the signal-to-noise ratio (S/N).

4. Results

At macroscopic scale (Fig. 3), the groundmass of the sample shows two types of coloured zones: reddish zones and dark-grey ones. Dark-grey zones are subspherical with a maximum diameter of 8 mm and are surrounded by reddish zones. The dark-grey groundmass contains 1-mm- to 8-mm-sized plagioclase phenocrysts (average of ~ 4 mm) with a preferential orientation, and dark-green pyroxene phenocrysts (~ 0.7 mm to 0.5 cm). It is worth noting that, regardless of the colour (dark-grey or reddish) of the zone examined, no signs of vesiculation are visible at macroscopic scale.

In thin section, the sample exhibits a mainly porphyric texture, but is locally glomeroporphyric. Glomeroporphyric aggregates constitute around 10% of the rock and show generally an orthocumulate textures. They contain the same phenocrysts observed in the groundmasses and are described in detail below.

4.1 Phenocrysts

Phenocrysts of plagioclase (15%), mica (2%), Fe-Ti oxides (1%), amphibole (1%), pyroxene (1%), olivine (1%), and fluoroapatite ($< 1\%$) were observed in thin section. The remaining amount of $\sim 78\%$ is constituted of two groundmasses.

Mica crystals are euhedral to subeuhedral and show a distinctive reddish-brown pleochroism (Fig. 4a). The chemical composition of the mica was:

as reported in Pignatelli et al. (2011) and falls in the phlogopite field of the classification diagram proposed by Tischendorf et al. (2007) (Fig. 4b). No significant difference was observed in the crystal-chemistry of micas found in the groundmass and glomeroporphyric aggregates (Fig. 4b). According to literature data, Ruiz Peak mica is highly oxidized, though the degree of oxidation can vary slightly from one crystal to another (Takeda and Ross, 1975; Pignatelli et al., 2011; Pignatelli and Nespolo, 2011). In previous studies, this mica was named either “oxybiotite”, because of the elevated amounts of the oxygen in the structure (Takeda and Ross, 1975; Otha et al., 1982), or “ferriphlogopite” (Pignatelli et al., 2011; Pignatelli and Nespolo, 2011) to emphasize the presence of Fe^{3+} and to distinguish it from tetra-ferriphlogopite, in which ferric iron occupies the tetrahedral sites according to Rieder et al. (1998). Mica crystals are often surrounded by a dark rim of thickness varying from ~ 10 to 50 μm (Fig. 4a). SEM-EDX analyses of these rims indicate that they are composed of Fe-oxides and Na-Ca pyroxenes close to omphacite. A few basal sections of micas were observed in the glomeroporphyric aggregates and some of these presented straight marks intersecting at 60° (Fig. 5a, b). Even though the resulting aspect is similar to a sagenitic texture (Gary et al., 1972; Shau et al., 1991), the SEM data indicate that these marks are not needle-like inclusions of titanite or rutile, but correspond to sets of fractures that are sometimes discontinuously filled by small inclusions of iron oxides. Micas containing these fractures also show chemical zoning corresponding to Mg-Fe substitutions and probably also to Fe^{2+}/Fe^{3+} ratio variations (higher FeO wt. % are associated with lower MgO in the brighter regions of the BSE images; Fig. 5c and 5d). Moreover, the total sum of oxides in the darker zones is lower, possibly suggesting greater amounts of Fe^{3+} (Fig. 5d).

Amphibole crystals are subeuhedral or rounded and also present a reddish-brown pleochroism. Like the micas, the amphiboles are surrounded by a breakdown rim of variable

thickness (from ~ 6 to 98 μm – Fig. 6) but this rim is instead composed of plagioclase, pyroxene and Fe-Ti oxides. The BSE images of the rims show a symplectite-like texture (Fig. 7), similar to that described by Pletchov et al. (2006, 2008). The mean composition of amphibole (Table 2) calculated using the Locock's spreadsheet (Locock, 2014), is:

According to the amphibole nomenclature approved by IMA2012 (Hawthorne et al., 2012) the compositions of the analysed crystals all fall within the Ti-rich magnesio-hastingsite field.

Rare euhedral, Mg-rich (Fo_{77}) phenocrysts of olivine were observed. These are all fractured and have destabilized into iddingsite (Fig. 8a). Given the high degree of destabilisation, it was only possible to determine the chemical composition in a few iddingsite-free zones in the crystals. However, high content of magnesium measured in olivines results probably from oxidization process that produce iddingsite (Delvigne et al., 1979; Smith et al., 1987; Cortés et al., 2006; Genareau et al., 2010). Olivine phenocrysts are characterised by ripening dendritic overgrowths. The SEM images show that the iddingsite has developed parallel to the {001} planes of the olivines, filling several fractures, also present in the overgrowths (Fig. 8b). In contrast, the longest fractures, which span the entire crystals, were unaffected by the destabilization process. Moreover, microphenocrysts of olivine display a particular kind of ripening skeletal morphology known as “closed hopper”, characterized by glass inclusions that are arranged symmetrically with respect to the centre of the crystal (Faure and Schiano, 2004) (Fig. 8c).

Plagioclases are characterised by a seriate texture consisting of phenocrysts (> 300 μm) and large (~ 150-10 μm) and small microlites (< 10 μm). The phenocrysts have a preferential orientation, as noted in the macroscopic description. They are euhedral to subhedral with complex reaction textures, similar to those described in the Mt. Etna lava

(Nicotra and Viccaro, 2012; Lanzafame et al., 2013). Two types of plagioclase phenocrysts can be distinguished in both groundmasses with equivalent amounts. Type I phenocrysts are normally zoned and characterized by a core with lighter colour in BSE images ($An = 49-55\%$, Table 3). The core is mantled by a plagioclase rim that appears darker in the BSE images (low An content; $An = 36-39\%$) (Fig. 9a). The clear core sometimes contains partially-crystallized glass inclusions with clinopyroxene and cristobalite as daughter minerals (Fig. 9b). Type II plagioclases exhibit an irregular-shaped core rimmed by a sieve-zone (Fig. 9c) or only a sieve-texture. This sieve-texture perfectly meets the description given by Tsuchiyama (1985), i.e. a mixture of micrometer-scale channels of crystal and melt. In both cases, the cores of Type II plagioclases are dark in the BSE images due to their low An contents ($An = 39-41\%$). The boundary between the sieve zone and the overgrowth that restores the polyhedral morphology is generally irregular and truncates the sieve zone. The overgrowth rim is always lighter in the BSE images, and is thus richer in An ($An = 53-59\%$) than the sieve zone. However, the overgrowth rims are sometimes more complex and show compositional zoning. Similarly, An -rich plagioclase has developed on wavy resorption horizons composed of An -poor plagioclase (Fig. 9d).

Clinopyroxene phenocrysts are pale green and show interference colours up to high second order. They are present in both types of groundmass and in the glomeroporphyric aggregates. Thick prismatic sections, typical of pyroxene habitus, are observed in the groundmass (Fig. 10b), whereas the crystals are anhedral in aggregates (Fig. 10a). No difference in composition was observed between the pyroxenes in the groundmass and those in the aggregates (Figure 10c). The mean pyroxene composition, calculated using the WinPyrox program (Yavuz, 2013), is close to that of diopside and augite (Table 4):

where Fe^{2+}/Fe^{3+} ratio was estimated using the Dropp's method (1987).

Fluorapatites are present as inclusions in mica or plagioclase, and are also dispersed in the groundmass and glomeroporphyric aggregates. The fluorapatite crystals are surrounded by a reddish rim, but the rim was too small to determine its composition (its thickness varies from 1 to 3 μm).

Oxides in groundmass and glomeroporphyric aggregates were examined using an optical microscope with reflected light and analysed by SEM. Two types of iron oxide can be distinguished (Fig. 11a): an isotropic oxide, which is lighter in the BSE images and characterised by a low Ti content (Table 5); and an anisotropic oxide, which is darker in the BSE images and is Ti-rich (Table 5). The compositions of these oxides are close to titanomagnetite and ferri-ilmenite, respectively (Fig. 11b).

4.2 Microlites and groundmass

The microlites observed in both reddish and dark-grey groundmass are irregular-shaped oxides, plagioclases and pyroxenes. The larger plagioclase microlites (10-150 μm) have a characteristic fork-shaped skeletal morphology (Lofgren, 1974) (Fig. 12) and a preferential orientation, whereas the smaller ones (< 10 μm) are randomly orientated and have a lath morphology. BSE images show that the larger microlites exhibit compositional zoning with a bright core (An-rich; An = 54-57%) and a thinner, dark rim (An-poor) (Fig. 12b). In contrast, the small microlites have a homogeneous dark colour (An-poor) that is similar to the rims of larger microlites. Like the plagioclases, the pyroxene microlites are of various sizes and shapes. Microlites larger than 10 μm are usually hollow, as is characteristic of skeletal growth (Kouchi et al., 1983) (see Fig. 13b), and are preferentially orientated, while the smaller microlites appear to be elongated and randomly orientated.

The two coloured groundmasses can be distinguished by their different porosity in SEM images (Fig. 13a): the dark-grey groundmass displays small vesicles (average size of $\sim 1 \mu\text{m}$)

trapped between the crystals (Fig. 13b), while the reddish groundmass is characterized by larger (average size of $\sim 40 \mu\text{m}$) irregular-shaped vesicles (Fig. 13c). The chemical compositions of each type of groundmass are reported in Table 6. Analyses were performed using different probe diameters (15 to $60 \mu\text{m}$) in order to take into account the different sizes of crystals and vesicles; however, oxides wt.% show little variation and are independent of the probe diameter. The two types of groundmass do not differ significantly in chemical composition and mineralogy, except for the occurrence of a SiO_2 phase only in the reddish groundmass and mainly next to large vesicles (Fig. 13a-c and 14a). Raman spectroscopy allows to identify this silica phase as the tetragonal α -polymorph of cristobalite (Bates, 1972) (Fig. 14b). The typical “fish-scale” cracking exhibited by cristobalite (Fig. 13a-c and 14a) is consistent with a decrease in volume following the displacive phase transition from β to α -cristobalite (Carpenter et al., 1998; Horwell et al., 2013).

4.3 Thermobarometry

Pressures and temperatures during phenocryst crystallization can be estimated from crystal chemistry. The applicability of thermobarometric models assuming that the crystal growth occurs in near-equilibrium conditions can be critical when the crystals show disequilibrium textures such as in this rock (Hammer et al., 2016). A combination of different thermometers/barometers is used here to obtain more reliable data.

Two recently developed amphibole thermobarometers (Ridolfi et al., 2010; Ridolfi and Renzulli, 2012) allow the physical-chemical conditions during amphibole formation to be calculated from Al contents. The reliability of these thermobarometers is discussed in details by Erdmann et al. (2014), who showed that both give incorrect values of the pressure but reasonable estimations of temperature. The authors also stated that the calculated oxygen fugacity appears reasonable only when the calibration of Ridolfi et al. (2010) is used.

277 According to the measured Al_2O_3 concentrations (Table 2), the Ruiz Peak rhyodacite
278 amphiboles formed at a temperature of around 1000°C with a relative oxygen fugacity
279 (ΔNNO) ranging from 0.6 to 1 (respectively for Al_{max} and Al_{min} contents - Ridolfi et al.,
280 2010).

281 Pressure estimation can be obtained using other geobarometers (Hammarstrom and Zen,
282 1986; Hollister et al., 1987; Schmidt, 1992) which suggest a range of $\sim 4\text{--}7$ kbar, in agreement
283 with the results obtained using the equations of Ridolfi and Renzulli (2012). It should be
284 noted that the pressure estimation were obtained using core compositions to avoid post-
285 crystallization alteration, considering that amphibole crystals in this rock present
286 disequilibrium textures (i.e. breakdown rim).

287 Thermobarometric estimations can be obtained also on Ti contents of mica that increase with
288 Fe and/or $f\text{O}_2$ (Arima and Edgar, 1981; Abrecht and Hewitt, 1988). According to the
289 experimental results of Trnnes et al. (1985), the mean TiO_2 of 6.5 wt. % in the Ruiz Peak
290 rhyodacite micas suggests that Ti was incorporated into the mica structure at a temperature of
291 $\sim 1000^\circ\text{C}$ and a pressure of < 10 kbar.

292 Thermobarometric estimations obtained on pyroxene using the WinPyrox program (Yavuz,
293 2013) yield pressures ranging from ~ 4 to 10 kbar (Nimis and Ulmer, 1998; Nimis, 1999),
294 consistent with the results of the mica and amphibole barometry, and temperatures of around
295 930°C (Nimis and Taylor, 2000).

296 In summary, thermobarometric estimations based on amphibole, mica and clinopyroxene
297 phenocrysts suggest pre-eruptive conditions of around $900\text{--}1000^\circ\text{C}$ and $< 10\text{ kbar}$.

298 On the basis of Fe-Ti oxide compositions (Table 5), it also possible to estimate the
299 temperatures of $\text{Fe}^{2+} + \text{Ti}^{4+} \leftrightarrow 2\text{Fe}^{3+}$ and $\text{Fe}^{2+} \leftrightarrow \text{Mg}^{2+}$ exchanges between oxides. Using the
300 thermobarometry proposed by Ghiorso and Evans (2008) we estimate temperatures of $\sim 750^\circ$
301 C and $\sim 1160^\circ\text{C}$, respectively for Fe-Ti and Fe-Mg exchanges. These results are also

indicative of oxidizing conditions ($\log f\text{O}_2$ relative to $\text{NNO} = 1.4$) and elevated TiO_2 activity ($a_{\text{TiO}_2} = 0.8$) in the magma prior to eruption.

5. Discussion

The petrographic and textural characteristics described above indicate the complexity of the emplacement and crystallization of the Ruiz Peak rhyodacite. The disequilibrium textures, the presence of two different colours of groundmass, as well as the presence of two plagioclase populations were initially interpreted by Guilbeau (1982) and Guilbeau and Kudo (1985) as reflecting mixing between mafic (basaltic andesite) and felsic (rhyolite) magmas. However, our detailed petrographic study has revealed two lines of evidence which contradict this hypothesis. First, the absence of quenched margins around the dark grey groundmass indicates that the associated magmas were at similar temperatures, which is inconsistent with mixing between basaltic andesite and rhyolite magmas (Tepley et al. 1999; Couch et al., 2001). Second, the two types of groundmass have similar chemical compositions (Table 6). In order to understand how they are formed, we first discuss the origin of cristobalite, the principal phase that allows the two groundmasses to be distinguished. We then put forward an alternative model in which a self-mixing magma process is invoked to explain the observed disequilibrium textures and the extraordinary presence of numerous long period polytypes of mica in this rock (Fig. 15). In this model, the different degree of vesiculation in the reddish and dark-grey groundmasses is not associated with a magma mixing process, but with the strain localization at the conduit wall or even across the conduit during the magma ascent as described by Shea et al. (2014).

5.1 Origin of cristobalite

Cristobalite is a high-temperature polymorph of the silica phase that occurs at low pressure. The cristobalite stability field is located between 1470°C and 1713°C and at pressures below to 10 kbar (Deer et al., 1996). The presence of cristobalite crystals in the reddish groundmass therefore implies late metastable crystallization of this mineral at low temperature. According to Deer et al. (1996), the persistence of this silica polymorph at ambient temperature and pressure is due to the high energy needed for the reconstructive transformation of cristobalite to quartz. Horwell et al. (2013) have recently reviewed where and how cristobalite crystallizes in silicic volcanic systems. Two mechanisms are generally invoked to explain the presence of this metastable phase in lava domes: devitrification of groundmass glass and precipitation from a vapour-phase fluid (Baxter et al., 1999; de Hoog et al., 2005). The precipitation is controlled by the degassing at atmospheric pressure and low temperature (Boudon et al. 2015). The first mechanism seems improbable for Ruiz Peak because silica crystals that have been produced by devitrification are characterized by unusual textures that are not observed in the Ruiz Peak rhyodacite such as spherulites (Swanson et al., 1989) or finely crystalline intergrowths of feldspar and silica. In contrast, cristobalite that has been precipitated from a vapour-phase fluid is generally found within fractures or open pore-spaces (Horwell et al. 2013), as it is in the Ruiz Peak sample (Fig. 14a). The fluid could have two origins, either hydrothermal circulation (i.e. a meteoric fluid that passes through the permeable dome) or a magmatic origin, in which case it would correspond to the ultimate differentiated liquid (Horwell et al., 2013). However, feldspars in the Ruiz Peak sample show no signs of alteration by a meteoric fluid, and the presence of cristobalite daughter crystals inside feldspar-hosted melt inclusions (Fig. 9b) is further evidence for a magmatic origin for this fluid.

5.2 Pre-eruptive magma history

The presence of glomeroporphyric aggregates composed of minerals (plagioclase, mica, amphibole, oxide, clinopyroxene, fluorapatite) that have the same compositions as the phenocrysts found in the two types of groundmass, suggests that the original magma was partially crystallized. These aggregates correspond, thus, to a crystal mush forming on the margins of the magmatic chamber. The difference in temperature and viscosity between the cooler margins and the hotter interior of the chamber could lead to the development of convection cells similar to those proposed for the Soufrière Hills volcano, Montserrat (Couch et al., 2001) and for the Iwodake rhyolitic cone of the Satsuma-Iwojima Volcano, Japan (Shinohara and Tanaka, 2012). The complex crystal morphologies observed in the Ruiz Peak rhyodacite are consistent with the development of convection cells and the associated self-mixing of magma (Fig. 15). For example, the presence of both euhedral morphology and ripening dendrites on a single olivine crystal suggests the occurrence of cooling-heating cycles brought about by thermal convection in the magmatic chamber (Faure and Schiano, 2004; Welsch et al., 2009). Similarly, the two types of plagioclase textures observed in the sample seem consistent with the existence of a convective mechanism. Type I plagioclase, which has an An-rich core mantled by An-poor rim, corresponds to a crystal that formed within the hotter zone and moved into a colder, shallow region. In contrast, Type II plagioclases show an An-poor core with a sieve texture due to immersion of the crystal in hotter magma, as demonstrated experimentally by Tsuchiyama (1985). In addition, the complex compositional zoning suggests that the crystals moved quickly from one growth medium to another. Thus, the presence of plagioclase phenocrysts with different degrees of resorption and presenting compositional zoning could be due to the existence of a convective self-mixing mechanism (Kuo and Kirkpatrick, 1982). However, unlike Satsuma-Iwojima Volcano, where magma convection is considered to be a steady state process that allows

continuous degassing of the magma (Kazahaya et al., 2002), short transient magma convection is proposed for Ruiz Peak volcano. The convection is triggered only when the cooler crystal-rich mush on the walls of the chamber has more than 50% of crystals and a critical thickness that causes its sink (Welsch et al., 2009). This therefore implies the presence of small and short-lived convection cells in the vicinity of the chamber margins, as has previously been proposed for mafic magmas (Welsch et al., 2009; Colin et al., 2012).

5.3 Ascent of magma in a conduit

During the ascent of a magma with uniform composition and crystal content, shear-zones form in the conduit to accommodate the velocity differences according to the model of Shea et al. (2014). This favours the H₂O migration towards the zone of maximum strain, which became richer in H₂O. In these zones the vesicles are more deformed and interconnected in comparison to zones where the strain is lower (Fig. 15). It follows that the different degree of vesiculation observed in the reddish and dark grey groundmasses (Fig. 13) is not due to the mixing of magmas but it is degassing-induced.

Preferential orientations of plagioclase phenocrysts, pyroxene microlites and 10-150 µm-sized plagioclase microlites were acquired during the ascent of the magma in the conduit (Fig. 15). In addition, the presence of relatively large, irregular vesicles in the reddish groundmass and the skeletal morphology exhibited by the microlites are both consistent with a degassing during magma ascent. The loss of dissolved water is due to decompression, which produces an increase in liquidus temperature and thus can induce the high degree of undercooling required for a new nucleation event and a rapid rate of growth (Hammer and Rutherford, 2002; Calzolaio et al., 2010; Mollard et al., 2012; Shea and Hammer, 2013). The crystallisation that takes place in the conduit is therefore driven by the syn-eruptive degassing (Applegarth et al., 2013).

Rim formations around amphiboles and micas (Fig. 15) can be related to the decompression process (Pichler and Schnitt-Riegraf, 1997; Garcia and Jacobson, 1979; Rutherford and Hill, 1993; Tepley et al., 1999; Rutherford and Divine, 2003; Buckley et al., 2006). In particular, variation in the width of the amphibole rim can be related to variation in the velocity of the magma ascent in the conduit (Rutherford and Hill, 1993). As it is not possible to calculate the depth of magmatic chamber using the equation of Ridolfi et al. (2010) due to the unreliable estimation of pressure, we cannot estimate the ascent time. A short ascent time can be conjectured, because the magma consistently displays water heterogeneities (water-rich and water-poor zones) and there is no textural ripening of skeletal plagioclases (Couch et al. 2003; Blundy et al. 2006). However, since the amphibole rims have generally a thickness < 60 µm (Fig. 6) and crystal size < 20 µm (Fig. 7), we cannot exclude that they are heating-induced (De Angelis et al., 2015) or that they can be caused by both decompression and heating. Finally, the sets of fractures observed in micas in glomeroporphyric aggregates can be interpreted as “percussion figures”, as described by Xu and Ji (1991). The development of such a texture may indicate that the onset of degassing was abrupt enough to have caused a brittle deformation in micas at the beginning of magma ascent in the conduit (Fig. 15). It is worth noting that the inclusions in micas are small iron oxides rather than the sphene reported in Xu and Ji (1991). The difference in composition can be related to iron release according to the following Ti-oxy substitution (Takeda and Ross, 1975):

5.4 Eruption

The final eruption event corresponds to the formation of a dome, consistent with a high magma viscosity due to the high SiO₂ and crystal contents (Fig. 15). At this stage, cristobalite crystallises in the pore spaces within the reddish groundmass where the water vapor pressure

is highest (Tortorelli and More, 2003). Magma emplacement typical of a dome would explain the random orientations of the smallest microlites of plagioclase and pyroxene observed in thin section. Magma emplacement is also a favourable context for oxidation and would explain the formation of exsolution marks in the oxides, which generally occurs when temperature decreases and/or fO_2 increases (Lindsley, 1962; Buddington and Lindsley, 1964; Haggerty, 1976; Venezky and Rutherford, 1999; Turner et al., 2008). In a dome, the extent of oxidation-induced exsolution is controlled by variations in oxygen fugacity and cooling rates, which reflect the different conditions in the outer surface and the core of the dome (Saito et al., 2004). According to Boudon et al. (2015) exsolution in oxides is favoured by fluid migration due to the cristobalite precipitation that is a systematic process during a dome formation. In addition, the presence of iddingsite both in the cores and in the overgrowths of olivine phenocrysts, as well as the characteristic pleochroism of amphibole and mica, are features that result from oxidation processes (Smith et al., 1987; Tepley et al., 1999; Takeda and Ross, 1975). Oxidation is also likely the cause of the difference in colour of the two types of groundmass, with the reddish colour reflecting a higher degree of oxidation.

5.5 Formation of long-period polytypes of mica

A self-mixing mechanism, which is related to the formation of convection cells in the magmatic chamber, provides a favourable dynamic context for the crystallization of long period polytypes of mica. When this mineral undergoes cooling-heating cycles in the magma chamber, its spiral growth mechanism can be stopped by a decrease in magma temperature. In contrast, in the parts of the magmatic chamber where the temperature is higher, the spiral growth mechanism is dominant and other spirals can be activated. This interpretation is in agreement with Baronnet (1980), who related high temperature conditions during mica formation to the occurrence of long period polytypes in a vapour-rich system. Moreover, the

convection cells let the thin platy crystals of mica move freely as “*flying magic carpet*” during their growth (Sunagawa and Tomura, 1976). Both the number of active spirals and the mobility of mica crystals increase after the formation of convective cells, enhancing the possibility of spiral-spiral interactions, crystal-crystal and spiral-crystal interactions. These interactions “perturb” the growth and are, thus, related to the formation of the complex stacking sequence of long period polytypes according to the “*perturbative theory*” of Nespolo (2001). This theory states that the interaction of two or more spirals can generate a polytype with a new stacking sequence when at least one of spirals corresponds to a polytype different from *1M*. Otherwise, if all spirals correspond to the *1M* polytype, the only way to produce a new stacking sequence is thought the interaction of twinned spirals, i.e. rotated of $n60^\circ$ (Sunagawa and Tomura, 1976). In the crystal-crystal interaction, the coalescing crystals are imperfectly attached and one or more spirals develop along the boundary. Consequently, few layers of one crystal crystallize onto the surface of the other crystal, generating a new stacking sequence. The spiral-crystal interaction, also called “*platelet mechanism*” (Takeda and Ross, 1995) is due to the deposition of a small crystal on the spiral growth of another crystal; in this case a new polytype can form from the interactions of the stacking sequences of the two crystals.

It should be noted that, although the spiral-spiral, crystal-crystal and spiral-crystal interactions are favoured, long period polytypes are only one third of micas in the Ruiz Peak rhyodacite because the self-magma convection is a short transient mechanism in comparison with studies of other magmas (Welsch et al., 2009; Colin et al., 2012).

6. Conclusion

The uniqueness of Ruiz Peak rhyodacite is due to the presence of both long period and short period polytypes of mica. The coexistence of these polytypes and, in particular, the elevated

number of long period polytypes (one third of micas) are certainly related to the complex geological history of the rock. Petrographic and textural characteristics indicate that the complexity is not due to the mixing between mafic and felsic magmas, but to the development of convection cells associated with the self-mixing magma. This mechanism explains both occurrence and rarity of long period polytypes of mica, as well as some of observed disequilibrium textures (i.e. ripening dendritic overgrowths on olivine and seriate texture of plagioclase phenocrysts). Rims around mica and amphibole phenocrysts, “percussion figures” of mica, and skeletal microlites are formed later, during the ascent of magma in a conduit. At this stage, the degassing induces different degrees of vesiculation in the magma and, thus, the formation of two groundmasses. The differentiation of these groundmasses is completed after the eruption, during the magma emplacement in a dome: one of them becomes reddish due to a higher oxidation degree and cristobalite crystallises in its pores.

To conclude, the study of the Ruiz Peak rhyodacite has shown that, like reaction textures presented by other minerals, the occurrence of long period polytypes of mica in magmatic rocks can be considered witnesses of the complex history of magmas.

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Figure captions

Fig. 1. (a) OD vectors indicating the orientation of M layers. Stacking sequences of two short period polytypes, i.e. 1M and 3T (b) and a long period polytype (c) (32A₁ - Kogure and Nespolo, 1999). 1M, 3T and 32A₁ are Ramsdell symbols (1947) written as NS_n, where N is the number of layers, S is the symmetry and n generally indicates the order in which polytypes are discovered.

Fig. 2. Map of Jemez Mountains (after Kudo, 1974). The sample comes from the formation called “rhyodacite of La Jara Canyon” in the southern part of Ruiz Peak area (red circle).

Fig. 3. Photograph of rock sample N. 3149-8 illustrating the preferential orientation of plagioclase phenocrysts and regions of reddish and dark-grey coloured groundmass. Some regions of dark-grey groundmass are encircled to better recognize them.

Fig. 4. (a) A mica crystal surrounded by a dark rim. (b) Composition of mica crystals plotted in the classification diagram of Tischendorf et al. (2007), where the occupancy of the octahedral sheet is described using the parameters Mg minus Li (= *mgli*) and $^{VI}\text{Fe}_{\text{tot}} + \text{Mn} + \text{Ti}$ minus ^{VI}Al (= *feal*) a.p.f.u. Green circle = mica observed in groundmass; Red diamond = mean composition of micas in groundmass; Grey square = mica in glomeroporphyric aggregates.

Fig. 5. (a) Optical image of a crystal of mica in a glomeroporphyric aggregate. This basal section exhibits fractures that intersect at 60°, known as a “percussion figure”; (b) SEM image of iron-oxide-filled fractures in a mica crystal. (c) BSE image and (d) chemical analysis (wt.%) of zoning in mica with “percussion figures”.

Fig. 6. Frequency *versus* thickness of amphibole rims.

Fig. 7. (a) SEM image showing symplectite-like texture (Syp) surrounding an amphibole crystal (Am). (b) Magnification of symplectite-like texture.

Fig. 8. (a) Optical image of an euhedral crystal of olivine destabilized into iddingsite; (b) SEM image of an olivine phenocryst displaying the “closed hopper” ripening morphology; (c) SEM image of dendritic overgrowth observed on an olivine crystal.

Fig. 9. (a) SEM images of Type I plagioclase phenocryst; (b) cristobalite inclusion in Type I plagioclase; (c) Type II plagioclase phenocryst; (d) An-rich plagioclase that has developed on a wavy resorption horizon composed of An-poor plagioclase.

Fig. 10. Optical images of (a) an anhedral pyroxene in a glomeroporphyric aggregate and (b) euhedral pyroxene observed in the groundmass. (c) Pyroxene composition plotted in the Ca-Mg-Fe diagram (after Rock, 1990). Diamonds = pyroxene in groundmass; circles = pyroxene in glomeroporphyric aggregates.

Fig. 11. (a) BSE image of Fe-Ti oxides; (b) Ternary diagram for iron oxides (modified from Spencer and Lindsley, 1981). Diamonds = high-Ti oxides; circles = low-Ti oxides.

Fig. 12. (a) Optical and (b) BSE images of plagioclase microlites with skeletal morphology.

Fig. 13. SEM images of the Ruiz Peak rhyodacite groundmass. (a) Adjacent areas of reddish and dark-grey groundmass; (b) dark-grey groundmass and (c) reddish groundmass. (Pl = plagioclase; Px = pyroxene; Cris = cristobalite).

Fig. 14. (a) SEM image of cristobalite (Cris) in reddish groundmass; (b) Raman spectra of (1) cristobalite observed in the Ruiz Peak sample and (2) a reference α -cristobalite from the RUFF database (RUFF ID R060648).

Fig. 15. Schema showing at which stage of magmatic history the disequilibrium textures developed. (a) In the magmatic chamber, the development of convection cells is related to the formation of ripening dendrites on olivine crystals, of two types of plagioclase textures and of long period polytypes. (b) Formation of shear-zones in the conduit: the zone with vesicles more deformed and interconnected (red portion) are those of maximum strain in comparison with zones of lower strain where the vesicles are smaller and more regular-shaped (grey portion). During the ascent of the magma, plagioclase phenocrysts acquired preferential orientations, rims formed around amphiboles and micas, and “percussion figures” developed in micas. (c) Magma emplacement typical of a dome is related to the random orientations of small microlites, the formation of exolutions in oxides and the crystallization of cristobalite. At this stage, oxidation caused the different colour of two groundmasses: the groundmass with more deformed and interconnected vesicles became reddish due to a higher degree of oxidation.

Tables

Table 1. Keres Group stratigraphy after Guilbeau (1982) and Kudo & Guilbeau (1985). *

Studied sample.

Era	System	Series	Lithologic units				
Cenozoic	Tertiary	Middle Pliocene	Keres Group	Bearhead Rhyolite	Flow and dome member		
					Peralta Tuff member		
				Paliza Canyon Formation	Rhyodacite member		
					Dacite member		
					Upper andesite member		
					Lower member		
		Rhyodacite of La Jara Canyon*					
		Canovas Canyon Rhyolite		Dome member			
				Flow and tuff member			
		Basalt of Chamisa Mesa			Olivine basalt flows		

Table 2. EPMA analyses of amphiboles and mean formula.

Oxide (wt. %)	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	MnO	Cr ₂ O ₃	F	Total
	41.91	3.63	11.94	11.69	15.15	11.28	2.75	0.52	0.18	0.1	0.28	99.32
	41.72	3.54	12.38	11.35	14.97	11.01	2.60	0.45	0	0.06	0.07	98.14
	42.04	4.08	11.49	11.85	14.65	11.49	2.55	0.64	0.16	0	0	98.95
	41.65	3.94	12.33	11.82	14.88	11.27	2.50	0.47	0.09	0	0.18	99.12
	40.87	3.95	12.56	12.35	14.40	11.25	2.89	0.48	0.07	0.01	0.31	99.12
	40.96	4.01	11.71	11.81	14.65	11.26	2.50	0.48	0.29	0	0	97.66
	41.78	4.07	11.41	11.12	15.38	10.94	2.71	0.40	0.11	0	0.39	98.30
	42.28	3.75	11.58	11.92	15.23	11.10	2.64	0.43	0.21	0.01	0.52	99.65
	43.33	4.23	9.73	11.00	15.18	11.12	2.58	0.96	0.34	0	0.55	99.02
a.p.f.u.	Si	Ti	Al _{tot}	Fe _{tot}	Mg	Ca	Na	K	Mn	Cr	F	
	6.02	0.39	2.02	1.41	3.25	1.74	0.76	0.10	0.02	0.01	0.13	
	6.03	0.39	2.11	1.37	3.23	1.71	0.73	0.08	0	0.01	0.03	
	6.08	0.38	1.96	1.50	3.16	1.78	0.72	0.12	0.02	0	0	
	5.99	0.43	2.09	1.42	3.19	1.74	0.70	0.09	0.01	0	0.08	
	5.92	0.43	2.15	1.50	3.11	1.75	0.81	0.09	0.01	0	0.14	
	5.98	0.41	2.02	1.44	3.19	1.76	0.71	0.09	0.04	0	0	
	6.06	0.44	1.95	1.35	3.33	1.70	0.76	0.07	0.01	0	0.18	
	6.06	0.41	1.96	1.43	3.26	1.71	0.73	0.08	0.03	0	0.23	
	6.31	0.46	1.67	1.34	3.29	1.73	0.73	0.18	0.04	0	0.25	
Average	6.05	0.42	2.00	1.42	3.22	1.73	0.74	0.10	0.02	0	0.12	
σ	0.11	0.03	0.14	0.06	0.07	0.03	0.04	0.03	0.01	0	0.09	

Table 3. EPMA analyses of the most representative plagioclases (Type I and II phenocrysts and skeletal microlites).

	Oxide (wt. %)	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	MnO	Cr ₂ O ₃	Total	Ab%	An%	Or%
Type I	Core	53.60	0.01	28.77	0.42	0.03	11.23	4.94	0.20	0.08	0.01	99.48	43.47	54.66	1.87
Type I	Core	54.07	0.08	28.20	0.74	0.02	10.98	5.13	0.38	0.05	0	100.02	44.81	52.99	2.21
Type I	Core	55.19	0.05	27.74	0.52	0.06	10.44	5.53	0.38	0	0	100.21	48.44	49.08	2.48
Type I	Rim	57.70	0.07	25.88	0.38	0.03	7.59	6.97	0.63	0.05	0	99.35	60.16	36.24	3.60
Type I	Rim	58.07	0	26.06	0.47	0.01	7.94	6.93	0.63	0.04	0	100.24	59.07	37.42	3.52
Type I	Rim	56.96	0.09	25.28	0.64	0.03	7.96	6.51	0.70	0	0.05	98.30	57.25	38.71	4.04
Type II	Core	57.60	0.02	26.46	0.31	0.01	8.58	6.52	0.52	0	0.02	100.09	56.20	40.87	2.92
Type II	Core	57.91	0.02	26.29	0.31	0.03	8.37	6.29	0.67	0.03	0	100.01	55.42	40.71	3.87
Type II	Core	57.63	0.12	26.02	0.43	0.02	8.20	6.61	0.58	0	0.01	100.12	57.36	39.32	3.32
Type II	Rim	53.76	0.04	28.18	0.61	0.11	10.94	5.09	0.40	0.09	0	99.43	44.67	53.03	2.30
Type II	Rim	54.02	0	28.04	0.61	0.06	10.46	5.13	0.49	0.09	0	99.32	45.66	51.45	2.89
Type II	Rim	52.93	0.03	29.18	0.54	0	11.76	4.32	0.29	0.08	0	99.19	39.23	59.02	1.75
Microlite		54.19	0.08	28.54	0.73	0.11	11.24	5.01	0.33	0.03	0	100.59	43.78	54.34	1.89
Microlite		52.93	0.10	28.57	0.49	0.11	11.65	4.68	0.30	0	0.03	98.98	41.38	56.89	1.73
Microlite		54.50	0.09	28.96	0.74	0.05	11.35	4.74	0.35	0.06	0.01	100.84	42.17	55.77	2.06

890 **Table 4.** EPMA analyses of clinopyroxene and mean formula (*: pyroxene in
891 glomeroporphyric aggregates).

Oxide (wt. %)	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	MnO	Cr ₂ O ₃	Total
	52.41	0.39	0.98	6.95	16.19	21.71	0.53	0	0.62	0	99.81
	53.06	0.34	1.03	7.33	16.33	21.61	0.54	0	0.74	0.02	100.99
	52.36	0.45	1.19	7.34	16.09	20.99	0.53	0	0.63	0	99.57
	52.21	0.62	1.73	7.61	15.64	20.89	0.63	0	1.03	0.02	100.40
	52.19	0.55	1.49	7.94	15.49	20.41	0.66	0.01	0.75	0.09	99.58
	53.52	0.41	1.08	6.91	16.05	21.70	0.47	0	0.61	0	100.74
	53.29	0.32	1.05	6.95	16.25	21.40	0.49	0	0.75	0	100.49
	52.94	0.28	0.85	7.24	15.81	21.73	0.52	0	0.81	0	100.19
	53.00	0.34	1.10	6.87	15.95	21.51	0.48	0	0.63	0	99.90
	52.31	0.39	1.25	7.12	15.85	21.62	0.56	0	0.82	0	99.94
	53.09	0.33	0.87	7.14	16.13	21.65	0.52	0.01	0.81	0	100.54
	52.53	0.33	1.13	7.40	15.78	21.46	0.64	0.07	0.43	0	99.76
	52.85	0.36	0.95	7.01	16.05	21.42	0.49	0	0.71	0.03	99.92
	53.46	0.34	0.87	6.98	15.69	22.09	0.68	0.01	0.78	0	100.93
*	53.39	0.32	1.00	7.30	16.10	21.36	0.59	0	0.67	0	100.74
*	53.39	0.33	0.99	7.44	15.9	21.60	0.71	0	0.84	0	101.32
*	52.35	0.36	1.29	7.38	15.95	20.83	0.52	0.01	0.65	0.09	99.53
*	52.51	0.44	1.51	7.71	15.77	21.03	0.59	0.01	0.77	0.01	100.36
*	52.42	0.49	1.30	7.37	15.93	21.69	0.69	0	0.55	0	100.44
*	51.96	0.39	1.17	7.05	16.21	21.77	0.54	0	0.75	0.02	99.89

a.p.f.u.	Si	Ti	Al _{tot}	Fe _{tot}	Mg	Ca	Na	K	Mn	Cr
	1.93	0.01	0.04	0.21	0.89	0.86	0.04	0	0.02	0
	1.93	0.01	0.04	0.22	0.89	0.84	0.04	0	0.02	0
	1.93	0.01	0.05	0.23	0.89	0.83	0.04	0	0.02	0
	1.92	0.02	0.07	0.23	0.86	0.82	0.05	0	0.03	0
	1.93	0.02	0.07	0.25	0.86	0.81	0.05	0	0.02	0
	1.96	0.01	0.04	0.21	0.87	0.85	0.03	0	0.02	0
	1.95	0.01	0.05	0.21	0.89	0.84	0.03	0	0.02	0
	1.95	0.01	0.04	0.22	0.87	0.86	0.04	0	0.03	0
	1.95	0.01	0.05	0.21	0.88	0.85	0.03	0	0.02	0
	1.93	0.01	0.05	0.22	0.87	0.85	0.04	0	0.03	0

	1.94	0.01	0.04	0.22	0.88	0.85	0.04	0	0.03	0
	1.94	0.01	0.05	0.23	0.87	0.85	0.05	0	0.01	0
	1.95	0.01	0.04	0.22	0.88	0.85	0.03	0	0.02	0
	1.95	0.01	0.04	0.21	0.85	0.86	0.05	0	0.02	0
*	1.95	0.01	0.04	0.22	0.88	0.84	0.04	0	0.02	0
*	1.94	0.01	0.04	0.23	0.86	0.84	0.05	0	0.03	0
*	1.94	0.01	0.06	0.23	0.88	0.83	0.04	0	0.02	0
*	1.93	0.01	0.07	0.24	0.86	0.83	0.04	0	0.02	0
*	1.92	0.01	0.06	0.23	0.87	0.85	0.05	0	0.02	0
*	1.91	0.01	0.05	0.22	0.88	0.86	0.04	0	0.02	0
Average	1.94	0.01	0.05	0.22	0.87	0.84	0.04	0	0.02	0
σ	0.01	0	0.01	0.01	0.01	0.01	0.01	0	0	0

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Table 5. SEM-EDX analysis of Fe-Ti oxides (wt. %). FeO and Fe₂O₃ in oxides were calculated using ILMAT (Lepage, 2003).

Lighter zone	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ measured	MgO	Total	FeO calculated	Fe ₂ O ₃ calculated	Darker zone	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ measured	MgO	Total	FeO calculated	Fe ₂ O ₃ calculated
	5.69	1.20	91.68	1.44	96.33	33.20	54.80		39.31	1.76	56.90	2.03	96.40	31.70	21.60
	0.52	0.56	98.72	0.20	96.68	30.30	65.10		34.79	5.62	57.32	2.28	96.99	27.20	27.10
	5.70	1.17	91.51	1.63	96.40	32.90	55.00		43.24	1.69	52.13	2.94	96.47	33.70	14.90
	5.23	1.47	91.52	1.63	96.23	32.70	55.20		38.43	2.13	57.46	1.84	96.40	31.50	22.50
	4.73	1.29	92.27	1.71	96.43	31.90	56.80		36.61	2.00	60.18	1.20	96.61	30.80	26.00
	3.52	1.18	93.94	1.37	96.47	31.30	59.10		38.10	1.65	59.32	0.93	96.38	32.60	23.10
	2.15	1.45	95.42	0.98	96.58	30.70	61.30		43.97	1.51	50.77	3.75	96.33	32.90	14.20
	1.76	1.73	95.95	0.56	96.55	31.00	61.50		39.49	2.04	56.03	2.43	96.56	31.20	21.40
	5.17	1.03	92.43	1.38	96.38	32.80	56.00		40.26	1.77	55.55	2.41	96.44	31.90	20.10
	6.33	1.19	91.09	1.39	96.21	33.80	53.50		38.22	2.09	57.93	1.76	96.47	31.20	23.20
	8.22	0.57	91.00	0.21	95.90	37.20	49.70		33.27	2.29	63.57	0.87	96.83	28.40	32.00
	5.78	0.91	91.73	1.58	96.37	33.00	55.10		35.34	2.17	61.91	0.59	96.50	30.70	27.70
	6.66	1.49	89.52	2.34	96.39	32.80	53.10		38.42	1.55	57.73	2.30	96.67	30.50	23.90
	7.45	1.95	88.43	2.17	96.27	33.80	50.90		38.94	1.74	56.66	2.66	96.64	30.30	23.00
	8.37	2.08	87.39	2.16	96.21	34.70	48.90		35.95	2.03	61.33	0.68	96.46	31.10	26.70
	5.50	1.85	90.98	1.67	96.42	32.80	54.60		35.79	2.02	61.49	0.70	96.51	30.90	27.10
	4.68	1.40	93.04	0.88	96.26	33.10	56.20		38.55	2.20	57.28	1.97	96.52	31.20	22.60
Average	5.14	1.32	92.15	1.37	96.36	32.82	55.69		38.16	2.13	57.86	1.84	96.54	31.05	23.63
σ	2.16	0.43	2.75	0.63		1.62	4.23		2.78	0.93	3.32	0.80		1.52	4.43

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Table 6. Chemical compositions of groundmass (wt. %).

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Dark grey groundmass	Probe diameters	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	Total
1	60 µm	68.58	0.42	14.64	1.87	0.51	2.06	4.95	4.69	97.72
2	60 µm	68.82	0.43	15.48	1.61	0.45	1.85	5.32	4.53	98.49
3	30 µm	69.95	0.97	14.4	2.06	0.50	1.46	4.77	5.19	99.30
4	30 µm	68.45	0.30	15.3	1.36	0.36	2.15	4.93	4.82	97.67
5	15 µm	68.20	0.27	15.44	1.63	1.04	2.10	5.72	3.90	98.30
6	15 µm	69.55	0.52	15.17	1.17	0.25	1.57	4.68	5.71	98.72
Average		68.94	0.49	15.07	1.62	0.52	1.87	5.06	4.81	
σ		0.70	0.25	0.45	0.32	0.27	0.29	0.39	0.61	

Reddish groundmass	Probe diameters	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	Total
1	60 µm	67.24	0.57	15.13	1.71	0.81	2.03	5.30	4.19	96.98
2	60 µm	66.24	0.32	14.19	1.44	2.24	1.89	4.68	4.48	95.48
3	30 µm	67.94	0.50	14.30	2.03	0.96	1.97	4.95	4.19	96.84
4	30 µm	64.13	0.57	13.96	2.14	2.77	2.22	4.95	4.49	95.23
5	15 µm	69.12	0.17	14.93	1.53	0.98	1.99	4.80	4.46	97.98
6	15 µm	62.02	0.75	14.32	2.53	2.34	2.00	4.66	4.99	93.61
Average		66.14	0.48	14.47	1.90	1.68	2.02	4.89	4.47	
σ		2.62	0.21	0.46	0.41	0.86	0.11	0.24	0.29	

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	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MgO	Na ₂ O	K ₂ O	Tot.
1	39.90	6.42	14.76	13.07	15.02	0.63	9.70	99.50
2	38.59	6.96	14.08	16.21	13.51	0.71	9.46	99.52
3	38.91	6.79	13.83	13.74	15.21	0.80	8.72	98.00
4	38.72	6.67	13.62	11.21	16.90	0.93	9.06	97.11
5	38.55	6.72	13.79	11.99	16.50	0.89	8.55	96.99
6	38.36	6.42	13.49	11.41	16.53	0.92	8.49	95.62
7	38.27	6.72	13.76	11.59	16.33	0.93	8.89	96.49

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