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Integrating field, textural and geochemical monitoring to track eruption triggers and dynamics: a case-study from Piton de la Fournaise

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Abstract

The 2014 eruption at Piton de La Fournaise (PdF), la Reunion, which occurred after 41 months of quiescence, began with surprisingly little precursory activity, and was one of the smallest so far observed at PdF in terms of duration (less than 2 days) and volume (less than 0.4 Mm^3). The pyroclastic material was composed of spiny-opaque, spiny-iridescent, and fluidal scoria along with golden pumice. Density analyses performed on 200 lapilli reveal that the spiny-opaque clasts are the densest (1600 kg/m^3) and richest in crystals (54 vol%), and the golden pumices are the lightest (400 kg/m^3) and poorest in crystals (14 vol%). The connectivity data indicate that the fluidal and golden (Hawaiian-like) clasts have more isolated vesicles (up to 40%) than the spiny (Strombolian-like) clasts (0-5%). These textural variations are linked to primary pre-eruptive magma storage conditions. The golden and fluidal fragments track the hotter portion of the melt, in contrast to the spiny fragments which mirror the cooler portion of the shallow reservoir. Progressive tapping of these distinct

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portions leads to a decrease in the explosive intensity from early fountaining to Strombolian activity. The geochemical results confirm the absence of new hot input of magma and confirm the involvement of a single, shallow, differentiated magma source, possibly related to residual magma from the November 2009 eruption. We found that the eruption was triggered by water exsolution, favoured by the shallow depth of the reservoir, rather than cooling and chemical evolution of the stored magma.

Key words : Piton de La Fournaise, Hawaiian activity, Strombolian activity, shallow reservoire, texture, petrology, geochemistry

1. Introduction

A detailed characterization and understanding of eruptive dynamics and of processes driving and modulating volcano unrest is crucial in monitoring active volcanoes and fundamental for forecasting volcanic eruptions (Sparks, 2003). Many studies suggest that eruptive phenomena are strongly dependent on the physico-chemical properties of ascending magma in the conduit (e.g., temperature, viscosity, porosity, and permeability) (e.g. Sparks, 1978; Rust and Cashman, 2011; Gonnermann and Manga, 2013; Polacci et al., 2014). Integrating petrographic, chemical and textural data can thus provide critical information to constrain both the pre-eruptive storage conditions, and the processes related to magma ascent, degassing and cooling (e.g., reference in Table 1 in Gurioli et al., 2015). This multidisciplinary approach is of even greater importance in the monitoring of volcanoes which emit relatively unchanging magma compositions over time, like basaltic volcanoes (e.g. Di Muro et al., 2014; Gurioli et al., 2015; Coppola et al., 2017). As a result, monitoring of textures, and petrochemical properties of lava fragments and pyroclasts is now routinely carried out on a daily basis at active volcanoes such as Kilauea, Etna, and Stromboli (e.g., Thornber et al., 2003; Polacci et al., 2006; Swanson et al., 2009; Tadeucci et al., 2002; Colo’ et al., 2010; Houghton et al., 2011; 2013; 2016; Carey et al., 2012; 2013; Lautze et al., 2012; Andronico et al., 2013a; b; 2014; Corsaro and and Miraglia, 2014; Di Muro et al., 2014; Eychienne et al., 2015; Gurioli et al.; 2014; Leduc et al., 2015; Le Pennec and Leibrandt, 2015; Kahl et al., 2015). In the past, time series of petrographic and geochemical data have been measured for Piton de la Fournaise (PdF) basalts and particularly for effusive products. The aim of these datasets was to constrain potential magma evolution for one of the most active basaltic volcanoes of the world (e.g. Albarède et al., 1997; Vlastélic et al., 2005; 2007, 2009; Vlastèlic

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and Pietruszka, 2016; Schiano et al., 2012; Boivin and Bachèlery, 2009; Peltier et al., 2009; Lénat et al., 2012; Di Muro et al., 2014; 2015). However, this type of approach has seldom been coupled with detailed textural studies at PdF and instead has mostly focused on crystal textures and crystal size distribution (Welsch et al., 2009; 2013; Di Muro et al., 2014; 2015). Moreover, only sporadic data exist on the textures of pyroclasts ejected by the PdF (Villemant et al., 2009; Famin et al., 2009; Michon et al., 2013; Vlastelic et al., 2013; Welsch et al., 2009; 2013; Morandi et al., 2015; Di Muro et al., 2015; Ort et al., 2016).

Within this paper, we present a multidisciplinary textural, chemical and petrological approach to quantify and understand the short-lived 2014 PdF eruption. This approach combines detailed study of the pyroclastic deposit (grainsize and componentry) with bulk texture analysis (density, vesicularity, connectivity, morphology, vesicle distribution and crystal content) and a petro-chemical study (bulk rock, glass, minerals, melt inclusions) of the same clasts. This integrated approach has now been formalized within the National Observation Service for Volcanology (SNOV) as routine observational systems (DynVolc, Dynamics of Volcanoes, http://wwwobs.univ-bpclermont.fr/SO/televolc/dynvolc/ and GazVolc, Observation des gaz volcaniques, http://wwwobs.univ-bpclermont.fr/SO/televolc/gazvolc) to provide data for the on-going activity at PdF (Harris et al., 2017).

In spite of being the first of a series of eruptions, the June 2014 event was preceded by only weak inflation and by a rapid increase in number of shallow (< 2 km below volcano summit) volcano tectonic earthquakes that happened only 11 days before the eruption (Peltier et al., 2016). The eruptive event was dominantly effusive, lasted only 20 hours and emitted a very small volume of magma (ca. 0.4 Mm^3, Peltier et al., 2016), which makes this event one of the smallest, in terms of duration and volume, observed at PdF up to now. In addition, the eruption started during the night and very little direct observation exists for the first few hours of the activity, when the lava effusion was associated with very weak fountaining activity and Strombolian explosions.

This eruption occurred just outside the southern border of the summit Dolomieu caldera, at the top of the central cone of PdF (Fig. 1). This is a high risk sector because of the high number of tourists. Identification of precursors of this kind of activity represents an important challenge for monitoring systems (Bachélery et al., 2016).

Therefore this eruption represents an ideal context to apply our multidisciplinary approach, with the aim of addressing the following key questions:

(i) why was such a small volume of magma erupted instead of forming an

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intrusion?

(ii) what caused the rapid trigger and the sudden end to this small volume eruption?

(iii) which was the source of the eruption (shallow versus deep, single versus multiple small magma batches)? what was the ascent and degassing history of the magma? what was the time and space evolution of the eruptive event?

Furthermore, this eruption provides an exceptional opportunity to study processes leading to the transition from mild Hawaiian (<20 m high fountains, following the nomenclature proposed by Stovall et al., 2011) to Strombolian activity, whose products are little modified by post-fragmentation processes because of the very low intensity of the activity.

Finally, with these results we want to stress how combined textural and petro-chemical quantification of the eruptive products can be used to characterize on-going activity, and to provide valuable information to understand both the causes and the dynamics of potentially harmful eruptions.

2 The 2014 activity

2.1 Precursory activity

The 20 June 2014 summit eruption represents the first eruption at PdF after 41 months of quiescence. The last previous eruption had been on 9 December 2010, with a shallow (above sea level) intrusion on 2 February 2011 (Roult et al., 2012). From 2011, the deformation at PdF was constant with two distinct types of behaviour: (i) a summit contraction of a few centimetres every year; (ii) a preferential displacement of the east flank at a rate of 1-3 centimetres per year (Brenguier et al., 2012; Staudacher and Peltier, 2015). The background microseismicity was very low (< 5 shallow events/day below volcano summit) and low-temperature summit intracaldera fumaroles emitted very little sulphur (H_2S or SO_2) and carbon (CO_2) (Di Muro et al., 2012; 2016). After 41 months of rest, a new intense cycle of activity (June 2014, February 2015, May 2015, July 2015, August-October 2015; May 2016; September 2016; January 2017 and the on-going eruption that had started the 14 of July 2017) began with surprisingly little precursory activity.

The 2014 summit eruption started during the night of June 20/21, at 21h35 GMT (0h35 local time) and ended on June 21 at 17h09 GMT (21h09 local time), after less than 20 hours of dominantly effusive activity. The volcano reawakening was preceded, in March and April

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2014, by deep (15-20 km below sea level) eccentric seismicity and increase in soil CO_2 flux below the western volcano flank, 15 km NW of the volcano summit (Liuzzo et al., 2015; Boudoire et al., 2017). Background microseismicity and inflation of the central cone increased progressively starting on 9 June 2014. On June 13, 17 and 20, three shallow (above sea level) intense seismic crises occurred below the summit Dolomieu caldera (Fig. 1), with hundreds of events located in a narrow depth range between 1100 and 2100 metres below the volcano summit. These seismic crises consisted of swarms of low magnitude (M: 1-2) volcano tectonic events which increased in number from the first to the third crisis. On June 20, seismicity increased progressively and a final seismic crisis started at 20h20 GMT, only 75 minutes before the eruption. This seismic crisis was coupled with acceleration in the deformation of the summit area, which began only 60 minutes before the eruption. Interestingly, only slight inflation of the central cone (< 2 cm of dilatation) was detected 11 days before the 2014 eruption with a maximum of 1 cm and 1.6 cm enlargement at the summit and the base of the cone, respectively (Peltier et al., 2016). A moderate increase in CO_2 and H_2S emissions from summit intracaldera fumaroles was detected starting on June 2, but only very minor SO_2 emissions occurred before the eruption (mostly on June 7 and 15, unpublished data). Therefore, the increase in both geophysical and geochemical parameters was mostly related to the late phase of injection of the dyke towards the surface. Following the end of the June 20-21 eruption, a long-term continuous inflation of the edifice began, at a moderate rate, and mostly at the base of the volcano. More than one year after this first eruption, the long-term deformation trends showed that the 2014 eruption marked a kink between the deflation trend which followed the caldera-forming 2007 eruption (Staudacher et al., 2009) and the currently ongoing continuous inflation trend (Peltier et al., 2016; Coppola et al., 2017).

2.2 Chronology of the events and sampling strategies

We reconstructed the chronology of events by combining a distribution map of the fissures, pyroclastic deposits and lava flows (Fig. 1) with a review of available images (visible and IR) and videos extracted from the observatory data base, the local newspapers, and web sites (Fig. 2). The 2014 eruption occurred at the summit and on the SE slopes (Figs. 2a and 2b) of the central cone and evolved quickly and continuously over 20 hours. The full set of fractures opened during a short period of time (minutes) and emitted short (<1.7 km long) lava flows

(Fig. 1 and Figs. 2c and 2d). Feeding vents were scattered along a 0.6 km long fissure set (Fig. 1) and produced very weak (low) Hawaiian to Strombolian activity (Fig. 2).

Fissures opened from west to east, initially sub-parallel to the southern border of Dolomieu caldera and then propagated at lower altitude (Fig. 1). The summit part of the fractures (ca. 2500 m asl, Western Fracture, Fig. 1) emitted only small volumes of lava and pyroclasts. This part of the fracture set was active only during the first few hours of the eruption, at night. The eastern part of the fractures (Upper Fracture, Fig. 1) descended to lower altitude (between 2400 and 2300 m asl, Middle Fracture, Fig. 1) along the SE flank of the summit cone and emitted most of the erupted volume (Figs. 2a and 2b). As often observed in PdF eruptions, the activity progressively focused on a narrow portion of the fractures at low altitude and finally on a single vent located at the lower tip of the fracture system (Main Vent, at 2336 m asl, Figs. 1, 2). The first in situ observations in the morning of June 21 (ca. 04h00 GMT) showed that weak Strombolian activity was focused on a narrow segment of the lower fractures and that a'a lavas had already attained the elevation of 1983 m asl (0.2 km before maximum runout). A small, weak gas plume was also blowing northwards. A single sample of partially molten lava was collected from the still active lava front and partially water quenched. During most of June 21, the activity consisted of lava effusion in three parallel lava streams merging in a single lava flow and mild-weak "Strombolian" explosions at several closely spaced spots along the lower part of the feeding fracture. At 13.00 (GMT), only weak explosions were observed within a single small spatter cone (Figs. 2e and 2f). Most of the lava field was formed of open channel a'a lavas. The total volume of lava was estimated by MIROVA service (https://www.sites.google.com/site/mirovaweb/home), with the use of the MODIS images and the analyses of the flux from the spectral properties, to be within 0.34 (+/- 0.12) million m^3, (Coppola et al., 2017). Satellite derived volume estimates are consistent with independent photogrammetric estimates (0.4 ± 0.2 Mm^3; Peltier et al., 2016) and rank the 2014 eruption at the lower end of the volume range typically emitted by Piton de la Fournaise (Roult et al., 2012).

Apart from the sample from the front of the still active lava flow, all other samples were collected in two phases: 3 days after the eruption (pyroclasts on June 24, Fig. 3a; lavas on July 2) and three months later (pyroclasts from the Main Vent; November 18) (Table S1). June 24 samples were collected both from the main fractures, the Main Vent and the active lava flow (Fig. 1 and Table S1). Scattered scoriaceous bombs and lapilli were collected from the discontinuous deposits emplaced close to the Western Fracture, active only at the beginning of the eruptive event (Figs. 3c and 3d). In contrast, the sustained and slightly more energetic activity at the lower tip of the fractures built a small spatter cone and accumulated a small volume of inversely graded scoria fallout. This deposit is 10 cm thick at 2 m from the vent and covers an area of about ~1000 m^2 (Main Vent, Fig. 1). For this fall deposit we collected two bulk samples, one from the base (within the lower 5 cm) and the other from the top (within the upper 5 cm), for the grain size and componentry analyses. The sample at the base was collected in November because on June 24 the loose proximal lapilli blanket was still very hot (405 °C; thermocouple measurement, Fig. 3a) and fumaroles with outlet temperatures in the range 305-60 °C were observed all along the fractures several weeks after the eruption. Both in June and in October, more than 200 clasts of similar size (maximum diameter between 16 and 32 mm, see Gurioli et al. 2015) were collected, both close to the Main Vent and in the 'distal' area (30 metres away from the vent) for density, connectivity, permeability, petrological and geochemical analysis.

3. Methodology

3.1 Grain size, componentry and ash morphology

We performed grain size analyses on the two bulk samples collected from the Main Vent, following the procedure of Jordan et al. (2015) (Table S2). The samples were dried in the oven at 90°C and sieved at ½ phi intervals in the range of -5 φ to 4 φ (Fig. 3c); the data are also shown in full phi for comparison with the deposits of the 2010 PdF fountaining episode (Hibert et al., 2015; Fig. 3f). Sieving was carried out by hand and for not longer than three minutes to avoid breaking and abrasion of the very vesicular and fragile clasts. For the scattered scoria sampled from the Western Fracture (Figs. 1, 3c and 3d), we followed the grain size strategy proposed in Gurioli et al. (2013). Within this procedure we sampled each fragment and we recorded the weight and the three main axes (a being the largest, b, and c). To allow comparison with the sieving grain size analyses (Inman, 1952), we used the intermediate b axis dimension to obtain $\varphi = -\log_2 b$.
Following the nomenclature of White and Houghton (2006) the componentry analysis is the subdivision of the sample into three broad components: i) juvenile, ii) non-juvenile particles, and iii) composite clasts. The juvenile components are vesicular or dense fragments, as well as crystals, that represent the primary magma involved in the eruption; non-juvenile material includes accessory and accidental fragments, as well as crystals that predate the eruption from which they are deposited. Finally, the composite clasts are mechanical mixtures of juvenile and non-juvenile (and/or recycled juvenile) clasts. In these mild basaltic explosions, the non-

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juvenile component is very scarce, so we focused on the juvenile component that is characterized by three groups of scoria: (i) spiny-opaque, (ii) spiny-glassy, and (iii) fluidal, along with golden pumice (Fig. 4). The componentry quantification was performed for each grain size fraction between -5 φ to 0.5 φ (Figs. 5a and 5b), where a binocular microscope was used for the identification of grains smaller than -1 phi (Table S2). For the coarse ash fraction (250-300 µm size) of the two bulk deposits collected at the Main Vent, we also performed a morphological quantification using the Morphologi G3 at Laboratoire Magmas et Volcans (LMV) of Clermont-Ferrand following the procedure of Leibrandt and Le Pennec (2015) to distinguish between smooth versus spiny clasts within the coarse ash (Fig. 5c).

In the following, we will use the crystal nomenclature of Welch et al. (2009), with the strictly descriptive terms of macrocrysts (> 3 mm in diameter) mesocrysts (from 0.3 to 3 mm in diameter), and microcrysts (<0·3 mm in diameter).

3.2 Particle bulk texture (density, porosity, connectivity) and microtexture

For each sample site, we selected 27 to 146 juvenile particles within the 8-32 mm fraction for density and pycnometry measurements (Table S3). This is the smallest granulometric fraction assumed to be still representative of the larger size class in terms of density (Houghton and Wilson, 1989; Gurioli et al., 2015), and has been used in previous textural studies (e.g., Shea et al., 2010). In addition, this size range is ideal for vesicle connectivity measurements (e.g. Formenti and Druitt, 2003; Giachetti et al., 2010; Shea et al., 2012; Colombier et al., 2017a, b). Textural measurements (density/porosity, connectivity) were performed at the LMV. Density of juvenile particles was measured by the water-immersion technique of Houghton and Wilson (1989), which is based on Archimedes principle. A mean value for the vesicle-free rock density was determined by powdering clasts of varying bulk densities, measuring the volumes of known masses using an Accupyc 1340 Helium Pycnometer, then averaging. The same pycnometer was also used to measure vesicle interconnectivity for each clast using the method of Formenti and Druitt (2003) and Colombier et al. (2017a). Vesicle size distribution and crystal content were performed following the method of Shea et al. (2010) on nine clasts picked up from each component-density distribution mode (Fig. 6c). These data are presented in Figure 4 and we followed the strategy of Leduc et al. (2015) for the quantification of the vesicle size distribution.

A full description of the protocol for density and connectivity measurements is available at http://wwwobs.univ-bpclermont.fr/SO/televolc/dynvolc/index.php, while the textural data are available at http://wwwobs.univ-bpclermont.fr/SO/televolc/dynvolc/bdd.php.

3.3 Bulk geochemistry

For the determination of the bulk chemistry (Table S4 and Fig. 7) of the different pyroclasts we selected the largest pyroclasts of golden pumice and the largest fluidal, spiny glassy and spiny opaque scoriae. We also analyzed two fragments of lava, from the beginning and the end of the eruption. Samples were crushed into coarse chips using a steel jaw crusher and powdered with an agate mortar. Major and trace element compositions were analyzed using powder (whole rock composition). In addition, for a sub-set of pyroclasts, glass chips (2-5 mm in size) were hand-picked under a binocular microscope and analyzed separately for trace elements. For major element analysis, powdered samples were mixed with $LiBO_2$, placed in a graphite crucible and melted in an induction oven at 1050 °C for 4.5 minutes, resulting in a homogeneous glass bead. The glass was then dissolved in a solution of deionized water and nitric acid (HNO_3), and finally diluted by a factor of 2000. The final solutions were analyzed by ICP-AES. Trace element concentrations were analysed following a method modified from Vlastelic et al. (2013). About 100 mg of sample (powder and chip) were dissolved in 2 ml of 28M HF and 1 ml of 14M HNO_3 in teflon beaker for 36 hours at 70°C. Solutions were evaporated to dryness at 70°C. The fluoride residues were reduced by repeatedly adding and evaporating a few drops of concentrated HNO_3, before being fully dissolved in ca. 20 ml of 7M HNO_3. These solutions were diluted by a factor of 15 with 0.05M HF (to reach rock dilution factor of ca. 4000) and trace element abundances were determined by quadrupole ICPMS (Agilent 7500). The analyses were performed in plasma robust mode (1550 W). The reaction cell (He mode) was used to reduce interference on masses ranging from 45 (Sc) to 75 (As). The signal was calibrated externally (every 4 samples) with a reference basaltic standard (USGS BHVO-2) dissolved as for the samples and using the GeoRem recommended values (http://georem.mpch-mainz.gwdg.de/). For elements that are not well characterized in literature (As, Bi, Tl), or which show evident heterogeneity (e.g. Pb) in BHVO-2 powder, the signal was calibrated using the certified concentrations of a synthetic standard, which was also repeatedly measured. The external reproducibility (2σ error) of the method is 6% or less for lithophile elements and 15% or less for chalcophile elements.

3.4 Glass and crystal chemistry

Spot analyses of matrix glass and crystal composition (Table S5) were carried out using a Cameca SX100 electron microprobe (LMV), with a 15 kV acceleration voltage of 4 nA beam current and a beam of 5 µm diameter for glass analyses. However, for the spiny opaque scoria, characterized by abundant crystals with rapid growth textures, a voltage of 8 nA beam current and a beam of 10 µm diameter were used. For this latter sample, 10 analyses per sample were performed due to the heterogeneity within the highly crystallised glass (Fig. 8a), while for the other samples 6 analyses per sample were enough to characterize the clean homogeneous glass. For crystal analysis, a focused beam was used. For the characterization of the meso- and micro-crysts, due to their small size, only two measurements were performed, one at the edge and one at the core of the crystals, to check for possible zonation.

3.5 Melt inclusions

Melt inclusions (MIs; Table S6, Figs. 8b and 9) were characterized in the olivine mesocrysts from the three groups of scoriae (fluidal, spiny glassy and spiny opaque). Crystals in the pumice group were too rare and small to be studied for melt inclusions.

Olivine crystals were handpicked under a binocular microscope from the 100– 250 and 250– 600 µm grain size fractions of crushed tephra. Crystals with MIs were washed with acetone, embedded in epoxy and polished individually to generate adequate exposure of the MIs for *in situ* electron probe microanalysis. The MIs are spherical to oblate in shape and range in size from 10 to 200 µm. Some of the MIs contain shrinkage bubbles but all of those studied are totally deprived of daughter minerals. Major elements were measured on a Cameca SX-100 microprobe at LMV (Table S6). For major elements, the larger MIs were analyzed with a spot diameter of 10-20 **µ**m and sample current of 8 nA, whereas the smaller MIs were analyzed with a beam of 5 **µ**m and a sample current of 4 nA. The results are given in Table S6, and analytical details and uncertainties are listed in Óladóttir et al. (2011) and Moune et al. (2012).

4 Results

4.1 Deposit texture (grain size, componentry, morphology) and petrological description of the samples

The pyroclastic deposits at the Western and Upper fractures sites (Fig. 1a) are formed by scattered homogeneous smooth fluidal (Figs. 3d) bombs and lapilli. The average dimension of the fragments is around 4 cm (maximum axis) with bombs up to 10 cm and scoria lapilli up to 2 cm in size.

At the Main Vent, the reversely graded deposit is made up of lapilli and bombs, with only minor coarse ash (Fig. 3c). The lower 5 cm at the base are very well-sorted and show a perfect Gaussian distribution with a mode at 4 mm. In contrast, the grain size distribution of the upper 5 centimetres is asymmetrical with a main mode coarser than 22 cm and a second mode at 8 mm. This upper deposit is negatively skewed due to the abundance of coarse clasts. The dataset show a similarity between the grain size distributions of the basal tephra ejected from the 2014 main vent and the ones for the lava fountaining of the 2010 summit event (Hibert et al., 2015). On the contrary, the top of the 2014 fall differs from fountain deposits, being coarser and polymodal, and it is ascribed to dominantly Strombolian activity.

In terms of componentry of the deposits, four types of clasts were distinguished (Fig. 4): (i) golden pumice, (ii) smooth or rough fluidal scoriae, (iii) spiny glassy scoria, (iv) spiny opaque scoria. The pumices are vesicular, light fragments, characterized by a golden to light brown color, sometimes with a shiny outer surface (Fig. 4a). They are usually rounded in shape. Golden clasts studied for textures contain a few microcrysts of plagioclase (up to 0.1 mm in diameter), clinopyroxene up to 0.05-0.06 mm in diameter, and small olivine up to 0.03 mm in diameter (Fig. 4), together with large areas of clean, light brown glass. The fluidal scoria fragments have dark, smooth or rough shiny surfaces (Fig. 4b). They can be more or less elongated in shape and have spindle as well as flattened shapes. The fluidal fragments are characterized by rare mesocrysts of plagioclase and clinopyroxene and microcrysts of plagioclase, clinopyroxene and olivine (Fig. 4b). The spiny glassy fragments are dark, spiny scoria that range in shape from subrounded to angular (Fig. 4c). These fragments contain abundant glassy areas, while the spiny opaque fragments lack a glassy, iridescent surface. Both groups of spiny clasts are characterized by the presence of dark and light brown glass. The spiny opaque fragments are the densest fragments and have the largest amount of crystals. They contain, as the most abundant phase, relatively large meso- and micro-crysts of plagioclase, up to 3 mm long, together with meso- and micro-crysts of clinopyroxene and olivine (Figs. 4c and 4d). In the dark portions of their matrix, tiny fibrous microcrysts of olivine + clinopyroxene + plagioclase + Fe-Ti oxides occur. The spiny glassy fragments have the same crystal populations as the spiny opaque ones, but their plagioclases are much smaller and attain a maximum length of only 0.3 mm. Clusters of plagioclase and clinopyroxene are present in both the spiny opaque and the spiny glassy fragments, as well as rare macrocrysts of olivine. The olivine macrocrysts exhibit the typical compositional (Fo 84.2) and petrographic features of olivine phenocrysts described in previous studies (Clocchiatti et al., 1979; Albarede and Tamagnan, 1988; Bureau et al., 1998a and b; Famin et al., 2009; Welsch

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et al., 2013). They are automorphic, fractured with oxides (mostly chromite) and melt inclusions (Fig. 4c). Fluidal and pumice fragments studied for textures contain rare macrocrysts and mesocrysts of olivine, and the crystals are essentially microlites. The pumice and some fluidal fragments have lower contents of microlites than some fluidal and spiny fragments, with the latter having the highest microlite content (Table S4).

The componentry results are reported in Figure 5 for the Main Vent deposits; the deposits from the Fractures are characterized exclusively by fluidal clasts (Fig. 3). At the base of the Main Vent deposit, the coarse fraction of the deposit is rich in golden and fluidal components that represent more than 60-70vol% (Figs. 5a and 5b). The proportion of the two groups is similar. If we look at the Morphologi G3 results (Fig. 5c) for the coarse ash fragments, this population is formed exclusively by smooth fragments that correspond to fluidal and golden pumice. In contrast, in the upper, coarse grained fall deposit, the clasts bigger than 8 mm are dominated by the spiny scoria fragments, while the fraction smaller than 8 mm show a dramatic increase in the golden and fluidal fragments, with the fluidal ones always more abundant than the golden ones (Figs. 5a and 5b). The small amount of coarse ash fraction in the top deposit, however, is dominated by the presence of spiny fragments (Fig. 5c). Abundant light, golden, coarse lapilli pumice and bombs have been found scattered laterally up to 30 metres from the main axis and were not found in the proximal deposit. On the basis of the high amount of pumice in the lower part of the deposit, we correlate the large, light clasts with the base of the proximal deposit, and consequently we interpret them as material emitted at the beginning of the June 2014 eruptive event.

4.2 Particle density, porosity connectivity and texture

Density analyses performed on 200 coarse lapilli (Table S3) reveal a bimodal distribution, with a main population of light fragments having a mode at 800 kg/m^{-3}, and a second and denser population centered at 1400 kg/m^{-3} (Fig. 6a). The fluidal fragments, mostly collected at the Western Fracture (Fig. 1a), have a density range from 600 to 1400 kg/m^{-3} and a mode at 1000 kg/m^{-3} (Fig. 6b). The bulk deposit collected close to the Main Vent has a bimodal density distribution, with the golden and fluidal fragments forming the lower-density population and the spiny fragments being dominant in the denser population (Fig. 6c). For these samples there is a marked correlation between porosity and morphology, so that the spiny-opaque clasts are the densest (up to 1600 kg m^{-3}, with a vesicularity of 45 vol%) and the golden pumice are the lightest (minimum density of 400 kg m^{-3} with a vesicularity of up to 86 vol%; with a Dense Rock Equivalent density of 2.88 x 10^3 kg m^{-3}). The increase in vesicularity correlates with an increase in the amount of small (0.1 mm), medium (0.5-1 mm) and large (up to 4 mm) vesicles. In the fluidal clasts, these vesicles have a regular rounded or elliptical shape and are scattered throughout the sample. The lightest pumices are often characterized by the presence of a single, large central vesicle (10 – 15 mm) with the little vesicles and a few medium vesicles distributed all around it (Fig. 4). The spiny glass texture is characterized by a lower amount of large vesicles than in the pumice and by the presence of mostly medium sized vesicles, while the spiny opaque has more irregular shaped, very large (up to 10 mm) vesicles with a small and a medium sized bubble population. In the spiny glass samples, the glass is more or less brown, with the dark brown portions being the ones with the lowest vesicle content and the highest microlite content. The opaque samples have a central, very dark glass portion, with low vesicle content, and a more vesicular glassy portion at the outer edges (Fig. 4).

The connectivity data (Fig. 6d) also indicate that the fluidal and golden clasts have a larger amount of isolated vesicles (up to 40% in volume). The fluidal clasts from the Western Fracture are the most homogeneous with an average percentage of isolated vesicles around 30% in volume. Both the pumice and the fluidal fragments with high vesicularity are characterized by fewer amounts of isolated vesicles. Finally the spiny fragments have the lowest content of isolated vesicles (0-5% in volume).

4.3 Chemistry and geochemistry of the products

Major and trace element concentrations of whole-rock and hand-picked glass samples are reported in Table S4. Whole rock major element composition is very uniform (e.g., 6.5<MgO<6.7 wt%) and well within the range of Steady State Basalts (SSB), the most common type of basalts erupted at Piton de la Fournaise (Albarède et al., 1997). However, compatible trace elements, such as Ni and Cr, are at the lower end of the concentration range for SSB (<100ppm) indicating that the June 2014 eruption sampled relatively evolved melts. Ni and Cr generally show higher concentrations in 2014 bulk rocks (79<Ni<92ppm and 71<Cr <87ppm) compared to the 2014 glass chips (66<Ni<73ppm and 54<Cr <59ppm for all but two chips). In the Cr vs Ni plot (Fig. 7a), whole rocks plot to the right of the main clinopyroxene +/- plagioclase–controlled melt differentiation trend. This trend is controlled by the addition of Ni-rich olivine (Albarède and Tamagnan, 1988). We estimate that the Ni excess results from the occurrence of a low amount (0.7 to 1.3 wt%) of cumulative olivine in whole rocks, consistent with thin section observations. The composition of olivine macrocrysts (ca. Fo84) is too magnesian to be in equilibrium with the low-MgO evolved composition of the 2014 magma. Using our estimate for the amount of cumulative olivine, we recalculate the olivine–corrected MgO content of the 2014 magma at 6.2 wt%. The June 2014 melt is thus only moderately depleted in compatible elements compared to the previous eruption of December 2010 (MgO~6.6 wt%, Ni~80 ppm, Cr~120 ppm). Conversely, the June 2014 melt is significantly depleted in compatible elements compared to the earlier November 2009 eruption, which sampled relatively primitive magmas (average MgO~7.7 wt%, Ni~135 ppm, Cr~350 ppm) (Fig. 7a). The 2014 evolved composition plots at the low-Ni-Cr end of Piton de la Fournaise historical differentiation trend (Albarède and Tamagnan, 1988), near the composition of lavas erupted on 9 March 1998 after 5.5 years of quiescence (1992-1998). Note that olivine accumulation at PdF generally occurs in melt having ca.100 ppm Ni (Albarède and Tamagnan, 1988). Olivine accumulation in evolved melts (Ni < 70 ppm) seems to be a distinctive feature of many small post-2007 eruptions (e.g. this event and the three 2008 eruptions, see Di Muro et al., 2015).

A closer inspection of Ni-Cr variability in June 2014 whole rock samples (Fig. 7b) reveals that scoria from the Western Fracture (140624-9b-6, Table S4) and early erupted lavas (1406-21-1, Table S4) have the lowest amount of olivine (<0.9%) whereas scoria from the Upper Fracture (140624-13a) and late erupted lavas (140324-12) have a slightly higher amount of olivine (>1.2%). This is consistent with the general trends observed at PdF of olivine increase from the start to end of an eruption (Peltier et al., 2009).

The so called “olivine control trend” in Ni-Cr space cannot be explained either by addition of pure olivine (which contains less than 500 ppm Cr (Salaün et al., 2010; Di Muro et al., 2015; Welsch et al., 2009), or by the addition of olivine plus pyroxene (which would require ca. 50% pyroxene with 970 ppm Ni and 4800 ppm Cr, see Fig. 7 caption). Instead, addition of olivine hosting ca. 1% Cr-spinel (with 25 wt.% Cr) accounts for data and observations, and is consistent with crystallization of olivine and Cr-spinel in cotectic proportions (Roeder et al., 2006). The fact that some samples (golden pumice) plot off the main, well-defined array, can be explained either by addition of more or less evolved olivine crystals (with the range of Fo 80-85 measured in June 2014 samples) and/or slight variations (± 0.02%) in the proportion of Cr-spinels (Fig. 7b).

The glass chemistry of the four clast types allows us to correlate porosity and oxide contents and shows an increase in MgO from the spiny opaque to fluidal and golden fragments (Fig. 8a). Consistent with petrological and textural observations, the spiny opaque

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is the most heterogeneous type of clast in terms of glass composition (Fig. 8). The glassy portion at the edge of the clast is similar to the spiny glass, while the interior, characterized by dark areas rich in tiny fibrous microcrysts, shows scattered glass compositions with very low MgO content as well as a decrease in CaO (Fig. 8). We attribute the significant variation in glass composition within the different components to variable degrees of micro-crystallisation as the bulk chemistry of all clasts is very similar and globally homogeneous.

4.4 Melt inclusions

MI analyses must be corrected for post-entrapment host crystallisation at the MI - crystal interface. We used a Kd = $(FeO/MgO)_{ol}$ / $(FeO/MgO)_{melt}$ = 0.306 (Fisk et al., 1988; Brugier, 2016) and an average $Fe^{3+}/\Sigma Fe_{total}$ ratio of 0.11 (Bureau et al., 1998a; Di Muro et al., 2016 and references therein) defined for PdF magmas. For the June 2014 melt inclusions, the post entrapment crystallization (PEC) ranges from 2.9 to 10.5 wt%. Raw and corrected major and volatile element concentrations of MIs are reported in Table S6.

Host olivines span a large compositional range from Fo_{80} to Fo_{86}. Despite the evolved bulk composition of the magma, most olivines are quite magnesian (Fo_{83-85}) and are not in equilibrium with the evolved host magma. On the contrary, Mg-poor olivines (Fo_{80-81}) can be considered as being in equilibrium with the bulk rock composition. The corrected compositions of MIs in phenocrysts from the different samples partly overlap with the evolved bulk rocks (MgO_{wr}: 6.1-7.2 wt%) and extend to higher MgO contents of up to 8.8 wt% (Table S6). MIs display a narrow range of transitional basaltic compositions (K_2O= 0.5-0.9 wt%) and show no significant difference between the three types of scoriae. The major element composition of melt inclusions correlates with that of the host olivines. Melt inclusions in the high Fo-olivines have the highest MgO, CaO and TiO_2 and lowest K_2O concentrations (Table S6). It is interesting to note that the June 2014 products contain two populations of magnesian ($Fo_{>83}$) olivines hosting melt inclusions with two distinct Ca contents. Most of the magnesian olivines contain MIs with unusually high CaO contents (11.6 – 12.9 wt%) and high CaO/Al_2O_3 ratios (0.8-0.9), higher than that of the bulk rocks (0.8) (Fig 8). The occurrence of olivines with "high Ca" melt inclusions has been observed in all three different types of scoriae. A few magnesian olivines and all Mg-poor olivines ($Fo_{80.5-83.6}$) host MIs with lower CaO contents (11.4 wt%). This latter composition overlaps with that of the bulk rock (Fig 8). The "high Ca" population of inclusions is also enriched in TiO_2 and Al_2O_3 and depleted in MgO, FeO_T and Na_2O for a given olivine Fo content with respect to the "low

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Ca" population. Both low- and high-Ca populations of melt inclusions have similar K_2O contents and total alkali content increases from 3 wt% at 12.6 wt% CaO, to 3.5 wt% at 10.8 wt% CaO. However, we remark that high Ca melt inclusions from the June 2014 activity record a significant scattering in K_2O contents, which range from 0.55 to 0.9 wt%.

MIs in olivines from June 2014 can best be compared with those of other recent small-volume and short-lived eruptions which emitted basalts with low phenocryst contents, like those in March 2007 (0.6 Mm^3) and November 2009 (0.1 Mm^3) (Roult et al., 2012). March 2007 aphyric basalt has a bulk homogeneous composition with intermediate MgO content (MgOwr: 7.33 wt%; K_2O: 0.67 wt%). Their olivines (Fo 81) are in equilibrium with the bulk rock and their composition is unimodal (Di Muro et al., 2014). November 2009 products are the most magnesian lavas emitted in the 2008-2014 period, slightly zoned (MgOwr: 7.6-8.3 wt%; K_2O: 0.75 – 0.62 wt%) and contain a few percent of normally zoned olivine macrocrysts with bimodal composition (Fo81 and Fo83.5, see Di Muro et al., 2016). June 2014 bulk rocks (MgOwr: 6.7 wt%; K_2O: 0.75 wt%) and melt inclusions in Fo_{80-81} olivines are quite evolved. Their composition is close to that of products emitted by summit intracaldera eruptions in 2008, ca. 1.5 years after the large 2007 caldera forming eruption (Di Muro et al., 2015) (Fig. 8). As already reported for 2008 products, many olivine macrocrysts of 2014 are clearly too magnesian to be in equilibrium with the relatively evolved host melts. Overall, MgO content in 2007-2014 melt inclusions tends to decrease with decreasing Fo content of the host olivines. MIs in olivines also exhibit a trend of linear decrease in MgO and increase in FeO from April 2007 to 2009-2014 products (Fig. 9). Melt inclusions in March 2007, November 2009 and June 2014 follow the same trend of FeO enrichment (Fig. 9). In the large-volume and olivine-rich April 2007 products, MIs in magnesian olivines with $Fo_{>82}$ have distinctly higher MgO, FeO and lower SiO_2 and Al_2O_3 than MIs in 2009-2014 products. The distinctive FeO enrichment of many of the MIs from the April 2007 oceanite has been interpreted by Di Muro et al. (2014) as a result of post-entrapment modification during long lasting magma storage.

Two populations of low- and high-Ca melt inclusions are also found in the November 2009 olivines. Low-Ca melt inclusions from the November 2009 and June 2014 eruptions indicate a single trend of chemical evolution (Fig. 8), consistent with bulk rock compositions. June 2014 products have lower MgO and CaO contents than those from November 2009. Significant scattering in K_2O content (0.6-0.9 wt%) is found in low-Ca inclusions from 2009, as observed in high-Ca inclusions from the 2014 eruption, but they share similar K_2O contents. In 2009 and 2014 products, K_2O content of melt inclusions is partly anti-correlated with the olivine Fo content. This observation has been attributed to moderate heterogeneity of primary melts feeding the plumbing system of PdF. Rapid temporal changes of K_2O content in PdF basalts have been reported (Boivin and Bachelery, 2009).

4.5 Mineral composition and glass – plagioclase equilibrium

All 2014 scoriae (spiny, fluidal, golden) contain the same paragenesis of olivine, clinopyroxene and plagioclase. The composition of minerals found in golden, fluidal and spiny scoriae is indistinguishable.

In olivines, average MgO content decreases from macrocrysts ($Fo_{84.1}$) to mesocrysts ($Fo_{79.6}$) to microlites. Olivine microlites (Table S5) are normally zoned. Their composition ranges from $Fo_{78.0-75.3}$ in the cores to $Fo_{74.3-70..5}$ in the rims. Overall, olivines in 2014 products span the full range of typical Fo contents of recent Piton de la Fournaise magmas (Boivin and Bachèlery, 2009; Di Muro et al., 2014; 2015). Clinopyroxene composition (augites) ranges from $En_{53}Fs_{15}Wo_{32}$ to $En_{41}Fs_{14}Wo_{45}$. Their average composition ($En_{45}Fs_{14}Wo_{41}$) is consistent with that found in other recent evolved melts like those emitted by the 2008 eruptions (Di Muro et al., 2015) and more generally in recent Piton de la Fournaise products (Boivin and Bachèlery, 2009). Clinopyroxenes are unzoned, the composition of cores and rims is very similar and close to that found in microcrysts and mesocrysts. Plagioclase composition ranges from $An_{79.5}Ab_{19.9}Or_{0.6}$ to $An_{63.1}Ab_{35.7}Or_{1.2}$ with a bimodal distribution ($An_{76.5-79.5}$ and $An_{63.1-72.9}$). Similar bimodal distribution was observed in the evolved 2008 products, in particular from the November 2008 eruption (Di Muro et al., 2015). Mesocrysts ($An_{75.5}Ab_{23.8}Or_{0.7}$ on average) are more calcic with respect to microcrysts ($An_{65.7}Ab_{33.1}Or_{1.2}$ on average). Normal zoning is commonly found from plagioclase cores to rims, which contrasts with the complex zoning patterns previously reported for the 2008 products (Di Muro et al., 2015).

Plagioclase-melt equilibrium and melt composition in pyroclastic rocks and water-quenched lavas were used to estimate both temperature and water content dissolved within the melt (Table S5). Temperature estimates are based on the equation of Helz and Thornber (1987) modified by Putirka (2008). Water content was calculated from the plagioclase hygrometer of Lange et al. (2009) at 50 MPa corresponding to the maximum CO_2-H_2O pressure estimation (recalculated with Papale et al., 2006), typically recorded in melt inclusions from central products at PdF (e.g. 1931 eruption; (Di Muro et al., 2016) and references therein). This pressure roughly corresponds to the sea level depth, which is inferred to be the location of the potential main shallow magmatic reservoir (Peltier et al., 2009;

Lengliné et al., 2016; Coppola et al., 2017). The application of the hydrous thermometer of Putirka (2008) makes it possible to estimate the dissolved water content in the melt with a nominal uncertainty of 0.15 wt% and is only slightly dependent on pressure. Plagioclase compositions not in equilibrium with the melt (glass or bulk rock) are those mesocryst cores with the highest ($An_{>76.5}$) anorthite content (Table S5). Such compositions are more in equilibrium with CaO-richer magnesian melts than those measured in matrix glasses and bulk rocks of 2014 eruption.

In order to determine pre-eruptive conditions, calculations were performed only on paired plagioclase rims and matrix glasses in equilibrium, using the plagioclase-melt equilibrium constant of Putirka (2008) calibrated for melts whose temperature exceeds 1050°C ($Kd_{An\text{-}Ab}$ = 0.27±0.05). Our review of published and unpublished data shows that melt temperature progressively decreases from April 2007 (1188+/-16 °C) to January-October 2010 (1147+/-9°C) and positively correlates with K_2O content in melts which increases from 0.70 to 0.96 wt% (Fig. 10). The melts from the June 2014 eruption record the lowest temperatures in post-2007 eruptions (1131±15 °C) together with the highest K_2O-enrichment (K_2O: 0.90±0.12 wt%). The lowest temperatures are recorded by spiny scoriae, while the temperature of golden scoriae overlaps with that of 2010 products emitted before the 2010-2014 phase of quiescence. In spite of the large variability in melt composition and temperature, average pre-eruptive water content dissolved in the melts (0.5 +/- 0.2 wt%) is quite homogeneous for the whole 2008-2014 period. In 2014, the lowest estimated dissolved water content (down to 0.38 wt%) is for the golden and some fluidal scoriae, while the maximum amount (0.68 wt%) is for the spiny opaque scoriae. However, water content estimated from core-bulk rock equilibrium (0.3±0.1 wt%) is slightly lower than that estimated from rim and microlite-matrix glass equilibrium (0.5±0.2 wt%), but the difference broadly overlaps the nominal uncertainty related to calculations. Dissolved water contents in melts of the pyroclasts are thus lower than those measured in 2007 melt inclusions (H_2O: 0.8 +/- 0.15 wt% and up to 1.1 wt%) and higher than typically found in degassing matrices of lava and Pele's hairs of 2007 (Fig. 10; 0.2 wt%; see Di Muro et al., 2015; 2016).

5 Discussions

5.1 The activity

The activity fed by the uppermost Western Fractures (Fig. 1) was very short-lived, as shown by the presence of only scattered bombs and coarse lapilli (Fig.s 3c and 3d). The homogeneity of these clasts, their coarse grained nature and the fluidal smooth texture are in agreement with very short-lived fire-fountaining/magma jets. Glassy outer surfaces of clasts have been interpreted as a late-stage product of fusion by hot gases streaming past the ejecta within the jet/fountain (*Thordarson et al.,* 1996; *Stovall et al.,* 2011). However, the occurrence of this process is not supported by the homogeneous glass composition in our fluidal clasts.

At lower altitude and close to the Main Vent (Fig. 1), the 5 cm layer at the base of the fall deposit is fine-grained (Fig. 3b), rich in fluidal and golden fragments, with a perfect Gaussian grain size curve (Fig. 5), and similar to that reported from the weak 2010 fountaining event (Fig. 3e and Hibert et al., 2015). Therefore we interpret this deposit as being due to weak Hawaiian like fountaining (sustained, but short-lived) activity. The top of the same deposit is coarse grained, bimodal, has a lower content in coarse ash (Table S2) and is rich in spiny opaque and spiny glass fragments. The reverse grain size likely records the transition from early continuous fountaining to late discrete Strombolian activity, typical of many eruptions at PdF (Figs. 2a and 2b and Hibert et al., 2015). The reverse grading of the whole deposit (Fig. 3b) is thus not correlated with an increase in energy of the event, but with two different eruptive dynamics and fragmentation processes. The decrease in coarse ash, which correlates with the decrease in energy of the event, highlights the most efficient fragmentation process within the Hawaiian fountaining with respect to the slow gas ascent and explosion of the Strombolian activity. These conclusions are consistent with (i) the continuous and progressive decrease in intensity of Real time Seismic Amplitude Measurement recorded by the OVPF seismic network (unpublished data), and (ii) satellite derived TADR (Coppola et al., 2017) which suggest continuous decay of magma output rate after an initial short-lived intense phase.

5.2 Interpretation of the different textural signatures and the meaning of the 4 typologies of clasts.

The first microtextural analysis of Hawaiian ejecta was performed by Cashman and Mangan (1994) and Mangan and Cashman (1996) on pyroclasts from 1984 to 1986 Pu‘u ‘Ō‘ō fountainings. The authors defined two clast types: 1) ‘scoria’ consisting of closed-cell foam of ≤85% vesicularity, with round, undeformed, broadly-sized vesicles, and 2) ‘reticulite’, an open-cell polyhedral foam with ~1 μm thick vesicle walls with >95% vesicularity. They stated that the scoria to reticulite transition is a consequence of Ostwald ripening, where larger bubbles grow at the expense of smaller bubbles due to post-fragmentation expansion of clasts

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within the fountain. According to this model, scoria preserves textures closer to conditions at fragmentation, whereas continued vesiculation and clast expansion in the thermally-insulated core of the fountain results in reticulate. This model was confirmed at lava fountains at Etna (Polacci et al., 2006), Villarrica (Gurioli et al., 2008), Kīlauea Iki, (Stovall et al., 2011 and 2012), Mauna Ulu (Parcheta et al., 2013) and Al Madinah (Kawabata et al., 2015). These last authors also measured the connected and isolated porosity in the AD1256 Al-Madinah Hawaiian fountaining eruptions. They found that the reticulite-like textures from the central part of these very high fountains showed isolated vesicles in agreement with low shear rates and low viscosity melts, where bubbles may grow spherically and remain isolated. In contrast, at margins of the fountains, high shear may lead to stretching and mechanical coalescence of bubbles, forming the common, fluidal types of particles seen also in the deposits. They also stated that lower vesicularity and greater isolated porosity were found in some tephra interpreted as resulting from Strombolian eruptive phases.

The data that we found in our study of the typical activity of PdF agree only partially with all these interpretations. The reason is that we sampled and measured products of very weak Hawaiian to Strombolian activities. If we plot the approximate durations and masses of these events on the Houghton et al. (2016) diagram, the 2014 activity of PdF falls into the two fields for transient and fountaining activity, but at the base of the diagram. We here show for the first time that short lived and weak fountaining can preserve pyroclast textures that can be considered as representing a valid approximation to magma ascent and fragmentation conditions before the explosions. The occurrence of time-variable ascent conditions is also reflected in the time evolution of explosion dynamics, with the golden and fluidal emitted from the low fountaining episodes and the spiny fragments from the Strombolian-like explosions. This is quite evident when, following Stovall et al. (2011), we use a plot of vesicle-to-melt ratio (Vg/Vl, after Gardner et al., 1996) and vesicle number density (Nv). As explained by Stovall et al. (2011), addition of small bubbles leads to an increase in Nv and only a slight increase in Vg/Vl. Bubble growth by some combination of diffusion and decompression leads to an increase in Vg/Vl at constant Nv. Nv decreases while Vg/Vl increases during bubble coalescence, whereas loss of bubbles via collapse or buoyant rise leads to a reduction in both parameters. Intermediate trends on the diagram reflect combinations of more than one of these processes. The pumice and the scoria from the Main Vent of PdF show the highest Vg/Vl, but also the highest Nv, suggesting preservation of small vesicles and growth by some combination of diffusion and decompression. The presence of the small vesicles confirms that the weak PdF activity leads to only limited post-fragmentation expansion inside the hot portions of the short-lived fountains. These data contrast with the data from the more energetic fountaining events observed at Kilauea or elsewhere, where pre-eruptive information is basically erased because pumice textures are dominated by expansion effects due to their longer permanence within the long-lived fountaining. In contrast, the densest, spiny scoriae and the scoria from the Fractures activity show the lowest values of Nv and Vg/Vl, due to incipient coalescence and/or loose/lack of small bubbles.

The golden and fluidal fragments, from the Main Vent, show the highest vesicularity, variable proportions of isolated vesicles and high, but variable, Nv numbers (Figs. 4, 6d and 11). They are also characterized by a uniform vesicle population with clear evidence of incipient expansion (Fig. 4). From the connectivity graph, there is a clear decrease in isolated vesicles with the increase in vesicularity which we interpret as an increase in expansion of the clasts and a reduction in volume of the isolated vesicles. We interpret the fluidal and golden fragments, at the Main Vent, to be the faster and less degassed magma, which experienced only a very short residence time in the magma transport system (dyke+vent). According to previous work, these fragments should be derived from the central part of the fountains, but they do not show the strong post expansion signatures reported in the literature (Fig. 11). In fact they might represent the gas-rich magma with foam accumulation. In contrast, the spiny fragments from the Main Vent and the fluidal fragments from the Fractures show low Nv and low Vg/Vl. However, the presence of large numbers of isolated vesicles within the fluidal scoria from the Fractures agrees with their provenance from a fast hot ejection of relatively degassed magma (low Nv). The higher percentage of microlites, the lack of isolated vesicles, the coalescence signature and the low Nv values of the spiny fragments (Fig. 4) are indicative of a extensively degassed and cooled magma. These clasts likely record the slowest ascent velocity and the longest residence time in the reservoir+dyke+vent system compared to the golden/fluidal counterpart (as suggested by similar clasts found also at Kilauea Iki, (Stovall et al., 2011).

Our findings are in full agreement with the recent review of Colombier et al. (2017b). According to these authors, connectivity values can be used as a useful tool to discriminate between the basaltic scoria from Hawaiian (fire fountaining) and Strombolian activity. The broad range in connectivity for pumice and scoria from fire fountaining is interpreted simply as being due to variations in the time available before quenching due to differences in location and residence time inside the fountain. The fluidal fragments from the Western Fractures are the richest in isolated vesicles because they are transported by very short lived hot lava jets. In contrast, the higher connectivity observed in scoria from Strombolian activity is probably related to their higher average crystallinity, and more extensive degassing prior to the eruption, (Colombier et al., 2017b). The spiny surface of these Strombolian fragments is due to the fact that these weak explosions emit only a small solid mass fraction and the partially quenched dense clasts land quickly after a short cooling path through the surrounding atmosphere (e.g. Bombrun et al., 2015).

Among spiny fragments, the opaque ones are the densest, they lack a uniform glassy surface, and they are characterized by i) very high microlite content, ii) strong coalescence signature (Fig. 4), iii) heterogeneous glass chemistry, and iv) mingling with hotter magma at the edges (Fig. 8a). All these features reveal the composite nature of these clasts. We interpret the spiny opaque as spiny glass fragments recycled inside the eruptive vent during the explosions, being the densest portion of the magma prone to fall back in the vent/fracture (Fig. 2b).

Longitudinal variation in eruptive style along the fracture system produces a spatial variability in the proportions of the four typologies of clasts. The uppermost fractures are characterized solely by fluidal fragments; they lack both the spiny and the golden components. In addition, these fluidal clasts are the ones showing the smoothest surfaces (indicative of rapid quenching in a very hot environment), the lowest vesicularity, the highest content in isolated vesicles, but also by low vesicle numbers, comparable to the spiny fragments. These features are related to the emissions of only small amounts of partially degassed magma by short lived jets, which are clearly different from the sustained activity occurring at the Main Vent. The scoriae from the Fractures underwent rapid quenching in a very hot environment (jet of magma) that prevented any expansion or further vesiculation and preserved a very high number of isolated vesicles (Fig. 6d).

The four typologies of clasts were also found at the Main Vent, but with vertical variations in abundance from the base to the top, reflecting the variation in intensity and dynamics of the activity with time. It is interesting to note that the fluidal fragments at the Main Vent are less smooth (Fig. 4), more vesiculated, and have a lower content of isolated vesicles than the fluidal scoria from the uppermost Fractures (Fig. 6). Therefore fluidal fragments at the Main Vent could represent clasts that have been partly modified during their residence in the external part of the fountains, while the golden samples could come from the central part (Stovall et al., 2011 and 2012). However, the slight differences in crystallinity and glass chemistry between the fluidal and golden fragments support the idea that each of these fragments has an imprint from the pre-fragmentation setting. The spiny fragments are the most degassed, densest and the most crystal rich magma that was emitted during low-energy activity by Strombolian explosion, where recycling phenomena were also very frequent (Fig. 2f).

5.3 Integration between the physical and textural characteristics of the products and their geochemical signature: insight into the feeding system

According to Peltier et al. (2016), the June 2014 eruption emitted magma from a shallow pressurized source located only 1.4-1.7 km below the volcano summit. Coppola et al. (2017) suggest that the 2014 event was fed by a single shallow and small volume magma pocket stored in the uppermost part of the PdF central plumbing system. All 2014 clasts show homogeneous and evolved bulk compositions, irrespective of their textural features. June 2014 products are among the most evolved products erupted since at least 1998 and are moderately evolved with respect to those emitted in 2010, just before the 2010-2014 quiescence. The different type of scoria and pumice show significant variations in glass composition (Fig. 8b) due to variable degrees of micro-crystallization. In theory, microlites and microphenocrysts can reflect late stage (during magma ascent and post-fragmentation) crystallization. In this case, their variable amount within, for instance, the glassy and opaque parts of the spiny scoria might reflect slower ascent velocity or longer residence time in the system (e.g. Hammer et al., 1999, Stovall et al., 2012; Gurioli et al., 2014) in agreement also with the vesicle signature. However, the four typologies of clasts differ in terms of mesocryst content (from rare to 5 vol% for the golden and fluidal and 14-23 vol% for the glassy spiny and spiny opaque, respectively). Equilibrium plagioclase-melt pairs record an almost constant and moderate dissolved water content, intermediate between that expected for melts sitting in the main shallow reservoir (located close to sea level) and the degassed matrix of lavas. Dissolved water contents are thus consistent with pre-eruptive magma water degassing at shallow level, as suggested by geophysical data, and suggest that the plagioclase mesocrysts and some of the microlites in the spiny scoria grew during magma storage. Melt composition records a potential pre-eruptive thermal gradient of ~30 °C between the hotter (pumice and fluidal) and the cooler (spiny) magma.

Tait et al. (1989) suggest that magma evolution can lead to oversaturation of volatile species within a shallow reservoir and trigger a volcanic eruption. At PdF, the golden and the fluidal clasts might represent the portion of magma sitting in the shallow reservoir and accumulating bubbles of water rich fluids, released by the cooler, more crystallized and more degassed "spiny" magma. The small volume of magma, its constant bulk composition and the very small inflation recorded prior to the eruption could be consistent with an internal source of pressure related to volatile exsolution. Larger inflation rates over a broader area are expected when reservoir pressurization is related to a new magma input from a deeper source. Textural heterogeneity of the 2014 products partly reflects a pre-eruptive thermal gradient recorded by the variability in crystal and bubble contents in the shallow reservoir feeding this eruption. Our results are consistent with processes of mechanical reservoirs/dyke stratification, as already observed experimentally by Menand and Phillips (2007). The golden and fluidal fragments are the bubble richer and hotter portion of the melt. The spiny fragments are the degassed and cooler portion of the reservoir, whose progressive tapping led to a decrease in explosive intensity (from fountaining to Strombolian activity). The spiny opaque clasts can be considered as being recycled material that fell back into the system. Accumulation of olivine crystals out of equilibrium with the host magma produces minor variations in phenocryst contents as observed within the same type of clasts sampled at different times/locations during the eruption, with the scoria from the Western Fracture and early erupted lava being the ones with the lowest amount of olivine (Table S4 and Fig. 7b). Again, this temporal variation supports a slight increase in heavy crystals within the most degassed magma emitted toward the end of activity, as observed as a general trend at PdF (Peltier et al., 2009).

Melt inclusion results allow us to confirm the involvement of a single and only slightly heterogeneous magma source in 2014, possibly related to cooling and fractional crystallisation of an older magma batch (November 2009). Interestingly, this latter short lived summit eruption was also characterized by the same large range of pyroclastic products in spite of the less evolved magmatic composition. The main difference with respect to 2014 is that the 2009 products contain a slightly larger amount of mm-sized olivine macrocrysts in the lava, scoria and pumice. This suggests that bubble accumulation and source pressurisation is mostly controlled by the shallow storage depth, which allows water exsolution (Di Muro et al., 2016), rather than by a trend of magma cooling and evolution (Tait et al., 1989).

6. Conclusions In this paper we show that a combination of textural and petro-chemical quantification of the eruptive products can be used to characterize the on-going activity at PdF and provide valuable information to understand both the causes and dynamics of these very short-lived and small-volume eruptions.

The June 2014 summit eruption occurred after a relatively long phase of quiescence and was preceded by only weak and short geophysical precursors. This multidisciplinary approach provides new constraints on the mechanisms triggering such short-lived, small eruptions. First, we found that this kind of eruption can be triggered solely by bubble accumulation and source pressurisation at a very shallow storage depth. We suggest that it is the shallow depth of the reservoir itself that allows exsolution, rather than magma cooling and evolution or recharge from a deep source. Second, these small, summit eruptions are usually related to small pockets of magma left behind following previous eruptions. Third, the thermal-mechanical stratification at the reservoir level between the bubble rich portion and the more degassed and cooler one modulates the style of the explosions. Therefore, in terms of ascent and degassing history of the magma the golden and fluidal fragments represent the bubble richer and hotter portion of the melt with faster ascent rate, while the spiny fragments are the degassed, cooler portion of the reservoir, whose progressive tapping lead to a decrease in explosive intensity (from fountaining to Strombolian activity). Finally, an accumulation of olivine crystals out of equilibrium with the host magma produces minor variations in phenocryst contents with a slight increase in heavy crystals within the most degassed magma emitted toward the end of activity, as observed as a general trend at PdF (Peltier et al., 2009).

To conclude, these results highlight the importance of petrological monitoring, which can provide complementary information regarding the ongoing volcanic activity to other geophysical and geochemical monitoring tools commonly used on volcanoes.

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Figure captions

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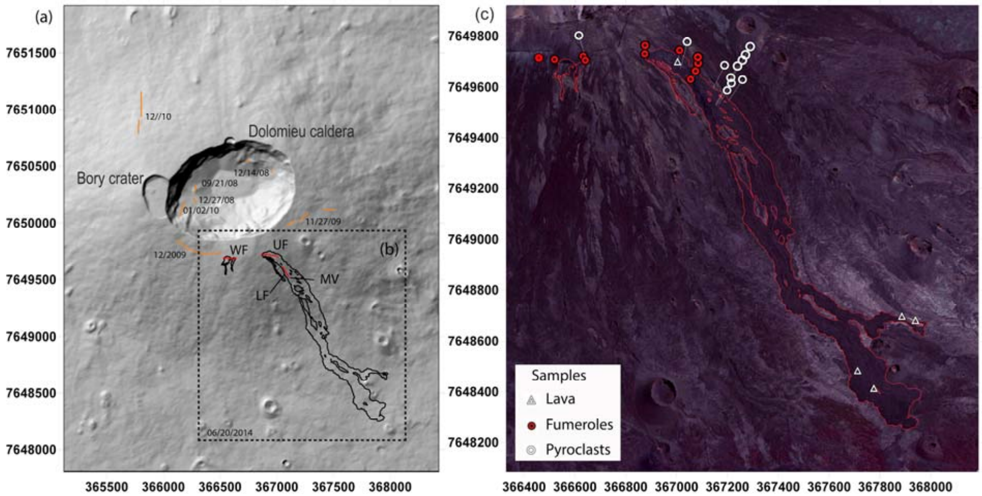

Figure 1 a) Digital elevation model of the summit crater area at Piton de La Fournaise, la Réunion, France; orange = fractures generated by pre-2014 eruptions; b) red = fractures active during the 2014 eruption: WF (western fracture), UF (upper fracture), LF (lower fracture), MV (Main Vent). Black= outline of the 2014 lava field; c) location of sample collection points

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Figure 2 Photo collection from the web site. From a to f: evolution of activity from early morning to evening, June 21. All the photos are from the Main Vent (see Fig. 1). The images in a, b and the inset in b are from Laurent Perrier; c) http://www.rtl.fr/actu/sciences-environnement/la-reunion-eruption-du-piton-de-la-fournaise-apres-4-ans-de-sommeil-7772778861; d) http://www.ipreunion.com/volcan/reportage/2014/06/21/eruption-du-piton-de-la-fournaise-actualise-a-17h-la-lave-coule-sur-1-5-kilometre,26023.html; e) http://www.zinfos974.com/L-eruption-du-Piton-de-la-Fournaise-Le-point-de-17h_a72981.html; f) http://nancyroc.com/eruption-a-la-reunion

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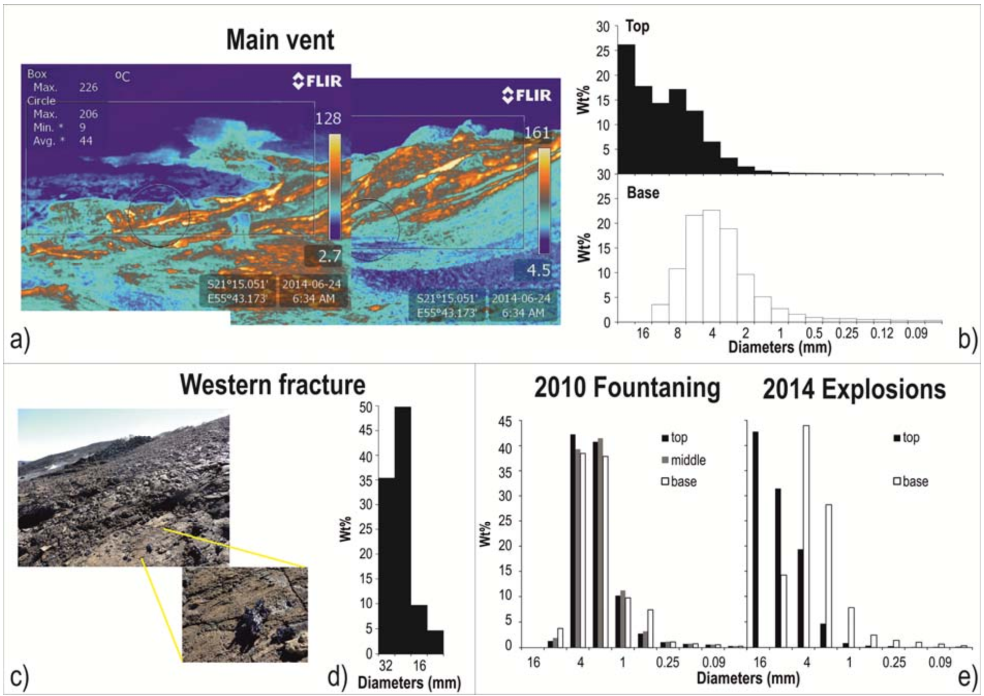

Figure 3 a) Thermal photo of the scoria fall out area in proximity to the Main Vent (see Fig. 1 for the location); b) grain size histograms for the base and top deposit of the Main Vent; c) scattered scoria from the Western Fracture (see Fig. 1 for the location); d) grain size histogram of the scoria deposit at the Western Fracture; e) comparison between the grain size histograms for the 2010 Hawaiian fountaining and the 2014 Main vent activity

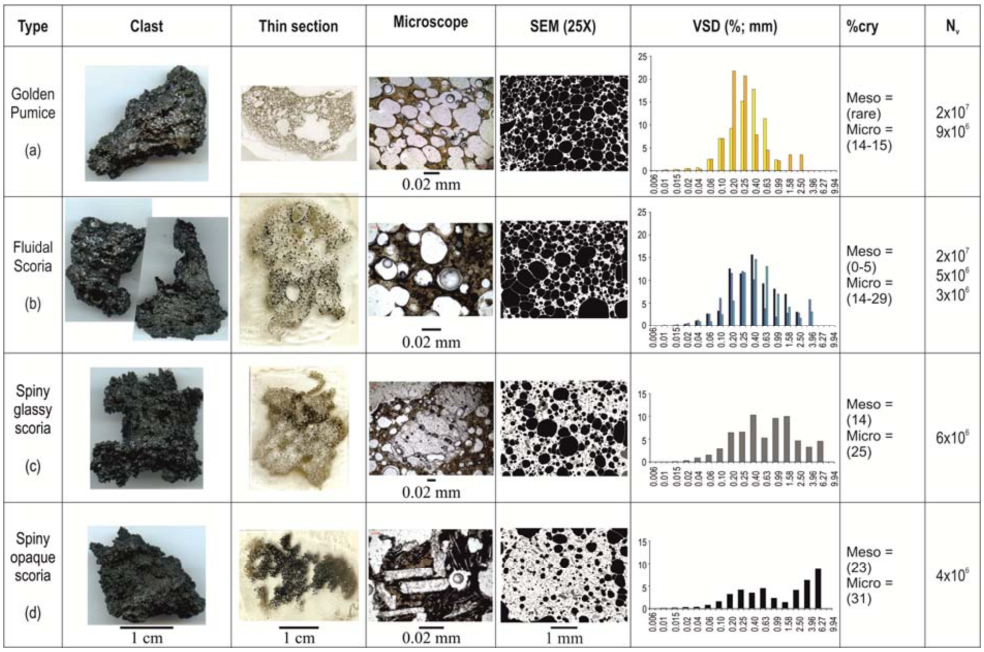

Figure 4 Textural features of the 2014 pyroclasts. Clast = photo of the different types of juvenile pyroclasts. Thin section = thin section imaged with a desktop scanner. Microscope = photo taken with an optical microscope using natural light; SEM (25X) = photo captured using a scanning electron microscopy (SEM), in BSE mode at 25x magnification: black are vesicles, white is glass, grey are crystals. VSD = vesicle volume distribution histograms, where the diameter, in mm, is plotted versus the volume percentage. %Cry = is the total percentage of crystals corrected for the vesicularity. Nv = number density

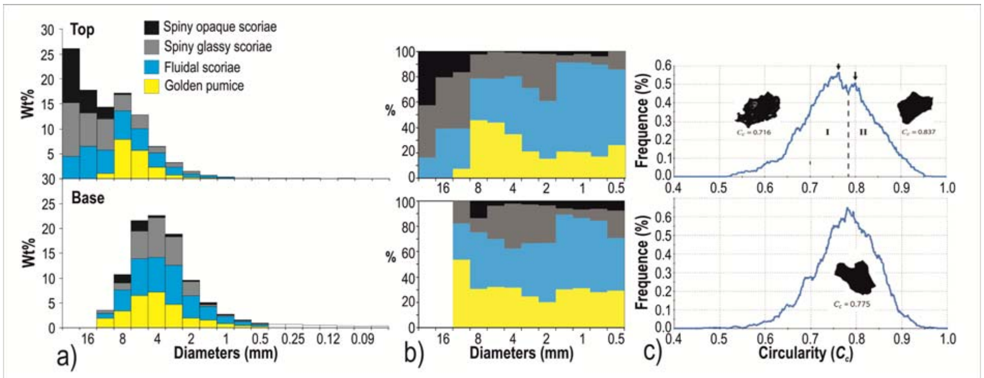

Figure 5 Proportion of each type of clast measured from the Main Cone for the 2014 eruption: (a) componentry within the different grain size classes from the base to the top of a 10 cm thick scoria deposit; b) normalized componentry composition from the base to the top of the deposits; (c) Morphologi G3 results for the coarse ash fragments (350 micron), where the population is formed exclusively of smooth fragments that correspond to fluidal and golden pumice.

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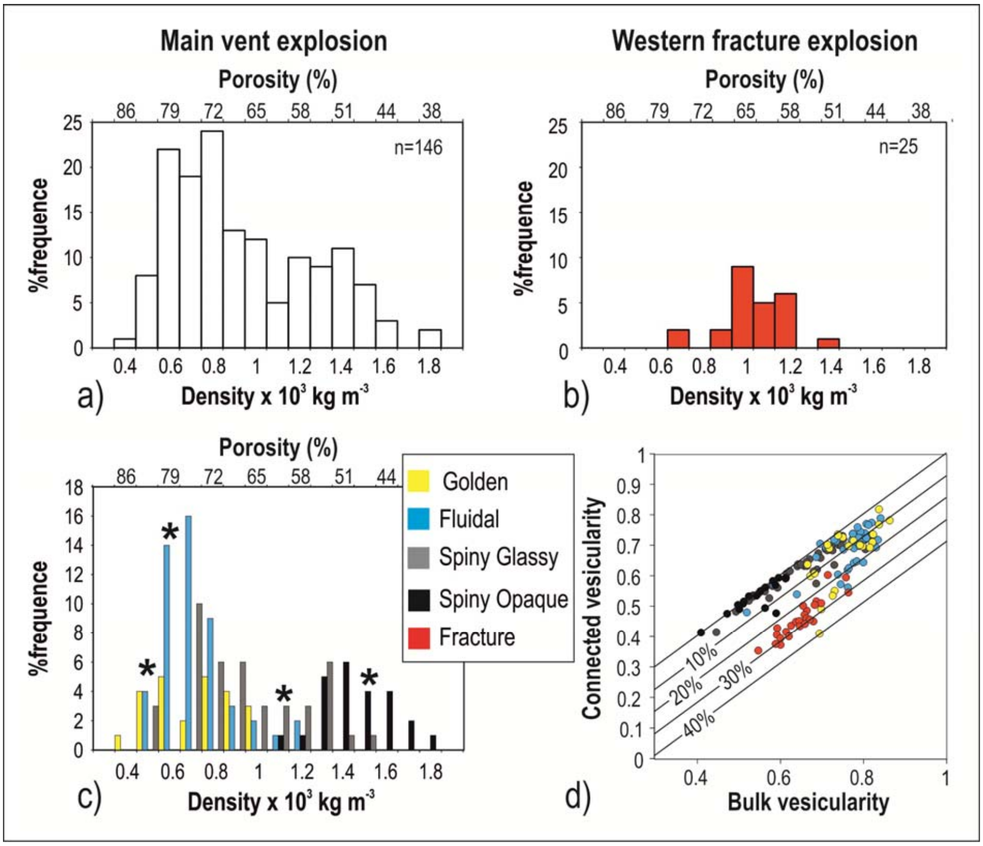

Figure 6 Density and connectivity data of the 2014 pyroclast fragments: a) density distribution histogram for all the pyroclast fragments measured for the 2014 activity from the Main Vent; b) for the Western fracture; and c) for different typologies of clasts from the Main Vent; d) graph of connected vesicularity versus total vesicularity. The diagonal line represents equality between the connectivity and vesicularity, beneath this line the samples have isolated vesicles

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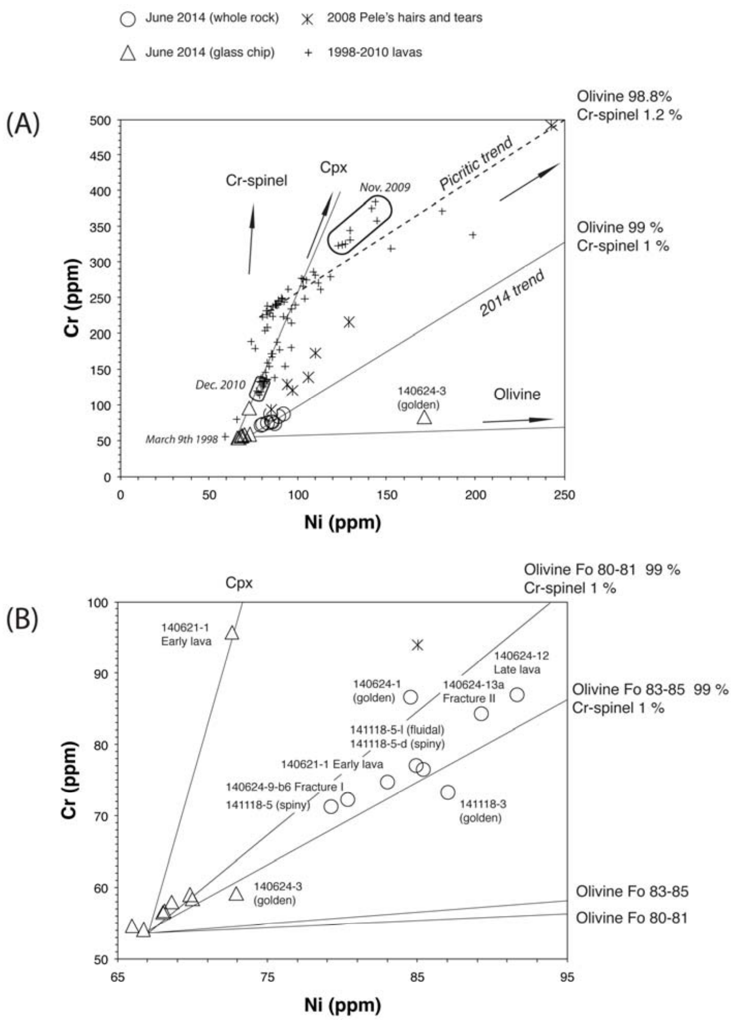

Figure 7 Ni-Cr concentration plot. (a) Ni-Cr signature of the June 2014 lavas compared to that of recent eruptions (Di Muro et al. (2015) and unpublished data). Whole-rock (circles) and glass (triangles) compositions are shown for the June 2014 eruption. Olivine controlled lines are indicated for olivine hosting 1.2 and 0.6 wt.% Cr-spinel. Compositions used for olivine (Ni=1900 ppm, Cr=300ppm), clinopyroxene (Ni=970 ppm, Cr=4800 ppm), and Cr spinel (Ni=1500 ppm, Cr=25%) are inferred from Salaün et al. (2010), Di Muro et al. (2015) and Welsch et al. (2009). (b) Zoom of the Ni-Cr relationship between glass (triangles) and whole-rock (circles) samples from the June 2014 eruption. Fracture I = Western fracture, Fracture II = Upper fracture

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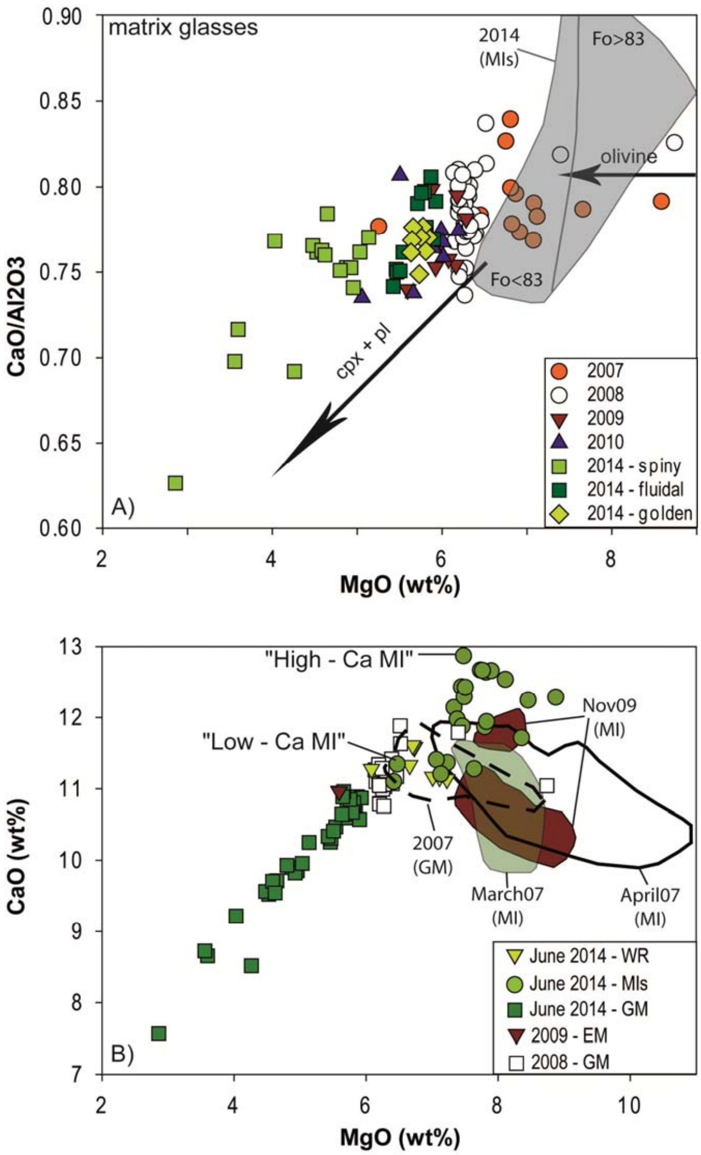

Fig. 8 (a) Evolution of CaO/Al_2O_3 ratio in the matrix glasses of recent eruptions at Piton de la Fournaise as a function of MgO content (directly proportional to melt temperature). MI = Melt inclusions (grey area for the 2014 samples). (b) CaO versus MgO content for Piton de la Fournaise products. WR = whole rock, GM = ground mass; MI = melt inclusion, EM = embayment glass

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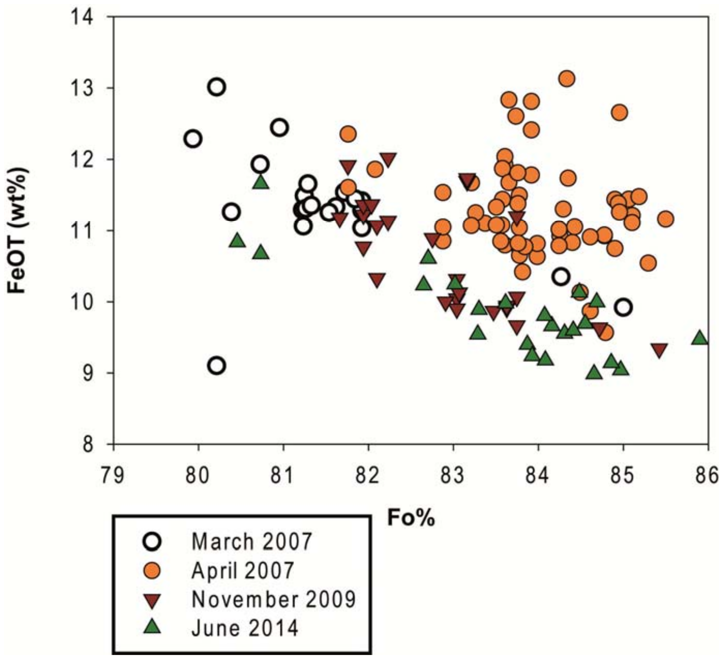

Figure 9 FeOT in melt inclusions as function of Fo content of the olivine host for recent eruptions at Piton de la Fournaise

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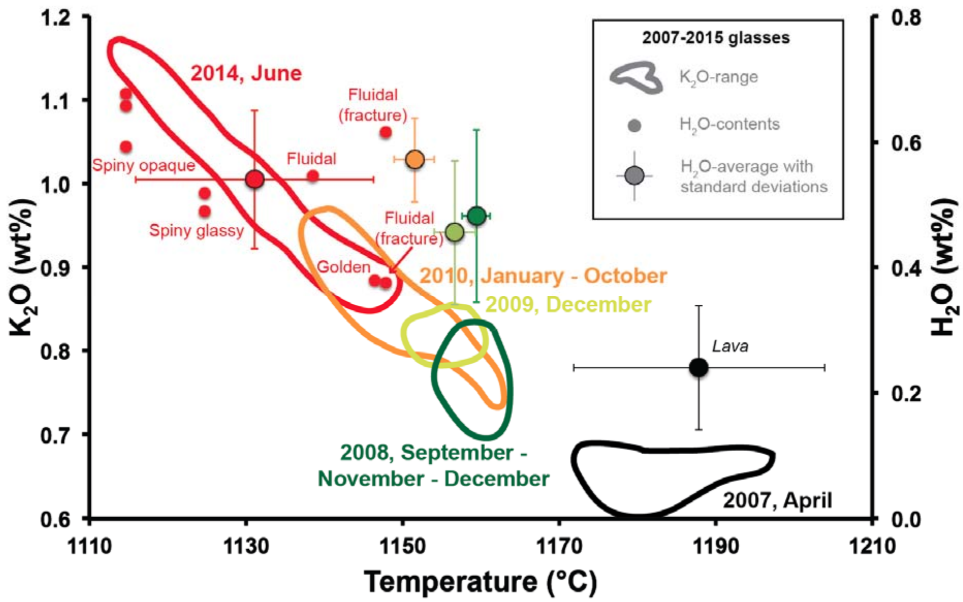

Figure 10 Temperature, composition (K2O) and dissolved water content (H2O) for the evolution of 2007-2014 melts from glasses. Temperature estimation based on the MgO-thermometer of Helz and Thornber (1987) modified by Putirka (2008). Water content from the plagioclase hygrometer of Lange et al. (2009). Only plagioclases in equilibrium with melts are considered, following the procedure described by Putirka (2008) for >1050°C melts (Kd = 0.27±0.05).

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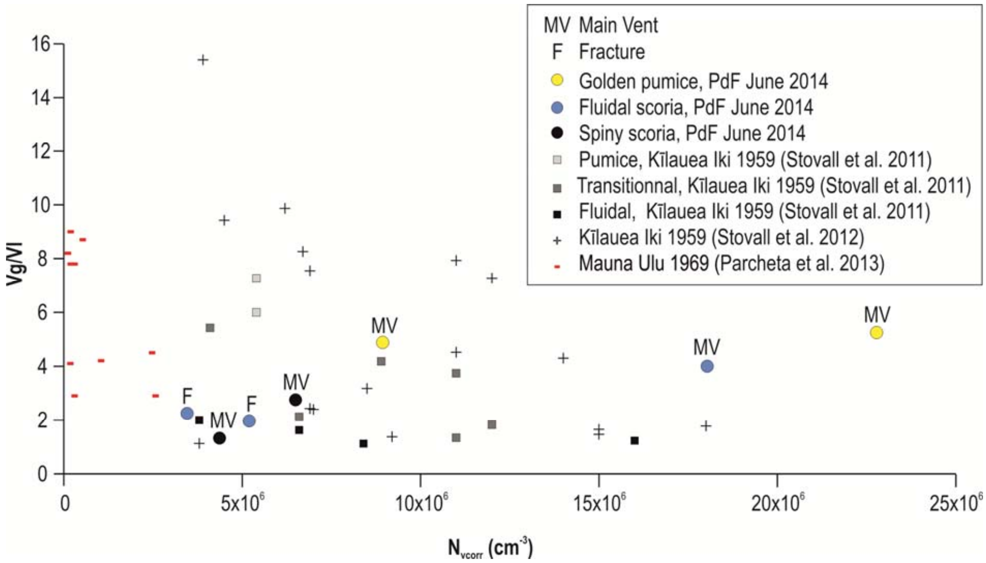

Figure 11 Volumetric ratio of vesicles to melt (VG/VL) versus vesicle number density