The effects of rock heterogeneities on dyke paths and asymmetric ground deformation: The example of Piton de la Fournaise (Réunion Island)

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Research paper

Abstract

During pre-eruptive periods and eruptions, Piton de la Fournaise volcano shows an asymmetric magma-induced pattern of deformation. The origin of this asymmetry is not well constrained and the pre-eruptive deformation does not have a satisfactory explanation. Here we present data on the past history and complex structure of the volcano. We also provide new field data on exposed dykes in the volcano. The field data confirm that the eruptive centre has migrated through time. Our field data indicate average dyke dips of about 80°. By contrast, inferred dips of recent feeder dykes using geodetic data are 53°–75°. We explain this difference as being partly due to the models used to infer the dyke dips from geodetic data, which assume the host rock to be homogeneous and isotropic. When fractures and the other heterogeneities are taken into account in the models, the inferred dyke dips become similar to those of the exposed dykes. New numerical models indicate that when the complex internal structure of the volcano is taken into account, part of the asymmetry of the ground deformation can be explained. Part of the deformation asymmetry, however, is due to the curved structure of the rift system, resulting in the injection of curved dykes, and the dyke dips. © 2008 Elsevier B.V. All rights reserved.

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1. Introduction

The ground surface deformation of active volcanoes has been observed for a long time (e.g. Murray et al., 1977). The deformation occurs mostly within the short timescale of a magmatic event (eruptive and/or intrusive). More specifically, the deformation reflects stress changes at the surface induced by magmatic processes occurring within the volcano, such as the growth of the magma chamber under internal pressure (e.g. Lundgren et al., 2003; de Zeeuw van Dalfsen et al., 2004; Bonaccorso et al., 2005) and the injection of dykes (e.g. Pollard et al., 1983; Sigmundsson et al., 1999; Fukushima et al., 2005). The deformation intensity and the area affected depend partly on the depth and the shape of the source of pressure (Pollard et al., 1983) and partly on the mechanical properties of the rocks (Gudmundsson, 2003, 2006). Another type of deformation can also be observed on active volcanoes, and is interpreted as a more continuous behaviour of the edifice. It is the large-scale deformation induced by volcanic spreading (Borgia, 1994; Delaney and Denlinger, 1999; Owen et al., 2000; Bonforte and Puglisi, 2003; Lundgren et al., 2003), or by the regional stress field (de Zeeuw van Dalfsen et al., 2004). As the magma path is determined by the state of stress in the volcano (Gudmundsson, 2006), the regional stress field or volcanic spreading may influence the magma path so as to generate orientated rift zones (Walter et al., 2006).

Usually the regional stress field, volcanic spreading, and distribution of injections responsible for volcanic deformations are mechanically linked. Stresses induced by tectonic or volcano-tectonic events may favour magmatic injections (Neri et al., 2005). Conversely, magma-induced stresses can reactivate...
volcano-tectonic structures (Ando, 1979; Walter and Amelung, 2006). The result of these interactions is often a complex pattern of deformation (e.g. Lundgren et al., 2003). For example, the pattern of deformation of Etna is interpreted as an eastward motion of the east flank, combined with magmatic deformations (Bonforte and Puglisi, 2003; Lundgren et al., 2003). At Kilauea, the asymmetric pattern of deformation is related to the spreading of the south-eastern flank (Delaney and Denlinger, 1999; Owen et al., 2000).

In other volcanoes, such as Stromboli and Piton de la Fournaise, the asymmetric pattern of deformation observed is supposed to be entirely due to magmatic effects (Zlotnicki et al., 1990; Cayol and Cornet, 1998; Sig mundsson et al., 1999; Battaglia and Bachély, 2003; Froger et al., 2004; Mattia et al., 2004; Fukushima et al., 2005; Peltier et al., 2007). However, for Piton de la Fournaise, some authors suggest that the instability of the eastern flank can lead to overestimate the dip of the modelled dykes (Fukushima, 2005; Peltier et al., 2007). Mainly, these authors obtain their results by using a homogeneous, isotropic medium, in which they simulate the injection of dykes. This assumption allows a simplification of the models which, thereby, give quick solutions, but does not reflect the complexity of the volcano.

By contrast, other authors (e.g. Gudmundsson, 2006) suggest that the layering may modify the state of stress in a volcano and thus the observed deformation. Using 2-D numerical models, it can be shown, for examples, that dyke-induced stresses at the free surface of a volcano depend on the mechanical model used: homogeneous or layered (Gudmundsson, 2003). It follows that the inferred geometry of a volcano-deformation source could, because of the layering, be mistaken. Another implication of these results is the potential effect that all kinds of heterogeneities may have on the stress distribution. Layering in a volcano is mainly the result of exogenous growth processes, which tend to accumulate stiff (high Young’s modulus) lava flows and soft (low Young’s modulus) pyroclastic and sedimentary layers. Endogenous growth processes, such as dyke injections, create stiff heterogeneities (Gudmundsson, 2006) that may partly control the magma paths in rift zones. Such heterogeneities may modify the state of stress in their surroundings during periods of unrest and thus affect the ground deformation. For example, the evolution of Piton de la Fournaise has been complex and many old intrusions have been identified within the volcano (Bachély and Mairine, 1990). The especially strongly asymmetrical west–east structure indicates that the edifice would in any case be subject to asymmetric deformation.

This study is partly based on a field campaign on Piton de la Fournaise volcano and partly on numerical modelling. The main aims are to explore, first, the structure of the plumbing system of the volcano and, second, how the plumbing system as well as fracturing contribute to the asymmetric surface deformation of the volcano observed during most recent eruptions.

2. Geological settings

Piton de la Fournaise is the active volcano on Réunion Island, an oceanic island located in the south-western part of the Indian Ocean (Fig. 1). Réunion Island is the emerged part of a ~7000 m high and ~220 km wide edifice composed of 3 volcanoes: Piton des Neiges (PN) and Les Alizés volcano, both inactive, and the active Piton de la Fournaise volcano (PF) (Malengreau et al., 1999). The eruptive history of PF began ~530 ka ago (Bachély and Mairine, 1990). The growth of the volcano has been affected by several major volcano-tectonic events (Bachély and Mairine, 1990); the most recent one led to the creation of a summit collapse, the Enclos Fouqué Caldera (EFC — see the location in Fig. 1 — Bachély and Mairine, 1990), 4.5 ka ago. The Grand Brûlé structure (located in Fig. 1) joined the summit caldera to create a single, continuous depression, limited by a 200 to 400 m high scarp, corresponding to the boundary fault, referred to, in this paper, as the Enclos Fouqué Caldera scarp (EFC scarp). The origin of the depression is still debated. Three main hypotheses as to the origin of the depression are discussed by various authors: (1) a landslide (Lénat et al., 1989; Gillot et al., 1994; Labazuy, 1996; Oehler et al., 2004); (2) a caldera collapse associated with a landslide (Bachély, 1981; Bachély and Mairine, 1990); and (3) a hybrid event with a calderalike depression and a landslide (Merle and Lénat, 2003). Recent studies (Bachély and Mairine, 1990; Merle and Lénat, 2003) tend to agree on a vertical collapse of the upper part of the depression.

A study of the past eruptive activity reveals that the eruptive centre has gradually migrated from the Plaine des Sables (Fig. 1) towards its present location (Bachély and Mairine, 1990), which is marked by a steep cone topped by 2 summit craters, Bory in the west and Dolomieu in the east (Fig. 1). The active
eruptive centre is fed by two feeding paths: (1) a central plumbing system, allowing the transport of magma from a shallow magma chamber, located at around sea level to the summit crater (Lénat and Bachèlery, 1990; Nercessian et al., 1996; Peltier et al., 2005, 2006); and (2) a system of rift zone segments connected to the central plumbing system and allowing the lateral transport of magma (Bachèlery, 1981; Lénat and Bachèlery, 1990; Fukushima, 2005; Peltier et al., 2005). At the edifice scale, mapping of eruptive features, from a set of aerial photographs obtained by the Institut Géographique National (IGN) in August 2003, shows the existence of a NNE–SSE trending main rift zone, confirmed in recent seismic studies (Brenguier et al., 2007), as well as a rift zone trending N120° joining the summit area of both PN and PF (PN–PF axis) (Figs. 1 and 2), mentioned by Bachèlery (1981) and rarely active in recent times. At the scale of the central cone, superimposed on the NNE–SSE path, a SW-intrusion path is present as well as a N120° intrusion zone southeast of the cone (Fig. 2) (Michel and Zlotnicki, 1998; Michon et al., 2007).

Many authors (Briole et al., 1998; Froger et al., 2004; Fukushima et al., 2005; Peltier et al., 2007) have noticed an asymmetric pattern of deformation during recent magma injections and eruptions (Fig. 3), as well as during the pre-eruptive inflations (Fig. 3), both affecting a limited area close to the summit and along the eruptive fissures. During dyke emplacement, the western part of the volcano shows little surface deformation, except close to the eruptive fissures, while the eastern part shows considerable deformation. The surface deformation is normally interpreted as being attributable to the geometry of the injected dykes (Zlotnicki et al., 1990; Sigmundsson et al., 1999; Battaglia and Bachèlery, 2003; Froger et al., 2004; Fukushima et al., 2005; Peltier et al., 2007).

3. Field results

3.1. General description

The EFC scarp provides a natural 200–300 m cross-section through the uppermost products of the volcano, but the vegetation

![Fig. 2. Eruptive fissures and cones visible on aerial photographs from August 2003 (provided by the French Institut Géographique National — IGN). The distribution allows the characterisation of the main dyke paths.](image)

![Fig. 3. A) Cumulative displacements recorded from September 2004 to January 2006. This period includes 2 eruptions occurring in the Plaine des Osmondes, an eruption on the northern flank of the summit cone and a summit eruption, and B) inter-eruptive displacements recorded between June and October 2005, as detected by the GPS network. On the left are the horizontal displacements, on the right the vertical displacements. Coordinates in Gauss–Laborde Réunion.](image)
coverage prevents detailed observations in some subsections. The only areas available for observations are the Bellecombe scarp, and the cliff surrounding the Plaine des Osmondes (Fig. 4). This cliff shows a tendency to a partial collapse, so we avoided taking measurements in the disturbed areas. Due to inaccessibility of the cliffs, some field measurements had to be made from a great distance and are subject to an error of ±5°.

The studied cliffs show a comparatively homogeneous pile of two main lava types, aa and pahoehoe. The variation of thickness among the lava flows is from ~1 m in the slopes to 5–10 m in the flat areas and coincides often with change in lava type. A high density of dykes correlates with highly fractured lava flows and pyroclastic layers (Fig. 5), mainly due to the creation of small scoria cones along the eruptive fissures.

3.2. Dyke studies

There are two main petrological types of magma emitted recently at PF; (1) largely aphyric basalt with less than 5% of phenocrysts, and (2) more primitive, picritic, olivine-rich basalt with more than 30% of olivine xenocrysts, locally referred to oceanite (Bachèlery, 1981; Albarède and Tamagnan, 1988). We found aphyric basalt dykes in both the areas of study, while olivine-rich dykes were found only in the Plaine des Osmondes. The dyke thicknesses are from 0.5 to 2 m and show no clear relation with the petrology.

The studied areas show different dyke trends. In the northern part of the Bellecombe scarp (Fig. 4) the most common trend is N120–125° (Fig. 6), whereas in the southern part of the area dykes have a radial distribution (Figs. 4 and 6), with a point of convergence 2–3 km west of the actual summit craters. The

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Fig. 4. Location of the study areas. In the Plaine des Osmondes area, note that the dip directions are presented through rose diagrams. The shaded circles represent the previous eruptive centre with their period of activity. The actual eruptive centre is located beneath the summit crater. The dashed lines mark the main rift zones. The thin white lines correspond to the strike of the dyke observed in the Bellecombe scarp. The northern part of the Bellecombe scarp is in shaded dark, while the south part is in shaded white. Coordinates in Gauss–Laborde Réunion.
dykes observed in the Bellecombe scarp, measured at a maximum depth of 150–200 m beneath the surface, are steeply dipping (70° to 90°) with a mean values of 77.5° (southward dipping) and 79.9° (northward dipping) (Fig. 7).

In the Plaine des Osmondes, the density of dykes increases on approaching of the rift zone (Fig. 4). Some 90% of the dykes trend N0–40° (Fig. 6), half of which are eastward (seaward) dipping, and are mostly steep (82.1°±12.9°; Fig. 7). One-fourth of these dykes dip westward and are more gently dipping (74.8°±15.8°; Fig. 7). The remaining one-fourth of the N0–40° trending dykes are vertical. A second trend, N85°, is constituted by only 4% of the dykes (Fig. 6), sub-parallel to the scarp and mainly northward dipping. Some 6% of the dykes do not belong to either of the main trends but strike in various directions without any dominating orientation. The dykes in the Plaine des Osmondes were measured at a depth of 400 m beneath the actual surface.

3.3. Application to the recent activity

Our field data provide new results as regards the dyke distribution in PF. Various studies have used deformation data to infer the geometry of associated dykes, but do not consider similar, exposed dykes. The dykes observed in the EFC scarp are clearly older than the event that created the scarp itself (>4.5 ka) and correspond to a volcanic centre located west of the present one (Bachély and Mairine, 1990). Therefore, to allow a comparison between our field data on exposed dykes and the inferred geometries of dykes using recent deformation data, we need to be sure that the state of stress at present is similar to that which existed when the exposed dykes were emplaced.

The age of the EFC collapse is estimated at 4.5 to 5 ka (Bachély, 1981; Bachély and Mairine, 1990). Gillot et al. (1994) provide dating of 13±6 ka for a lava flow at the base of Piton de Crac (Fig. 1 for location), a remnant block of the pre-EFC collapse topography (Bachély, 1981). Despite an uncertain origin of the Plaine des Osmondes, the consistent altitude between the Plaine des Osmondes and Piton de Crac indicates a similar age. Another dating from Gillot et al. (1994) provides an age for the base of the Bellecombe scarp of 11±4 ka. Thus, intrusions in both areas appear to be older than 4.5 ka and younger than 13±6 ka.

Maintaining a state of stress within the volcano implies maintaining the parameters that can influence the stress. Normally, topography and geometry of the plumbing system, its host rock mechanical properties, as well as the regional stress field, affect the local state of stress. No major plate-scale change has occurred in the past 20 ka around Réunion Island, suggesting that the regional stress field has been essentially constant through this time. The topography, however, has changed in the past due to the creation of the EFC depression and the growth of the central cone. Based on the height of the caldera scarp, there is an elevation difference of ~200 m between the present topography and the pre-EFC collapse topography. Furthermore, Bachély and Mairine (1990) have shown that the eruptive centre has migrated in the
past, from the Plaine des Sables (see Fig. 1 for location) towards its current position (Fig. 4). Our observations confirm that an intermediate stage of migration occurred (Bachèlery and Mairine, 1990), as is also supported by gravimetric studies (Rousset et al., 1989; Malengreau et al., 1999). The first stage, at around 150 ka (Bachèlery and Mairine, 1990), generated a central plumbing system west of the present summit craters (Fig. 4). The present location of the central plumbing system is the result of a migration that probably occurred some 4.5–5 ka ago (Bachèlery, 1981) (Fig. 4). The present northern branch of the rift zone seems to coincide with the pre-EFC collapse rift zone northern branch. Location of recent eruptions shows that the same path used before the EFC collapse is still used by dykes (Fig. 4 and 8). Only the central part of the rift system appears to have been reorganised because of the shifting of the volcanic centre.

Through the time considered, the loading variations affecting the state of stress are at a volcano-scale. The present position of the volcanic centre is closer to the eastern free flank and thus more subject to instabilities. As suggested by Cayol and Cornet (1998), reorientation of stresses may occur in the depth range of a few hundred meters (<250 m) below the free surface (Fig. 9). Such effects on dyke propagation will thus be limited to the summit cone, or close to the eastern flank, and cannot be ruled out for the dyke observed. Thus, the comparison between field data on exposed dykes and recent models on dyke geometries from geodetic data is, despite these variations, justified.

The main difference between geodetic and field results concerns the dip values obtained through modelling feeder dykes of recent eruptions and the dyke dips observed in the field. Models give generally shallow seaward dipping dykes on their entire dip dimension (53°–75°) (Cayol and Cornet, 1998; Froger et al., 2004; Fukushima, 2005; Fukushima et al., 2005; Peltier et al., 2007) while field results show a mean dyke dip of around 80° at maximum depth of exposure of 400 m (Fig. 7).

4. Modelling

Several models have been proposed to explain the observed asymmetric deformation (Fig. 3), such as shallowly eastward dipping dykes (e.g. Cayol and Cornet, 1998; Froger et al., 2004; Fukushima et al., 2005; Peltier et al., 2007). These shallow dips are not supported by the field data provided in this study.

We have made many 3-D numerical models, using the code ANSYS (www.ansys.com; Logan, 2002) including fractured medium and presence of stiffer rocks within the edifice to test the influence of heterogeneities within the volcano on the asymmetric ground deformation. We have modelled the dyke-induced deformation associated with dykes following the most common path since 1998 (Fig. 8). In the models, we also used various models for the dyke-propagation style. Peltier et al., 2005 suggest that a vertical dyke starts from a magma chamber at around 300 m a.s.l. Some dykes propagate laterally, along parts of their paths. However, a lateral injection does not always

Fig. 7. Histogram of dip distribution for the Bellecombe scarp and the NE rift zone.

Fig. 8. Eruptive fissures since 1998 (white). The dashed black lines correspond to the main trends observed. Coordinates in Gauss–Laborde Réunion.
occur (Peltier et al., 2006): it may depend on the existence or not of a large-scale level of neutral buoyancy (Ryan, 1987; Pinel and Jaupart, 2004; Fukushima, 2005). The local stress, which depends on the presence of recent intrusions, increasing temporarily the compression, can also influence whether lateral injection occurs, and, so can the magma pressure.

In the models, we take the elevation of 1800 m a.s.l. to be the level of initiation of lateral dyke injections (1600 m a.s.l. in Fukushima, 2005). This altitude coincides with that of the lowest eruptive fissures observed in the surroundings of the cone. Dykes are modelled as vertical fluid-filled cracks, opening with a magma overpressure of 2 MPa (Fukushima, 2005; Peltier et al., 2007) applied horizontally to the dyke walls. The western and bottom boundaries of the models have a zero displacement since PF volcano is buttressed in its western part by the Piton des Neiges volcano (Fig. 1).

### 4.1. Influence of the superficial fracturing

To test the influence of pre-existing fractures on dyke-induced ground deformation, we have included vertical fractures. The fractures are modelled as open cracks within a homogeneous medium and extend to a depth of 1400 m a.s.l. (1200 m under the summit craters and 600–800 m under the floor of the caldera). The maximum depth for tension fractures can be obtained using an equation based on the Griffith criterion (Gudmundsson, 1992).

Taking the maximum principal compressive stress to be vertical ($\sigma_1$), the maximum depth is:

$$d_{\text{max}} = \frac{3T_0}{\rho_1 g}$$

where $\rho_1$ is the rock density and $T_0$ is the tensile strength.

Numerical applications for PF give us a range of 300 to 800 m, while Fukushima (2005) suggests a depth of 1000 m beneath the ground for open fractures. The pressure source is a vertical dyke that stalls at a shallow depth in the northern branch of the rift zone (Fig. 8). The en échelon structures of the dyke thought to be present only at a shallow level and correspond to the reorientation of stresses close to and at the surface (Cayol and Cornet, 1998; Fukushima, 2005). Ignoring the en échelon pattern allows us to observe the general pattern of deformation without the short wavelength deformation due to the en échelon structures.

For this set of models, we consider a homogeneous elastic medium with a Young’s modulus (elastic stiffness) of 5 GPa and a Poisson’s ratio of 0.25 (Fukushima et al., 2005). It also takes into account the topography, which is extracted from a 25 m step Digital Elevation Model (DEM), filtered to a 400 m step. The main structure and topography are kept, the filtering occurring on the small features such as small cones.

A degree of asymmetry (DA) for each model is defined by the maximum deformation in the eastern (or northern) side of the dyke, divided by the maximum deformation in the western (or southern) side. Fig. 10 compares the total ground displacement for a fully, non-fractured homogeneous medium, and models considering a homogeneous medium including sets of fractures west or east of the intrusion, as well as on both sides (Fig. 10). The results show a slight increase in the asymmetry of deformation by a few percent (8%) between the homogeneous model and the two models including fractures to the east of the intrusion. When fractures are located to the west of the intrusion, the deformation is favoured on the western side of the dyke, reducing the natural asymmetric pattern of deformation.

### 4.2. Influence of the rift zone system

Since remnants of the pre-EFC plumbing system could be present in the western part of the volcano, such as dykes and sills, we tested the influence of stiffer rocks within the volcano (Fig. 11) both during the pre-eruptive inflation and the dyke injection. We also tested the influence of the actual rift zone system by including the structural control that it has on the magma injection paths (Fig. 8), characterised by curved dykes.
The first set of models presents the resultant total deformation for the injection of vertical and planar dykes in a fully homogeneous medium (Fig. 12) along the three main dyke paths observed since 1998 (Fig. 8). The second set of models tests separately the effect of the presence of an old plumbing system with vertical and planar dykes. In the third set, the curvature of vertical dykes is tested. We then combine both effects in a last set of models (Fig. 13). For the comparison we also use a degree of asymmetry.

For these sets of models, the same program is used, and the dykes considered also stall at shallow depths. Three different media were used to define the different models (Fig. 11). Two are elastic isotropic media with a stiffness (Young’s modulus) of 5 GPa corresponding to the lava pile constituting the main part of the volcano (Fukushima, 2005), and 8 GPa corresponding to the central part of the old plumbing system. The third medium is an elastic orthotropic medium with a stiffness of 8 GPa in the X and Y direction and of 4 GPa in the Z direction that corresponds to the branches of the old plumbing system’s rift zone.

Results presented in Fig. 13, when compared with the homogeneous models (Fig. 12), show that in all the cases the presence of the old plumbing system to the west of the actual active part tends to increase slightly the natural asymmetry existing between the east and the west of the intrusion. The curvature of the dyke appears to have a greater effect on the asymmetry of deformation. Both mechanisms combined show a strong increase (by up to 42%) of the asymmetry, which does not correspond to the simple addition of the effects.

Fig. 10. Comparison of the deformation induced by a vertical dyke in a: A) homogeneous medium, and in a fractured medium with fractures; B) to the east of the dyke; C) with both set of fractures; and D) to the west of the dyke. The trace of the dyke is marked by the black dashed line, and the fractures are marked by the thick black lines. Percentages indicate the change of DA from the homogeneous model. Coordinates in Gauss–Laborde Réunion.

Fig. 11. Location of the rift zone within the model (dashed line), and limit of the computed area (thin black line) using Ansys. The numbers correspond to the various medium used. 1 and 2 are elastic isotropic medium of stiffness (Young’s modulus) 5 GPa and 8 GPa, respectively. 3 is an orthotropic medium with X and Y=8 GPa and Z=4 GPa. The Poisson’s ratio is common for all the materials and equal to 0.25. Coordinates in Gauss–Laborde Réunion.

Fig. 14 shows the deformation triggered by an overpressure of 10 MPa in a shallow magma chamber located at sea level (Nercessian et al., 1996; Peltier et al., 2005). It allows modelling...
of the pre-eruptive pattern of deformation. The first model is homogeneous with the same characteristics as the other model (stiffness of 5 GPa), while the second model includes the old plumbing system defined above. Its presence leads to a shift of the peak of deformation slightly to the east of the crater, in agreement with the pre-eruptive pattern of deformation observed at PF (Fig. 3). The deformation is also 20% larger because of the rift zone.

5. Discussion

5.1. Complex plumbing system

The petrological characteristics of an erupted magma provide information on the path followed by the magma. In our studied areas, olivine-rich basalts have only been observed in the Plaine des Osmondes. Recent eruptions generally support
our observations, although olivine-rich basalts can occasionally be observed in other places, along the SSE rift zone (Albarède and Tamagnan, 1988) or in the east of Dolomieu crater, and along the PN—PF axis, in the Plaine des Sables. According to Bachèlery (1999), olivine-rich basalt magmas are likely to derive from a deep-seated reservoir and are associated with the refilling of the associated shallow magma chamber.

As suggested by Peltier et al. (2006), and in the absence of eruptive fissures in the summit area (Fig. 14), olivine-rich basalt eruptions occurring in the Plaine des Osmondes are fed by lateral dykes propagating at greater depths. The 1998 eruption shows that aphyric dykes can also propagate, at shallow depths, far from the central plug (Fig. 15) (Bachèlery, 1999; Peltier et al., 2006). This eruption is considered as a refilling event but does not reflect the common style of an oceanic eruption (low effusion rate and no olivine crystals for 1998, while oceanitic eruptions have a high effusion rate) nor aphyric eruptions occurring at PF, because of its duration and the large volume of evolved magma. Aphyric basalts in the recent period are mainly erupted in the summit craters or on the flank of the cone and its close proximity (Peltier et al., 2006). The PN—PF axis is represented mainly by aphyric basalts or aphyric basalts bearing olivine nodules (Bachèlery, 1981; this study). Bachèlery (1981) suggests that the lavas emitted along the PN—PF axis have a deep origin, the nodules of olivine coming from the ascension of magma through a shallow (partially) crystallised magma chamber.

This difference in lithology seems to reflect different magma paths as well as different depths of origin. The N120° direction is found at all scales on the island and the oceanic crust (Michon et al., 2007), and therefore might control the transport of magmas coming from great depths. The NNE—SSE trending rift zone likely reflects gravitational processes (Walter et al., 2006) and controls the paths of magma at shallow depths, in the active plumbing system. Within the NNE—SSE rift zone, long distance or short distance dyke propagation reflect transport at different depths (Aki and Ferrazzini, 2000; Peltier et al., 2006). The depth of propagation reflects a difference in overpressure and/or density of the magma; the higher the pressure, the deeper origin of the magma. Existence of a complex storage system can also explain partly the different depths of magma injection (Aki and Ferrazzini, 2000). Fig. 16 illustrates the various magma/dyke paths within the volcano.

5.2. Rift zone, fractures and structural control

Field results confirm that the magma follows some preferential paths in the volcano, such as NNE—SSE, SW, and N120°, marking the rift zones where most of the recent volcanic vents occur (Figs. 2 and 8). In the field, rift zones coincide with an increase in the density or intensity of dykes, as well as with an increase in the proportion of pyroclastic (scoria) deposits (Fig. 5). The number of tectonic fractures is also much greater within a rift zone than outside the zone. By decreasing the large-scale elastic stiffness and the tensile strength of the host rock, fractures make dyke paths easier to form. This effect on magma paths is observed within the rift zones as well as in association with all the large fractures of the volcano. Field data in the Plaine des Osmondes show a second trend of dykes coinciding with the orientation of the scarp. Recent eruptions demonstrate the structural control that the faults associated with the EFC scarp exert on the magma path.
On 5th of January 2002, an eruption began. After a period with a low seismic tremor, the tremor increased. On 12th of January 2002, a new eruptive fissure opened at the foot of the EFC scarp, off the apparent axis of propagation of the dyke, suggesting propagation along the EFC scarp following the pre-existing fractures (Fig. 15). On 17th of February 2005, another eruption occurred in the Plain des Osmondes. On 25th of February, one day before the end of 17th February fissure eruption, a new eruptive fissure opened, in the Trou de Sable (Fig. 15), at 500 m a.s.l., again along the ECF scarp. The new opening was accompanied by an increase in the seismic tremor, and without any evidence of new magma injection from the summit area, suggesting a further propagation of the initial dyke along the pre-existing fracture boarding the EFC scarp. The lavas emitted in the Trou de Sable are olivine-rich basalts and strongly degassed. In December 2005, GPS data showed that the dyke, responsible for the opening of a new eruptive fissure within the EFC scarp, propagated further outside the EFC depression. The load difference due to the difference of altitude may have prevented the opening of an eruptive fissure outside the EFC depression as suggested by Acocella et al. (2006) for Vesuvius.

However, in April 1977, an eruption occurred far away from the EFC scarp, outside the EFC depression. Thus, it appears that the EFC scarp and the associated faults did not control all the eruption sites. Injected dykes with a high overpressure, such as the April 1977 dyke (effusion rate of the April 1977 eruption was 3 times higher than the other considered eruptions — Global Volcanism Program website), managed to propagate through the fractures associated with the EFC scarp.

The pre-existing fractures play an essential role in forming the magma path in the rift zone as well as allowing propagation of magma along the EFC scarp. The scarp and associated faults also have an important role as they act as structural barriers, limiting the opening of fractures outside the depression (Acocella et al., 2006) and exert a strong structural control on the magma paths.

5.3. Influence of rift system on deformation

Our field results, especially from the northern branch of the NNE–SSE rift zone, and the computed dykes (Froger et al., 2004; Fukushima, 2005; Peltier et al., 2007), show a marked difference. The calculated dyke dips, characterising the entire dip dimension of each dyke, are too shallow in comparison with our field observations of dykes propagating at shallow depths (<400 m). Moreover, although calculated dykes may explain the asymmetric deformation during the injection, they offer no satisfactory explanation for the asymmetric deformation occurring prior to dyke injection (Fig. 3). We have explored the influence of various heterogeneities on the ground deformation.

The fully homogeneous models show that a natural asymmetry exists as the deformation in the east is greater than that in the west, demonstrating the influence of the topography on the deformation. In a first set of models, we show that few tectonic fractures within the rift zone have a limited influence on the ground deformation, increasing it by a maximum of 8% the DA (Fig. 10). By decreasing the elastic stiffness of the rocks (Bell, 2000), fractures make the medium more easily deformed and increase the natural asymmetry of deformation that exists. As suggested by Fig. 2, the extension of fractures is much greater than in the models. Increasing the number of fractures tends to reduce further the elastic stiffness and thus increase their influence on the ground deformation.

We have also considered the influence of an old plumbing system, located to the west of the actual volcanic centre, on dyke paths (Fig. 11). Studies (Roussel et al., 1989; Malengreau et al., 1999) suggest that this old plumbing system was possibly preserved during the vertical collapse of the EFC (Bachelery, 1981; Merle and Lénat, 2003). The results of numerical modelling indicate that the presence of heterogeneities (old plumbing system) increase, somewhat (<7%) the asymmetric deformation both during dyke injection (Fig. 13) and pre-eruptive inflation (Fig. 14).

We have also tested the effects of curvature of dykes on the surface deformation. The complex evolution of the rift system through time, associated with the gradual migration of the eruptive centre, has resulted in a curved rift system (Figs. 2 and 8). Magma paths follow this curvature (Michon et al., 2007). Our models show that the dykes curvature has potential effects on deformation in some cases (Fig. 13), but is far less than actually observed (e.g. Fukushima, 2005). However, the combination of both the curvature of dykes and the presence of the old plumbing system west of the actual intrusion zone have great effects on the ground deformation and make it closer to what is observed.

The presence of the old plumbing system (dense and stiff rocks) west of the actual plumbing system tends to limit the deformation in this direction. The stiff rocks create a barrier and isolate the eastern flank from the rest of the edifice. A similar model has been proposed for Kilauea in Hawaii (Delaney and Denlinger, 1999). However, in the case of Hawaii, the existence of a basal décollement allows a more efficient decoupling between the two sides of the rift zone. Curvature of a dyke tends to have two effects. First, it increases the area affected by the loading variation on the convex part of the dykes, and thus dissipates the stress. Second, it concentrates the stresses in the concave part of the dyke, increasing the compression (Walter et al., 2005) and thus the deformation.

For dyke injections to the east of Dolomieu crater along a direction N120°, our models explain the kind of asymmetry observed during some but not all the eruptions. The maximum deformation observed in this area is not always on the same side of the eruptive fissures (Fukushima, 2005). The irregular pattern of deformation along this direction seems to reflect mainly variation in dyke dip that can be partly the effect of the topography of the summit cone (Fig. 9). However, a more complex structure in the eastern part of Dolomieu at depth, linked with older intrusive system, cannot be ruled out. In the other two cases studied, the asymmetry obtained in our models does not properly fit the deformation data. A complex internal structure, or the variation in the dyke dip are parameters that allow a better fit of the models and observed deformation. Our models consider vertical dykes, while the stress field associated with the free flank is likely to interfere with the propagating dykes (Fig. 9). However, the dyke dip needed to explain the
asymmetry will be less in a heterogeneous medium and thus in a better agreement with our field observations of exposed dykes and the inferred stress field (Fig. 9).

6. Conclusions

In this study we provide new data on the magmatic plumbing system of Piton de la Fournaise, and new results on the deformation of the volcano before and during an eruption. Our main conclusions may be summarised as follows:

- The distribution of exposed dykes confirms that the volcanic centre has migrated in the past, its current position being the result of gradual migration from the Plaine des Sables, to close to the Bellecombe scarp (Fig. 4), and then towards its current location. The first migration occurred ~150 ka ago (Bachèlery and Mairine, 1990), followed by the second migration which was contemporaneous with the formation of Enclos Fouqué Caldera.

- The lithology of dykes, observed along the rift zone outside the central cone, indicates a complex plumbing system. Multiple paths are dependent on the depth of origin of the magma as well as the location within the volcano. The PN–PF axis reflects a deep structure (Michon et al., 2007) and controls the transport of magma from deep-seated sources. The NNE–SSE trending rift zone, perhaps of gravitational origin (Walter et al., 2006), controls the magma paths at shallower depths (Fig. 16).

- Fractures induced by recurrent intrusions and long-term deformation of the volcano influence the magma paths. Fractures linked with volcano-tectonic processes such as collapse or landslide influence the propagation of magma along the Enclos Fouqué Caldera scar. The effects of fractures and topographic load variations have a direct impact on the assessment of volcanic hazards as it prevents the opening of fractures outside the Enclos Fouqué Caldera (Fig. 15). Propagation and opening outside the Enclos Fouqué Caldera are linked with highly overpressured magma.

- Our field data (Fig. 7) indicate much steeper dyke dips than geodetic models previously published to explain the asymmetric deformation (Fig. 3; e.g. Zlotnicki et al., 1990; Cayol and Cornet, 1998; Sigmundsson et al., 1999; Battaglia and Bachélery, 2003; Froger et al., 2004; Fukushima et al., 2005; Peltier et al., 2007). We provide new numerical models that take into account existing fractures and intrusions (Figs. 10 and 13). These new models allow an explanation of the asymmetric deformation (Figs. 10, 12, 13 and 14) and provide a better fit with the field data.

- Our models show that complex structure, fracturing and structural control of magma paths allow to explain part of the asymmetric pattern of deformation observed at PF. The same kind of behaviour and influence can be expected in many volcanoes that have a long and complex eruptive history. A better understanding of the past history and internal structures of volcanoes is essential for reaching a better understanding of their present deformation.

Fig. 16. Schematic representation of the plumbing system of Piton de la Fournaise. The depth of shallow magma chamber for the two past volcanic systems (530–150 ka and 150–5 ka) are not well constrained. Thick lines correspond to a plug supposed to be use recurrently to feed eruptions at the summit or on the summit cone. By analogy, we extend the presence of the plug to the two past volcanic systems. At greater depths in the plug, olivine-rich magmas are laterally injected to feed eruptions in the Plaine des Osmondes. The deep magma chamber is inferred form seismic data (Battaglia and Bachélery, 2003), while the remnants of Les Alizés volcano are based on gravimetric studies (Rouset et al., 1989; Malengreau et al., 1999). Coordinates in Gauss–Laborde Réunion.
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