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RESEARCH ARTICLE

The interplay between collapse structures, hydrothermal systems, and magma intrusions: the case of the central area of Piton de la Fournaise volcano

Jean-François Lénat · Patrick Bachèlery · Aline Peltier

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Abstract We explore the possible relationships between a structural heterogeneity, the hydrothermal system, and the intrusive activity at Piton de la Fournaise volcano. Geological and geophysical data show that as the result of repeated collapses (the last one in 2007), a cylinder of faulted, fractured, and crumbled rocks must exist between the surface and the top of a magma reservoir at about sea level. This structure constitutes a major geological heterogeneity. An obvious spatial correlation exists between this column of fractured and brecciated rock and the location of (1) most of the seismic activity, (2) a low-resistivity dome, (3) a huge self-potential anomaly, (4) thermal evidence of hydrothermal activity, and (5) the root of magma intrusions. The dominant factors that make this structural heterogeneity a trap for the activity are probably its higher permeability and its weaker mechanical strength. Evidence exists for the presence of an active hydrothermal system confined in this permeable zone. The long-term stability of the activated zone above sea level and the similarity of the pre-eruptive crises, in spite of the inferred large perturbation of the magmatic system in 1998, suggest a common triggering mechanism for all the eruptions since at least the first data recorded by the observatory in 1980. This mechanism can be purely magmatic, resulting from the pressurization of a reservoir, but we also propose that the hydrothermal system may play a role in the development of volcanic instabilities. A qualitative model is proposed to explain the triggering of magma intrusions by hydrothermal processes, and its speculative aspects are discussed. This work represents a first attempt to integrate the structural and dynamic information in a unified framework at Piton de la Fournaise.

Keywords Collapse · Hydrothermal system · Piton de la Fournaise · Eruption · Pre-eruptive crisis · Magma transfer

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Introduction

Volcanic activity is controlled by many factors, some related to deep processes and some related to specific characteristics of the edifices. Among the latter, structural and lithologic heterogeneities, local tectonic and gravitational stresses, and hydrothermal systems can play a major role in the location, frequency, and style of the eruptions. In this work, we focus on the case of a well-studied and monitored volcano, Piton de la Fournaise, to explore the possible relationships between a major structural heterogeneity, the hydrothermal system and intrusive activity.

Analysis of historic volcano-tectonic activity in the summit zone leads us to postulate that because the summit has collapsed repeatedly, a brecciated column exists beneath the summit craters. This structure constitutes a major geological heterogeneity. It acts as a “trap” for the seismic, intrusive, and hydrothermal activities which are remarkably concentrated in this column.

A synthesis of the present knowledge of the internal structure and behavior of Piton de la Fournaise is first carried out to provide the basis for a discussion of the role played by the fractured column on the internal activity. We then extend the discussion to possible interactions between the hydrothermal system in the column and the generation of seismic activity and intrusions from a shallow magma reservoir.

Geological and geophysical structure of the central zone of Piton de la Fournaise

Piton de la Fournaise, on the southeastern part of La Réunion Island, is one of the world's most active basaltic shield volcanoes. Its activity began about 500,000 years ago (Gillot and Nativel 1989). Bachèlery and Mairine (1990), and Bachèlery and Lénat (1993) distinguish two phases of construction forming the “Ancient Shield” (>0.15 Ma) and the “Recent Shield” (<0.15 Ma), marked by several volcano-tectonic events which are detailed in Merle et al. (2010). The presence of a buried volcano, named Les Alizés, on the eastern coast area of Piton de la Fournaise, has been inferred from gravity (Malengreau et al. 1999; Gailler et al. 2009), magnetic (Lénat et al. 2001; Gailler and Lénat 2010), and drillhole (Rançon et al. 1989) data. The Alizés activity would pre-date, or be partially contemporaneous with that of Piton des Neiges, the other volcano of La Réunion (Smietana et al. 2010). It would also be the base on which Piton de la Fournaise has grown.

The recent activity of Piton de la Fournaise has been restricted mainly to effusive eruptions on the Central Cone (Fig. 1) and along the NE and SE rift zones, with some prehistoric eruptions taking place outside of these main structures. The Central Cone, built in the depression of the Enclos Fouqué, is 3 km in diameter and 400 m in height, with a mean slope of 15°–20°. The summit is occupied by a western crater, the Bory crater (350 by 200 m), and a larger eastern one, the Dolomieu crater (1,000 by 700 m) (Fig. 2) (Carter et al. 2007).

Formation and evolution of the summit craters

Dolomieu is a collapse crater. Although the smaller Bory crater, west of Dolomieu crater, is regularly intersected by eruptive fissures (e.g., 1937, 1942, 1956, 1979, 1981, 2003), it has probably not been active as a central crater since the end of the 18th century (Bory de Saint-Vincent 1804; Lacroix 1936, 1938). The zone occupied by the present-day Dolomieu crater has been the summit focus of eruptive activity from at least the 1766 eruption (Bory de Saint-Vincent 1804; Lacroix 1938). A series of craters have repeatedly formed and disappeared by refilling since 1776 (1791, 1844–51, 1860), but the observations are incomplete and poor before the 20th century. Lacroix (1936, 1938) published a synthesis of the available historical information and a more detailed description of the changes at the summit between 1911 and 1936. In 1911, the zone of the Dolomieu crater was virtually filled by lava flows, and only the western upper rims of a previous collapse were visible. In 1936, aerial photos (Fig. 2a) show that an area, 700–800 m in diameter, on the eastern part of the summit had collapsed by about 150 m. The accounts from sporadic visits to the summit (1927, 1929, 1930, 1933, 1934, and 1935) describe a progressive collapse of this area with a paroxysm in 1934–35 (Jean 1935). Note that in 1931, a huge eruption, located at an elevation of about 1,400 m near the northern rim of the Enclos Fouqué, emitted about 130×10^6 m^3 of lava (the second largest volume, after that of the April 2007 eruption (Bachèlery et al. 2010), for a single eruption since 1920, the starting date for the present eruption volume database).

Since that period, three smaller pit crater collapses have occurred (1953, 1986, and 2002) inside the Dolomieu crater, near its border faults (Ducrot 1958; Delorme et al. 1989; Hirn et al. 1991; Longpré et al. 2007). The Dolomieu crater has been gradually filled by the products of subsequent eruptions (Fig. 2b) resulting in its complete replenishment at the end of the August 2006–January 2007 eruption, when lava flows spilled over the rim at its lowest elevation, to the east. The 340-m deep April 2007 collapse (Fig. 2c) of the Dolomieu crater was associated with exceptional eruptive and seismic activity extensively described in several articles (Michon et al. 2007; Urai et al. 2007; Peltier et al. 2009c; Staudacher et al. 2009). In contrast to the 1929–36 collapse, that in April 2007 was less than 24 h and occurred during an eruption, located at about 600 m in elevation, near the SE border of the Grand Brûlé depression, discharging 220×10^6 m^3 of lava flows (Bachèlery et al. 2010). Visual observations and seismic, geodetic, and petro-geochemical data acquired during the 2007 crisis indicate a direct relationship between the eruption of a large volume of magma at low elevation and the collapse of the summit. The key element is the inferred presence of a magma reservoir located beneath Dolomieu at about sea level. The presence of such a reservoir, first proposed by Lénat and Bachèlery (1990), has been more recently discussed and documented by Aki and Ferrazzini

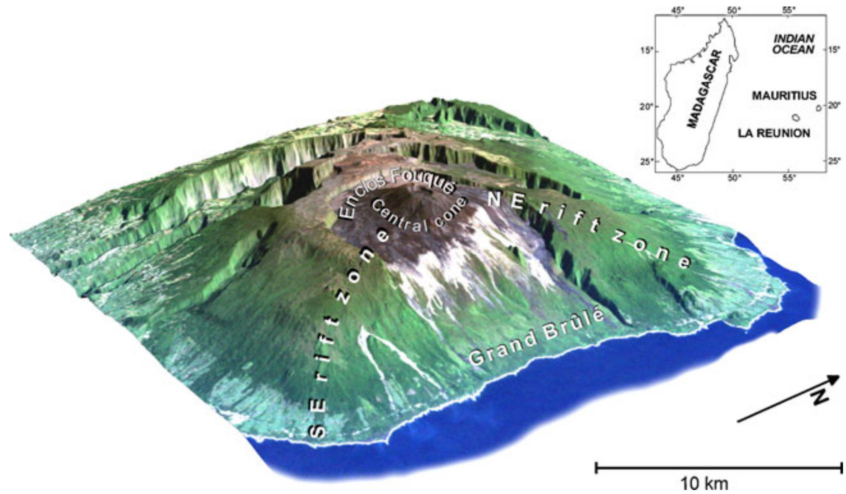

Fig. 1 3D oblique view (DEM+Landsat image) of Piton de la Fournaise from the southeast

(2000), Peltier et al. (2009a), and Prôno et al. (2009). In general terms, the collapse of the summit may be explained by the draining of the reservoir associated with the huge distal low-elevation eruption (Michon et al. 2009b; Peltier et al. 2009c; Staudacher et al. 2009).

Internal structure of the summit area

Geophysical information

The structure of the central area of Piton de la Fournaise can be constrained using both static and dynamic geophysical data.

Available static geophysical data comprise resistivity, self-potential (SP), gravity, and seismic information. Figure 3a shows the resistivity distribution established by Lénat et al. (2000). A dome of low-resistivity values, culminating at a depth of a few hundreds of meters beneath the surface, has been interpreted as a hydrothermal system (note that the bottom of the low-resistivity zone is not defined owing to the limited depth investigation of the methods used—see Lénat et al. (2000)). The fact that this structure is also associated with a huge positive SP anomaly (Malengreau et al. 1994; Lénat 2007) indicates that the inferred hydrothermal system is active (Zlotnicki and Nishida 2003). The gravity pattern of this area is dominated by a short wavelength anomaly attributed to the construction of the Central Cone by low density materials (Gailler et al. 2009). The seismic tomographies (all based on data acquired before the 2007 collapse) (Nercessian et al. 1996; Brenguier et al. 2007; Prôno et al. 2009) show a positive seismic velocity anomaly beneath the summit area, extending from the near surface down to about 1,000 m a.s.l (Fig. 3b) (note that the horizontal location of this structure slightly fluctuates in the different tomographies, probably as a result of the different data-sets used).

This velocity anomaly could indicate the presence of: (1) a pile of thick lava flows which have recurrently filled the summit crater before being periodically dragged down into the collapses, (2) a high density of dikes, or (3) small cooled magma bodies emplaced at shallow depth, or a combination of these structures. Such layers should have a higher seismic velocity than the average material of the Central Cone (see below). Below 1,000 ma.s.l., down to about sea level, Prôno et al. (2009) describe a relatively low-velocity zone. There, the pile of thick lava flows may not exist or may become progressively more brecciated as a result of the successive collapse episodes and the frequent fracturing by intrusions. Most of the pre-eruptive earthquakes occur in this volume (see below). A second low-velocity anomaly is located between about 1,000 and 2,000 m b.s.l. The two low-velocity zones are separated by a relatively high-velocity volume, at about sea level with a strong velocity gradient suggesting the presence of a prominent mechanical (and thus probably lithologic) vertical transition. In addition, Battaglia (2001) shows that at least during the March 1998 crisis, the base of the two low-velocity zones correspond to low seismicity regions (Fig. 3b) (named D2 at 1.5–2 km b.s.l and D3 at sea level) and that a high attenuation zone exists above D2 (at about 1.5 km b.s.l.).

The dynamic geophysical information is provided by the monitoring data which have been collected by the Piton de la Fournaise Volcanological Observatory since 1980. It has been summarized in early works by Lénat and Bachèlery (1988, 1990) and more recently by Peltier et al. (2009a). Among the main results, we can highlight the following: (1) before every eruption or shallow intrusion, a volume between a few hundreds of meters beneath the surface and about sea level, in an area corresponding to that of about the summit craters, is seismically active (Fig. 3a, b), with most of the events located in the upper low-velocity zone, and (2) the pre-eruptive deformation pattern, when present,

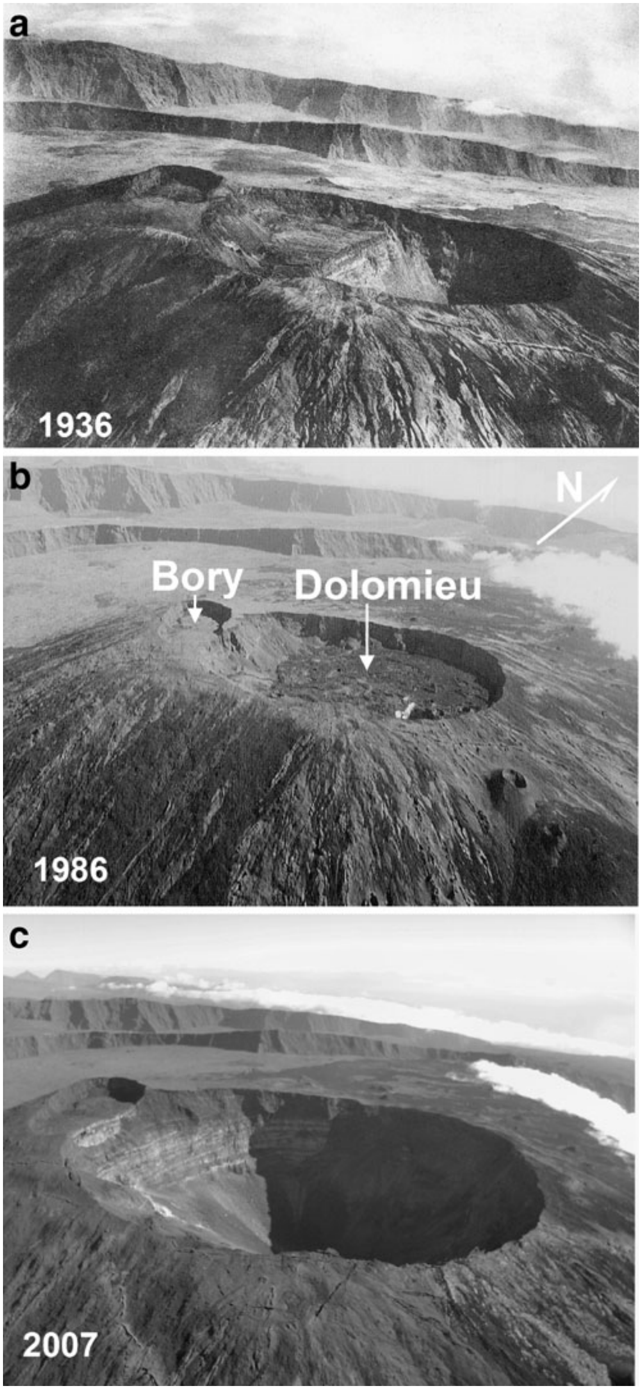

Fig. 2 Evolution of the summit area between 1936 and 2007. **a** 1936 photo from Lacroix (1938); **b** 1986 photo from Lénat, and **c** 2007 photo from Guillaume Levieux (person. com.)

is mostly restricted to the summit area or to the Central Cone. In addition, joint analyses of the deformation, seismicity, and petrology–geochemistry of the lavas from the eruptions have led different authors to infer the presence of a magma reservoir beneath the central area at about sea level or above (Lénat and Bachèlery 1988; Lénat et al. 1989a, b; Lénat and Bachèlery 1990; Sigmarsson et al. 2005; Boivin and Bachèlery 2009; Peltier et al. 2009a; Prôno et al. 2009). The exact depth, volume, and shape of this reservoir are yet to be definitely established, as well as if there are several separate reservoir units. On the base of ground deformation and seismic data, Peltier et al. (2009a) and references herein favor the presence of a reservoir near sea level, during at least the 1998–2007 period, within the upper low-velocity zone. Prôno et al. (2009) suggest that the deeper low-velocity zone should also be considered as a possible reservoir location. In addition, some (infrequent) summit eruptions (e.g., July 1986) are preceded by extremely brief seismic crises (10–15 mn), suggesting that magma could have been stored at a very shallow depth, as suggested by the model proposed by Lénat and Bachèlery (1990). In this study, for simplicity, we will infer that a magma reservoir exists near sea level, because the exact elevation of the reservoir is not critical for our model. Also, it is not known to what extent the 2007 events may have changed or perturbed the plumbing system (the seismic tomographies predate the 1998 events, and the work by Peltier et al. (2009a) is based on post-1998 to 2007 deformation signals). Preliminary results suggest that the increase in the frequency of pure dyke intrusions in 2008 and 2009 has been strongly influenced by the Dolomieu caldera collapse (Peltier et al. 2010). Stress changes in the volcanic edifice have, temporarily at least, favored the arrest of dykes at depth after 2007. During the successive dyke intrusions, magma stopped at shallow depth (>1,000 m a.s.l.) to form transitory bodies of stored magma before erupting during successive small summit eruptions.

Geological information

The 340-m collapse of the Dolomieu crater in 2007 (Staudacher et al. 2009) created excellent exposures of the crater wall, enabling the geological structure of the summit area to be inferred. Five main types of features are observed: piles of thin lava flows, tephra accumulations, pit crater fillings, dikes, and hydrothermal features.

A large part of the walls show uniform piles of thin lava flows, from the surface to the base of the crater wall (Fig. 4a). They are similar to the lava flows which have been observed during historical summit eruptions. They do not have a thick massive core, and their scoriaceous, vesiculated, and brecciated base and top constitute at least half of each flow body. In some areas, accumulations of scoriaceous tephra are associated with cinder cones. On the south–southwestern wall, a particularly large outcrop of reddish tephra is interpreted by Michon et al. (2009a) as the top of a “hidden strombolian pyroclastic cone” which would constitute most of the central cone. As a whole, the products that have built the Central Cone have a low density, a fact that corresponds to the negative gravity anomaly described by Gailler et al. (2009).

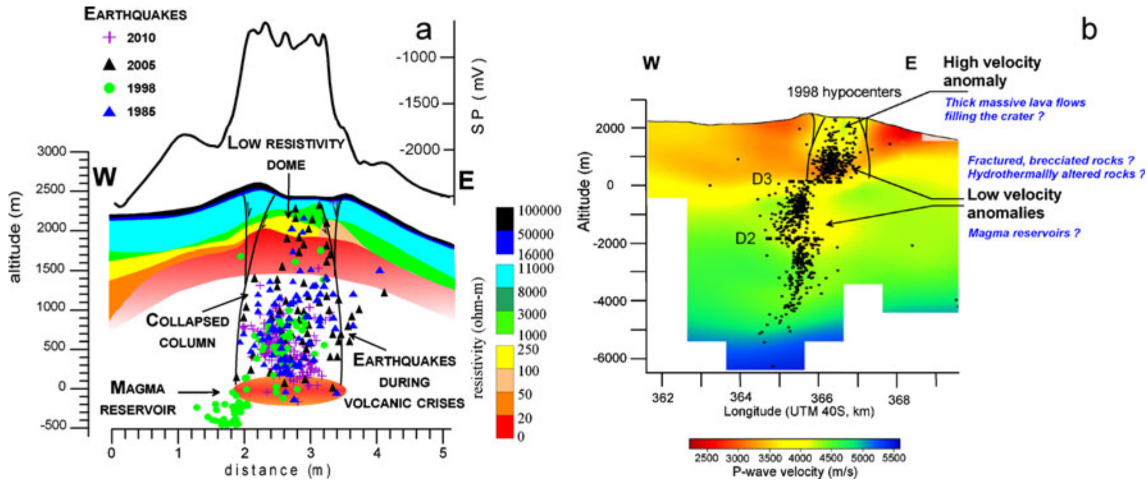

Fig. 3 Structure of the summit zone. **a** Resistivity distribution (adapted from Lénat et al. 2000), SP anomaly (adapted from Malengreau et al. 1994; Levieux 2004) and earthquake hypocenters from several crises between 1985 and 2002 (Piton de la Fournaise Volcanological Observatory data). Hypocenter datasets have been selected in order to illustrate the pre- and post-1998 activity as well as the March 1998 one, while keeping the clarity of the figure. For the same reason, the location errors bars are not shown. They are variable between the old and more recent data, but they do not cast doubt on the fact that the earthquakes are concentrated in the collapse column. **b** W-E section in the P wave tomography from Prôno et al. (2009) showing March 1998 hypocenters. D2 and D3 are low seismicity regions located at the base of the two low-velocity zones (Battaglia 2001)

In two areas, accumulations of thick horizontal lava flows are associated with the filling of pit craters. To the south, a 150–200 m diameter filled crater appears to be a young, possibly historic, structure, because its top is very near the surface. To the west (Fig. 4a, b), a much larger pit crater, with an estimated diameter of about 800 m, has been exposed by the collapse. The minimum depth to its top is about 50 m, and its vertical extent is greater than the 200 m exposed in the wall. A similar pile of horizontal thick lava flows was visible in the middle of the Dolomieu crater in 1936 (Fig. 2 in Lacroix 1939).

Numerous dikes are observed in the walls of the crater. Their usual width is less than 1 m. A halo of hydrothermal alteration is locally associated with some dikes. However, large patches of hydrothermal alteration are also observed in areas much greater than the distance of influence of individual dikes. This is the case, for example, at the outer limits of the pit craters where the permeability contrast between the thick massive lava flows filling the pit craters and the more permeable surroundings seems to act as a drain for hydrothermal fluids. All these observations indicate the influence of an underlying active hydrothermal system, as suggested by the resistivity and SP data (Fig. 3). We believe that the top of this system now crops out low in the new Dolomieu crater. Indeed, a ring of steam and wet soil can be observed in the walls of the crater (Fig. 4). The wet soil is caused by condensation of water in the fumaroles. This ring of fumaroles coincides with the contact between the screes at the base of the crater and the wall of the crater. Staudacher (2010) suggests that the hydrothermal fluids flow along the peripheral fault of the collapse, but we cannot rule out the possibility that they flow upwards over all the surface of the crater and that, near the surface, they are drained by the zone with the highest permeability. Note the good agreement between the elevation of the fumaroles and the top of the low-resistivity dome (Figs. 3a and 4).

The 2007 collapse has led several authors (Michon et al. 2007; Gailler et al. 2009; Peltier et al. 2009c; Staudacher et al. 2009) to realize that, as a consequence of the subsidence, a cylinder of faulted, fractured, and crumbled rocks must exist between the surface and the top of the magma reservoir. This interpretation is well supported by both modeling results and field observations. The analog models from Roche et al. (2001) show that pit crater collapses are caused by the upward propagation of faults. In the case of a roof thickness/roof width ratio of about 2, comparable to the one expected in the case of Piton de la Fournaise, stopping occurs above the emptied reservoir and propagates upwards, thus generating strong fracturing and brecciation within the subsided part.

Geophysical signals accompanying small pit crater collapses at Piton de la Fournaise (Hirn et al. 1991; Longpré et al. 2007), or the larger 2007 Dolomieu collapse (Michon et al. 2007; Peltier et al. 2009c; Staudacher et al. 2009) support such an incremental collapse mechanism where stopping may play a large role. Michon et al. (2009a) suggest that the incremental character of the collapse is influenced by the interplay between the deflation of the edifice, the evolution of the pressure in the magma reservoir, and the gravitational stress exercised by the collapsing column. The apparently progressive collapse of the Dolomieu between about 1929 and 1936 may be

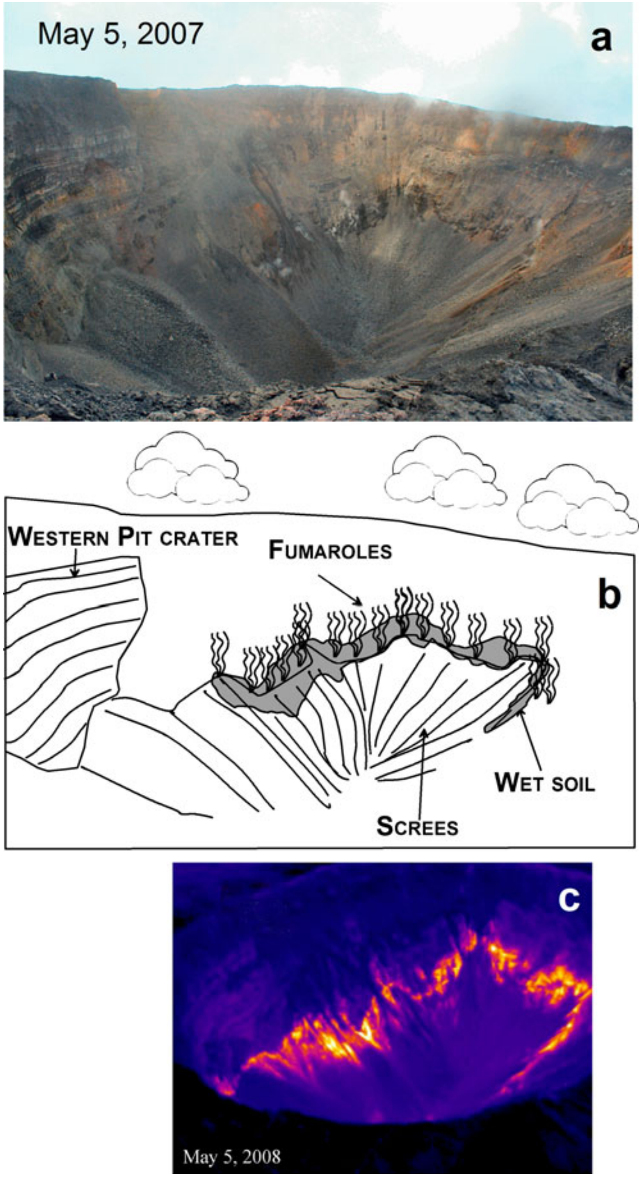

Fig. 4 Evidence of the presence of an underlying hydrothermal system. A ring of steam and wet soil can be observed in the walls of the crater (**a** and **b**). The wet soil is caused by condensation of the water of the fumaroles. The temperatures (Staudacher 2010) range between 95°C (*yellow*) and 40°C (*red*) (**c**). Dark areas correspond to temperatures between 1°C and 5°C

additional evidence for such an incremental mechanism of collapse. Similar phenomena have been observed elsewhere as, for example, at Mount Etna with the progressive opening of the Bocca Nuova crater (Murray 1980) or at Miyakejima volcano with the gradual collapse of a small caldera (Kumagai et al. 2001; Geshi et al. 2002). The pit craters of Masaya volcano (Rymer et al. 1998; Harris 2009) exhibit a dense fault system in the collapsed blocks and clearly show the heterogeneous character of the collapsed column, which comprises different units of lava flow piles, breccias, and screes. Similarly, a brecciated column is often observed in subsided rocks overlying mine or civil engineering underground excavations (e.g., Jeremic 1994). On the base of these results and observations, and taking into account the fact that episodes of collapse are recurrent in the summit area of Piton de la Fournaise, we can confidently postulate that the breccia column beneath the summit craters represents a volume with a significantly lower strength and higher permeability than those of the surrounding rocks. The periodic episodes of collapse prevent compaction or hydrothermal sealing from achieving a significant reduction in the permeability. The fact that the boundary of the Dolomieu crater has remained practically unchanged despite the 340 m collapse of its interior (Michon et al. 2009b) is additional evidence for the existence of a well established fault system limiting the column.

Figure 5 summarizes the main information on the internal structure of the central area of Piton de la Fournaise derived from geophysical and geological observations.

Inter-eruptive behavior of Piton de la Fournaise

The Piton de la Fournaise Volcanological Observatory was established in 1980. Since that date, nearly 60 eruptions and 20 shallow intrusions have been monitored. The seismicity of Piton de la Fournaise since 1980 can be assigned to three zones. One, extending from about −6 km b.s.l. to sea level, located slightly west of the summit, appeared in March 1998 (Battaglia et al. 2005). A second zone exists beneath the eastern flank. It first appeared in July 1985 (Lénat et al. 1989a) and has been activated during some crises since then. This second zone is generally regarded as caused by the tendency of the edifice to move seaward, in the direction of its eastern free flank. The third zone is located between about sea level, or slightly deeper, and the surface beneath the summit craters. This is the most active zone and is systematically activated during both inter-eruptive periods and crisis periods.

The inter-eruptive and eruptive behavior of Piton de la Fournaise has been analyzed by Lénat (1988), Lénat and Bachèlery (1988, 1990) and, more recently, by Peltier (2007) and Peltier et al. (2009a). Two main periods are identified, separated by the 1998 crisis. Before 1998 (Fig. 6a), the precursors of the eruptions and shallow intrusions were mostly limited to between a few days and a few weeks of moderate volcano-tectonic seismicity, with 5–10 volcano-tectonic events per day located in the seismic zone below the summit craters; summit deformation was weak or absent. During the intrusive phase which preceded each eruption by a few tens of minutes to a little less than 3 h, rapid inflation was measured, and tens to hundreds of volcano-tectonic earthquakes were recorded. Most of the inflation remained as a permanent deformation and was

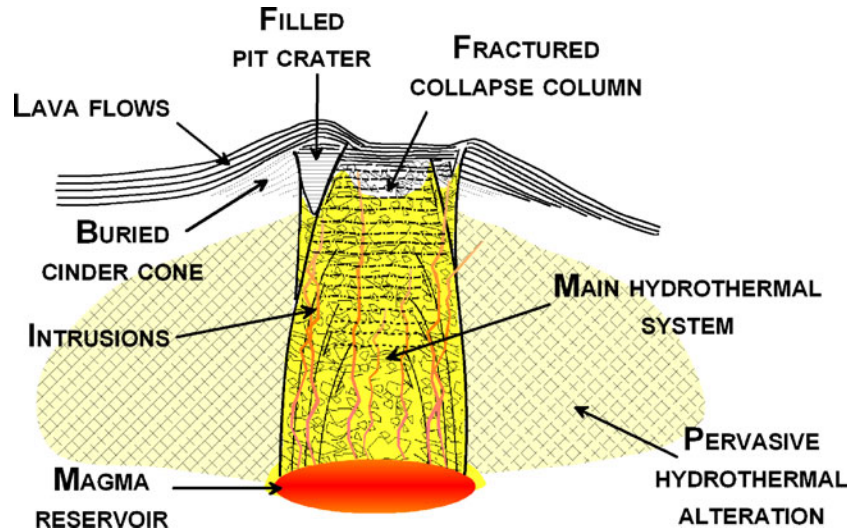

Fig. 5 Interpretative geological scheme. A fractured column is created by the periodic collapses of the summit. Its high permeability enables the circulation of hydrothermal fluids

interpreted as resulting from the emplacement of the feeder dike of the eruption (Zlotnicki et al. 1990). The accompanying seismic activity was also located in the same depth range as the pre-eruptive events ~0.5–2.5 km beneath the summit. No time/depth relationship that could have indicated a migration of the fracturing was observed during any of the crises; although the seismic network allowed only a relatively low accuracy for earthquake location in the first years of the Observatory, systematic upward migrations during crises would have been at least coarsely identified.

No eruption occurred from 1992 to 1998, whereas the mean eruption rate since 1972 had been about two eruptions per year. During this repose period, three noticeable events occurred in 1996–97 (Battaglia et al. 2005). In September 1996, a deep (~16 km) seismic swarm occurred about 10 km north-northwest of the summit. In November 1996, a swarm of shallow events beneath the summit was interpreted as an intrusion of magma (Peltier et al. 2009a). A similar event occurred in August 1997. From 1996, the rate of volcano-tectonic earthquakes above sea

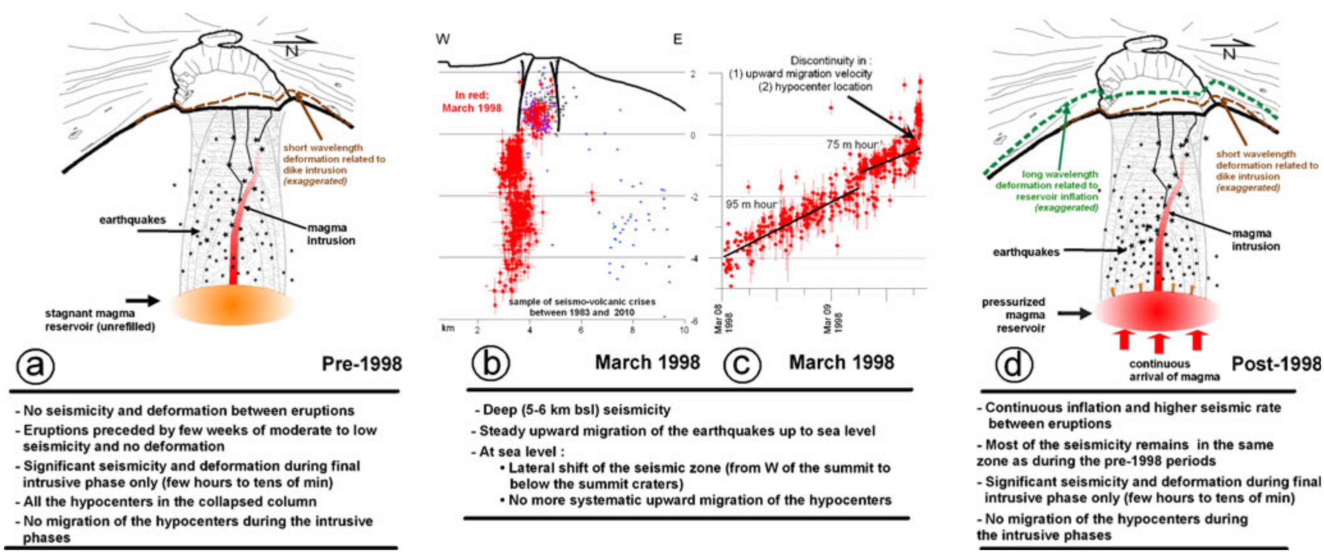

Fig. 6 Summary of the behavior of Piton de la Fournaise since 1980. **a** Sketch summarizing the main characteristics of the volcano before 1998 (based on works by Lénat 1988; Lénat et al. 1989a, b; Lénat and Bachèlery 1990; Sapin et al. 1996 and Volcanological Observatory records; Peltier et al. 2009a). **b** Location of the March 1998 earthquakes (in *red*) (Battaglia 2001; Battaglia et al. 2005). For comparison, some epicenters from 1983 to 2010 are also shown. **c** Depth versus time plot of the hypocenters during the most active period of the March 1998 pre-eruptive crisis (Battaglia 2001; Battaglia et al. 2005). **d** Sketch summarizing the main characteristics of the behavior of the volcano after 1998 (based on works by Longpré et al. 2007; Massin 2009; Peltier et al. 2009a; Fukushima et al. 2010; Rivemale et al. 2010 and Volcanological Observatory records). Note that the relative scales of the long and short wavelength deformation curves are different (the short wavelength deformation related to the intrusion is significantly larger than the long wavelength deformation)

level beneath the summit started to increase; it accelerated during 1998, announcing the March 1998 eruption.

The seismic precursors of the March 1998 eruption were drastically different from those of all previously monitored eruptions, both in location and in the time evolution of their focal depths (Battaglia et al. 2005). The seismic crisis (Fig. 6b) started on March 6 around midnight (Universal Time). Between March 7 and 8, the seismic activity dramatically increased, and an eruption started on March 9. For the first time, a significant seismicity was observed beneath the western part of the central zone from depths ranging 5–6 km b.s.l. up to sea level. The events clearly migrated upward at about 25 mh^{-1} on March 6 and 7, then at 75–95 mh^{-1} on March 8 and 9 (Battaglia et al. 2005) (Fig. 6c). When the swarm reached sea level, the upward migration of the hypocenters stopped, the seismicity shifted eastward beneath the summit, and the events spread randomly between sea level and shallow depths (Fig. 6b, c). The eruption was not preceded by significant deformation, except during the shallow intrusive phase for about 1 h before the beginning of the eruption (Battaglia and Bachèlery 2003).

The clear migration of the seismicity, from 5–6 km b.s.l. to sea level was interpreted as a transfer of magma from a reservoir at 5–6 km b.s.l. (Battaglia et al. 2005). This new magma may have fed the shallow reservoir inferred to exist near sea level beneath the summit. The March 1998 eruption was noticeably long (196 days) and emitted a significantly larger volume of lava (60×10^6 m^3) than average eruptions at Piton de la Fournaise. It was clearly the start of a new and different period of activity, which would in fact become discernible from 2000 onward (Peltier et al. 2009a).

The main changes in behavior of the volcano between the pre- and post-1998 epochs are in the deformation regime and in increase of the level of seismicity between the eruptions (Fig. 6d). The eruptions following the March 1998 eruption were preceded by months of inflation, which became continuous from about 2000. The long-term inflation is compatible with a pressurized source beneath the summit at about sea level (Peltier et al. 2009a). As during the pre-1998 period, most of the seismicity occurred within the same volume, beneath the summit craters (Massin 2009; Rivemale et al. 2010). However, unlike before 1998, some earthquakes are located at 5–6 km b.s.l. (e.g., Apr–Oct 2005, Apr 2007). A few earthquakes also occurred on the eastern flank during some crises. The new deformation and seismic pattern suggests that the dynamics of the storage and transfer of magma in the shallow reservoir has changed following the 1998 events. As suggested by Peltier et al. (2009a), a continuous supply of magma to the shallow reservoir can explain the inflation pattern. In this scheme, the deep intrusion of 1998 opened a conduit between a deep (5–6 km b.s.l.) reservoir and the shallow one. The latter was apparently drained during the April 2007 eruption (Peltier et al. 2009c; Staudacher et al. 2009; Villemant et al. 2009), as attested by the huge volume of lava flows (220×10^6 m^3) associated with a high effusion rate (~50 m^3 s^{-1}), the high percentage of large olivine phenocrysts in the lava flows at the end of the eruption, and the presence of blocks of hydrothermally altered rocks from the edifice carried by the lava flows when the Dolomieu crater collapsed. However, we note a remarkable similarity of the pre-eruptive patterns with those from the pre-1998 period (Longpré et al. 2007; Massin 2009; Peltier et al. 2009a; Rivemale et al. 2010).

The role of the fractured collapse column in the volcano-tectonic, hydrothermal and volcanic activity

The fractured column constitutes a major heterogeneity within the edifice. It is a long-lived structure (at least at the scale of a few centuries) which is kept alive by periodic episodes of collapse at different scales, from small pit craters to the entire column. An obvious spatial correlation exists between this structure and the locations of (1) most of the seismic activity, (2) a low-resistivity dome, (3) a huge SP anomaly, (4) thermal evidence of hydrothermal activity, and (5) the early location of the magma intrusions.

Collombet et al. (2003) first called attention to the stable location of the seismogenic volume beneath the summit since the beginning of monitoring, but they only noted that it was regarded as “the main path for the magma flow toward the surface” and did not discuss further its structural significance. On the other hand, Grasso and Zaliapin (2004) and Traversa and Grasso (2009) suggest that the location of the volcanic seismicity does not reflect directly that of the magma intrusion tip or path, but rather a diffuse release of the background stresses in the edifice under the effect of the perturbation caused by the injection of magma. Similar observations have been reported by Rubin et al. (1998) at Kilauea, Pedersen et al. (2007) in Iceland and by Toda et al. (2002) in the Izu islands. This can account for the observed lack of apparent migration of the earthquakes between about sea level and the surface during upward magma propagation at Piton de la Fournaise. However, it cannot explain why the seismicity remains concentrated in the collapsed column and why a clear migration (Fig. 6b) of the seismicity was observed in March 1998 between about 5 km b.s.l. and sea level, but not above sea level, when it shifted eastward to the column (Battaglia 2001; Battaglia et al. 2005). In this latter structure, the upward migration pattern disappears or the upward migration velocity becomes unrealistically rapid, inconsistent with the ground deformation velocity associated with dyke propagation.

The shallow seismicity, mainly above sea level, is always "trapped" in the column, the lower part of which corresponds to a low-velocity zone (Fig. 3b). Without the fractured collapse column, one would expect to observe a more diffuse seismicity within the cone, even if it remained denser in the central area. On the contrary, however, the areas around the column are virtually aseismic. Therefore, it can be argued that this structure is a dominant attractor for the seismic activity or acts as a stress guide. The abrupt shift of the March 1998 seismic crisis from the near vertical seismic path west of the summit to the column (Fig. 6b) is a striking illustration of this role. Recent studies (Rivemale et al. 2010; Battaglia et al. 2011a, b; Massin et al. 2011) suggest the presence of a network of small fractures (with lateral dimensions of tens of meters) in the lower part of the fractured column, some reactivated during successive crises. They are deduced from the presence of clusters of similar micro-earthquakes, which represent only a portion of the recorded seismicity. Note also that the zone below the western margin of the Dolomieu crater is the most seismogenic one in the seismic volume. This could indicate that stresses and/or fluid circulation (hydrothermal and/or magmatic) tend to concentrate along these pre-existing fractures.

Although the depth of investigation of the available electrical resistivity soundings in the summit area (Lénat et al. 2000) does not extend to more than about 1–1.5 km in depth, the resistivity values, the SP anomaly (Malengreau et al. 1994; Michel and Zlotnicki 1998) and the thermal anomalies (Staudacher 2010) in the crater show the presence of strong hydrothermal activity, spatially correlated with the collapse structure of the summit. Before the 2007 collapse, the hydrothermal activity could not be observed at the surface except for short-lived fumaroles associated with the cooling of recent eruptive vents. This unusual lack of surface hydrothermal activity on such a highly active volcano may be caused by the huge flux of meteoric water (rainfall rate >10 my^{-1}) that percolates into the edifice (Join et al. 2005), possibly buffering the hydrothermal flux at depth or by hydrothermal sealing. The mechanism of sealing is probably not important because the exposures in the walls of the new Dolomieu crater do not show zones of extreme hydrothermal alteration capable of forming impermeable barriers; nonetheless, the presence of such zones at depth beneath the Dolomieu crater cannot be ruled out.

The close association of hydrothermal activity evidence with the Dolomieu crater shows that the fractured column acts as a drain for hydrothermal circulation. This spatial concentration of the main hydrothermal activity of Piton de la Fournaise results from the combination of magmatic heat sources at depth, a large amount of fluids from magmatic and meteoric origin, and mainly the presence of a structure (the fractured column) with a higher permeability than that of the surrounding edifice.

Modeling of ground displacements caused by dikes feeding the eruptions (e.g., Peltier et al. 2005 and references therein; e.g., Fukushima et al. 2010) shows that all the intrusions start at depth and rise vertically in the fractured column. At shallower depth, they can migrate laterally, probably because they encounter a level of neutral buoyancy (Peltier et al. 2009a; Fukushima et al. 2010). The column therefore also acts as a guide for the intrusions. The reason for the deeper part of the intrusion starting in the fractured column may be that it is easier for magma to intrude the broken, more vertically fractured rocks of the column or because magma is captured at the boundary between the column and the surrounding rocks because of the contrast in rock stiffness.

Possible role of the hydrothermal system in eruption triggering

Since 1998, the inter-eruptive and pre-eruptive behavior of Piton de la Fournaise resembles that of other basaltic volcanoes such as Kilauea (Wolfe et al. 1987), Mauna Loa (Lockwood et al. 1987), and Krafla (1975–1984) (Sturkell et al. 2006), with inter-eruptive slow inflation and moderate seismic activity (up to 100 earthquakes per day), and with intrusive episodes accompanied by rapid deformation and intense seismicity. This type of behavior is usually interpreted in terms of magma transfers. The inter-eruptive inflation corresponds to the pressurization of a reservoir by the arrival of magma from deeper levels and/or by increase of gas pressure, and then the intrusive phase starts when the wall of the reservoir is fractured, allowing the upward or lateral migration of magma. This model is, however, difficult to apply to the period preceding 1998, because no evidence for the pressurization of a reservoir was observed (Fig. 6a). Considering the inferred structure of the volcano, with a magma reservoir surmounted by a hydrothermal system in a permeable column, we can examine the possible interactions between these units and their effects on the volcanic activity for both the pre- and post-1998 periods.

It is beyond the aim of this work to derive a quantitative model of the interactions between the magmatic and hydrothermal systems. Because the exact geological structure, the physical properties of the units, and the thermal and fluid fluxes are not accurately known, this would require exploration of various conditions. Existing numerical models such as HYDROTHERM (Hayba and Ingebritsen 1994) or TOUGH2 (Pruess et al. 1999) can be used to investigate different scenarios. A qualitative model can, nevertheless, be inferred for the interactions between the magmatic and hydrothermal systems (Fig. 7). Near sea level, the magma reservoir complex, whatever its shape or exact vertical position, is in contact with the hydrothermal system. The fact that the latest 2007 lava flows, thought to have drained the reservoir, transported many xenoliths of hydrothermally altered blocks can be direct evidence of this contact. The hydrothermal system is therefore heated by the reservoir (Fig. 7a) and, to a lesser extent, by the cooling of nearby recent intrusions. Reversely, a flux of cold meteoric water moves downward in the edifice. Close to the magmatic heat source near sea level, super heated liquids or supercritical (374°C and 22 MPa for water) conditions can exist. The heated fluids increase pore-fluid pressure in the permeable column. Depending on fissure sealing by precipitation and rock resistance, higher-than hydrostatic pore pressures can occur and induce hydraulic fracturing or nucleation of earthquakes in the deeper parts of the hydrothermal system. Alternatively, transient-localized decompression produced by water boiling and degassing near active fractures can bring the fluids near-spinodal conditions (Thiéry and Mercury 2009) and induce explosive vaporization. Published simulations of geothermal or volcanic systems (e.g., Yano and Ishido 1998; Hurwitz et al. 2003; Reid 2004; Croucher and O'Sullivan 2008; Pashkevich and Taskin 2010) show that these mechanisms are plausible for the case of Piton de la Fournaise. The studies generally stress the necessity to have a medium with a high permeability between the heat source and the fluids' plume to enhance the efficiency of the system. These mechanisms can provide an alternative explanation for the background seismicity observed in the fractured column during the inter-eruptive periods, especially in its lower part. At some threshold, the thermal perturbation of the hydrothermal system may evolve to a sustained fracturing crisis. This hydrothermal instability may affect the magmatic system and possibly trigger a magma intrusion as shown on Fig. 7c. How such an interaction can initiate a magmatic crisis is not well established. One potential

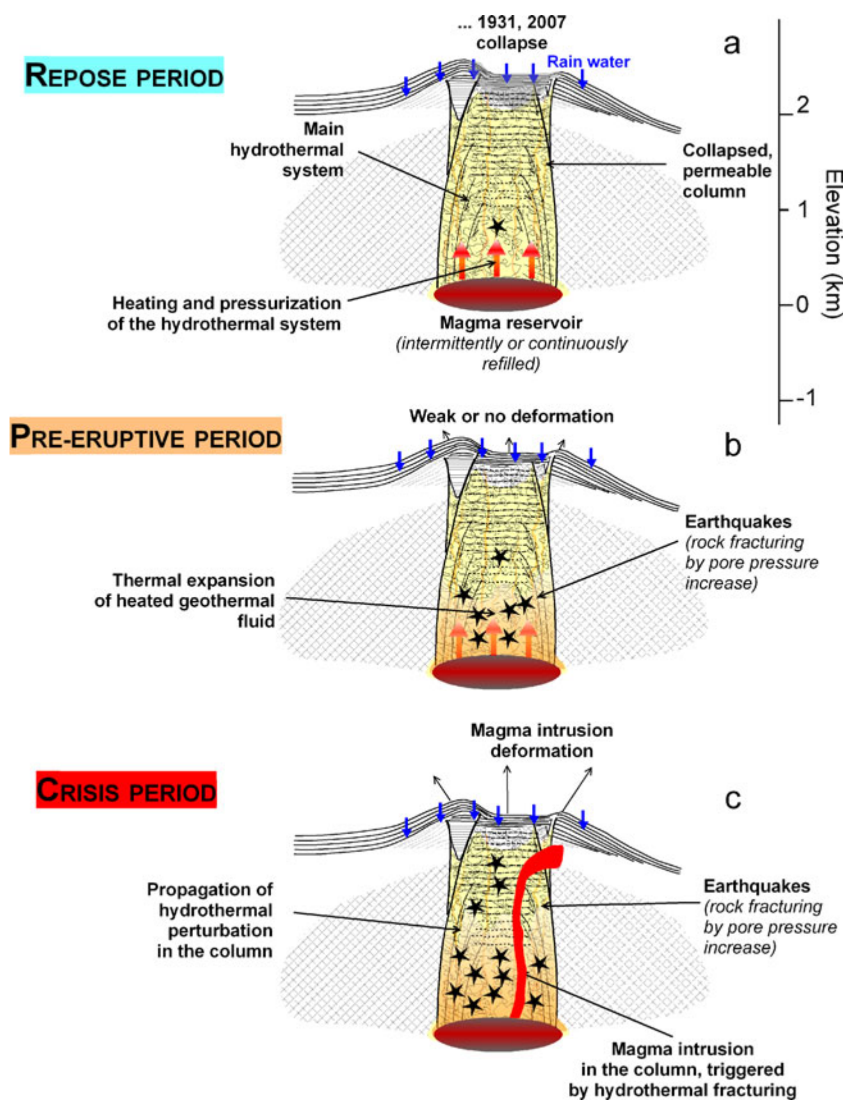

Fig. 7 Sketches illustrating the possible role of the hydrothermal system in eruption triggering. **a** Repose period. The hydrothermal system is heated by the magma reservoir and by the cooling recent intrusions (*light-red broken lines* in the collapse column). **b** Pre-eruptive period. The high temperature front moves upward as more heat is absorbed. Pore-fluid pressure increase leads to rock fracturing. **c** Crisis period. The thermal perturbation of the hydrothermal system evolves to a sustained fracturing crisis. Hydrothermal fracturing cause transient decompressions on the wall of the reservoir, thus triggering vesiculation of the magma decompression and starting the intrusion process. Magma intrudes first upward in the column and may later move laterally

mechanism is that the hydrothermal fracturing can cause transient decompressions at the wall of the reservoir, leading to local zones of vesiculation which will act as nuclei for the development of a magma intrusion. This would be in agreement with the observations made by Battaglia et al. (2011a) who note that clusters of similar micro-earthquakes near sea level are systematically active during recent pre-eruptive crises and not during the intrusive phases, suggesting that "a significant part of pre-eruptive seismicity is related to rupturing a specific interface to allow dike propagation rather than to the propagation itself". The periodicity of the crises, if examined in the framework of this geothermal model, might be compared with geyser activity, in that it requires a recharge phase between the crises. Several processes can collaboratively contribute to bring back the geothermal system in a rest or recharge phase. During the crises, the thermal energy tends to be spread in a larger volume when the hot fluids propagate upward in the column and convection can carry cold surface meteoric water to depth. The contact surface between the hydrothermal system and the magma can also be disturbed. It is obvious, however, that future quantitative modeling and monitoring data are necessary to refine the hydrothermal hypothesis.

The above model provides an alternative interpretation for the seismicity observed during pre-eruptive and intrusive crises at Piton de la Fournaise. If the behavior of Piton de la Fournaise is analyzed in the framework of this model, the following interpretation can be proposed. Before 1998, when the magma reservoir was not recharged and, therefore, had a lower heating power, the phase corresponding to Fig. 7a was seismically quiet, because the hydrothermal system was only heated slowly. When the hydrothermal system reached a critical thermal state, a cascade of explosive depressurizations (hydraulic fracturing and/or boiling-induced decompressions) would initiate a magma intrusion. With the continuous recharge of the reservoir following the 1998 events, the heating power has been significantly increased. As a result, the hydrothermal system is more active and generates more or less continuous fracturing in the column. Accordingly, the eruptions became more frequent after 1998 (Peltier et al. 2009a). We also note that, on the basis of petrological and geochemical observations (Boivin and Bachèlery 2009; Peltier et al. 2009a; Vlastélic et al. 2009), an episode of recharge of the reservoir is inferred to have occurred in 1986. Following this event, the frequency of the eruptions also increased for several years.

Discussion and conclusions

In our synthetesis and analysis of the behavior of Piton de la Fournaise, two main issues have been raised: (1) the presence of a major heterogeneity below the summit area and (2) the role of the interactions between the magmatic and the hydrothermal systems in the volcanic crises. The first issue leads to a well-supported new structural model of the central area of the volcano that implies specific characteristics for the volcanic activity of Piton de la Fournaise. The second issue gives rise to a more speculative hypothesis, suggesting that the hydrothermal system could be primarily responsible for the initiation of the volcanic crises.

Our new analysis leads us to consider that the seismic, hydrothermal, and intrusive activities are concentrated in the volume inferred to be a column of rock fractured and brecciated by repeated collapse; this provides a new framework in which to interpret the activity of Piton de la Fournaise. The dominant factors that make this major structural heterogeneity a trap are probably its higher permeability and its lower mechanical strength. Future work should be devoted to characterizing more accurately the geometry and the physical properties of the collapsed column as well as investigating its role in the evolution on the volcanic edifice. For example, the interpretation of seismic tomographies has to take into account the presence of this collapse column, and electrical resistivity surveys with a deeper depth of investigation than the presently available ones might resolve the root of the inferred main hydrothermal zone. Similarly, the models used to interpret the deformation should incorporate this major mechanical heterogeneity.

The stability of the activated zone above sea level and the similarity of the pre-eruptive crises before and after 1998, suggest a common triggering mechanism for all the eruptions. This mechanism could be purely magmatic, resulting from the pressurization of a reservoir, but we propose that the hydrothermal system may also play a major role in the development of volcanic instabilities beneath the summit area. The contribution of the hydrothermal systems to the seismicity and deformation is increasingly scrutinized for various types of volcanoes (Bonafede 1991; De Natale et al. 2001; Waite and Smith 2002; Chiodini et al. 2003; Hill et al. 2003; Howle et al. 2003; Nakaboh et al. 2003; Todesco et al. 2004; Battaglia et al. 2006; Ukawa et al. 2006; Hole et al. 2007; Hurwitz et al. 2007; Saccorotti et al. 2007; Cusano et al. 2008; Gambino and Guglielmino 2008; Peltier et al. 2009b), but it has not yet been suggested that it plays a role in triggering basaltic eruptions, as we suggest here. This hypothesis has to be further evaluated in terms of the observed signals in order to estimate its consistency, but the idea can also stimulate complementary observations.

An apparent weak point for the hydrothermal model at Piton de la Fournaise is the lack of typical hydrothermal seismicity, which usually includes long-period (LP) signals (e.g., Chouet 1996; Matoza and Chouet 2010). Because our model implies intense fluid activity (circulation, thermal expansion) in the fractured column, LP events would be expected, along with the VT events. For comparison, a large hydrothermal crisis in 1991 at the basaltic Karthala volcano (Savin et al. 2005) was accompanied by both VT and LP events. At Piton de la Fournaise, however, LP events are not frequent (Aki and Ferrazzini 2000; Battaglia et al. 2005). Aki and Ferrazzini (2000) noted that most of the LP events are located beneath the summit at about sea level. However, recent and ongoing seismic studies at Piton de la Fournaise (Zecevic et al. 2009; F. Brenguier and L. De Barros, pers. comm.) show the presence of LP events preceding recent eruptions. Preliminary analyses indicate that they are shallow (a few hundreds of meters) and have a relatively high dominant frequency (2–3 Hz) for LP events. A possible explanation for the apparent scarcity of LP events in the fractured column might be that it is mechanically weak and would therefore be a poor resonator.

The hydrothermal model assumes that eruptions are preceded by a crisis, or intensified activity, in the hydrothermal system that could possibly be detected at the surface as an increase of fumarolic temperature or flux. However, as described above, no hydrothermal activity was observed at the surface before the 2007 collapse of the Dolomieu crater. Now, fumaroles exist at depth in the crater (Staudacher 2010) (Fig. 4), issuing from the top of the hydrothermal system. An example of enhanced fumarolic activity reported by the Piton de la Fournaise Volcanological Observatory in October 2009, about 2 weeks before the November eruption (http://www.undervolc.fr/RapportActivite2009-ThS_20091029-10h44.pdf), when the VT activity in the fractured column was sustained, could be evidence of a perturbation of the geothermal system preceding eruption.

Deformation of the summit area would be expected when the hydrothermal system is heated because expansion of hydrothermal fluids exerts a mechanical pressure on their surroundings (Thiéry and Mercury 2009). The quantitative models proposed for the Phlegrean Fields or other calderas (e.g., Bonafede 1991; De Natale et al. 2001; e.g., Battaglia et al. 2006; Hurwitz et al. 2007) show that increased temperature and pressure in the hydrothermal system will cause inflation. At Piton de la Fournaise prior to 1998, no long-term (weeks to months) inflation of the summit preceded the eruptions, but since 1998, continuous inflation of the central area has been observed (Peltier et al. 2009a). If we infer a common hydrothermal process for triggering all eruptions, we have to reconcile these apparently contradictory observations. The structural setting of the hydrothermal system of Piton de la Fournaise appears to be specific, with the relatively permeable column in a less permeable surrounding. The hydrothermal system is thus more or less impounded laterally but may have a free surface, so that a volume increase of the hot fluids is accommodated by oscillations in upper level of the hydrothermal system. Such oscillations in a geothermal system have been shown, for example, by changes in elevation of an acid water lake inside the crater of Karthala volcano (Savin et al. 2005). In such a situation, practically no deformation would be generated by perturbation of the geothermal system, and the deformation observed since 1998 could be attributed only to pressurization of the magma reservoir. The lack of large hydrothermal explosions during the 2007 collapse also suggests that the hydrothermal system was not sealed and therefore pressurized. The rupture of a sealed system during the collapse would probably have resulted in vigorous phreatic explosions. However, Gouhier and Coppola (2011) suggest that a large release of SO_2 from the Dolomieu crater coincided with the collapse of the summit in 2007.

In summary, our work draws attention to two features of Piton de la Fournaise, a mechanical heterogeneity and a hydrothermal system, and analyzes the possible interactions between these structures and the volcanic activity. As a first attempt to integrate these elements in a unified framework, it raises several issues that require further study by experts in different fields to refine or invalidate the new interpretations of the crises at Piton de la Fournaise. We think that the ideas expressed here can contribute to improving the interpretation of the geophysical data recorded at Piton de La Fournaise and at similar volcanoes, and they possibly can be adapted by monitoring networks to allow for the specific surveillance of the hydrothermal systems.

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