Morphology of Piton de la Fournaise basaltic shield volcano (La Réunion Island): Characterization and implication in the volcano evolution

Laurent Michon, Francky Saint-Ange

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 113, B03203, doi:10.1029/2005JB004118, 2008

Morphology of Piton de la Fournaise basaltic shield volcano (La Réunion Island): Characterization and implication in the volcano evolution

Laurent Michon[1] and Francky Saint-Ange[1]

Received 24 October 2005; revised 24 May 2007; accepted 13 December 2007; published 5 March 2008.

[1] The topography of Piton de la Fournaise volcano (PdF) differs from the classic view of basaltic shield volcanoes as it is characterized by (1) several steep slope zones on its flanks and (2) a large U-shaped caldera, the Enclos-Grand Brûlé structure (EGBS). Most of these structures were previously interpreted as the scars of lateral landslides, the deposits of which cover the submarine flanks of PdF. We carried out a detailed analysis of the morphology of PdF, which reveals that the steep slope zones form two independent, circumferential structures that continue into the caldera. The development of circumferential steep slopes on volcano flanks may have several origins: constructive, destructive, and deformation processes. We interpret those processes acting on PdF as caused by the spreading of the volcanic edifice above a weak hydrothermal core, leading to outward displacements and a summit extensive stress field. The continuity of the steep slope on both sides of the EGBS escarpments suggests that this structure was not caused by a 4.5 ka old giant landslide as it is usually proposed but is due to a mainly vertical collapse. The recent debris avalanche deposits east of the island indicate that this event likely destabilized part of the submarine flank. We propose that the collapse of the Grand Brûlé, the lower half of the EGBS, was due to the downward drag related to the dense intrusive complex of the Alizés volcano, which is located 1 km below the Grand Brûlé. The collapse of the Enclos is interpreted as the consequence of the deformation of the hydrothermal system of the pre-Enclos volcano. Although the continuity of the geological and morphological structures between the Enclos and the Grand Brûlé suggests a narrow link between these two collapse events, their chronology and relationship are still uncertain. Finally, we hypothesize that the persistence of the NE and SE rift zones during the last 150 ka, despite the large changes of the topography related to the recurrent flank destabilizations, is linked to a deep sources, which can be either underlying crustal faults or the continuous downward subsidence of the Alizés intrusive complex.

Citation: Michon, L., and F. Saint-Ange (2008), Morphology of Piton de la Fournaise basaltic shield volcano (La Réunion Island): Characterization and implication in the volcano evolution, *J. Geophys. Res.*, *113*, B03203, doi:10.1029/2005JB004118.

1. Introduction

[2] The morphology of volcanoes results from construction, destruction and deformation processes that interact during their evolution [e.g., *Moore*, 1964; *Moore and Mark*, 1992; *Rowland and Garbeil*, 2000; *Merle and Borgia*, 1996; *Cecchi et al.*, 2005]. Hence the analysis of the morphology allows the identification of structures the development of which is related to internal processes and/or a specific eruption history [e.g., *Tort and Finizola*, 2005]. At Piton de la Fournaise, one of the world's most active shield volcanoes [e.g., *Lénat and Bachèlery*, 1987], the succession of construction and dismantling phases (i.e., erosion and landslide) led to the development of a complex morphology [e.g., *Bachèlery*, 1981; *Rowland and Garbeil*, 2000]. One of the most striking features is the E-W elongated horseshoe-shaped collapse structure in which the currently active cone developed (Figure 1). The structure is composed, from west to east, by the Enclos depression, the Grandes Pentes, and the Grand Brûlé and is bounded by 100- to 200-m-high escarpments, the Bois Blanc, Bellecombe and Tremblet escarpments in the north, west and south, respectively. The formation of the Enclos-Grand Brûlé structure (EGBS) is one the greatest scientific controversies on PdF. One interpretation is that the Enclos depression results from a polyphase caldera collapse whereas the Grand Brûlé exhibits the scars of lateral landslides with the Grandes Pentes as their headwall [*Bachèlery*, 1981] (Figure 2a). Another interpretation is that the entire EGBS results from a giant landslide that was also partly responsible for the debris

[1]Laboratoire GéoSciences Réunion, Institut de Physique du Globe de Paris, Université de la Réunion, CNRS, UMR 7154, Géologie des Systèmes Volcaniques, Saint Denis, France.

0148-0227/08/2005JB004118$09.00

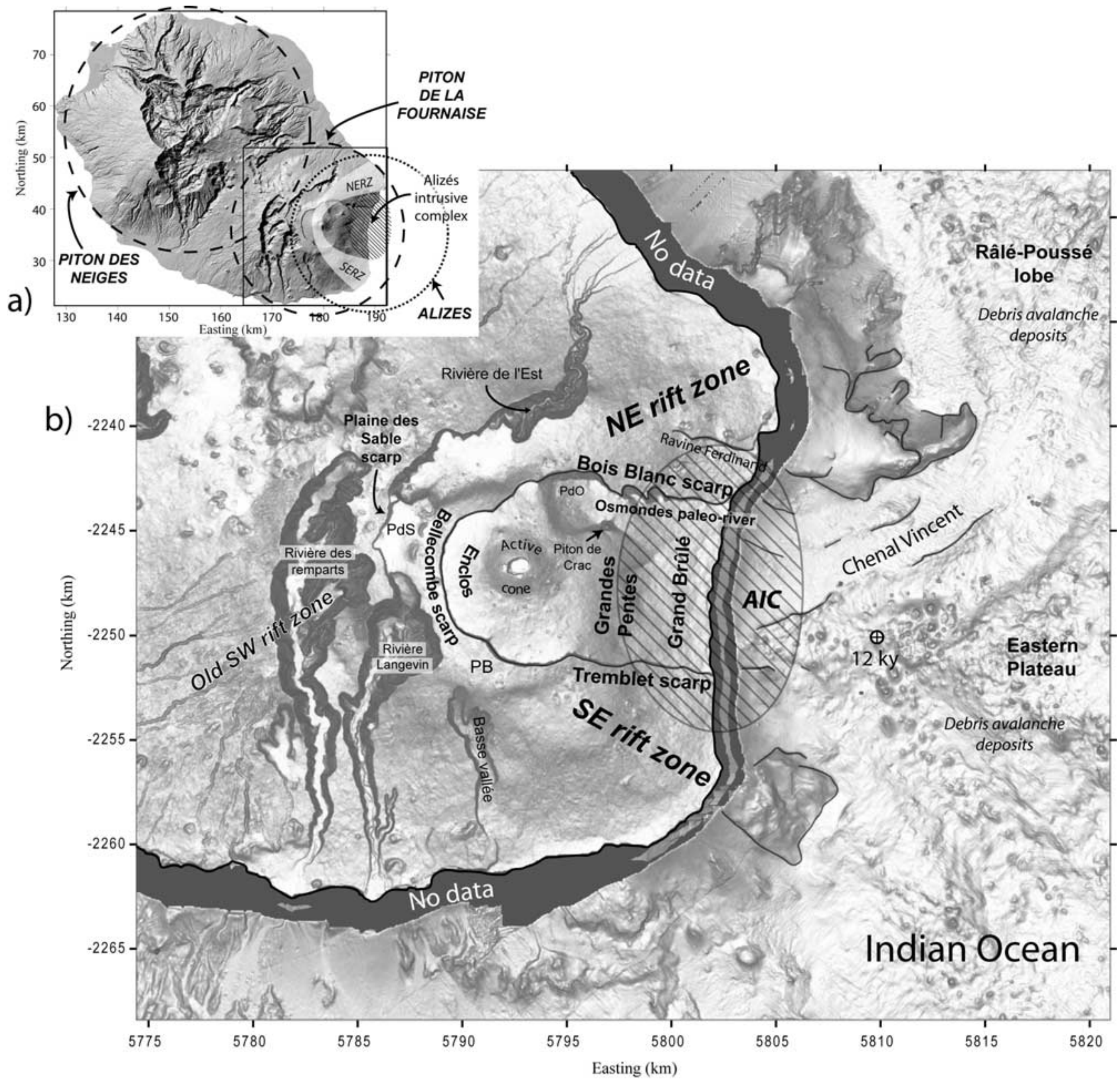

Figure 1. (a) Shaded relief image (illumination from the NW) presenting Piton de la Fournaise in the setting of La Réunion (NERZ, northeast rift zone; SERZ, southeast rift zone). Coordinates in Gauss Laborde Réunion. (b) Digital terrain model of the subaerial and submarine parts of Piton de la Fournaise showing the structures discussed in the text (i.e., scarps, valleys, rift zones, debris avalanche deposits). The dating at 12 ka of a sample dredged on the proximal submarine flank is also reported [*Labazuy*, 1996]. AIC, Alizés intrusive complex; PB, Planèze du Baril; PdO, Plaine des Osmondes; PdS, Plaine des Sables. Coordinates are in Mercator.

avalanche deposits on the submarine flanks [*Duffield et al.*, 1982; *Gillot et al.*, 1994; *Labazuy*, 1996] (Figure 2b). A third interpretation was recently proposed by *Merle and Lénat* [2003], who take into account the role of both hydrothermal systems and deep décollement levels in the volcano's deformation. In their model, a lateral movement (the Grand Brûlé) triggered a vertical collapse in the summit area (the Enclos depression; Figure 2c). Several additional large-scale structures may be related to the same mechanism, resulting in 100-m-high escarpments and steep slope zones outside the structures mentioned above [e.g., *Bachèlery*, 1981; *Oehler et al.*, 2004].

[3] The present study aims at describing in detail the morphology of PdF and at understanding the origin of the main structures, i.e., the steep slope zones on the volcano flanks and the EGBS. We use different digital elevation models (with 25- and 50-m resolution) to determine the detailed and first-order morphologies of the edifice. Integration of the available geological and geophysical data, and of the different published conceptual models, allows us

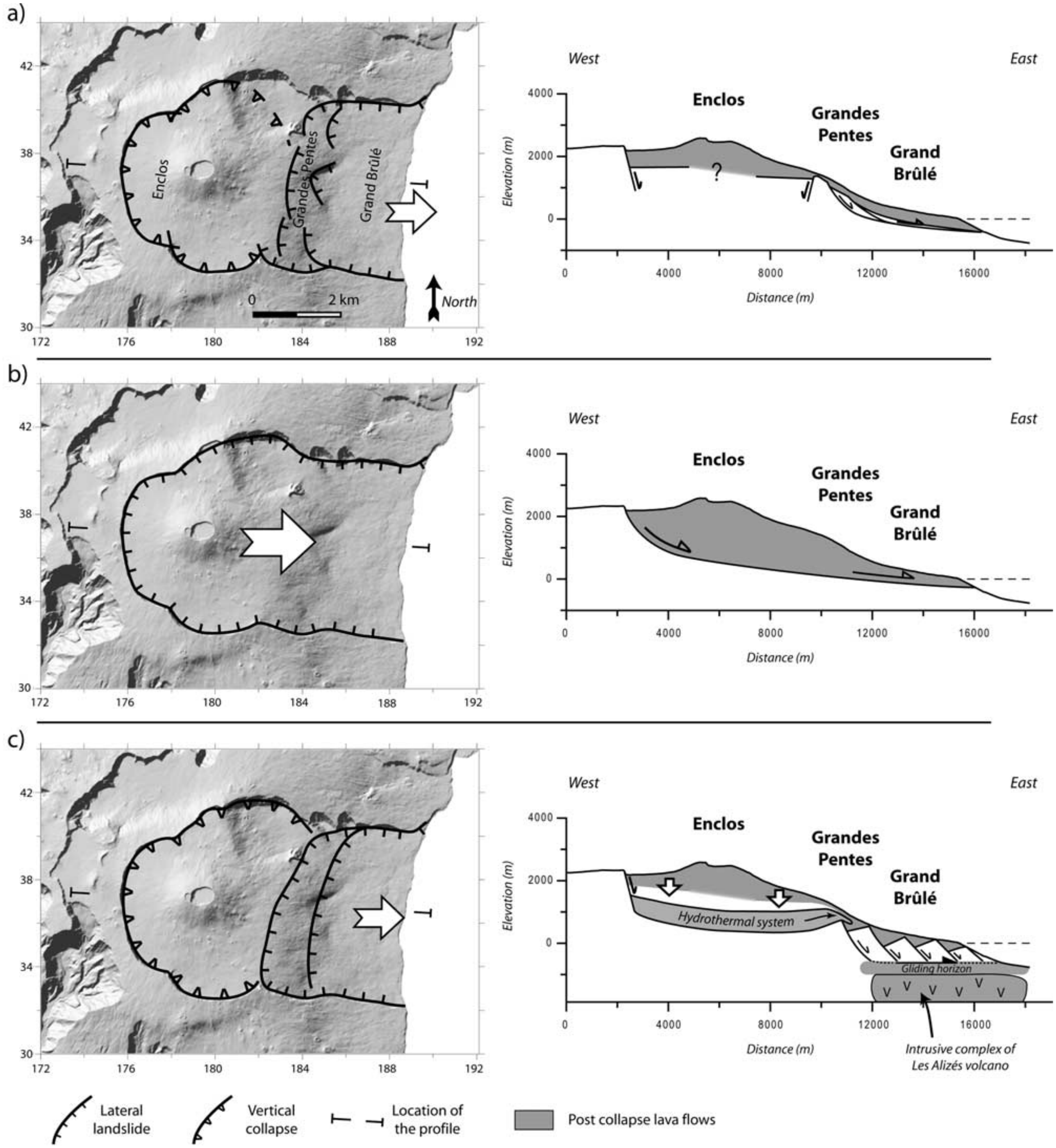

Figure 2. Schematic views and cross sections of the three different models previously proposed to explain the formation of the large Enclos-Grand Brûlé structure (EGBS). (a) A polyphase caldera collapse and the Grand Brûlé is the scar of one or several lateral landslides, the Grandes Pentes being the headwall of these slides [after *Bachèlery*, 1981]. (b) The entire structure is the scar of a single, landslide [after *Labazuy*, 1996; *Oehler et al.*, 2004]. (c) A slide of the Grand Brûlé due to a deep décollement induced a collapse of the Enclos caldera [after *Merle and Lénat*, 2003]. Coordinates are in Gauss Laborde Réunion (km).

to constrain their spatiotemporal relationship and their potential origin. Finally, we present a chronology for the last 150 ka of PdF in which we consider the potential role of the rift zones and the intraedifice density contrasts in the volcano evolution.

2. Evolution of Piton de la Fournaise in the Geological Setting of La Réunion Island

[4] Measured from the seafloor, La Réunion Island is a 7-km-high oceanic shield volcano with a diameter of 220–240 km. Considering the historical magma production rate and the oldest dated subaerial basalts (2.1 Ma [*MacDougall*, 1971]), an age of around 5 Ma was estimated since the beginning of the edifice growth [*Gillot et al.*, 1994]. The initial magmatic evolution was characterized by the development of two adjacent volcanoes (the Piton des Neiges and Alizés), which encountered recurrent flank destabilizations [*Lénat et al.*, 2001; *Bachèlery et al.*, 2003; *Oehler et al.*, 2004]. The Alizés volcano is now completely dismantled and the only evidence of its past existence is the large intrusion complex discovered by drilling below the Grand Brûlé (Figure 1) [*Rançon et al.*, 1989], the old submarine remnants [*Labazuy*, 1996] and the pre-Brunhes reverse magnetic anomalies, which were determined as magmatic formations older than those of Piton de la Fournaise [*Lénat et al.*, 2001]. The destabilization of the Alizés volcano led to the development of the Eastern Plateau, which corresponds to a submarine relief composed of several hundreds of km^3 of debris avalanche deposits [*Labazuy*, 1991; *Oehler*, 2005, available at http://tel.archivesouvertes.fr/tel-00010498/en/]. Around 530 ka ago, Piton de la Fournaise appeared west of the center of Alizés, which had stopped its activity. Between 530 and 12 ka (the date of Piton des Neiges' last eruption [*Deniel et al.*, 1992]), Piton des Neiges and Piton de la Fournaise showed contemporaneous activity. Finally, for the last 12 ka, eruptions are restricted to Piton de la Fournaise.

[5] Geochronological [*Gillot and Nativel*, 1989] and geological data [*Mairine and Bachèlery*, 1997] indicate that PdF results from two main building phases, 0.53–0.29 Ma and 0.15 Ma to present-day, separated by a period during which erosion prevailed. The first construction period led to the formation of the "ancient" PdF, which was likely centered on the present-day Plaine des Sables [*Bachèlery and Mairine*, 1990]. The distribution of dike swarms in the Rivière des Remparts indicates that at least one rift zone developed during this first phase (i.e., the SW rift zone [*Mairine and Bachèlery*, 1997]). Debris flow units in the eastern scarp of the Rivière des Remparts and western scarp of the Rivière Langevin intercalated with 0.22 Ma old lava flows indicate that the "ancient" Fournaise suffered intense erosion, which induced the incision of the paleo-Rivière des Remparts [*Bachèlery and Mairine*, 1990]. This valley was progressively filled in until the collapse of the Morne Langevin caldera 0.15 Ma ago. This event led to an eastward shift of the volcanic center. As a result, the geometry of the feeding zone changed: two NE and SE rift zones developed from the volcanic center [*Bachèlery*, 1981] (Figure 1). Part of the debris avalanche deposits covering the Eastern Plateau and forming the Ralé Poussé lobe are interpreted as related to this large collapse [*Oehler*, 2005]. During the last 0.15 Ma, the "recent" PdF was affected by at least two caldera collapses whose origins are still controversial.

[6] 1. The Plaine des Sables caldera resulted from several collapse events starting 60 ka ago [*Bachèlery and Mairine*, 1990]. Several datings of the debris avalanche deposits (between 110 and 45 ka) dredged on the Eastern Plateau show that these events were associated with large flank landslides, the deposits of which spread over the proximal part of the submarine plateau and the Râlé Poussé (Figure 1b) [*Labazuy*, 1996].

[7] 2. The Enclos caldera formed around 4.5 ka ago [*Bachèlery and Mairine*, 1990]. Pyroclastic deposits around the Enclos indicate that the collapse was simultaneous to a large explosive eruption [*Abchir et al.*, 1998]. However, as for other basaltic calderas [e.g., *MacDonald*, 1972; *Munro and Rowland*, 1996], the volume of the deposit (around 0.5–1 km^3 [*Abchir et al.*, 1998]) is 1 order of magnitude lower than that of the Enclos caldera. The summit part of the Râlé Poussé debris avalanche lobe, which age is uncertain is considered as to be related to the slide of the Grand Brûlé [*Labazuy*, 1996; *Oehler*, 2005].

[8] During the "recent" Fournaise evolution (post-0.15 Ma), intense erosion led to the formation of deep valleys (Rivière des Remparts, Rivière Langevin, Rivière Basse Vallée, and Rivière de l'Est; Figure 1). An additional erosional structure was discovered on the eastern flank of PdF. Geophysical data and a drill hole in the northern part of the Grand Brûlé revealed the presence of the Osmondes paleoriver [*Courteaud*, 1996], where more than 170 m of alluvial formations were encountered below 60 m of recent lava flows. The relationship and the chronology between this valley and the Grand Brûlé are poorly understood.

3. Analysis of the Piton de la Fournaise Morphology

[9] Piton de la Fournaise is characterized by three main morphological features: (1) The flanks, which are continuous from the summit to the sea in the north, east and south, whereas the western flank is buttressed by Piton des Neiges, (2) two visible calderas, the Plaine des Sables and the EGBS, bounded by 100-m-high escarpments, and (3) very deeply incised valleys (the Rivière des Remparts, Rivière Langevin, Rivière de l'Est, and Rivière Basse Vallée), which dissect the western part of the edifice. We focus our analysis on the two first morphological structures.

3.1. Characterization of the Flanks of PdF

[10] Our analysis of the morphology of PdF is mainly based on the study of a 25-m step digital elevation model (DEM) developed by Institut Géographique National (IGN). This DEM was calculated by stereophotogrammetry from a set of aerial photographs taken in 1997. Stereotriangulation and 967 ground control points were used to control the accuracy of the model (vertical error of ± 4.5 m at 2σ [*Villeneuve*, 2000]). The DEM being calculated from aerial photographs, its accuracy slightly decreases in densely vegetated areas (i.e., the flanks of the edifice) and increases around the Enclos depression. As a consequence, minor topographic structures are disregarded.

[11] The flanks of PdF are characterized by different slope domains from the coast to the summit [*Rowland*

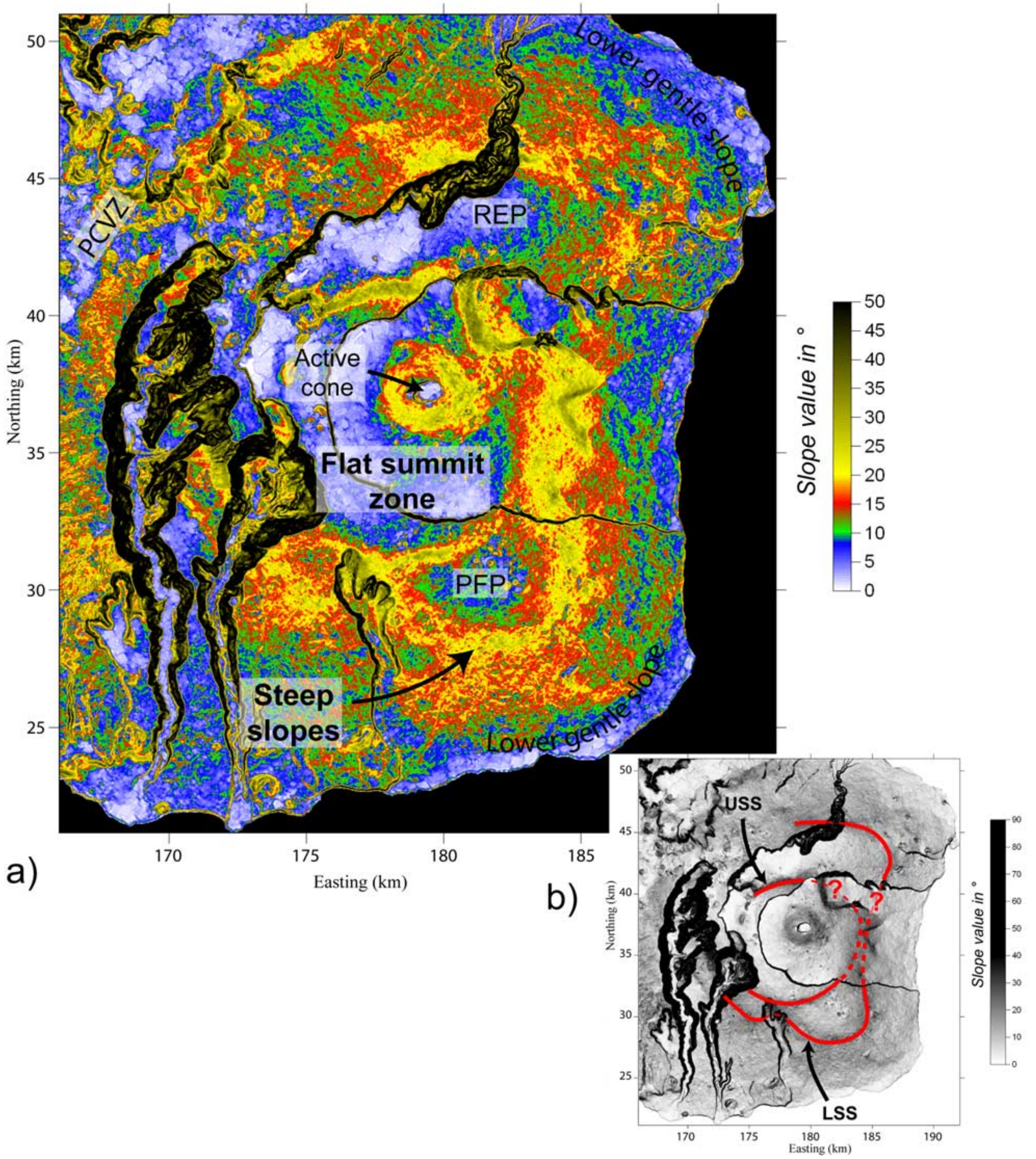

Figure 3. (a) Slope map of PdF showing the variable slopes on the volcano flanks. PFP, Piton de Fourche Plateau; REP, Rivière de l'Est Plateau; PCVZ, Plaine des Cafres Volcanic Zone. (b) Gray scale slope image with location of the upper and lower steep slope zones highlighted in red (USS and LSS) characterized by slope values ranging between 20° and 30°. Coordinates are in Gauss Laborde Réunion.

and Garbeil, 2000] (Figure 3a). At low elevations, the slope value range between 8° and 15°, corresponding to typical values of basaltic oceanic volcanoes [*Mark and Moore*, 1987; *Hürlimann et al.*, 2004]. At higher elevations, the topography of the southern, eastern and northern flanks is characterized by steep slope zones ranging between 20° and 35° (Figures 3 and 4), while the summit shows slopes between 2° and 8°, if the escarpments and the active cone are disregarded. Such a slope distribution is strikingly similar to what is observed on Volcán Fernandina, Wolf, and Cerro Azul in the western Galapagos archipelago [*Rowland*, 1996; *Rowland and Garbeil*, 2000]. In contrast,

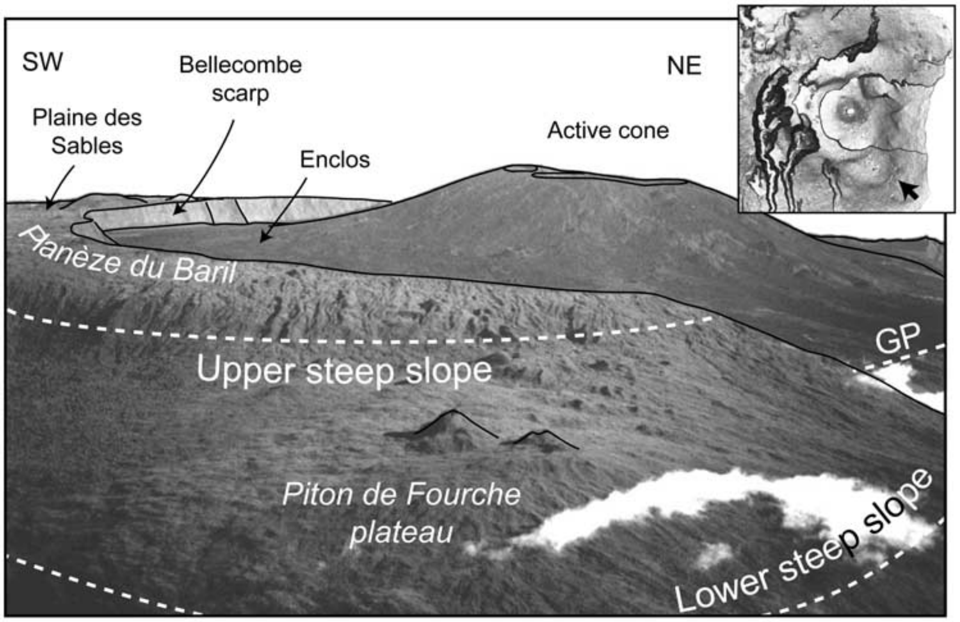

Figure 4. Northwestward view of the PdF summit. The USS separates two areas of low slope values (i.e., the Piton de Fourche Plateau below and the Planèze du Baril above). GP, Grandes Pentes. The black arrow in the insert indicates the location and the viewing direction.

this slope distribution is not observed on the western flank of PdF where the edifice is buttressed by Piton des Neiges.

[12] Radial topographic profiles allow a better characterization of the flank geometry (Figure 5). On the southern and eastern flanks, single linear profiles were produced from digital elevation models, but a composite profile was required for the northern flank in order to avoid complications of Rivière de l'Est and Plaine des Osmondes (Figure 5a). Topographic profiles, except for profile 3 (the east flank), reveal a seaward tilted stair-like topography with two steep slope zones separated by inclined plateaus such as the Piton de Fourche and the Rivière de l'Est (Figure 5a). Each steep slope zone corresponds to one step between flank segments that are characterized by constant slope values. The elevation changes across the upper steep slope (USS) and lower steep slope (LSS) are similar with 160, 230, and 210 m for the upper slope and 240, 285, and 220 m for the lower slope, respectively. Moreover, for each profile, the elevation change across the USS is always smaller than that across the LSS. In the SW, SE and north flanks, the USS and LSS are located at a nearly constant elevation at around 2000 m for the USS and 1000–1200 m for the LSS (Figure 5b). The resulting morphology is characterized by a geometry resembling an "overturned soup plate." Combining the slope map and topographic profiles suggests that the USS and LSS form two independent morphological structures on the southern and northern flanks of PdF. Concerning profile 3 on the east flank, only one steep slope zone is observed. This exception raises the problem of the relationship between the steep slope zones outside the EGBS and in the Grandes Pentes. The Grandes Pentes could correspond either to an independent structure whose development is related to the formation of the EGBS [e.g., *Bachèlery*, 1981; *Merle and Lénat*, 2003], or to the continuity of the upper and lower steep slope zone, which would merge in the east flank. Whatever the relationship between the Grandes Pentes and the USS and LSS, two morphological structures can be determined in the south, SE, NE, and north flanks (Figure 3b). The upper morphological structure is subcircular, whereas the lower structure presents two lobes at the location of the NE and SE rift zones. Along the NE rift zone, the lower structure presents relatively low slope values, which could be due to burial by lava flows such as during the 1977 eruption.

3.2. Analysis of the Enclos-Grand Brûlé Structure

[13] The EGBS is the most recent large-scale structure of Piton de la Fournaise. From east to west, it is composed of the Enclos depression, the Grandes Pentes and the Grand Brûlé (Figure 6). Secondary structures correspond to the Plaine des Osmondes and Piton de Crac. The Plaine des Osmondes is interpreted as the most recent collapse structure of the Enclos [*Bachèlery*, 1981]. The similar geometry (dip and thickness of the lava flows), the petrology of the lava flows of Piton de Crac and Bois Blanc scarp, and the orientation of the only dike observed on Piton de Crac toward the active summit cone suggests that Piton de Crac is a remnant part of PdF isolated by collapse events and erosion [*Bachèlery*, 1981].

[14] For our analysis of the EGBS, we consider three different zones that help reinterpreting the evolution and the formation of the overall collapse structure.

3.2.1. Enclos

[15] The floor topography of the Enclos reveals a series of relatively flat plateaus at different elevations: 2050–2200, 1750–1900, and 1900–2000 m for the western, southern, and northern plateau, respectively, which are separated by steeper slope zones (Figure 7a). The transition between the western and southern plateaus consists of a N65° trending steep slope zone, which is associated with a subtle kilome-

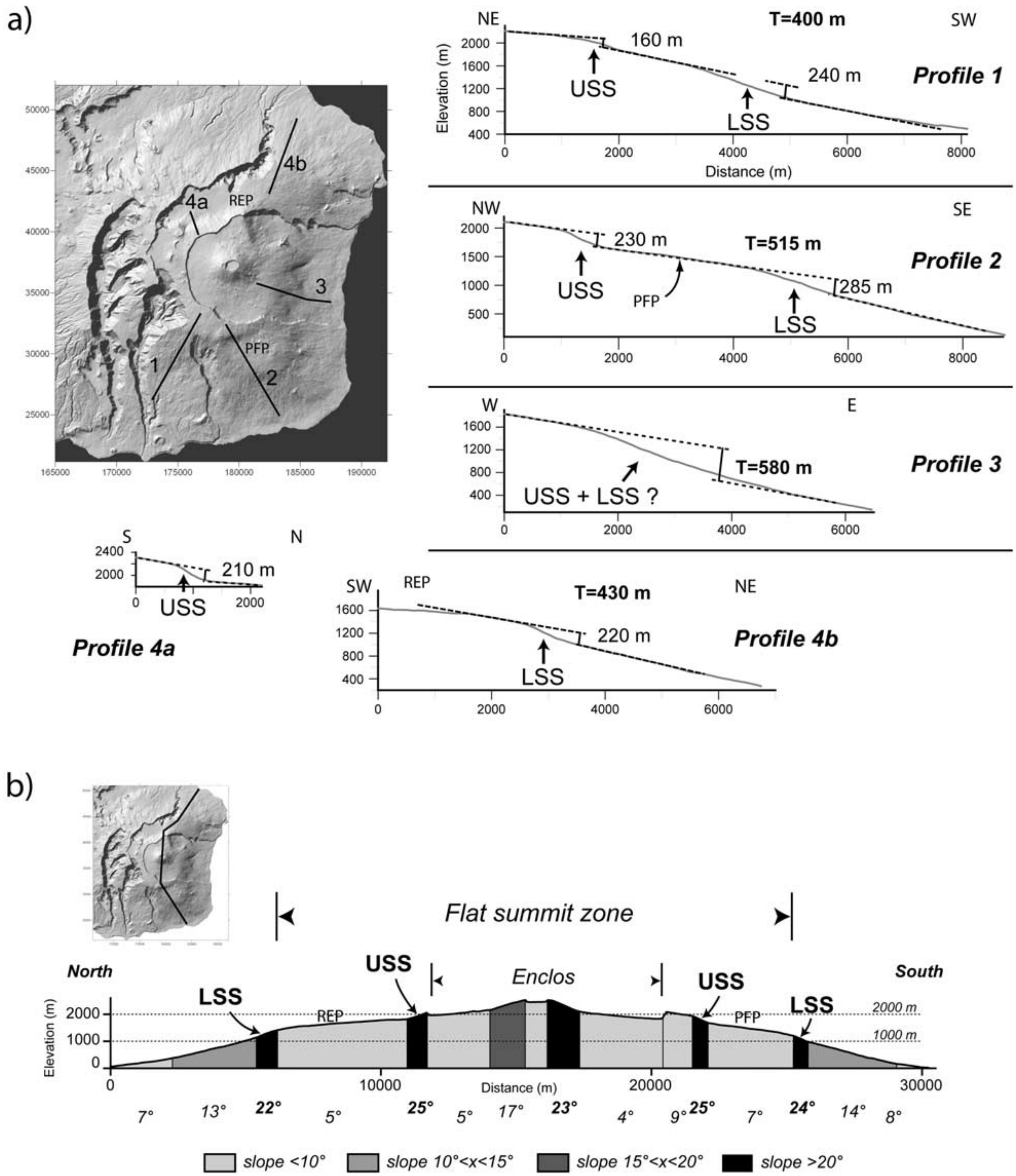

Figure 5. (a) Radial topographic profiles showing the succession of steep slopes and domains of gentle slopes. Elevation changes between each domain of gentle slopes are determined. No vertical exaggeration. T corresponds to the total elevation difference for each profile. USS, upper steep slope; LSS, lower steep slope; REP, Rivière de l'Est Plateau; PFP, Piton de Fourche Plateau. (b) North-south topographic profile illustrating the shield morphology of PdF and its variable slope domains. Note the nearly identical elevation of both the LSS and USS on the north and south flanks.

ter-long lineament, located at the slope change (Figure 7b). It appears that this lineament is restricted to the EGBS, i.e., it cannot be traced south of the Enclos scarp, and continues northeast toward the Grandes Pentes (Figure 7c). In a volcanic setting, a lineament can have several origins: the margin of a lava flow, the trace of a lava channel or lava tube, a spatter rampart, a fluvial gully or a fault. In the present case, the orientation of the lineament, oblique to the

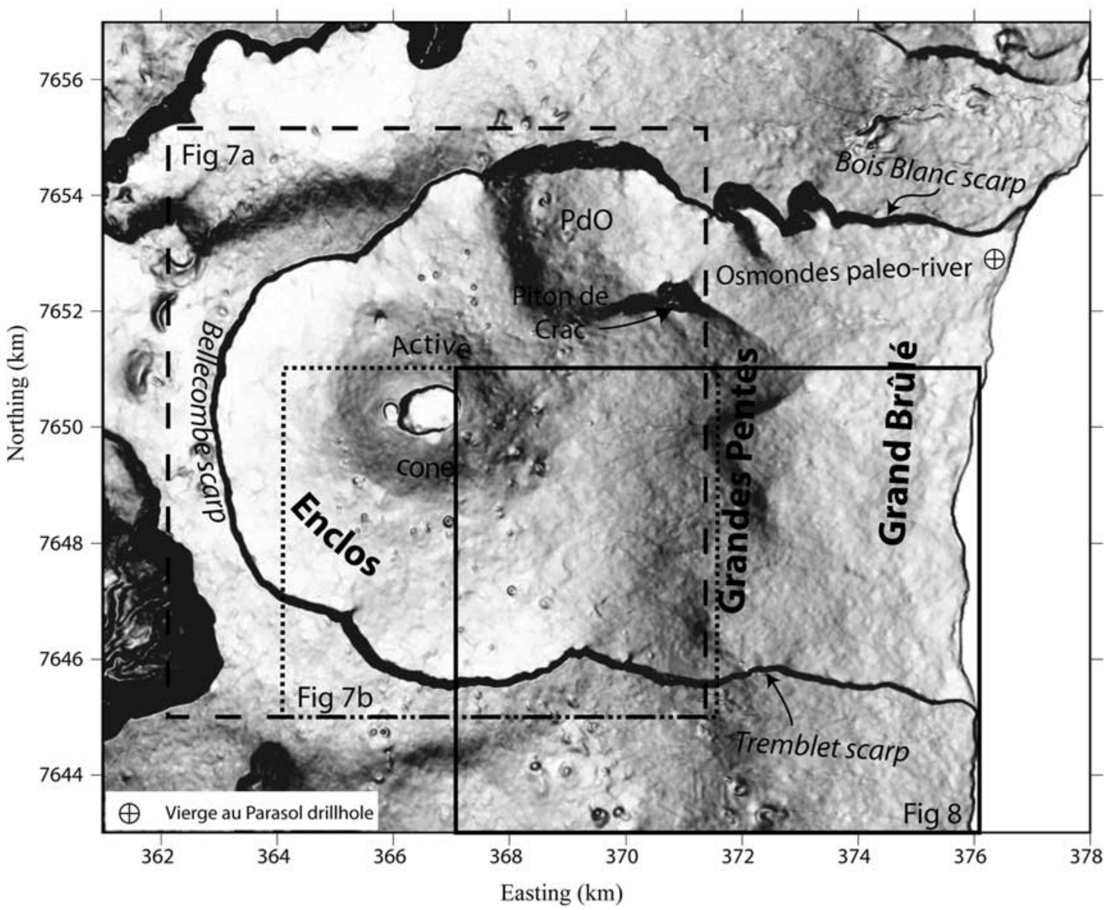

Figure 6. Gray scale slope map representing the structures of the EGBS. Figures 7a, 7b, and 8 are located. Coordinates are in UTM WGS84.

general slope, and its length are difficult to explain as lava structures or river patterns. Field observations reveal that long and linear extensive fractures cut several units of lava flows of different ages at the upper break-in-slope, which corresponds to the location of the lineament (Figure 7b). The older the lava flows, the wider the fractures. The development of such a fracture network is in agreement with a continuous process of deformation. Moreover, this part of the Enclos is characterized by recurrent tectonic earthquakes confined to the region south of the lineament [*Lénat et al.*, 1989; *Sapin et al.*, 1996] (Figure 7a). We suggest that the alignment of the extensive fractures, the lineament and the seismicity results from the activity of a normal fault (Figure 7b). We also propose that the 100-m-high offset of the Enclos floor results from tectonic activity along this large fault during or after the EGBS formation.

3.2.2. Tremblet Scarp

[16] Analysis of the morphology of PdF reveals the existence of steep slope zones both inside and outside the EGBS. The possible link between these morphological structures across the Tremblet scarp in the south and across the Bois Blanc scarp in the north is still unclear even though the different steep slope zones are characterized by identical slope values. The detailed analysis of the slope distribution on both sides of the Tremblet scarp provides new constrains on the relationship between the morphological structures outside and inside the EGBS.

[17] The lowest part of the Grandes Pentes is characterized by a V-shaped structure (VSS), which was commonly interpreted as the trace of a small-scale landslide that has occurred after the Grand Brûlé landslide [e.g., *Bachèlery*, 1981] (Figures 7c and 8). This structure is limited by two NE and SE linear steep slope zones where slope values exceed 30° (Figure 8). The slope map suggests that another SE trending steep slope zone seems to be continuous south of the Tremblet scarp (Figure 8b). However, the irregularity of the topography does not allow establishing firmly this continuity outside the EGBS. We calculated a 50-m low-pass-filtered DEM from the original 25-m step DEM (Figure 8c) in order to smooth the rough topography and to keep the first-order morphology. Three different slope domains can be distinguished from this DEM: domain A with an average slope of 10°, domain B with 14° and domain C exceeding 18°. Geometrically, these areas are continuous on both sides of the Tremblet scarp. Domain A is at low elevation. Inside the EGBS, it corresponds to the upper part of the Grand Brûlé. South of the Tremblet scarp, domain A is situated between the seacoast and domain B. The lowest part of the Grand Brûlé, which is characterized by gentle slopes between 4° and 8° corresponds to the coastal plain formed by the overlap of multiple lava deltas

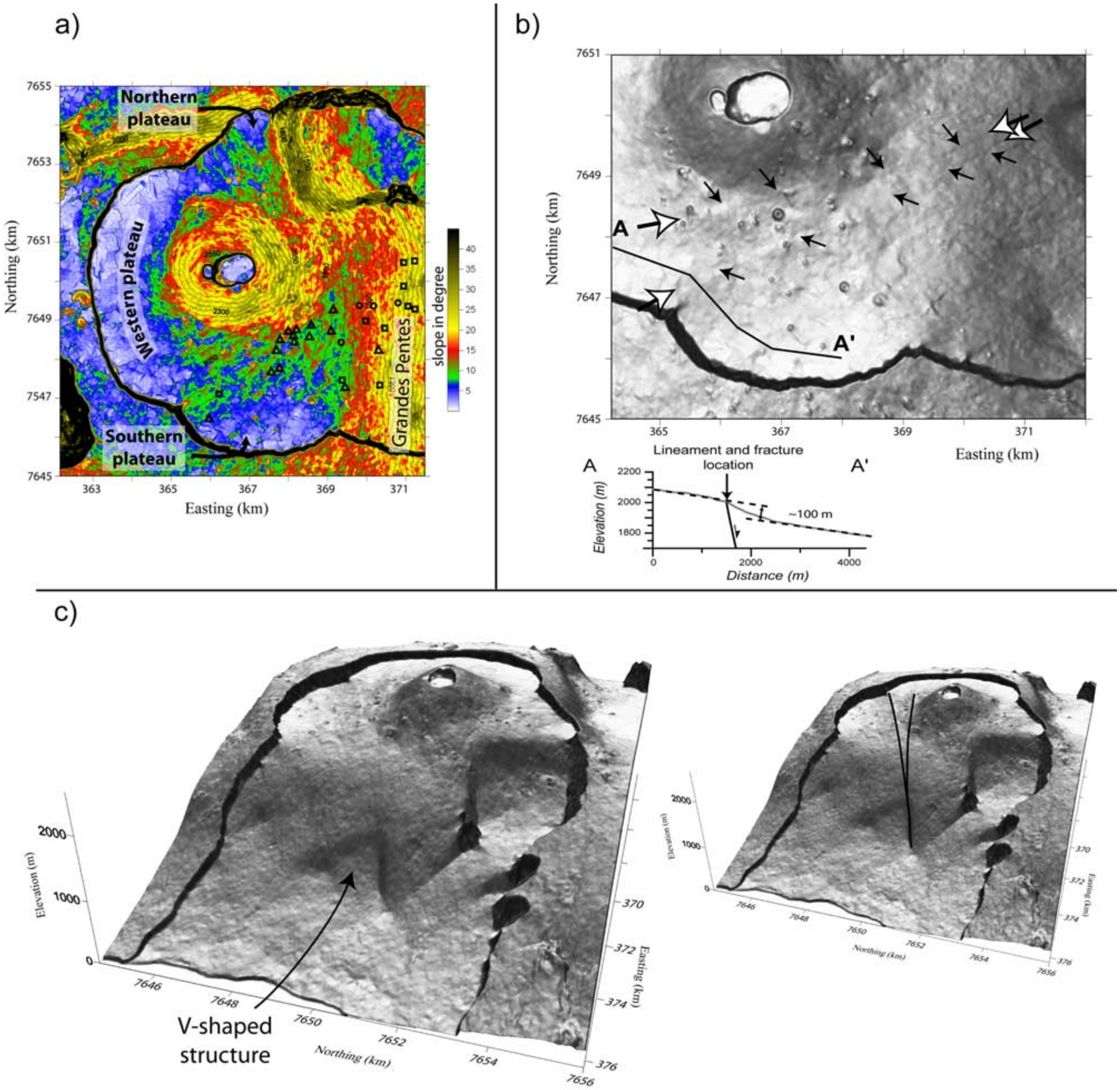

Figure 7. (a) Slope map of the Enclos showing the presence of three different plateaus east of the Grandes Pentes and the Plaine des Osmondes. (b) A main N65° and kilometer-long lineament is observed between the western and southern plateaus. Topographic profile A-A′ reveals that the lineament and the fractures are located at the slope change. The distribution of the nonmagmatic earthquakes in the lineament vicinity strongly supports a fault origin for the lineament (circles, post-1997 events, data of the Piton de la Fournaise Volcano Observatory; squares, July 1985 events [after *Lénat et al.*, 1989]; triangles, 1985–1988 events [after *Sapin et al.*, 1996]). (c) Three-dimensional representation of the EGBS slope map with the inferred N65° faults. Coordinates are in UTM WGS84.

[see *Rowland and Garbeil*, 2000] after the collapse of the EGBS. Domain B is restricted to a small area, which overlaps the Tremblet scarp. Part of its upper limit coincides with the margin of the SE trending steep slope located south of the VSS. Laterally, domain B stops in the north, whereas it continues southward between domains A and C. Note that Piton Takamaka increases the slope of this domain (Figure 8c). Domain C is characterized by the steepest slopes and corresponds to the Grandes Pentes in the EGBS divided in two branches south of the Tremblet scarp (i.e., the LSS and USS). It is noteworthy that the lowest part of the Grandes Pentes and LSS present the same morphological features that form the steep SE trending slopes.

[18] The clear continuity of the slope domains across the scarp into the EGBS suggests a continuity of the USS and LSS in the EGBS where they coalesce in a main steep slope

(the Grandes Pentes). Furthermore, the two morphological structures related to the USS and LSS are continuous in the north, east and south flanks of PdF. The classic interpretation in which the Grandes Pentes are restricted to the EGBS and are due to its formation [e.g., *Duffield et al.*, 1982; *Gillot et al.*, 1994; *Labazuy*, 1996; *Merle and Lénat*, 2003; *Oehler et al.*, 2004] is therefore in disagreement with the present data. Finally, the continuity of the steep slopes across the Tremblet scarp raises the questions of the chronology between the EGBS formation and the development of the USS and LSS, and of the direction of motion (vertical versus lateral), along the Tremblet scarp.

3.2.3. Bois Blanc Scarp

[19] The Bois Blanc scarp is usually considered to be the limit of the Grand Brûlé landslide, which is considered to have occurred 4.5–10 ka ago [*Bachèlery*, 1981; *Duffield et al.*, 1982; *Labazuy*, 1996; *Merle and Lénat*, 2003]. We have shown above that more than 170 m of alluvial formations, the Osmondes paleovalley, occur at the foot of the Bois Blanc scarp [*Courteaud*, 1996]. This suggests that the Osmondes paleovalley was deeply incised, below the present sea level, after the Grand Brûlé slide and was subsequently filled by fluvial deposits. Was the development of such a deep paleovalley possible after the Grand Brûlé slide?

[20] P. Mairine's unpublished data (2004) from drill holes in most of the main rivers of La Réunion Island reveal that the bottom of these valleys is always located 100–120 m below sea level. It is widely assumed that the incision depth is directly related to the elevation of the base level, which is considered to be the sea level in an island setting [e.g., *Schumm*, 1993]. Incision and sedimentation develop above and below the base level, respectively. The occurrence of alluvial sediments below sea level, at the foot of the Bois Blanc scarp clearly indicates that the Osmondes paleovalley was incised during a period of low sea level and was subsequently filled in. In the Indian Ocean, the last significant sea level low (-120 ± 5 m) occurred 17–18 ka ago during the last glacial period [*Camoin et al.*, 2004]. It was followed by a rapid sea level rise until 8–9 ka ago, when the sea level reached an elevation of 10 m below the present sea level. This evolution, which is common to the other oceans [e.g., *Hanebuth et al.*, 2000], implies that the incision of the Osmondes paleovalley cannot postdate 17–18 ka.

[21] The identical morphology (i.e., slope value of 55° to 60° and escarpment heights of 100 to 200 m) of the Tremblet and Bellecombe scarps, which bound the Grand Brûlé and Enclos structures, respectively, suggests a nearly similar age for the scarp formation [*Merle and Lénat*, 2003]. For comparison, the Plaine des Sables scarps which have been altered by erosion for about 60–45 ka, present slope values of 45°. Hence we argue that, the incision of the Osmondes paleoriver predates the formation of the Grand Brûlé, and the Bois Blanc scarp does correspond to the northern margin of the Osmondes paleovalley, rather than to the limit of a landslide. The northern limit of the Grand

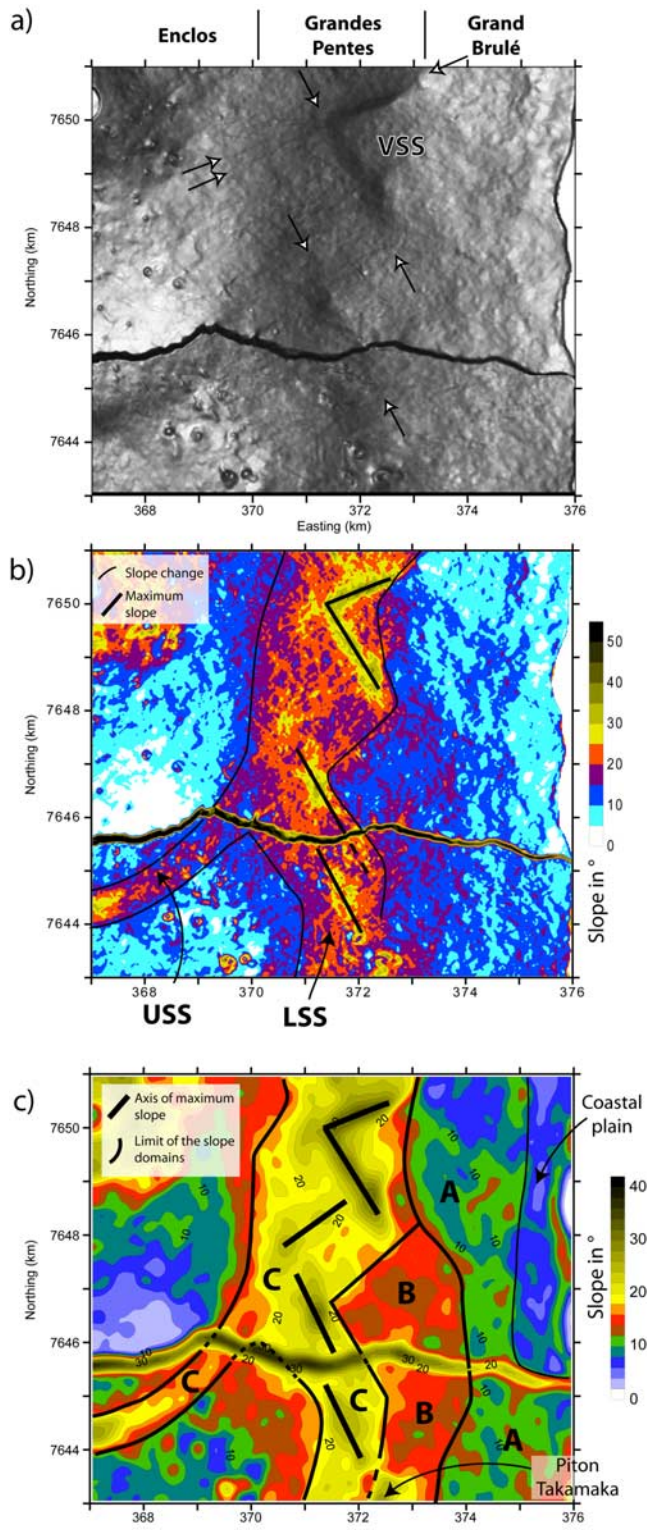

Figure 8. (a) Gray scale slope map illustrating the morphology of the Grandes Pentes. Several subtle lineaments limited by arrows are observed into the EGBS. Two of them bound a V-shaped structure (VSS) which is classically interpreted as to be the trace of a secondary landslide. (b) Slope map showing the maximum slope axes and the slope changes. (c) Slope map of a 50-m resolution filtered DEM illustrating the longer-wavelength morphology. The slope domains A, B, and C are continuous on both sides of the Tremblet scarp. The coastal plain of the Grand Brûlé likely results from the accumulation of lava deltas. Coordinates are in UTM WGS84.

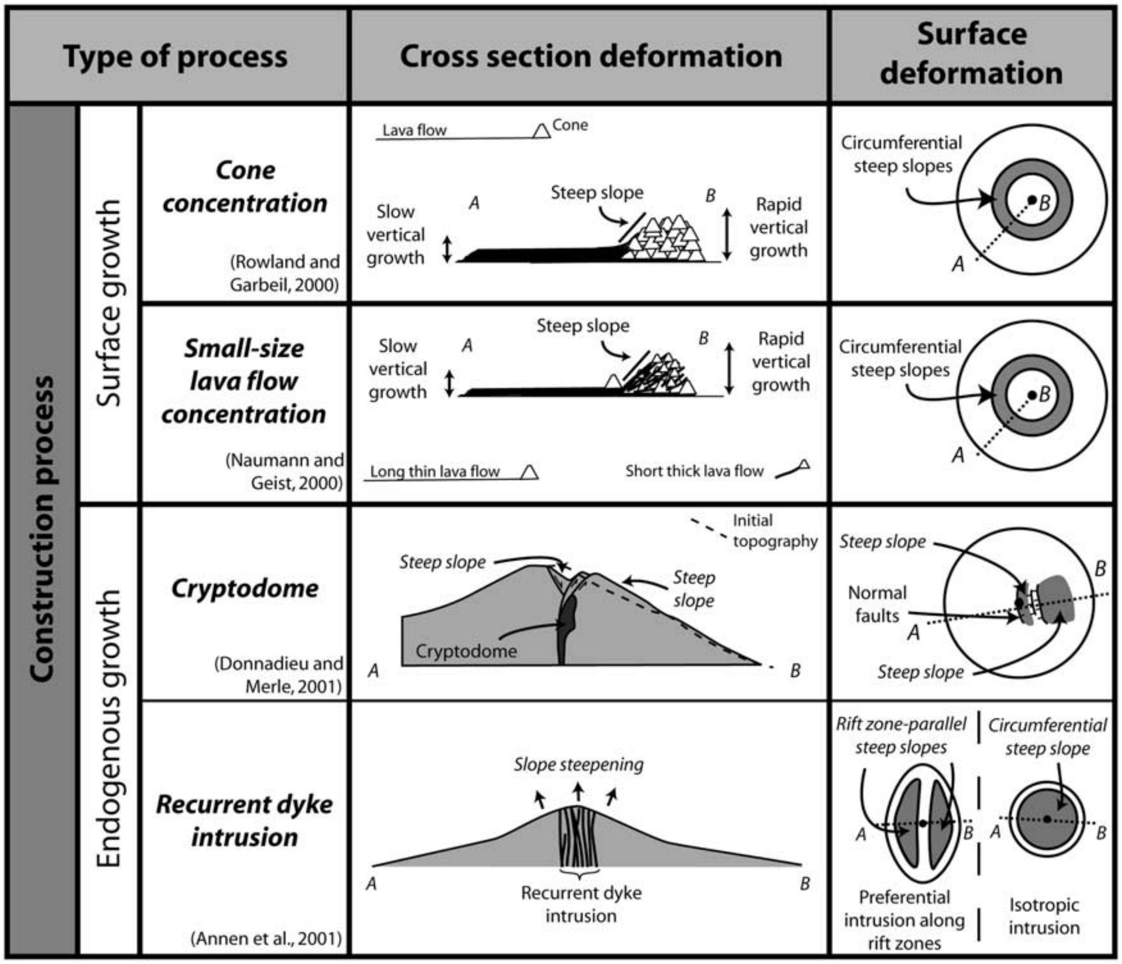

Figure 9. Construction processes able to develop steep slopes in basalt shield volcanoes.

Brûlé is subsequently located south of the Osmondes paleovalley axis and the Piton de Crac is a remnant part of the southern flank of the paleovalley.

4. Discussion

[22] Our study of the morphology of Piton de la Fournaise reveals the existence of two subcircumferential steep slope zones on the volcano's north, NE, SE, and south flanks, which are locally incised by deep valleys and, which are continuous on the east flank, into the Enclos-Grand Brûlé structure. In the following, we show the main implications of this geometric relationship that led to the recent evolution of Piton de la Fournaise and the development of the EGBS.

4.1. Steep Slope-Forming Processes

[23] Basaltic shield volcanoes are commonly characterized by a relatively flat summit zone and gentle oceanward slopes. This classic shape results from the superposition of thin low-viscosity lava flows the run out distances of which vary from few hundreds of meters to several kilometers. However, steep slopes are also observed on basaltic volcanoes [e.g., *Rowland and Garbeil*, 2000] and several processes can explain their development.

[24] Steep slopes may result from the concentration of eruptive vents and pyroclastic cones in the summit area, leading to a differential vertical growth between the summit and the volcano flanks (Figure 9) [*Rowland and Garbeil*, 2000]. The differential vertical growth may also originate from the development of small-size lava flows related to concentric summit vents [*Naumann and Geist*, 2000]. These two constructional processes lead to the formation of steep convex-outward slopes like at Cerro Azul, Wolf and Fernandina volcanoes, western Galapagos [*Rowland and Garbeil*, 2000; *Naumann and Geist*, 2000]. Endogenous growth through recurrent dike intrusion [e.g., *Annen et al.*, 2001], the growth of cryptodomes [*Donnadieu and Merle*, 2001] and an overpressurized magmatic chamber [*Cullen et al.*, 1987] are possible processes, which induce the formation of steep slopes on the summit or the flanks of the volcano. For dike intrusions, the resulting slope geometry ranges from elongated slopes along-strike of a rift zone to an isotropic slope distribution restricted to the summit [*Rowland and Garbeil*, 2000; *Annen et al.*, 2001]. The growth of a cryptodome leads to an asymmetric deformation and the development of concave- and convex-outward steep slopes [*Donnadieu and Merle*, 2001].

[25] Destruction processes such as erosion and landslides can also form steep slopes (Figure 10). Erosion is responsible for valleys the development of which can be controlled

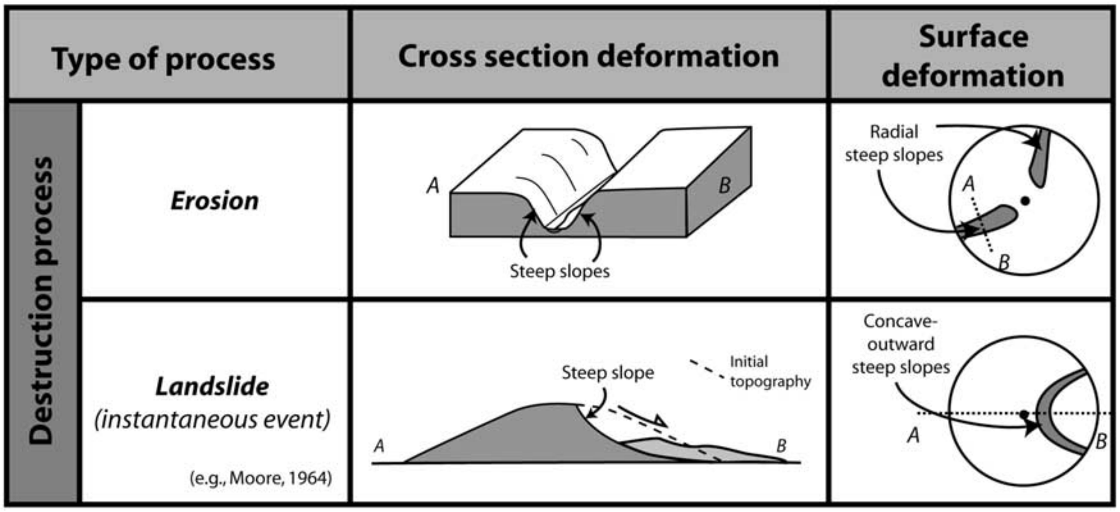

Figure 10. Destruction processes able to develop steep slopes in basalt shield volcanoes.

by preexisting faults or structural limits such as caldera walls [*Stearns and MacDonald*, 1946; *Bachèlery*, 1981]. Small-size and large-scale landslides also produce steep slopes on the volcano flanks presenting a concave-oceanward geometry [*Moore*, 1964; *Bachèlery*, 1981; *Oehler et al.*, 2004].

[26] Deformation processes have a strong influence on the volcano morphology (Figure 11). An edifice may deform above a basal layer composed of low-strength sediments, ultramafic cumulates or intrusive complexes [e.g., *Borgia et al.*, 1990; *Borgia*, 1994]. Circumferential thrust-fault-related steep slopes develop at the base of the edifice and normal faults accommodate the spreading motion [*Merle and Borgia*, 1996]. The lack of any deformation within the pelagic sediments below the submarine flanks and of any distal anticline, even several tens of kilometers away from the volcano suggests that this process did not occur at La Réunion [*Michon et al.*, 2007]. A hydrothermally altered interior of a volcano may also act as a décollement level. As shown by *Cecchi et al.* [2005], the sagging of the volcano's summit part results in the upward lateral extrusion of the altered interior. The resulting morphology presents circum-

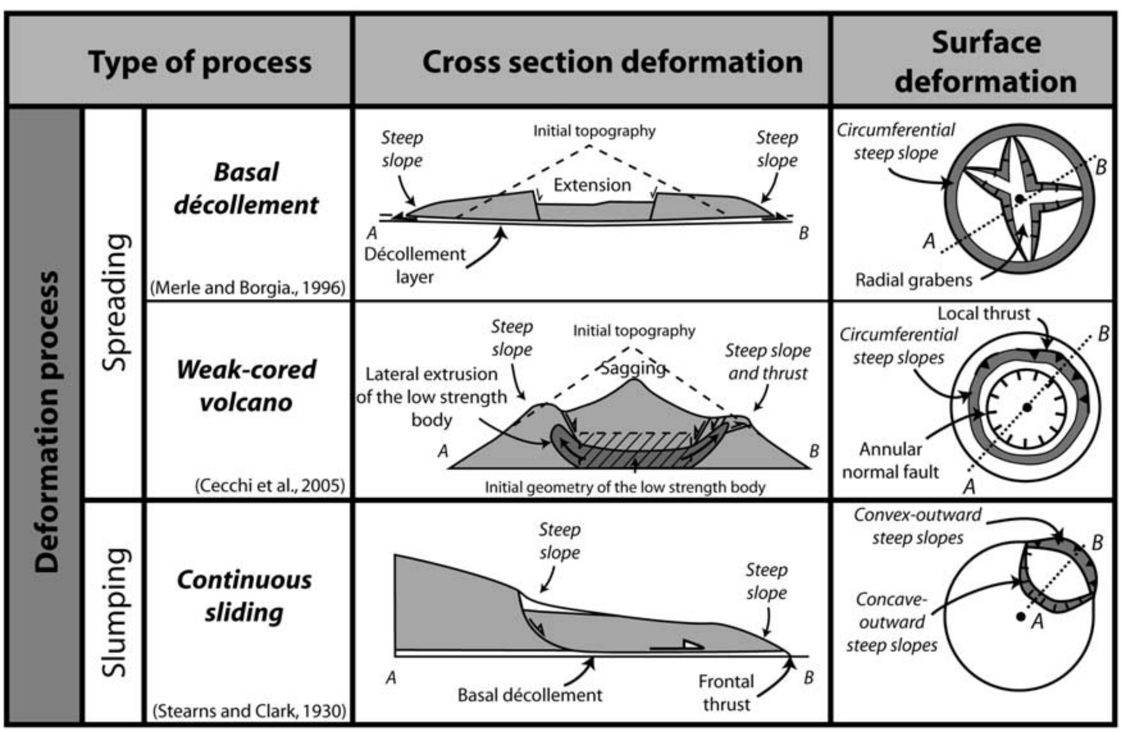

Figure 11. Deformation processes able to develop steep slopes in basalt shield volcanoes.

ferential steep slopes that are locally associated with thrust faults (Figure 11). Finally, slumping units above a local décollement level are bounded upward by normal faults, which also form steep slopes on the volcano flanks [*Stearns and Clark*, 1930; *Moore and Krivoy*, 1964; *Merle and Lénat*, 2003]. Steep slopes related to thrust faults present a convex-outward shape whereas concave-outward steep slopes are attributed to normal faults.

[27] In summary, circumferential steep slopes like those observed at PdF can only result from either construction processes or spreading processes, and not through erosion.

4.2. Caldera-Forming Processes

[28] Several mechanisms may lead to the development of circular to elliptical, and horseshoe-shaped calderas. Their geometry and size depend on the mechanisms encountered. On volcanoes other than basaltic shields, the development of large calderas is attributed to the collapse of the chamber roof during large explosive eruptions [e.g., *Lipman*, 1997; *Roche et al.*, 2000]. In that context, the size of the caldera is dependent on the volume of the erupted magma. On basaltic shield volcanoes, large calderas with diameters ranging from 5 to 20 km may also develop [*MacDonald*, 1965; *Bachèlery*, 1981; *Munro and Rowland*, 1996]. However, they usually lack of significant related pyroclastic deposits suggesting nonexplosive or minor explosive mechanisms to explain their development [e.g., *MacDonald*, 1972; *Munro and Rowland*, 1996]. In Hawaii, the caldera of Kilauea is interpreted as resulting from either the coalescence of several pit craters [*MacDonald*, 1965], a main collapse related to the lateral magma withdrawal during the large lateral eruption of 1790 [*MacPhie et al.*, 1990] or from the load of a cumulate complex at the base of the magma chamber [*Walker*, 1988]. In the western Galapagos, the combined effect of magma withdrawal and the load of cumulative bodies is considered to be responsible for the development of the large and deep calderas [*Munro and Rowland*, 1996]. Recently, *Cecchi et al.* [2005] showed that the deformation of the hydrothermal system, which is obviously larger that the magmatic reservoir may lead to the collapse of large calderas. It is important to note that whatever the model, the collapse of the volcano's summit is related to the active magmatic system.

[29] Horseshoe-shaped calderas are relatively common on island volcanoes like Fogo, El Hierro, Tenerife, and La Réunion islands. Their geometry is characterized by a flat summit zone and a large depression bounded by regular scarps that link the summit depression to the sea coast. The frequent occurrence of debris avalanche deposits on the submarine flanks suggests that their formation is due to destabilizations and subsequent landslides. Several models have been proposed to explain their origin. They can result from a single large landslide that affects both the volcano flank and the summit [*Duffield et al.*, 1982; *Labazuy*, 1996; *Cantagrel et al.*, 1999; *Day et al.*, 1999]. *Hürlimann et al.* [1999] proposed that the collapse of the summit was able to destabilize the volcano flank, leading to flank landslide. *Merle and Lénat* [2003] recently showed with analog models that the slide of a volcano flank may trigger a lateral flow in the hydrothermal system, which leads to the development of a large caldera in the summit area. They applied this model to PdF in order to explain the development of the EGBS.

4.3. Origin of the Steep Slopes and the EGBS

[30] One of the main observations made in the present study is the apparent continuity of the different slope domains across the Tremblet scarp into the EGBS without any visible lateral offset (Figure 7). Such a continuity suggests that the steep slopes and the EGBS are independent structures. We have used the geometry of both structures and their intersection to determine their temporal relationship and origin.

4.3.1. Age and Origin of the LSS and USS

[31] We put forward two hypotheses to explain the continuity of the slope domains on both sides of the Tremblet scarp:

[32] Hypothesis 1 is that the development of the LSS and USS post-dates the formation of the EGBS the lower part of which (i.e., the Grand Brûlé) has previously been considered as the trace of a landslide. Assuming that the EGBS formed 4.5 to 10 ka ago, the steep slopes developed quite rapidly. We showed that circumferential steep slopes result from two different processes: differential growth (construction process) or spreading (deformation process).

[33] Differential growth is highly unlikely for the following reasons. (1) In contrast to the volcanoes of western Galapagos, no concentric eruptive vents are found above the USS and LSS. (2) There is no concentration of pyroclastic cones in the two areas delimited by the steep slopes. (3) The length of lava flows is not constant and varies from hundreds of meters to several kilometers, making the lava flow accumulation hypothesis unrealistic. (4) Finally, 97% of the post-EGBS volcanic activity has been restricted to the Enclos. The total elevation changes along each topographic profile (T value in Figure 5a), which would reveal the differential growth should not be similar outside and inside the EGBS.

[34] Spreading is the second process capable of the development of circumferential steep slopes. If the elevation changes related to the LSS and USS are related to deformation, the total deformation is a combination of outward lateral and upward displacements. Considering the upward motion only, which corresponds to the elevation change measured for each profile, minimum displacement rates of 2 to 4 cm/a and 2.5 to 5 cm/a are calculated for the USS and LSS and are inferred for a deformation starting at 10 and 5 ka, respectively. These values are similar to the total deformation rates measured for the very active spreading at Kilauea where large earthquakes occurred [*Delaney et al.*, 1998]. At PdF the lack of seismicity in the vicinity of the steep slopes is in disagreement with a very active deformation. Moreover, geodetic data acquired outside the EGBS in the flat summit zone do not show any displacement [*Briole et al.*, 1998]. Consequently, the development of the LSS and USS due to spreading does not agree with the available data.

[35] Hypothesis 2 is that the development of the LSS and USS predates the formation of the EGBS and does not result from a large deeply rooted landslide. It seems unrealistic that a superficial landslide occurred on a décollement level presenting a topography characterized by similar slope domains than outside the EGBS. The exact motion along

the lateral scarps and subsequently of the Grand Brûlé cannot be determined precisely. However, only a predominantly vertical collapse is possible. Hence the formation of the EGBS results from a vertical to subvertical collapse rather than a giant landslide. Such a motion could explain the lack of clear continuity between subaerial and submarine structures [*Oehler*, 2005] (Figure 1b) in contrast to what is observed at large landslide deposits [e.g., *Le Friant et al.*, 2004; *Tibaldi*, 2001]. The preservation of the slope domains inside the EGBS would be explained by a low amount of overlapping lava flows in the Grandes Pentes and the Grand Brûlé since the collapse. Such a hypothesis is supported by the map of the historical lava flows, which shows that most of the lava flows are restricted to the upper part of the EGBS (i.e., the Enclos) and only few flows occur in the Grande Pentes [*Stieltjes et al.*, 1986].

[36] As for hypothesis 1, construction and deformation processes might form circumferential steep slopes. We showed above that construction processes like those proposed by *Naumann and Geist* [2000] and *Rowland and Garbeil* [2000] cannot be applied to PdF. The eruptive vents are not concentrically distributed above the LSS and USS, and the lava flow length strongly varies.

[37] In contrast, spreading may have occurred. The pre-EGBS edifice built up after the Plaine des Sables collapse about 60–45 ka ago. According to this potential onset of deformation, and given the elevation changes, expected deformation rates are of few millimeters per year. Volcanoes might spread under gravity if a low-strength body/layer exists at the base of the edifice, and if the load of the edifice acting on the low-strength material is high enough. Two different modes of spreading lead to the development of circumferential steep slopes. On the one hand, the spreading can be related to the deformation of a basal low-strength layer [e.g., *Merle and Borgia*, 1996; *Oehler et al.*, 2005]. In such a case, the steep slopes are located at the base of the edifice and radial normal faults develop. On the other hand, the spreading is induced by an internal low-strength layer or body [e.g., *Cecchi et al.*, 2005]. The resulting steep slopes are on the volcano flanks and normal faults are circumferential. At PdF, the exact topography of the pre-EGBS volcano cannot be firmly determined. Nevertheless, the morphology and the slope adjacent to the Enclos depression, which are preserved from the pre-EGBS period, and the shape of the active cone, allows estimating a pre-EGBS collapse topography of around 3000 m located at the place of the present active cone. The location of the circumferential LSS and USS on the volcano flanks supports the presence of an internal low-strength body rather than a basal décollement. In conclusion, we propose that the LSS and USS do result from neither construction processes nor a basal spreading, but from the deformation of the weak internal part of the volcano under gravity (Figure 12).

[38] The nature and geometry of the weak core are unknown. According to geoelectrical data acquired in the Enclos, the Plaines de Sables, and the Planèze du Baril [*Benderitter*, 1990; *Courteaud*, 1996; *Lénat et al.*, 2000], PdF is characterized by a deep widespread low-resistivity body which could correspond to the top of the hydrothermal system of the pre- EGBS volcano. The lack of hot springs [*Coudray et al.*, 1990] in the adjacent deep valleys and fumaroles would suggest that this hydrothermal system is no longer active. Nevertheless, these characteristics are not conclusive against the active hydrothermal system as despite a current intense volcanic activity and a large hydrothermal system [*Lénat et al.*, 2000], PdF presents only very few fumaroles restricted to the summit pit crater and no hot springs at all. It is subsequently hard to determine whether spreading is still active, but we suggest that the deformation stopped when the Enclos collapsed 4.5 ka ago. Indeed, the summit of the volcano, which was acting on the hydrothermal body was dismantled. The lack of steep slope on the west part of PdF could suggest that the remnant part of the ‘‘ancient’’ Fournaise blocked the deformation in the west.

[39] Contrary to the experiments of *Cecchi et al.* [2005] in which only one circumferential steep slope zone is described, two steep slope zones developed at PdF. Such a vertical complexity in the steep slope geometry also occurs on other volcanoes where several bulges are visible (e.g., Etna, Arenal [*Cecchi et al.*, 2005]). This can be explained by the complexity of the hydrothermal system compared to the simplified models. For instance, intra-edifice preexisting discontinuities might induce a deformation partitioning and subsequently several steep slope levels. Laterally the steep slope geometry can also be influenced by the rift zones along which recurrent magma intrusions decrease the strength of the rocks by hydrothermal alteration. Such a lateral variation could explain the NE and SE lobes of the LSS along the NE and SE rift zones, respectively. This indicates that the current rift zones have been inherited from the pre-Enclos volcano.

4.3.2. Origin of the EGBS

[40] Usually, vertical collapses are restricted to the area of the magmatic system (i.e., summit and rift zones). At PdF, besides the summit zone, the eastern flank also suffered a vertical collapse. Given the geomorphological and geological continuity between the Enclos and the Grand Brûlé (identical scarps slope and continuous faults in the EGBS), we propose that the collapse of the Enclos was associated to that of the Grand Brûlé. Although large lateral eruptions already caused the development of summit calderas on basaltic and andesitic volcanoes [e.g., *MacPhie et al.*, 1990; *Kaneko et al.*, 2005], the collapse of the Grand Brûlé, several kilometers away from any active magmatic system, is unlikely related to magmatic withdrawal. The vertical collapse of the EGBS might also be triggered by a deeply rooted slump in which the displacements in the upper part (close to the upper normal fault) can be predominantly vertical whereas they evolve to lateral away from the fault (above the low dipping décollement). Assuming a maximum depth of the décollement at the top of the oceanic crust (6 km below sea level (bsl) [*de Voogd et al.*, 1999]) and the Bellecombe scarp as the trace of the upper normal fault, the vertical motion would be restricted to the few first kilometers only, i.e., west of the Grandes Pentes, and not extended to the overall EGBS.

[41] We showed that the Grand Brûlé was located above the large intrusive complex of the Alizés volcano, the age of which is considered as to be older than 0.78 Ma [*Lénat et al.*, 2001]. This complex was drilled between 950 and 2850 m bsl, end of the drilling. It is composed of gabbros in the 1400 upper meters and dunites and wehrlite dunites below [*Rançon et al.*, 1989]. Gravimetric data suggest that the complex continues down to at least 4 km bsl

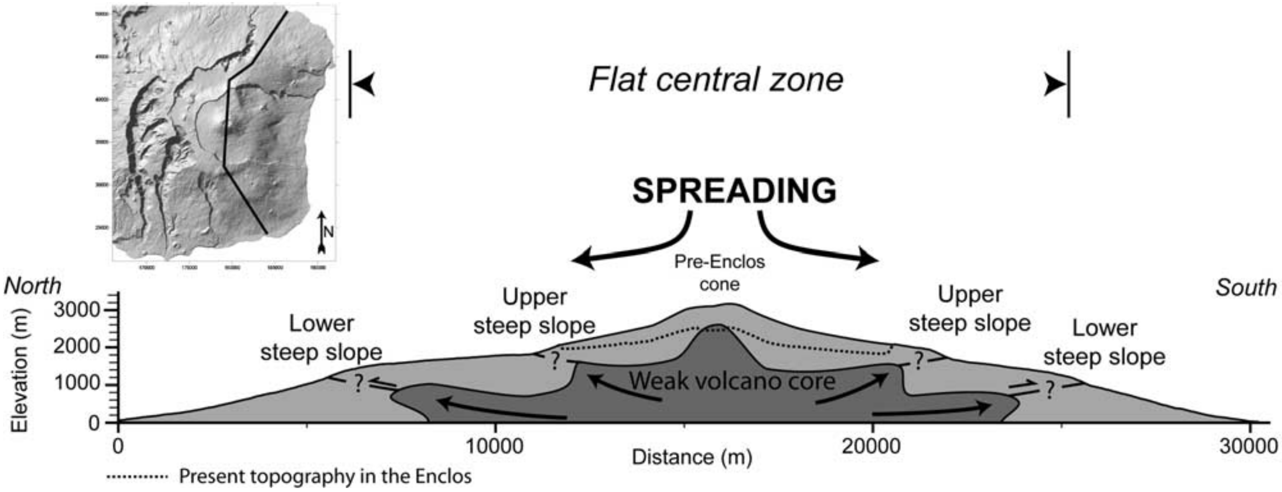

Figure 12. N-S cross section of PdF illustrating the potential origin of the upper and lower circumferential steep slopes by the spreading of the edifice above a weak hydrothermal core.

[*Malengreau et al.*, 1999]. Such a body may have two different implications in the evolution of PdF:

[42] 1. It has been shown that olivine cumulates at high temperature (>1100°C) may deform as ice [*Clague and Denlinger*, 1994]. The Alizés cumulative complex could have deformed under gravitational forces and subsequently induced the vertical collapse of the Grand Brûlé. However, as the drilling project aimed at determining the geothermal potential of the intrusive complex, it has been shown that the present temperature of the complex was far from hot (142°C at 3003.5 m below the surface [*Rançon et al.*, 1989]). Even if the 4.5 ka ago temperature cannot be firmly determined, analysis of secondary minerals within the gabbro and dunite reveals a progressive cooling of the complex from late magmatic biotite crystallization at 600–900°C to serpentine crystallization at 350°C maximum [*Lerebour et al.*, 1989]. Hence it is strongly unlikely that the olivine cumulates of the Alizés intrusive complex deformed in the recent times and triggered the collapse of the Grand Brûlé 4.5 ka ago.

[43] 2. *Walker* [1988] proposed that dense bodies, such as cumulates are able to trigger a downward drag and the collapse of the summit. Such a mechanism has been proposed as a source of subsidence to explain the development of the large calderas of the volcanoes of western Galapagos [*Munro and Rowland*, 1996]. Considering the presence of the large intrusive complex of the Alizés volcano, we hypothesize that the vertical collapse of the Grand Brûlé would correspond to a discrete event (i.e., collapse of 100–150 m of a 7-km-wide area) simultaneously to the long-term downward motion of this dense body. This model differs from *Walker* [1988] as the dense body is related to an ancient volcano (i.e., the Alizés) rather than an active volcano. The occurrence of the vertical collapse of the Grand Brûlé potentially destabilized the adjacent submarine flank and initiated debris avalanches which deposits could correspond to the youngest unit described by *Oehler* [2005].

[44] The Enclos caldera was interpreted in different ways: (1) an independent structure resulting from successive syneruptive collapses [*Bachèlery*, 1981]; (2) the upper part of the scar of a large flank landslide [*Duffield et al.*, 1982; *Labazuy*, 1991; *Gillot et al.*, 1994]; and (3) a summit deformation initiated by the slide of the Grand Brûlé [*Merle and Lénat*, 2003]. Although the occurrence of pyroclastic deposits around the Enclos indicates that the collapse was simultaneous to an explosive eruption [*Abchir et al.*, 1998], the volume of the deposit (around 0.5–1 km^3 [*Abchir et al.*, 1998]) is 1 order of magnitude lower than that of the Enclos caldera. This volume difference and the lack of any evidence of large submarine eruption suggest that the collapse of the Enclos does not result from the emptying of the magmatic reservoir. It has been recently proposed that the collapse of the Enclos has been caused by the lateral flow of the large summit hydrothermal system [*Merle and Lénat*, 2003]. This model, which can explain the collapse of a large structure without the same erupted volume faces one main problem. According to *Merle and Lénat* [2003], the deformation of the hydrothermal system was made possible by the lateral slide of the Grand Brûlé only, which headwall corresponds to the Grandes Pentes. Our data strongly suggest that this slide did not occur 4.5 ka ago and that the Grandes Pentes (domain C in Figure 8) are not restricted to the EGBS. Hence the deformation as the authors proposed cannot be applied to PdF. However, we follow *Merle and Lénat* [2003] on two points: (1) the deformation of the Grand Brûlé and the Enclos are linked and (2) the Enclos may result from the deformation of the summit hydrothermal system.

[45] Taking into account our results, we propose two distinct evolutions in which a first collapse initiated the second one. On the one hand, the collapse of the EGBS was initiated by the vertical collapse of the Grand Brûlé, due to the continuous downward drag of the Alizés intrusive complex. This event allowed the deformation of the hydrothermal system of the pre-Enclos volcano and subsequently the collapse of the summit zone. On the other hand, the collapse of the Enclos results from the deformation of the summit hydrothermal system under gravity, which also led to the development of the LSS and USS. This hypothesis is supported by analog models, which show that the spreading of a volcano related to a weak internal core entails the

coeval development of circumferential steep slopes on the volcano flanks and the vertical collapse of the summit zone [*Cecchi et al.*, 2005]. We propose that this event may have destabilized part of the eastern flank of PdF, which was continuously affected by a downward drag. The relationship between summit collapse and flank deformation has already been suggested for Tenerife [*Marti et al.*, 1997; *Hürlimann et al.*, 1999]. However, contrary to what geological data show for PdF, the summit collapse triggered a flank landslide at Tenerife.

4.4. Evolution of PdF During the Last 150 ka

[46] Considering the geological data and the interpretations presented above, we propose the following chronology for the last 150 ka of PdF. This date corresponds to the collapse of the Morne Langevin caldera [*Bachèlery and Mairine*, 1990]. According to *Oehler* [2005], this event corresponds to the first and most voluminous eastward destabilization of PdF, which spread over the Eastern Plateau. Afterward, the "recent" PdF (<150 ka) built up at a location west of the present active cone [*Bachèlery and Mairine*, 1990]. This edifice the shape and evolution of which are unknown suffered recurrent collapses between 60 and 45 ka, forming the Plaine des Sables calderas. Although the origin of the upper part of this caldera is still uncertain (vertical [*Bachèlery and Mairine*, 1990] versus lateral [*Duffield et al.*, 1982; *Gillot et al.*, 1994; *Oehler et al.*, 2004]), the debris avalanche deposits on the submarine Eastern Plateau indicate that the collapses were coeval to at least one major landslide [*Labazuy*, 1996; *Oehler*, 2005]. It is noteworthy that the northern limit of the landslide scar (the Ravine Ferdinand [*Merle and Lénat*, 2003]) is the only well-observed structure which is really continue in both the subaerial and submarine domains (Figure 1b).

[47] The pre-Enclos edifice built up at a location close to the present active cone. This volcano progressively spread above its weak hydrothermal core. The geometry of this low-strength body was likely controlled by preexisting structures such as the décollement level of previous slides, the magma chambers and the intrusion complexes. Two circumferential steep slopes (i.e., the LSS and USS) result from this deformation (Figure 13a). At the same time, the volcano was incised by deep valleys, among which is the Osmondes paleovalley in the east flank. The presence of the Osmondes paleovalley in the northern part of the EGBS, whose minimum age is of 18–19 ka clearly indicates that PdF did not subsequently suffered the slide of the Grand Brûlé as it was usually proposed. Around 4.5 ka ago, both the summit and east flank were cut by a vertical collapse that led to the development of the west and south parts of the EGBS; the present northern limit of EGBS consisting of the Osmondes paleovalley's northern flank (Figure 13b). This collapse event likely destabilized part of the coastal zone and the eastern submarine flank. The related debris avalanche covered the proximal Eastern Plateau as revealed by the age of a dredged sample (12 ka [*Labazuy*, 1996]) (Figure 1b) and flowed through the Chenal Vincent down to the Râlé Poussé [*Labazuy*, 1996; *Oehler*, 2005]. The coastal and submarine origin of the debris avalanche could explain the absence of clear continuity between the subaerial and submarine domains. Similar or larger submarine destabilizations occurred during the recent evolution of PdF. However, their importance was underestimated until recently. According to *Oehler* [2005], their deposits cover around half of the surface of the submarine flanks.

[48] The lack of lava flows intercalated in the alluvial sediments of the Osmondes paleoriver could suggest the existence of a relief between the eruption zone and the valley. This barrier disappeared when the Plaine des Osmondes collapsed (Figure 13c). Even though the age of this event is unknown, a vertical mode of deformation can be reasonably inferred. Indeed, the width of the lowest part of the Plaine des Osmondes (i.e., between the Piton de Crac and the Bois Blanc scarp) is more than twice lower than at its maximum width. Such a geometry disagrees with the classic landslide scar. After the collapse of the Plaine des Osmondes, the lava flowed through the Osmondes paleovalley and completely counterbalanced the erosion process that developed since tens of thousands of years ago.

[49] One of the most striking features of this 150 ka lasting evolution is the persistence of the NE and SE rift zones, despite the recurrent collapses and landslides, and the eastward migration of the eruptive center. This contrasts with the general observations made on other volcanoes, where flank destabilizations or large changes of the topography induce a reorganization of the rift zones [*Tibaldi*, 2003; *Walter and Troll*, 2003; *Acocella and Tibaldi*, 2005]. The persistence of both rift zones, which contributed to form the NE and SE submarine plateaus (Figure 1), suggests that their location and development are controlled by deep sources. *Walter et al.* [2006] proposed that the NE and SE rift zones results from the intermittent eastward spreading of PdF. However, the undeformed marine sedimentation east of the island [*de Voogd et al.*, 1999] and the lack of anticline structures in the bathymetry [*Oehler et al.*, 2005] suggest that PdF did not experience such a deformation.

[50] In what follows we put forward two different features, which could have played a key role in the development of the stable rift zones:

[51] 1. It has been recently shown that the main structures of La Réunion volcanoes are parallel to the oceanic lithospheric structures, suggesting a control of the crustal discontinuities in the tectonomagmatic evolution [*Michon et al.*, 2007]. The persistence of both rift zones could then result from a control of the crustal structures.

[52] 2. The lower part of the NE and SE rift zones, the submarine parts included, are located north and south of the Alizés intrusive complex, respectively. Assuming that a continuous downward drag related to a dense body entails a diffuse extension in its vicinity, we hypothesize that the dike lateral migration in the NE and SE flanks of PdF was controlled by this stress field. In a certain way, this model presents similarities with that of *Walter et al.* [2005] in which the gravity-driven flank movement leads to a radial adjacent extension controlling the magma lateral migration and the development of two concentric curved rift zones. Whatever the model, the rarity of the eruptions along the rift zone lower parts, i.e., outside the Enclos caldera (only 3% of the eruptions since the 18th century [*Villeneuve*, 2000]) and their fan-shaped geometry suggest the existence of a diffuse stress field, which does not efficiently favor the magma lateral migration on the flanks. Despite the slight lateral propagation of the magma intrusions along the rift zones, the combined effect of the summit frequent intrusion

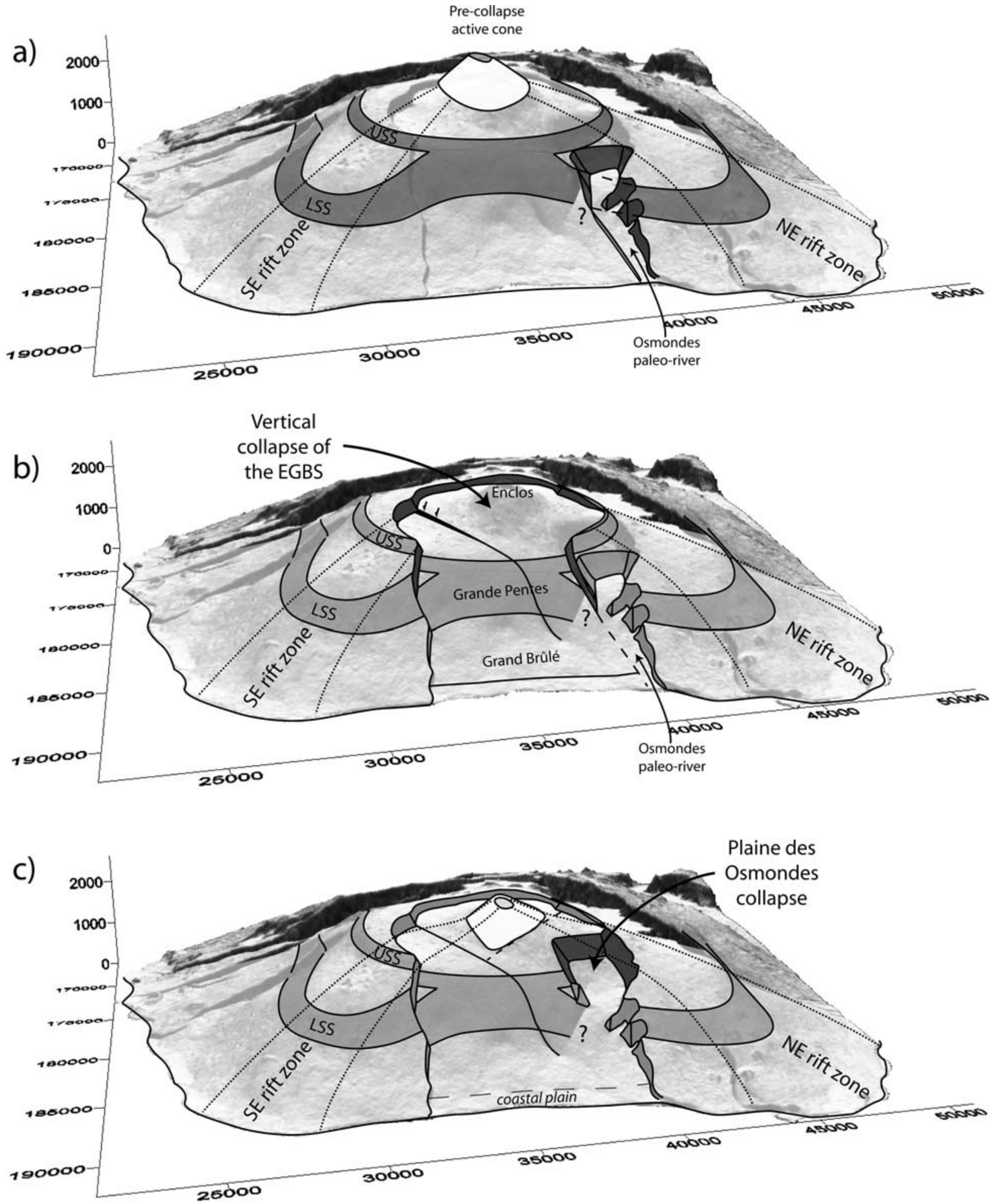

Figure 13. Chronology of the geological events that marked the last 60 ka of PdF. (a) After the collapse of the Plaine des Sables caldera, two circumferential steep slopes developed on the north, east, and south flanks. Simultaneously, the Osmondes paleovalley incised the east flank. (b) Around 4.5 ka ago, the EGBS collapsed vertically. The Osmondes paleovalley was likely separated from the magmatic system by a barrier as no lava flow is intercalated in the alluvial sediments drilled in the lower part of the valley. (c) The collapse of the Plaine de Osmondes, which occurred after 4.5 ka, allowed the lava to flow through the Osmondes paleovalley.

and the very likely existence of a décollement level within the edifice (the top of the Alizés volcano) potentially favored the large eastward destabilizations, which occurred between 45 and 60 ka.

5. Conclusions

[53] Our study of PdF aimed at the understanding of the present-day edifice morphology, which recorded the most recent geological events. The analysis was focused on the steep slope zones located on the volcano flanks and the Enclos-Grand Brûlé structure, which cut the upper part and eastern flank of the edifice. We took into account the different available geological data and tested them against the potential processes, which may form both structures. We presented new results allowing a reappraisal of the recent evolution of Piton de la Fournaise.

[54] 1. The steep slope zones were previously interpreted as slide headwalls and/or caldera margins [e.g., *Bachèlery*, 1981; *Oehler et al.*, 2004]. However, their distribution and morphological characteristics suggest that they form two independent circumferential structures (the USS and LSS), which coalesce in the east flank. Their development is interpreted as resulting from the spreading of the pre-Enclos volcano above a weak hydrothermal core.

[55] 2. The Enclos-Grand Brûlé structure (a U-shaped structure) is formed by the Enclos depression, the Grandes Pentes, and the Grand Brûlé. Although the Enclos origin was a matter of great debate during the last decades (vertical versus lateral collapse), there had been a general agreement on a sliding origin of the Grand Brûlé [*Bachèlery*, 1981; *Duffield et al.*, 1982; *Gillot et al.*, 1994; *Labazuy*, 1996; *Merle and Lénat*, 2003; *Oehler et al.*, 2004]. However, our analysis of the continuity of different slope domains inside and outside the EGBS suggests that the Grand Brûlé results from a mainly vertical collapse instead of a giant landslide. The entire EGBS subsequently underwent a vertical collapse ~4.5 ka ago. Among different potential sources for the collapse of the Grand Brûlé, we prefer the continuous downward drag induced by the dense intrusive complex of the Alizés volcano. Following *Merle and Lénat* [2003], we propose that the collapse of the Enclos was caused by the deformation of the hydrothermal system of the pre-Enclos volcano. Despite the clear continuity between the Enclos and the Grand Brûlé, which suggests a close relationship in the development of both structures, their chronology and exact links remain poorly understood.

[56] **Acknowledgments.** The authors want to thank Scott Rowland, Laszlo Keszthelyi, Jim Kauahikaua, Jurgen Neuberg, and the Associate Editor Susan Sakimoto, whose comments considerably improved an earlier version of the paper. Thanks are also given to Jean-Francois Lénat, Olivier Merle, and Philippe Labazuy for the stimulating discussion about Piton de la Fournaise. This work was partly funded by the BQR 2004 of the University of La Réunion provided to L.M. This is IPGP contribution 2299.

References

Abchir, M. A., S. M. Semet, G. Boudon, P. Ildefonse, P. Bachèlery, and R. Clocchiati (1998), Huge hydrothermal explosive activity on Piton de la Fournaise, Réunion Island: The Bellecombe ash member, 2700 BC, in *Volcanic Risk—The European Laboratory Volcanoes*, *Publ. EUR 18161 EN*, edited by R. Casal et al., pp. 447–455, Eur. Comm., Brussels.

Acocella, V., and A. Tibaldi (2005), Dike propagation driven by volcano collapse: A general model tested at Stromboli, Italy, *Geophys. Res. Lett.*, *32*, L08308, doi:10.1029/2004GL022248.

Annen, C., J.-F. Lénat, and A. Provost (2001), The long-term growth of volcanic edifices: Numerical modelling of the role of dyke intrusion and lava-flow emplacement, *J. Volcanol. Geotherm. Res.*, *105*, 263–289.

Bachèlery, P. (1981), Le Piton de la Fournaise (Ile de la Réunion). Etude volcanologique, structurale et pétrologique, Ph.D. thesis, 215 pp., Univ. Clermont-Ferrand II, Clermont-Ferrand, France.

Bachèlery, P. and P. Mairine (1990), Evolution volcano-structurale du Piton de la Fournaise depuis 0.53 Ma, in *Le Volcanisme de la Réunion, Monographie*, edited by J.-F. Lénat, pp. 213–242, Cent. de Rech. Volcanol., Clermont-Ferrand, France.

Bachèlery, P., B. Robineau, M. Courteaud, and C. Savin (2003), Avalanches de débris sur le flanc occidental du volcan-bouclier Piton des Neiges (Réunion), *Bull. Soc. Geol. Fr.*, *174*, 125–140.

Benderitter, Y. (1990), Etudes géoélectriques et électromagnétiques à la Réunion, in *Le Volcanisme de la Réunion, Monographie*, edited by J.-F. Lénat, pp. 29–42, Cent. de Rech. Volcanol., Clermont-Ferrand, France.

Borgia, A. (1994), Dynamic basis of volcanic spreading, *J. Geophys. Res.*, *99*, 17,791–17,804.

Borgia, A., J. Burr, W. Montero, L. D. Morales, and G. E. Alvarado (1990), Fault propagation folds induced by gravitational failure and slumping of the Central Costa Rica Volcanic Range: Implications for large terrestrial and Martian volcanic edifices, *J. Geophys. Res.*, *95*, 14,357–14,382.

Briole, P., P. Bachèlery, B. McGuire, J. Moss., J. C. Ruegg, and P. Sabourault (1998), Deformation at Piton de la Fournaise: Evolution of the monitoring techniques and knowledge acquired in the last five years, in *Volcanic Risk—The European Laboratory Volcanoes*, *Publ. EUR 18161 EN*, edited by R. Casal et al., pp. 467–474, Eur. Comm., Brussels.

Camoin, G. F., L. F. Montaggioni, and C. J. R. Braithwite (2004), Late glacial to post glacial sea levels in the western Indian Ocean, *Mar. Geol.*, *206*, 119–146.

Cantagrel, J. M., N. O. Arnaud, E. Ancochea, J. M. Fúster, and M. J. Huertas (1999), Repeated debris avalanches on Tenerife and genesis of Las Cañadas caldera wall (Canary Islands), *Geology*, *27*, 739–742.

Cecchi, E., B. van Wyk de Vries, and J. M. Lavest (2005), Flank spreading and collapse of weak-cored volcanoes, *Bull. Volcanol.*, *67*, 72–91, doi:10.1007/s00445-004-0369-3.

Clague, D. A., and R. P. Denlinger (1994), Role of olivine cumulates in destabilizing the flanks of Hawaiian volcanoes, *Bull. Volcanol.*, *56*, 425–434, doi:10.1007/s004450050052.

Coudray, J., P. Mairine, E. Nicolini, and J. M. Clerc (1990), Approches hydrogéologique, in *Volcanic Risk—The European Laboratory Volcanoes*, *Publ. EUR 18161 EN*, edited by R. Casal et al., pp. 307–355, Eur. Comm., Brussels.

Courteaud, M. (1996), Etude des structures géologiques et hydrogéologiques du Massif de la Fournaise par la méthode audiomagnétotellurique, Ph.D. thesis, 212 pp., Univ. Réunion, Saint Denis, France.

Cullen, A., A. R. MacBirney, and R. D. Rodgers (1987), Structural controls on the morphology of Galapagos shields, *J. Volcanol. Geotherm. Res.*, *34*, 143–151.

Day, S. J., S. I. N. Heleno da Silva, and J. F. B. D. Fonseca (1999), A past giant lateral collapse and present-day flank instability of Fogo, Cape Verde Islands, *J. Volcanol. Geotherm. Res.*, *94*, 191–218.

Delaney, P. T., R. P. Denlinger, M. Lisowski, A. Miklius, P. G. Okubo, A. T. Okamura, and M. K. Sako (1998), Volcanic spreading at Kilauea, 1976–1996, *J. Geophys. Res.*, *103*, 18,003–18,023.

Deniel, C., G. Kieffer, and J. Lecointre (1992), New ^{230}Th and ^{238}U and ^{14}C age determinations from Piton des Neiges volcano, Reunion. A revised chronology for the differentiated series, *J. Volcanol. Geotherm. Res.*, *51*, 253–267.

de Voogd, B., S. Pou Palomé, A. Hirn, P. Charvis, J. Gallart, D. Rousset, J. Dañobeitia, and H. Perroud (1999), Vertical movements and material transport during hotspot activity: seismic reflection profiling offshore La Réunion, *J. Geophys. Res.*, *104*, 2855–2874.

Donnadieu, F., and O. Merle (2001), Geometrical constraints of the 1980 Mount St. Helens intrusion from analogue models, *Geophys. Res. Lett.*, *28*, 639–642.

Duffield, W. A., L. Stieltjes, and J. Varet (1982), Huge landslide blocks in the growth of Piton de la Fournaise, La Réunion, and Kilauea Volcano, Hawaii, *J. Volcanol. Geotherm. Res.*, *12*, 147–160.

Gillot, P.-Y., and P. E. Nativel (1989), Eruptive history of the Piton de la Fournaise Volcano, Reunion Island, Indian Ocean, *J. Volcanol. Geotherm. Res.*, *36*, 53–65.

Gillot, P.-Y., J.-C. Lefèvre, and P. E. Nativel (1994), Model for the structural evolution of the volcanoes of Réunion island, *Earth Planet. Sci. Lett.*, *122*, 291–302.

Hanebuth, T., K. Statteger, and P. M. Grootes (2000), Rapid flooding of the Sunda Shelf: A late glacial sea level record, *Science*, *288*, 1033–1035.

Hürlimann, M., E. Turon, and J. Marti (1999), Large landslides triggered by caldera collapse events in Tenerife, Canary Islands, *Phys. Chem. Earth A*, *24*, 921–924.

Hürlimann, M., J. Marti, and A. Ledesma (2004), Morphological and geological aspects related to large slope failures on oceanic islands: The huge La Orotava landslides on Tenerife, Canary Islands, *Geomorphology*, *62*, 143–158.

Kaneko, T., A. Yasuda, T. Shimano, S. Nakada, T. Fujii, T. Kanazawa, A. Nishizawa, and Y. Matsumoto (2005), Submarine flank eruption preceding caldera subsidence during the 2000 eruption of Miyakejima Volcano, Japan, *Bull. Volcanol.*, *67*, 243–253, doi:10.1007/s00445-004-0407-1.

Labazuy, P. (1991), Instabilités au cours de l'évolution d'un édifice volcanique en domaine océanique: Le Piton de la Fournaise (ıcirc;le de la Réunion), Ph.D. thesis, 252 pp., Univ. Clermont Ferrand II, Clermont Ferrand, France.

Labazuy, P. (1996), Recurrent landslides events on the submarine flank of Piton de la Fournaise volcano (Réunion Island), in *Volcano Instability on the Earth and Other Planets*, edited by W. J. McGuire, A. P. Jones, and J. Neuberg, *Geol. Soc. Spec. Publ.*, *110*, 293–305.

Le Friant, A., C. L. Harford, C. Deplus, G. Boudon, R. S. J. Sparks, R. A. Herd, and J. C. Komorowski (2004), Geomorphological evolution of Montserrat (West Indies): Importance of flank collapse and erosional processes, *J. Geol. Soc.*, *161*, 147–160.

Lénat, J.-F. and P. Bachèlery (1987), Dynamics of magma transfer at Piton de la Fournaise volcano (Réunion Island, Indian Ocean), in *Modeling of Volcanic Processes*, edited by C.-Y. King and R. Scarpa, pp. 57–72, F. Vieweg, Brauschweig, Germany.

Lénat, J.-F., P. Bachèlery, A. Bonneville, and A. Hirn (1989), The beginning of the 1985–1987 eruptive cycle at Piton de la Fournaise (la Réunion): New insights in the magmatic and volcano-tectonic systems, *J. Volcanol. Geotherm. Res.*, *36*, 209–232.

Lénat, J.-F., D. Fitterman, D. B. Jackson, and P. Labazuy (2000), Geoelectrical structure of the central zone of Piton de la Fournaise volcano (Réunion), *Bull. Volcanol.*, *62*, 75–89.

Lénat, J.-F., B. Gibert-Malengreau, and A. Galdéano (2001), A new model for the evolution of the volcanic island of Réunion (Indian Ocean), *J. Geophys. Res.*, *106*, 8646–8663.

Lerebour, P., J.-P. Rançon, and T. Augé (1989), The Grand Brûlé exploration drilling: New data on the deep framework of the Piton de la Fournaise volcano. part 2: Secondary minerals, *J. Volcanol. Geotherm. Res.*, *36*, 129–137.

Lipman, P. W. (1997), Subsidence of ash-flow calderas: Relation to caldera size and magma chamber geometry, *Bull. Volcanol.*, *59*, 198–218.

MacDonald, G. A. (1965), Hawaiian calderas, *Pac. Sci.*, *19*, 320–334.

MacDonald, G. A. (1972), *Volcanoes*, 510 pp., Prentice-Hall, Englewood Cliffs, N. J.

MacDougall, I. (1971), The geochronology and evolution of the young island of Réunion, Indian Ocean, *Geochim. Cosmochim. Acta*, *35*, 261–270.

MacPhie, J., G. P. L. Walker, and R. L. Christiansen (1990), Phreatomagmatic and phreatic fall and surge deposits from explosions at Kilauea volcano, Hawaii, 1790 A.D.: Keanakakoi Ash Member, *Bull. Volcanol.*, *52*, 334–354.

Mairine, Ph., and P. Bachèlery (1997), Un grand épisode érosionnel dans l'histoire ancienne du Piton de la Fournaise (ıcirc;le de la Réunion), *C. R. Acad. Sci.*, *325*, 243–249.

Malengreau, B., J.-F. Lénat, and J.-L. Froger (1999), Structure of the Reunion Island (Indian Ocean) inferred from the interpretation of gravity anomalies, *J. Volcanol. Geotherm. Res.*, *88*, 131–146.

Mark, R. K., and J. Moore (1987), Slopes of the Hawaiian ridge, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 101–107.

Marti, J., M. Hürlimann, G. J. Ablay, and A. Gudmundsson (1997), Vertical and lateral collapses on Tenerife (Canary Islands) and other volcanic ocean islands, *Geology*, *25*, 879–882.

Merle, O., and A. Borgia (1996), Scaled experiments of volcanic spreading, *J. Geophys. Res.*, *101*, 13,805–13,817.

Merle, O., and J. Lénat (2003), Hybrid collapse mechanism at Piton de la Fournaise volcano, Reunion Island, Indian Ocean, *J. Geophys. Res.*, *108*(B3), 2166, doi:10.1029/2002JB002014.

Michon, L., F. Saint-Ange, P. Bachelery, N. Villeneuve, and T. Staudacher (2007), Role of the structural inheritance of the oceanic lithosphere in the magmato-tectonic evolution of Piton de la Fournaise volcano (La Réunion Island), *J. Geophys. Res.*, *112*, B04205, doi:10.1029/2006JB004598.

Moore, J. G. (1964), Giant submarine landslides on the Hawaiian ridge, *U.S. Geol. Surv. Prof. Pap.*, *501-D*, 95–98.

Moore, J. G., and H. L. Krivoy (1964), The 1962 flank eruption of Kilauea volcano and structure of the east rift zone, *J. Geophys. Res.*, *69*, 2033–2045.

Moore, J. G., and M. K. Mark (1992), Morphology of the Island of Hawaii, *GSA Today*, *2*, 257–262.

Munro, D. C., and S. K. Rowland (1996), Caldera morphology in the western Galapagos and implications for volcano eruptive behavior and mechanisms of caldera formation, *J. Volcanol. Geotherm. Res.*, *72*, 85–100.

Naumann, T., and D. Geist (2000), Physical volcanology and structural development of Cerro Azul Volcano, Isabela Island, Galapagos: Implications for the development of Galapagos-type shield volcanoes, *Bull. Volcanol.*, *61*, 497–514.

Oehler, J.-F. (2005), Les déstabilisations de flanc des volcans de l'Ile de La Réunion (Océan Indien): Mise en évidence, implications et origins, Ph.D. thesis, 422 pp., Univ. Clermont Ferrand II, Clermont Ferrand, France.

Oehler, J.-F., P. Labazuy, and J.-F. Lénat (2004), Recurrence of major flank landslides during the last 2–Ma-history of Reunion Island, *Bull. Volcanol.*, *66*, 585–598, doi:10.1007/s00445-004-0341-2.

Oehler, J. F., B. Van Wyk de Vries, and P. Labazuy (2005), Landslides and spreading of oceanic hot-spot and arc shield volcanoes and low strength layers (LSLs): An analogue modelling approach, *J. Volcanol. Geotherm. Res.*, *144*, 169–189.

Rançon, J.-P., P. Lerebour, and T. Augé (1989), The Grand Brûlé exploration drilling: New data on the deep framework of the Piton de la Fournaise volcano. part 1: Lithostratigraphic units and volcanostructural implications, *J. Volcanol. Geotherm. Res.*, *36*, 113–127.

Roche, O., T. H. Druitt, and O. Merle (2000), Experimental study of caldera formation, *J. Geophys. Res.*, *105*, 395–416.

Rowland, S. K. (1996), Slopes, lava flow volumes, and vent distributions on Volcán Fernandina, Galapagos Islands, *J. Geophys. Res.*, *101*, 27,657–27,672.

Rowland, S. K., and H. Garbeil (2000), Slopes of oceanic basalt volcanoes, in *Remote Sensing of Active Volcanism*, *Geophys. Monogr. Ser.*, vol. 116, edited by P. J. Mouginis-Mark, J. A. Crisp, and J. H. Fink, pp. 223–247, AGU, Washington, D. C.

Sapin, M., A. Hirn, J.-C. Lépine, and A. Nercessian (1996), Stress, failure and fluid flow deduced from earthquakes accompanying eruptions at Piton de la Fournaise volcano, *J. Volcanol. Geotherm. Res.*, *70*, 145–167.

Schumm, S. A. (1993), River response to baselevel change: Implications for sequence stratigraphy, *J. Geol.*, *101*, 279–294.

Stearns, H. T., and W. O. Clark (1930), Geology and water resources of the Kau District, Hawaii, *U.S. Geol. Surv. Water Supply Pap.*, *616*, 194 pp.

Stearns, H. T., and G. A. MacDonald (1946), Geology and ground-water resources of the Island of Hawaii, *Bull. 9*, 363 pp., Hawaii Div. of Hydogr., Honolulu.

Stieltjes, L., P. Bachèlery, A. Bacquet, A. de Villèle, M. Kasser, J.-F. Lénat, and P. Moutou (1986), Carte des coulées historiques du volcan de la Fournaise (Ile de la Réunion, Océan Indien), 43 pp., BRGM, Orléans, France.

Tibaldi, A. (2001), Multiple sector collapses at Stromboli volcano, Italy: How they work, *Bull. Volcanol.*, *63*, 112–125.

Tibaldi, A. (2003), Influence of cone morphology ion dykes, Stromboli, Italy, *J. Volcanol. Geotherm. Res.*, *126*, 79–95.

Tort, A., and A. Finizola (2005), Structural survey of Misti volcanic cone (southern Peru) combining elliptical Fourier function analysis of the volcano morphology and self-potential measurements, *J. Volcanol. Geotherm. Res.*, *141*, 283–297.

Villeneuve, N. (2000), Apports multi-sources à une meilleure compréhension de la mise en place des coulées de lave et des risques associés au Piton de la Fournaise: Géomorphologie quantitative en terrain volcanique, Ph.D. thesis, 378 pp., Inst. de Phys. du Globe de Paris, Paris.

Walker, G. P. L. (1988), Three Hawaiian calderas: An origin through loading by shallow intrusions, *J. Geophys. Res.*, *93*, 14,773–14,784.

Walter, T. R., and V. R. Troll (2003), Experiments on rift zone evolution in unstable volcanic edifices, *J. Volcanol. Geotherm. Res.*, *127*, 107–120.

Walter, T. R., V. R. Troll, B. Cailleau, A. Belousov, H.-U. Schmincke, F. Amelung, and P. v. d. Bogaard (2005), Rift zone reorganization through flank instability in ocean island volcanoes: An example from Tenerife, Canary Islands, *Bull. Volcanol.*, *67*, 281–291, doi:10.1007/s00445-004-0352-z.

Walter, T. R., A. Klügel, and S. Münn (2006), Gravitational spreading and formation of new rift zones on overlapping volcanoes, *Terra Nova*, *18*, 26–33.

L. Michon and F. Saint-Ange, Laboratoire GéoSciences Réunion, Université de la Réunion, BP 7151, F-97715 Saint Denis cedex 9, France. (laurent.michon@univ-reunion.fr)