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Volcano core collapse triggered by regional faulting

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Abstract

It has been proposed recently that large-scale caldera-like structures may result from the viscous lateral spreading of a hydrothermal system located in the central part of volcanic edifices. For this to occur, an unbutressed boundary resulting in reduced lateral stresses must exist or be created not far from the hydrothermal system. Different ways to achieve such a weak lateral confinement may be proposed. In this paper, we show that regional tectonics may provide an unconfined boundary if a normal fault lowers part of the volcanic edifice. A natural example is the island of Nuku Hiva in the Marquesas Archipelago (French Polynesia) for which geological data strongly suggest that a regional fault oriented N75°E is responsible for the present-day shape of the island, i.e. an ellipse cut in the middle by a straight side along which two nested calderas are breached. Analogue modelling shows the mechanical consistency of the model and reveals that very low velocity faulting is needed for the formation of a caldera-like structure.

Keywords: volcanoes; hydrothermal system; weak confinement; spreading

1. Introduction

The formation of caldera structures is generally ascribed to the emptying of a magma reservoir located at depth and ensuing roof collapse (e.g., Walker, 1984). This mechanism has been established for many edifices, especially andesitic and rhyolitic volcanoes. However, such a mechanism is unlikely for some large volcanoes where the emission of magma during single or short successive events is too small to match the observed collapse. In that case, the caldera cannot be directly linked to magma withdrawal and an alternative mechanism has to be found. According to recent studies showing that hydrothermal alteration may weaken the edifice and trigger sector collapse (Walker, 1992; Lopez and Williams, 1993; Day, 1996; Voight and Elsworth, 1997; Van Wyk de Vries et al., 2000; Reid et al., 2001; Cecchi et al., 2005), it has been proposed that caldera-like structures may result from the lateral spreading of hydrothermally deeply altered rocks in the central part of the edifice, followed by roof vertical collapse (Merle

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and Lénat, 2003). For such a process to occur, an unbutressed boundary should be present reducing lateral stresses and allowing the hydrothermal system to flow laterally in that direction. In Piton de la Fournaise volcano (Réunion Island), the Grand Brûlé slide may have provided that free boundary triggering the collapse of the Enclos caldera.

The basic idea of the model presented herein is that any deformation leading to an unbutressed boundary close enough to a large central hydrothermal system may be followed by vertical collapse of the central part of the edifice into the void left by the departing hydrothermal complex. In this paper, we show that regional tectonics may drastically change the boundary conditions around a volcano, reducing lateral stresses and promoting escape of hydrothermal system in a way similar to that observed in Piton de la Fournaise.

2. Nuku Hiva Island

Located in the South Pacific ocean, the Marquesas Archipelago is the northernmost volcanic island chain of French Polynesia and is limited to the north and south by two major fracture zones trending N75°E: the Galapagos and Marquesas Fracture Zones, respectively (Fig. 1). Its eight main volcanic islands form a linear chain 100 km wide and 350 km long, resulting from short-lived hotspot volcanism, which lasted from 5.5 Ma to 0.4 Ma. The oceanic crust, on which the archipelago was built, is of Eocene age (53–49 Ma) and is clearly thickened with a 15–20 km deep Moho (Filmer et al., 1993; Caress et al., 1995).

Nuku Hiva Island is located in the central part of the Marquesas and is the largest shield volcano of this linear chain with a dimension of 30 × 20 km and a surface area of about 380 km^2. It is characterised by two hemispheric nested calderas breached on the sea, which have been identified from their well-preserved ramparts (Chubb, 1930; Brousse and Guille, 1978; Le Dez et al., 1996). The northern (outer) one is 10 km wide and the southern (inner) one 16 km wide (Fig. 2). Inside the inner caldera, a smaller and almost circular depression opened towards the sea in Taiohae Bay is considered to result from rapid erosion of strombolian layers (Maury et al., 2005). The

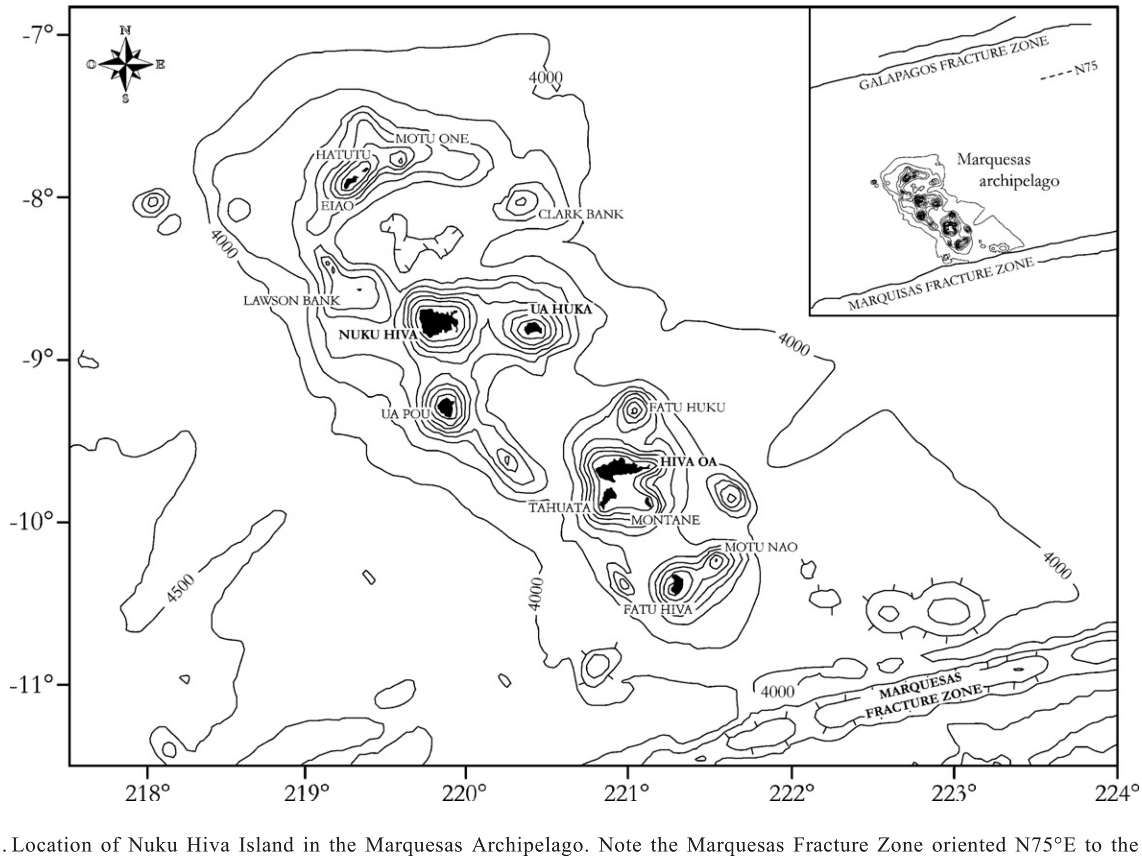

Fig. 1. Location of Nuku Hiva Island in the Marquesas Archipelago. Note the Marquesas Fracture Zone oriented N75°E to the south of the Archipelago. The insert shows the location of the parallel Galapagos Fracture Zone. Bathymetry taken from Wolfe et al. (1994).

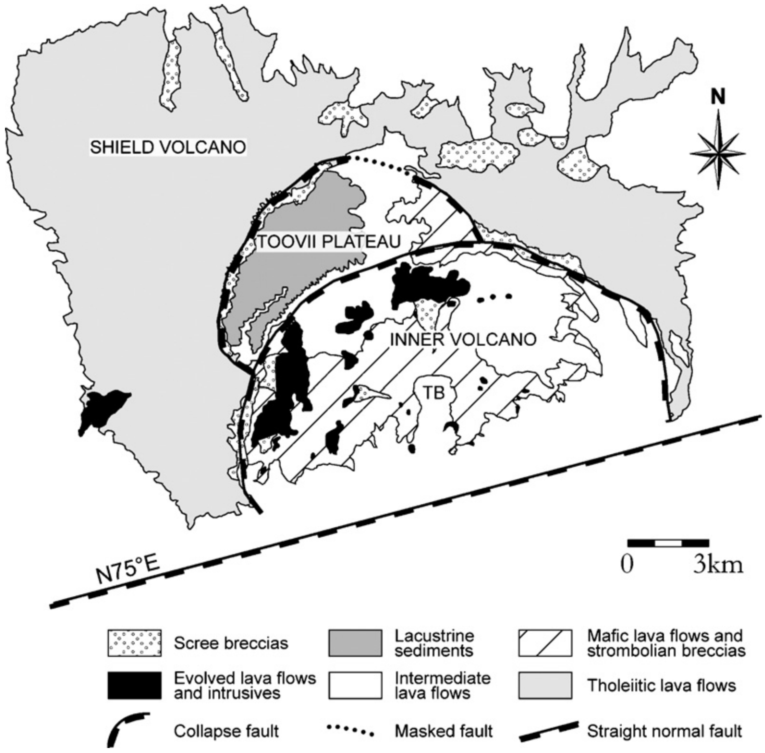

Fig. 2. Sketch map of Nuku Hiva showing the two nested calderas breached at the sea. Note the straight N75°E southward limit of the island interpreted as resulting from normal faulting. The dashed line south of the Taiohae Bay (TB) denotes the trend of trachytic domes and plugs. Modified and simplified from Savanier et al. (2005).

two nested calderas make it possible to define three geological zones in Nuku Hiva Island:

(1) the Tekao outer shield volcano, which is exclusively composed of a more than 1000-m thick tholeiitic lava flow sequence, emplaced between 4.5 and 4.0 Ma (Le Dez et al., 1996; Maury et al., 2005) and is clearly offset by the two caldera ring faults;
(2) the Toovii plateau, which represents the rather flat floor of the outer (northern) caldera. It exposes a part of the inner volcano hawaiitic/mugearitic lava flow sequence (up to 800 m thick), dated to 4.0–3.9 Ma and overlain by volcaniclastic lacustrine sediments. Both units are cut by the inner caldera ring fault;
(3) the inner Taiohae volcano, which was built between 4.0 and 3.6 Ma (Maury et al., 2005) in the central part of the inner caldera. It is composed of three main successive units: (i) basaltic lava flows and associated strombolian breccias which cover 25% of the surface of the volcano, (ii) prominent hawaiitic and mugearitic lava flows 4.0 to 3.8 Ma old, spread over 50% of the edifice as well as over the floor of the outer caldera, and finally (iii) numerous trachytic domes and flows emplaced between 3.9 and 3.6 Ma (Legendre et al., 2005). Its total exposed volume is ca. 60 km^3, i.e. less than that of the initial hemispheric caldera-like depressions, our minimal estimate of which is 100 km^3.

The main caldera collapse event of the Tekao shield volcano occurred at 4.05±0.10 Ma (Le Dez et al., 1996), i.e. just before the end of its building and the start of the edification of the inner Taiohae volcano at 4.0 Ma. These two dates are consistent with all the presently available K–Ar and Ar–Ar data (Maury et al., 2005), which overlap around 4.0 Ma, suggesting that there was no significant temporal gap between the end of the Tekao activity and the start of that of the Taiohae volcano. Field relationships (Fig. 2 and Savanier et al., 2005) show that the 4.05±0.10 Ma collapse event involved the formation of the two ring faults. Indeed, the Taiohae volcano grew at 4.0 Ma in the centre of the inner caldera. Its hawaiitic and mugearitic lavas flowed into the inner caldera, filled up the northern part of it and flooded the floor of the outer caldera (the Toovii plateau), locally sealing the outer caldera ring fault. However, field evidence also indicates that the inner (southern) caldera fault was re-activated after 3.9 Ma as shown by the clear offset along that fault of several hawaiitic and mugearitic lava flows, the youngest of which being dated at 3.90±0.06 Ma (Maury et al., 2005), in the western and central parts of the Toovii plateau. North of this fault, the altitude of the base of the Toovii lacustrine deposits is 790–810 m while south of the inner caldera fault it outcrops at 710 m, indicating thus an offset of 80–100 m during the second collapse event. In the northern part of the Toovii plateau, the hawaiitic and mugearitic flows seal the outer caldera fault without being offset, a feature which suggests that the outer fault was not re-activated after the main collapse event.

In search of a model to explain the formation of these two giant calderas on Nuku Hiva Island, it should be kept in mind that the lack of geophysical data and drillings onshore and offshore makes any model hypothetical and still to be confirmed. Starting from scratch, the classic model of caldera formation, however, is rather unlikely. Geological data do not support the occurrence of large individual eruptions consistent with the emptying of a reservoir large enough to explain the collapse of such giant calderas. Moreover, field studies as well as K–Ar and Ar–Ar dating show that the formation of the two nested calderas at 4.05±0.10 Ma pre-dated the eruption of the largest volume of intermediate/evolved lava exposed on the island, i.e. the ca. 30 km^3 hawaiitic to mugearitic flows of the inner Taiohae volcano, which poured into the calderas between 4.0 and 3.8 Ma and locally sealed the northern wall of the outer one. Such a sequence of events is not consistent with caldera formation by magma withdrawal.

The competing model of giant landslides on low angle failure planes should be envisaged. The geometry of those landslides as described on other islands as Fogo volcano (Cape Verde Islands) or Tenerife (Canary Islands) does not match that observed on Nuku Hiva Island (e.g. Carracedo, 1994). These landslides always display in their distal part (i) a channel with two more or less parallel limits and (ii) a slope going down from the main eruptive center to the sea. As opposed to this landslide geometry, the nested calderas in Nuku Hiva Island are breached directly on the sea but at an altitude significantly higher than sea level and the outer one displays a flat floor (the Toovii plateau) on which horizontal strata of lacustrine sediments were deposited. These sediments rest on hawaiitic and mugearitic lava flows which partly filled up the outer caldera where they display a rather constant thickness and cover wide areas, whereas in the southeastern part of the island similar flows were usually channelled into valleys. We interpret these features as suggesting that the floor of the outer caldera was already rather flat before the emplacement of these flows. The El Golfo collapse structure on El Hierro, Canary Islands, also displays a flat floor but at the difference of Nuku Hiva it is located near sea level (Carracedo et al., 1999), and possibly connected with a debris avalanche deposit offshore.

Likewise, no rift zones can be observed on Nuku Hiva Island making instability due to repeated dyke intrusions very unlikely (e.g. Dieterich, 1988; Carracedo, 1994; Walter and Schmincke, 2002; Walter et al., 2005). Furthermore, deformation due to volcano spreading has never revealed structures that mimic the large calderas of Nuku Hiva. Volcano spreading is mainly associated with the formation of leaf graben systems including sector collapse and strike–slip components along radial faults (e.g. Merle and Borgia, 1996; Van Wyk de Vries and Francis, 1997; Olher et al., 2005; Wooller et al., 2005).

3. The model

The following points are critical to provide a reasonable interpretation of the nested calderas:

(1) The rough shape of the island is that of an ellipse cut southward by a straight limit oriented N75°E (Fig. 2).
(2) Numerous trachytic domes and plugs are located along a N75°E trend, north of Taiohae Bay (Fig. 2; Legendre et al., 2005).
(3) At a larger scale, bathymetric records show a plurikilometric straight lineament oriented N75°E, which matches the southern limit of the island (Fig. 3).
(4) The N75°E direction is that of the Marquesas Fracture Zone which bounds the archipelago to the south. This major regional feature, as the Austral and Galapagos Fracture Zones and many others of same orientation, was associated with the activity of the Farallon oceanic ridge until 26 Ma in this part of the Pacific (Mammerickx et al., 1980; Sichoix and Bonneville, 1996; Sichoix et al., 1998; Bonneville and Sichoix, 1998; Munschy et al., 1998). During the growth and collapse of Nuku Hiva between 4.5 and 3.6 Ma, this fault system was no longer active as transform faults and only vertical adjustments may have occurred.

Put together, these data strongly suggest that the southern half of the initial island has subsided along a normal fault oriented N75°E. As an attempt to explain reactivation of the fault, one can note that thickening of the oceanic crust has been interpreted as resulting from

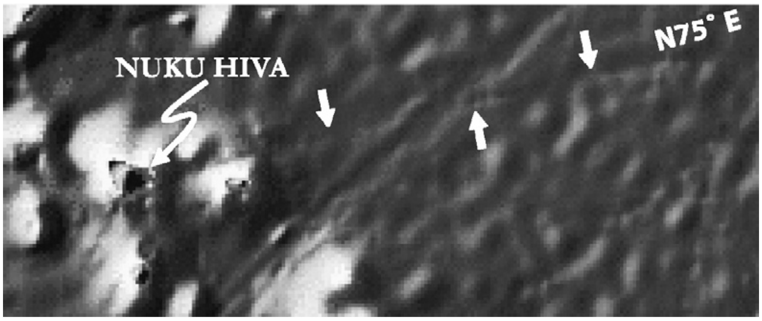

Fig. 3. Bathymetry around Nuku Hiva from Jules Verne Voyager. Despite the bad resolution, a N75°E lineament parallel to the Marquesas Fracture Zone is well visible and matches in the west the southern coast of the island.

recent magmatic underplating due to hotspot volcanism (Filmer et al., 1993; Caress et al., 1995; McNutt and Bonneville, 2000). Aside this bottom-loading process, top-loading also occurred due to volcano growth (Filmer et al., 1994; Wolfe et al., 1994). Bottom- and top-loading are accompanied by competing upward and downward deflection of the plate, respectively. Both upward and downward large-scale deflection of the plate can be locally adjusted by re-activation of previous vertical faults in the crust. Concerning our model, this makes no difference as what is needed is a vertical offset along the N75°E fault making the present-day emerged part of the volcanic edifice higher than the submerged one.

This hypothesis leads to propose an alternative model for the formation of the two large nested caldera-like structures of Nuku Hiva. Reducing lateral stresses at the vicinity of an active hydrothermal system may be achieved in different ways. It may arise from a giant landslide on the flanks of the volcano, as proposed at Piton de la Fournaise on Réunion Island (Merle and Lénat, 2003). In the case of Nuku Hiva, the re-activation of a regional fault oriented N75°E which cuts the island in the middle, not far away from the main eruptive vents, is an equally efficient way to provide a free boundary along which lateral stresses dramatically reduce. Lateral flow of the hydrothermal system towards the free boundary may occur, triggering the collapse of the caldera-like structure in the still emerged part of the island (Fig. 4). The fault is considered to be vertical as it belongs to the set of Farallon-related transform faults

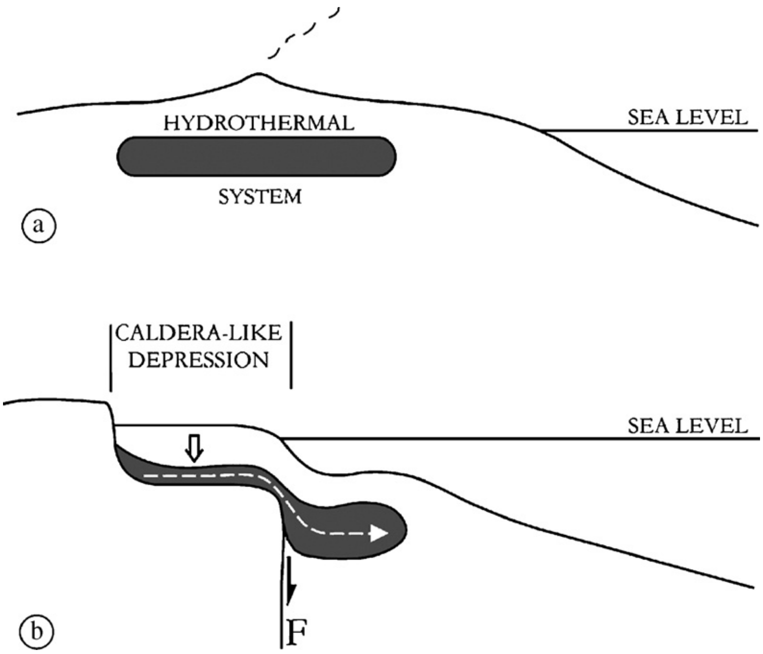

Fig. 4. Conceptual sketch of the model. The normal fault (F) which lowers the right compartment provides an unconfined boundary on the seaward side of the hydrothermal system, which spreads laterally.

of the Galapagos/Marquesas/Austral Fracture Zones group.

If the basic idea of the model is simple, it has several implications concerning the kinematic evolution of the deformation. Displacement along a fault may happen as a single catastrophic event, but a slow and continuous downthrow may also occur. Slow and continuous displacement which does not cut and split the hydrothermal system but extends it along the fault may also allow thinning of the hydrothermal system located on the stable part of the island. Likewise, the offset quantity necessary to allow lateral flow of the hydrothermal system is difficult to evaluate. As previously done for Piton de la Fournaise, a set of experiments has been conducted in order to check the mechanical consistency of the model and achieve a better understanding of its temporal evolution.

4. Experiments

Analogue materials are identical to those used in experiments on Piton de la Fournaise (Merle and Lénat, 2003), that is a dry sand/plaster mixture to simulate the brittle pile of volcanic rocks and silicone putty to simulate the clay/sulfate-rich hydrothermal system supposed to deform as a viscous fluid. Likewise, the process under consideration being identical to that inferred for the Piton de la Fournaise in Réunion Island, the scaling procedure is very close to that published by Merle and Lénat (2003) and is presented in the following section. Of course, geometric ratios are different and determined according to the dimension of the largest caldera in Nuku Hiva.

A positive residual gravimetric anomaly north of Taiohae Bay (Maury et al., 2005) suggests the occurrence at depth of a large intrusive complex. However, there is no available information on the location, depth and thickness of the fossil hydrothermal system which might have formed at its top. The unusually oxidised character of Taiohae volcano magmas, and their well-documented fractionation under high water pressures involving separation of kaersutitic/pargasitic amphibole, suggest deep interaction with water (Maury et al., 1978; Legendre et al., 2005). The corresponding magma reservoir must have been located at depths greater than 7 km to account for the stability of the abundant amphibole phenocrysts present in Nuku Hiva lavas. Such conditions could be consistent with the occurrence of a well-developed hydrothermal system above the magma reservoir, although there is presently no geological evidence of it. Indeed, the initial floors of the caldera-like structures which could have provided such evidence have been flooded by the hawaiitic to mugearitic flows emitted from the Taiohae inner volcano. According to what is known in Piton de la Fournaise from geophysical data, both depth and thickness of the hydrothermal system in experiments vary from 0.5 to 1 cm (i.e. 0.5 to 1 km in nature). The analogue volcano straddles two rigid and horizontal plates, one of which moving down with the means of a screw jack governed by a computer-controlled stepper motor (Fig. 5). The progressive evolution of model deformation is recorded by overhead time-lapse photography.

5. Scaling

A scale model is a simplified version of a natural prototype in terms of dimensions, rock rheology and boundary conditions. The lack or uncertainties of many natural data concerning Nuku Hiva make the scaling

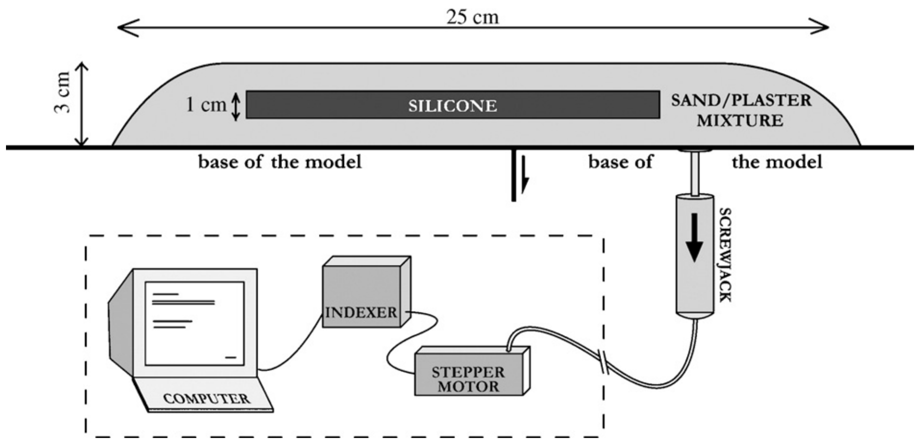

Fig. 5. Experimental device. The hydrothermal system is simulated by silicone putty embedded in a sand/plaster mixture, which simulates the cohesive, brittle rocks of the volcanic edifice. The right side of the base is moving down by the means of a computer-controlled screw jack.

procedure challenging. Basically, the experiments aim at reproducing a physical process where unconfined lateral boundary is obtained by tectonic collapse of a part of a volcanic island (i.e. by normal faulting). When natural data from Nuku Hiva are not available, reasonable values are chosen according to information gathered from other volcanic islands, the ultimate goal being to verify the mechanical consistency of the model under consideration.

The scaling procedure followed herein is identical to that discussed by Merle and Borgia (1996) where selected dimensionless numbers must be of the same order of magnitude in nature and experiments. According to the Buckingham Π theorem, there are 10 variables minus 3 dimensions that give 7 dimensionless numbers that need to maintain the same value in nature and experiments.

Of these, two numbers are the geometric ratios of the system (Table 1):

$$\Pi_1 = \frac{\text{Length of the whole collapse structure}}{\text{Width of the whole collapse structure}} = \frac{L}{W}$$

$$\Pi_2 = \frac{\text{Total offset along the normal fault}}{\text{Width of the whole collapse structure}} = \frac{H}{W}$$

Vertical displacement along the fault, which probably occurred in two steps (see discussion), is not available with accuracy but can be assumed from general considerations. The total offset quantity is supposed to be of the order of magnitude of the vertical distance from the highest peak in Nuku Hiva (1224 m) to the floor of the ocean south of the island (−500 m according to Wolfe et al., 1994), that is close to 2 km.

Concerning the scaling of brittle material, the angle of friction is the first dimensionless parameter:

$$\Pi_3 = \varphi$$

The angle of friction, being dimensionless, must be the same in nature and experiments. The angle of friction of the analogue material used in these experiments is in the range 33°–37°, which can be considered as similar to the friction angle of most rocks encountered in nature.

The second dimensionless number to scale brittle materials is given by the ratio of gravitational stress to cohesion:

$$\Pi_4 = \frac{\rho g H}{\tau_0}$$

This dimensionless number needs to maintain the same value in nature and experiments to guarantee similarities. In nature, we consider an average density of about 2500 kg m^{-3} for the pile of volcanic rock. The cohesion of intact and massive natural rocks is usually considered to be about 10^7 Pa (Jaeger and Cook, 1971; Watters et al., 2000). However, uncertainties remain regarding the values of cohesion in nature as it has been shown that the cohesion of lithified and/or fractured rock mass is one or two orders of magnitude less than this upper value and could be in the range of 10^{15}–10^{16} Pa (Hoshino et al., 1972; Hoek and Bray, 1981; Schultz, 1996). In Nuku Hiva, the Tekao outer shield volcano is composed of fractured tholeiitic lava flows, interbedded with scoriae levels and cross-cut by numerous dykes. Thus, we assume a cohesion of about 10^{16} Pa for the whole volcano, that is one order of magnitude less than the cohesion usually considered for intact rocks. The dimensionless number in nature yields a value of 50. In experiments, the density of the dry/plaster mixture is about 1300 kg m^{-3} with a cohesion of about 25 Pa. Using these physical quantities to calculate the value of the dimensionless number Π_4 number in the experiment yields a value of 10.

To scale ductile material, the principle dimensionless number which has to be considered is the ratio between the gravity force and the viscous force, that is:

$$\Pi_5 = \frac{\rho g h t}{\eta}$$

where t is the time during which motion occurs and h and η the thickness and the viscosity of the flowing material. In experiments, the time needed to form the caldera-like depression once the lateral side is unconfined is about 1 h. The density and the viscosity of the silicone used to simulate the clay-rich weak core are 1000 kg m^{-3} and 5×10^4 Pa s^{-1}, respectively. The thickness of the analogue hydrothermal system varies

Table 1
Average Π dimensionless number in the field and experiments

Dimensionless variable	Definition	Value	
		Field	Experiment
Π_1	Length/width of the whole structure	~0.6	0.66
Π_2	Fault offset/width of the whole structure	0.1–0.05	0.1
Π_3	Friction angle of brittle material	~30°	33°–37°
Π_4	Gravitational stress/cohesion	50	10
Π_5	Gravitational/viscous forces	8	3.6
Π_6	Inertial/viscous forces	10^{-18}	10^{-10}
Π_7	Volcano/hydrothermal system density	1.5	1.3

from 0.5 to 1 cm depending on the experiment. This yields a Π_5 dimensionless number of 3.6 with h=0.5 cm. The viscosity of clay in nature may vary by several orders of magnitude and can be in the range of 10^{14}–10^{18} Pa s, according to the wide variety of minerals, water contents and applied strain rates (Van Wyk de Vries and Matela, 1998; Carena et al., 2000; Arnaud, 2005). Due to the lack of data concerning the viscosity of the clay-rich weak core in Nuku Hiva, we choose an average value of 10^{16} Pa s. The density of the unconsolidated clay is estimated to be 1600 kg m^{-3} (Philipponat and Hubert, 1997). Assuming that a few hundred years is a reasonable time scale for the hydrothermal system to flow, as at Casita volcano (500 years) (Van Wyk de Vries et al., 2000), we select 10^{10} s (300 years) for the process under consideration. This yields a Π_5 dimensionless number of about 8 with a h=500 m.

It is well known that inertial forces may be neglected in most geological processes. This can be appreciated by the Reynolds number, which gives the ratio between the inertial and the viscous forces in a process:

$$\Pi_6 = \frac{\rho h^2}{\eta t}$$

In experiments, the Reynolds number is about 10^{-10}. In nature, the Reynolds number is about 10^{-18}. Clearly, the very small values indicate that inertial forces are negligible with respect to viscous forces both in nature and experiments and this number will receive no further consideration.

Finally, the last dimensionless number must be related to the difference in density between the volcano (ρ_v) and the clay-rich weak core (ρ_h) of the hydrothermal system:

$$\Pi_7 = \frac{\rho_v}{\rho_h}$$

This Π_7 dimensionless number yields 1.5 and 1.3 in nature and experiments, respectively.

6. Results

A set of experiments has been conducted varying two main parameters, the velocity of the downgoing compartment and the brittle/ductile ratio (i.e. the thickness ratio between the brittle roof and the underlying silicone). The two basic experiments shown herein, which give the best results, have been carried out with a very slow velocity (a few millimetres per hour) and a low brittle/ductile ratio of about 0.5 (Fig. 6).

In slow experiments, no fault forms at the surface along the limit between the two compartments. Due to the silicone underneath which prevents any fracturing, the fault is accommodated at the surface by a gentle slope between the two compartments. After 1 h, a ring fracture appears in the stable compartment all along the sides of the underlying silicone. With time, the brittle material delimited by the ring fracture slowly collapses forming a caldera-like structure (Figs. 6 and 7). In all experiments, the floor of the caldera-like structure is perfectly flat and overhangs the lowered compartment. No secondary landslides are observed on the surface of the caldera, which is remarkably stable, and the only visible motion of the floor is vertical downward, as a lift. The outline of the caldera-like depression always

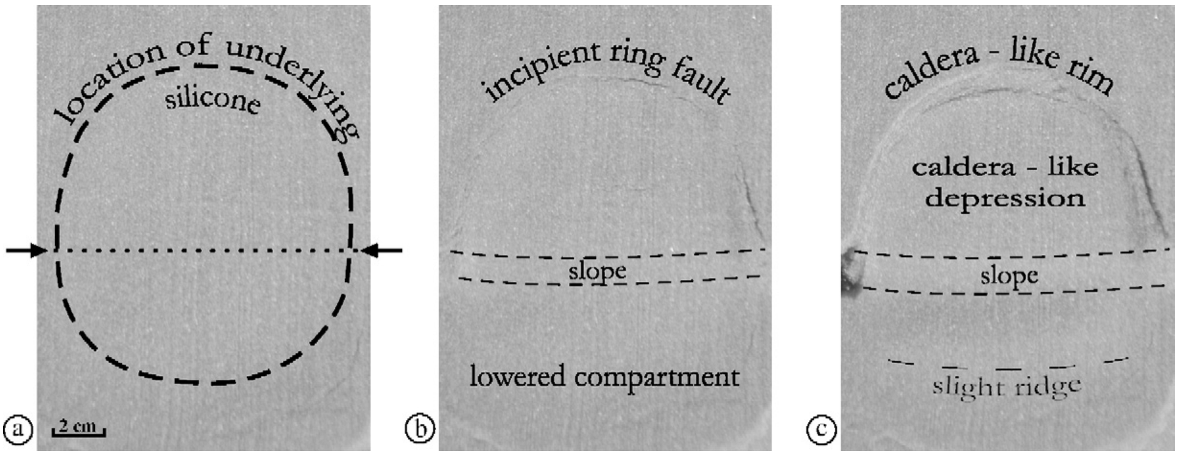

Fig. 6. Three evolutionary stages of experiment HS 10. Black arrows show the location of the E–W fault in the basement. A ring fault (b) develops in the upper compartment, which matches the shape of the underlying silicone (dashed line, a). Vertical collapse inside the ring fault leads to a caldera-like structure (c). In the lower compartment, a slight E–W ridge indicates that spreading silicone intrudes horizontally in front of the slope (see Fig. 4).

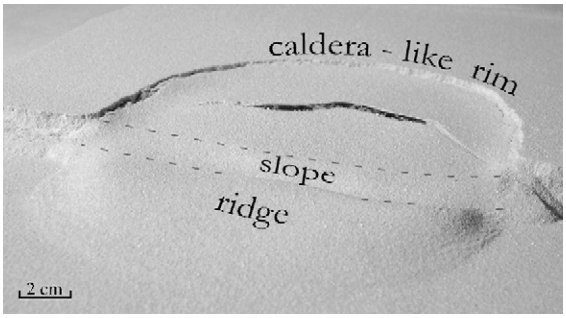

Fig. 7. Oblique view of the caldera-like structure in experiment HS 12. Note that a large fracture has developed on the floor of the caldera. When two nested calderas are observed in the field, the location of the inner caldera may be controlled by activation of such fractures into normal faults.

mimics the size and shape of the analogue clay-rich core underneath, a result already shown in experiments on the Réunion Island (Merle and Lénat, 2003).

The fact that no silicone outcrops at the surface of the model during the deformation suggests that the underlying silicone flows from the upper compartment to the dowthrowning compartment along the slope in between. This process results in the formation of a slight ridge located in front of the slope on the downthrown compartment (Fig. 6). Such a deformation is active, even if the downgoing movement is interrupted, until the flow of the silicone is rendered impossible due to its reduced thickness.

Increasing the brittle/ductile ratio leads to the formation of a large curved fracture on the flat floor of the caldera-like structure, roughly parallel to the adjacent slope and laterally connected to the ring fracture. When this fracture enlarges, the buoyant rise of the silicone may occur and silicone outcrops at the surface.

Performing experiments with a very high velocity of the downgoing compartment is similar to instantaneous faulting leaving a free boundary on the side of the stable compartment. The deformation is then completely different with the formation of faults parallel to the free boundary. Faults develop successively backward away from the free surface. Slightly later, a ring fracture also forms, matching the shape of the underlying silicone, but vertical collapse remains very limited.

In short, the best results were obtained for very low velocity faulting and a brittle/ductile ratio lower than 2. The velocity parameter stresses the balance between the reducing lateral stresses and the flow of the ductile material. Creating instantaneously a free boundary does not leave enough time for the silicone to achieve a slow horizontal spreading distributed within the entire ring fracture. Repeated faulting and sliding next to the free boundary precedes and then prevents the formation of a large caldera-like structure behind. This observation suggests that a large caldera-like structure in nature may form when normal faulting occurs with either a slow velocity or a limited offset. In other words, lateral stresses must be reduced but not suppressed to allow the collapse of the whole hydrothermal system.

The brittle/ductile ratio is probably less crucial than the velocity parameter for the mechanism under consideration. Fracturing the floor of the caldera may even be not visible in natural examples as it is frequently reshaped by periodic lava flows. A discontinuity like that visible in Fig. 7 might well be re-activated later and influence the location of a second caldera fit into the larger one.

7. Discussion

The overall geometry described in experiments matches some features observed on Nuku Hiva Island, including a ring fault delimiting a large flat depression overhanging the sea without intermediate landslide structures. Together with evidences of a fault bounding the south part of the island, these geometrical similarities are probably significant and indicative of a specific kinematic process different from the models usually proposed to explain volcano instabilities, like magma withdrawal (classic model of downsag caldera formation), landslides on low-angle fault planes or volcano spreading.

We are well aware that more data are needed to confirm such a hypothesis, especially geophysical studies which are critical to support the model. For instance, only a marine survey may show the implied tongue of weak hydrothermal altered material extruded onto or just below the seafloor. Likewise, deep drillings into the caldera floor (similar to the Grand Brûlé drilling in Réunion) are needed to document the importance and lateral extent of the hydrothermal clay-rich core supposed to have spread towards the sea. To this respect, it may be envisaged that the weak core was not entirely hydrothermal at the moment of the formation of the caldera-like depression and could include a soft, hot magmatic complex, able to spread laterally. Despite thermal limitation of flow that can occur with time, such a hot magmatic complex may undergo a similar kinematical process once an unbutressed boundary is present reducing lateral stresses.

Geological data show that the two nested caldera-like structures formed nearly at the same time at about 4.05 Ma, i.e. at the end of the building of the Tekao shield volcano. In most experiments, a single ring fault surrounds the flat depression which mimics a caldera structure. However, a large concentric fracture may develop in the floor of the flat depression as depicted in the experiment shown in Fig. 7. This fracture goes right through the entire brittle floor of the depression and reaches the ductile part of the model, that is the analogue hydrothermal system. Such a fracture, once laterally connected to the former ring fault, is likely to initiate a split of the flat depression into two individual compartments evolving separately, the inner one suffering stronger collapse due to its proximity to the regional normal fault. We argue that such an evolution may explain the two nested calderas in Nuku Hiva. In this model, uncertainties remain on the contour of the ring fault of the outer caldera-like structure as shown in Fig. 2. It is difficult to know whether or not the lateral limits of the inner caldera parallel that of the outer one and the possibility that the ring fault of the inner caldera have enlarged the outer caldera, as suggested by field data, cannot be ruled out.

Later on, hawaiitic and mugearitic lava flows from the Taiohae volcano filled up the inner caldera-like structure and flooded onto the outer one between 4.0 and 3.9 Ma, before the deposition of the Toovii lacustrine sediments. Subsequent vertical offset (i.e. 80–100 m) of these lacustrine sediments and underlying hawaiitic and mugearitic lava flows along the ring fault of the inner caldera may be linked to the emplacement of the 15 km^3 of trachytic domes and flows, which erupted from the Taiohae volcano between 3.9 and 3.6 Ma. Such a mass may have overloaded the floor of the inner caldera and this process, together with the emptying of underneath reservoirs, may have triggered further but limited collapse of that inner caldera through the reactivation of its ring fault.

The question arises whether or not such a mechanism can have controlled the tectonic evolution of other volcanic islands. Surprisingly, Nuku Hiva is not the only island in the Marquesas to show structural features similar to those discussed above. Although their caldera morphology is likely to have been modified by debris avalanches (Clément et al., 2002; Legendre et al., 2006), Ua Huka and Hiva Oa (Figs. 1 and 8) also display an asymmetrical shape limited to the south by a more or less straight limit oriented along a direction close to that of the major fracture zones of the area. Likewise, the occurrence of caldera-like structures breached at the sea along the southward straight limits of the islands (Fig. 8) may indicate that these systems have collapsed in a similar way, i.e. under the same regional tectonic control as for Nuku Hiva Island. In addition, the only known zone of hydrothermal alteration in the archipelago, the Fatueki “Soufrière”, is located along the southern coast of Hiva Oa (Guille et al., 2002). We suggest that the working hypothesis presented in this paper may serve as a tectonic guide to study the structural evolution of other islands of the Marquesas Archipelago.

8. Conclusion

Experiments show the mechanical consistency of the model proposed herein as an alternative to the process by which lateral stresses are reduced in Piton de la Fournaise (Merle and Lénat, 2003). Regional tectonics involving normal faulting may also provide a weak lateral confinement along which viscous spreading of the altered core of the edifice takes place. Geological data from Nuku Hiva may be interpreted in this way, and give a reasonable explanation to the two large nested calderas that cannot be explained either by magma withdrawal from a large underlying reservoir or by landslides on low-angle failure planes.

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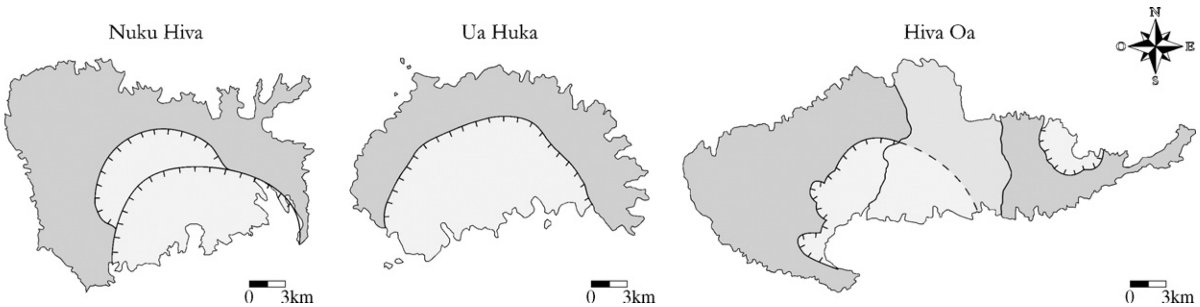

Fig. 8. Compared tectonic features of some Marquesas Islands. Explanations in the text. The location of the three islands is shown in Fig. 1.

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