Available online at www.sciencedirect.com

SCIENCE DIRECT®

Journal of Volcanology and Geothermal Research 144 (2005) 169–189

Journal of volcanology and geothermal research

www.elsevier.com/locate/jvolgeores

Landslides and spreading of oceanic hot-spot and arc shield volcanoes on Low Strength Layers (LSLs): an analogue modeling approach

Jean-François Oehler*, Benjamin van Wyk de Vries, Philippe Labazuy

Laboratoire Magmas et Volcans, UMR 6524 CNRS, OPGC, Université Blaise Pascal, 5 rue Kessler, 63038 Clermont-Ferrand Cedex, France

Received 16 November 2004; accepted 16 November 2004

Abstract

During their growth and evolution, oceanic hot-spot and arc shield volcanoes are destabilized and often dismantled by landslides. The origin of these phenomena can be found in the structure of volcanic islands marked by several low strength horizons such as a marine substratum, hyaloclastites, hydrothermally altered rocks, or volcaniclastic deltas, that weaken the edifice. We have tested the role of these Low Strength Layers (LSLs) with analogue modeling. We conclude that destabilization on these levels can lead to landslides or gravitational spreading. In our experiments, slides generated are only slump-type events, but they can evolve, in nature, into debris avalanches through mechanisms that could not be modeled in the laboratory. The role of fan-shaped volcaniclastic deltas included in the edifice structure is critical to slump formation. This process is more efficient if deltas are associated with another upper LSL, such as weak hyaloclastites or hydrothermally altered rocks. In this case, recurrent events generating collapses with increasing width are observed. Gravitational spreading of a volcanic island occurs when it is underlain by a ductile marine substratum, or includes an intra-edifice unconstrained hyaloclastic layer, 1/3 as thick as the height of the subaerial edifice. Spreading is characterized by edifice stretching and subsidence and does not generate major landslides. We also show that gravitational spreading on a marine substratum tends to inhibit slide generation, especially when the thickness of the sedimentary layer is larger than 1/10 of the edifice total height. However, if this value is lower, spreading occurs later and slumps can develop initially. We describe natural examples of oceanic hot-spot or arc shield volcanoes affected by both spreading and landslides, such as Reunion Island, Hawaii Island, James Ross Island (Antarctica) and Martinique Island (Antilles). The main structural features on these volcanoes can be explained solely by deformation on LSLs. We therefore propose a general model for the destabilization of volcanic islands that is applicable to all shield-like edifices.

Keywords: low strength layers; landslides; volcano spreading; analogue models; oceanic hot-spot shield volcanoes; arc shield volcanoes

* Corresponding author. Tel.: +33 04 73 34 67 21; fax: +33 04 73 34 67 44.
E-mail address: J.F.Oehler@opgc.univ-bpclermont.fr (J.-F. Oehler).

0377-0273/$ - see front matter
doi:10.1016/j.jvolgeores.2004.11.023

1. Introduction

Landslides are recognized to be common and recurrent during volcano growth. If steep stratovolcanoes have a greater probability to be destabilized, the largest landslide deposits, with volume up to 1000 km^3, have been detected around gently sloping volcanic shields. Numerous studies, especially those using swath bathymetry and sonar data, have shown the importance of landslides in hot-spot related oceanic volcano growth, for example, in the Hawaiian Islands (Moore et al., 1989, 1994), the Canary Islands (Ablay and Hürlimann, 2000; Krastel et al., 2001), Reunion Island (Lénat et al., 1989; Labazuy, 1996; Oehler et al., 2004a), the French Polynesia Islands (Filmer et al., 1994; Clouard et al., 2001; Clément et al., 2002) and Tristan da Cunha Island (Holcomb and Searle, 1991). Seabound arc volcanoes can be shield-like and also suffer landslides, for example in the Lesser Antilles Arc (Deplus et al., 2001; Le Friant et al., 2002).

The term landslide is retained as a general term for all slope failures, but following the classification of Varnes (1978), two general types of landslides, slumps and debris avalanches, are considered here. A slump is a slow and intermittently-moving event characterized by displacements of large undeformed masses, whereas a debris avalanche is a catastrophic event where material is largely disaggregated during transport. Several trigger mechanisms have been proposed to explain the generation of these landslides. These include dyke intrusions (Iverson, 1995; Elsworth and Day, 1999), oversteepening of slopes by new lava (Siebert, 1984), caldera collapse (Marti et al., 1997), seismicity (Tilling et al., 1976; Lipman et al., 1985), slope steepening by coastal erosion (Ablay and Hürlimann, 2000), but also effects of low strength layers (LSLs) existing in the volcano structure. Such layers can be hot olivine cumulates (Clague and Denliguer, 1994), volcanic residual soils (Hürlimann et al., 2001) or water-saturated rocks (Siebert, 1984). For this study, we particularly focus on the role of four LSLs, namely, (1) a marine substratum, (2) volcaniclastic deltas, (3) hydrothermalized zones and (4) hyaloclastites (Fig. 1):

(1) Volcanic islands generally grow on top of marine sediments. These clay-rich water-saturated pelagic sediments act as a décollement horizon, because of their low shear strength and because rapid compaction under the volcano load leads to overpressuring of pore fluids and low effective normal stress (Iverson, 1995; Smith et al., 1999). The presence of such a layer under volcanic islands causes spreading. Gravitational spreading is distinguished from volcanic spreading. Gravitational spreading is defined as the deformation of a weak strata under the weight of a volcano that induces the deformation of the strata and consequently the destabilization of the volcano (Merle and Borgia, 1996). The destabilization is made easier when a sufficiently magma influx is added to the process. Borgia (1994) uses the term volcanic spreading to describe this magma-driven spreading. Spreading of stratovolcanoes can result in disastrous flank landslides and debris avalanches. Several examples are known such as Mombacho (Van Wyk de Vries and Francis, 1997) or Socompa (Van Wyk de Vries

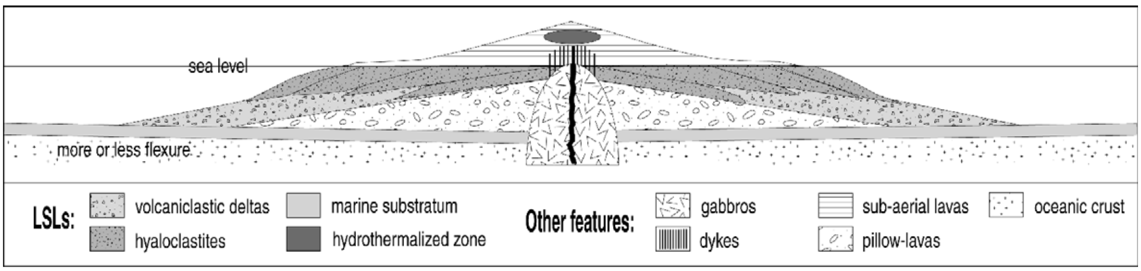

Fig. 1. Our idealized construction of volcanic island. 4 LSLs are indicated. This general model is made for a basaltic shield, but can equally be applied to more evolved hot-spot volcanoes, such as Piton des Neiges (La Reunion), or to arc volcanoes, such as Guadeloupe, Martinique or Montserrat (Antilles).

et al., 2001a). For volcanic islands, spreading is thought to contribute to the generation of slumps. The Hilina slump (Kilauea, Hawaii) is interpreted as a deep-seated gravitational slide due to spreading of the volcano south flank on the pre-Hawaiian seafloor (Lipman et al., 1985; Borgia and Treves, 1992).

(2) Volcaniclastic deltas (also called 'deltas' in the following) result from different sedimentary processes on volcano flanks. Landslide events are certainly an important process for delta formation since hundreds of km^3 of material can be displaced in a single event. Growing seamounts, still below sea level, are already affected by such mass-wasting events. Loihi seamount, the youngest Hawaiian volcano, is thus extensively modified by landsliding, even though its summit is still 1 km below sea level and the volcano is perhaps a hundredth the size of a mature shield (Moore et al., 1989). Seismic tomography reveals that the structure of volcanic islands is often marked by lateral and vertical heterogeneities, including low velocity zones, interpreted as landslide deposits (Weigel and Grevemeyer, 1999). Fan-shaped slide deposits are probably included within the structure of all volcanic islands. Landslide deposits, and in particular debris avalanche deposits, consist of a poorly sorted mixture of brecciated debris. They are characterized by the association of large coherent blocks and large chunks of various lithology and by a surrounding mixture of crushed heterogeneous fragments with a predominance of thin elements. Fluid circulation, subsequent erosion and weathering alter these deposits and produce a more fine-grained and clay-rich material that could play the role of a décollement zone. Debris flow deposits, detritic deposits and fine sediments resulting from the pelagic sedimentation and turbidite deposition are also components of deltas. Such materials contain mechanically weak subaerially weathered products. On arc volcanoes and evolved shields, pyroclastic flow materials dominate the volcanic pile. This material, either emplaced directly or reworked, will also form a weak component in the submarine construction. Deltas are thus a heterogeneous, potentially unstable layer of low strength material beneath or within the volcanic edifice that can influence island flank stability (Leslie et al., 2002).

(3) The alteration of volcanoes by hydrothermal circulation results in edifice weakening and can trigger sector collapses (Lopez and Williams, 1993; Day, 1996). Alteration by hot fluids produces extensive rock dissolution and generates minerals such as gypsum and clay. These are weak and ductile. Hydrothermally altered zones have therefore low strength and can flow, collapse and spread laterally under their own or the edifice weight. On stratovolcanoes, this can cause large flank landslides, such as at Casita volcano (Van Wyk de Vries et al., 2000; Kerle and van Wyk de Vries, 2001). Merle and Lénat (2003) underline, however, that this process will be less active for shield volcanoes, because of the buttress effect of a shallow slope. They add that if this confinement is breached in one sector, then the weakened zone can spread and create a caldera-like structure.

(4) Hyaloclastites are glassy, water-quenched, fragmental lava, generally of basaltic composition. Most hyaloclastites are produced when subaerial lava flows across a shoreline, where it is disrupted, quenched and shattered (Moore, 2001). The existence of hyaloclastites in the structure of volcanic islands is proved by drilling data (Moore, 2001; Garcia and Davis, 2001; DePaolo et al., 2001; Nielson and Stiger, 1996; Klingelhöfer et al., 2001) but also by submarine observations. The Hilo drill hole in Hawaii exhibits two types of hyaloclastites (Moore, 2001). Uppermost hyaloclastites are porous, permeable, and uncemented, whereas the deeper older hyaloclastites absorb less water and become better cemented due to the palagonitization of glass and addition of secondary minerals. Moore (2001) affirms that the zone of freshest uncemented hyaloclastites represents the weakest rock in the drill hole and concludes that it is a likely level for tectonic or landslide disruption. In general, we suggest that most LSLs have a ductile behavior, especially due to their clay-rich composition. The behavior of hyaloclastites is however more debatable. Our

observations on hyaloclastites in James Ross Island (Antarctica) (Van Wyk de Vries et al., 2003; Oehler et al., 2004b) show that they are a highly fractured, weathered and palagonitized material, with clear destabilization potential. If their behavior must therefore be considered as brittle on a small scale, on a large scale and under the weight of an overlying edifice, this level probably behaves as a ductile destabilization zone, although certainly with a higher viscosity than the other LSL, due to lower clay contents, and greater cementation.

We present analogue experiments that study the influence of the four LSLs (a marine substratum, volcaniclastic deltas, hydrothermalized zones and hyaloclastites) on the stability of oceanic hot-spot and arc shield volcanoes. We base our scaled models on the hot-spot Reunion Island edifice. This oceanic shield has in fact probably the best average morphology for generalizing models to other oceanic hot-spot and arc edifices. We attempt to understand how slide structures, that is to say horseshoe-shaped structures or amphitheatres, can be generated only from the deformation of LSLs and to evaluate the potential of spreading in generating landslides. It seems right that weak LSLs should favor the gravitational destabilization of volcanic islands by causing slumps. By combining structures observed in our models with those on real volcanoes, we construct a general model for LSL gravitational destabilization of shield volcanoes.

2. Analogue models

2.1. Scaling

We simulate a shield volcano by a model composed by sand and plaster modeling the brittle volcanic rocks. LSLs included in the structure of the edifice are considered ductile and are simulated by silicone layers. The choice of these materials is linked with the scaling procedure. A scaled model is a reduced representation of reality, which needs to be geometrically, kinematically and dynamically similar to nature (Hubbert, 1937; Ramberg, 1981). The scaling procedure described in previous studies on scaled models of volcanic edifice aims to respect these conditions (Merle and Vendeville, 1995; Van Wyk de Vries and Merle, 1996; Merle and Borgia, 1996; Donnadieu and Merle, 1998). For technical reasons, we set the model/nature length ratio L^* to 5×10^{-6} (1 cm in experiments represents 2 km in nature). The mixture of sand and plaster used in experiments has an average bulk density of 1400 kg m^{-3}, whereas volcanic rocks present an average bulk density of 2500 kg m^{-3}. The density ratio ρ^* is thus 0.56. Experiments were carried out in the terrestrial gravitational field so that the gravity ratio g^* is 1. The stress ratio is therefore $\sigma^*=\rho^*\times g^*\times L^*=2.8\times10^{-6}$. We can deduce from it that our models should be about 300,000 times less strong that the natural volcano. As the cohesion τ_0 has the dimension of a stress, we can write that $\tau_0^*=\sigma^*$. Consequently, if we take a general high cohesion for volcanic rocks (about 10^6–10^7 Pa), the analogue material must have a very low cohesion. The mixture of sand (95% in weight) and plaster (5% in weight) satisfies to this condition with a very low cohesion of about 10–20 Pa (calculated following Donnadieu, 2000). It moreover presents an angle of internal friction comparable to that of natural rocks (~30°). The material allows us to follow the morphologic evolution of structures, which appear as small scarps in experiments. To simulate LSLs, we generally use the transparent silicone PDMS (trade name: SGM 36; produced by Dow Corning, UK), mixed with less than 1% in weight of sand. We measured the viscosity of this mixture at about 10^5 Pa s. It is chosen as the reference ductile material for the scaling. In one series, we however use both PDMS and the pink silicone Rhodorsil Gomme (trade name: Silbione Gomme 70009; produced by Rhône-Poulenc, France) mixed with up to 5% in weight of sand (viscosity measured at about 10^6 Pa s). It allows us to simulate a contrast of viscosity between less ductile hyaloclastites and the other LSLs. The deformation of a ductile material is time-dependent and we can write that $t^*=\mu^*/\tau_0^*$, t^* being the time ratio and μ^* the viscosity ratio. If we consider that the viscosity of LSLs, strongly linked with that of clay, is about 10^{18} Pa s (Borgia, 1994; Merle and Borgia, 1996; Carena et al., 2000), then t^* is about 10^{-7}. One hour of experiments is then equivalent to about 1 ka. In a few hours, most experiments produced significant deformation. As 1 ka is well within the lifetime of large volcanic edifices, such deformation can be reasonably expected to occur in nature.

2.2. Experimental procedure

Our scaled models consist of 3-cm-high and 30-cm-diameter circular shields that aim to simulate an edifice comparable to Reunion Island (Fig. 2A,C). An important morphological aspect is the natural break of slope near the shore and which is reproduced on models (Fig. 2B,D). This break of slope is a general feature on shield-like volcanic islands. It is explained by a change in depositional environment across the shoreline. Subaerial lava flows generally form low slopes. In contrast, when they cross the shoreline, they are rapidly quenched and undergo cooling, contraction and granulation (Moore and Chadwick, 1995). The resulting breccia forms a steep scree slope. Pyroclastic flows entering the sea form similar morphologies. Models are directly built on a table with sand and plaster plus silicone layers for the different tested LSLs. We have first chosen to study the role of each type of LSL individually. The interactions between different LSLs are then tested in more complicated, but also more realistic models. A qualitative structural analysis is first done during experiments, and then quantitative deformation analysis is made using vertically acquired digital photos with standard methods (Donnadieu et al., 2003). Displacements are estimated using black silicon carbon grains arranged at the surface of the model at the initial stage and followed step by step through image processing.

2.3. Limitations

Our experiments only study the gravitational deformation of LSLs under their own and the weight of the overlying edifice. Therefore, for such passive models simulating shield volcanoes, only slump-type slide events can be observed. The triggering of catastrophic events (debris avalanches) would require to take into account changes in state not possible in the laboratory. An important limitation of our models is that we simulate LSLs with a similar viscosity (~10^5 Pa s) whatever their origin. We assume that while LSLs have different constituent materials, their bulk behaviour is roughly the same. This is because the proportion of weak elements such as clays can be similar in each situation. In nature, it is likely that there is a significant range of viscosity even within one LSL type.

In a series of experiments, we consider hyaloclastites one order of magnitude less ductile than the other LSLs (~10^6 Pa s) to simulate their lower clay content and better cementation. The aim is simply to apply a contrast of viscosity between LSLs. Another omission

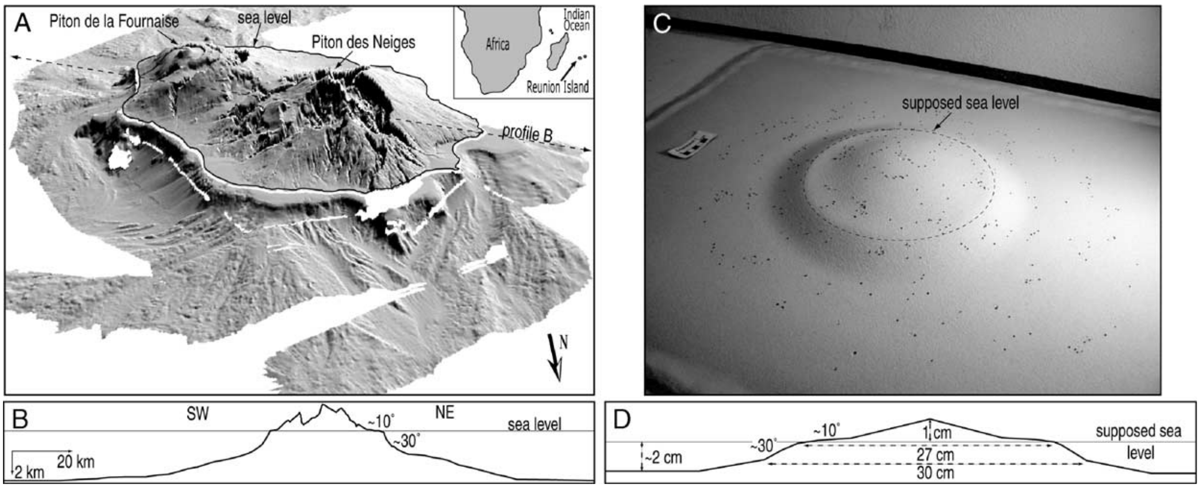

Fig. 2. Comparison between Reunion Island and our scaled model. (A) Three-dimensional representation of the emerged and submerged flanks of La Reunion (×4 vertical exaggeration). Global 100 m-gridded DTM with shaded relief overlay (apparent illumination from the south-west). Inset shows the location of Reunion Island in the Indian Ocean. From Oehler et al. (2004a). (B) Southwest northeast profile of La Reunion edifice (location is given in A). (C) Oblique view and (D) schematic section of the reference scaled model. Note that the natural break of slope of La Reunion edifice is respected in the scaled model.

in the models is the restraining load of the ocean. This would have to be modeled as a low-density (500 kg m^{-3}), low viscosity (10^{-19} Pa s^{-1}) material. The sea would have an effect of damping deformation induced by the emerged edifice, but would also load submarine deltas, possibly increasing delta-related deformation. Erosion processes and eruptions are not modeled, nor crustal flexure under the load of volcanic edifice, especially important for such large shield volcanoes. Our models are therefore a simplification of the reality, but observed structures are comparable to those described in nature.

3. Experimental results

Although the scaling is based on an edifice comparable to La Reunion, we try to generalize as much as possible our results. LSL main parameters are thus compared with parameters characterizing

Table 1
Synthesis of our analogue experiments

LSLs characteristic parameters	Examples of experiment	Characteristic times (ka)[a]	Displacement rates (cm/year)[b]	Main processes
Case A: Only a marine substratum				
e=1/10 H_e	Fig. 3	4–6	40	Spreading
e>1/10 H_e	–	3	–	Spreading
e<1/10 H_e	–	24	–	Spreading
Case B: Only deltas				
Type 1	Fig. 4B,C	<5	14	Slide
Type 2	Fig. 4D,E	<5	80	Slide
Type 3	Fig. 4F,G	<24	6	Slide
Case C: Only an upper LSL				
D_h>8/9 D_v, whatever E (hyaloclastites)	Fig. 5	3–5	16	Spreading
D_h≤8/9 D_v, whatever E (hyaloclastites or hydr. zone)	–	–	–	Nothing
Case D: Association of a marine substratum and deltas				
e>1/10 H_e	–	3	56	Spreading
e<1/10 H_e	–	–	–	Slide (see case B)
e=1/10 H_e, n=3 (type 1), 2 observed steps:	Fig. 6	2	52	Slide+
		9	35 deltas, 22 elsew.	Spreading
e=1/10 H_e, n=4 (type 1), 2 observed steps:	–	2	86	Slide+
		5	71 deltas, 42 elsew.	Spreading
e=1/10 H_e, n=2 (type 1), 2 observed steps:	–	2	85	Slide+
		5	88 deltas, 55 elsew.	Spreading
Case E: Association of an upper LSL and deltas				
No contrast of viscosity between both LSLs:				
D_h>8/9 D_v, whatever δ	Fig. 7B,C	3	10	Spreading
D_h≤8/9 D_v, δ≤1/27 D_v	Fig. 7D,E	2	23	Slide
	Fig. 7F,G	2	17	Slide
whatever D_h, δ>1/27 D_v	–	–	–	Slide (see case B)
With hyaloclastites more viscous:				
D_h=D_v, E≤1/3 H_v	Fig. 8B,C	2	14 deltas, 1.5 elsew.	Slide+ Spreading
D_h=D_v, E>1/3 H_v	Fig. 8D,E	<3	25 deltas, 4 elsew.	Slide+ Spreading
Case F: Association of an upper LSL, a marine substratum and deltas				
whatever all parameters	Fig. 9	1	80	Spreading

[a] Time needed to observe first significant structures.
[b] Maximum displacement rates on average on the total duration of experiment. See text and figures for more precisions.

the geometry of a volcanic island, such as the edifice total height H_e or the diameter and height of its subaerial part, D_v and H_v. The main results of our analogue experiments are synthesized in the Table 1. It classes the observed destabilizations into two main processes, spreading and slides, although intermediate forms exist where spreading and slide are intimately associated. This table can thus serve as a guide to evaluate the potential destabilization of oceanic hot-spot and arc-shield volcanoes.

3.1. Case A: only a marine substratum

First we modeled a brittle volcano lying on a ductile marine substratum of constant thickness e (Fig. 3A). This layer is considered as infinite and the silicone was constrained at its boundaries by a sand heap in order to restrict lateral creep. We varied the parameter e, defined as a fraction of the edifice total height (H_e), from e=1/10 H_e to e=1/3 H_e. In all the cases, the final morphology of the edifice is quite similar; the structures being generated more rapidly with increasing thickness (Table 1). The morphology is characterized by (a) radial stretching structures (triangular horsts and grabens) localized at the break of slope and (b) concentric anticlines and thrusts, all around (Fig. 3B). Displacements are mainly radial causing an overall subsidence of the edifice (Fig. 3C). The process clearly resembles gravitational spreading as described by Merle and Borgia (1996). We observed no potential slide structures.

3.2. Case B: only deltas

Here we included only fan-shaped ductile deltas within a brittle edifice (Fig. 4A). They are considered to lie on a brittle conic base simulating a pillow-lava foundation. Three types of deltas were tested (Table 2, Fig. 4A). For each experiment, three deltas of the same type were used and laid at 120° one from each other. Whatever the delta shape, the observed structures are quite similar (Fig. 4B–G), even if they develop at different rates (Table 1). Systems of normal faults develop directly facing the deltas. They present a horseshoe-shaped morphology and are organized in

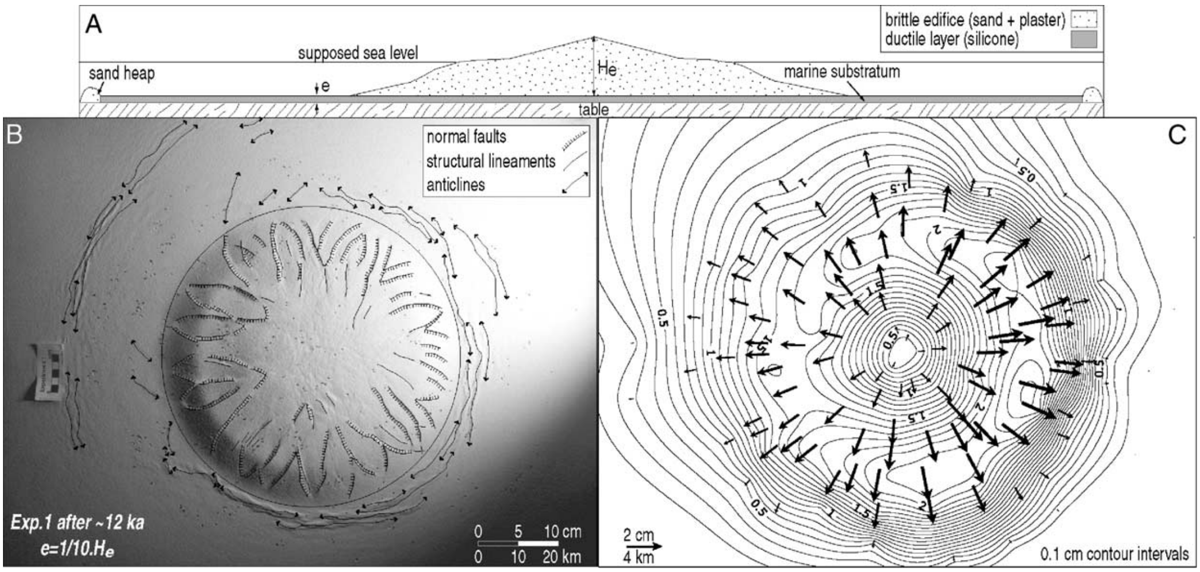

Fig. 3. The effect of a ductile marine substratum on the stability of a brittle shield volcano. (A) Schematic general representation of initial stage of experiments. (B) Experiment 1 photograph with structural map overlay. (C) Contour map of the calculated displacement field (in cm). Vectors illustrate the amplitude and direction of displacement. The larger the arrow is, the more important the displacement is. The model clearly illustrates gravitational spreading described by Merle and Borgia (1996). We have deliberately chosen to build an asymmetric model; the extent of the silicone layer being greater to the left than elsewhere. It explains why concentric anticlines and thrusts have developed at the base of the edifice to the right, while they are located further to the left.

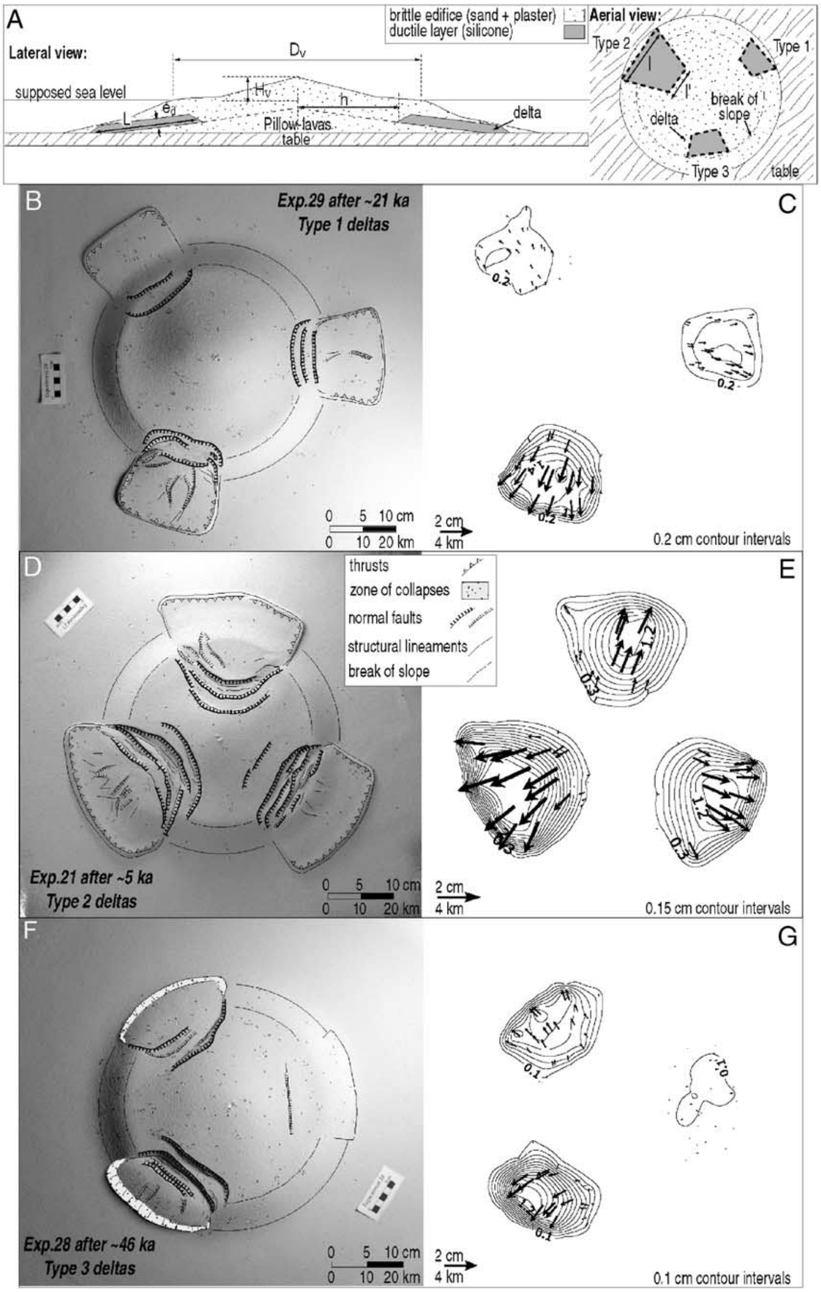

Fig. 4. The effect of ductile volcaniclastic deltas including in the structure of a brittle edifice on its stability. (A) Schematic general representation of initial stage of experiments. (B, D, F) Experiments 29, 21 and 28 photographs, where the three different types of delta are tested (see Table 2). Structural map is overlaid. (C, E, G) Contour maps of the calculated displacement field (in cm). Systems of normal faults have developed within the edifice, directly facing the deltas. Coastal (experiment 29), but also major subaerial slumps (experiments 21 and 28) are generated depending on the size and location of the deltas. Note that on the surface of the deltas grabens develop when deltas spread. They present a fan-shaped organization.

Table 2
Geometric characteristics of volcaniclastic deltas

Illustrations	Delta types	Delta sizes				Locations
		L/D_v	$1/D_v$	$1'/D_v$	e_d/H_v	h/D_v
Fig. 4B,C	Type 1	0.41	0.48	0.33	0.5	0.5
Fig. 4D,E	Type 2	0.55	0.74	0.37	0.5	0.25
Fig. 4F,G	Type 3	0.26	0.48	0.48	0.5	0.33

Parameters are illustrated on Fig. 4A.

steps, creating several collapse units. Grabens generally tangential to the break of slope are observed downhill from these faults. The observed structures and displacement fields show that deltas are not destabilized simultaneously. They spread one after the other under their own weight and that of the overlying edifice. The consequence is the development of slide structures that mainly affect edifice edge. We also varied the delta thickness and conclude that slide structures are generated more rapidly with increasing thickness.

3.3. Case C: only hyaloclastites or a hydrothermalized zone

Here a circular ductile section of diameter D_h and thickness E is incorporated at the supposed sea level within a brittle edifice, in order to simulate an upper LSL (Fig. 5A). This level could be regarded as either hyaloclastites or a hydrothermalized zone according to the ratio D_h/D_v; D_v being the subaerial edifice diameter. We considered that we simulated a hydrothermalized zone if D_h<1/3 D_v and hyaloclastites otherwise. We varied the parameters E and D_h and conclude that D_h mostly controls the deformation; E simply influencing deformation rate. We distinguish two situations:

(1) If D_h>8/9 D_v, the edifice surface is dissected by large but disorganized stretching structures (horsts and grabens) (Fig. 5B). Small collapses occur at the break of slope. Observations and displacement fields (Fig. 5C) indicate that the ductile silicone layer (hyaloclastites in this case) is deformed under the edifice load. Spreading is predominant and no slide structures are observed.
(2) If D_h≤8/9 D_v, nothing happens as the ductile layer is efficiently constrained. We thus agree with observations of Merle and Lénat (2003), for whom, destabilization of a low slope-angle shield volcano is unlikely to result from the collapse and spreading of a confined hydrothermally altered zone.

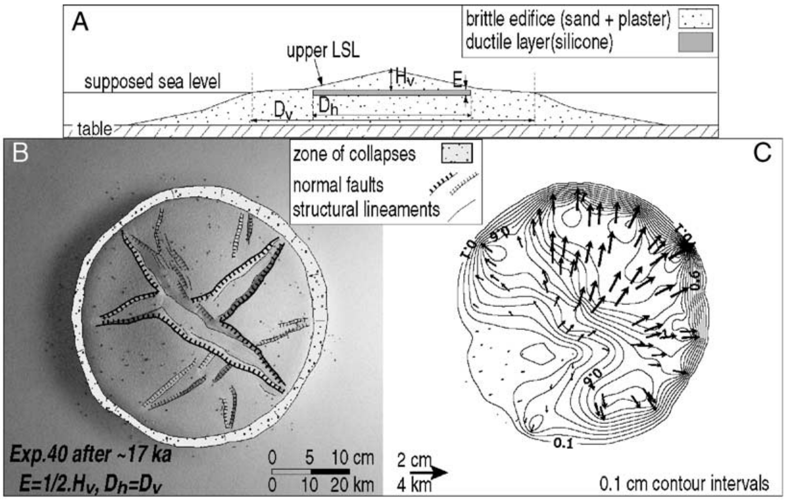

Fig. 5. The effect of a ductile hyaloclastic layer on the stability of a brittle shield volcano. (A) Schematic general representation of initial stage of experiments. (B) Experiment 40 photograph with structural map overlay. (C) Contour map of the calculated displacement field (in cm). The model illustrates the spreading of the hyaloclastic layer under the edifice load. Note that the edifice is mainly deformed towards the upper right corner. This dissymmetry is due to an imperfect construction; the ductile hyaloclastic layer being more confined by sand towards the lower left corner.

3.4. Case D: association of a marine substratum and deltas

Here we investigated the association of ductile marine substratum and fan-shaped deltas without contrast of viscosity between the LSLs (Fig. 6A). We varied the thickness e of the marine substratum, defined as a fraction of the edifice total height H_e, but also the type and the number n of deltas. We observed that the addition of an underlying ductile layer influences the generation of delta-related slide events, such as produced in delta only experiments (Fig. 4). In particular, the thickness e of this level is critical:

(1) If $e>1/10\ H_e$, the surface of the edifice is quickly marked by stretching structures. Spreading predominates and slides are inhibited.
(2) For $e<1/10\ H_e$, spreading is limited and slides predominate (Fig. 4).
(3) For $e=1/10\ H_e$, slides and spreading occur successively. The edifice is first affected by peripheral slumps (Fig. 4), then spreading becomes predominant with the formation of (a) concentric anticlines and thrusts all around the models and (b) major stretching structures (graben-horst-graben) at their surface (Fig. 6B). The development of these stretching structures is strongly linked with the type and the number of deltas. They are more marked for type 1 deltas. Grabens develop directly facing deltas and thus a 4-branch structure forms when $n=4$, a triangular structure when $n=3$ (Fig. 6B) and a 2-branch structure when $n=2$. Displacement analysis shows that these particular morphologies can be explained by a more rapid spreading of the deltas compared with the rest of the edifice (Fig. 6C).

3.5. Case E: association of hyaloclastites or a hydrothermalized zone and deltas

Here we associated ductile fan-shaped deltas and an upper ductile LSL (hyaloclastites or a hydrothermalized zone, see case C, Fig. 7A). We varied the thickness E and the diameter D_h of the upper LSL, but also the type of deltas. For hyaloclastites, we also studied the effect of a contrast of viscosity between the upper LSL and deltas.

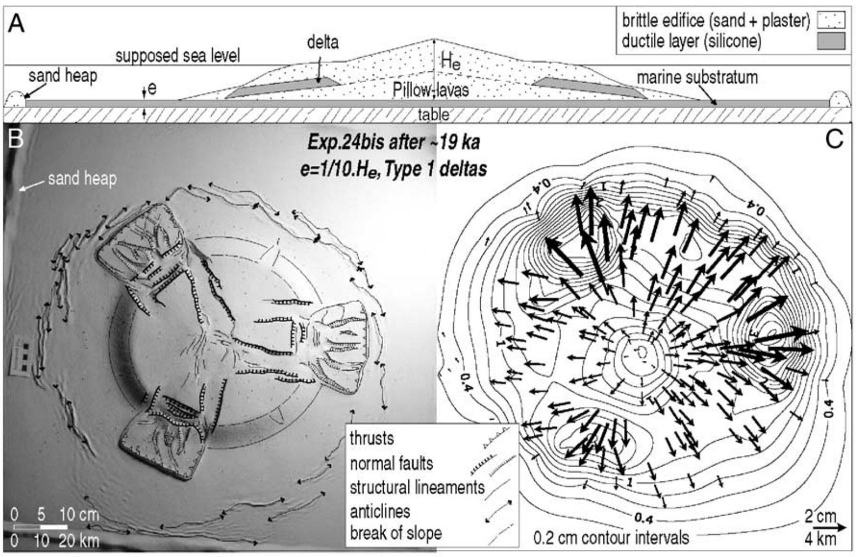

Fig. 6. The effect of the association of ductile volcaniclastic deltas and a marine substratum on the stability of a brittle shield volcano. (A) Schematic general representation of initial stage of experiments. (B) Experiment 24bis photograph with structural map overlay. (C) Contour map of the calculated displacement field (in cm). The model is marked by spreading-related structures, whose development is controlled by initial delta-related slumps. A system of graben-horst-graben has thus developed at the edifice surface, directly facing deltas, forming a remarkable triangular structure.

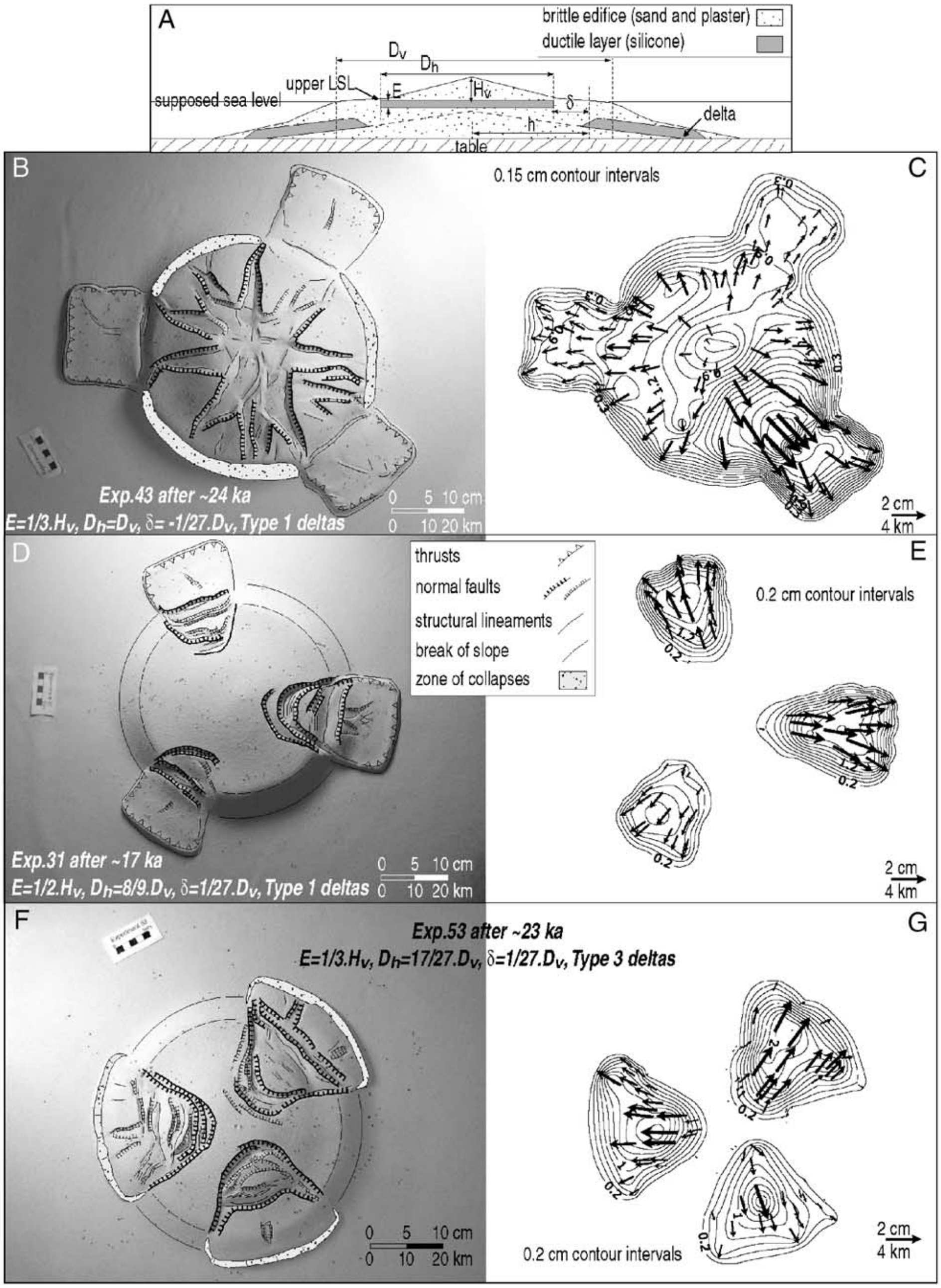

Fig. 7. The effect of the association of ductile volcaniclastic deltas and an upper ductile LSL (hyaloclastites or a hydrothermalized zone, depending on the ratio D_h/D_v) on the stability of a brittle edifice. Both LSLs have the same viscosity. (A) Schematic general representation of initial stage of experiments. (B, D, F) Experiments 43, 31 and 53 photographs with structural map overlay. (C, E, G) Contour maps of the calculated displacement field (in cm). Models show that slides and spreading are intimately linked. The predominance of one or another process depends on the diameter D_h of the upper LSL and the parameter δ.

3.5.1. 1st series: without contrast of viscosity

First we did experiments without contrast of viscosity. PDMS silicone is used to simulate both LSLs. In the models, the main parameters controlling deformation are the diameter D_h of the upper LSL and the value δ. δ is the distance from the uphill part of the deltas to the upper LSL (Fig. 7A). We define δ in the models as a fraction of the subaerial edifice diameter (D_v). Three situations are observed that strongly depend on the values D_h and δ; the thickness E simply influencing deformation rate:

(1) Whatever D_h and if $\delta>1/27\ D_v$, nothing happens apart from the peripheral edifice destabilization (Fig. 4). The upper LSL has no influence on the morphology of the model, as it stays constrained.

(2) If $D_h>8/9\ D_v$ and whatever δ, the upper LSL (hyaloclastites in this case) is free and quickly spreads. At the edifice surface, numerous stretching structures appear and form an alternation of triangular horsts and grabens with a star-shaped organization (Fig. 7B). At the break of slope, small collapses are observed. Horseshoe-shaped scarps are not clearly evident, but the late delta deformation nevertheless provokes collapses directly facing them.

(3) If $D_h\leq 8/9\ D_v$ and $\delta\leq 1/27\ D_v$, the upper LSL is efficiently constrained as it is buttressed at its boundaries. It can be destabilized only if it is liberated by delta deformation. The edifice is first destabilized by peripheral slumps (Fig. 4) that free the upper LSL on a length restricted to the deltas. The deformation of the upper LSL then allows migration of destabilization uphill and the subsequent formation of larger slide structures (Fig. 7D,F). Growing-size recurrent slumps are thus generated. Organized blocks sliding slowly downhill within slide structures are observed. The displacement field shows that only delta areas are affected by deformation (Fig. 7E,G). We therefore observe that similar morphologies characterizing growing-size recurrent slides can be generated considering different initial systems. The only necessary conditions are the association of two ductile LSL, an upper one, sufficiently confined (hyaloclastites or a hydrothermalized zone), and a lower one (peripheral deltas), both separated by a distance $\delta\leq 1/27\ D_v$. This observation agrees with those of Merle and Lénat (2003) for their analogue models simulating hybrid collapse in Piton de la Fournaise. They underline that the destabilization of an upper ductile layer is possible only if this layer is located near a lower one.

3.5.2. 2nd series: with a contrast of viscosity between hyaloclastites and deltas

If we only consider hyaloclastites, the most realistic model would be that where D_h=D_v, that is to say where the boundary of hyaloclastites is located exactly at the level of the break of slope. In this case, however, spreading is the predominant process and slides are secondary. This is clearly not the usual case in nature. The first solution was therefore to constrain the ductile hyaloclastic layer by sand at its boundary, increasing in effect the resistance to deformation. Another more realistic solution was to consider that hyaloclastites are less ductile than deltas and therefore to use two different silicones to model them. Such experiments were done considering hyaloclastites of diameter D_h=D_v, one order of magnitude more viscous than ductile deltas. Hyaloclastites are simulated with Rhodorsil Gomme silicone and deltas with PDMS silicone. For these experiments, the parameter controlling the deformation style and rate is the thickness E of hyaloclastites defined as a fraction of the subaerial edifice height (H_v). Two situations are observed:

(1) For $E\leq 1/3\ H_v$, the hyaloclastic level is subjected to only limited spreading and slide structures can develop (Fig. 8B). The edifice is mostly stable. The displacement field shows that mainly the peripheral part of the edifice facing the deltas is destabilized (Fig. 8C). Slide structures as already described (Fig. 4) are quickly generated. The delta deformation induces hyaloclastic layer destabilization, which collapses simultaneously. Peripheral slumps are therefore amplified by the induced destabilization of hyaloclastites and tend to migrate uphill to form growing-size recurrent events.

(2) For $E>1/3\ H_v$, spreading effects are noticeably more marked, with the formation at the

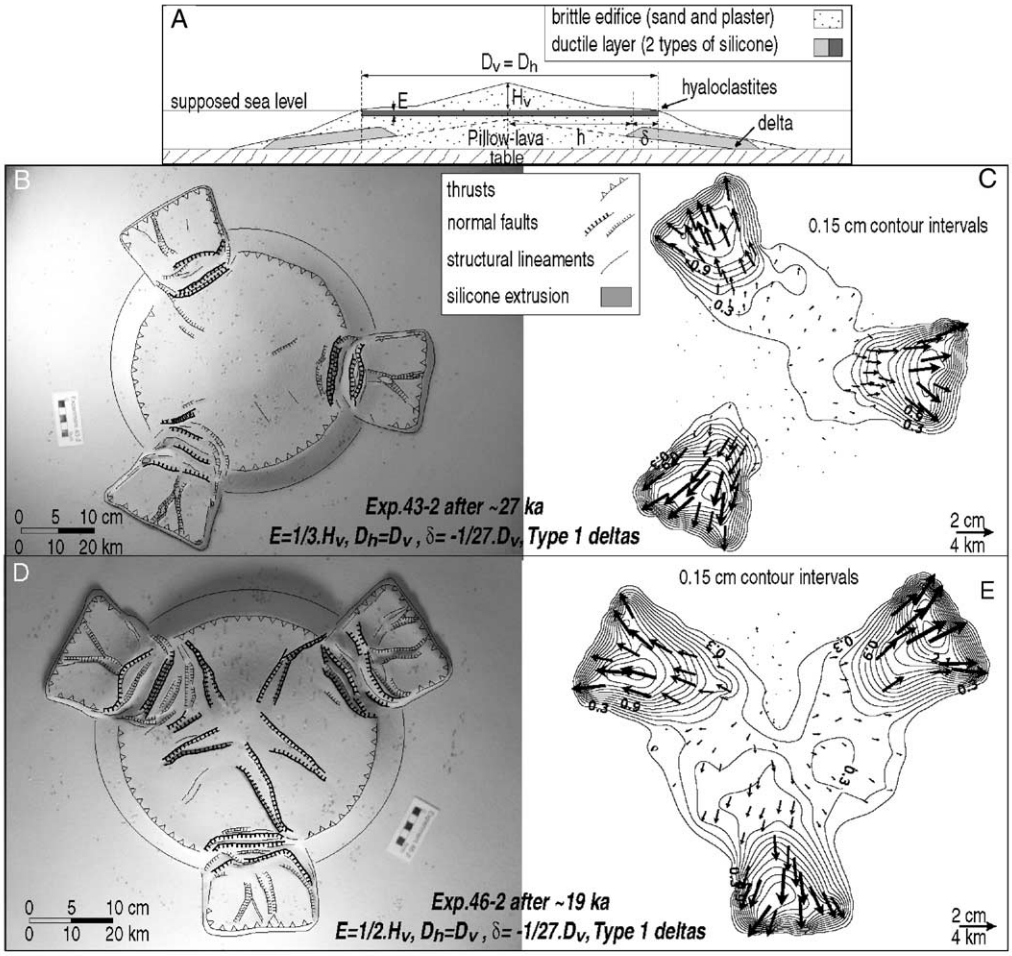

Fig. 8. The effect of the association of ductile volcaniclastic deltas and hyaloclastites (D_h=D_v) on the stability of a brittle edifice. Hyaloclastites are considered 1-order of magnitude more viscous than deltas. (A) Schematic general representation of initial stage of experiments. (B, D) Experiments 43-2 and 46-2 photographs with structural map overlay. (C, E) Contour maps of the calculated displacement field (in cm). The higher viscosity of the hyaloclastic layer is a key factor for limiting general edifice spreading. The main destabilization process can therefore be delta slide.

surface of the model of star-shaped grabens (Fig. 8D). We notice that grabens are not facing the collapses, but develop at their edges. Slide structures are however well developed and tend to migrate uphill, after the induced destabilization of the ductile hyaloclastic layer. Spreading and slide structures form simultaneously, as illustrated by displacement fields (Fig. 8E).

The higher viscosity of the hyaloclastic layer is therefore revealed to be a key factor for limiting general edifice spreading. Ductile deltas are however necessary to generate slide structures. The association of deltas and hyaloclastites allows the uphill migration of slumps to produce growing-size recurrent superficial events with a décollement high above the edifice base.

3.6. Case F: association of a marine substratum, an upper LSL and deltas

Finally, we simulated an edifice with 3 LSL types (Fig. 9A). We used experiments like in case

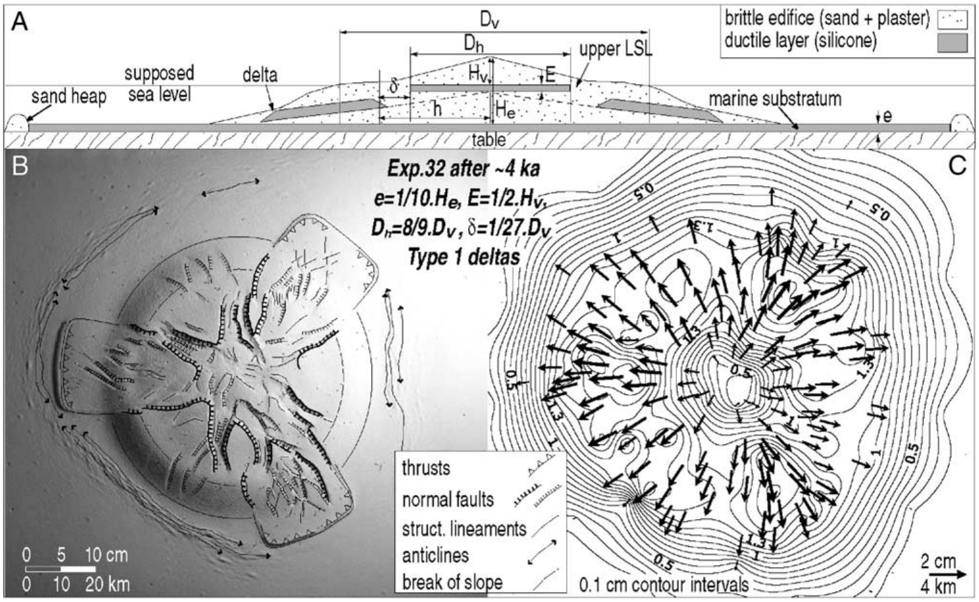

Fig. 9. The effect of the association of ductile volcaniclastic deltas, an upper LSL and a marine substratum on the stability of a brittle edifice. (A) Schematic general representation of initial stage of experiments. (B) Experiment 32 photograph with structural map overlay. (C) Contour map of the calculated displacement field (in cm). The edifice is clearly affected by spreading.

E and added a ductile layer under the models (marine substratum). The thickness e of this layer is fixed to 1/10 of the edifice total height (H_e), as we have already concluded that slides are inhibited with a larger thickness. Whatever the structure of the considered model or the chosen viscosity for LSLs, spreading is clearly predominant and slide structures almost non-existent (Fig. 9B). The surface of the model is marked by numerous radial horsts and grabens with a star-shaped organization. A belt of folds and thrusts developed all around. The displacement field is mainly radial (Fig. 9C). The presence of an underlying ductile layer in the structure of an edifice can therefore inhibit slide structure generation. The edifice spreads following the radial movement of the underlying ductile level, but also of the upper LSL. These experiments show that while the thickness of the pelagic layer is critical, the strength of the edifice resisting spreading is also important. The spreading of the edifice can probably be limited if the marine substratum is thinner than 0.3 cm (<1/10 H_e). It was difficult to do such experiments, as silicone films can not be produced.

4. Comparison with natural examples

4.1. Landslides

Our experiments show that landslides can form only by the deformation of LSLs under the edifice load. If small collapses identified with limited debris avalanches have been observed in some cases, most of the generated landslides are slumps. Hürlimann et al. (2001) underline that the causes of landslides include preparing factors contributing to the edifice destabilization and triggering mechanisms. We suggest that effects of LSLs constitute a preparing factor, as they contribute to the destabilization of the edifice by the development of slumps. Other mechanisms are needed to trigger debris avalanches such as seismicity, dyke intrusion, sea level changes, but they are not simulated here. We therefore propose that modeled slumps constitute the first destabilization stages of volcanic islands. These slumps can later evolve into debris avalanches when they become gravitationally unstable or because of other triggering mechanisms. This process was described by Labazuy (1996) for the east flank of Piton de la Fournaise and is also accepted for Canary Islands debris avalanche deposits (Urgeles et al., 1999).

Our scaled models suggest a major role of deltas in the formation of these slumps. The presence of a ductile fan-shaped delta in the structure of an edifice and its deformation under the weight of the volcano is in fact the best manner to generate slide structures. In our models, the larger and the thicker are deltas, the greater is the area affected by the slumps. Different-sized events can thus be observed:

(1) Coastal episodes are generated from the deformation of type 1 deltas (Fig. 4B and Table 2). They are small-size volcaniclastic deltas, whose uphill part is located under the break of slope of the edifice. We propose that they mainly result from the emplacement of sub-aerial material on the island flanks, either directly (for example, by a pyroclastic or lava flow entering in sea) or through erosion processes (sedimentary aprons). The destabilization of the edifice coastal part can thus be explained by the growth of the subaerial island over a weak delta. This can generate easily identifiable scarps in the lower flanks of an island. Reunion Island is marked by several breaks of slope in the coastal zone (Fig. 10A). Oehler et al. (2004a) proposed that they are source areas of debris avalanche events. These escarpments are similar to normal faults generated by experiment where type 1 deltas are deformed under their own weight and the load of the overlying edifice (Fig. 4B). We suggest that these scarps may result from the same process under the load of Piton des Neiges volcano. This may have first formed coastal slumps that may have then spawned debris avalanches. These structures are often located in the vicinity of mouths of La Reunion main rivers. We suspect that sedimentary aprons have developed directly offshore and may have contributed to the destabilization.

(2) Major subaerial slumps are generated by the deformation of type 2 and 3 deltas (Fig. 4D,F and Table 2). These deltas are deeply rooted in the edifice structure. They must be explained by processes having occurred during the early growth of the volcanic island, even before emergence. These LSLs may include primitive landslide deposits, turbidites and phreatomagmatic products. Another hypothesis is to consider the case of overlapping volcanoes. Large oceanic volcanoes are in fact often constructed by several overlapping edifices, where deltas of an earlier volcano are covered by a more recent edifice. These internal weak levels could have an influence on the stability of the younger volcano. Subaerial major events, including an entire section of the edifice can thus be generated, solely by deformation of such deltas under the volcano load. This may be an explanation for the formation of the Hilina slump (Kilauea, Hawaii). We have reproduced a morphology comparable to Hilina by deformation of a type 3 delta (compare Figs. 4F and 10B). The model has a similar organization with uphill normal faults, a massive displaced block and debris avalanches at the front. We have also calculated an average displacement rate of about 6 cm/year (Table 1) that is comparable with the rate of 10 cm/year estimated or measured in reality (Moore et al., 1989; Moore and Clague, 1992; Delaney et al., 1998). The Hilina slump may therefore be explained by delta deformation. A massive 2 km-thick section of volcaniclastic materials, such as sandstone and debris-flow breccias, has been recognized at the base of Kilauea's south flank (Naka et al., 1998). Lipman et al. (2002) note that several sampled clasts of this section have slickensided or faceted surfaces, indicating deformation. They add that these clasts mainly have a Mauna Kea and/or Kohala origin. All these observations agree with the existence of a paleo-volcaniclastic delta of Mauna Kea and/or Kohala origin that was covered by the young Kilauea volcano. The deformation of these weak materials under the load of Kilauea may explain the Hilina slump. Another example of shield-like volcano affected by delta-related destabilizations may be the island of Martinique (Antilles, Fig. 10C). The west side of this island is dominated by a series of concave, west-facing structures, the most recent of which are clearly faults and may be the source zones of major debris avalanches identified offshore (Deplus et al., 2001). The present volcanic topography is built on a set of Tertiary volcanoes that lie to the west of the

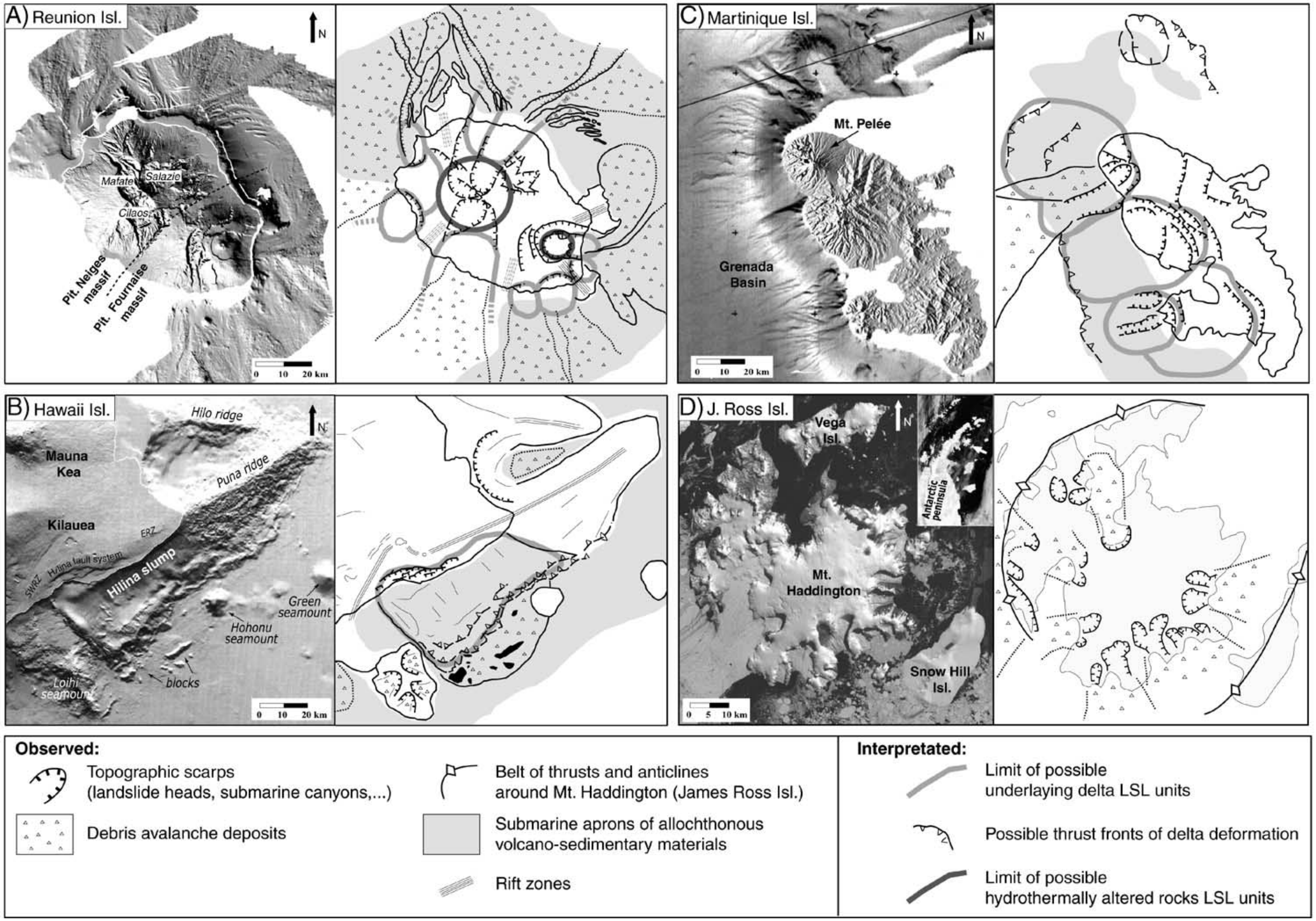

Fig. 10. Four shield-like volcanic islands. Well constrained structural features are shown in solid lines, more hypothetical ones in dotted lines. The location of possible LSL units is proposed. (A) Shaded relief map of Reunion Island edifice from Oehler et al. (2004a). Reunion Island appears to be mainly affected by landslides. Triangular horsts at the north-east of the subaerial part are thought to be spreading-related (van Wyk de Vries et al., 2001b). (B) Shaded-relief map of the south-east Hawaii region from Moore and Chadwick (1995). The area is characterized by the active Hilina slump, whose origin is to be found in the joint action of rift zones and deformation of the edifice on LSLs. (C) Shaded-relief map of Martinique Island. The submarine part is from Deplus et al. (2001). The subaerial DEM is from NASA SRTM archive. The hypothetical extent of the Tertiary volcanoclastic delta LSL sequences is shown. They are suspected to play a role in the generation of some large debris avalanches. (D) Satellite photographs of James Ross Island (Antarctica). Inset shows the location of the island in the Antarctic Peninsula region. The island is characterized by several U-shaped amphitheaters we propose to be related to landslides and surrounded by a broad annular outcrop resulting from edifice sagging.

island's axis. Erosion of these will have created large deltas to the west in the Grenada Basin. Martinique has therefore grown at least partially on a sloping set of deltas that may have played a role in its destabilization. Other large volcanoes in the Antilles have similar west-directed concave structures and major debris avalanche deposits in the Grenada Basin. We thus suggest that the main process in generating and locating these events is long term deformation on delta LSLs related to the Tertiary arc.

The growth of slumps is more efficient when deltas are associated with another upper LSL, such as hyaloclastites or hydrothermally altered rocks. In this case, the first coastal slumps due only to delta deformation migrate within the edifice and affect larger and larger sections of the volcano. If triggering mechanisms occur during this process, recurrent debris avalanches can be generated. Using these observations, we propose an explanation for the recurrence of mass-wasting events generally observed for volcanic islands. At La Reunion, Bachèlery et al. (2003) describe at least 4 debris avalanches having affected the west flank of Piton des Neiges. The lower thicker unit contains hydrothermally altered fragments and is therefore regarded as resulting from a landslide that cut the central altered part of Piton des Neiges. We propose that the observed slide deposits may be explained by the successive deformation of a lower delta and of an upper hydrothermalized zone (Fig. 7F). The destabilization of the edifice coastal part by the deformation of a delta may have freed the weak central hydrothermalized part of Piton des Neiges and contributed to the development of a main landslide. Recurrent slide episodes were later generated by the continuous deformation of the Piton des Neiges central core. We have also found field evidence for the recurrence of slides at Mount Haddington volcano on James Ross Island (Antarctica) (Van Wyk de Vries et al., 2003; Oehler et al., 2004b). The morphology of this Antarctic shield-like volcano is characterized by several large well-formed U-shaped structures (Fig. 10D). Such amphitheatres have been reproduced by the association of deltas and hyaloclastites (Fig. 7D). We propose that these structures are related to slides and suggest that they result from the deformation of these LSLs. Hyaloclastites in fact constitute a main part of the volcanic section of this island and field observations have demonstrated that this level is potentially very weak.

4.2. Spreading

For the case of shield-like volcanic islands, it appears from our experiments that gravitational spreading, as described by Merle and Borgia (1996), is not at the origin of landslides. In fact, if an edifice is clearly destabilized on a marine substratum, only stretching and subsidence occur (Fig. 3). We also conclude that gravitational spreading tends to inhibit slide formation. Spreading is thus predominant and sliding absent if underlying sediments are thicker than 1/10 of the volcanic edifice total height (Table 1). For a lower sediment thickness, landslides can be generated, as spreading is slow compared with deltas deformation. For exactly 1/10 of the total height, landslides can occur first, but are rapidly inhibited as spreading takes over. This case is encountered in Reunion Island, where an about 600-m-thick preexisting marine substratum (de Voogd et al., 1999) underlies an about 6000-m-high edifice. Theoretically, the Reunion edifice could therefore be affected by landslides first and spreading later. For landslides, Oehler et al. (2004a) have shown that at least 15 main events have affected the edifice, while some possible spreading-related normal and thrust faults have been described by Van Wyk de Vries et al. (2001b) from field observations and digital terrain model analysis. In particular, triangular horsts appear in the morphology of La Reunion and are similar to those developed in models clearly dominated by the spreading process (compare Figs. 3 and 10A). A mechanical decoupling between the edifice and the top of the oceanic basement is confirmed by seismic data, however the imaged sediments appear undeformed (de Voogd et al., 1999). Gravitational spreading results in deformation of the marine substratum with the development of concentric anticlines and thrusts all around the edifice as shown by our experiments (Figs. 3, 6, 9). However, we observed that these compressive structures do not necessarily appear at the base of the volcano. Their appearance is dependant on the constraints applied to the underlying ductile level. They are generally located about 20 km from the volcano in our models (Fig. 3), but could occur further, if the marine substratum is unconstrained. For Hawaii, the marine substratum is estimated to be 0.08 to 0.1 km-thick (Leslie et al., 2002) for an about 9 km-high edifice. In these conditions, gravitational spreading should be non-existent. However, works by Morgan et al. (2000) and Morgan and Clague (2003) support the existence of spreading for Mauna Loa and Kilauea volcanoes. This can be explained by volcanic spreading (Borgia, 1994); the low thickness of the marine substratum being compensated by the action of rift zones. James Ross Island (Antarctica) is built on a very thick Cretaceous sedimentary substratum (possibly up to 6 km-thick, Van Wyk de Vries et al., 2003; Oehler et al., 2004b). According to our scaled models, the about 1.5 km-high volcano must therefore spread. However, we have found field evidence of slumps that appear not to be compatible with our general model (Fig. 10D). Van Wyk de Vries and Matela (1998) observed that a volcanic edifice spreads laterally when underlain by a thin ductile layer, but they noticed that thicker ductile layers lead to inward flexure (or sagging). Sagging of an edifice is marked mainly by vertical movements and edifice compression that could increase delta-related deformation and the formation of slumps. We therefore suggest that if a threshold is exceeded for the marine substratum thickness, spreading is replaced by sagging allowing the development of slides.

The origin of the large Salazie, Mafate and Cilaos "cirques" of Piton des Neiges (La Reunion) (Fig. 10A) is still not understood. Several hypotheses have been proposed to explain their formation including subsidence of dense bodies (Lénat and Merle, pers. com.), erosion (Kieffer, 1990), but also spreading (Van Wyk de Vries et al., 2001b) and landslides (Oehler et al., 2004a). We suggest an alternative including the last three processes with a predominant role of landslides. We have in fact reproduced a morphology comparable to Piton des Neiges with the development of large star-shaped grabens in the central part of the edifice by associating, in our models, deltas with either a marine substratum or a hyaloclastic level (Figs. 6 and 8D respectively). We propose that the "cirques" of Piton des Neiges can be explained by: (1) large delta-related slumps with a possible role of a central hydrothermalized zone (Fig. 7F), which have evolved in debris avalanches as indicating by deposits identified off the island (Oehler et al., 2004a), (2) subsequent development of spreading-related structures controlled by the first-generated slides and (3) continual rapid torrential tropical erosion. Our experiments support that spreading of the edifice (Piton des Neiges in the case of La Reunion) may be related to the deformation of the volcano on its marine substratum (Fig. 6), but also be caused by deformation on hyaloclastites (Fig. 8D). For deformation, the hyaloclastic layer must be unconstrained and at least 1/3 as thick as the height of the subaerial edifice (Table 1). This kind of deformation can theoretically affect Ascension Island where at least 1 km-thick hyaloclastites have been identified by drilling (Nielson and Stiger, 1996; Klingelhöfer et al., 2001) under an about 1 km-high subaerial edifice. At La Reunion, we suspect the existence of such weak hyaloclastic layer although no field evidence exists.

4.3. Genesis and influence of rift zones

Gravitational destabilization of a shield-like volcano creates extensional structures. Spreading forms alternations of horsts and grabens with a star-shaped organization at the edifice surface (Figs. 3, 5, 7B, 8D, 9), whereas delta-related landslides develop grabens generally tangential to the break of slope downhill from horseshoe-shaped headwalls (Fig. 4). These zones of extension are preferential pathways for dike intrusion and constitute potential areas for rift zone development (Walter, 2003). Thus, in nature, rift zones could form, (1) either radially, if the edifice is affected by spreading, (2) or parallel to the headwall of an active slump, as observed for Hilina (Fig. 10B). The development of spreading-related structures is moreover intimately controlled by delta destabilization. Grabens thus often develop at delta edges (Figs. 6, 7B, 8D, 9). This may lead to the formation of rift zones surrounding the entire delta-sliding sector, reminiscent for example Piton de la Fournaise morphology (Fig. 10A). We consequently agree with Walter (2003) and Walter and Troll (2003) that rift zone location is strongly controlled by edifice gravitational destabilization.

4.4. Submarine canyons

Submarine canyons are observed around several islands including La Reunion (Oehler et al., 2004a), La Palma (Canary Islands) (Urgeles et al., 1999) and Antilles (Deplus et al., 2001). These canyons are mainly thought to have an erosive origin. However, fan-shaped grabens described at the surface of the deltas in our models are morphologically like these submarine canyons (compare Fig. 10A, C with, for example, Fig. 6B). Their development during the deformation of deltas therefore suggests that a tectonic component may have been added in their formation process. They may result of the spreading of slide deposits after their emplacement.

5. Summary and conclusions

The main conclusions of this experimental study are the following:

(a) The gravitational destabilization of an oceanic hot-spot or arc shield volcano on LSLs creates landslides and/or spreading of the affected edifice.
(b) The formation of slides is strongly linked with the presence in the edifice structure of fan-shaped volcaniclastic deltas. In our experiments, generated events are only slumps. Debris avalanches could not be triggered due to laboratory constraints. The development of these slumps is more efficient if deltas are associated with another upper LSL, such as hyaloclastites or hydrothermally altered rocks. In this case, recurrent events generating collapses with increasing width are observed.
(c) Gravitational spreading occurs when the edifice is underlain by a ductile marine substratum or marked by the presence of an intermediate, unconstrained hyaloclastic layer, thick as at least 1/3 of the height of the subaerial edifice.
(d) The spreading process is not the origin of large landslides at shield-like volcanic islands. It only provokes stretching and subsidence of the affected edifice. Gravitational spreading of a volcano on its marine substratum actually tends to inhibit slide formation, especially when the thickness of the sedimentary layer is larger than 1/10 of the edifice total height. Slumps can only develop if spreading occurs later in the destabilization process that is to say if the thickness of the marine substratum is equal or lower to this value.
(e) Structural features observed on several oceanic hot-spot or arc shield volcanoes can be attributed to the deformation of edifices on LSLs. One structural control is the location of rift zones (Walter and Troll, 2003). LSLs may be important at Reunion Island which is affected by both spreading and landslides. The formation of its “cirques” can be explained by the joint action of these processes, and with erosion. The Hilina slump (Kilauea, Hawaii) may be attributed to the deformation of the south flank of the island on a volcaniclastic delta. The particular morphology of James Ross Island (Antarctica) is certainly linked with a simultaneous deformation of deltas and hyaloclastites made easier by edifice sagging. The role of delta deformation may be important in the destabilization of Martinique Island and other Lesser Antilles islands. Our model for the destabilization of volcanic islands is therefore applicable to all shield-like edifices.

Acknowledgments

The paper benefited from helpful reviews and comments by Marco Bonini, Valentin Troll and Valerio Acocella. The authors moreover thank Olivier Merle who contributed to its improvement. This work was supported by a French PNRN grant to Benjamin van Wyk de Vries. Works on Haddington volcano (James Ross Island, Antarctica) was supported by the British Antarctic Survey.

References

Ablay, G., Hürlimann, M., 2000. Evolution of the north flank of Tenerife by recurrent giant landslides. J. Volcanol. Geotherm. Res. 103, 135–159.
Bachèlery, P., Robineau, B., Courteaud, M., Savin, C., 2003. Avalanches de débris sur le flanc occidental du volcan-bouclier Piton des Neiges (Reunion). Bull. Soc. Geol. Fr. 174, 125–140.
Borgia, A., 1994. Dynamic basis of volcanic spreading. J. Geophys. Res. 99, 17791–17804.
Borgia, A., Treves, B., 1992. Volcanic plates overriding the oceanic crust: structure and dynamics of Hawaiian volcanoes.

In: Parson, L.M., Murton, B.J., Browning, P. (Eds.), Ophiolites and their modern oceanic analogues. The Geological Society, London, pp. 277–299.

Carena, S., Borgia, A., Battaglia, A., Pasquare, M., Ferraris, M., Martelli, L., De Nardo, M.T., 2000. Gravity synclines. J. Geophys. Res. 105, 21819–21833.

Clague, D.A., Denliguer, R.P., 1994. Role of olivine cumulates in destabilizing the flanks of Hawaiian volcanoes. Bull. Volcanol. 56, 425–434.

Clément, J.P., Legendre, C., Caroff, M., Guillou, H., Cotton, J., Bollinger, C., Guille, G., 2002. Epiclastic deposits and "horseshoe-shaped" calderas in Tahiti (Society Islands) and Ua Huka (Marquesas Archipelago), French Polynesia. J. Volcanol. Geotherm. Res. 120, 87–101.

Clouard, V., Bonneville, A., Gillot, P.Y., 2001. A giant landslide on the southern flank of Tahiti Island, French Polynesia. Geophys. Res. Lett. 28, 2253–2256.

Day, S.J., 1996. Hydrothermal pore fluid pressure and the stability of porous, permeable volcano. In: Mc Guire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. Geol. Soc. London, pp. 295–306.

Delaney, P.T., Denliguer, R.P., Lisowski, M., Miklius, A., Okubo, P.G., Okamura, A.T., Sako, M.K., 1998. Volcanic spreading at Kilauea, 1976–1996. J. Geophys. Res. 103, 18003–18023.

DePaolo, D.J., Stolper, E., Thomas, D.M., 2001. Deep Drilling into a Hawaiian Volcano. Eos, Am. Geophys. Union 82 (13), 149–155.

Deplus, C., Le Friant, A., Boudon, G., Komorowski, J.C., Villemant, B., Harford, C., Ségoufin, J., Cheminée, J.L., 2001. Submarine evidence of large-scale debris avalanches in the Lesser Antilles Arc. Earth Planet. Sci. Lett. 192, 145–157.

de Voogd, B., Pou Palomé, S., Hirn, A., Charvis, P., Gallart, J., Rousset, D., Danobeitia, J., Perroud, H., 1999. Vertical movements and material transport during hotspot activity: Seismic reflection profiling off shore La Reunion. J. Geophys. Res. 104, 2855–2874.

Donnadieu, F., 2000. Déstabilisation des édifices volcaniques par les cryptodômes: Modélisation analogique et approche numérique. PhD thesis, Univ. Blaise Pascal, Clermont-Ferrand, France.

Donnadieu, F., Merle, O., 1998. Experiments on the indentation process during cryptodome intrusions; new insights into Mount St. Helens deformation. Geology 26, 79–82.

Donnadieu, F., Kelfoun, K., van Wyk de Vries, B., Cecchi, E., Merle, O., 2003. Digital photogrammetry as a tool in analogue modeling: applications to volcano instability. J. Volcanol. Geotherm. Res. 123, 161–180.

Elsworth, D., Day, S.J., 1999. Flank collapse triggered by intrusion: the Canarian and Cape Verde Archipelagoes. J. Volcanol. Geotherm. Res. 94, 323–340.

Filmer, P.E., McNutt, M.K., Webb, H.F., Dixon, D.J., 1994. Volcanism and Archipelagic Aprons in the Marquesas and Hawaiian Islands. Mar. Geophys. Res. 16, 385–406.

Garcia, M.O., Davis, M.G., 2001. Submarine growth and internal structure of ocean island volcanoes based on submarine observations of Mauna Loa volcano, Hawaii. Geol. Soc. Amer. Bull. 29, 163–166.

Holcomb, R., Searle, R., 1991. Large landslides from oceanic volcanoes. Mar. Geotechnol. 10, 19–32.

Hubbert, M.K., 1937. Theory of scale models as applied to the study of geologic structures. Geol. Soc. Amer. Bull. 48, 1459–1519.

Hürlimann, M., Ledesma, A., Marti, J., 2001. Characterisation of a volcanic residual soil and ist implications for large landslide phenomena: application to Tenerife, Canary Islands. Eng. Geol. 59, 115–132.

Iverson, R.M., 1995. Can magma-injection and groundwater forces cause massive landslides on Hawaiian volcanoes? J. Volcanol. Geotherm. Res. 66, 295–308.

Kerle, N., van Wyk de Vries, B., 2001. The 1998 debris avalanche at Casita Volcano, Nicaragua; Investigation of structural deformation as the cause of slope instability using remote sensing. J. Volcanol. Geotherm. Res. 105, 49–63.

Kieffer, G., 1990. Grands traits morphologiques de l'Ile de la Reunion. In: Lénat, J.F. (Ed.), Le volcanisme de la Reunion, monographie. Cent. Rech. Volcanol. Clermont-Ferrand, France, pp. 75–114.

Klingelhöfer, F., Minshull, T.A., Blackman, D.K., Harben, P., Childers, V., 2001. Crustal structure of Ascension Island from wide-angle seismic data: implications for the formation of near-ridge volcanic islands. Earth Planet. Sci. Lett. 190, 41–56.

Krastel, S., Schmincke, H.U., Jacobs, C.L., Rihm, R., Le Bas, T.M., Alibés, B., 2001. Submarine landslides around the Canary Islands. J. Geophys. Res. 106, 3977–3997.

Labazuy, P., 1996. Recurrent landslides events on the submarine flank of Piton de la Fournaise volcano (Reunion Island). In: Mc Guire, W.J., Jones, A.P., Neuberg, J. (Eds.), Volcano Instability on the Earth and Other Planets. J. Geol. Soc. London, pp. 295–306.

Le Friant, A., Boudon, G., Komorowski, J.C., Deplus, C., 2002. L'île de la Dominique, à l'origine des avalanches de débris les plus volumineuses de l'arc des Petites Antilles. C. R. Géosci. 334, 235–243.

Lénat, J.F., Vincent, P., Bachèlery, P., 1989. The off-shore continuation of an active basaltic volcano: piton de la Fournaise (Reunion Island, Indian Ocean): structural and geomorphological interpretation from Sea Beam mapping. J. Volcanol. Geotherm. Res. 36, 1–36.

Leslie, S.C., Moore, G.F., Morgan, J.K., Hills, D.J., 2002. Seismic stratigraphy of the Frontal Hawaiian Moat: implications for sedimentary processes at the leading edge of an oceanic hotspot trace. Mar. Geol. 184, 143–162.

Lipman, P.W., Lockwood, J.P., Okamura, R.T., Swanson, D.A., Yamashita, K.M., 1985. Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea volcano, Hawaii. U. S. Geol. Surv. Prof. Pap. 1276 (45 pp.).

Lipman, P.W., Sisson, T.W., Ui, T., Naka, J., Smith, J.R., 2002. Ancestral submarine growth of kilauea volcano and instability of its South flank. In: Takahashi, E., Lipman, P.W., Garcia, M.O., Naka, J., Aramaki, S. (Eds.), Hawaiian Volcanoes: Deep Underwater Perspectives. Geophys. Monograph, vol. 128. Am. Geophys. Union, pp. 161–191.

Lopez, D.L., Williams, S.T., 1993. Catastrophic volcanic collapse: relation to hydrothermal processes. Science 260, 1794–1796.

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

Merle, O., Borgia, A., 1996. Scaled experiments of volcanic spreading. J. Geophys. Res. 101, 13805–13817.

Merle, O., Lénat, J.F., 2003. Hybrid collapse mechanism at Piton de la Fournaise volcano, Reunion Island, Indian Ocean. J. Geoph. Res. 108, 2166.

Merle, O., Vendeville, B., 1995. Experimental modeling of thin-skinned shortenng around magmatic intrusions. Bull. Volcanol. 57, 33–43.

Moore, J.G., 2001. Density of basalt core from Hilo drill hole, Hawaii. J. Volcanol. Geotherm. Res. 112, 221–230.

Moore, J.G., Chadwick, W.W., 1995. Offshore geology of Mauna Loa and adjacent areas, Hawaii. Mauna Loa Revealed: Structure, Composition, History and Hazards. Geophysical monograph, vol. 92. AGU.

Moore, J.G., Clague, D.A., 1992. Volcano growth and evolution of the island of Hawaii. Geol. Soc. Amer. Bull. 104, 1471–1484.

Moore, J.G., Clague, D.A., Holcomb, R.T., Lipman, P.W., Normark, W.R., Torresan, M.E., 1989. Prodigious submarine landslides on the Hawaiian Ridge. J. Geophys. Res. 94, 17465–17484.

Moore, J.G., Normark, W.R., Holcomb, R.T., 1994. Giant hawaiian landslides. Annu. Rev. Earth Planet. Sci. 22, 119–144.

Morgan, J.K., Clague, D.A., 2003. Volcanic spreading on Mauna Loa volcano, Hawaii: evidence from accretion, alteration, and exhumation of volcaniclastic sediments. Geology 31, 411–414.

Morgan, J.K., Moore, G.F., Hills, D.J., Leslie, S., 2000. Over-thrusting and sediment accretion along Kilauea's mobile south flank, Hawaii: evidence for volcanic spreading from marine seismic reflection data. Geology 28, 667–670.

Naka, J., Shipboard Scientists, 1998. Preliminary results of the KAIREI KR98-08, 09 cruises around the Hawaiian Islands. Eos, Trans. Am. Geophys. Union 79 (45), F1022 (Suppl).

Nielson, D.L., Stiger, S.G., 1996. Drilling and evaluation of Ascension #1, a geothermal exploration well on Ascension Island, South Atlantic Ocean. Geothermics 25, 543–560.

Oehler, J.F., Labazuy, P., Lénat, J.F., 2004a. Recurrence of major flank landslides during the last 2 Ma-history of Reunion Island. Bull. Volcanol. 66, 585–598.

Oehler, J.F., van Wyk de Vries, B., Smellie, J., 2004b. How oceanic islands become destabilized? Ideas from landslides on James Ross Island (Weddell Sea, Antarctica). EGU 1st General Assembly, 25–30th April 2004. Nice.

Ramberg, H., 1981. Gravity, deformation and the Earth's crust. Academic, San Diego, CA. 452 pp.

Siebert, L., 1984. Large volcanic debris avalanches: characteristics of source areas, deposits and associated eruptions. J. Volcanol. Geotherm. Res. 22, 163–197.

Smith, J.R., Malahoff, A., Shor, A.N., 1999. Submarine geology of the Hilina slump and morpho-structural evolution of Kilauea volcano, Hawaii. J. Volcanol. Geotherm. Res. 94, 59–88.

Tilling, R.I., Koyanagi, R.Y., Lipman, P.W., Lockwood, J.P., Moore, J.G., Swanson, D.A., 1976. Earthquakes and related catastrophic events, Island of Hawaii, November 29, 1975: a preliminary report. U.S. Geol. Surv. Circ. 740 (33 pp.).

Urgeles, R., Masson, D.G., Canals, M., Watts, A.B., Le Bas, T., 1999. Recurrent large-scale landsliding on the west flank of La Palma, Canary Islands. J. Geophys. Res. 104, 25331–25348.

van Wyk de Vries, B., Francis, P.W., 1997. Catastrophic collapse at statovolcanoes induced by gradual volcano spreading. Nature 387, 387–390.

van Wyk de Vries, B., Matela, R., 1998. Styles of volcano-induced deformation: numerical models of substratum flexure, spreading and extrusion. J. Volcanol. Geotherm. Res. 81, 1–18.

van Wyk de Vries, B., Merle, O., 1996. The effect of volcanic constructs on rift fault patterns. Geology 24, 643–646.

van Wyk de Vries, B., Kerle, N., Petley, D., 2000. Sector collapse forming at Casita volcano, Nicaragua. Geology 28, 167–170.

van Wyk de Vries, B., Self, S., Francis, P.W., Keszthelyi, L., 2001a. A gravitational spreading origin for the Socompa debris avalanche. J. Volcanol. Geotherm. Res. 105, 225–247.

van Wyk de Vries, B., Cecchi, E., Robineau, B., Merle, O., Bachèlery, P., 2001b. Factors governing the volcano-tectonic evolution of La Reunion Island: a morphological, structural and laboratory modeling approach. EUG XI, 8–12th April 2001, Strasbourg.

van Wyk de Vries, B., Oehler, J.F., Smellie, J., 2003. Gravitational tectonics at Mt Haddington, an antarctic submarine/subglacial volcano. EGS-AGU-EUG Joint Assembly, 6–11th April 2003. Nice.

Varnes, D.J., 1978. Slope movement types and processes. Landslides, Analysis and Control, Spec. Rep. 176. In: Schuster R.L., Krizek R.J. (Eds.), National Academy of Sciences, Washington, D. C., 1973, pp 11–33.

Walter, T.R., 2003. Buttressing and fractional spreading of Tenerife, an experimental approach on the formation of rift zones. Geophys. Res. Lett. 30 (6), 1296.

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

Weigel, W., Grevemeyer, I., 1999. The Great Meteor seamount: seismic structure of a submerged intraplate volcano. Geodynamique 28, 27–40.