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Gravitational spreading controls rift zones and flank instability on El Hierro, Canary Islands

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Abstract – Ocean island volcanoes frequently develop local rift zones associated with flank movement and flank collapses. The ocean island El Hierro grew by coalescence and collapse of three volcanic edifices, which are an elongated topographic ridge (the Southern Ridge) and two semi-circular volcanic cones (Tiñor volcano, El Golfo volcano). During edifice growth and volcano coalescence, eruption fissures nucleated into rift zones that developed a complex triangle pattern. In scaled analogue experiments we could successfully reproduce the geometry of rift zones and unstable flanks as observed on El Hierro. The experimental results suggest that the rift configuration on El Hierro is the result of gravitational volcano spreading over deformable basal substrata, rather than of deep-seated magma updoming as thought previously. This paper elucidates the importance of the basal substratum and gravitational spreading, and the relationship to rifting and flank instability on El Hierro Island, and may help in understanding similar volcano architectures elsewhere.

Keywords: rift zones, volcano-tectonic structure, gravity tectonics, structural geology, experimental studies.

1. Introduction

Growing ocean island volcanoes are often structurally unstable and subject to gravity-driven deformation and giant landslides. Processes of gravitational deformation occur in various dimensions and time scales, including gradual (slow) movement of large edifice sectors, as well as catastrophic (rapid) flank collapses that may form far-travelling debris avalanches (Siebert, 1984; Holcomb & Searle, 1991; McGuire, 1996; Voight & Elsworth, 1997; Carracedo, 1999). Gradual flank movement, hereafter referred to as 'volcano spreading', is often associated with the mechanical decoupling of parts of the volcanic edifice from its basal substratum (Borgia, 1994; Denlinger & Okubo, 1995; van Wyk de Vries & Borgia, 1996; Borgia, Delaney & Denlinger, 2000). Preferred detachment zones may be provided by weak hydrothermally altered rock or marine sediments at the volcano base, as suggested for the volcanoes Kilauea and Mauna Loa on Hawaii (e.g. Delaney *et al.* 1998; Morgan & Clague, 2003; Morgan & McGovern, 2005). Associated with volcano spreading, dyke intrusions are often parallel and group into swarms to form rift zones (Fiske & Jackson, 1972; Nakamura, 1980; Dieterich, 1988; Walker, 1992; Walter, 2003; Klügel *et al.* 2005). Structural studies of the Hawaiian and Canary Islands showed that many unstable flanks and giant landslides are enclosed by rift zones (Lipman, 1980; Lipman *et al.* 1985; Moore *et al.* 1989; Carracedo, 1994, 1996*a*). This suggests a two-way interaction between volcano spreading and rift zone growth, where lateral flank displacement causes extension needed for sustained rift zone intrusions and, in turn, forceful intrusions push the unstable flanks further seaward and accelerate flank movement (Dieterich, 1988; Denlinger & Okubo, 1995). Volcano spreading is thought to be essentially driven by gravity and is controlled by the morphology of a growing volcano (Dieterich, 1988; Borgia, Delaney & Denlinger, 2000). Likewise, rift zone orientations are largely influenced by morphology, reportedly following pronounced topographic ridges and scarps (Fiske & Jackson, 1972; Walter & Schmincke, 2002). In order to test further the interaction between volcano growth, spreading and rifting, we investigate how rift zones and flank instability develop on a consecutively growing and reshaping volcano. This paper examines how a volcanic edifice's growth history influences (1) the direction of flank displacement, (2) the geometric development of rift zones and (3) the susceptibility of flanks to form giant landslides. The edifice studied is the island of El Hierro, Canary Islands.

The paper is organized as follows. First, we review the structure of El Hierro by focusing on four major units relevant for this study, which are (a) the volcano basement, (b) merging volcanic centres, (c) rift zones, and (d) unstable flanks. Second, we employ analogue

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Table 1. Sequence and age of constructive and destructive periods during the evolution of El Hierro

	Construction	Destruction	Age (ka)
		El Golfo collapse	134–13
Rift volcanism	Simultaneous activity at all rift zones		158–recent
		Las Playas collapse San Andres collapse El Julan collapse	176–134 545–176 545–200
El Golfo volcanism	Rift zone development in directions WNW and ENE		545–176
		Tiñor collapse	882–545
Tiñor volcanism	Rift zone development in N–S and NE directions		1120–882
Southern Ridge formation	Rift zone development in direction N–S	?	?

The ages of the rift zones were constrained by stratigraphy; dykes themselves have not been dated. The dyke swarms exposed in the area of the San Andres fault system are from the Tiñor volcanic phase (> 882 ka) as evidenced by the absence of El Golfo lavas in that area and by the minimum age of 176 ka for the aborted San Andres giant landslide. The dykes in the Las Playas embayment have been uncovered by the Las Playas giant landslide and thus have a minimum age of 145 ka. The dyke swarms inside the El Golfo scarp pre-date the El Golfo giant landslide and thus must be older than 13 ka. Includes data from Abdel-Monem, Watkins & Gast (1972); Guillou *et al.* (1996); Masson (1996); Day, Carracedo & Guillou (1997); Carracedo *et al.* (1999); Szérémeta *et al.* (1999); Gee *et al.* (2001).

experiments implying that on El Hierro the units (a) and (b) are capable of organizing the units (c) and (d). Periods of constructive and destructive phases on El Hierro are summarized in Table 1.

The models predict where rift zones ideally develop in time and space, reproducing the natural pattern as observed on El Hierro. Numerous implications arise from the geometric development of rift zones and eruption centres and related flank instability, which are also applicable to a variety of other volcanic edifices subject to gravitational spreading and rifting.

2. The structure of El Hierro

The island of El Hierro, the westernmost and youngest of the Canary Islands, represents the summit of a volcanic shield elevating from the surrounding seafloor at depth of 4000 m to up to 1501 m above sea level (Fig. 1a). The development of El Hierro was governed by its basal architecture, as well as by episodes of intrusive and effusive growth and gravitational collapses.

2.a. Volcano base

The island of El Hierro is situated on 156 Ma old Jurassic crust (Roeser, 1982). Seismic profiles of the surrounding seafloor show subhorizontal layers of various reflectivity above the basaltic crust (Urgeles *et al.* 1998; Collier & Watts, 2001). An about 1.0 km thick unit of low reflectivity was interpreted to represent pre-volcanic sediments (Units I and II, after Collier & Watts, 2001). An approximately 0.7–0.8 km thick layer is overlying, defined by higher reflectivity, which has been interpreted as syn-volcanic deposits dominated by volcaniclastic sediments (Units III, IV and V, after Collier & Watts, 2001). Significant lithospheric flexure due to volcanic loading of El Hierro Island could not be recognized.

2.b. Merging volcanic centres

El Hierro formed by successively growing and overlapping volcanic centres (Carracedo *et al.* 2001). The southern tip of the island extends into a submarine topographic ridge, hereafter referred to as the 'Southern Ridge'. It is about 40 km long and has a curved hinge line. The elevation of the Southern Ridge exceeds 2000 m above the seafloor. A saddle at longitude 27°30′ separates two segments of the Southern Ridge, one oriented NE–SW and one oriented N–S. South of 27°30′ the ridge is characterized by intense erosional morphology and steep flanks of up to > 40°; it was therefore described as a comparatively old structure (Gee *et al.* 2001), confirmed by interpretations of data from a dredging campaign (Schmincke & Graf, 2000) and radiometric dating (P.v.d. Bogaard, unpub. data, 2005). The Southern Ridge merges in the north with two main volcanic centres, which make up most of the subaerial part of the present El Hierro Island. The northeastern volcanic centre is Tiñor volcano, the lavas of which formed the earliest subaerial sequences identified (Carracedo *et al.* 2001). Apart from erosional scars and valleys, Tiñor shows an approximately circular basal outline in the NE (Figs 1b, 2a) despite an inferred rift zone in the northeastern island sector (Carracedo, 1994). Based on the overall morphology and bathymetry, the centre of the Tiñor volcano is located around 17°55′ W, 27°48′ N. The southwestern part of the Tiñor edifice is overlain by the El Golfo volcano, which is the youngest volcanic centre on El Hierro Island (Carracedo *et al.* 2001). The submarine northern and western flanks of the El Golfo volcano also approximate a circular outline with a centre at 18°01′ W, 27°44′ N (Fig. 2a). For both Tiñor and El Golfo volcanoes the bathymetry suggests basal diameters of at least 35 km. The last eruptive stage of El Hierro was characterized by synchronous activity at various locations forming abundant pyroclastic cones

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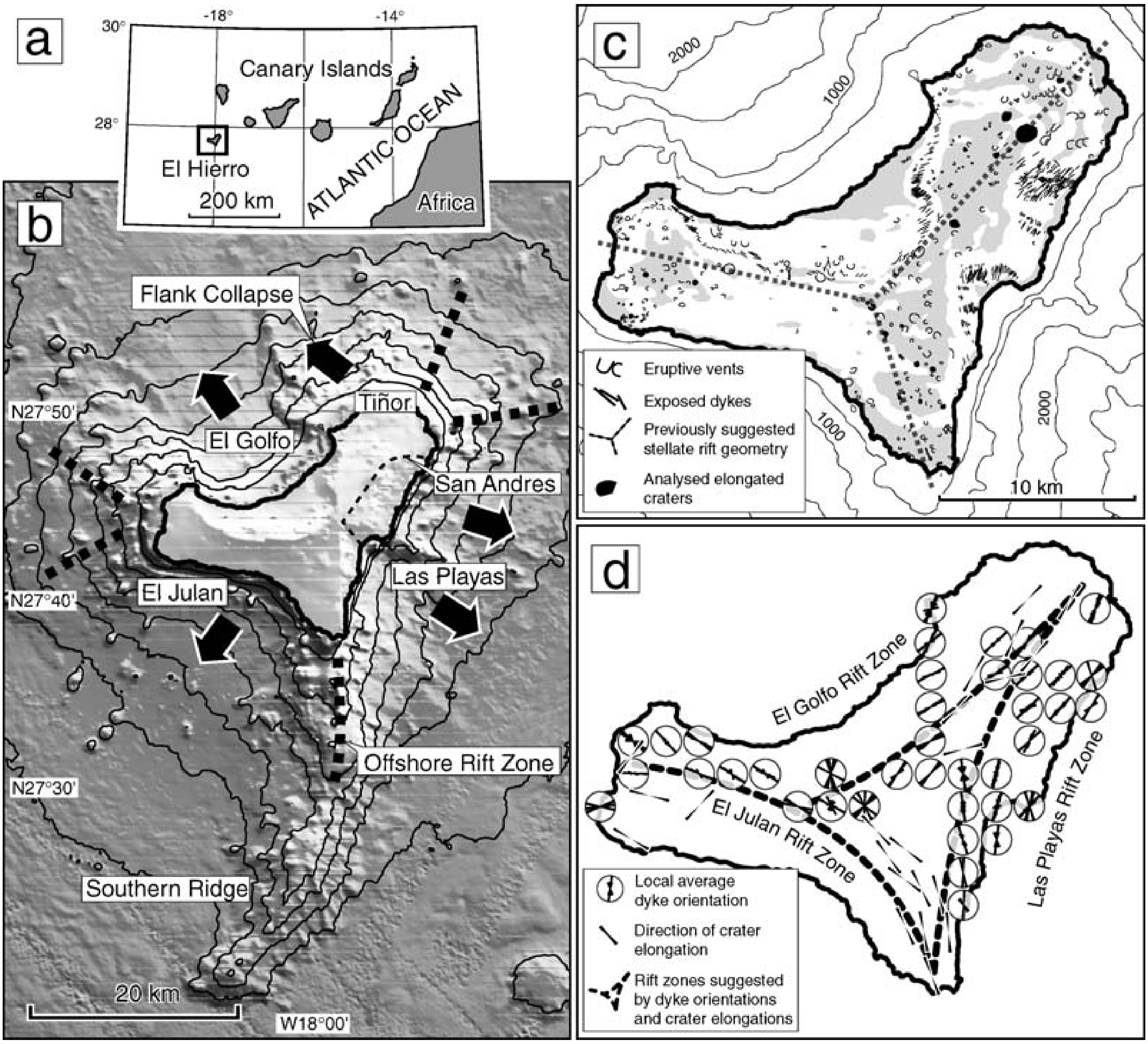

Figure 1. (a) Location of the island of El Hierro in the Canarian Archipelago. (b) Topographic map (bathymetry contours are 500 m) showing the volcano complex of El Hierro composed of three overlapping volcanoes: the Southern Ridge, the Tiñor volcano and the El Golfo volcano. The present island shape is the result of two giant landslides leaving large amphitheatres in the northwest and southwest of the island. These and three further giant landslides are thought to be related to rift zones. The irregular morphology of the western and northeastern flanks indicates diverging offshore rift zones (Gee *et al.* 2001). (c) Pyroclastic cones and dykes exposed on El Hierro used in structural analyses (Carracedo *et al.* 2001). Areas inclined $\leq 10^{\circ}$ are shaded in grey. (d) Local average dyke orientations (rose diagrams) and directions of crater elongations reflect the presence of three conjugate rift zones: the Las Playas rift zone, the El Golfo rift zone and the El Julan rift zone. These rift zones form a rift system in the shape of a hyperbolic triangle. Bathymetric data are based on Krastel *et al.* (2001) and Masson *et al.* (2002).

and lavas that cover most of the island (‘rift volcanism phase’ after Guillou *et al.* 1996; Carracedo *et al.* 2001; Fig. 1c). The evolutionary stages of edifice growth and mergence of volcanic centres were repeatedly interrupted by massive erosional events at unstable flanks.

2.c. Unstable flanks

The three large amphitheatres on El Hierro represent the headwall scarps of giant landslides into the sea (Fig. 1b). The earliest constrained landslide involved a large part of the Tiñor volcano and was directed to the northwest (Tiñor Landslide); it was identified by a geomagnetic discordance between the Tiñor and El Golfo sequences (Guillou *et al.* 1996; Carracedo *et al.* 1999). Another giant landslide at El Julan produced a 15 km wide embayment in the southwest of the island and of the El Golfo volcano. The subaerial headwall scarp is covered by younger lavas, obscuring the exact location and dimension of the collapse structure. The El Julan Landslide was approximately 60–120 km^3 in

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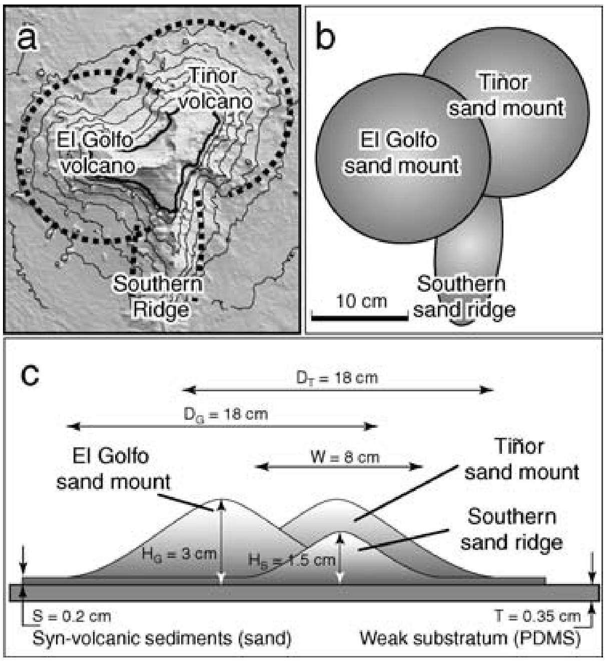

Figure 2. Sketches of the experimental set-up to simulate spreading of overlapping volcanoes over viscous deformable substrata. (a) Boundaries of the major El Hierro volcanoes estimated from the mean maximum length of submarine channels. (b) Plan view sketch of the sand edifices in the experiment, placed and shaped in a simplified geometry of El Hierro (see a). (c) Schematic side view of the sand edifices on viscous substratum. Geometric dimensions of the model are defined by: the diameters of the El Golfo cone (D_G) and the Tiñor cone (D_T), the width of the Southern ridge (W), the heights of the El Golfo cone and Tiñor sand cone ($H_G = H_T$), the height of the Southern ridge (H_S), the thickness of the viscous substratum (T) and the thickness of the syn-volcanic sediments (S).

volume, and is evidenced by debris avalanche deposits covering more than 1600 km^2 of the adjacent seafloor (Holcomb & Searle, 1991; Carracedo, 1996*a*; Gee *et al.* 2001; Krastel *et al.* 2001). The east of the island has been affected by two further collapse events. A 10 km wide embayment marks the headwall scarp of the Las Playas Landslide, which is a SE-directed landslide 25–35 km^3 in volume (Hausen, 1972; Day, Carracedo & Guillou, 1997; Carracedo *et al.* 1999). The submarine scarp forms a distinctive landslide valley down to 2500 m water depth (Krastel *et al.* 2001; Masson *et al.* 2002). To the northeast, a pronounced fault system encloses a flank area of about 50 km^2. This fault system is assumed to represent the headwall structure of the San Andres Landslide, which is an incomplete or aborted slump-like failure of the Tiñor volcano (Day, Carracedo & Guillou, 1997). Having a 15 km width and a scarp as high as 1000 m, the most prominent and most recent amphitheatre formed in the north of El Hierro as a result of the El Golfo giant landslide, depositing 150–180 km^3 of material on the seafloor (Hausen, 1972; Masson, 1996; Urgeles *et al.* 1998). The scarp of this landslide continues to more than 3000 m water depth (Fig. 1b).

2.d. Rift zones

El Hierro was regarded to be a classic example of a triaxial stellate rift system formed by three narrow rift zones. This rift geometry was interpreted from the distribution of eruptive centres, which, in a simplistic view, are clustered along the three island-centered axes (Fig. 1c). The arrangement of the three linear rift zones was supposed to reflect the geometry of stellate 120° least-effort fractures caused by magmatic upward pressure (Luongo *et al.* 1991; Carracedo, 1994, 1996*a*,*b*; Guillou *et al.* 1996; Carracedo *et al.* 2001).

Also, bathymetric studies around the island of El Hierro revealed the submarine existence of rift zones (Gee *et al.* 2001). Towards the south the rift continuation is well defined by the Southern Ridge (Fig. 1b; Gee *et al.* 2001). At the submarine western and northeastern flanks of the island, broad areas of pinnacles and ridges have been detected. These have been interpreted to reflect either seaward bifurcations of the onshore rift zones (Fig. 1b) or a broad zone of rifting (Gee *et al.* 2001). The latter is supported by the vent distribution at the western and northeastern island parts that show a broadening of the vent cluster from the islands crest regions seaward. Carracedo *et al.* (2001) considered gravitational stress in later volcanic stages to have affected the vent clustering.

2.d.1. *Dyke orientations*

Because dykes in homogeneous rocks are typically oriented perpendicular to the least compressive stress axis $\sigma 3$, preferred dyke orientations in rift zones that are made up by hundreds or even thousands of parallel dyke intrusions depend on sustained local or regional stress fields (e.g. Shaw, 1980 and references therein). On El Hierro, direct subaerial hints of preferred dyke paths are given in erosional valleys that uncover some of the older sequences, with dyke orientations that were mapped in detail by Carracedo *et al.* (2001) and reanalysed by us (Fig. 1c,d). We note that measurements were from subaerial dykes only. In order to analyse the dyke measurements spatially, we defined a grid of regular quadrangles of 3 km^2 in area and grouped the data accordingly. Within each quadrangle we plotted the dykes in a stereoplot and calculated the mean of the measured dyke orientations. The results are shown in Figure 1d.

Dykes in the Tiñor volcano are principally oriented NE–SW (N45° ± 15°). Dykes in the vertical cliffs of the Las Playas embayment have an orientation that curves from NNE–SSW (N15° ± 10°) in the northern wall to NNW–SSE (S170° ± 10°) in the southern wall. Dykes in the El Golfo embayment are curved from WNW–ESE (S120° ± 30°) in its western part to ENE–WSW

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(N60° ± 20°) in the eastern El Golfo scarp. The coastal areas in the northeast and southwest of the island reveal only a few dykes.

2.d.2. Pyroclastic cones

Following Tibaldi (1995), we used the direction of crater elongation as a further stress indicator, because craters tend to elongate parallel to the orientation of their supplying eruptive dykes. Assuming that dyke orientations depend on the surrounding stress field (e.g. Shaw, 1980), the elliptic crater of a pyroclastic cone is elongated perpendicular to the least compressive stress axis $\sigma 3$. However, the stress field reflected by the crater elongation at pyroclastic cones is indicative only if the cone is located on subhorizontal slopes ($\leq 10°$: Tibaldi, 1995). Therefore, we consider only those pyroclastic cones that are located on slopes shallower than 10°. The crater elongation in the northeastern part of the island is generally NE–SW. Crater elongation are NNW–SSE in the most southern area and NW–SE in the northeastern El Julan area. Further to the west only a few craters are located on slopes $\leq 10°$, and show directions varying from NE–SW to WNW–ESE (Fig. 1d).

Combining geomorphologic and structural datasets, that is, the crater elongation, the mean dyke direction and the bathymetry, we can distinguish between (1) a rift zone along the Southern Ridge that is oriented N–S, (2) a rift zone joining Tiñor volcano and the Southern Ridge that is oriented NE–SW in the north and N–S in the south, (3) a rift zone joining El Golfo volcano and Tiñor volcano that is oriented WSW–ENE in the eastern edifice and NE–SW towards the Tiñor volcano and (4) a rift zone joining El Golfo volcano and the Southern Ridge. According to their location, we hereafter refer to those rift zones as (1) the Southern Rift Zone, (2) the Las Playas Rift Zone, (3) the El Golfo Rift Zone and (4) the El Julan Rift Zone, respectively (Fig. 1d). These rift zones are gently curved rather than linear and thus define a conjugate rift system in a shape similar to a hyperbolic triangle (Fig. 1d). In the following section we study this geometry in analogue experiments and demonstrate the interconnectivity between volcano basement, volcano coalescence, flank instability and rift zones.

3. Understanding rift zones and unstable flanks through gravitational spreading experiments

Using analogue experiments scaled to El Hierro, we investigate how the process of gravitational spreading of adjacent volcanoes is associated with the development of rift zones and flank instability.

Previous experiments simulating volcano deformation due to gravitational spreading demonstrated that (a) extensional faults have approximately radial orientation if the volcano is ideally cone-shaped (Merle & Borgia, 1996; Borgia, Delaney & Denlinger, 2000) and (b) extensional faults group into a preferred direction if two volcanoes overlap (Walter, 2003; Walter, Klügel & Münn, 2006). Because dyke intrusions are oriented perpendicular to the direction of least principal stress, we assume that rift zones will form along a system of extensional faults. We utilize the experimental method by Merle & Borgia (1996), Borgia, Delaney & Denlinger (2000), Walter (2003) and Cecchi, van Wyk de Vries & Lavest (2005) in order to understand volcano spreading and rifting on El Hierro. The models imitated the principal conditions of the volcano complex including a pre-volcanic substrata and the incremental loading by overlapping volcanic centres.

3.a. Experimental set-up and scaling

The process of gravitational volcanic spreading is accomplished through loading on a deformable pre-volcanic substratum. We simulated this substratum with a layer of viscous silicone material (Polydimethylsiloxane, PDMS), on top of which we incrementally constructed sand mounts according to a simplified geometry of El Hierro (Fig. 2). We first constructed the Southern Ridge and the Tiñor volcano, which are the oldest edifices of El Hierro (Carracedo *et al.* 2001; Gee *et al.* 2001). After seven hours of spreading, we constructed the El Golfo volcano that partly overlapped the older volcanoes. At this time, the spreading process of the Tiñor volcano and the Southern Ridge was not completed yet, initiating simultaneous spreading of all the three sand edifices. Around the edifices a layer of sand imitated syn-volcanic sediments. For a better visualization of deformation structures the sand volcanoes were powdered with a very thin layer of flour. The experiments were analysed in detail after 34 hours, when no further deformations were observable. Duration of spreading should ideally mimic the real chronological relations known from El Hierro Island, but was constrained by the viscosity of the silicone material in our experimental setup (see below and Tables 2, 3).

In order to compare the experimental results with data from El Hierro, we scaled the geometric and dynamic parameters of El Hierro into laboratory dimensions (Hubbert, 1937; Ramberg, 1981). The scaling procedure was adopted from Merle & Borgia (1996) and Borgia, Delaney & Denlinger (2000) and is summarized in Table 2. The dimensional analysis was made according to the Buckingham Π−theorem (Buckingham, 1914), which is generally used for comparisons of physically different systems with equivalent terms. We set up seven dimensionless Π-numbers for calculating adequate values for the material properties and geometric dimensions of the model. The geometric parameters of the model were scaled to nature with a ratio of $2.7–5.5 \times 10^{-6}$, that is, 1 cm in the model represented 1.8–3.7 km in nature.

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Table 2. Geometric and dynamic variables of the model scaled to El Hierro

Parameter	Variable	Model	Nature	Dimension	Model/nature
Height of volcano	H	3.0E−02	5.5E+03	m	5.5E−06
Diameter of volcano	D	1.8E−01	3.6E+04	m	5.0E−06
Thickness of viscous substratum	T	3.5E−03	1.0E+03	m	3.5E−06
Thickness of syn-volcanic sediments	S	2.0E−03	7.5E+02	m	2.7E−06
Density of volcano	ρv	1.4E+03	2.7E+03	kg/m^3	5.2E−01
Density of viscous substratum	ρs	1.1E+03	2.2E+03	kg/m^3	5.2E−01
Viscosity of viscous substratum	μ	1.9E+04	1.0E+18	Pa ∗ s	1.9E−14
Angle of internal friction	ϕ	3.0E+01	3.0E+01		1.0E+00
Gravity	g	9.8E+00	9.8E+00	m/s^2	1.0E+00
Time span of deformation	t	5.4E+04	7.9E+12	s	6.8E−09

Geometric variables of the southern sand ridge (see Fig. 2c) are not indicated because have a similar geometric ratio as the sand cones and thus give similar Π−numbers.

Table 3. Dimensionless analysis of geometric and dynamic parameters

Π	Equation		Model	Nature	Model/nature
Π1 =	Height of volcano/diameter of volcano	H/D	0.17	0.15	1.1
Π2 =	Thickness of syn-volc. sed./height of volcano	S/H	0.067	0.14	0.49
Π3 =	Thickness of visc. substr./thickness of syn-volc. sed.	T/S	1.8	1.3	1.4
Π4 =	Density of volcano/density of visc. substr.	$\rho v/\rho s$	1.2	1.2	1.0
Π5 =	Gravitational force/viscous force	$\rho v * g * H * t / \mu$	1.2E+03	1.1E+03	1.0
Π6 =	Failure resistance force/viscous force	$\tan\phi *$ (Π5∗ 2/3 ∗ (1+Π2) +1)	0.048	0.05	0.96
Π7 =	Inertial force/viscous force	$\rho s * T^2 / \mu / t$	1.4E-11	2.8E-22	4.9E+10

Dry fine sand deforms according the Navier-Coulomb criterion and is a well-known analogy to brittle rock. The low viscosity μ of 1.9×10^4 Pa s of the PDMS substratum demanded a time span t of about 15 hours in which most of the deformation in the experiments occurred. This time span represents 250 000 years for spreading in nature. If scaling is correct, the Π-ratios ($= \Pi_{model}/\Pi_{nature}$) for geometric proportions (Π1, Π2 and Π3), density proportions (Π4) and dynamic proportions (Π5 and Π6) should be near unity. As indicated in Table 3, Π1–Π6 approach 1, showing that the models are well scaled. The low values of the Reynolds numbers Π7 reflect the dominance of viscous forces over inertial forces, where the relatively low viscosity of the silicone material causes the deviation between model and nature. However, this is an inevitable but accepted limitation because it enables the execution of such experiments in effective time frames (see Merle & Borgia, 1996 for more details). We did not simulate forceful intrusions, thus regarding the orientation of fracture systems as being controlled by the gravitational stress field only.

3.b. Experimental results

All results were similarly reproduced in repeated experimental runs. Gravitational spreading of the overlapping Southern Ridge and Tiñor volcano caused several prominent extensional structures. Spreading started within a few minutes and after seven hours, distinct zones of extension were visible, which were different in the south and north (Fig. 3b). In the south, a major graben developed along the ridge, with maximum extension just between the edifices. The zone of maximum extension is a consequence of the mutual buttressing effect of both sand mounts and indicates that a rift zone would preferably form along a NNE–SSW direction. On the Tiñor sand mount, in contrast, tensile fractures formed in a radial pattern. These radial fractures are best expressed on the northern half of Tiñor and become smaller towards the major graben faults in the south–southwest. After seven hours of spreading we constructed a new cone (El Golfo sand mount) overlapping the western Tiñor sand mount and to a lesser degree the Southern Ridge (Fig. 3c). We allowed spreading of the coalesced edifice for a further 27 hours, during which earlier fractures were partly reactivated and new fracture zones developed (Fig. 3d). Continuing deformation along the former major graben faults was observed, and a new graben developed between the Tiñor cone and the El Golfo cone in a WSW–ENE direction. A third graben, much less well expressed, developed between El Golfo and the Southern Ridge. On the western flanks of El Golfo, radial fractures formed.

The final pattern is a system of conjugate axial graben structures connecting the topographic maxima of the sand edifices and radial fractures on the unbuttressed WNW and NE flanks of El Golfo and Tiñor volcano, respectively. The flank segregation by these faults thus indicated radial spreading of the unbuttressed flanks. Close-up views show that many graben faults have slight strike-slip components, especially the main radial graben on the free sides of

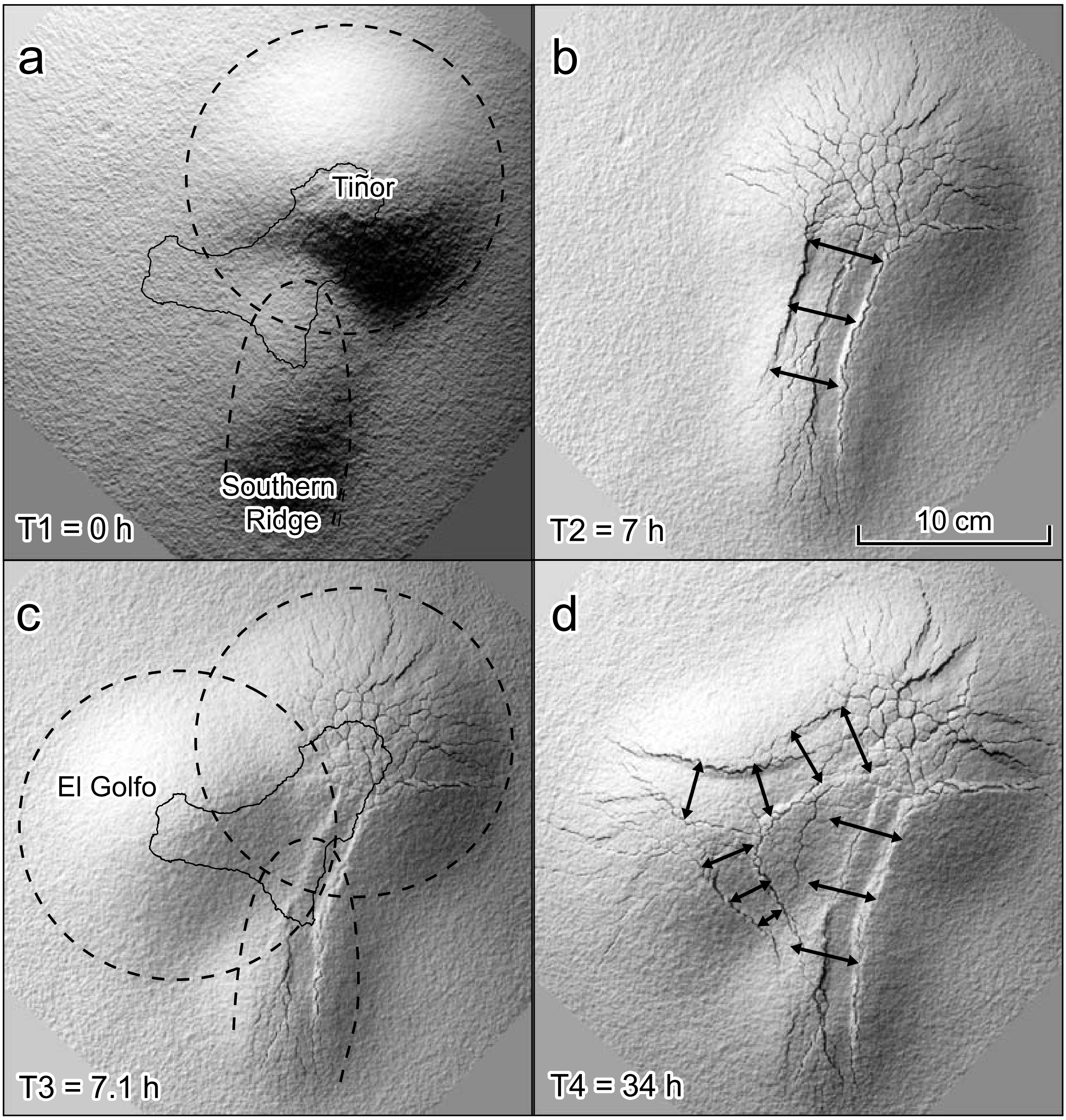

Figure 3. Plan view photographs of a representative experiment. (a) The overlapping 'Tiñor cone' and the 'Southern Ridge' were emplaced simultaneously. (b) After 7 hours of simultaneous spreading, a major graben formed between the edifices. (c) After 7.1 hours, the 'El Golfo cone' was added overlapping the 'Tiñor cone' and the ridge. (d) The two cones and the ridge have spread for 27 hours showing a triangular system of conjugate graben axes.

the El Golfo and Tiñor cones (Fig. 3d). Large sectors of the edifice are enclosed by (1) the major graben (potential rift zone) in the central part, and (2) radial transtensional graben on both flanks. No prominent fractures formed within the large sectors, implying that no major internal deformation occurred and that displacement was coherent in each sector. Moreover, from the absence of apparent deformations we infer that the flank steepness of these large sectors does not decrease during spreading. This contrasts to the volume-preserving lengthening and flattening of the edifice at the radially spreading edifice sectors.

4. Discussion

The orientation of exposed dykes and of crater elongations on El Hierro reflects a structural architecture in the form of three main rift zones. We found that the rift zones are gently curved, and conjugate near the centres of the Tiñor volcano and the El Golfo volcano rather

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than in the centre of the island. Submarine extensions of these rift zones diverge and form radial alignments of eruptive cones and ridges (Gee *et al.* 2001). Applying the gravitational spreading hypothesis (e.g. Borgia, Delaney & Denlinger, 2000), we better understand the sequential development of the El Hierro volcanic centres, dyke intrusions and rift zone geometry, as well as flank instability, which is a fundamental improvement to the previously supposed hypothesis of triaxial linear rift zones that formed by updoming (e.g. Carracedo, 1996*a*; Guillou *et al.* 1996; Day, Carracedo & Guillou, 1997). The importance of gravitational spreading is supported by analogue experiments, where we simulated deformation of overlapping sand edifices. In experiments, three rift axes formed in between the 'volcanic' edifices, resembling the real situation on El Hierro in stunning detail.

4.a. A structural comparison of the sand edifices to El Hierro

The grabens in the experiments display zones of maximum extension. In nature, such extensional zones are likely to develop into rift zones. Based on this perception, the natural and the experimental structures compare well (Fig. 4); in the experiments, the first rift zone formed between the Southern Ridge and the Tiñor sand cone. Likewise, the oldest dykes on El Hierro exposed at the Las Playas Rift Zone point towards the Southern Ridge and towards Tiñor volcano. Upon construction of the El Golfo sand cone, new rift zones developed in the experiments, one in the SW–NE direction and a less-expressed one in the NW–SE direction. Moreover, spreading of the older rift between the Southern Ridge and the Tiñor sand mount continued. On El Hierro Island, indeed two rift zones developed on El Golfo volcano extending from its centre towards the NE (El Golfo Rift Zone) and towards the NW (El Julan Rift Zone). Parallel dyke orientations of the Tiñor and El Golfo sequences in the Las Playas embayment suggest continued activity of the Las Playas Rift Zone during El Golfo volcanism, similar to what was observed in the experiments.

The experimental rift zones diverge radially at the outer edifice flanks (Fig. 4a), similar to the observed alignment of submarine cones and ridges off El Hierro (Gee *et al.* 2001; Fig. 4b). The overall similarity of the rift zones in the experiments and in nature suggests that gravitational forces (spreading) rather than endogenous forces (updoming) played a major role in the structural development of El Hierro.

4.b. Flank instability

Volcano flank instability and the susceptibility to form giant landslides are controlled by the steepness of the flank and forceful intrusions that push the flanks

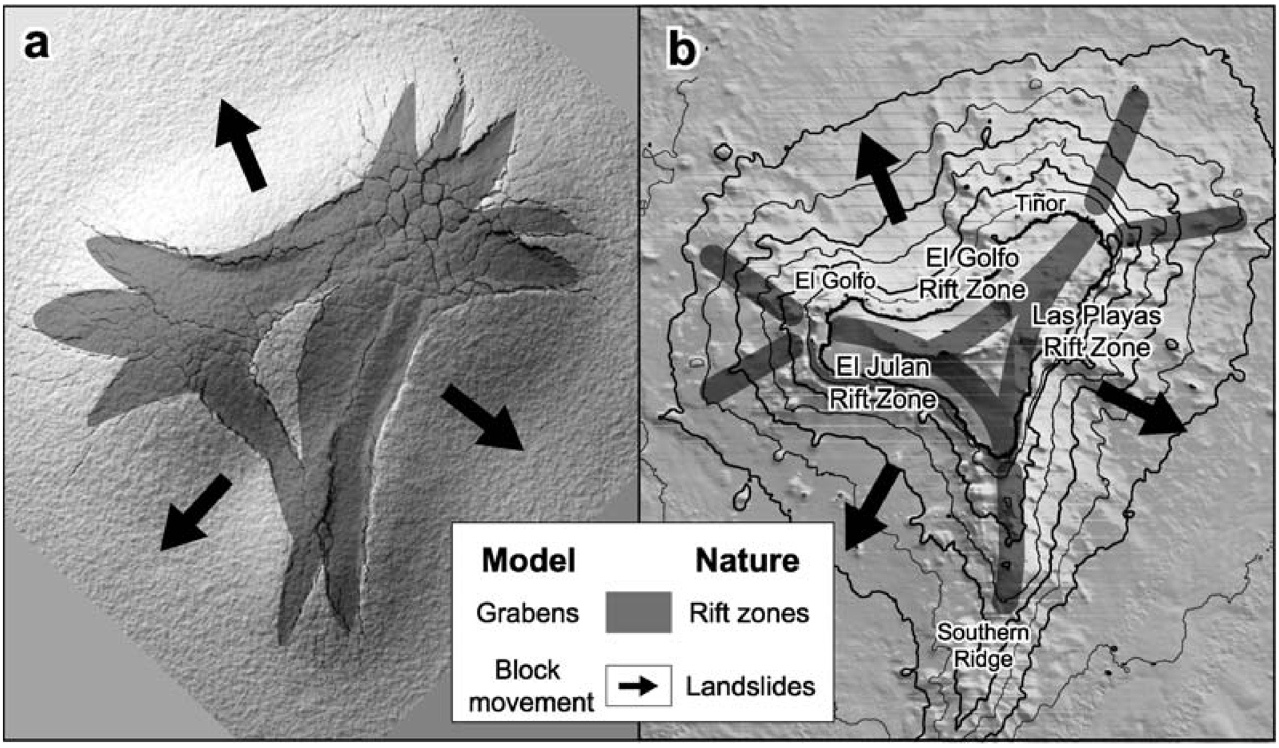

Figure 4. Graphical comparison of the deformation structures formed in the experiment (a) and the rift architecture and giant landslide structures on El Hierro (b). Note the geometric congruence of (1) the rift system on El Hierro with the graben system formed in the experiment and (2) the locations and directions of the block movement in the experiment with locations and directions of giant landslides on El Hierro.

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outward (Swanson, Duffield & Fiske, 1976), as well as other often related factors such as pore pressure changes or hydrothermal weakening (Elsworth & Voight, 1995; Day, 1996). Our work also provides constraints on the development of coherent intrusive complexes and the slope gradient. The hypothesis of forceful intrusions and flank push (Swanson, Duffield & Fiske, 1976) necessitates the formation of rift zones in order to develop unstable sectors. Radial spreading is associated with two further features evident in both the experiments and the free volcano flanks of Tiñor and El Golfo volcanoes, i.e. (1) radial extension at these flanks favours radial diverging distribution of dykes and eruption centres, and (2) lateral spreading decreases slope gradients. Radial lateral extension and the development of radial dykes imply that no pronounced dyke-related displacement of large sectors developed. Therefore, the free western flank of El Golfo and the free northeastern flank of Tiñor volcanoes were more stable than other parts of the coalesced edifice. Indeed, the corresponding western and northeastern flanks on El Hierro show no evidence for giant landslides.

Steep volcano flanks are unstable and susceptible to giant landslides, especially when they are enclosed by pronounced rift zones (e.g. Carracedo, 1999; Morgan & McGovern, 2005). Also, on El Hierro the flanks enclosed by rift zones were the most unstable, as shown by localities of the giant landslides that had occurred (Carracedo, 1994, 1999; see Fig. 1). This agrees with our experimental observations that the corresponding flanks apparently did not deform internally and thus gravitational stress at these flanks was not relieved. Other experimental studies demonstrated that at constant normal stress the threshold for brittle failure can be reached simply by strain accumulation within slow creeping flanks (Petley & Allinson, 1997).

4.c. Cone distribution and dyke orientation on El Hierro

The inferred directions of the subaerial rift zones of El Hierro are based on the distribution of eruptive centres, their crater elongations, and measured dyke orientations (Carracedo *et al.* 2001). Outcropping dykes, however, do not equally represent all stratigraphic units, and large parts of the island lack such outcrops altogether. Dykes are visible mainly in deeply eroded valleys and landslide headwalls in the north and the east of the island, as well as around the coastline. Where dykes intrude near a collapsed sector, their orientation is influenced by the near-surface topography and parallels the headwall scarp enclosing the collapsed sector. Such circumscribing dykes have been reported from various volcanoes, such as Mauna Loa (Lipman, 1980), Mount Etna (McGuire & Pullen, 1989), Tenerife (Walter & Schmincke, 2002), and Stromboli Island (Tibaldi, 2003). The rift zones at El Hierro may have been similarly influenced. However, we conjecture that their curved trend reflects a long-term pattern rather than one influenced by a single event for several reasons. Dykes of the Las Playas Rift Zone (Tiñor age) and the much younger dykes emplaced during the Rift Volcanism have similar orientations. In the north, a large number of dykes crop out around the El Golfo embayment, which is about 15 ka or even younger (Masson, 1996). Because the dykes crop out in the near-vertical cliffs without the presence of eruptive centres, most of them must have been emplaced before the 15 ka giant landslide event. This implies that also the El Golfo Rift Zone was curved and volcanically active before the northern El Golfo giant landslide occurred.

As there are no dyke outcrops around the El Julan embayment, the inferred direction of the El Julan Rift Zone is based on the distribution of pyroclastic cones (Carracedo *et al.* 2001), their crater elongations and the comparison to our experimental findings.

In experiments we moreover observed that the three main rift axes enclose an area in the centre of the edifice that is relatively undeformed. On El Hierro, however, this area shows pronounced fracturing and eruptive centres. This contradiction may be due to our experimental setup construction, where the geometric centre of the sand edifice is enclosed by topographic highs and thus buttressed on all sides. In nature, in contrast, such a zone is likely to be infilled by volcanic and erosional activity and thus subjected to a different state of stress.

4.d. Basal décollement or triple junction under El Hierro?

Our model is based on the assumption that the island of El Hierro is decoupled from the oceanic basement by viscous deformable substrata, as has also been suggested for the island of Hawaii (e.g. Nakamura, 1980). Experimental studies (e.g. Merle & Borgia, 1996; Cecchi, van Wyk de Vries & Lavest, 2005) as well as seafloor observations (e.g. Morgan *et al.* 2000) show that volcano spreading over a near-horizontal décollement may produce thrust faults and a bulge in the volcano-surrounding sediments. On El Hierro, submarine data do not show such a frontal bulge or fracture zone. However, van Wyk de Vries *et al.* (2003) showed that for spreading volcanoes that have very gentle slopes without a slope break, no such frontal bulge develops. Krastel *et al.* (2001) discussed the idea that the steep slopes of the Canary Islands are a result of higher differentiated magmas, but this cannot be the case for El Hierro, lacking differentiated rocks.

By comparison to Hawaii, Morgan & McGovern (2005) suggested that the steep slopes and shallow landslide scars on the Canary Islands reflect only a minor role of a weak basal décollement. Our study, in contrast, qualitatively supports the spreading hypothesis for El Hierro Island, well explaining the sites of flank instability and the subaerial and submarine rift zone geometry. The depth and dip of such a décollement, however, remains unclear. A detailed quantitative study may solve the key questions of the décollement location, dip and amount of displacement during spreading.

The idea that gravitational spreading is a dominant cause of rift zone development and triaxial rift arrangement on El Hierro is new. An additional influence of lithospheric flexural stress caused by intrusive and volcanic loading cannot be ruled out, however. For instance, the arrangement of adjacent volcanoes may be influenced by island loading when a plate moves over a hotspot (Carracedo *et al.* 1998; Hieronymus & Bercovici, 1999, 2001). Intrusive forces updoming the crust (similar to large-scale triple junctions) would imply that the locations of rift zones are defined in very early stages of volcano growth and reactivated thereafter (Carracedo, 1996*b*). Our model, in contrast, allows rift zones and sites of flank instability to rearrange with changing gravity load of a growing volcanic edifice.

4.e. Lessons from other Canary Islands

Pronounced rift zones can be found on five out of seven of the Canary Islands. The coalescence of distinct volcanic edifices may have had an important influence on the rift zone development. Tenerife, the largest of the islands, is made up of four volcanoes (Roque del Conde, Teno, Anaga and Cañadas; Ancochea *et al.* 1990). Gravitational spreading of the partly overlapping edifices of Tenerife caused buttressing and partial spreading, developing into zones of extension and the formation of rift zones (Walter, 2003). La Palma Island was formed by an old volcanic centre in the north (Garafia and Taburiente), and a rift zone to the south (Cumbre Nueva, Cumbre Vieja; Carracedo *et al.* 2001). It was proposed that the southern rift zones of La Palma change and migrate as a consequence of flank instability and flank movement (Day *et al.* 1999; Walter & Troll, 2003). In addition, the formation of the N–S La Palma rift zone might have been facilitated by the presence and buttressing effect of an old (pre-Miocene) submarine volcano (Bogaard, pers. comm. 2001) south of the island (Klügel & Walter, 2003). These studies suggest that ocean island rift zones depend upon the growth history of merging volcanic centres, and further evidence was added by this study.

5. Conclusions

The structural evolution of El Hierro has been strongly controlled by the formation of volcanic rift zones. The distribution of pyroclastic cones, the orientation of dykes on the island, and the flank morphology offshore reflect a rift system with a triangular geometry in its central part and diverging rift-zone continuations at the northeastern and western submarine flanks. This study suggests that the rift configuration is controlled by gravitational spreading of the consecutively overlapping volcanic edifices. In analogue sand box experiments we reproduced the formation of both the subaerial and submarine rift zone geometries. Furthermore, the experiments demonstrated that during the spreading process the free unbuttressed flanks of El Hierro have been stabilized by radial spreading, whereas the flanks enclosed by rift zones maintained their potential to collapse in the form of giant landslides.

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