Setting the scene for self-destruction: From sheet intrusions to the structural evolution of rifted stratovolcanoes

Alessandro Tibaldi  
*Dipartimento di Scienze Geologiche e Geotecnologie, Università di Milano-Bicocca, Piazza della Scienza 4, Milan 20126, Italy*

Derek Rust  
School of Earth and Environmental Sciences, University of Portsmouth, Burnaby Building, Burnaby Road, Portsmouth PO1 3QL, UK

Claudia Corazzato*  
*Dipartimento di Scienze Geologiche e Geotecnologie, Università di Milano-Bicocca, Piazza della Scienza 4, Milan 20126, Italy*

Andrea Merri  
*Dipartimento di Scienze della Terra “A. Desio” Università di Milano, Via Mangiagalli 34, Milan 20133, Italy*

ABSTRACT

We investigate the processes of growth and consumption of rifted stratovolcanoes to better understand their morphostructural and geological evolution in relation to sheet intrusions. Field data collected from four rifted volcanoes, located in Italy and Chile, and representing structural end members, are integrated with scaled physical models and numerical three-dimensional (3D) modeling. Due to preferential along-rift magma pathways, volcanoes grow perpendicular to rift strike mostly by intrusions, and parallel to the rift mostly by effusions. These constructive processes are partially or fully balanced by destructive processes that occur both perpendicular to the rift and along the rift zone, and that are initiated by the sheet intrusions. Field data indicate that sheets and/or flank eruptions increase in number approaching the summit part of a cone. Since dyking within a volcano causes a magmatic driving overpressure and lateral displacement/deformation of the cone flank, this displacement is assumed to be greater where intrusions have the highest frequency. Lateral deformation can also lead to flank failure along a deep-seated slip surface. Symmetrical rifts may produce deep-seated collapses in opposing directions, while asymmetrical rifts can preferentially lead to destruction on the tectonically downthrown side. If the collapse depression is refilled by new volcanism, the cone attains a new critical height (and mass), setting the scene for renewed failure. This sequence inhibits the cone from widening normal to the rift zone. The field examples studied, plus analog and numerical modeling, also indicate that dyking within the rift zone destabilizes the volcano flanks along the rift: where dyke tip stresses approach the slope, surface landsliding might be induced. This in turn produces debuttressing above the dyke, favoring effusive eruptions and lavas flowing within the landslide depression. In this case a balance can occur between new lava outpourings and gravity mass removal, or the volcano can grow along the rift if the emitted magma volume is higher than the collapsed material. From a mechanical point of view, the different behavior of volcanic slopes, both parallel to and perpendicular to dyke strike, is a function of magma forces and the cumulative dyke-induced displacement. The combination of along-rift-strike landsliding and normal-to-rift deep-seated large-scale lateral collapse, characteristically produces in map view four main zones of "consumption" and "rebuilding" on the volcano. These alternate with four triangular sectors where the volcanic slopes are relatively more stable. When volcanic activity ends, cone consumption is concentrated along the rift and is favored by fault slip or channeled erosion.

1. INTRODUCTION

Several polygenetic volcanoes, both island and continental, are affected by volcanotectonic rifts. These rifts are linear or slightly curvilinear features characterized by swarms of parallel dry (noneruptive) dilational fractures and eruptive fissures. Dykes or other sheetlike intrusions are the subsurface equivalents of eruptive fissures, and swarms of these minor intrusions may contain tens to hundreds of individual dykes (Fiske and Jackson, 1972; Walker, 1999). In some cases, rift fissures can evolve into normal faults with two main arrangements: (1) a symmetrical rift with parallel inward-facing fault scarps (i.e., with fault planes dipping toward each other), as in the case of N-S Rift on Mount Etna, Italy (Groppelli and Tibaldi, 1999), or (2) an asymmetric rift with parallel normal faults dipping in the same direction, as in the case of the Ollagüe volcano on the Chile-Bolivia border (Tibaldi et al., 2006).

Nakamura (1977) and Nakamura et al. (1977) showed that in many arc volcanoes both island elongation and rifts are parallel to the trajectory of motion of the plate on which they are situated. Consequently, rift orientation is tectonically controlled. Other detailed studies of volcanoes in the Azores and Samoa also show the positions and orientations of rifts to be mainly tectonic in origin (Walker, 1999). Small sea-mounts aligned with basement faults also imply some influence of tectonic features on eruption geometry (Batiza and Vanko, 1983; Fornari et al., 1987). Fiske and Jackson (1972), on the other hand, consider that rift zones in large volcanoes do not generally coincide with basement structures, highlighting instead the importance of the volcano load and stress distribution for...
rift formation. Where a weak substratum is present, gravity or volcanic spreading are other processes that have been proposed to control rift development and structural evolution of many large volcanoes (e.g., Borgia et al., 2000; Walter, 2003).

Whatever the origin of rifts, they are fundamental features since they are the locus of eruptions and thus localize volcanic hazards. Moreover, they may influence, or interfere with, groundwater circulation; thus acting as heat exchangers concentrating geothermal activity and offering economic potential. They are also generally regarded as controlling the geometry of sector collapses to a direction perpendicular to the rift trend, as shown by numerical modeling and field studies (Moore and Albee, 1981; Lipman et al., 1981; Voight et al., 1981; Siebert et al., 1987; Voight and Elsworth, 1997; Voight, 2000; Donnadieu et al., 2001; Tibaldi, 1996, 2001; Apuani and Corazzato, 2009). As a result, rifts and sector collapses together help to control the form and structure of a volcano. The relationship between rifts and lateral instability of volcanoes has also been widely studied by analog modeling, as well as the propagation of normal faults from the edifice substrate upwards into the cone (van Wyk de Vries and Merle, 1996; Vidal and Merle, 2000; Merle et al., 2001; Acocella, 2005; Tibaldi et al., 2006).

In spite of several papers dealing with the origin of volcanic rifts, or with the relationship between rifts and sector collapse, more data are needed in order to understand the morphostructural evolution of volcanic rift zones and rifted volcanoes, particularly in relation to along-rift sheet intrusions. A specific work was done by Fialko and Rubin (1999) in order to explain the along-rift low profiles of basaltic shield volcanoes such as the Hawaii. Unresolved questions include the following: If magma is channeled along rifts, why are not all rifted volcanoes elongated? What are the differences in evolution of a volcano crossed by an asymmetric rift versus a symmetric rift? Does flank instability take place along the direction of a rift? What happens to a volcano when volcanic or tectonic activity along the rift ends? This paper will try to answer these questions and to contribute to understanding the morphostructural evolution of rifted volcanoes, with special emphasis on the relationship with sheet intrusions within the rift zones of stratovolcanoes. This structural setting will be analyzed in relation to detailed field data from four selected volcanoes (Stromboli in Italy, and Ollagüe, Peineta, and Inacaliri in Chile) that highlight different stages of morphostructural evolution and examples of asymmetric and symmetrical rifts. Field data will be integrated with analog and numerical 3D modeling designed to simulate these phenomena, and will be presented first to analyze the growth of volcanoes, and then their consumption. Our work essentially focuses on polygenetic stratovolcanoes affected by a single rift, and thus does not refer to radial dykes clustered in triple-armed rift zones (Marinoni and Gudmundsson, 2000; Acosta et al., 2003; Walter, 2003).

2. GROWTH OF RIFTED VOLCANOES

An example of a very active stratovolcano crossed by a rift is Stromboli, located in the Aeolian Arc, southern Tyrrenhenian Sea, Italy (Fig. 1). This zone is characterized by the presence of Late Quaternary NE-striking normal faults in the eastern part of the arc, and transpressional structures in the central and western part (Neri et al., 2003; Goes et al., 2004; Argnani et al., 2007). The least principal stress (σ3) trends NW-SE in the eastern part of the arc, with a vertical maximum principal stress (σ1), whereas in the central and western part σ1 is horizontal and trends from NNE-SSW to NNW-SSE. Stromboli is 2.9 km high and is composed dominantly of lavas and subordinate pyroclastic deposits. Composition spans from basaltic andesite to shoshonite and latite-trachyte (Hornig-Kjarsgaard et al., 1993). The oldest dated rocks have an age of ~100 ka (Gillot and Keller, 1993), and since that time the feeding system for the various growth phases has been strongly dominated by sheet intrusions and fracturing along a NE-trending axial zone of weakness crossing the entire volcano summit (Rift zone in Fig. 2; Tibaldi, 1996, 2001). Within this rift zone some intrusive sheets fed eruptions, as evidenced by their transition to scoria deposits (Corazzato et al., 2008; Tibaldi et al., 2009a). From a total of 109 sheets in outcrop, only a couple of subcylindrical conduits have been found (Mallo Passo and Serro Monaco necks, Fig. 1), suggesting that the inner growth of the cone occurred essentially along the rift zone in the form of planar intrusions. This NE-SW rift zone is also consistent with regional tectonics of the eastern Aeolian Arc, as portrayed in

Figure 1. Shaded relief bathymetry of the southeastern Tyrrenhenian Sea showing the island volcanoes of the Aeolian Arc (Italy), the studied Stromboli Island volcano (box), the main late Quaternary faults, and the late Quaternary state of stress (compression: converging arrows, extension: diverging arrows). A—Alicudi; F—Flicudi; S—Salina; L—Lipari; V—Vulcano; P—Panarea; St—Stromboli (bathymetry from Di Roberto et al., 2008; structures from Neri et al., 2003; Goes et al., 2004; Argnani et al., 2007).
Figure 1, where $\sigma$, has a NW-SE trend. Another zone of sheeting developed after the occurrence, 13 ka ago, of a large sector collapse toward the NW (Fig. 2; Tibaldi, 2001). Since that time sheets have injected along the NE rift, and also parallel to the amphitheater walls of the collapse (Tibaldi, 2003). There are no data, however, to assess whether these circum-collapse sheets have fed eruptions. Instead, several prehistoric eruptions have taken place at the summit crater and at NE-striking eruptive fissures located along the NE part of the rift. Historic eruptions have occurred at the summit crater, as well as at NE-striking fissures located close to this crater (Casagli et al., 2009). One historic eruption occurred 2 ka ago (Arrighi et al., 2004; Speranza et al., 2004; 2008) along the NE part of the rift. The summit crater is usually composed of 3–5 vents, mostly aligned NE-SW. The present activity is also fed by a dyke striking NE-SW (Chouet et al., 2003; Mattia et al., 2008). Most lavas emitted by these features flowed toward the NW, NE, and SE, in decreasing order of frequency (Tibaldi, 2010).

An example of a Quaternary stratovolcano (presently in a fumarolic stage) crossed by an asymmetric rift is Ollagüe, on the Chile-Bolivia border (Fig. 3). This part of the volcanic arc is characterized by the presence of Quaternary NW-striking normal faults and eruptive fissures (red lines in Fig. 3), consistent with a horizontal $\sigma_3$ trending NE-SW and a vertical $\sigma_1$ (Tibaldi et al., 2009b). Other volcano-tectonic features that strongly indicate a geometric control by NW-striking normal faults include the large Pastos Grandes caldera (white lines in Fig. 3). This depression, first classified as a caldera by Baker (1981), de Silva (1989), and de Silva and Francis (1991), shows a very strong asymmetry, with an intact and continuous northeastern rim that is characterized by a series of SW-dipping, rectilinear slip planes ~30 km long. By contrast the other parts of the caldera wall indicate no clear structural control. The main growth phases of Ollagüe, which is 2.2 km high above the substrate, are represented by Pleistocene andesitic-dacitic lavas and subordinate pyroclastic deposits (Feeley et al., 1993; Feeley and Davidson, 1994; Feeley and Sharp, 1995; Wörner et al., 1994). In the last 100 ka, activity has been essentially in the form of a summit dome that presently emits a steam jet. Pleistocene lavas are emplaced in the form of flows and domes, and the related vents are located in the summit part of the cone as well as along its NW and SE flanks. Moreover, these vents are generally concentrated along a NW-SE zone; while the coeval vents are also aligned NW-SE and most of the single domes are NW-SE–elongated (Fig. 4). Several NW-SE–striking fractures and faults are also present along this zone. All these data indicate the presence of a preferred zone of magma ascent and fracturing crossing the entire volcano, suggesting the presence of a volcanic rift. A NW-SE rift zone is also consistent with the regional tectonic data portrayed in Figure 3. NW-SE–striking sheets have been recognized in the summit part of the volcano, whereas at lower altitudes their possible presence is obscured by extensive alluvial deposits. The main difference in comparison to Stromboli is that along the Ollagüe rift extensional tectonic motion has taken place along SW-dipping normal faults (Tibaldi et al., 2006), producing an asymmetric rift. Younger lavas flowed toward the SW, NW, and SE in decreasing order of frequency.

In other examples of stratovolcanoes, their location and growth is also controlled by linear extensional tectonic structures in the substrate but, after the effusive activity ended, basement normal faulting further propagated across the cone creating two main types of morphostructures: asymmetrically and symmetrically offset cones. For the asymmetrical end member, we take the example of the Pliocene Cerro Peineta volcano, located in Chile immediately south of Ollagüe. A new geological map of the cone (Fig. 5B; Uttini, 2005) shows 22 lithostratigraphic units that have been mapped at 1:50,000 scale. We group these into 14 main lithostratigraphic units of volcanic or sedimentary origin, with Cerro Peineta activity grouped into nine units that are in turn combined into three successions of lava and pyroclastic deposits of andesitic to basaltic-andesitic composition, separated by prominent stratigraphic unconformities. The map shows the presence of units 10, 12, and 13 on only the SW side of the volcano, whereas other units, such as 8, 9, 11, and 14, are distributed all over the cone. This distribution suggests synvolcanic tectonic down-sagging of the hanging-wall block of a NW-SE–striking normal fault passing through the cone summit. This fault trace is also indicated by a clear rectilinear scarp facing SW, consistent with a SW-dipping fault (see Figs. 5A and 5B). This normal fault orientation is also consistent with the regional tectonic data portrayed in Figure 3, where...
NW-SE–striking extensional features dominate with a NE-SW–trending \( \sigma_1 \). Units 10, 12, and 13 have been deposited on the down-sagged block and were confined to the NE by the fault scarp. This is also consistent with flow directions of lavas measured in the field; these indicate that some phases of volcanic activity (e.g., units 10, 12, and 13) were accompanied by lavas flowing preferentially along the southwestern half of the cone, which should have clearly been topographically lower than the opposite NE side. After the fault scarp was onlapped by volcanic products, units 8, 9, and 11 were also able to mantle the NE side of the cone. Successive increments of fault displacement also offset the main stratigraphic unconformities such as those illustrated in photos C, D, and E of Figure 5. In particular, the stratigraphic surface above unit 11 is located at 4700 m a.s.l. on the NE side of the volcano (Fig. 5D), whereas this surface is at 4550 m a.s.l. on the opposing SW side (Fig. 5E), again indicating normal offset along a SW-dipping fault. Apart from the main SW-dipping normal fault, two other normal faults, dipping SSW, cut units 8 and 9 (Fig. 5B), indicating new faulting events at the close of Cerro Peineta volcanic activity or in a postvolcanic stage.

Another volcanic center from the same Chilean region, the Pliocene Inacaliri volcanic complex, provides an example of symmetrical rifting. Here, a set of NW-SE–striking, parallel normal faults dip toward each other, indicating the propagation of extensional stresses from the basement through the cones to form a symmetrical rift (Fig. 6A). This NW-SE rift zone is consistent with the regional tectonic data portrayed in Figure 3. The Inacaliri volcanic complex is made up of three main stratovolcanoes whose products are partially interlayered, suggesting a general coeval growth. The three main crater areas are aligned NW-SE, as are the tectonic faults of the region. All the products were emitted by these summit craters, suggesting stable central conduits. After cessation of volcanic activity, the substrate faults propagated across the entire complex and volcanism resumed only in late Quaternary times with the emplacement of the Pabellon lava dome along the main northeastern fault (Fig. 6B).

3. CONSUMED RIFTED VOLCANOES

Let us now examine the way in which volcanoes consume themselves, using the same field examples outlined above. At Stromboli, during a period dominated by summit caldera collapse that forced most lavas to emplace within the caldera walls, the SE flank of the edifice was affected by a collapse (labeled 2 in Fig. 2) dated between 35 ka and 26 ka B.P. (Pasquaré et al., 1993; Tibaldi, 2001). The products of the subsequent cone growth emplaced largely within this amphitheater, flowing southeastwards. From 13 ka B.P., four lateral collapses occurred along the NW volcano flank, alternating with phases of cone growth, and formed horseshoe-shaped amphitheaters open to the northwest (Fig. 2). These collapses involved huge sectors of the volcano (up to 1–2 km\(^2\)), Tibaldi, 2001) and were characterized by deep-seated slip surfaces. Apart from these known large collapses, several others of smaller magnitude have taken place at Stromboli, leaving curved headscars up to 100 m wide and involving rock masses some tens of meters thick; suggesting that they are more superficial landslides. These landslides are mostly located on the NE and SW flanks of the volcano and some are aligned in the NE-SW trend (Figs. 2 and 7). The events that accompanied the evolution of one of these landslides have been reconstructed. On the NE flank of the volcano the San Bartolo (SB) lavas, with an age of 2 ka B.P. (Arrighi et al., 2004; Speranza et al., 2004, 2008), were emitted by an eccentric vent located at 600 m a.s.l., and flowed northeastwards down to the sea (Laiolo and Cigolini, 2006; Fig. 7). This unit is composed of a series of lavas with a total thickness of more than 30 m. The downslope part of the flows is located at the same topographic level as much older stratigraphic units and is bounded by two vertical and subparallel, NE-striking scarpas that connect upslope by a curved headscarp to form a horseshoe shape in plan view (Fig. 7B). Around

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**Figure 3.** Satellite image (Landsat TM) of northern Chile and western Bolivia showing the studied volcanoes (boxes), the main Quaternary extensional structures (red lines), the scars of the Pastos Grandes Caldera (white lines), and the orientation of the Quaternary tectonic least-principal stress (diverging arrows) (structures and stresses after Tibaldi et al., 2009b).
Figure 4. (A) At Ollagüe volcano, Chile-Bolivia border, the vents are generally concentrated along a NW-SE zone, and most of the single domes are NW-SE–elongated. Several fractures and faults are also present along this zone and strike NW-SE as well. These data indicate the presence of a volcanic-rift, and since the related normal faults dip to the SW, the final architecture is an asymmetric rift. (B) Photo of Ollagüe looking NE with traces of along-rift landslides and normal-to-rift deep lateral failure.

The growth of Ollagüe volcano was also interrupted by a series of lateral collapses. The largest of these occurred toward the WSW and involved a huge sector of the volcano, producing a late-Pleistocene debris avalanche (Francis and Wells, 1988; Feeley et al., 1993; Clavero et al., 2005) (Fig. 4). Postcollapse lavas were erupted within and around the collapse amphitheater. More superficial landslides took place mostly toward the NW and SE, and were characterized by high length/width ratios. Most of them have been guided by the rift structures. Since volcanic activity has been very low over the past hundreds of thousands of years, the effects of normal surface erosion processes have not been hidden by volcanic products. Scree tongues surround the entire volcano base, but the widest are present toward the SW, the SE, and the NW, in descending order. Gullies have developed and most of them are rectilinear and oriented toward the NW (Fig. 4B). Field observations show that they are guided by rift system fractures. The largest erosional valley is aligned toward the SE on the SE flank of the volcano and clearly follows the main normal fault that crosses the cone.

The other volcano case studies show linear gullies and deeply eroded elongated depressions dominantly aligned NW-SE, parallel to the trend of all major volcanoes in this area of the Chilean Andes (Tibaldi et al., 2009b). Cerro Peineta shows deeply eroded valleys that follow
Figure 5. Original data from the Pliocene Cerro Peineta volcano, located in Chile immediately south of Ollagüe. (A) Aerial photograph with orientation of photos C, D, and E. (B) Geological map showing the main lithostratigraphic units and normal faults. These faults have a dominant dip to the SW and propagated from the basement. (C) This rock succession made of units 10 and 12 is present only on the SW side of the volcano, indicating synvolcanic tectonic down-sagging along the normal fault hanging-wall block. (D) The stratigraphic surface above unit 11 is located at 4700 m a.s.l. on the NE side of the volcano, whereas in photo (E) this surface is at 4550 m a.s.l. on the opposing SW flank. This again indicates normal offset along a SW-dipping fault.
the NW-SE–striking normal faults and that have developed on opposing NW and SE flanks of the volcano (Figs. 5A and 5B). As can be seen in Figure 5A the scree tongue from the valley on the SE flank widens southward because it is deflected by the presence of another volcano. Due to downfaulting of the southwestern half of Cerro Peineta, the two main NW-SE–trending valleys are asymmetric; the northeastern valley slope being higher than the opposite southwestern slope.

The highest erosion rates on the Inacaliri volcanic complex occur within the symmetric rift, rather than on the outer volcano flanks (Fig. 6A). Zones of preferential erosion have developed along the main fault scarps and debris is drained along the rift floor; the floor of the southeastern rift is completely filled with debris. Other channeled zones of enhanced erosion have developed at the tips of the Inacaliri rift and trend mostly NW-SE. The northeastern and the southwestern flanks of the volcanic complex are relatively poorly eroded.

4. ANALOG MODELING

Scaled models have been carried out in order to test the relationships between dyke intrusions and landsliding along- and normal-to-dyke strike. Models have been scaled both kinematically and dynamically in order to be a valid representation of the natural prototype (Hubbert, 1937; Ramberg, 1981) following the classical scaling methodology for volcano deformation (e.g., Merle and Borgia, 1996; Vidal and Merle, 2000; Lagmay et al., 2000; Acocella, 2005; Tibaldi et al., 2006). In order to achieve dynamic similarity in our study, the Ramberg number for brittle deformation $(R = \rho g L / c)$, where $\rho$ is the density, $g$ is the acceleration of gravity, $L$ is the characteristic length, and $c$ is the cohesive strength) must be the same in the model and in nature $(R_{\text{prototype}} / R_{\text{model}} = 1)$. This condition is achieved in natural gravity $(g_{\text{prototype}} = g_{\text{model}})$ choosing the right linear scale factor $(L_{\text{prototype}} / L_{\text{model}})$. Under brittle conditions deformation is not time-dependent, provided that inertial forces are negligible (Hubbert, 1937). Three different analog materials have been selected (Q150 sand, Q100 sand, and flour), simulating the main types of volcanic materials observed at the field-studied volcanoes. Using these materials, whose properties are described in detail by Tibaldi et al. (2006) and reported here in Table 1, dynamic similarity is achieved for a linear scaling ratio of $1/25,000$. The apical angle of the cone has been maintained similar in the model and in nature. This is in accordance to the principle that angular quantities, being dimensionless, should be equal in the prototype and in the model (Hubbert, 1937).

The analog cone $(D = 40 \text{ cm}, h = 10 \text{ cm})$ was composed of alternating layers of different colors of the above-described analog materials, so as to simulate internal lithologic variations with different cohesion, since all the field-studied volcanoes are made of interlayered breccia, lava, and pyroclastic deposits. A series of experiments have also been done with the central portion of the cone made of SQS, in order to simulate only the case of highest frequency of dykes and eruptive centers observed in the field-studied volcanoes, at different depths. We performed this series of experiments in order to simulate only the case of thin dykes growing too thick. We retain that this is the best simulation representing the case of a lateral displacement produced by the sum of several dyke emplacements, as usually occurs at volcanic rifts. We are aware that other types of experiments might be done, in the attempt of

Figure 6. (A) In the Pliocene Inacaliri volcanic complex, a set of NW-striking parallel normal faults dipping toward each other indicates propagation of extensional stresses from the basement across the cones to form a symmetrical rift. (B) After cessation of the volcanic activity, the substrate faults propagated across the entire complex, and volcanism resumed only in Late Quaternary times with the emplacement of the Pabellon lava dome along the main northeastern fault.

Geosphere, June 2010 195
simulating dykes that propagate with different directions, but they were out of the scope of the present work.

Figures 9A and 9B show the case of a shallow dyke intrusion; upon initial dyking, a deformation zone developed along a cone flank perpendicular to dyke strike. This deformation zone developed above the dyke tip and was elongated as the dyke strike, with an elliptical shape in plan view. Deformation occurred first as a bulging, and rapidly developed into a small, elongated, surficial landslide. The landslide scar is located exactly above the dyke tip and the landslide deposit glided down along the dyke strike.

In other series of experiments, deeper and thicker dyking along the same zone of intrusion was attained by two ways: (1) incremental pumping of air into the same single vertical plastic sheet, or (2) successive air pumping in different parallel, plastic sheets in contact with each other. In both cases, the response of the analog cone to these thicker intrusions was first the development of a wide bulge zone on the cone flank, perpendicular to the dyke zone; this was followed by the failure of a wide sector of the cone (sector collapse s.s.) with a debris avalanche in a direction perpendicular to dyke strike (Figs. 9C and 9D).

5. NUMERICAL MODELING

5.1. Conceptual Model

A stress-strain analysis of volcano-intrusive sheet interaction was performed using the FLAC 3D code (Fast Lagrangian Analysis of Continua in 3 Dimensions; ITASCA, 2005), a three-dimensional explicit finite difference numerical code for engineering and geomechanical simulations. The conceptual model of the volcano was discretized by a grid adjusted to fit the shape of the object to be modeled (Fig. 10). Each grid element has the physical-mechanical properties of the corresponding material in the geotechnical model (Table 2) and behaves according to a prescribed linear or nonlinear stress-strain law in response to the applied forces and boundary restraints.

The code allowed us to simulate the effects of dyke emplacement below the summit craters on the deformation of the volcano flanks, in terms of magma pressure and overpressure applied to the intrusion walls. The simulation was hence implemented to model the deformation pattern in the host rock due to a dyke that propagated from the central conduit toward the NE.

The simulation was implemented in successive stages, in order to overcome the inability of the code to reproduce a continuous evolutionary process. The geometry and the magma pressure for each successive stage was defined using the results of the previous stages as conditions for the next one, as well as using field information collected at the case-study volcanoes (Fig. 11). As a real case, both for dimension and parameters, we simulated the Stromboli volcano that was discretized by a regular grid with a cell size of 100 m (Fig. 10). The dykes are represented as discontinuities in the grid, and are modeled by means of “interfaces.” Such interfaces have been placed along a NE-striking vertical plane passing through the summit craters (Fig. 10). Physical-mechanical properties were assigned to the interfaces, as reported in Table 3. The interface portions representing the dyke filled with magma were set “unglued” (no bond between the two sides of the interface), while a traction bond equal to 10 MPa was imposed elsewhere: this simulates the strength of the rock against the propagation of the fracture. Magmatic pressure is modeled as a normal pressure applied to both sides of the interfaces. Since the model does not reproduce the transient injection process, the shear stress applied by the magma during dyke emplacement is not simulated.

The setup of the numerical model required a simplification of the geological model. No superficial loose materials were considered, and the volcano was modeled considering one
lithotechnical unit, whose physical and mechanical parameters are mediated from the different geological units (“lava-breccia unit,” represented by alternating lava and breccia layers, after Apuani et al. [2005a, 2005b] and Apuani and Corazzato [2009]), as summarized in Table 2.

Based on the observation that discontinuities and geological complexity at outcrop scale result in a reduction of rock mass strength and deformability modulus, compared to the intact rock properties (Hoek and Brown, 1997), strength and deformability of the volcano rock mass were obtained (Apuani et al. 2005a) by combining intact rock properties with the field-based assessment of the “rock mass quality” (Hoek and Brown, 1980), using rock mass rating (ISRM, 1981) and Geological Strength Index (GSI, Hoek, 1994; Sonmez and Ulusay, 1999; Hoek et al., 2002). This was done according to the Hoek-Brown criterion (Hoek et al., 2002), which enables property scaling from intact-rock scale to rock-mass scale.

In the model, an elasto-plastic constitutive law was adopted and a homogeneous Mohr-Coulomb strength criterion was chosen for the volcanic cone. Mohr-Coulomb equivalent strength and elastic parameters for the rock mass (friction angle $\varphi$, cohesion $c$, tensile strength, modulus of deformation) were obtained (Apuani et al., 2005a) by fitting an average linear relationship to the curve generated by solving the generalized Hoek-Brown nonlinear equation (Hoek et al., 2002), mediated for the edifice. The Mohr-Coulomb properties of the material influence principally the deformation of the volcano flanks in response to the stress conditions due to the dyke; otherwise, the propagation of the dyke is mainly controlled by the interface bond strength. We consider this procedure as the best reliable in view of the complexity of the studied volcano; we also recognize that a stratovolcano is a composite structure and a given layer may be considerably stronger than implied by the Mohr-Coulomb criterion when fracturing, due to its possible very high toughness (Gudmundsson, 2009).

Groundwater conditions were imposed assuming the volcanic edifice was saturated in the portion below the sea level, and dry above, consistently with the general absence of springs on the Stromboli Island. Effective stresses were computed accordingly. On the contrary, possible superficial water tables into the superficial deposits were not taken into account: they are below the representability of the grid cell size (100 m) and are neglectable in the issue of flank collapses.

The results of each simulation stage, expressed in terms of shear strain increments or displacements, represent the equilibrium state under the imposed conditions. Simplifications and assumptions, necessary for numerical modeling procedures, allowed us to obtain results that can be compared to field observations in terms of distributions and trends, rather than expecting equivalence in absolute values.

### 5.2. Procedure and Results

The first step in the simulation is designed to model the initial state of equilibrium and to determine the stress configuration resulting from volcano geometry. The stress field is a crucial factor in determining the intrusion directions of the sheets, generally developed along the $\sigma_1-\sigma_3$ plane, perpendicular to the $\sigma_3$.

The second step is aimed at simulating the partial filling of the magmatic conduit, modeled as a vertical plane located below the active craters and oriented in a NE-SW direction, consistent with the structural data on the summit rift zone previously described. Magma pressure was applied for a horizontal range of 400 m from the bottom of the edifice to the magma raising level assumed, in this stage of the simulation, equal to 500 m a.s.l.

The applied value of magma pressure was calculated as a sum of two components: the
TABLE 1. PHYSICAL PROPERTIES OF ANALOG MATERIALS USED IN THE EXPERIMENTS AND VOLCANIC ROCK MASSES OF OLLAGÜE

<table>
<thead>
<tr>
<th>Material</th>
<th>Density (t/m³)</th>
<th>Cohesion (Pa)</th>
<th>Characteristic length (m)</th>
<th>R = ρgL/c</th>
<th>Refs.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SQS sand (0.150–0.210 mm)</td>
<td>1.500</td>
<td>25</td>
<td></td>
<td>0.1</td>
<td>58.885</td>
</tr>
<tr>
<td>Q 100 sand (0.100 mm)</td>
<td>1.200</td>
<td>69</td>
<td></td>
<td></td>
<td>17.068</td>
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<tr>
<td>Flour</td>
<td>1.200</td>
<td>330</td>
<td></td>
<td></td>
<td>3.569</td>
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<td>Prototype</td>
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<tr>
<td>Pyroclastic deposits</td>
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<td>1.00E+06</td>
<td></td>
<td>2500</td>
<td>58.885</td>
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<tr>
<td>Fractured andesitic lavas</td>
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<td>5.00E+06</td>
<td></td>
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<td>13.004</td>
</tr>
<tr>
<td>Poorly fractured andesitic lavas</td>
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<td>2.00E+07</td>
<td></td>
<td></td>
<td>3.251</td>
</tr>
</tbody>
</table>

Note: Density (ρ) and cohesion (c) values of model and natural materials have been deduced from laboratory tests by (a) Krantz (1991), (b) Saotome et al. (2002), (c) Lohrmann et al. (2003), and (d) Apuani et al. (2005a, 2005b), carried out on materials with comparable composition, grain size, texture, and structure. Given a linear scale of 1:25,000, the characteristic length L is 10 cm for the model and 2.5 km for the prototype. R is the Ramberg number. Gravity acceleration g = 9.81415 m/s².

Figure 9. Experimental results from physically scaled analog models mimicking dyking in a central volcano. (A) Cone after intrusion of a thin dyke. (B) Interpreted sketch; note the development on the cone flank along the dyke path of a deformation zone with small landslide instability. (C) Cone after intrusion of a thick series of dykes. (D) Interpreted sketch: note the development of a sector collapse on the cone flank perpendicular to the dyke sheets. Inflation was produced via a plastic bladder, mimicking basic magmas, which was placed into the model cone during construction and then inflated with a controlled air supply.
first, due to the weight of the magma filling the vertical dyke, is linearly increasing with depth, and was calculated as \( P_m = \gamma h \), where \( \gamma \) is the magma unit weight, assumed equal to 26 kN/m\(^3\), and \( h \) is the length of the overstanding magma column. The second component is related to the driving pressure of the dyke and was hypothesized in consideration of the fact that, during an eruption, the magma has a driving pressure enough to reach the summit crater, located at a height of ~750 m a.s.l. Magma overpressure in the dyke was then calculated as equal to the overpressure necessary to lift the magma until the summit craters. This means an overpressure equal to 6.5 MPa (Fig. 11A).

The results of this stage, expressed in terms of deformation and shear strain increment contours, show a pervasive deformation path with a highly deformed zone in the upper part of the volcano cone. Deformation can be interpreted as a bulging of the lower part of the volcano flanks, with a main deformation component in the NW direction for the northwestern flank, and in the SE direction for the southeastern one. The pattern of deformation is then quite parallel to the magmatic pressure applied in the dyke. The upper part of the cone shows a deflation deformation pattern, with deformation vectors mainly directed downward (Figs. 11C and 11D). The amount of deformation developed by the model is less than 10 m at the surface of the model, and ~15 m at the bottom of the dyke, where magma pressure is higher. Figure 11B shows the separation induced by the magma pressure at the two sides of the interface representing the active dyke. The mushroom shape of the separated portion denotes the tendency toward lateral propagation of the dyke in the upper part of the model. The distribution of the shear strain increment evidences the presence of a highly stressed portion in correspondence with the Sciara del Fuoco flanks (Fig. 11E). This horseshoe-shaped region can represent a preferential location for the intrusion of dyke swarms emplaced with direction parallel to the flanks of the Sciara del Fuoco. This is in accordance with field structural data (Corazzato et al., 2008; Tibaldi et al., 2009a, and references therein). The intrusion of these dykes is controlled by the stress field due to the geometry of the Sciara del Fuoco sector collapse depression. In Figure 11E in the vertical section offshore Sciara del Fuoco it is important to note the presence of a zone of shear strain exactly placed between the submarine trace of the Sciara del Fuoco lateral scarps. This shear strain zone coincides with the possible emersion of the sliding surface of the major sector collapses of Stromboli as suggested on the base of structural and geological data by Tibaldi (2001).

**Figure 10.** FLAC 3D numerical model setup. The grid has a cell size of 100 m, and each element has the physical-mechanical properties reported in Table 2. Dykes are represented by means of interfaces, whose properties are reported in Table 3. See text for details.
Figure 11. (A) Model NE-SW cross section showing the interface and the region in which the magma pressure has been applied (in purple). The applied magma pressure has a magmastic component, increasing with depth, represented by a triangle (white background), and an overpressure represented by a rectangle (yellow background). The amount of overpressure was assumed equal to 6.5 MPa (see text for details). (B) Contours of the separation of the two sides of the interface (representing the active dyke) induced by the magma pressure acting on the dyke. The mushroom shape of the separated portion denotes the tendency toward lateral propagation of the dyke in the upper part of the model. (C) Surface displacements after the application of the magma pressure in the vertical dyke. Sea level and Sciara del Fuoco depression rim are highlighted. (D) Displacement contours at depth. Maximum displacements at the dyke bottom are 15 m. Gray vectors shows displacements at model nodes. (E) Contours of shear strain increment at model surface. The application of the magma pressure in the dyke induces shear strain concentration along the flanks of the Sciara del Fuoco.
Figure 12. (A) Modeling of lateral propagation of the dyke toward the NE for 500 m. The dark purple region represents the portion of the interface in which the magma pressure was added to model lateral dyke propagation. (B) Interface separation confirms the aperture of tension cracks and eruptive vents above the dyke. (C) Surface displacements show the effects of the lateral propagation of the dyke. The highly deformed region is the upper NE side of Sciara del Fuoco. (D) SE-NW cross section showing the displacement contours at depth. Displacements are triggered by the magma pressure acting on the dyke and show a bulging with a downward sliding of the flank portion in front of the dyke. (E) Contours of shear strain increment in cross section show a highly stressed region above and in front of the lateral propagation of the dyke.
The model reaches equilibrium conditions and the global effects are not sufficient to expect any large flank collapse. But, in dependency to both the amount of magma pressure applied on the dyke and the geometrical relations between the dyke and the volcano flanks, the magma pressure in the dyke could represent a triggering effect for the development of deep-seated flank collapses with movement direction perpendicular to the dyke strike.

We then modeled the lateral propagation of the dyke for 500 m toward the NE. Magma pressure was added to the northeastern part of the interface, as shown in Figure 12A. This modeling was aimed at simulating an intrusion event comparable to the one that produced the outpouring of the 2002 lavas, which occurred along the rift in the Sciara del Fuoco between the active crater zone and the Sciara del Fuoco northeastern escarpment. Contours of shear strain increment at the surface and in cross section (Fig. 12E) show a highly stressed region developed around the active dyke. This stress field induced by the magma pressure acting on the dyke walls determines yield conditions, both in shear and in tension, for the model’s blocks located above the active dyke, at the volcano surface. This is consistent with the opening of surface tension cracks and sink holes that usually develop before and during eruptive crises. Deformation is concentrated in the upper northeastern flank of the Sciara del Fuoco and reaches, in this model stage, a maximum value of 3 m (Figs. 12C and 12D). Flank deformation, induced by magma pressure acting on the dyke, could contribute to the development of tensional stress conditions at the top of the volcanic cone, and then promote the upward propagation of the dyke. The two processes, namely the intrusion of the dyke and development of deep-seated flank instability, are then interdependent processes, mutually influenced and reciprocally triggered. The evolution of rift fissures into normal faults could be determined by evolution of the magma-induced flank instability, under the effects of gravitational forces, contemporarily and after the intrusive phase. Interface separation (Fig. 12B) confirms the aperture of surface tension cracks and eruptive vents above the dyke, as observed during the 2002 eruptive crisis.

Successively we modeled the lateral propagation of the dyke toward the NE for a further 100 m (Fig. 13A). In this stage, the dyke tip reaches the topography simulating the aperture of a vent similar in position to the emission center of the San Bartolo lavas. Surface displacements developed by the model show the effect of the dyke lateral propagation. Deformation concentrates on the northeastern volcano flank, with the most deformed region located outside the border of the active dyke zone.

Figure 13 (continued on following page). (A) Lateral propagation of the dyke toward the NE for further 100 m. The purple region represents the portion of the interface in which the magma pressure was added to model dyke propagation. (B) Interface separation suggests the aperture of tension cracks and eruptive fissures on the NE volcano flank, in accordance with the historical eruption of the San Bartolo lavas. (C) Surface displacements show the effects of the lateral propagation of the dyke. Deformation concentrates on the NE volcano flank, with the most deformed region located outside the border of Sciara del Fuoco. (D) Subsurface displacements.
of the Sciara del Fuoco depression (Figs. 13C–13E). Contours of x-displacements (Fig. 13G) show a highly deformed zone on the northeastern flank of the volcano. X-displacements are a minor component of the 3D displacement, but are useful to highlight the high rate of deformation developed at the emersion of the dyke on the northeastern volcano flank. Surface and deep shear strain increment contours show the development of a highly stressed band above the dyke (Figs. 13F–13H). In this region, the model blocks are in tensile conditions, in accordance with tension cracks parallel to the dyke.

The geometrical relations between the dyke intersecting the ground and the slope orientation determine the characteristics of the induced flank instability. Models clearly show that when the dyke is emplaced parallel to the volcano flank, deformation is maximized and instability conditions for deep-seated collapses could be reached. On the contrary, when the dyke and the slope are normal to each other (dyke strike perpendicular to slope orientation), displacements are lower and there is no evidence (from numerical models) of deep-seated instability. But it is important to note that in both cases the stress field developed at the surface could generate superficial instability, especially on weak volcanioclastic deposits and heavy fractured rocks. As previously exposed, the developed numerical models represent a simplified geological model in which strength and deformability of the rocks are mediated by the volcanic units actually

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**Figure 13 (continued).** (E) Subsurface displacements on two perpendicular directions evidences the deformation pattern induced by the propagation of the dyke. (F)–(H) Surface and deep shear strain increment contours show the development of a highly stressed band above the dyke. Model blocks in this region are in tensile conditions, in accordance with tension cracks parallel to the dyke. (G) Contours of x-displacements show a highly deformed zone on the NE flank of the volcano. X-displacements are a minor component of the 3D displacements (vectors).
In spite of the differences among these models, et al., 2003), or they can intrude actively, pushing rift due to lateral spreading of the cone (Branca can intrude passively, following dilation in the ent models of rift zone origin imply that dykes (Borgia, 1994); and (3) tectonic control exerted (Merle and Borgia, 1996; Munn et al., 2006); Consequently the destabilization of the volcano mation of a weak substrate under the weight of a (Borgia et al., 1992; Clague and gia, 1996; Borgia et al., 2000; Walter, 2003), or merly by steeper slopes, while distant downrift eruptions noes, it has also been demonstrated that erup- tions near the volcano summit are favored by erup- tions within the cones, as seen at other volcanoes by Walker (1992). This is also consistent with the possibil- ity that dykes propagate not only from below, but also horizontally from the main central conduit (Accocella and Neri, 2009, and refer- ences therein). In this latter case, dykes should be concentrated within a certain distance of the central conduit, thus maximizing the number of dykes lying below the central summit zone of a cone. Since dyking within a volcano causes a magmatic overpressure and lateral displacement of the cone flank along deep-seated slip surfaces (Voight and Elsworth, 1997), possibly triggering flank failure, this lateral displacement should be greater where intrusions have the highest fre- quency. Thus, as at Stromboli, inception of deep flank failure occurs along two directions that are orthogonal to the segment of the rift where the maximum dyking is reached. This is also consistent with the wide recognition of sector col- lapses that developed perpendicular to dyke or rift zones (Siebert, 1984; Siebert et al., 1987; McGuire et al., 1990; Elsworth and Voight, 1995; Voight and Elsworth, 1997; Voight, 2000; Donnadieu et al., 2001; Tibaldi, 1996, 2001), as well as with instances of sector collapses in opposed outward directions normal to a rift zone, such as at Nisyros in Greece (Tibaldi et al., 2008), at Tahtı-Nui in French Polynesia (Hildbrand et al., 2004), and at Stromboli. On the inception of flank failure, the initial collapsed mass is likely to have sufficient mobil- ity to travel laterally far enough to promote fur- ther slices of material failing in rapid succession in a retrogressive fashion. In the case of volca- noes affected by asymmetric rifts, as at Ollágüe, lateral collapse can occur preferentially on the tectonically downhill side (i.e., here to the SW). This is due to a series of co-causes such as fault slip that creates a gravity gradient in the same direction of the fault dip, the forma- tion of other mechanical discontinuities in the present and do not consider the weak superficial deposits. The effects of the dyke propagation on superficial instability are then underestimated by this numerical model, and superficial insta- bility on the northeastern flank would be larger than that developed by the models.

6. DISCUSSION

6.1. Rift Zones and Volcano Growth

In analog experiments, Fiske and Jackson (1972) showed the importance of the volcano load and stress distribution for the formation of rift zones on volcanoes. Their results suggest that the topography of a volcano directs dyke paths where rifting is parallel to a topographic ridge, but do not explain why the volca- no grows with an elongated shape in the case before rift initiation. Fiske and Jackson’s (1972) analog model was recalculation by Dieterich (1988), who found that volcano flanks must dilate laterally to overcome a plumbing effect. Additionally, the presence of weak substrata (e.g., deformable ocean sediments, lacustrine deposits, etc.) provides décollement zones at the base of a volcano (Van Bemmelen, 1949; Nakamura, 1980; Delaney et al., 1998; Bor- gia et al., 2000). Similarly, a viscous substrate upon which volcanoes may be built can cause volcanoes to spread laterally, driven purely by their own weight (Borgia, 1994; Merle and Bor- gia, 1996; Borgia et al., 2000; Walter, 2003), or favored by the growth of the intrusive cumu- litic complex (Borgia et al., 1992; Clague and Denlinger, 1994). The southern Kilauea flank expands on such a basal sliding surface, creat- ing space for dykes in the Kilauea rift (Diet- erich, 1988). Rift zones can thus originate from: (1) gravitational spreading, defined as the deformation of a weak substrate under the weight of a volcano that induces deformation of the cone and consequently the destabilization of the volcano (Merle and Borgia, 1996; Munn et al., 2006); (2) volcanic spreading, defined as the destabiliza- tion of the cone when sufficient magma input is added to the process of gravitational spreading (Borgia, 1994); and (3) tectonic control exerted by a substrate that is locally or regionally subject to horizontal tensional stresses related to plate tectonic motions (Nakamura et al., 1977), or to a dipping substrate (Wooller et al., 2004). Different models of rift zone origin imply that dykes can intrude passively, following dilation in the rift due to lateral spreading of the cone (Branca et al., 2003), or they can intrude actively, pushing apart the volcano flank, or both modes can oper- ate alternatively (Tibaldi and Groppelli, 2002). In spite of the differences among these models, all of them imply that growth of volcanoes must occur along the rift zones. This growth is attributed to the concentration of adjoining cinder cones, lavas erupted from fissures, and intru- sion of parallel dyke swarms within an elong- ated area (Fiske and Jackson, 1972; Dieterich, 1988; Carracedo, 1994; Moore et al., 1994; Walker, 1999; Walter and Schmincke, 2002). This along-rift concentrated activity can occur also at relatively small polygenetic lava volca- noes and stratovolcanoes, such as those shown by Pasquaré et al. (1988) in the Mexican Volcanic Belt. Based on all this, we consider that rift zones can also develop in the earliest stages of growth of a stratovolcano. Once a rift is established, magma that reaches the surface of the rift flows nearly parallel or perpendicular to the rift zone axis, the latter case being more common at all basaltic shield volcanoes. For example, at the S-N rift zone of Mount Etna, most lavas flowed parallel to the rift zone (Behncke et al., 2008). At the upper stratovolcanoes here studied, most lavas outpoured at rift zones flow parallel to the rift axis. Stratovolcanoes should grow in directions perpendicular to the rift strike mostly by lateral expansion due to intrusions and parallel to the rift mostly by effusions. This might explain why elongated volcanoes do occur, but it is also necessary to understand how their aspect ratio might evolve. We will contribute to this topic in the next section.

6.2. Localization of Volcano Consumption in Relation to Dyking

In a common view, the along-rift growth of a volcano should be mainly dictated by the dis- tance from the cone summit reached by dyke- fed eruptions. Surface eruptions associated with dykes mostly concentrate near the tip of the dyke (Fiske and Jackson, 1978), and the tip may propagate vertically, horizontally, or at any oblique angle. Dykes will tend to propagate laterally if the difference between magma pressure and the total horizontal stress in the surrounding rock at the dyke center exceeds the difference at the top and bottom of the dyke (Rubin andPollard, 1987; Rubin, 1995). At shield volca- noes, it has also been demonstrated that eruptions near the volcano summit are favored by shallow slopes, while distant downrift eruptions may be encouraged by steeper slopes, although along-rift variations in the stress state may interfere with this process (Fialko and Rubin, 1999). In Japan’s stratovolcanoes it has also been demonstrated that steeply sloping edifices may encourage lateral injection (Geshi et al., 2009). In the present section we wish to demon- strate that another important factor in shaping stratovolcanoes is that along-rift constructive processes are balanced by destructive processes that closely interact, both perpendicular to the rift and along the rift strike. In our view, a stra- tovolcano can be consumed in two directions at right angles to each other, within a process that is strictly controlled by the rift geometry and magma upwelling. This holistic view is high- lighted, for example, by the data outlined above for Stromboli and Ollagüe. In both volcanoes, dykes and/or flank vents and eruptive fissures increase in number approaching the summit part of the cone. This indicates that dykes should increase in number below the central part of the cone, as seen at other volcanoes by Walker (1992). This is also consistent with the possibil- ity that dykes propagate not only from below, but also horizontally from the main central conduit (Accocella and Neri, 2009, and references therein). In this latter case, dykes should be concentrated within a certain distance of the central conduit, thus maximizing the number of dykes lying below the central summit zone of a cone. Since dyking within a volcano causes a magmatic overpressure and lateral displacement of the cone flank along deep-seated slip surfaces (Voight and Elsworth, 1997), possibly triggering flank failure, this lateral displacement should be greater where intrusions have the highest fre- quency. Thus, as at Stromboli, inception of deep flank failure occurs along two directions that are orthogonal to the segment of the rift where the maximum dyking is reached. This is also consistent with the wide recognition of sector col- lapses that developed perpendicular to dyke or rift zones (Siebert, 1984; Siebert et al., 1987; McGuire et al., 1990; Elsworth and Voight, 1995; Voight and Elsworth, 1997; Voight, 2000; Donnadieu et al., 2001; Tibaldi, 1996, 2001), as well as with instances of sector collapses in opposed outward directions normal to a rift zone, such as at Nisyros in Greece (Tibaldi et al., 2008), at Tahtı-Nui in French Polynesia (Hildbrand et al., 2004), and at Stromboli.
downthrown block that favor the failure, and bulging in the downthrown flank (Tibaldi, 1995; Merle et al., 2001). The depression created by such a lateral failure will concentrate subsequent lava emission and, after the depression is filled, the cone attains a new critical height (and mass) that promotes renewed lateral failure. The result of this process is that volcanoes are unlikely to be widened normal to the rift zone. Multiple lateral failures of volcanoes are being recognized worldwide (see Tibaldi and Lagmay, 2006), up to the extreme example of the St. Augustine volcano in the Aleutian Arc, which suffered 11 lateral collapses interspersed with constructive phases in which the cone rebuilt to about the same elevation (Beget and Kienle, 1992).

A dyke within a rift zone can either propagate to the surface, or become arrested at depth. The latter case can be produced by the presence of some sort of a stress barrier, a transverse discontinuity, or an abrupt change in Young’s modulus that arrests the vertical propagation of the dyke (Gudmundsson, 2003). On the other hand, when a dyke propagates to the surface, it can produce surface deformations also along the rift. This phenomenon has been mainly explored assuming a horizontal topography below which a planar vertical intrusion is numerically simulated (Pollard et al., 1983; Rubin and Pollard, 1988; Rubin, 1992; Roth, 1993; Bonafede and Danesi, 1997; Gudmundsson, 2003; Gudmundsson and Loetveit, 2005; Behn et al., 2006). Pollard et al. (1983) showed the development of two symmetrical elongated bulges bounding a zone of down-sagging, parallel to the dyke. Gudmundsson and Loetveit (2005) indicated that the main effects of a dyke propagating vertically upwards within a graben are, first, to close its boundary faults and, then, to encourage reverse slip on the faults. In all these cases, the upward-directed magma force is smaller than the force acting normal to the rift zone, and only localized bulging of the surface cover can occur. Our work, by contrast, has focused on the along-rift deformation induced on a quite steep sloping topography, such as the flanks of a stratovolcano like Stromboli and Ollagüe. On these slopes, surface bulging induced by dykes locally increases the downslope angle, while decreasing the angle upslope of the future magma emission point, as observed at the San Bartolo eruption center at Stromboli. Field evidence from Ollagüe (Vezzoli et al., 2008) and Stromboli, as well as our analog modeling, indicate that one or more aligned landslides can be initiated from such a zone of bulging. These landslides form parallel to the rift: the slide planes can exploit lateral ramps from rift-parallel fissures, faults, and fractures that usually characterize volcanic rift zones. Additionally, our numerical modeling

Figure 14. (A) General geological-structural sketch map of Mount Etna (Italy) showing the rift zones on the volcano (modified after Corazzato and Tibaldi, 2006, and Neri et al., 2004). The inset shows the location. (B) Map of the main latest Pleistocene-Holocene faults and of the July-August 2001 fractures affecting the summit zone of Mount Etna. Here the N-S rift zone is expressed by normal faults with converging dips and by a prehistoric-historic lava succession (gray area) channeled within the graben depression. The dyke that fed the July-August 2001 eruption induced a surface incipient landslide expressed by the horseshoe-shaped headwall scarp and the lateral slip zones; this unstable zone has been here evidenced in orange (redrawn after Tibaldi and Groppelli, 2002).
shows deformation and the development of shear stresses near the surface at the propagating dyke tip. These zones of deformation represent the most plausible localization of future sliding surfaces. The landsliding, in turn, unbuttresses the dyke tip, promoting effusive eruption and lava flows within the landslide depression. In this case we can obtain a balance between new outpoured lava and gravity mass removal, or the volcano can grow along the rift if the emitted magma volume is higher than the collapsed material. Our findings are also consistent with other dyke-triggered surface landslides that occurred along volcanic rift zones at Mount Etna (Tibaldi and Groppelli, 2002; Billi et al., 2003) and at Nyiragongo volcano (Komorowski et al., 2002). At Mount Etna (Fig. 14), a dyke emplaced during the July–August 2001 eruption along the North-South rift zone produced an incipient surface landslide. This is expressed by a horseshoe-shaped headwall scarp and the lateral slip zones with kinematics compatible with a southward (i.e., downward) slide (large red arrow in Fig. 14B). At Mount Etna, the cumulative displacement due to repeated dyke intrusions along the same North-South rift zone has also produced deep-seated flank slip of the entire eastern Mount Etna flank at a right-angle respective to the rift (Borgia et al., 1992).

From a mechanical point of view, the different behavior of volcanic slopes parallel and perpendicular to the dyke strike is a function of magma force and cumulative displacement. Theoretical models indicate that any significant magmatic overpressure in a dyke propagating through a homogeneous, isotropic crust, always produces very high stresses at the dyke-tip (Pollard et al., 1983). From typical aspect ratios (along-strike dimension/thickness) of dykes and the estimated static overpressures of several mega-Pascals or more, numerical models show theoretical values of the tensile stresses at dyke tips as ~10E3–10E4 MPa (Gudmundsson, 1990, 2000, 2003). Even if the radius of curvature referred only to individual dyke segments, with lengths of tens of meters to a hundred meters, theoretically dyke tips show tensile stresses of ~10E2–10E3 MPa (Gudmundsson, 2003). The in situ tensile strength of solid rocks is 0.5 – 6 MPa (Schultz, 1995; Amadei and Stephansson, 1997), so the theoretical tensile stresses at dyke tips are orders of magnitude greater than the tensile strengths of typical host rocks. When magma approaches a volcano slope, dyke propagation produces tensile stresses around the dyke tip that fracture the host rock, or produce offset of unconfined surface layers. By these means magmatic stress can deform the topographic surface in a zone whose geometry, assuming other parameters are the same, is a function of the angle between the dyke tip line and the volcano slope. As one endmember, the dyke tip line approaches the slope at a highly oblique angle to slope contours, as a result magmatically induced deformation should be concentrated in a small zone (smaller rift-parallel landslide in Fig. 14). At the other extreme the dyke tip line is parallel to slope contours, so a narrow elongated deformation zone can be produced. In both cases, deformation generally occurs at different altitudes along the slope intersected by the rift, and the cumulative deformation gives rise to an elongated deformed zone along the rift. Additionally, since only part of the dykes originating from the central conduit zone can extend far enough to intersect the slope in any way, the cumulative deformation on the slopes intersected by the rift (i.e., the slope perpendicular to dyke strike) is produced only by a percent of intrusions. By contrast, all the parallel to subparallel dykes of a rift zone produce compression normal to the rift and offset in the same direction of host rock. Thus cumulative displacement of the volcano flanks at the sides of the central rift segment will be the sum of most, if not all, intrusions (Fig. 15). Moreover, the magmatic force exerted by each dyke is acting along the entire dyke wall, also involving the deeper part of the cone flank, as clearly demonstrated by our numerical modeling (Fig. 11). Together the above factors ensure that cumulative deformation induced by magmatic stresses on the volcano flanks parallel to the rift zone involves a much larger slope area than on the flanks aligned normal to the rift zone.

6.3. Morphological Evolution of Stratovolcanoes versus Rift Zones

Here we focus on the effect of rift zones on stratovolcano erosion, also taking into account that a detailed review on the geomorphology of volcanoes has already been given by Thouret (1999). In a symmetric stratovolcano, the combination of along-rift landslides and normal-to-rift deep-seated large lateral collapses produces four main zones of “consumption” and “rebuilding” of the volcano, alternating with four remaining triangular sectors where the volcano slopes are more stable (RSS in Fig. 16A). In these RSS zones, relatively older rocks and fewer flank vents can be found, whereas erosion is enhanced outside these zones. A similar symmetric pattern can be seen at Nisyros volcano (Fig. 16B) where consumption at the two opposite collapsed areas was followed by replenishment of the amphitheater depressions by subsequent volcanic activity (Tibaldi et al., 2008). This deposition of volcanic products perpendicular to the NE-SW rift zone produced a NW-SE growth of the volcano that in this case balances the along-rift effusions, configuring a circular shape to the cone. A similar symmetry can be found also at the Tahiti-Nui volcano (Fig. 16C) (Hildenbrand et al., 2004) where intrusions and effusions along an E-W rift have been compensated first by opposite N-S lateral collapses and then by lava flowing in the same directions. Opposite to Nisyros, the Tahiti-Nui volcano is no longer active, and here it can be appreciated that erosion has been proceeding.
Sheet intrusions and evolution of rifted volcanoes

Both along the collapse zones and along the rift, maintaining the circular shape of the volcano in plan view. More rarely, long-rift consumption can also occur by huge landslides up to deep-seated flank collapses like the one that occurred at Mount Rainier volcano 5.6 ka ago (John et al., 2008) (Fig. 16D). Deep-seated flank collapses and dyke intrusion zones may coincide in the case of emplacement above a transcurrent zone (Lagmay et al., 2000). The Mount Cameroon volcano can be taken as an end member example, since it has a high aspect ratio with a well-developed, long NE-SW–trending rift zone (Fig. 16E). Along the rift the dominant behavior is growth by dyking and opening of vents with lava effusion and emplacement of pyroclastic cones (Suh et al., 2001), whereas consumption takes place in the opposite NW and SE directions by accelerated erosion and landslides.

When volcanic activity ends, rifting can progress as slip continues along parallel inward-facing fault scarps (symmetrical rift), as at the Inacaliri example, or along faults with the same dip (asymmetric rift), as at the Cerro Peineta example. In the first case, erosion develops along the rift fractures and faults producing gullies and valleys aligned along the rift trend. Erosion is concentrated within the rift floor and escarpments, preserving the outer flanks of the volcano. While, by contrast, the cone flanks intersected by the rift are the loci of surface landsliding and deep erosion. The morphostructural evolution of this type of volcano is characterized by a gradual deepening of the erosion level along the rift, giving rise to the gradual formation of two parallel ridges separated by the rift depression. Resumption of volcanic activity in this case will be controlled by the presence of the rift structures and will probably occur by craters located at the rift margins, as in the case of the Pabellon lava dome (Fig. 6). Lava effusion may fill the rift depression without additions to the external volcano flanks.

Asymmetric rifting will also favor preferential erosion along the rift structures, but also, at a minor rate, along the offset blocks. The combination of normal-to-rift and parallel-to-rift

Figure 16. (A) Sketch model of primary and secondary collapse directions on rifted volcanoes, with Stromboli Island as an example (shaded relief topography as in Fig. 2). Note that deep-seated lateral collapses can occur perpendicular to the main direction of dykes and in correspondence of their maximum concentration, whereas more surficial landslides can take place in the same direction of along-rift dyke intrusions. Between these zones of flank instability there are triangular sectors less prone to develop gravity failures and vent openings (RSS). A similar situation is observed at Nisyros, Greece (B) (redrawn after Tibaldi et al., 2008) and at Tahiti-Nui (C) (redrawn after Hildenbrand et al., 2004). At Mount Rainier, USA, a large lateral collapse developed along the main zone of dyking (D) (redrawn after John et al., 2008), whereas at Mount Cameroon, Cameroon, the NE-SW rift is in a major growing stage, and relatively small slope deformations occur along the NW and SE flanks (E).
erosion produces a characteristic morphological reshaping of the volcano flank located on the hanging-wall fault block, which will attain a conical shape. Resumption of volcanic activity will produce preferential deposition on the downthrown flank, with consequent growth of the cone in that direction. This explains why not extinct, asymmetrically rifted volcanoes affected by normal faults, as Ollagüe, can grow with an approximate circular shape. In this case, in fact, along-rift growth can be balanced by growth on the downthrown side.

Since deep-seated lateral collapse and more surficial landslide phenomena are favored by steep slopes, we should expect that a balance between growth and consumption of volcanoes is most pronounced in steep edifices. In fact, basaltic volcanoes with gentle slopes, such as, for example, Hawaii, can reach huge dimensions, and rift zones can increase in size, particularly for cones higher than 3 km (Mitchell, 2001).

The process of consumption of volcanoes is much more complicated and important than recognized up to now, since it may occur at two levels (i.e., deep-seated and surface), it takes place frequently, controls the position of subsequent lavas, creates weak strata interlayered in the rock succession, and controls the general shape of volcanic edifices. We consider this process to be more effective at active volcanoes, whereas at extinct rifted edifices slow erosion occurs following the same trends.

7. CONCLUSIONS

Using field data, scaled physical models, and numerical 3D modeling, we analyzed the processes of growth and consumption of rifted stratovolcanoes in order to better understand their morphostructural and geological evolution in relation to sheet intrusions. The main conclusions are as follows.

Outer growth of rifted volcanoes mostly occurs through fissure eruptions and localized vent eruptions along the rift zone, both at the cone summit and at the two opposing flanks of the rifted cone. Inner growth mostly occurs by the injection of a series of parallel dykes that tend to displace or bulge the two cone flanks located at the sides of the rift.

These constructive processes are compensated by destructive processes that closely interact, both perpendicular to the rift and parallel to rift strike. Deep-seated flank failure is likely to develop normal to the rift. Symmetric rifts may produce collapses that verge outwards in opposing directions, given the volcano lies on a horizontal substrate. If the collapse depression is infilled by subsequent volcanism, the cone attains a new critical height (and mass) and failure can occur again, thus limiting the width of the cone normal to the rift zone.

Field data and modeling show the development of along-rift landslides due to localized bulging and shear stress above a propagating dyke. Surface landsliding, in turn, produces debudderting above the dyke, which can induce an effusive eruption with lava flows within the landslide depression. In this case a balance between newly produced lava and gravity mass removal can occur, or the volcano can grow along the rift if the emitted magma volume is higher than the collapsed material.

From a mechanical point of view, the different behavior of volcanic slopes parallel to and perpendicular to dyke strike is a function of magma forces, their orientation, rock strength, and cumulative dyke-induced displacement.

The combination of along-rift landsliding and normal-to-rift deep-seated lateral collapse produces four main zones of “consumption” and “refilling” on the volcano, alternating with four triangular sectors where the volcano slopes are relatively more stable and the hazard of vent opening is lower.

When volcanic activity ends, cone consumption concentrates along the rift and is favored by fault slip and channelled erosion.
Sheet intrusions and evolution of rifted volcanoes


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