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Volcán Ecuador, Galapágos Islands: erosion as a possible mechanism for the generation of steep-sided basaltic volcanoes

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Abstract. Volcán Ecuador (0°02′ S, 91°35′ W) consists of two strongly contrasting components: the eroded and vegetated remnant of a once-circular main volcano with a probable caldera, and a prominent rift zone extending to the northeast that is neither strongly eroded nor weathered. There are about 20 young-looking flows and vents on this caldera floor but only one on the higher remnant of the main volcano. The southwest half of the main volcano is faulted into the ocean. The main part of Volcán Ecuador possesses steep erosional slopes (average 30–40°) that cut into a sequence of flows that dip radially outward at $<10°$. In contrast, the northeast rift zone consists entirely of young flows and vents. The upper 10 km of the rift zone forms a peninsula about 7.5 km wide that connects Volcán Ecuador to Volcán Wolf. The rift zone bends to the southeast and the lower 8 km is tangential to the coast of Volcán Wolf. The rift zone axis dips away from the northeast edge of the main volcano, and its flanks slope roughly northwest and southeast at $<4°$. The rift zone is the Galápagos structure that most closely resembles a Hawaiian rift zone because it is constructed of lavas from subparallel linear vents, shows evidence of a deep feeder conduit, and has changed its direction to avoid a direct intersection with neighboring Volcán Wolf. The steep erosional slopes extending around the perimeter of the main volcano (except to the southwest where slumping occurred) were probably generated by marine erosion during a prolonged period of eruptive inactivity (perhaps 20000–30000 years). Only a few post-erosional eruptions have taken place at the main volcano in and near what was once the caldera. The entire rift zone postdates the period of prolonged erosion. Using the evidence for prolonged inactivity at Volcán Ecuador, we propose that erosion may have helped to produce steep slopes on the other western Galápagos volcanoes. On these more active volcanoes, however, numerous subsequent eruptions have completely mantled the erosional slopes with lava. The mechanism by which the volcanoes may shut off for long periods of time is unknown, but the fact that the Galápagos hotspot is presently supplying nine active volcanoes suggests that the magma supply at an individual volcano could vary greatly over periods of (tens of?) thousands of years.

Key words: Galápagos – erosion – steep slopes – eruption hiatus – rift zone – magma supply – caldera

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Introduction

The Galápagos Islands (at 89–92°W on the equator; Fig. 1) are oceanic hotspot shield volcanoes. They differ in a number of ways from the better-studied Hawaiian shields. One of the most obvious differences between the two groups is their topographic profiles (e.g. McBirney and Williams 1969; Fig. 2). The gentle slopes (3–6°) of Hawaiian volcanoes during their tholeiitic shield-building stages (e.g. Macdonald et al. 1983; Mark and Moore 1987; Peterson and Moore 1987) are usually cited as typical of basaltic shields, and can be attributed to the low viscosity and/or usually high effusion rate of the erupted lavas which allow flows to travel long distances from their source vents (e.g. Moore and Mark 1992). Additionally, in Hawaii there is only minor pyroclastic activity which might otherwise contribute a large percentage of steep slope-forming material. The steepest subaerial slopes on the active Hawaiian shields occur on the west flank of Mauna Loa and the south flank of Kilauea (Moore and Mark 1992). In both of these locations, normal faults (downdropped towards the ocean) have been identified (or proposed) to account for the steepness (e.g. Swanson et al. 1976; Normark et al. 1979; Lipman et al. 1985; Mark and Moore 1987; Moore and Mark 1992).

In comparison, the active volcanoes on Isla Fernandina (Volcán Fernandina) and Isla Isabela (Volcanes Ecuador, Wolf, Darwin, Alcedo, Sierra Negra, and Cerro Azul) have been described as unusual because

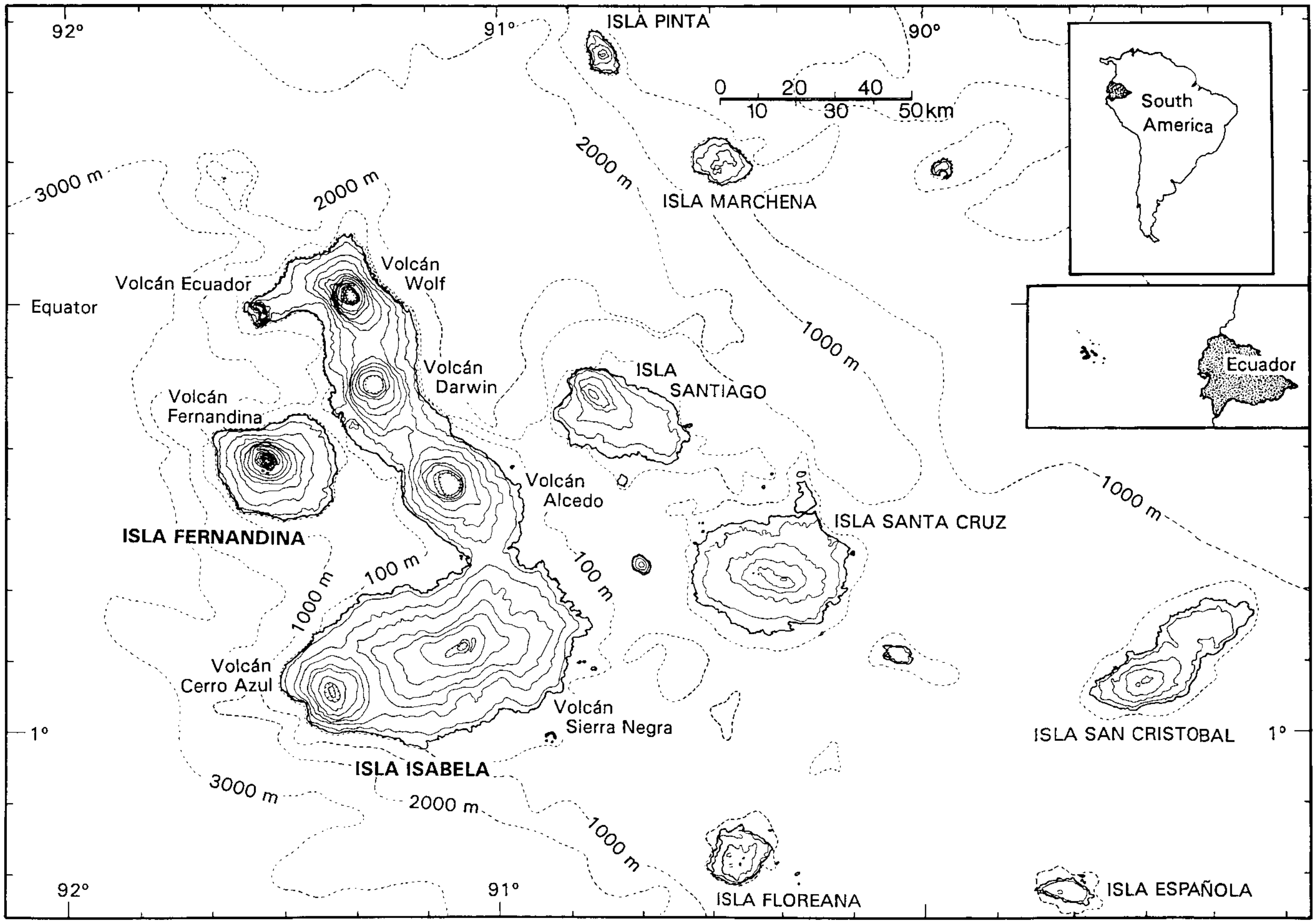

Fig. 1. Location map of the Galápagos islands (adapted from a map published by Libreria Internacional, Quito, Ecuador). Volcán Ecuador forms the northwest-most tip of Isla Isabela. Subaerial contours are at 50 m and then multiples of 200 m; submarine contours (*dashed*) are at 100 m and then multiples of 1000 m

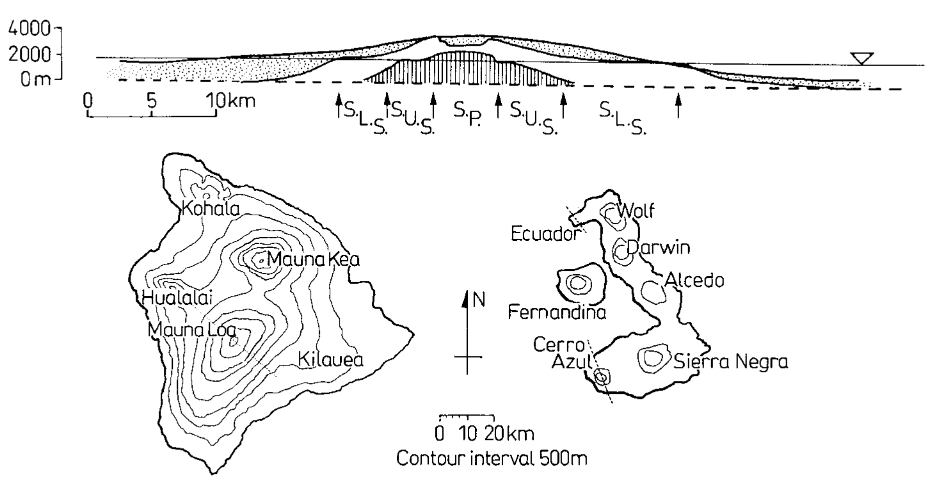

Fig. 2. (top) Profiles without vertical exaggeration across Volcán Cerro Azul, (*unpatterned*; from Nordlie 1973), Volcán Ecuador (*vertical lines*), and part of the subaerial portion of Mauna Loa, Hawaii (*stippled*; caldera not shown). For the Cerro Azul profiles, *S.U.S.* (subaerial upper segment) and *S.L.S.* (subaerial lower segment) are geomorphic sections from Nordlie (1973), and *S.P.* is the summit platform. *Inverted triangle* is Galápagos sea level, *dashed line* is both Hawaii sea level and Galápagos ocean bottom. (bottom) Maps drawn to the same scale of the island of Hawaii and the islands of Fernandina and Isabela. *Dotted lines* indicate traces of topographic profiles

although they are also constructed dominantly of (presumably) low-viscosity tholeiitic basalts, they possess distinctly steeper slopes (20–26° in some areas; McBirney and Williams 1969; Nordlie 1973) than their Hawaiian counterparts (Fig. 2). Such 'distinctive' steep slopes are not found for all profiles of all the active Galápagos volcanoes (Nordlie 1973); there are Galápagos profiles that are steeper than, equal to, or more gradual than those of Hawaiian shields. Nordlie (1973) proposed a nomenclature for the slope sections of Fernandina and Isabela which we have adopted (Fig. 2). Using this nomenclature, it is the steep 'subaerial upper segments' that are enigmatic in comparison to Hawaiian shields.

Several hypotheses have been proposed to explain the formation of steep slopes on the Galápagos shields. Banfield et al. (1956) suggested that the central part of each volcano was preferentially built up by an early period of dominantly pyroclastic eruptions. McBirney and Williams (1969) did not find significant pyroclastic deposits in the caldera walls and suggested that sill complexes were preferentially emplaced in the central regions of the volcanoes, causing them to be updomed. Nordlie (1973) included an episode of updoming in his interpretive chronology of events that led to the present Galápagos volcano shape. Importantly, he also noted that the flows that make up the present surface are not accentuating or creating the steep slopes. Simkin (1972, 1984) suggested that if eruptive fissures near and parallel to the caldera rim preferentially erupted viscous and/or small-volume flows, they could build up a steep-sided annular region near the center of the volcano. By building up this central region while caldera collapse continued, the relatively flat summits could be maintained. Cullen et al. (1987) returned to the inflationary mechanism and suggested that the permanent inflation of magma chambers with different aspect ratios could explain the detailed topographic and caldera-shape differences among the Galápagos volcanoes (although ignoring the fact that the topographic profiles are not radially symmetric). Chadwick and Howard (1991) preferred a constructional model by noting that the steepest slopes are being mantled by flows from arcuate fissures whereas flows that comprise the lower (more gradual) slopes issue from radial vents outboard of the steep slopes. Flows with steep primary dips in the walls of the Alcedo caldera were used by Geist et al. (1994) as evidence of a constructional origin for the steep slopes.

The previous ideas fall into two general categories: (1) preferential construction of the central region by lava and/or pyroclastics; and (2) intrusive updoming of the central region. In this paper we present erosion of the outer flanks as a possible third mechanism. This

Fig. 3. SPOT scene of Volcán Ecuador collected 27 April 1988. Note the obvious albedo and geomorphic differences between the crescent-shaped higher part of the main volcano and the rift zone. Note also that the rift zone is much younger than most of the northwest flank of Volcán Wolf. *CB* and *PVR* are Cabo Berkeley and Punta Vincente Roca, respectively. *Arrows* indicate rift-parallel scarps (see text). *B*: downdropped block (Chadwick and Howard 1991), S_1 and S_2: tops of outermost and innermost arcuate scarps, respectively (McBirney and Williams 1969). The outermost scarp is the proposed caldera wall. *Right angles* indicate three corners of Fig. 9

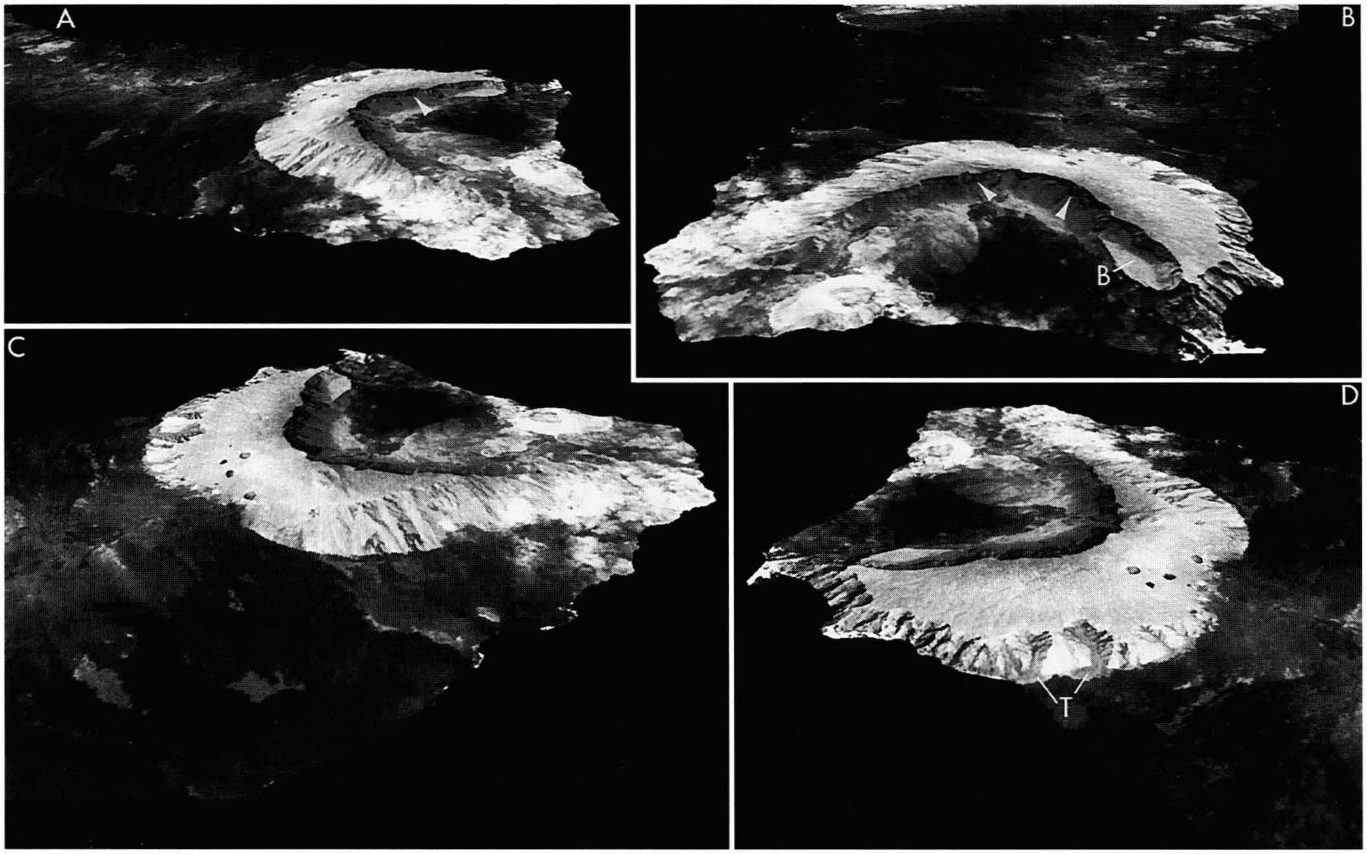

Fig. 4 A–D. Perspective views of Volcán Ecuador produced from the SPOT scene and a DEM derived by digitizing US Defense Mapping Agency map #22XCO22543. The view directions are from: **A** northwest; **B** southwest; **C** northeast; and **D** southeast. The viewpoint elevation in all four is the equivalent of 6000 m, and there is so vertical exaggeration. In **A** and **B**, *arrows* indicate boundary between intact flows and talus in the caldera. In **B**, *B* indicates intra-caldera bench. In **D**, *T* indicates talus from large gullies

hypothesis is suggested by the obviously eroded form of Volcán Ecuador and its similarity to the profiles of the other (more active) western Galápagos volcanoes. Through the analysis of high resolution images from the SPOT and LANDSAT satellites, air photographs, synthetic aperture radar (SAR), and topography data, we investigate the geologic history of Volcán Ecuador and the possible implications for the other Galápagos volcanoes.

Volcán Ecuador

Volcán Ecuador forms a peninsula extending westward from Volcán Wolf at the northern end of Isla Isabela (Figs. 1, 3). It consists of two distinct components. In this paper we will refer to the semi-circular western component as the main volcano, and the linear extension to the northeast as the northeast rift zone. Volcán Ecuador differs from the other western Galápagos volcanoes in a number of ways: (1) it is smaller; (2) it lacks the distinctive bimodal arrangement of radial and arcuate vents (Chadwick and Howard 1991); (3) it shows evidence of an eruptional hiatus (this paper); and (4) no eruptions have been recorded since the archipelago was discovered in 1535 (Simkin et al. 1981; Perry 1984).

Volcán Ecuador was first described by McBirney and Williams (1969) who referred to it as Cape Berkeley Volcano. They suggested that the missing southwestern half of the main volcano had been downfaulted on two concentric faults (Fig. 3). Presumably downdrop on a third fault further west lowered the southwest half of the main volcano below sea level. McBirney and Williams (1969) also noted (without elaborating) that the main volcano had once been separated by ocean from Volcán Wolf. They observed that there is only one youthful scoria cone on the high part of the main volcano, and that it and a group of pit craters are aligned with the northeast rift zone.

Simkin (1984) presented photographs of Volcán Ecuador, and also noted that the ‘primary’ slopes of the main volcano had been eroded. He interpreted that the volcano once had a caldera, and that the outermost (easternmost) of the two arcuate scarps identified by McBirney and Williams (1969) represents the caldera wall (Fig. 3). Chadwick and Howard (1991) cast some doubt on the caldera interpretation. They noted that a large block (‘B’ in Fig. 3) had slumped to its present position and was not the level of a former caldera floor, and perhaps the entire arcuate scarp was the headwall of a large slump or landslide, rather than a (now bisected) caldera. We find evidence (see below)

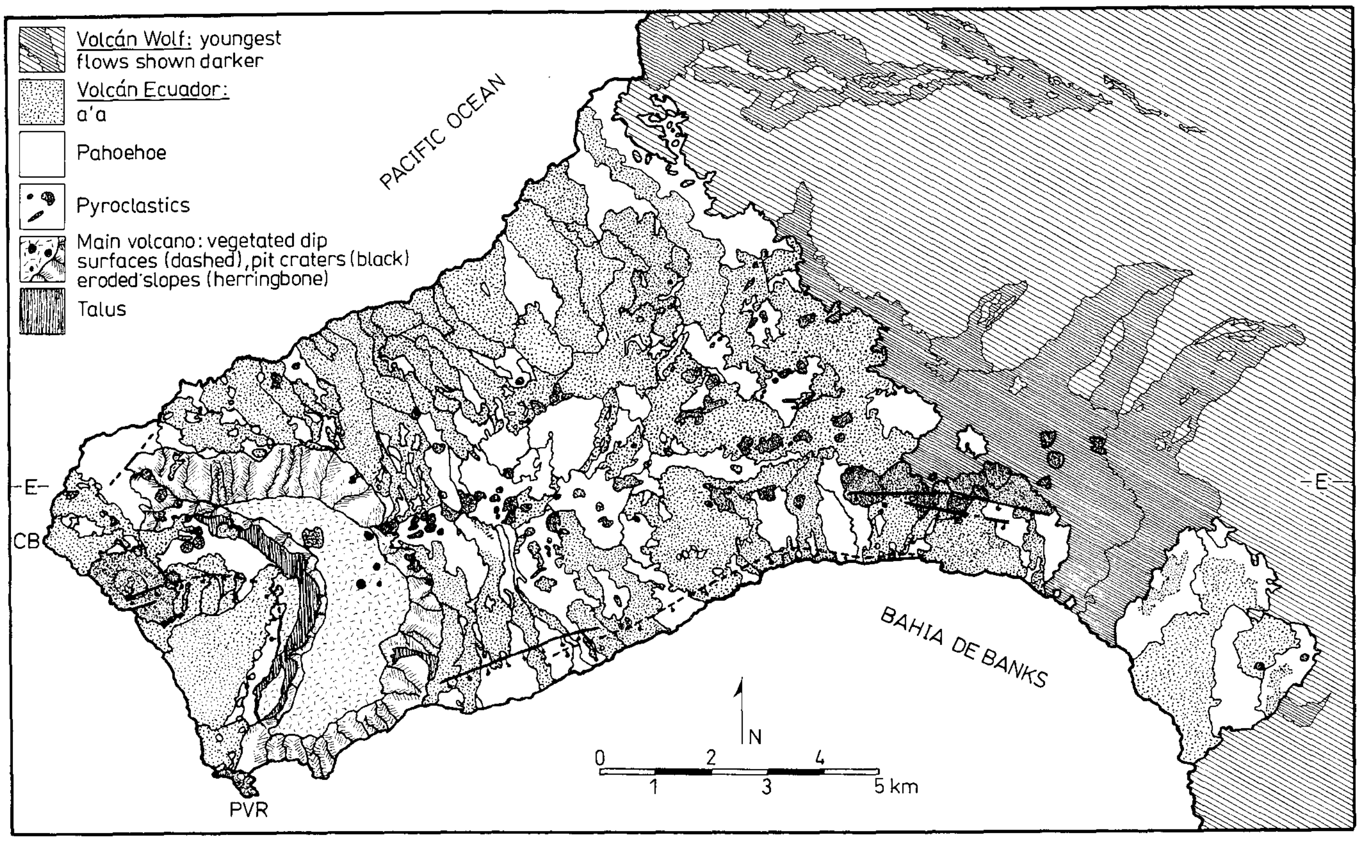

Fig. 5. Geologic map showing surface units determined from the SPOT, LANDSAT TM, and air photo data. *Heavy lines with tik-marks* are inferred faults with the balls on the downdropped side. *CB* is Cabo Berkeley, *PVR* is Punta Vincente Roca, *E* is the equator

that the structure is indeed a caldera, and from here on will refer to it as such.

We present remotely sensed data and maps drawn from them to interpret the structure and history of Volcán Ecuador. We did not conduct field-checks. The data consist of a panchromatic (black and white) SPOT image collected on 29 April 1988 (Fig. 3), which we have combined with a digital elevation model to create Fig. 4, a LANDSAT Thematic Mapper (TM) image taken on 31 May 1985 (not presented here), as well as a number of oblique aerial photographs taken during a NASA research flight in May 1993. The higher spatial resolution (10 m) of the SPOT scene allowed for more accurate mapping (Fig. 5). However, the multispectral nature of the lower-resolution (30 m) LANDSAT scene allowed for differentiation of some flow margins and vegetation in ambiguous areas of the SPOT scene.

The Main Volcano

The two parts of the volcano (the main volcano and the northeast rift zone) are clearly differentiated in the remotely sensed images (Figs. 3, 4). The main volcano is mostly light-toned with a surface consisting of vegetation and/or bare soil and weathered rock. The summit of the main volcano is a plateau bounded on the southwest by the outermost arcuate scarp (the caldera rim). The plateau is quite featureless except for the large scoria cone and pit craters noted by McBirney and Williams (1969; Fig. 6). The top of the plateau slopes gently away from the caldera rim in all directions (Fig. 4) with a dip of ~10°. At a distance of about 1.5 km from the caldera rim (measured radially), the gentle slopes are cut by steep erosional scarps that form the north, east, and south boundaries of the plateau (Fig. 4). These convex-outward scarps (with slopes averaging 30–40°) descend either to the ocean (to the south) or are abutted against by young lava flows (to the east and north). These steep scarps are heavily gullied (Figs. 3, 4, 6), and small deposits of talus have collected at the bases of some of the larger gullies (Figs. 4d, 5). The flows exposed in the top of the cliff section have shallow dips (Fig. 6) that parallel the surface of the plateau and from this we infer their outward dips of ~10°. Lower in the section (Fig. 6) are steeply angled erosional forms that may indicate steeper-dipping flows (K Howard, personal communication) or arcuate dikes (T Simkin, personal communication). Lavas dipping at ~20° are exposed above Punta Vincente Roca (Fig. 7).

The caldera wall has slopes of about 30°, and the top ~1/3 consists of intact lavas whereas the lower 2/3 is talus (e.g. Figs. 4a, b). A single intra-caldera bench ('B' in Figs. 3, 4b) occurs near the southern end of the

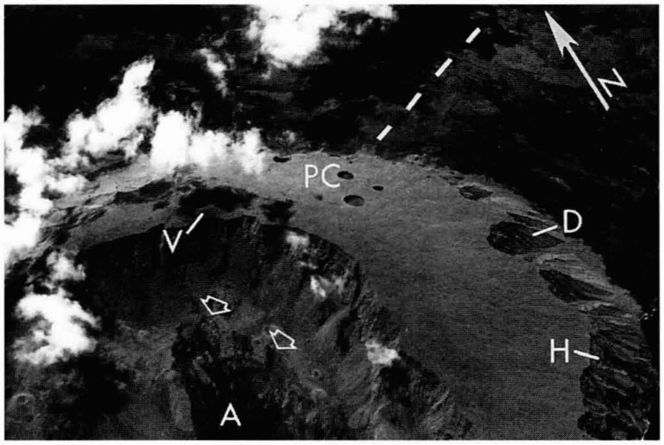

Fig. 6. Oblique photo (view to the NE) showing most of the old part of the main volcano. The one confirmed vent structure on this high part of the volcano is a large cinder cone (*V*), somewhat obscured by the shadow of a cloud. The pit craters (*PC*) fall on a line that extends the rift zone axis (*dashed*) to the edge of the old caldera. Note the numerous cinder cones around the base of the caldera wall and the arcuate source region (*arrows*) for the large a'a flow (*A*) that covers most of the caldera floor. In the lower right corner, note lavas with shallow dips (*H*) exposed near the top of the eroded section. Slightly to the northeast, steeper-dipping patterns (*D*) may be due to colluvial processes or underlying structure (implying steeply dipping flows or dikes)

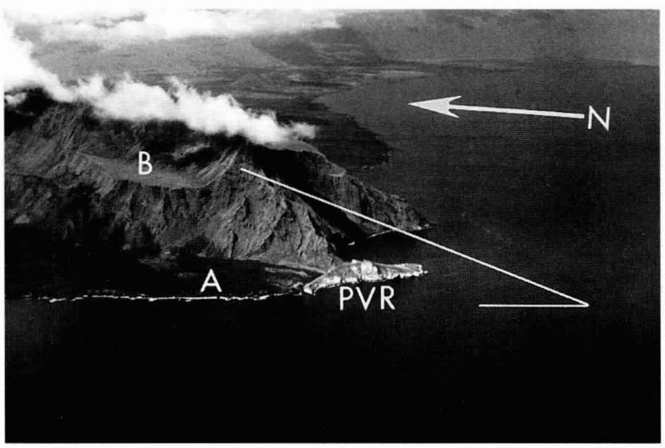

Fig. 7. Oblique photo (view to the NE). Note the flows visible in the eroded caldera wall. They have been projected out to sea level and measured with a protractor, yielding apparent dips of ~20°. The large downdropped block (*B*), Punta Vincente Roca (*PVR*), and young a'a flow (*A*) are labeled

caldera wall, and it has a nearly horizontal top surface about 350 m below the rim. The westernmost (innermost) arcuate scarp of McBirney and Williams (1969) is located at the boundary between a large young flow and vegetated talus and older flows ('S_2' in Fig. 3). Around to the northwest this scarp is less obvious and appears as a slight albedo change within older lavas.

Approximately half of the caldera floor is covered by a single dark a'a flow. This flow and several other dark lavas issued from vents on the two arcuate scarps. Eighteen scoria cones are found within the area of the

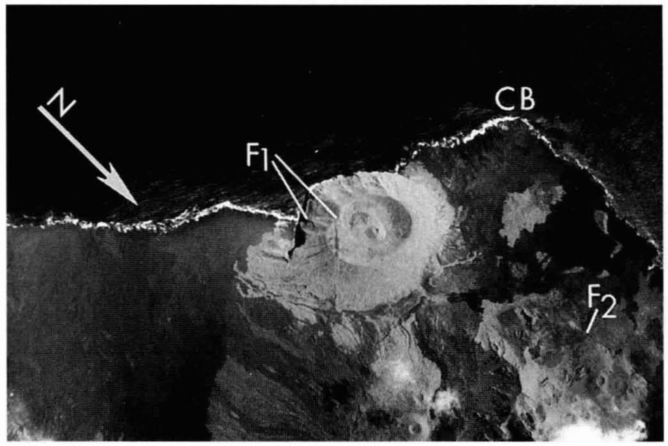

Fig. 8. Near-vertical photo showing Cabo Berkeley (*CB*), and the large tuff cone just to the southeast. Note the numerous dark (a'a) flows to the west of the cone. Note also the lineations (interpreted as faults) that cut the tuff cone (F_1). A lineation within the flows N of Cabo Berkeley (F_2) may be a fault or straight-flow boundary

caldera. They have basal diameters between 200 and 300 m. Two prominent tuff cones are considerably larger than the scoria cones. The northernmost of these (~1500 m in diameter) is about 1 km southeast of Cabo Berkeley and is situated at the point where the innermost arcuate scarp (Fig. 3) would project to the coastline. This cone is cut by two lineations and a cut face within the tephra (Fig. 8), interpreted as faults. A remnant of the second tuff cone forms Punta Vincente Roca (original diameter ~800 m). Its center would have been on an extension of the outhermost arcuate scarp. This observation that the arcuate scarps have a connection to magmatic plumbing leads us to the conclusion that they are caldera walls rather than landslide slip faces.

A number of young flows have built out the coastline to form Cabo Berkeley, and they combine with the westernmost flows of the northeast rift zone to form a continuous low-angle shelf at the base of the northern cliffs. The vents for these flows (Chadwick and Howard 1991) are both inboard and outboard of the old caldera wall. There is a barely discernible northeast-trending lineation in this area (Figs. 5, 8) but we could not determine its nature. According to McBirney and Williams (1969), most of these Cabo Berkeley flows are pahoehoe; however, at least half of them appear to be a'a (based on their low albedos; Fig. 8).

In summary, the high crescent-shaped part of the main volcano is constructed of outwardly dipping lavas, appears not to have been the site of any recent activity (with the possible exception of a single scoria cone), and has been heavily eroded (mostly inward from its periphery). Within what may once have been a caldera, however, recent activity has produced lava flows, scoria cones, and tuff cones from arcuate fissures. All of these recent flows appear to be much younger than any lavas that make up the crescent-shaped high plateau.

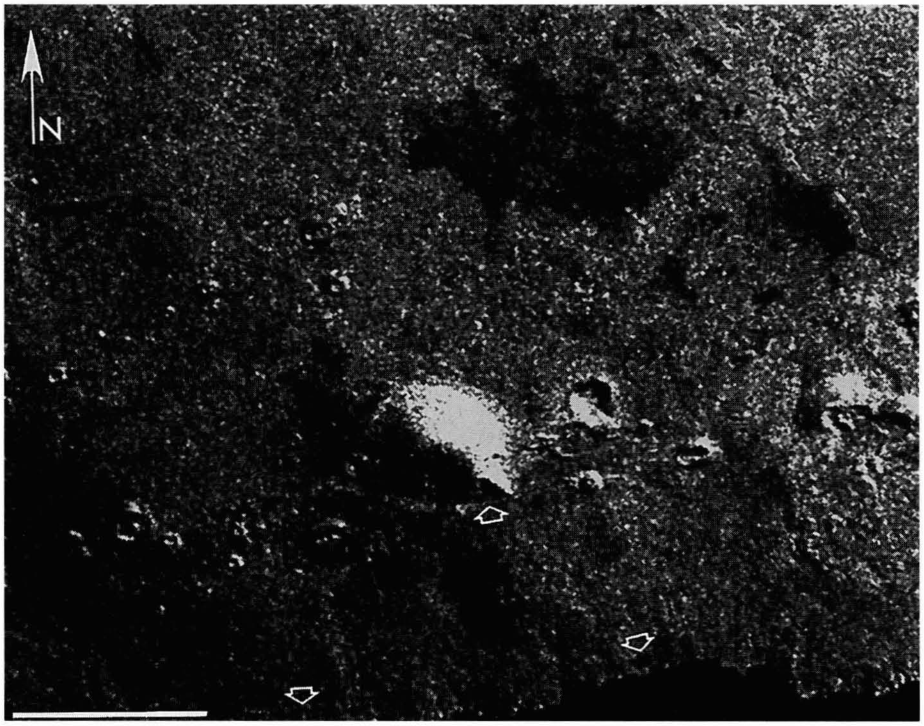

Fig. 9. SAR image (radar illumination from the northeast) showing faint linear features that may be tensional fissures (*arrows*). As noted in the text these do not appear to be associated with eruptive structures (although there are many such structures nearby). If suitably oriented, a topographic feature such as a fissure can often produce a signal in a radar image even if it is significantly smaller than the resolution of the image. *Scale bar* is 1 km, see Fig. 3 for location. TOPSAR data courtesy of Howard Zebker, Jet Propulsion Laboratory

The Northeast Rift Zone

The northeast rift zone extends from the northeast flank of the main volcano toward Volcán Wolf. As defined by the center line between the north and south coasts, it has an azimuth of ~60° but most vents fall along an azimuth of ~75°. The rift zone forms a broad ridge about 7.5 km wide subaerially with slopes of approximately 4°. The crest of the rift plunges to the northeast and is nearer to the south coast than to the north coast. About 10 km form the main volcano, the rift changes its orientation and narrows considerably, extending another 8 km to the east-southeast (Figs. 3, 5). The change in orientation occurs about 2 km from the contact with Volcán Wolf lavas.

Numerous linear vents and fissures parallel the general trend of the northeast rift zone, changing to a more east-southeast orientation where the rift zone similarly changes (Fig. 5; Chadwick and Howard 1991; Munro 1992). However, no vents are visible (nor were they mapped by Chadwick and Howard 1991) beyond about 4 km of the bend. We have interpreted faint linear albedo changes along the south coast to possibly be lava-mantled normal faults downdropped on the oceanward side, and they parallel the trends of nearby eruptive fissures. The SAR data (at 10 m resolution and collected by the NASA TOPSAR instrument in May 1993) provides additional evidence of cross-rift tension. In the SAR image (Fig. 9), faint lines indicating radar-facing structures can be found along the coast and in the vicinity of a large group of cinder cones. These linear features do not have a positive topographic form (they are not spatter ramparts), and thus are probably open tensional fractures.

Although there are a few vegetated kipukas (inliers), most of the flows and cones that make up the surface of the rift zone have a young appearance, especially when compared to the high part of the main volcano. Based on our mapping (Fig. 5) we estimate that there are almost 100 separate flows. On the north flank of the rift there are so many dark flows that it was difficult to differentiate them in the satellite images. About 75% of the flows are a'a. Their average lengths and areas are ~3 km and ~2 km^2, respectively, but because most entered the ocean these are minimum values.

The flows of the northeast rift are much younger looking than most of the north and northwest flanks of Volcán Wolf. However, the very youngest flows from the poorly defined northwest rift of Volcán Wolf (Munro 1992) postdate the northeast rift of Volcán Ecuador and a few of these managed to flow completely across it on their way to the coast after first being deflected to the southeast (Figs. 3, 5). This deflection and the albedo differences indicate the presence of the (topographically subtle) Volcán Ecuador rift zone out to 8 km from the bend. Even though Volcán Ecuador has not erupted historically, the northeast rift zone does appear to have been active during the past few thousand years.

The steep erosional cliffs of the main volcano are <100 m high adjacent to the axis of the rift zone. To the north and south of this point, young lavas bank against the base of the steep erosional scarps. Where the steep scarps drop to the ocean at the south coast, they are nearly 500 m high. The pit craters (Fig. 10) on the main volcano in the vicinity of its intersection with the rift zone contain no young lava; they are vegetated and appear to be purely collapse-derived.

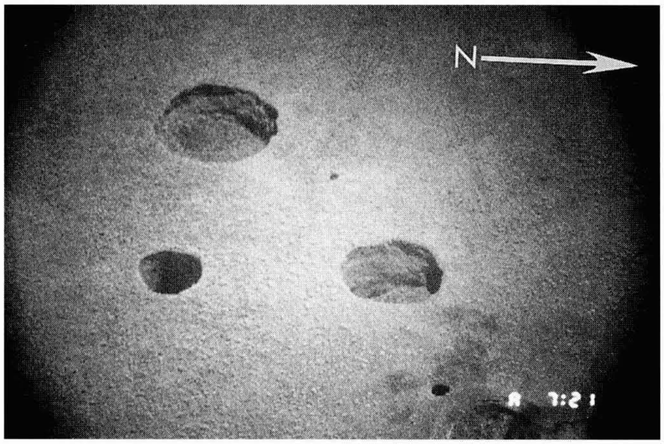

Fig. 10. Oblique photo showing pit craters formed in the main volcano (view is to the southwest). Note that these are not the sources of erupted material: rather they are derived by collapse. The largest crater (upper left) is 180 m across

Although Galápagos volcanoes have previously been noted to lack rift zones similar to those of the Hawaiian shields (Nakamura 1977, 1982; Chadwick and Howard 1991; Munro 1992), the northeast rift of Volcán Ecuador is a structure that strongly resembles them: (1) it is linear; (2) it has closely spaced lines of vents with relatively consistent azimuths; (3) there are normal faults and tensional fissures that strike parallel to the rift azimuth; (4) there is evidence of a deep conduit (the pit craters; Walker 1988); (5) there are numerous high-albedo flows, interpreted to be pahoehoe; and (6) its orientation veers southeast, apparently in response to the volcano-volcano boundary with Volcán Wolf (e.g. Fiske and Jackson 1972).

Eruptive History

The two major parts of Volcán Ecuador are unusual in their juxtaposition of different relative ages. Based on the images and descriptions, we propose a general eruptive history for Volcán Ecuador. Initially, a nearly circular shield with slopes of around 10° was constructed. It had a caldera about 4 km in diameter. The volcano then ceased erupting and erosion, both catastrophic and gradual, took over as the dominant geologic process. Wave erosion worked its way inward forming the steep cliffs but preserving the uppermost surfaces (the high plateau). The presence of steep slopes around the whole convex margin of the main volcano (e.g. Fig. 4c, d) indicates that the volcano was not part of Isla Isabela during this period. The southwest half of the volcano slumped or was downfaulted into the ocean but it is nor clear whether this occurred before, during, or after the period of prolonged erosion.

Most recently, the volcano became active again. A few eruptions took place within the remnant of the old caldera. Vents were aligned parallel to, and on extensions of, the caldera-bounding faults and magma was probably utilizing them to reach the surface. At the oceanward ends of these faults, magma interacted with water to produce the large tuff cone southeast of Cabo Berkeley as well as Punta Vincente Roca. A few eruptions took place around the northern base of the now steep main volcano, but most of the activity since the long erosional period formed the northeast rift zone. The rift zone partially buried the erosional scarps and eventually extended to Volcán Wolf, ending Volcán Ecuador's isolation as a separate island. During the formation of the Volcán Ecuador northeast rift, the northwest part of Volcán Wolf was relatively inactive. This produced the strong contrast between its smooth vegetated slopes and the surface of the Volcán Ecuador northeast rift (Fig. 3). As noted above, however, the few most recent flows from this northwest part of Volcán Wolf postdate the northeast rift of Volcán Ecuador.

Magma plumbing at Volcán Ecuador

The northeast rift of Volcán Ecuador is aligned along one of the two directions of 'Darwinian' lineaments (e.g. Darwin 1896 p 131; McBirney and Williams 1969; Nordlie 1973; Chadwick and Howard 1991). Two sets of these lineaments occur in roughly northwest-southeast and northeast-southwest directions, and they are defined by preferred locations of islands, orientations of elongate islands, bathymetric features, and (in some instances) vent concentrations. Why the higher part of the main volcano never reactivated is not clear. Perhaps magmatic pressure was unable to intrude dikes high into the edifice, or the loss of the southwest half disturbed the plumbing in some way so that the higher part of the volcano was no longer supplied with magma.

Magma feeding the northeast rift zone, however, did leave its mark as it passed through the high plateau as evidenced by the pit craters. Pit craters along the rift zones of Hawaiian volcanoes (e.g. Macdonald et al. 1983) develop by upward stoping of a void formed over a long-lived magma conduit (Walker 1988). A warm conduit is required if a rift zone is going to support low discharge-rate pahoehoe eruptions (Rowland and Walker 1990), and the presence of pahoehoe flows up to 3 km long at the farthest end of the Volcán Ecuador northeast rift zone is further evidence of such a conduit.

An added complication to rift zones in Hawaii occurs where the stress regimes of adjacent volcanoes interact with each other (Fiske and Jackson 1972). A younger volcano growing through the flank of an older neighbor will be forced to adopt a pre-existing stress orientation as it develops its own rift zones. Thus rift zones on adjacent volcanoes tend to become parallel. Volcán Wolf has a poorly developed northwest-trending rift zone evidenced by a slight topographic ridge and a concentration of old eruptive fissures (Chadwick and Howard 1991; Munro 1992). Dikes propagating to the farthest end of the Volcán Ecuador rift zone intrude through the flanks of Volcán Wolf, and by turning to the southeast they can most easily accommodate the slope-parallel extensions stress.

The generation of steep-sided Galápagos volcanoes

Previous models

Unequivocal support for any previously proposed model for the generation of steep slopes is lacking, possibly because extensive field observations have thus far been limited by the remote nature of the Galápagos volcanoes. Those ideas that rely on intrusive updoming require that hundreds of meters of permanent uplift take place and that it be strongly concentrated in the central part of the volcano. Cullen et al. (1987) cite two recent episodes of rapid uplift that occurred near the coast of Fernandina and the coastal boundary of Darwin and Alcedo (Couffer 1956) as support for their intrusive model. However, both of these episodes took place out on the volcano flanks and, if anything, would tend to act against the production of centrally located steep slopes. The recent development of a cryptodome at Usu Volcano in Japan has also been suggested as a possible Galápagos analog (D Swanson personal communication 1992). However, the Usu cryptodome reached eventual dimensions of 160 m in height and 700 m in width (McClelland et al. 1989), both at least an order of magnitude smaller than what would be required for a Galápagos volcano. The Usu cryptodome probably also consisted of viscous magma compared to the fluid basalt of the Galápagos; surface domes at Usu are rhyolite and dacite (Katsui et al. 1985).

The model of Cullen et al. (1987) requires single intrusions to fill magma chambers having horizontal and vertical dimensions of around 10 and 3 km, respectively. The pioneering work of Fiske and Kinoshita (1969) has shown that basaltic magma chambers are almost certainly not single voids but rather interconnected plexes of smaller chambers. It is not clear how such a complication would affect the modeling of Cullen et al. (1987), but they did note that multiple sill-like intrusions might be more likely. Experience at Kilauea and Mauna Loa has also shown that intrusive events are relatively small and, importantly, separated by periods of deflation (accompanying magma excursions); the net inflation and deflation is near zero. Furthermore, Geist et al. (1994) have shown that the lowest flank flows exposed in the caldera of Volcán Alcedo have not been tilted, even though they dip outwards at up to 40°. Thus the emplacement of the entire 250 m section of caldera wall lavas exposed at Volcán Alcedo was unaccompanied by permanent inflation.

The scenarios that involve preferential build-up are not supported by large amounts of pyroclastic material exposed in the caldera walls (McBirney and Williams 1969) or any obviously more viscous flows occurring around the caldera. However, Simkin (1972) and Naumann and Geist (1993) showed that in some instances there is a concentration of short flows on the steeper slopes. Most of these are fed by arcuate fissures which are perpendicular to the slope and therefore generate many short flows from numerous points along their length rather than a few larger flows. This mechanism will help to maintain steep slopes only if the flows come to rest before the subaerial lower segment. The diagram to illustrate this mechanism, however, (Fig. 2 of Simkin 1972) shows such hypothetical flows reaching the base of the steep slope and adding considerably more lava to the near-horizontal subaerial lower segments than to the steep subaerial upper segments. This same situation was observed by Nordlie (1973) to be occurring with the actual flows and for this reason he suggested that rather than creating or enhancing the steep slopes, the present activity of the volcanoes is obscuring them (Fig. 11).

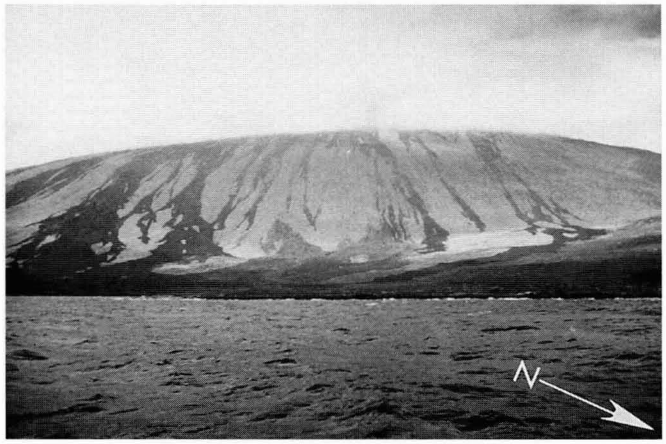

Fig. 11. Photograph of the northeastern slope of Volcán Wolf showing that although some present-day flows are restricted to the steeper slopes, most only mantle the steep slopes and pool at their bases

Erosion

Wave erosion during a long eruptive hiatus was responsible for the formation of steep slopes at Volcán Ecuador. Figure 12 shows a cross section of Volcán Ecuador. The sub-horizontal platforms that extend from the bases of the north and south steep slopes strongly resemble wave-cut terraces. The fact that the northern of the two is above sea level can be explained by this area having been resurfaced by young lavas. To the south where no such flows were emplaced, the terrace is a few tens of meters below sea level. Seaward of these terraces, the submarine slopes drop off more steeply.

The scarcity of isotopic age dates on western Galápagos lavas and the difficulty of determining erosion rates currently leave unconstrained the duration of the proposed erosional period at Volcán Ecuador. In Hawaii, it has been estimated that basalt shorelines retreat at an average rate of 5 cm/year (cited in Macdonald et al. 1983). Reconstruction of the pre-erosional profile of Volcán Ecuador (Fig. 12a) yields a shoreline retreat of 1–1.5 km, which at the Hawaiian coastal re-

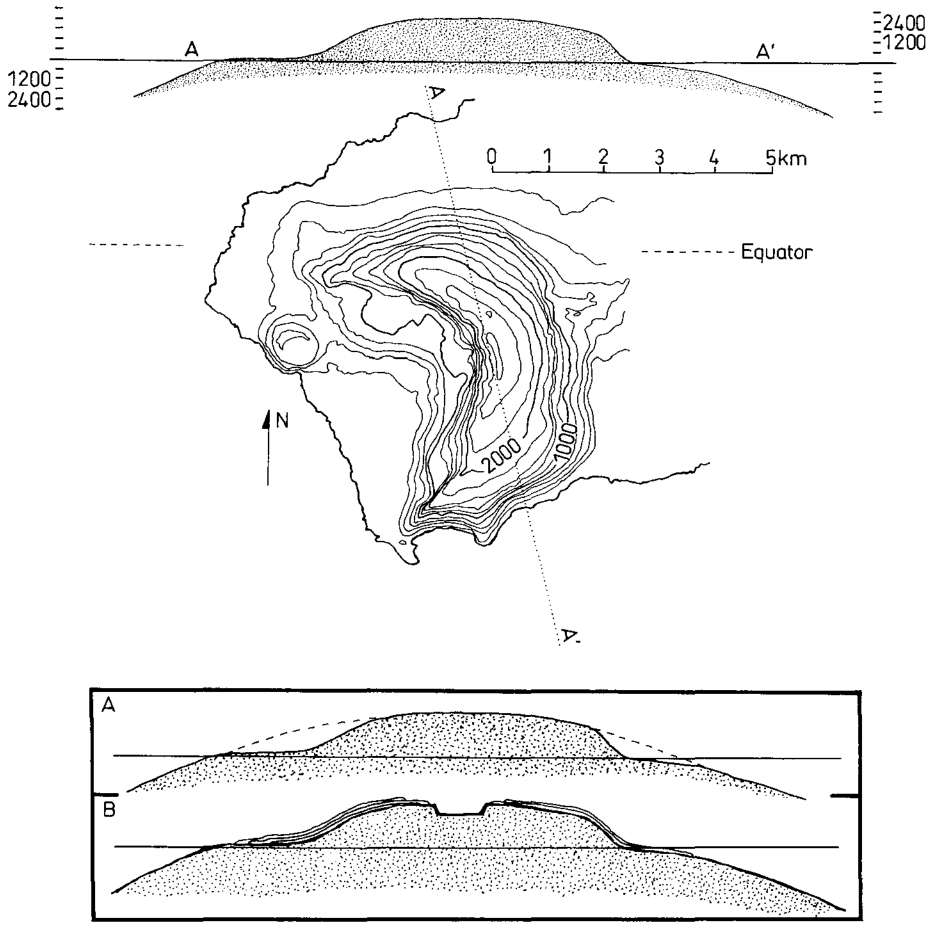

Fig. 12. Cross-section (elevations in feet: 1200 feet ≅366 m) and topographic map (contour interval 200 feet (61 m) of Volcán Ecuador (from US Defense Mapping Agency map #22XCO22543 (1985)). Note that the north and south slopes are equally steep despite their facing into different ocean environments. Note also the shallow submarine bench on the south side that resembles a wave-cut terrace. Presumably a similar shelf on the north side underlies young lavas. Box: **A** extension of the upper surface of the main volcano (*dashed*) to show the amount of material proposed to have been removed by erosion; **B** same profile with caldera and hypothetical lava flows added to mantle erosional slopes (compare with profile of Cerro Azul in Fig. 2)

treat rate would require ~27000 years to accomplish. The total volume that we propose was removed (see Appendix) was roughly 7.5 km^3 (assuming it occurred prior to the large-scale westward downfaulting). This converts to a removal rate of about 3×10^{-4} km^3/year. The submarine flanks of Volcán Ecuador have a surface area of ~320 km^2, and the erosion calculations yield a layer about 25 m thick deposited at a rate of about 1 mm/year.

We propose that perhaps the steep slopes of the other western Galápagos volcanoes were also generated during extended periods of eruptive hiatus and marine erosion. On these volcanoes, however, the return to life since then has been much more vigorous than at Volcán Ecuador (where it was confined almost entirely to the single rift zone). An erosional model that could generate steep-sided Galápagos volcanoes is presented in Fig. 13. Initially, each volcano has gradual slopes more typical of basalt shields. It then stops erupting and undergoes a long period of mostly marine erosion. After the erosion, renewed eruptions completely mantle the erosional slopes. Nordlie (1973) stated that with time, outer ... segments lengthen and flatten..., and many of the lava flows being erupted at present are merely mantling the steep slopes (Fig. 11). Draping the profile of Volcán Ecuador with a hypothetical layer of lava flows produces the classic Galápagos volcano shape (Fig. 12b).

In the case of a Galápagos volcano that is not an isolated island, the steep slopes would form only on the flanks where wave action could attack, leaving an asymmetric distribution of steep slopes. Such asymmetry was observed by Nordlie (1973). The concentrations of radial vents (McBirney and Williams 1969; Nordlie 1973; Munro and Mouginis-Mark 1990; Chadwick and Howard 1991; Munro 1992) with their accompanying excesses of erupted material show the least evidence (if any) of underlying steep slopes. If these rifts were present prior to the proposed erosional period they may have been massive enough to maintain their ridge-like quality throughout the hiatus. Alternatively, they may have formed since the erosion (as did the rift zone of Volcán Ecuador) and have completely buried the underlying erosional slopes. It is interesting to note that (particularly on Isla Isabela), the steep slopes tend to be those which do not face a volcanic neighbor. For example, Volcán Darwin, with Volcán Wolf to the north and Volcán Alcedo to the south, possesses its steepest slopes to the east and west. The more gentle slopes of the isthmuses betwen volcanoes may indicate that erosion did not attack from these directions, possibly because the volcanoes were not completely separated by ocean. The poorly developed rift zones also tend to be somewhat aligned with these isthmuses (Chadwick and Howard 1991). Fiske and Jackson (1972) proposed that once a topographic ridge

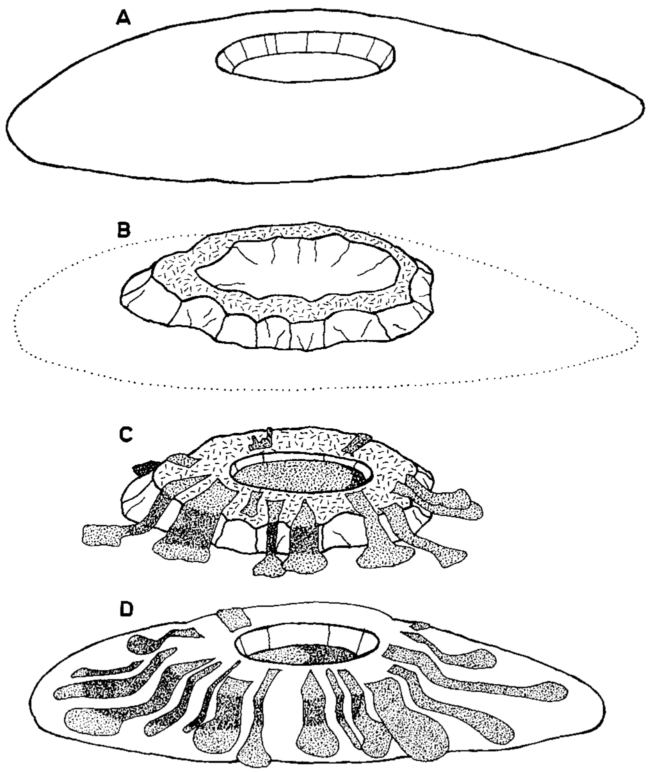

Fig. 13 A–D. Diagrams illustrating the erosional hypothesis presented in this paper. **A** A typical shield volcano grows. The low viscosity of the basalt lavas yields low-angle slopes. **B** The volcano shuts off and wave erosion begins to cut inward from the coastline (original surface *patterned* and original coastline *dotted*). Large-scale slumping of half the volcano at this point would produce a structure similar to the main volcano of Volcán Ecuador. **C** Volcanic activity starts again. Lava flows (*stippled*) mantle the steep erosional slopes and build the nearly horizontal subaerial lower segment (Nordlie 1973) atop a wave-cut terrace. **D** The volcano as seen today. Evidence of the long erosional event is seen only in the steep slopes now mantled by recent lavas flows (*stippled*)

forms it becomes a favorable site for intrusions, and the fact that these isthmuses might have survived the erosion as ridges may have made them more favorable sites for dike injection after the hiatus.

Volcán Sierra Negra is the most similar in profile and size to Hawaiian shields and has possibly erupted enough since its proposed resumption of activity so that any erosion-generated slopes are almost completely hidden. An alternative is that Sierra Negra did not go through an erosional stage at all. In its particular position on the Galápagos platform, it is almost completely surrounded by shallow water and even if its eruption rate diminished, wave energy to attack its flanks might have been too low for noticeable effects.

A number of problems exist with applying the erosional processes that have obviously taken place on Volcán Ecuador to the other western Galápagos shields. Probably the greatest is the size difference (Figs. 1, 2). The much larger volcanoes would require a considerably longer time period to allow erosion to cut what would have to be nearly 1000 m high cliffs, although (as noted above) these need not necessarily have extended around the entire circumference of each edifice. We also do not have an obvious reason of why the volcanoes would switch off and on. However, on a much shorter (and less significant) scale, strong activity at either Mauna Loa or Kilauea seems to correlate with quiescence at the other (e.g. Macdonald and Abbot 1970; Klein 1982). Additionally, the Ninole Hills (e.g. Macdonald et al. 1983; Lockwood and Lipman 1987; Lipman et al. 1990) indicate that at least part of Mauna Loa received no erupted lava and underwent considerable erosion. Finally, we are proposing a model that obscures evidence of its own earlier stages. Future fieldwork will concentrate on evaluating our hypothesis.

Conclusions

We have used remotely sensed data to examine the structure of Volcán Ecuador and interpret that the striking difference in appearance between its main volcano and its rift zone is due to a long period of eruptional hiatus and erosion that separated their formation. The Volcán Ecuador rift zone closely resembles rift zones of the active Hawaiian shield volcanoes.

Earlier hypotheses presented to explain the steep geomorphic profiles of Galápagos volcanoes have required preferential construction of the volcano centers by inflation or eruption. The inflational hypotheses do not fit well with what is known about better-studied basalt volcanoes such as those in Hawaii, namely the unlikelihood of hundreds of meters of permanent intrusive updoming, and the lack of radial symmetry of the volcanoes. The models of preferred central construction by lava flows are probably more realistic, and detailed mapping of the flows will help to constrain this idea (e.g. Naumann and Geist 1993). By examining the eroded form of Volcán Ecuador, we have developed an alternative to construction, namely that the subaerial outer margins of formerly gently sloping shields were preferentially removed by marine erosion during an eruptive hiatus of probably a few tens of thousands of years. On the western Galápagos volcanoes other than Volcán Ecuador, we infer that eruptions have since coated the erosional slopes with lava to form the characteristic Galápagos volcano profiles. Poorly developed rift zones either survived the erosion as ridges, formed later, or persisted throughout; they are the major reason for the lack of radial topographic symmetry.

Our model thus calls upon an eruptive hiatus marked by extensive marine erosion for parts or all of each of the western Galápagos volcanoes. The cause of the eruptive hiatus can only be speculated on but it may have to do with the inability of the Galápagos hotspot to supply the many eruptive centers at constant rates. Nine of the Galápagos volcanoes have erupted within the past 200 years (Simkin et al. 1981; McClelland et al. 1989; Global Volcanism Network 1991), and there does not seem to be a clear age progression within these active western volcanoes. It may be that without the controlling effect of fast plate motion (as in

Hawaii), the magma supply is free to become established in different locations under the influence of variable (and undefined) factors. This process could be envisioned to include the selective (and perhaps recurring) lack of supply to one or more eruptive centers for periods of time long enough for the proposed erosional episodes. Dating of lavas on the flanks and within the caldera walls of Galápagos volcanoes might be a way to test our hypothesis, including whether the volcanoes shut off at the same time or sequentially.

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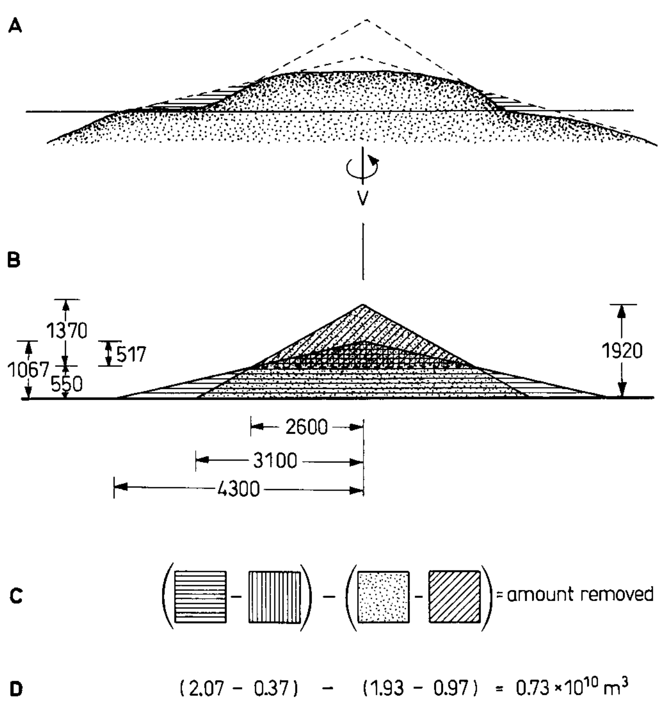
Appendix: Estimation of the volume removed by erosion at Volcán Ecuador



The volcano was modeled as a series of overlapping cones with dimensions derived from the topographic profile in Fig. 12. In A, the dashed lines extend the slopes of the topographic profile to show the construction of the cones formed by rotating around the vertical axis *V*. It is the *horizontally lined* portion (also rotated around the axis V) that we wish to determine the volume of. In B, the cones have been given different patterns (dimensions are in meters). In C, the patterns are presented as an equation, and in D, volumes of the cones (in units of $10^{10}\,m^3$) are inserted to reach the final volume removed of $0.73 \times 10^{10}\,m^3$.

If the downdropping of the southwest half of the main volcano took place before any of this erosion (see text), then the volumes are all halved.

References

Banfield AF, Behere CH Jr, StClair D (1956) Geology of Isabela (Albermarle) island. Geol Soc Am Bull 67:215–234

Chadwick WW, Howard KA (1991) The pattern of circumferential and radial eruptive fissures on the volcanoes of Fernandina and Isabela islands, Galapagos. Bull Volcanol 53:257–275

Couffer JC (1956) The disappearance of Urvina Bay. Natural History 65:378–383

Cullen AB, McBirney AR, Rodgers RD (1987) Structural controls on the morphology of Galapagos shields. J Volcanol Geotherm Res 34:143–151

Darwin C (1896) Geologic observations on the Volcanic Islands and parts of South America visited during the voyage of H.M.S. "Beagle" (third ed.). D Appleton, New York, 648 pp

Fiske RS, Jackson ED (1972) Orientation and growth of Hawaiian volcanic rifts: the effect of regional structure and gravitational stresses. Proc R Soc London A-329:299–326

Fiske RS, Kinoshita WT (1969) Inflation of Kilauea volcano prior to its 1967–1968 eruption. Science 165:341–349

Geist D, Howard K, Jellinek AM, Rayder S (1994) The Volcanic history of Volcán Alcedo, Galápagos Archipelago: A case study of rhyolitic oceanic volcanism. Bull Volcanol 56:243–260

Global Volcanism Network (1991) Reports for August and September 1991, pp 15 and 4, respectively

Katsui Y, Komuro H, Uda T (1985) Development of faults and growth of Usu-shinsan cryptodome in 1977–1982 at Usu Volcano, North Japan. J Faculty Science Hokkaido Univ. ser IV: Geology and Mineralogy 21:339–362

Klein FW (1982) Patterns of historical eruptions at Hawaiian volcanoes. J Volcanol Geotherm Res 12:1–16

Lipman PW, Lockwood JP, Okamura RT, Swanson DA, Yamashita KM (1985) Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea volcano, Hawaii. US Geol Surv Prof Pap 1276:45 pp

Lipman PW, Rhodes JM, Dalrymple GB (1990) The Ninole basalt – implications for the structural evolution of Mauna Loa volcano, Hawaii. Bull Volcanol 53:1–19

Lockwood JP, Lipman PW (1987) Holocene eruptive history of Manna Loa Volcano. US Geol Surv Prof Pap 1350:509–535

Macdonald GA, Abbot AT (1970) Volcanoes in the sea: the geology of Hawaii. University of Hawaii Press, Honolulu, 441 pp

Macdonald GA, Abbot AT, Peterson FL (1983) Volcanoes in the sea: the geology of Hawaii (2nd ed). University of Hawaii Press, Honolulu, 517 pp

Mark RK, Moore JG (1987) Slopes of the Hawaiian Ridge. US Geol Surv Prof Pap 1350:101–107

McBirney AR, Williams H (1969) Geology and petrology of the Galápagos islands. Geol Soc Am Mem 118:197 pp

McClelland L, Simkin T, Summers M, Nielson E, Stein TC (1989) Global volcanism 1975–1985. Prentice Hall, Englewood Cliffs, 655 pp

Moore JG, Mark RK (1992) Morphology of the island of Hawaii. GSA Today 2:257–262
Munro DC (1992) Applications of remotely-sensed data to studies of volcanism in the Galapagos Islands. PhD Dissertation, University of Hawaii, 306 pp
Munro DC, Mouginis-Mark PJ (1990) Eruptive patterns and structure of Isla Fernandina, Galapagos islands from SPOT-1 HRV and large format camera images. Int J Remote Sensing 11:1501–1509
Nakamura K (1977) Volcanoes as possible indicators of tectonic stress orientation – principle and proposal. J Volcanol Geotherm Res 2:1–16
Nakamura K (1982) Why do long rift zones develop better in Hawaiian volcanoes – a possible role of thick oceanic sediments. Bull Volcanol Soc Japan 25:255–267
Naumann T, Geist D (1993) Morphologic development of Galapagos shield volcanoes: evidence from Cerro Azul, Isabela Island. EOS 74: 43 (supplement):643
Nordlie BE (1973) Morphology and structure of the western Galápagos volcanoes and a model for their origin. Geol Soc Am Bull 84:2931–2956
Normark WR, Lipman PW, Moore JG (1979) Regional slump structures on the west flank of Mauna Lao volcano, Hawaii. Hawaii Symp on Intraplate Volcanism and Submarine Volcanism (Hilo, Hawaii), Abstracts with programs, p 72
Perry R (1984) The islands and their history. In: Perry (ed) Galapagos (Key Environments). Oxford, Pergamon Press:1–14
Peterson DW, Moore RB (1987) Geologic history and evolution of geologic concepts, island of Hawaii. US Geol Surv Prof Pap 1350:149–189
Rowland SK, Walker GPL (1990) Pahoehoe and a'a in Hawaii: volumetric flow rate controls the lava structure. Bull Volcanol 52:615–628
Simkin T (1972) Origin of some flat-topped volcanoes and guyots. Geol Soc Am Mem 132:183–193
Simkin T (1984) Geology of Galápagos islands. In: Perry (ed) Galápagos (Key Environments). Oxford, Pergamon Press:15–41
Simkin T, Siebert L, McClelland L, Bridge D, Newhall C, Latter JH (1981) Volcanoes of the world. Hutchinson Ross, Stroudsberg, 232 pp
Swanson DA, Duffield WA, Fiske RS (1976) Displacement of the south flank of Kilauea volcano: The result of forceful intrusion of magma into the rift zones. US Geol Surv Prof Pap 963:30 pp
Walker GPL (1988) Three Hawaiian calderas: an origin through loading by shallow intrusions? J Geophys Res 93:14773–14784

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