Marine Geophysical Researches (2003) 24: 59–90
DOI 10.1007/s11001-004-1162-6

Comparison of volcanic rifts on La Palma and El Hierro, Canary Islands and the Island of Hawaii

J. Acosta[1], E. Uchupi[2], D. Smith[2], A. Muñoz[1], P. Herranz[1], C. Palomo[1], P. Llanes[3], M. Ballesteros[1] & ZEE Working Group[3]
[1]*Instituto Español de Oceanografía. Grupo de Cartografía Multihaz. Corazón de María, 8, 28002 Madrid*
[2]*Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA*
[3]*A. Carbó, A. Muñoz-Martín, Univ. Complutense, Madrid; J. Martín-Dávila, M. Catalán and J.A. Marín, Real Observatorio de la Armada. S. Fernando, Cádiz; F. Pérez-Carrillo, C. Maté, Instituto Hidrográfico de la Marina, Cádiz*

Key words: Canary Islands, Hawaii, La Palma and El Hierro, multibeam bathymetry, volcanic rifts

Abstract

The meso-scale (km) morphology of the well-studied volcanic rift zones on the Island of Hawaii is compared to the morphology of the lesser known rift zones of La Palma and El Hierro, Canary Islands. We find that there are both differences and similarities in their morphologic characteristics. In general, the rift zones on La Palma and El Hierro are shorter (a few tens of km in length) than those on Hawaii (ranging up to >100 km in length), perhaps reflecting both magma supply and composition. Many of the rift zones on Hawaii have well defined axial zones, both on-and offshore. In contrast, the rift zones on La Palma and El Hierro display various geometries ranging from linear ridges having smooth to irregular crests to structures with a broad fan-like morphology in plan view. The pronounced fanning may be a reflection of: 1) the stress field within the rift being insufficient to trap dikes within a narrow region, 2) dike injection and volcanism shifting laterally through time, 3) volcanoes building nearly one atop of another in the Canary Islands, superimposing the stress field of one structure on the other and thus yielding a more complex distribution of gravitational stresses, and 4) low rate of magma supply producing low magma pressures and thus randomly oriented dike injections. Irregularities and curvature along the axes of the rifts on La Palma and El Hierro may be a reflection of differences in the rate of magma production. Unlike the volcanoes on the Island of Hawaii there may be insufficient volumes of lavas erupted on La Palma and El Hierro to smooth out irregularities. The superposition of rifts from different volcanoes may also add to topographic irregularities in the Canary Islands, especially if eruption rates are low.

Introduction

Of all the oceanic island chains inferred to have been generated by hotspot activity, the best known are the Canary and the Hawaiian Islands (Figures 1 and 2). Although both are formed by magmatism associated with a hotspot, the volcanoes within the two island groups display both similar and, of primary interest to us, distinctly different geologic and geophysical characteristics. We are not sure why their characteristics vary and thus the aim of this work is to illuminate possible controls on the formation, interaction, and evolution of the volcanoes and rift zones that make up these two island groups.

Previous work comparing the morphology of the Canary and Hawaiian Islands includes that by Carracedo (1999), who examined the structure and evolution of individual volcanoes, and Mitchell et al. (2002), who investigated the effect of landslides on submarine flank morphology and the role of volatiles in controlling the meso-scale morphology. In this paper we focus on the meso-scale (km) morphology of the subaerial and submarine sections of the volcanic rift zones of the two youngest Canary Islands, La

Palma and El Hierro, and compare their characteristics with those of the well-studied rift zones on the youngest of the Hawaiian Islands, Hawaii.

Our work shows that the rift zones on La Palma and El Hierro are typically shorter (10 s of km) than those on Hawaii (ranging up to >100 km). In addition, the rift zones on La Palma and El Hierro display various geometries ranging from linear ridges with smooth to irregular crests to ridges with a fan-like morphology in plan view. In contrast, the rift zones on Hawaii tend to consist mostly of well-defined linear ridges. Although the toes of some of the Hawaiian rifts do partially fan out, they do not to the extent displayed by the Canary Island rifts. Irregularities and curvature along the axes of the rift zones is also more common on La Palma and El Hierro than on Hawaii. We examine the possible causes for these observations.

Data

Comparison of the subaerial sections of the volcanic rifts in La Palma, El Hierro, and the island of Hawaii is based on published sources. Investigation of the offshore sections of the rifts in the Canary Islands is based on swath bathymetric data collected by the Instituto Español de Oceanografia and the Instituto Hidrográfico de la Marina in 1998 to 2002. Description of the submarine rifts in Hawaii is based on available swath bathymetry data from Monterey Bay Aquarium Research Institute (MBARI, 2000), side scan sonar data from a deep towed vehicle and bottom photographs (Smith et al., 2002a), and a multitude of published sources as cited in that study.

Background

Regional setting

The Canary archipelago is located on the continental rise off Cape Juby, northwest Africa. Fuerteventura and Lanzarote, at the eastern end of the chain, are 100 km from the African coast, and El Hierro and La Palma at its western end are 500 km from the coast (Figure 1). The inner islands, Lanzarote and Fuerteventura, are built on seafloor at a water depth of about 2000 m, just west of the transition from attenuated continental to oceanic crust (Emery and Uchupi, 1984). Gran Canaria, Tenerife, La Gomera, La Palma and El Hierro, built on seafloor with water depths of 2000 to 4000 m, are located on oceanic crust of Jurassic age (Uchupi et al., 1976). The boundaries of this crust are defined by magnetic anomaly S1 (180 Ma; along the transition from continental to oceanic crust) east of Lanzarote-Fuerteventura (Canary Ridge) and anomaly M-25 (156 Ma) west of La Palma and El Hierro (Emery and Uchupi, 1984).

Volcanism in the Canary Islands ranges from Late Cretaceous in the eastern islands to the present (Le Bas et al., 1986; Carracedo, 1994). Carracedo (1994) has divided the islands into three groups, those that have had eruptions in historic times (<500 yr; Tenerife, La Palma; Lanzarote and probably El Hierro), those with a history of Quaternary volcanism (Fuerteventura and Gran Canaria) and those lacking evidence of Quaternary volcanism (La Gomera). Lanzarote, Fuerteventura and Gran Canaria are in the post-erosional phase, La Gomera is in the repose stage and Tenerife, La Palma and El Hierro are in the shield stage of development (Carracedo, 1999). In Gran Canaria the post-erosional phase is also characterized by catastrophic mass wasting (Mehl and Schmincke, 1999). The seamount phase of construction is represented in the Canary Islands by the Basal Complex that is formed of basic plutonic rocks, marine sediments and volcanic materials, cut by a very dense dike network. It crops out in Fuerteventura (Stillman, 1999), in La Gomera (Bravo, 1982; Cendrero, 1970), and in the Caldera de Taburiente and the Barranco de las Angustias in La Palma (Staudigel and Schmincke, 1984).

At the southeast end of the Hawaiian Ridge are the islands of the Hawaiian chain (Figure 2), built on crust older than magnetic anomaly 34 and younger than anomaly M0, between 90 and 80 Ma (Figure 2) (Clague and Dalrymple, 1989). Some of these islands are formed by a single volcano and others consist of two or more coalescing volcanic edifices (Langenheim and Clague, 1987). The ages of the volcanoes associated with the islands get older to the northwest with historical volcanic activity being limited to Maui and Hawaii (Figure 2). Hawaiian volcanoes, like those in the Canary Islands, go through four stages of evolution: a seamount phase, the best example being Loihi Seamount sitting at the leading edge of the island chain, a shield phase during which a caldera may form, a long period of volcanic quiescence and erosion, and a post erosional volcanic phase during which a caldera also may form (Clague and Dalrymple, 1989). During the shield phase and post-shield phase eruptions occur at the summit of the volcano and along rift zones extending down the flanks of the volcanoes.

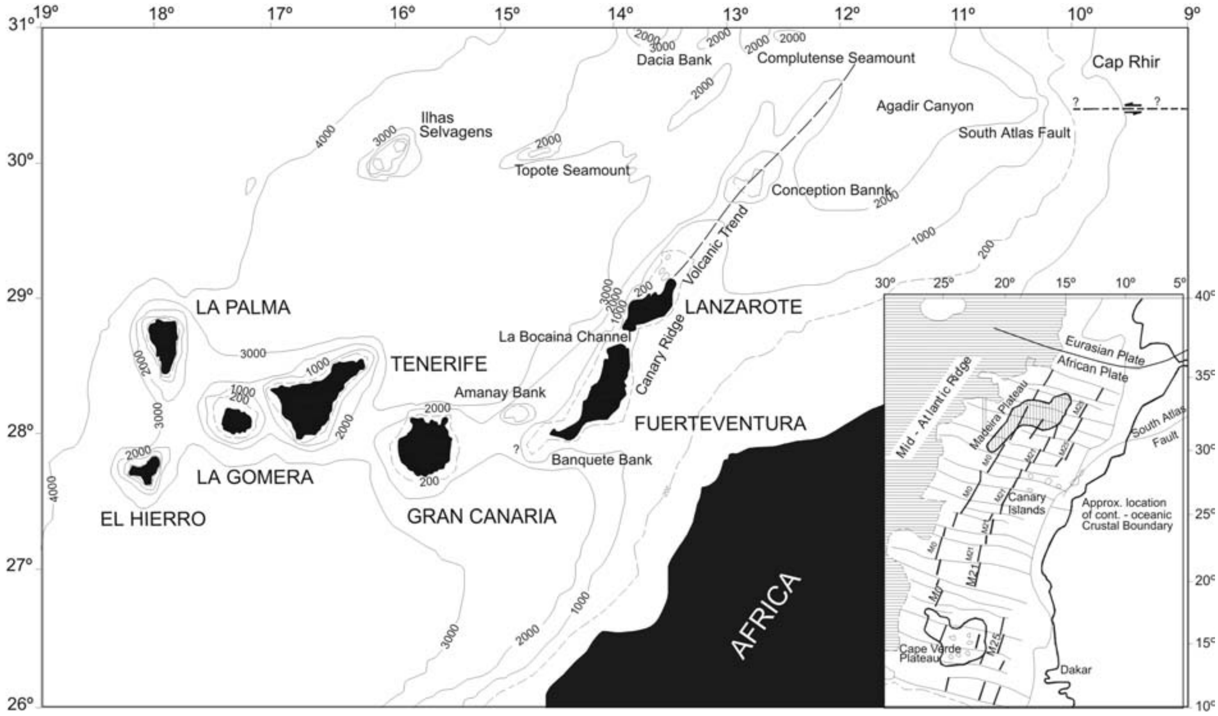


Figure 1. Geographic setting of the Canary Islands. Compiled from Emery and Uchupi (1984) and Uchupi et al. (1976).

Age progression

The volcanoes associated with the Emperor Seamounts-Hawaiian Ridge are an obvious hotspot trace superimposed on the fast moving Pacific plate ~70 mm/yr (Morgan, 1972). In contrast, the Canary Islands, on the slow-moving African plate (~25 mm/yr), form a less convincing hotspot trace (Carracedo, 1999). The Canary Islands do display a general age progression, from oldest in the east to the youngest in the west, but this trend displays several anomalies in the age-distance plot. Carracedo (1999) proposed that these anomalies may be a consequence of the islands having been created by a combination of a low-activity hotspot and a slow-moving plate. He noted that a similar age anomaly occurs in the Cape Verde Islands, an island group in the Eastern Atlantic also formed by hotspot activity. In the Cape Verde Islands, Quaternary volcanism shifted back to the east from Brava to Fogo, a situation comparable to the late Miocene shift in volcanism from La Gomera east to Tenerife. Carracedo (1999) also pointed out that Lanzarote is not really a separate island, but a northeast extension of Fuerteventura. The islands are separated by only the narrow 50-m-deep La Bocaina Channel. Differences in the age of volcanic activity in these two islands, therefore, may simply reflect the duration of their composite construction.

The lack of a simple east-west age progression in the older radiogenic age data led Anguita and Hernán (1975) to question a hotspot origin for the Canary Islands. They proposed, instead, that the islands were formed by magmatic intrusion along propagating fracture zones associated with seafloor spreading. The age of the islands (Late Cretaceous-Quaternary) is much younger than the age of the crust on which the islands are superimposed (Jurassic; Uchupi et al.,1976), however, and is not consistent with this theory. Another possibility is that magmatic activity along existing fracture zones may have been triggered by Alpine tectonism in western Africa (Uchupi et al., 1976). Price (1980), for example, proposed that volcanism along the Canary Ridge, capped by Lanzarote and Fuerteventura, was due to motion along a southwest extension of the sinistral South Atlas Fault (Figure 1). Another idea is that Lanzarote and Fuerteventura were emplaced on a structural weakness along the continental-oceanic crustal boundary. To date there are no data to verify any of these models.

New radiometric dating, however, appears to support a plume model for the Canary Islands with Fuerteventura-Lanzarote forming 20–21 Ma, Gran

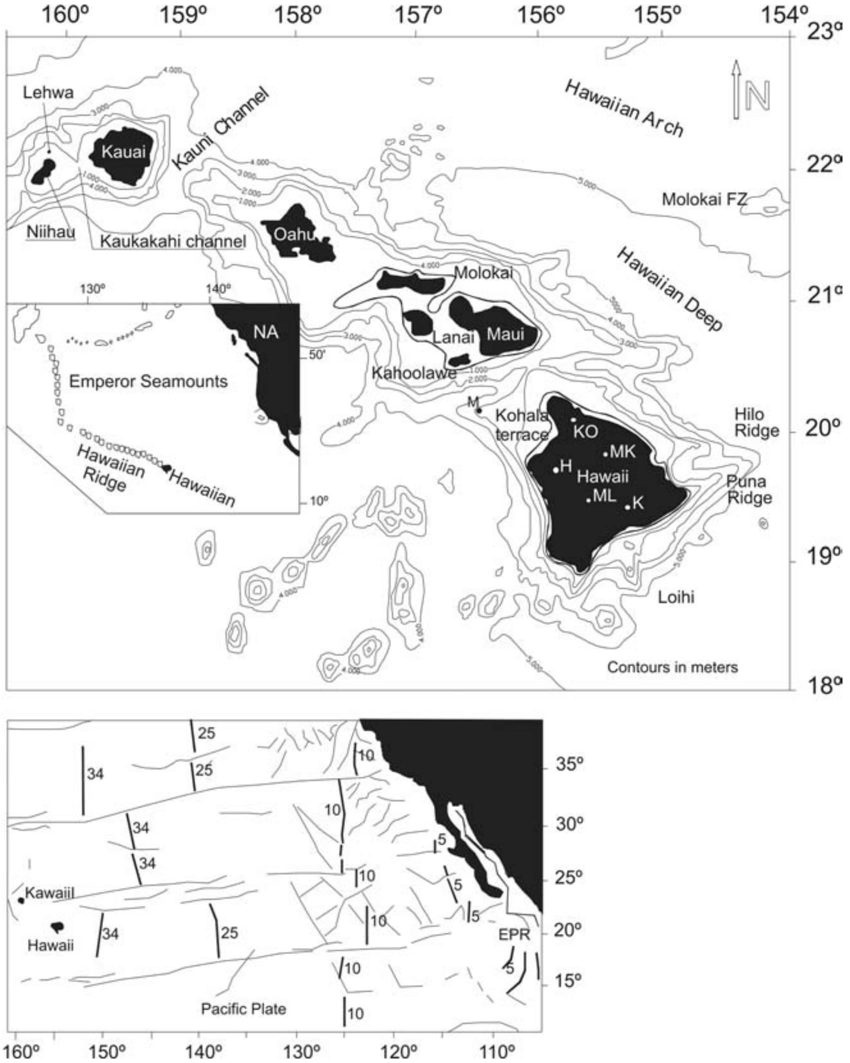
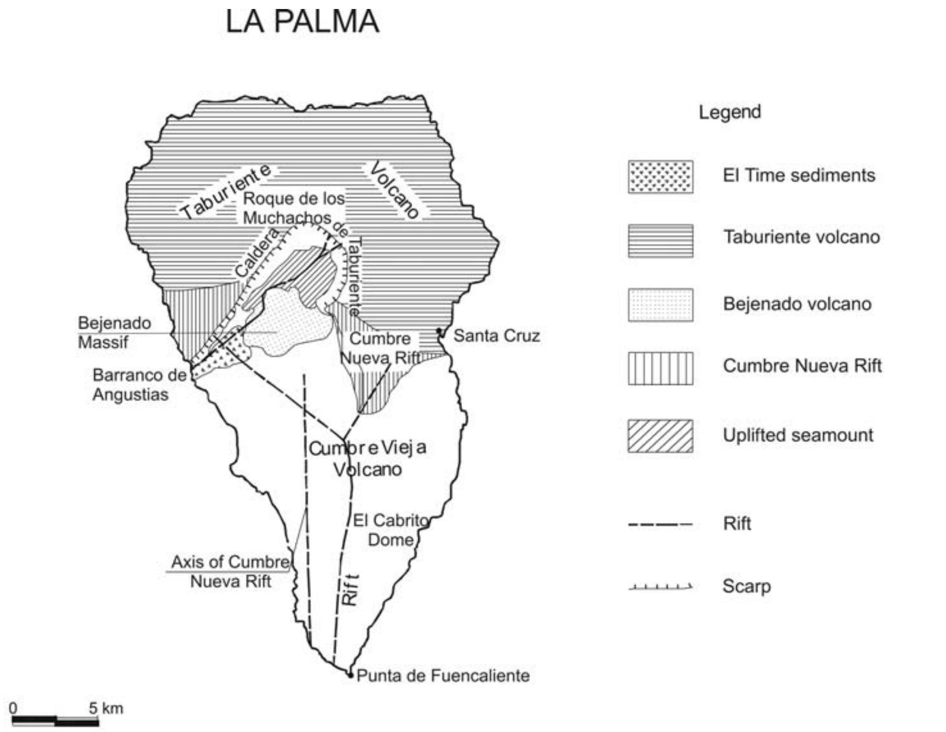


Figure 2. Top panel: geographic setting of the Hawaiian Islands. Modified from Fornari and Campbell (1986). Volcanoes are labeled as: KO-Kohala; H-Hualalai; MK-Mauna Kea; ML-Mauna Loa; KI-Kilauea. Lower panel: tectonic map of the eastern Pacific Ocean. Compiled from Atwater and Severinghaus (1989) and Decker (1989). Insert Map show the relationship of the Hawaiian Ridge and the Emperor seamounts. Modified from Clague and Dalrymple (1989).

Canaria 14–15 Ma, Tenerife 11–12 Ma and La Gomera 9–10 Ma. After the formation of La Gomera the east to west hotspot activity split into a north-south component, with dual hotspot activity forming La Palma and El Hierro < 2 Ma. Such dual volcanism is generally associated with changes in plate motion, but to date no such change in plate motion has been noted in the Canaries (Carracedo et al., 2001). On the basis of the new radiometric data Carracedo et al. (2001) concluded that the extension of the Atlas Fault System to the Canary Islands is a geographic coincidence and did not control the formation of the islands. They further proposed that the hotspot produced the islands, and that the hotspot is either spreading westward beneath the lithosphere or is fixed beneath the plate, which is slowly moving eastward.

Magma supply

The Hawaii and Canary Islands groups differ in their rates of magma supply. For example, eruption rates in Kilauea Volcano, Hawaii, from 1956 to 1983 suggest that the magma supply rate to this volcano is on the order of 86 km^3/1000 yr (e.g., Tilling and Dvorak,



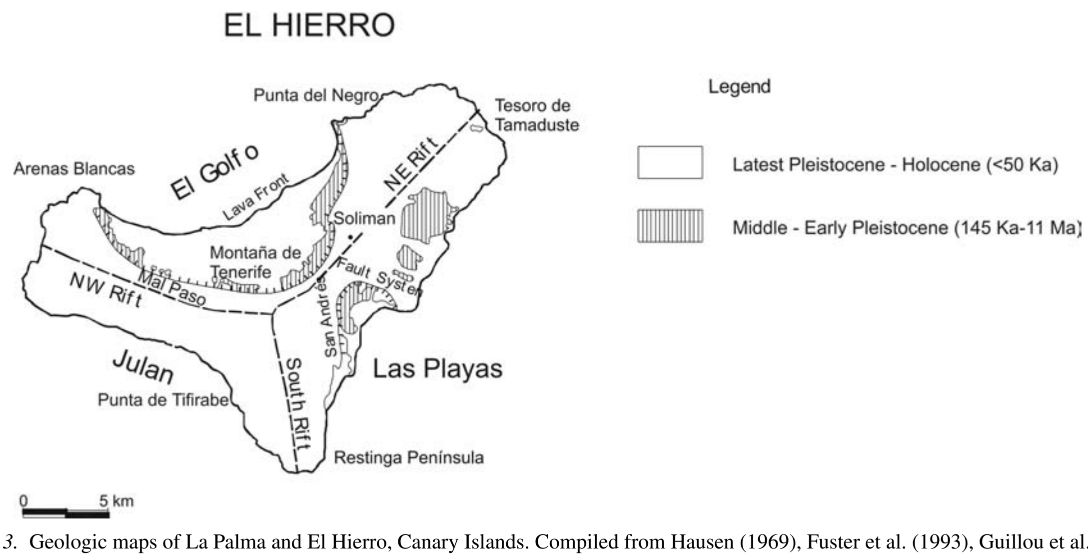


Figure 3. Geologic maps of La Palma and El Hierro, Canary Islands. Compiled from Hausen (1969), Fuster et al. (1993), Guillou et al. (1998) and Carracedo et al. (1999a).

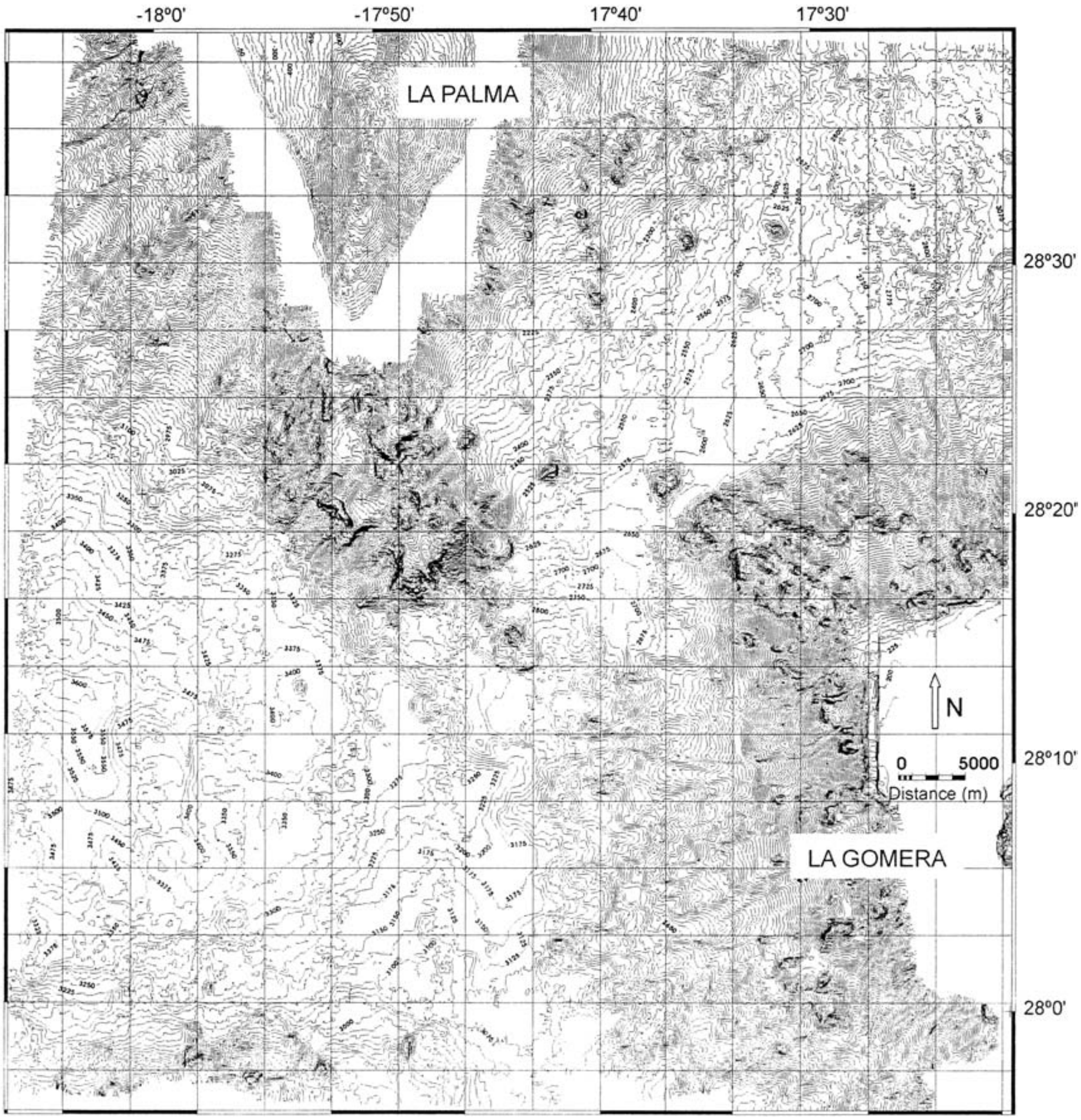


Figure 4. Topography of southern La Palma and swath bathymetry of the rift south of La Palma.

1993 and references therein). Magma supply rate can be estimated for El Hierro, which is probably the youngest and the best constrained geochronologically of any the Canary Islands (Guillou et al., 1996). The present emerged volume of the island, of about 140–150 km^3, has been produced in the last 1.2–1.5 Ma, giving an apparent average magma production rate of 0.12–0.13 km^3/1000 yr (Carracedo, 1999). Carracedo took into account the volume lost during the three giant lateral collapses that have affected the island, and concluded that the volume of the island would only be increased by $\sim$100 km^3 for each collapse, yielding only a slightly greater magma supply rate of about 0.40 km^3/1000 yr. The fact that magma supply is much greater in Hawaii is reflected in the differences in the areas of the islands. The Island of Hawaii has an area of 10,458 km^{2}whereas La Palma and El Hierro have areas of 687 and 273 km^2 respectively and even the whole Canary Archipelago only has a total area of 7,273 km^2 representing 70% of the area of the Island of Hawaii.

Cooling and subsidence

The Canary and Hawaii Islands vary in their rate of subsidence due to lithospheric cooling as they move

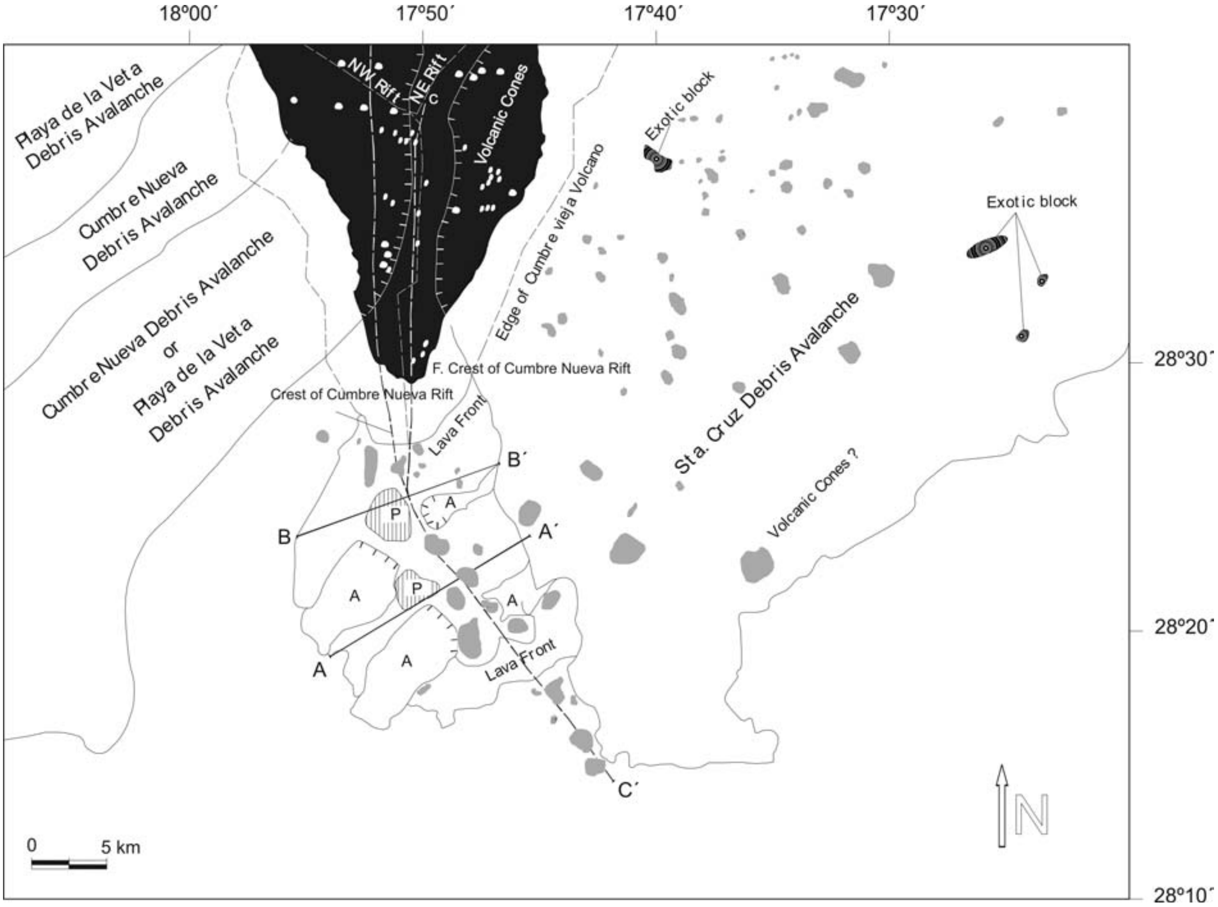


Figure 5. Morphologic interpretation of the swath bathymetry in Figure 4. A = Avalanche; P = Platform; A-A′, B-B′ and C-C′ = Profiles.

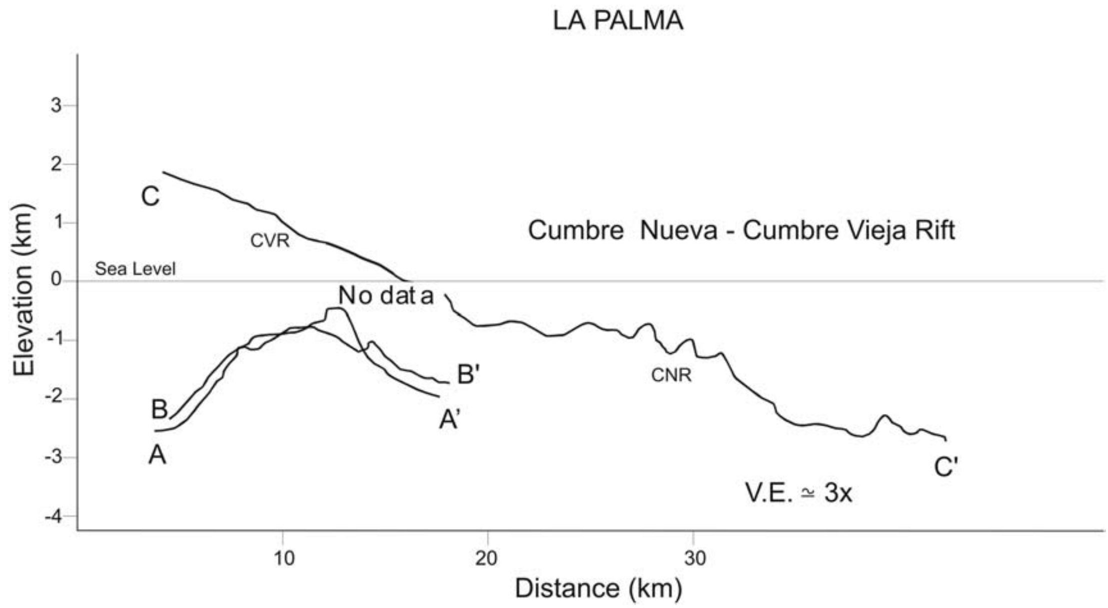


Figure 6. Profiles of the rift (Cumbre Vieja) south of La Palma. Compiled from Figure 4. See Figure 5 for locations of profiles.

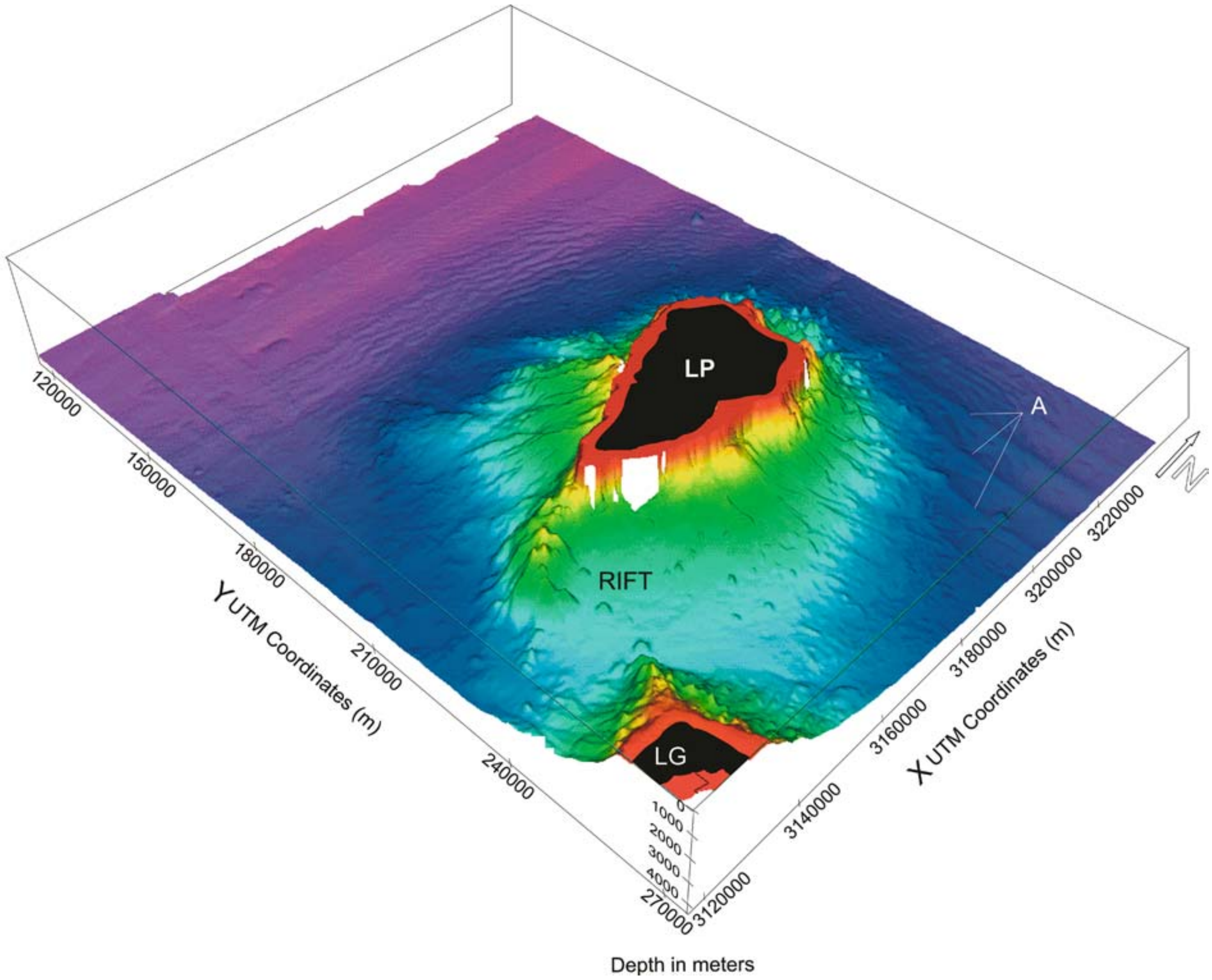


Figure 7. Relief diagram of the Cumbre Vieja Rift south of La Palma. Compiled from swath bathymetry. LG = La Gomera; LP = La Palma; A = Artifacts.

away from the hotspot (Morgan, 1970; Clague and Dalrymple, 1989 and references there in). In the Hawaiian Islands, there is a gradual progression from active volcanoes, Mauna Loa and Kilauea, to erosional remants (Niihau, Nihoa, Necker), growing atolls, (French Frigates Shoal and Midway Islands) to submerged guyots, (Ojin and Suiko) (Clague et al., 1988). Moore and Campbell (1987) estimated that most of the Hawaiian volcanoes have subsided 2–4 km and Moore and Campbell (1987) proposed that the bulk of this subsidence is rapid and concludes within a million years of the end of the shield phase. A much reduced subsidence rate due to regional processes continues beyond this phase. Dating of reefs on the gentle slope northwest of the Island of Hawaii indicates that it has been subsiding at a rate of 2.6 ± 0.4 mm/yr during the last 463 k.y. (Moore and Fornari, 1984; Moore and Clague, 1992). This rate of subsidence is such that one of the oldest islands in the chain, Kauai, is less than 6 Ma (McDougall, 1979).

No such subsidence trend is seen in the Canary Islands, with over 20 m.y. of volcanic rocks being exposed on some of the islands (Carracedo, 1999). The islands did, however, experience subsidence during the shield phase. Bathymetric data indicate a change from gently sloping ($\sim$87 m/km) subaerial lavas to more steeply sloping ($>$176 m/km) submarine lavas off shore Gran Canaria. The depth of the now submerged subaerial lavas suggests that during the shield building phase this island subsided 600–800 m (Funck and Schmincke, 1998). In the post-shield phase of island development, Gran Canaria and Tenerife have remained relatively stable with respect to sea level since about 14 Ma (Funck and Schmincke, 1998; Carracedo, 1999). Funck and

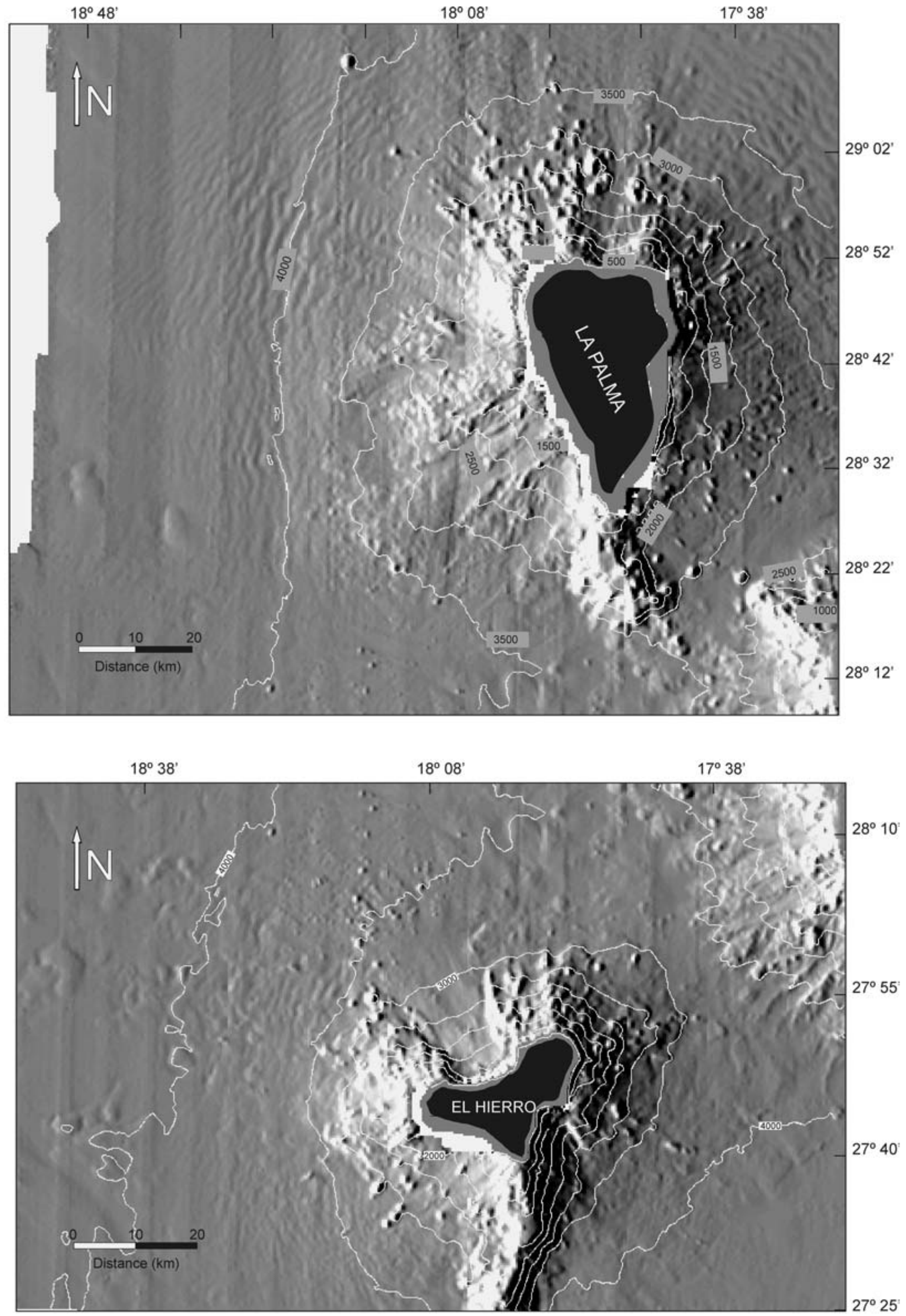


Figure 8. Relief diagrams of La Palma and El Hierro compiled from swath bathymetry. From Acosta et al. (this volume).

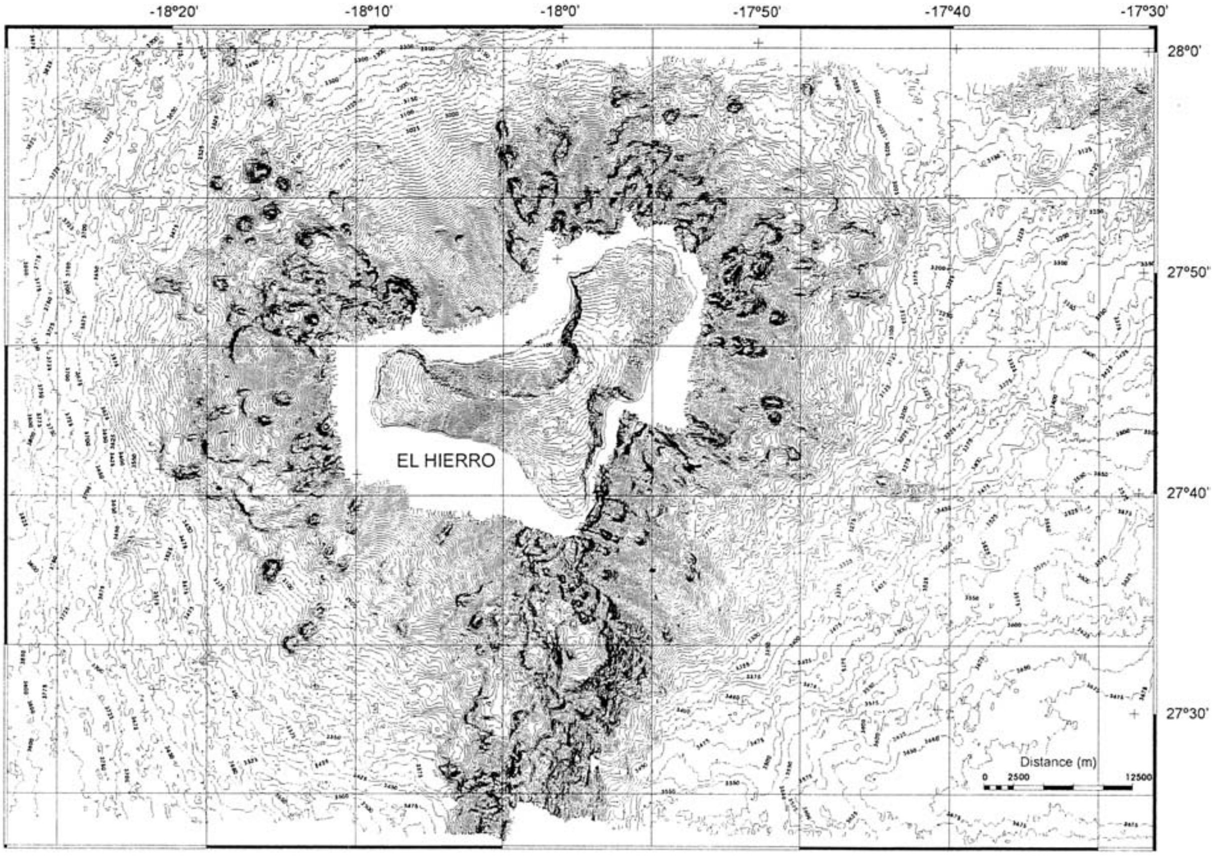


Figure 9. Bathymetry of the El Hierro offshore area based on swath bathymetry.

Schmincke (1998) proposed that this stability is due to a combination of minor volcanic loading during the post-shield phase and the strength of the Jurassic lithosphere, beneath the islands.

There is also evidence that at least two of the older islands, Lanzarote and Fuerteventura, have undergone recent uplift, as documented by raised marine terraces, beaches and Tertiary marine and lacustrine limestone at elevations of 55–60 m, 20 m and 10 m above sea level (Hausen, 1959; Coello et al., 1992; Stillman, 1999; Zazo et al., 2002). The raised Pleistocene terraces indicate an average uplift of 1.7 cm/1000 years since 1 Ma, with the present elevation of the last interglacial deposits suggesting that during the last 300,000 years Fuerteventura has been stable and that Lanzarote has subsided at about 0.7 cm/1000 years.

Swell and deep

The regional bathymetry of the Hawaiian Islands displays a moat, the Hawaiian Deep, with a maximum depth of over 5500 m. Beyond the moat is a broad swell, the Hawaiian Arch, a zone of tension whose crest is less than 4500 m deep, 500–1000 m shallower than the adjacent deep. These features on Cretaceous oceanic crust are the manifestations of lithospheric flexure caused by volcanic loading. Straddling the arch off the island of Oahu is the 24,000 km^2 active North Arch Volcanic Field (Clague et al., 2002). According to Clague et al. this field surrounds Cretaceous volcanic ridges, flat-topped volcanoes, and low relief sediment covered seafloor. They inferred that the volcanic field is younger than the 0.5–1.5 Ma age previously assigned to the field. South of the Island of Hawaii the arch is covered by flat sheet flows whose thin sediment cover and thin palagonite rinds on the lava surface suggest that the flows were erupted 1–10 Ka (Lipman et al., 1989).

No such topographic moat and swell or volcanic features are exposed on the seafloor surrounding the Canary Islands. Canales and Dañobeitia (1998), who studied the coherence between gravity and topography, proposed that a swell may be present, but likely masked by a thick sedimentary and volcanic cover.

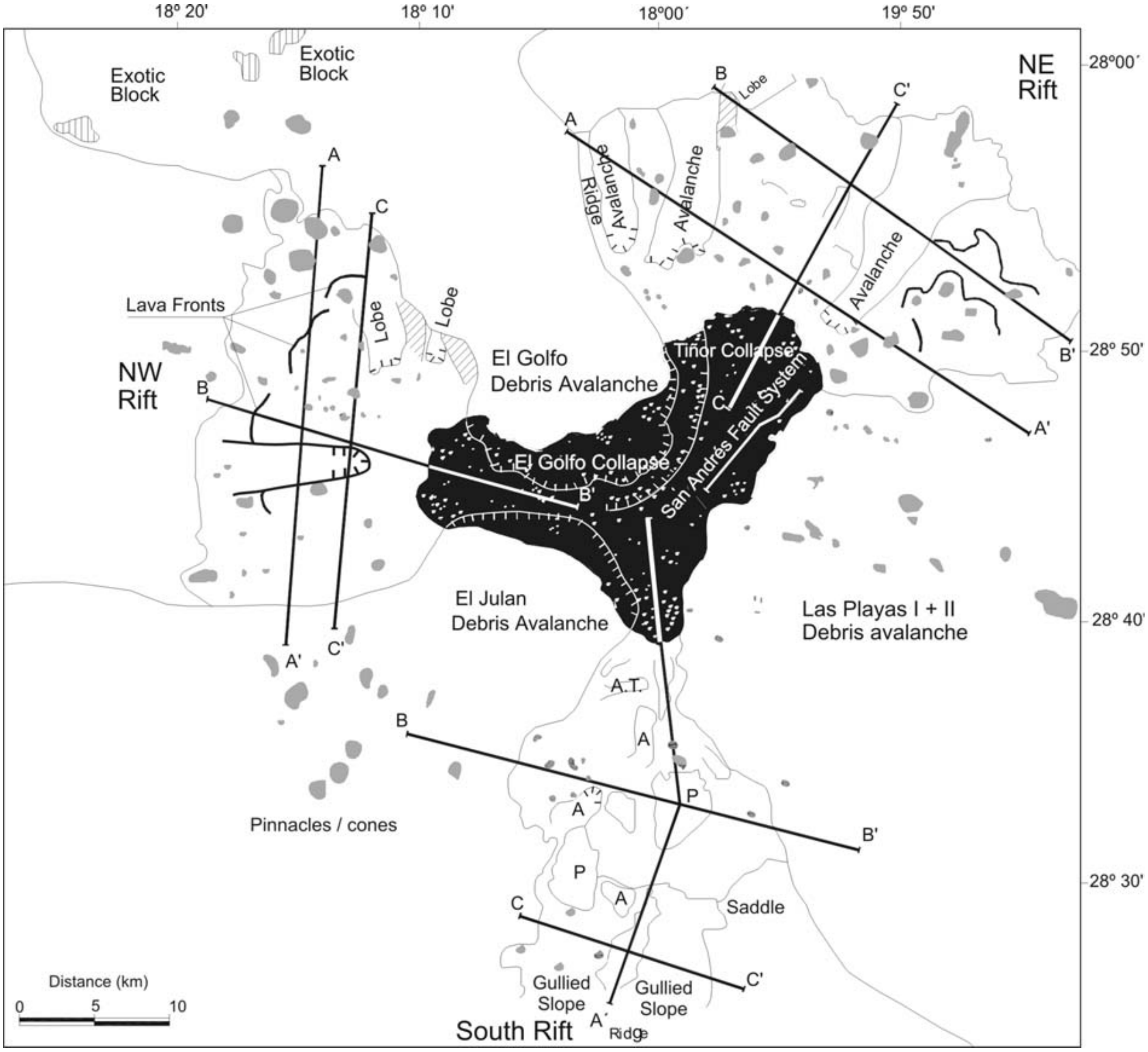


Figure 10. Morphologic interpretation of swath bathymetry in Figure 9. A = Avalanche; P = Platform; A–A′, B–B′ and C–C′ = Profiles.

This was verified by seismic reflection profiles (Watts et al., 1997) which showed a buried moat off Tenerife with a relief of 2–3 km and filled with sediments in part derived from the islands (Watts et al., 1997).

Volcanic and tectonic features of volcanoes

The geologic similarities between the Canary Archipelago and the Hawaiian Islands are as striking as their differences. As mentioned, both island groups underwent four volcanic phases: a seamount phase, a shield-building submarine and subaerial phase, characterized by rapid growth and massive slope failures, a period of quiescence and deep erosion (erosional gap), and a post-erosional stage of volcanic activity.

Volcanic rift zones in Hawaii tend to form during the seamount phase and grow above sea level as the edifice matures (e.g., Moore and Chadwick, 1995). They tend to form along zones of extension propagating from a magmatic center with the three principal rift arms arranged with angles of about 120° between them, possibly controlled by tangential stresses produced by a rising plume (Wyss, 1980). This origin also has been proposed for the rift zones in the Canary Islands (Carracedo, 1994). However, not all the volcanoes in the Hawaiian chain and the Canary Islands

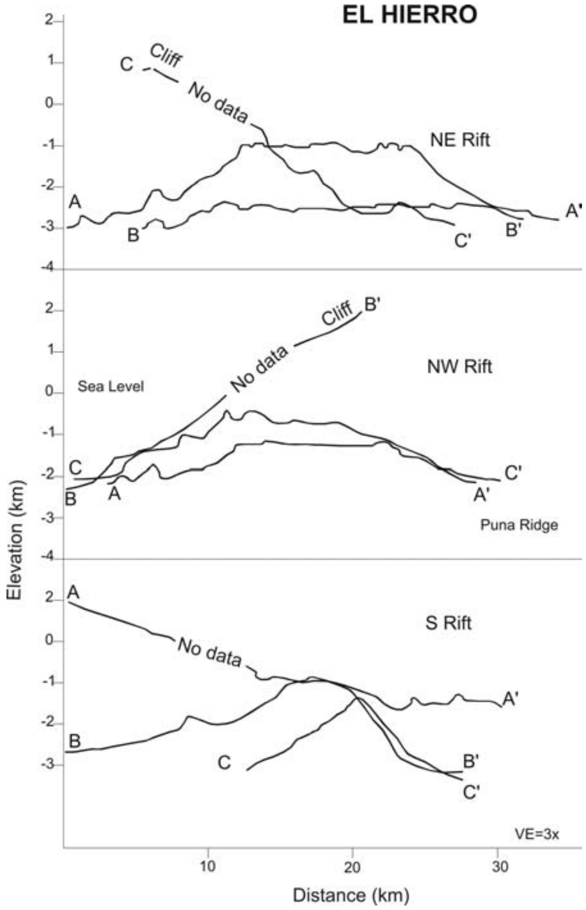


Figure 11. Profiles of Northeast, Northwest and South rifts of El Hierro. Compiled from Figure 9. See Figure 10 for location of profiles.

have three rift arms. Fiske and Jackson (1972) suggested that the locations of rift zones in Hawaii, and supposedly those in the Canary Islands, also are controlled by gravitational stresses within the volcanic edifice, which are influenced by the shapes and positions of preexisting volcanoes whose flanks serve as foundations for the younger structures.

On a small scale, rift zones in both the Hawaii and the Canary Islands are marked by eruptive vents and fissures aligned along the rift, open cracks, pit craters and grabens that confirm that the rift zones are the location of dike intrusions and fissure eruptions. It is thought that the development of the rift systems and the subsequent concentration of dikes injected into individual rifts promotes destabilization of the volcano flanks. Magma overpressure and mechanical and thermal overpressure of pore fluids leads to gravitationally unstable flanks (Carracedo, 1994, 1996; Elsworth and Day, 1999). In Hawaii, the rifts form the boundaries for zones of slumping that may ultimately lead to catastrophic collapse (Moore et al., 1989). Typically, however, the headwalls of the landslides are seaward of the rift zone axis (e.g., Moore and Clague, 1992). The wide arcuate landslide depressions in the Canary Islands often tend to be located between the two most active rifts, some with the third rift acting as a buttress for the landslide.

At a larger scale caldera formation at the summit of the volcanoes is common in both island groups. In the Canary Islands the calderas have been ascribed to several processes including: 1) collapse due to mass wasting (Ancochea et al., 1994; Masson et al., 2002), 2) fluvial erosion (Carracedo et al., 1999b), 3) magma chamber collapse (Martí et al., 1994, 1997), 4) a combination of collapse and post-collapse incision and retrogressive erosional collapse (Carracedo et al., 2001), and 5) a complex interplay of volcano inflation and deflation cycles (Troll et al., 2002). In Hawaii it has long been recognized that the formation of summit calderas is linked primarily to magma movement in

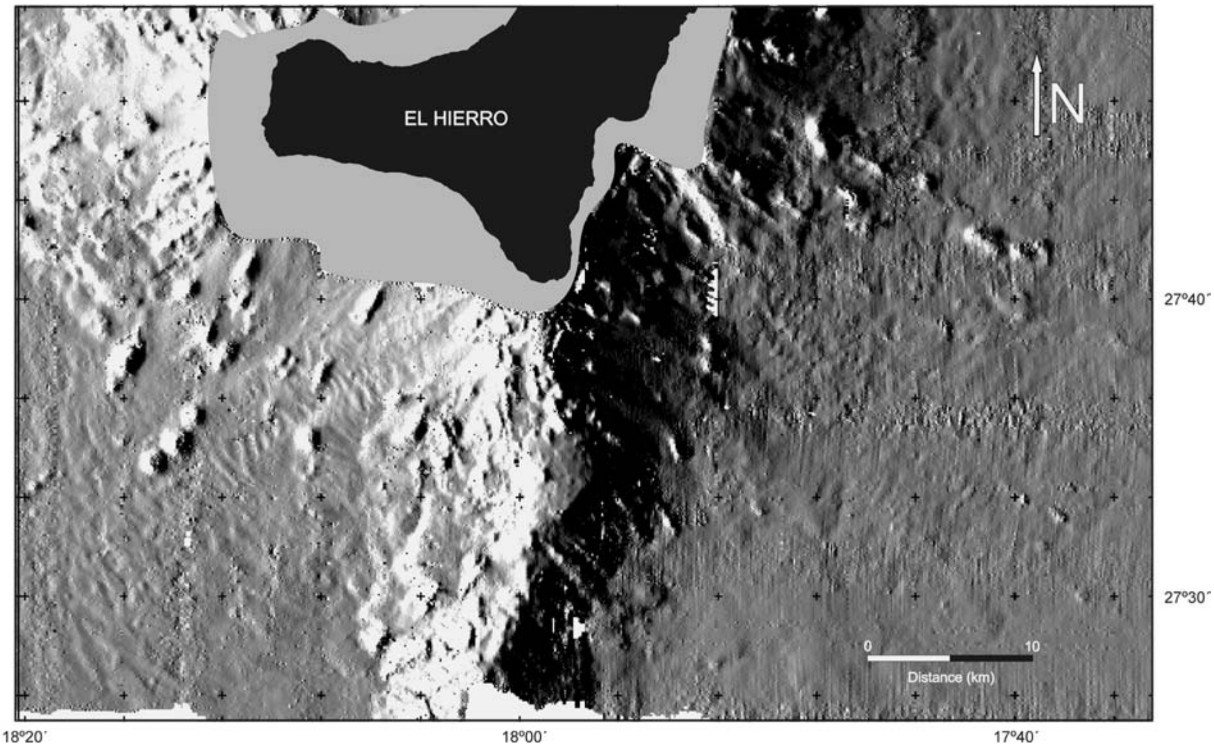


Figure 12. Relief diagram of South Rift off El Hierro. Compiled from swath bathymetry.

the summit reservoir. The roof of the reservoir is left unsupported when the magma is withdrawn leading to caldera collapse (e.g. Peterson and Moore, 1987; Tilling and Dvorak, 1993). Collapse may be incremental and recurring (Peterson and Moore, 1987). Walker (1988) pointed out, however, that the relationship between caldera subsidence and magma movement in Hawaiian volcanoes may not be as straightforward as we think. He suggested that subsidence may be caused locally by the excess load of intrusives. Either way it is clear that caldera formation in Hawaii is associated with magmatic processes compared to caldera formation in the Canary Islands, which can be ascribed to several different processes.

Canary Islands

In this section we examine the characteristics of La Palma and El Hierro, Canary Islands, and the Island of Hawaii. We then focus on the volcanic rift zones and compare and contrast their characteristics. The results will be used to help constrain the magmatic process that form these islands.

La Palma

General geology

The pendant-shaped, north-south trending island of La Palma is the second youngest island of the Canary Island archipelago (Figure 3). It has a maximum width of 28 km in an east-west direction, a maximum length of 47 km in a north-south direction, a maximum elevation of 2426 m above sea level at Pico de Los Muchachos, along the rim of Caldera de Taburiente, and an area above sea level of 687 km^2. The surface of the island is cut by many barrancos, the largest of which is the westward-draining Barranco de las Angustias (Figure 3). Barranco de las Angustias originates in the Caldera de Taburiente, whose vertical walls are up to 800 m high (Hausen, 1969) and whose floor has a maximum elevation of 1500 m above sea level (Hausen, 1969; Ancochea et al., 1994). The outer flanks of the caldera are incised by a radial system of barrancos, draining way from the center of the dome whose collapse, incision and retrogressive erosion (Carracedo et al., 2001) formed the Caldera de Taburiente. Extending southward from the rim of the caldera to Punta de Fuencaliente at the southern tip of La Palma is a narrow ridge, Cumbre Vieja, whose

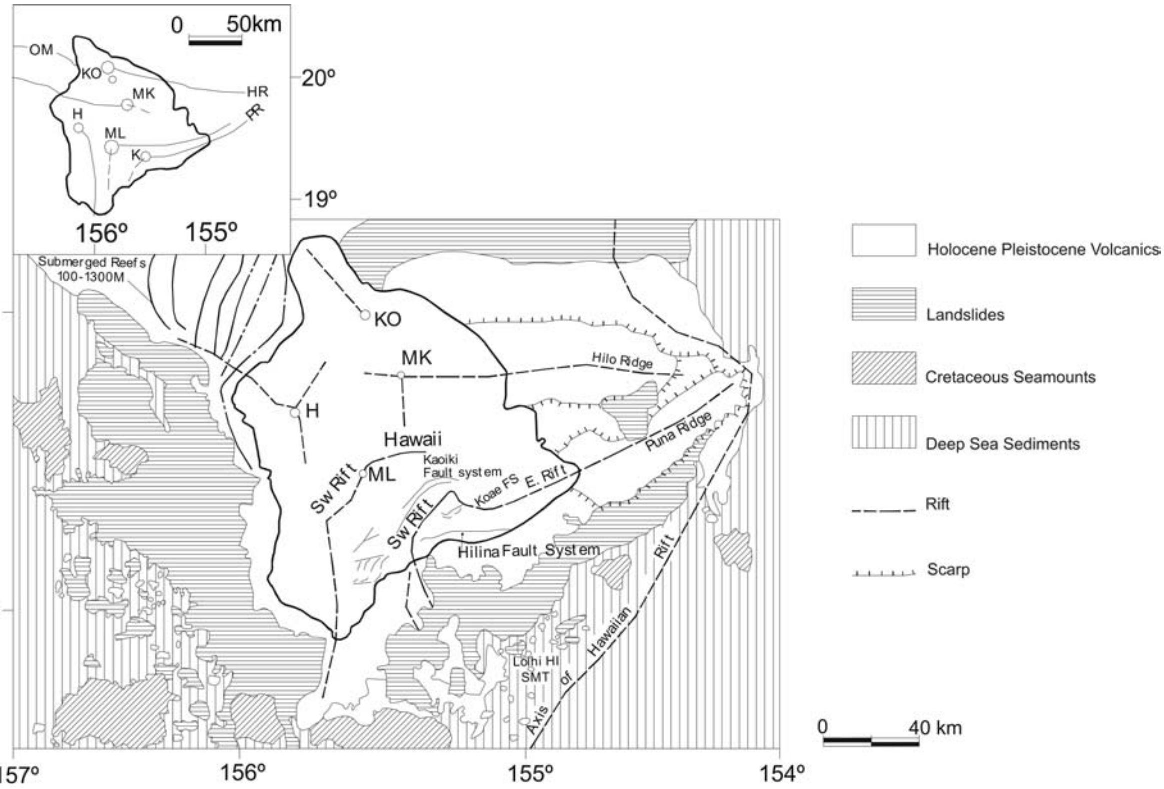


Figure 13. Geologic map of Hawaii compiled from Moore and Campbell (1987), Moore and Chadwick (1995) and D.J. Fornari (written communication, 2003). Insert shows distribution of rift zones postulated by Holcomb et al. (2000). Volcanoes are labeled as: KO-Kohala; H-Hualalai; MK-Mauna Kea; ML-Mauna Loa; KI-Kilauea. The east rift zone (ERZ) and southwest rift zone (SWRZ) of Kilauea are labeled.

crest is at an elevation of nearly 2000 m above sea level (Figure 3). Along the crest and slopes of Cumbre Vieja are many relatively young cinder cones. Lava flows originating from the cones extend to the east and west coasts of La Palma, filling many of the barrancos in the southerly slopes. The Cumbre Vieja has had seven eruptions in historical times, with the last taking place ~500 years ago (Guillou et al., 1996). La Palma's coasts are generally steep and rocky and except for Santa Cruz (Figure 3) on the east coast, which to a great extent is artificial, the rocky coasts lack natural harbors.

Geologically, La Palma can be divided into two sectors (Figure 3). The northern two-thirds of the island consists of the basement complex and the lavas of the Cobertera Series (1.7 to 0.8 Ma according to Ancochea et al., 1994 and 1.77 to 1.20 Ma according to Carracedo et al., 2001). The southern third of the island is dominated by the Cumbre Vieja volcano whose older lavas have been dated to 123 Ka (Carracedo et al., 2001). The northern complex consists of a basal plutonic sequence made up of small intrusions of gabbro, leucogabbro and cumulative ultramafics overlained by sills, and an alkali basaltic dike swarm cutting through the complex. Also included within and resting on the basement complex also are four debris avalanches. Staudigel and Schmincke (1984) inferred that the section was formed during the building of a seamount as it rose from deep-water to above sea level. The age of the seamount remains to be resolved. Hernández Pacheco and Fenández-Santin (1975) described Miocene microfossils in the seamount's submarine series, Staudigel and Schmincke (1984) found Pliocene fossils in the complex and Feraud et al.(1985) reported an age of 9 Ma and younger for the dikes in the complex.

The Garafía Volcano, with a diameter of ~20 km and a relief of 2500 and 3000 m, was constructed in the northern part of La Palma over the basement complex during the emplacement of the Lower Series Basalts (2.0–1.3 Ma according to Ancochea et al. (1994) and 1.77 to 1.20 Ma according to Carracedo et al. (2001). About 1.2 Ma, a paleocaldera formed on Garafía Volcano, as a consequence of the gravitational sliding of the southern slope or collapse and subsequent subaerial erosion (Carracedo et al., 2001). Avalanches created by the collapse of the Garafía Volcano are represented inshore by a several hundred meter-thick breccia (Carracedo et al., 1999a). An uplift of ∼1000 m of the central part of La Palma and westward tilting of large blocks of the basement complex were associated with the caldera formation. The uplift was subdued by emplacement of the 0.89 Ma Cumbre Nueva Series of Taburiente Volcano. This volcanic activity ceased around 0.4 Ma having formed three coalescing volcanoes to make up the northern shield of La Palma (Carracedo et al., 2001).

The Cumbre Nueva Series formed a volcanic edifice and a north-south trending rift, the Cumbre Nueva Ridge, along the southern flank of the volcano during its last stages of growth (Figure 3) (Day et al., 1999; Carracedo et al., 2001). Ancochea et al. (1994) inferred that the Cumbre Nueva Ridge rift only extended a short distance south of its present outcrop next to the Caldera de Taburiente. In contrast, Day et al. (1999) proposed that the rift extended further southward to nearly the southern tip of La Palma. Whether it extends offshore is not known. As it grew, the Cumbre Nueva Ridge became progressively steeper until its western flank collapsed and initiated the formation of the Caldera de Taburiente. The Cumbre Nueva Avalanche off the west coast of La Palma represents the debris avalanches created during this collapse (Urgeles et al., 1999; Masson et al., 2002). Another extensive debris flow, the Santa Cruz Debris Avalanche, off the east coast of La Palma, also could be a product of this collapse, or possibly the flow is older and is due to the collapse of the Garafía Volcano, unfortunately Carracedo et al. (2001) stated that there is no geological evidence for this debris on land.

Ancochea et al. (1994) proposed that the Caldera de Taburiente was formed by large landslide events. They suggested that the debris avalanches at the bottom of the Caldera de Taburiente, the age of the lava flows at the top and within the caldera, and the absence of materials from the caldera wall, between the Basal Complex and the Bejenado Massif, a volcano that fills the intersection of the Cumbre Nueva and Taburiente Calderas (Figure 3), supports such an origin. Bejenado Volcano has been dated by Ancochea et al. (1994) as being ∼0.70–0.71 Ma. In contrast, Carracedo (1999) speculated that Bejenado Volcano began to form soon after the formation of the Caldera de Taburiente about 560 Ka and ended 400 Ka. Guillou et al. (1998) proposed that the Caldera de Taburiente was formed from fluvial erosion by a drainage system trapped between the growing Bejenado volcano and the 558 Ka Cumbre Nueva collapse scar. Carracedo et al. (2001) speculated further that the morphology of the Caldera de Taburiente was formed by series of events: (1). Gravitational collapse of the Cumbre Nueva Volcano, (2) Development of the Bejenado volcano in the collapse structure, and (3) Incision of the Barranco de Las Angustias to form the initial Caldera de Taburiente.

Cumbre Vieja Rift: Onshore section

Dominating the southern third of La Palma is the north-south trending Cumbre Vieja rift whose construction took place in the last 123 Ka (Figure 3) (Carracedo et al., 2001). Day et al. (1999) inferred that the Cumbre Vieja Ridge volcanic rocks rest on the collapse scar formed by the failure of the western flank of the Cumbre Nueva rift about 560 Ka. Some authors have inferred that the Cumbre Vieja is a single rift somehow associated with the older volcanoes in northern La Palma (Ancochea et al., 1994). Others workers, (Carracedo (1994) and Carracedo et al. (1999a, b), have proposed that the construction of the Cumbre Vieja Volcano was controlled by a triple rift system, a north-south and northeast and northwest rift system that was re-organized into a single north-south rift at 7 Ka.

During the eruptive cycles the 24 km long and 17 km wide Cumbre Vieja Volcano grew into a steep sided ridge with side slopes of 300–400 m/km and a crest nearly 2000 m above sea level. Its west slope is partially covered by basaltic streams originating upslope, some of which date to historical times and its eastern side is strewn with cinder cones extending to the coast (Hausen, 1969). At its southern extremity is the 1860 m high El Cabrito lava dome from where the ridge descends to sea level and a field of cones (Hausen, 1969). According to Carracedo et al. (1999a, b) a Pleistocene glacially induced regression led to the formation of seacliffs on the western flank of the rift that in places are as high as 700 m. During the subsequent Holocene rise in sea level, the cliffs were partially buried between 20 to 15 Ka by scree-forming lavas. We question such a scenario, as seacliffs are indicative of transgressions, not regressions and in fact they are one of the features that are used to distinguish shores of submergence from ones of emergence. Thus it seems more geologically reasonable to infer that the cliffs were formed during a Pleistocene transgressive cycle and the lavas cascading over the cliffs were emplaced during a subsequent regression.

The Cumbre Vieja Ridge appears to be unstable and in state of collapse. During the 1949 eruptions, for example, normal fault ruptures developed along its crest (Klügel, 1997). Day et al. (1999), Moss et al. (1999) and Ward and Day (2001) infer that these ruptures are not the surface expression of a dike, but may represent surfaces of separation along which Cumbre Vieja volcano may collapse westward in the near future. Ward and Day (2001) inferred that this collapse could occur in historical time.

Cumbre Vieja Rift: Offshore section

The offshore ridge on strike with the onshore Cumbre Vieja Ridge is made up of two topographic segments. From the coast to a depth of about 800–110 m the offshore extension of Cumbre Vieja Ridge takes the form of a 15 km long platform that terminates along a 300 m high scarp. The scarp has a gradient of 200 m/km and probably represents a lava flow front overlapping the main part of the offshore ridge. In the absence of chronological data we have inferred from morphologic data that the ridge segment south of the platform is an older, lower part of the Cumbre Vieja rift. If so then the Cumbre Vieja rift may consist of a series of volcanic sequences overlapping one another in a southward direction. Such a model presupposes that rifting associated with the Cumbre Vieja propagated southward, first along the Cumbre Nueva Volcano and further south on oceanic crust. However if Day et al. (1999) contention that the Cumbre Nueva extends in the subsurface as far south as the southern tip of La Palma is correct, then it is not geologically unreasonable to suggest that the section of the offshore rift south of the platform is part of Cumbre Nueva, not the Cumbre Vieja Volcano. If so then the seaward edge of the platform is the southern end of the Cumbre Vieja rift resting on the older rift. That the Cumbre Nueva rift is much longer than the Cumbre Vieja also makes sense geologically as construction of Cumbre Nueva rift took place since of 330 Ka, whereas the Cumbre Viejo rift was constructed during the last 123 Ka (Carracedo et al., 2001).

We assume that individual eruption cycles are superimposed on one another. This implies that the main offshore extension of the Cumbre Vieja Ridge, an older segment of the ridge, is about 38 km long and has a maximum width of about 27 km, with its crest plunging 800 m southeastward for a distance of about 15 km. It is convex westward in plan view (Figures 4, 6 and 7). Its morphology appears to be the creation of mass wasting events and volcanic construction along its crest and northeast, south, and southwest flanks. The east flank of Cumbre Vieja Ridge has a topographic gradient ranging from 300 m/km between 1300 to 1700 m water depth, 350 m/km from 1800 to 2400 m and 300 m/km from 2600 to 3200 m. Its west flank has a gradient that ranges from 300 m/km from 1600 to 1900 m water depth and 200 m/km at depths of 1900 to 2100 m. Its southeast flank has a relief of 1200 m and a gradient of 500 m/km.

Along the crest of Cumbre Vieja are 21 volcanic cones (Figures 5, 7 and 8) ranging in relief from 25 to 300 m and surrounded by aprons suggestive of lava flows. The volcanic structures are conical in shape, with diameters of about 1 km to 100 m. The crest of the ridge is also disrupted by a 200 m high step at a water depth of 1900–2100 m (Figure 8) that we interpret as a lava front. Southeast of this step are two volcanic edifices separated by a saddle whose relief is about 50 m. The high south of the saddle has a tail-like structure on its west side that resembles a lava flow. To the east of this high is another volcanic edifice whose construction has led to a pronounced protrusion of the Cumbre Vieja Ridge's eastern flank.

The flanks of the ridge have the appearance of being mass wasted, that is, characterized by embayments separated by narrow divides and volcanic lobes. On the ridge's east side northeast progradation of one of these volcanic lobes led to the construction of an amphitheater low constrained to the north by one of the north-south trending ridges. The relatively steep southeast side of the Cumbre Vieja Ridge resembles a lava front whose slope may have been enhanced by mass wasting. Two short ridges extending from the southwest of the front, one 75 m high trending south and the other, 300 m high trending southwest. Near the southeast edge of the 300 m tall ridge is another southeast-trending high made up seamounts with heights ranging between 100 to 300 m. East of the ridge are three widely spaced seamounts forming a chain extending to the northwest tip of the La Gomera's margin (Figures 4, 7 and 8).

El Hierro

General geology

El Hierro is the youngest of the Canary Islands and may be located over the present site of the hotspot (Holik et al., 1991; Hoernle et al., 1991). However, Carracedo et al. (2001) inferred that volcanism in La Palma and El Hierro alternated between them, a scenario that may be due to changes in the stress fields triggered by gravitational collapses at the peak of volcanic construction. Carracedo et al. (2001) also concluded that the volcanic shields in El Hierro overlap concentrically, whereas those in La Palma are aligned in a north-south direction as independent shield volcanoes. As a consequence of this large overlap El Hierro formed over a stationary magma source characterized by a well defined rift system.

El Hierro has an area of 273 km^2, a maximum relief of 1500 m above sea level at Mal Paso (Figure 3), and displays a few barrancos (Hausen, 1973; Fuster et al., 1993). In plan view the island has three lobes whose intersection forms a central tableland on which are superimposed numerous extinct cinder cones (Figure 3). The island's periphery consists of steep cliffs indented by three embayments (Figures 3 and 8): El Golfo in the island's northwestern sector, Las Playas opening to the southeast, and El Julan opening to the southwest. These cliff-bound embayments resulted from large mass wasting events (Ridley, 1971; Hausen, 1973; Bravo, 1982; Holcomb and Searle, 1991; Carracedo, 1994; 1996).

El Hierro was built during the Tiñor and El Golfo volcanic phases. The Tiñor volcano at the eastern part of the island was built between about 1.12 and 0.882 Ma and El Golfo from about 550 to 130 Ka (Guillou et al., 1996; Carracedo et al., 2001). The Intermediate and Holocene rift volcanic series are dated at 158-76 and $<$ 50 Ka respectively (Fuster et al., 1993; Guillou et al., 1996). During the collapse of Tiñor, about 882 Ka, more than half of the northwest flank of the edifice was removed (Carracedo et al., 1999a, b). This collapse was followed by the construction of El Golfo Volcano, which filled the depression between 550-130 Ka (Guillou et al., 1996) or 545 to 126 Ka (Carracedo et al., 2001). Carracedo et al. (2001) also inferred that the construction of El Golfo coincided with the maximum development of the Cumbre Nueva rift zone on La Palma.

The collapse of El Golfo Volcano gave rise to the El Golfo Embayment (Figure 9). This embayment has 1400 m high walls partially covered by younger lavas that run down to the sea (Hausen, 1973). Offshore is the El Golfo Debris Avalanche (Figure 10). The age of the collapse of the El Golfo Volcano is yet to be resolved. Masson (1996) and Urgeles et al. (1997) have proposed that the collapse took place between 9–15 Ka or 10–17 Ka. Carracedo et al. (1999b) suggested that the failure occurred much earlier, $\sim$130 Ka.

Holcomb and Searle (1991) interpreted the El Julan Embayment as a collapse structure associated with an event that Masson (1996) suggested took place about 500 to 300 Ka. Carracedo et al. (1999a, b) proposed that failure took place when El Golfo Volcano was well developed at 130 Ka. More recently Carracedo et al. (2001) argued the presence of lavas belonging to the rift volcanism in the water galleries in the Julan Embayment places the minimum age of its collapse about 150 Ka.

Masson et al. (2002) suggested that Las Playas Embayment is not the result of a landslide, but of massive slumping along the San Andrés Fault system (Figure 10) between 545 and 261–176 Ka. Although inactive, the lateral collapse structure is constrained to have formed between the last emissions of El Golfo volcano at about 176 Ka and the rift lavas cascading down the collapse scarp dated at 145 Ka (Guillou et al., 1996).

The ages of the offshore extensions of the rift zones of El Hierro are determined from the debris avalanches constrained by them. We assume that Carracedo (1999a, b) is correct in stating that the El Golfo and El Julan failures took place about 130 Ka and that the Playas I and II slumps and debris avalanches took place at 546–178 Ka and between 176 and 145 Ka respectively (Masson et al., 2002). These avalanches appear to onlap the offshore rift extensions and do not extend beneath them. This then implies that the submarine sections of the rift zones must have been in existence by at least 130 Ka and possibly as early as 546 Ka or even earlier. Such ages suggest that the rift zones originated during the formation of Tiñor Volcano at 1.12–0.882 Ma, and were subsequently rejuvenated during the construction of El Golfo edifice 550 to 130 Ka.

Tiñor and El Golfo Volcanoes may have been built under different stress regimes, with the older one affecting the gravitational stress field of the younger one. Processes associated with the collapse of Tiñor and post-collapse erosional events also possibly affected the morphology of the submarine rifts. Since rift zone volcanism continued after the collapse of El Golfo at 130 Ka, the main magma chamber associated with Tiñor and El Golfo Volcanoes must continue to persist and marks the location of the supposed Canary Island hotspot.

The greatest concentration of eruptive vents in the Canary Islands is observed on the central tableland formed by the intersection of the three rift zones on El Hierro. All of the cones consist of loose volcanoclastic material. There are no obvious set of fractures that controlled the sites of extrusion. The cones are found at different elevations, and most are of Pleistocene age with only one of them having been active in historical time (Hernández Pacheco, 1982). Carracedo et al. (2001) dated the building of the cone at 145 Ka to 2.5 Ka with some activity extending to historical time. They further suggested that this period of volcanism is coeval with the maximum development of the Cumbre Vieja rift zone in La Palma.

The Quaternary volcanic cones are on average 100 m high, have flanks that are furrowed by erosion, and lavas and tuffs that are covered by small bushes or have been converted into tilled fields. Scattered over the tableland are a few cones free of vegetation and a few young cones such as Montaña de Tenerife and Mal Paso (Figure 3). Most of the latter cone, located along the scarp of El Golfo Embayment is missing, having slid into deep water (Hausen, 1973).

Northeast and Northwest Rifts: Onshore sections

The central volcanic highland developed at the intersection of the three rift zones, descends westward to a plain inclined to the north. This plain represents the flank of the former El Golfo Volcano. It is covered by lavas originating from cones in the central upland to the south and is cut by the 1400 m high cliffs (Figure 3) on the west. The east slope of the highland is divided into steps that are covered with cones and descend to the sea. According to Hausen (1973) the concentration of cones on the 12 km long by 7 km wide northeast rift attests to the intensity of eruptions along this rift. Most of the cones are Quaternary and have undergone erosion so that relief of the region is a composite of volcanism, erosion and faulting. Hausen (1973) speculated that there is evidence of more recent volcanism along the northeast rift. Among these recent volcanic edifices is Tesoro de Tamaduste (Figure 3) located along the north coast. Lavas from this double-crater volcano, open to the north, cascaded down the seacliff covering a platform at the foot of the cliff. Another young volcano is Solimán (Figure 3) found along the crest of the rift. A carbonized pine chunk enclosed in the lavas from on the crest of the rift yielded an age of 2900 $\pm$ 130 years B.P. Guillou et al. (1996) obtained an age of 2500 $\pm$ 70 years B.P. for another volcano along the crest stated that there are many vents of similar and younger ages in the rifts of El Hierro.

The southern side of the 10 km long x 4 km wide northwest rift is composed of ash, lapilli, and young lavas that flowed south from vents located in the central tableland. These flows extended to the shoreline, accounting for the smooth slopes of El Julan Embayment. The absence of barrancos and weathering indicate that the lavas of El Julan are young (Hausen, 1973). According to Carracedo et al. (2001) the lava flows filling the El Julan Embayment have been dated between 41 and 31 Ka. West of the ash-covered southern slope of the northwest rift is an extensive field of volcanic cones of Quaternary and Holocene age extending from the crest of the central tableland to the rocky west coast of El Hierro. Between the cones are lavas, the youngest of which are to the west. The rift slopes down to the west coast along a series of steps of older lava sheets in the upper steps and younger lavas and cones in the lower ones. Hausen (1973) stated that a cone near the coast looks so fresh that it must be recent; possibly from the time the Spaniards came to the region in the 15th century. A southeast valley in the region of this cone is filled with recent lavas and spatter cones whose outflows have caused progadation of the west coast. At the northwest corner of the northwest rift is the western end of the cliff of the El Golfo Embayment. The cliff is composed of lavas erupted along fractures at its crest. Here the cliff is 200 m high and lavas from young cones along its crest have cascaded down the cliff to the terrace at its foot. One of the cones along the top of the cliff has been so eroded by mass wasting that only the cone's lava-filled conduit has survived. The southwestern peninsula of El Hierro descends toward the coast and on reaching it is broken into a steep cliff. Hausen (1973) speculated that this cliff was formed prior to the construction of the volcanic cones scattered throughout the region. Also present in the region are boulders displaced from the adjacent central tableland.

Northeast and Northwest Rifts: Offshore sections

The offshore extensions of the Northeast and Northwest Rifts display a different morphology from the submarine South Rift and the submarine rift off southern La Palma. The Northeast and Northwest Rifts are lobate in plan view compared to the others, which are well defined ridges (Figures 9 and 10). They both widen down slope, becoming more diffuse with distance from the shoreline.

The submarine section of the Northeast Rift is about 26 km wide and 15 km long and extends to a water depth of 3100 m. It has an average along-axis topographic gradient of about 400 m/km compared to the onshore section that has a gradient of half that, 200 m/km (Figure 11). Local steeper gradients range from 600 m/km to 300 m/km and occur in a series of steps that extend down to about 2600 m water depth. Scattered over the rift surface are 30 volcanic cones whose diameters range from 300 m to 3 km and heights range from 25 to 250 m. Pinnacles, less than 50 m wide and up to 150 m in height, also occur along the rift. According to Gee et al. (2001) these pinnacles, which are sometimes in groups of two or three elongated or aligned down the rift, represent dike injections radiating from a central volcano. Wide expanses of sea floor lacking any topographic irregularities are also observed in the submarine rift (Figure 9). In the absence of bottom photographs or side-scan sonar it is impossible to discern the nature of this terrain, whether it represents a sedimented seafloor or lava flows. However, its embayment-like morphology at their heads suggest that their form is due to gravitational mass-wasting processes.

The flank of the submarine Northeast Rift, adjacent to the El Golfo Debris Avalanche, is linear and has the appearance of a multi-peaked, narrow, steep-sided ridge with a flank gradient of 400 m/km. In contrast the contact between the Northeast Rift and Las Playas Debris Avalanche is more discontinuous, with one ridge section extending from a water depth of 600 to 2200 m, separated by a gap from another ridge section offset to the north that extends from a depth of 2200 to 3100 m. The rift lacks a well defined front and instead takes the form of a series of lobes. Most of them appear to represent lava flows similar to the lava terraces seen along the flanks of Hawaiian rift zones (Smith and Cann, 1999). Others may be debris flows associated with landslides.

The submarine Northwest Rift zone is 26 km wide and 16 km long with its distal end located at a depth of 3100 m. Beyond this front is a several-kilometer-wide apron containing what we infer to be a few volcanic cones. The along-axis slope of the offshore extension of the Northwest Rift has a gradient ranging from 300 m/km near the coast, a gradient of 200m/km in its mid-section, and a gradient of 200–300 m/km at its distal end. Its onshore segment has a gradient of only 200 m/km. The flank of the submarine northwest rift bordering the El Golfo Debris Avalanche has a gradient of 300 m/km, whereas its gradient toward El Julan Debris Avalanche (Figure 10) to the south is much gentler with a gradient of 200 m/km. Twenty five volcanic cones ranging in diameter from 0.5 to 1.8 km and ranging in height from 50 to 275 m have formed within the rift zone. As in the submarine Northeast Rift, the volcanic edifices string together to form irregular lineations downslope. Most of the edifices are concentrated on the north side of the rift, off the field of recent lavas at the northwest tip of the island. As mentioned, Hausen (1973) inferred that the lavas on land were probably erupted along fractures associated with the western end of the cliff located along the south side of El Golfo Embayment. The lobes associated with the volcanic edifices can be traced to a water depth of 3500 m (Figure 9).

The topography of the southern side of the submarine rift is smoother than its northern side and has fewer cones and those that are present are smaller in size. The north side of the rift is characterized by northwest-trending lineaments capped by volcanic cones hundreds of meters high. A west-trending, poorly defined landslide that has a topographic expression to a water depth of 3100 m separates the terrains along the north and south sides of the rift. The distal edge of this side of the rift is marked by flat-topped lobes trending southwestward ending at a scarp with a gradient of 300 m/km.

South Rift: Onshore section

The 6 km wide by 8 km long subaerial South Rift is dominated by the Restinga Peninsula, a broad highland whose elevation decreases southward. The Las Playas Scarp forms the eastern edge of the peninsula and the El Julan Scarp marks the western edge (Figure 3). The peninsula is covered by cinder cones and lava fields (Hausen, 1973). The volcanic features vary in age, with some being weathered and in part forested, whereas others have a fresh appearance with black lava and ash fields. The northern part of the Restinga Peninsula is overgrown with bushes and tilled fields, but farther south the peninsula becomes more desert-like and still farther south it is covered by barren fresh lava fields. To the west the lavas have reached the coast, and to the east eruptions occurred before the final formation of the Las Playas Scarp (Hausen, 1973). Apparently the peninsula was created by repeated lava flows from the north. Hausen (1973) speculated that the numerous volcanic extrusions were a consequence of fractures that served as passageways for the emerging magmas.

South Rift: Offshore section

The ridge-like submarine South Rift, also known as the Southern Ridge, curves southwestward from its mid-section outwards and is asymmetrical in cross-section with its steeper side facing southeast (Figure 11). The width of the rift ranges from 3 km on its proximal end to 18 km at its distal end. Gee et al. (2001) described the crest of the South Rift as a remnant dike swarm. According to Gee et al. the tip of the south rift zone at 27°20′ N is over 1 km high, its flanks have slopes of over 600 m/km and its base is at a depth of 3700 m. Morphologically it resembles the southern tip of the submarine Cumbre Vieja Rift off La Palma.

A saddle near 27°30′ N divides the South Rift in two. The ridge section north of the saddle consists of a sequence of narrow volcanic lobes trending southeast to southwest that extend to a depth of 2500 m (Figure 12). These lobes have prograded over a probable lava terrace exposed on the southeast edge of the north segment. This terrace probably represents an older rift construction comparable in age to the southern segment, whereas the lobes are recent lava flows, probably offshore extensions of the young lava flows at the southern tip of the Restinga Peninsula. Both of these flow units superimposed one upon another are probably connected with the same source in El Hierro. Southeast of the lava terrace is a southeast trending apron whose base is at a water depth of 3200 m. Extending southwestward from the southern edge of the lava terrace is a several kilometer wide ridge, with a minimum depth of less than 1500 m.

The east-west trending, 400 m wide saddle at 27°30′ S is flanked on the east side by a smooth slope with a gradient of 500 m/km and whose base is at a depth of 3200–3400 m. On its western side the saddle is blocked by a 6 × 3 km wide terrace whose western side is gullied. South of the saddle is a 1.0–1.5 km wide ridge whose flank gradients are on the order of 400–500 m/km. The western side of the ridge is irregular with a <1 km wide east-trending ridge that can be traced from the west flank of the South Rift at a depth of 1700 m to a depth of 2600 m, over a distance of 3 km. The eastern side of the submarine South Rift is cut by gullies about 1.5 km long and 75 m wide.

Hawaiian Islands

General geology of the Island of Hawaii

The dimensions of Hawaii are roughly 127 km by 148 km; it has a surface area of about 10,000 km^2, and according to Moore and Clague (1992) the island has grown at a rate of 0.02 km^2/yr in the past 600 k.y. The Island of Hawaii consists of five coalescing volcanoes that in chronological order are: Kohala at the northwest tip of the island that completed its shield phase 245 Ka, Hualalai along the west coast which completed its shield phase 130 Ka, Mauna Kea southeast of Kohala which completed its shield phase 130 Ka, Mauna Loa slightly south of the center of the island which rose above sea level about 300 Ka and Kilauea located east of Mauna Loa which rose above sea level about 200 Ka (Moore and Clague, 1992). Hualalai, Mauna Loa and Kilauea volcanoes are historically active.

The rift zones on the Island of Hawaii (both on- and offshore components) have been well studied compared to these of the Canary Islands. Therefore, more types of data and information exist for the Hawaii rifts and more rigorous conclusions can be made about the nature of the morphologic features we observe. The locations of the rift zones (Figures 2 and 13) associated with the five volcanoes that make up Hawaii are controlled by gravitational stresses within each individual volcano, which in part are controlled by the morphologies and positions of pre-existing volcanoes (Fiske and Jackson, 1972). The characteristics of individual rift zones are variable since they are influenced by the distribution of neighboring volcanoes (Fornari, 1986). For example, the Southwest Rift Zone of Kilauea is sandwiched between Loihi Seamount and Mauna Loa's Southwest Rift Zone. It is only 32 km long on land and poorly developed offshore (Holcomb, 1987). As another example, the 55 km long subaerial portion of the East Rift Zone of Kilauea is buttressed to the north by Mauna Loa's flank, but free to spread to the south. Thus intrusion of magma into Kilauea's currently active East Rift Zone (ERZ) is leading to slumping and displacement of the south flank of the island towards the sea (Swanson et al., 1976; Morgan et al., 2000; Hills et al., 2002 and references there in).

As in the Canary Islands, the subaerial rift zones in Hawaii have seaward extensions forming salients extending from the shoreline to the base of the volcano at abyssal depths. The offshore sections of the rifts can vary in length: 55% of the 130 km long ERZ is below sea level; 61% of the 107 km long south rift of Kohala is submerged; ∼28% of the 108 km long south rift of Mauna Loa also is below sea level (Moore and Chadwick, 1995). As in the Canary Islands the longitudinal and lateral slopes of the submarine extensions of the rift zones are steeper than their subaerial counterparts (Figures 6, 11 and 14).

Along the crest of the onshore portions of the rifts, volcanic cones are commonly built and then eroded by subsequent eruptions. Lavas transported down the flanks of the subaerial rift zones form smooth,

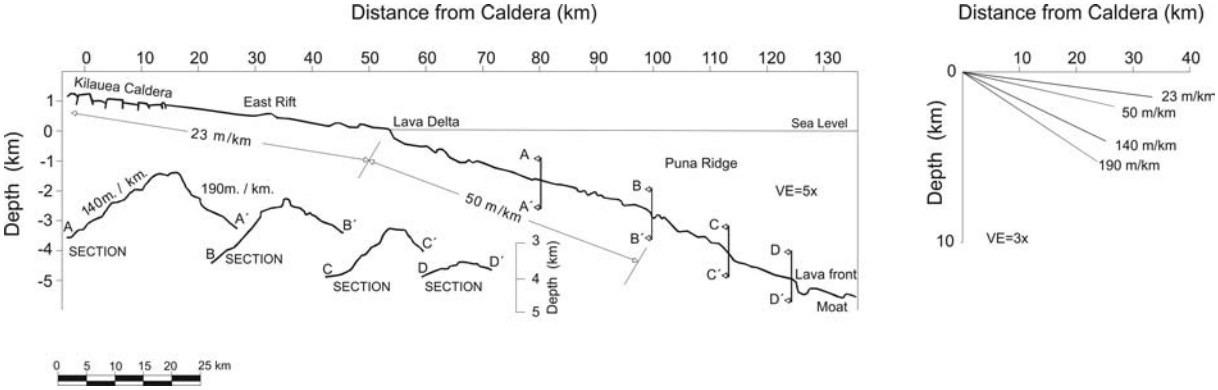


Figure 14. Profiles of Kilauea's subaerial East Rift and its seaward extension, the Puna Ridge. Vertical exaggeration 5X. Modified from Lonsdale (1989). Also included is a diagram showing the gradients displayed by the rift at a vertical exaggeration of 3X, an exaggeration similar to that of the profiles f the rifts of La Palma and El Hierro in Figures 6 and 11.

low-angle slopes, while lavas that cross the shoreline are believed to feed submarine debris flows (e.g., Moore et al., 1973). By contrast, the crest and flanks of the submarine extensions are covered by cones and terraces that have diameters of up to 1 km or more and sides up to several hundreds of meters high (Clague et al., 2002; Smith et al., 2002b). Many of the cones have craters in their tops that are 100s of meters in diameter.

Kilauea

East Rift Zone: Onshore section

Kilauea's rift zone system is one of the best studied in the world (Tilling and Dvorak, 1993, and references therein). The volcano is fed from a central magma chamber (or system of magma conduits) beneath the summit (Ryan et al., 1981; Ryan, 1988). Lava is erupted at the summit and/or one of the volcano's two rift zones, the Southwest Rift Zone (SWRZ) and the ERZ. Typically, the onset of a rift zone eruption is marked by seismicity that migrates from the summit region down a rift zone to the site of eruption. The early phase of eruption is normally through a fissure as long as several hundreds of meters (e.g., Klein et al., 1987; Wolfe et al., 1987). If fissure eruptions persist, they normally become confined to a single vent.

Since 1983, eruptions have occurred continuously along the ERZ centered at either the Pu'u 'O'O or Kupaianaha vents with more than 1 km^3 of lava being erupted since 1983 (Wolfe et al., 1987; Mangan et al., 1995). Surface deformation associated with seismic activity (Pollard et al., 1983), the fissure eruptions, and the observation of dikes within the eroded cores of Hawaiian volcanoes (Walker, 1988) indicate that rift zone eruptions are dike-fed and that the seismic activity is associated with magma moving through the underlying magma conduit system (Rubin and Gillard, 1998). The zone of eruptive fissures ranges in width from 1.5 to 3.0 km (Holcomb, 1987; Moore and Trusdell, 1991).

The ERZ of Kilauea extending from the southeast side of the summit caldera trends southeastward for ~15 km and then changes its trend towards the northeast (N65°E), an orientation that it maintains to the shoreline. The along-axis gradient is about 23 m/km, a uniform gradient that led Lonsdale (1989) to infer that as a result of high eruption rates the rift has been able to maintain a more or less constant, gentle profile. Superimposed on the long-axis profile are secondary features such as pit craters, cones, spatter ridges, low mounds, and phreatomagmatic ash near the rift's axial zone. Parfitt et al. (2002) classified eruptions along the ERZ into five categories: short-lived fissure eruptions that last less than five days and often less than one day, consisting of linear spatter ramparts and lava flows; longer-lived fissure eruptions of ~5 to 15 days duration, consisting of linear spatter ramparts and lava flows; fissure eruptions evolving into a central vent; episodic eruptions; long-lived steady state eruptions. They ascribe these different eruption types to differing thermal and driving pressure behavior in the feeder dikes. Epp et al. (1983) proposed that ERZ eruptions tend to drain the summit magma reservoir to pressure levels corresponding to the elevations of the eruptive vents.

East Rift Zone: Offshore section

The 70 km long Puna Ridge is the submarine extension of the onshore ERZ (Figures 15 16 and 17). Unlike

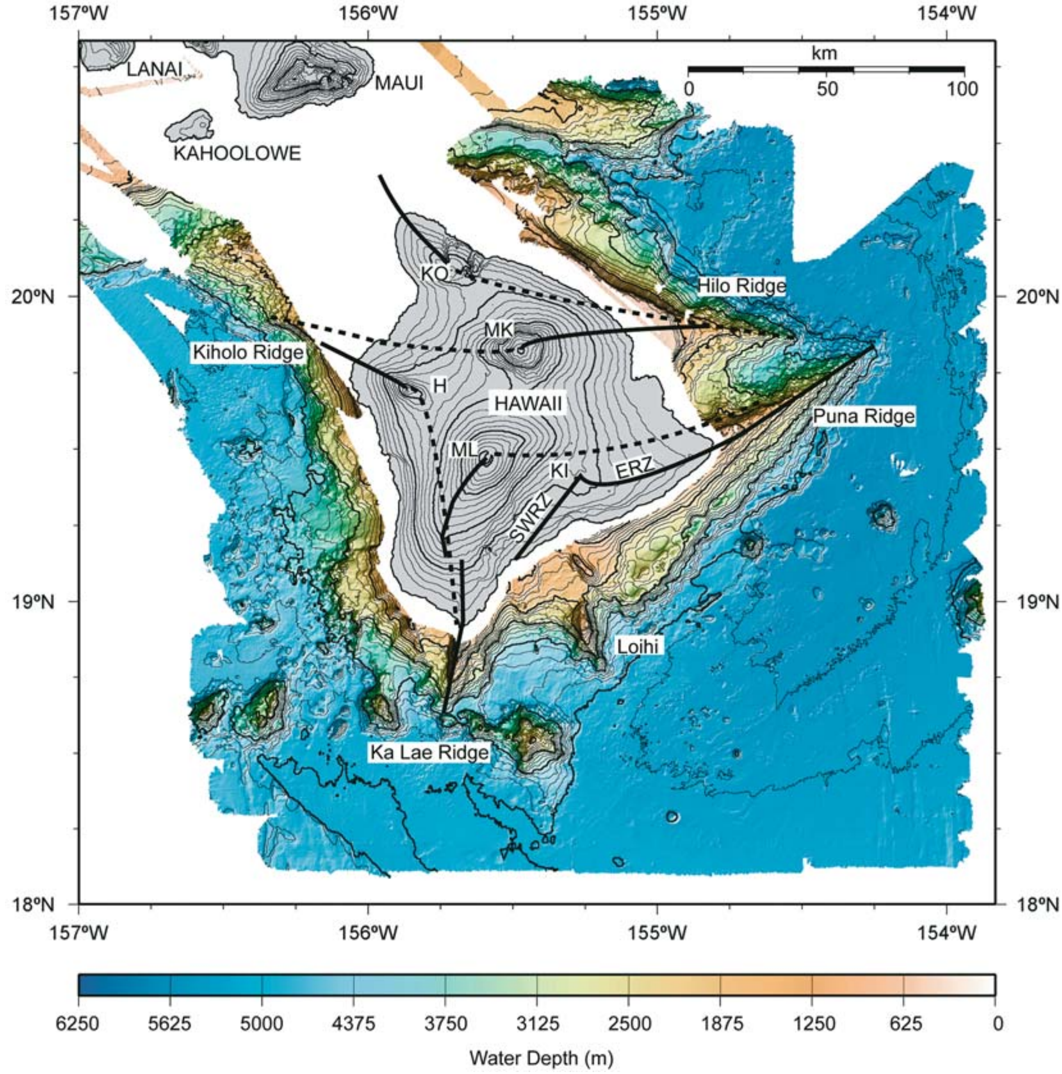


Figure 15. Topographic/bathymetric map of Hawaii and the surrounding seafloor, modified from Smith et al. (2002b). Volcanoes are labeled as: KO-Kohala; H-Hualalai; MK-Mauna Kea; ML-Mauna Loa; KI-Kilauea. Solid lines: distribution of rift zones from Fornari (1986). Dashed lines: distribution of rift zones from Holcomb et al. (2000). The East Rift Zone (ERZ) and Southwest Rift Zone (SWRZ) of Kilauea are labeled. Offshore extensions of the rift zones also are labeled. Loihi, the youngest Hawaiian volcano is located to the south of Kilauea.

the submarine extensions of the rift zones off the Canary Islands, the submarine rift zones of Hawaii have recently been the site of several high-resolution studies, and we are much more confident about the origin of submarine features here than at the submarine rift zones off shore La Palma and El Hierro. Multibeam bathymetry data for the entire Puna Ridge (Clague et al., 1994; Monterey Bay Aquarium Research Institute (MBARI), 2000), along with high resolution side-scan sonar images (Lonsdale, 1989; Smith et al., 2002a), photographic imagery (Moore and Fiske, 1969; Clague et al., 1988; Lonsdale, 1989; Smith et al., 2002a; Parfitt et al., 2002; Gregg and Smith, 2003), submersible dive observations (Fornari et al., 1978; Clague et al., 2000; Johnson et al., 2002), and GLORIA side-scan sonar data (Holcomb et al., 1988), verify that the Puna Ridge crest is the location of dike intrusions and fissure eruptions. Sea-surface magnetic data show a normally polarized anomaly centered over the ridge axis, consistent with the presence of a 11 km wide, 70 km long, nearly-vertical magnetic source, presumably representing a dike complex at depth (Malahoff and McCoy, 1967). High-resolution magnetic data show regional highs in the crustal magnetization along the ridge axis, perhaps

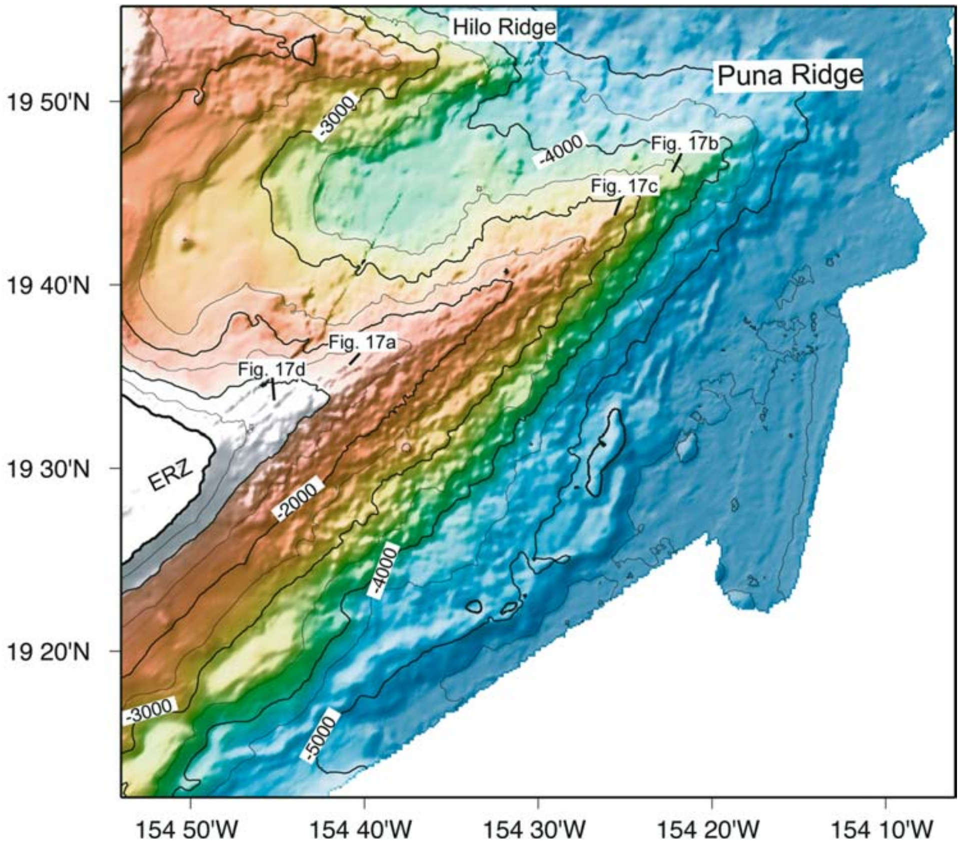


Figure 16. Color shaded relief bathymetry of the Puna and Hilo Ridges showing locations of the photomosaics in Figure 17. Contours in meters. The East Rift Zone (ERZ) is labeled. Modified from Smith et al. (2002a).

indicating areas of recent lava deposition (Smith et al., 2001).

Relatively few dike intrusions have been recorded from the summit of Kilauea to the Puna Ridge in historic time. Submarine lavas from the crest and flanks of the Puna Ridge appear to be older than most subaerial lavas. Holcomb (1987) estimated that 70% of the subaerial portion of Kilauea is younger than about 500 years. Based on palagonite thicknesses, Clague et al. (1995) estimated that dredged lavas from the Puna Ridge are 700–24000 years old, but mostly 2000–7000 years old. The most recent submarine eruptions are thought to have occurred in 1790, 1884 and 1924 (Stearns and Macdonald, 1946). The 1884 eruption was witnessed just offshore at 20 m water depth. In 1790 and 1924, magma withdrawal from the summit reservoir is inferred to have fed submarine eruptions on the Puna Ridge (Stearns and Macdonald, 1946). There has been a lack of seismicity along the ridge since 1960, suggesting recent inactivity (Klein et al., 1987).

The average along-axis slope of the Puna Ridge is more than three times as steep as that of the onshore section (∼73 m/km vs. 23 m/km, respectively) (Figure 14).The flank slopes of the Puna Ridge range between 160–275 m/km (Fornari et al., 1978; Smith and Cann, 1999), also about three times steeper than on the onshore ERZ (∼51 m/km). The slopes on the south flank of the Puna Ridge do not change significantly with distance from the shoreline until the crest is below 3500 m water depth. The slopes are steepest on the north flank between about 1200–2000 m water depth. Multibeam bathymetry data (e.g., Swanson et al., 1976; Moore and Chadwick, 1995) and high-resolution side-scan sonar imagery (Smith et al., 2001, 2002a) suggest a landslide scarp in this region.

Moore and Chadwick (1995) identified slump deposits at the southern base of the Puna Ridge. The deposits consist of anomalous terraced terrain that extends from about 4000 m water depth down to the base of the ridge where it meets the Hawaiian Deep at about 5500 m. They suggested that the terraces represent the

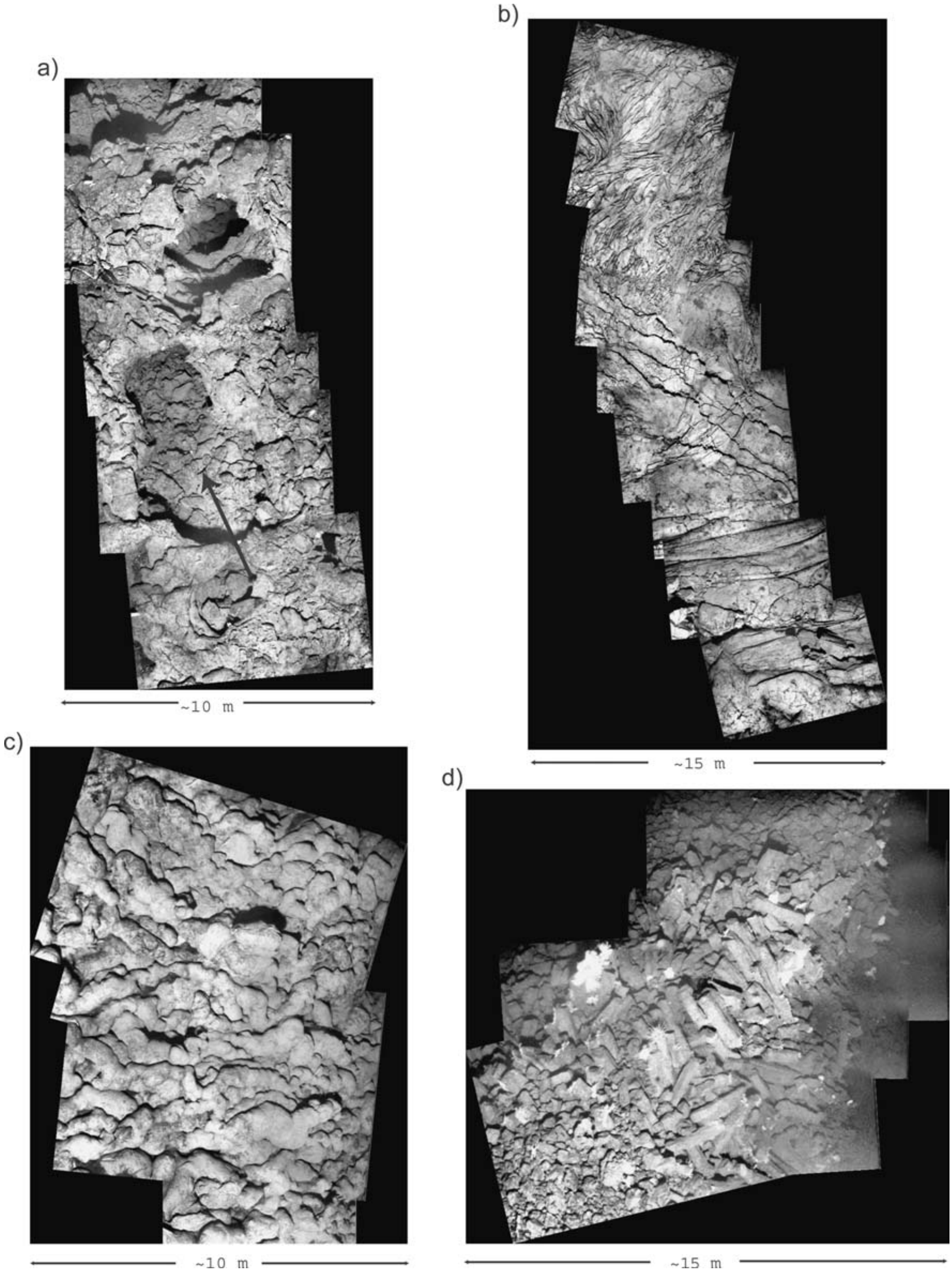


Figure 17. Photomosaics of bottom photographs obtained using a deeply towed photo-imagery system. Figure modified from Smith et al. (2002a). Locations of photomosaics are marked on Figure 16. (a) skylights in a lava tube, surface flows on top or a lava terrace. Black arrow marks a large skylight. (b) drapery folds in lava on the top of a terrace located at about 3300 m depth. The morphology is similar to that observed on the subaerial ERZ. (c) flanks of a pillow cone, covered with tubular pillows; downslope is to the right. (d) columnar joints that have tumbled out of a breach in the crater of a cone located at about 900 m water depth.

distal margin of a massive slump block that is now mostly covered by volcanic flows from the summit of the Puna Ridge. No slump deposits have been identified at the base of the north flank of the ridge (Moore and Chadwick, 1995) although they may be covered or modified by lavas from either the Puna Ridge or the submarine Hilo Ridge to the north (Moore and Chadwick, 1995).

From sea level to about 2100 m depth, the axis of the Puna Ridge trends 65° (Lonsdale, 1989), in line with the lower ERZ. The axis has a broad, 3-4 km wide crest, with a small (about 1 km) right-stepping offset near a water depth of 2000 m. Volcanic constructions, up to 140 m high, are scattered along the ridge crest, especially at the shallower end. From about 2100–3400 m depth, the axis of the Puna Ridge bends to 45° (Lonsdale, 1989), and the crest narrows. At about 2700 m depth, the axis has a right-stepping offset and the along-axis slope increases. Using a towed side-scan instrument, Lonsdale (1989) observed tightly clustered fissures and collapse pits concentrated in an inflated portion of the ridge above the offset, and suggested that intrusive rather than extrusive magmatism predominates here. Below 3400 m depth the ridge crest is poorly defined. The axis is composed of 'steps' that Lonsdale (1989) interpreted as flow fronts steepened by mass wasting.

The styles of volcanic features on the lateral slopes of the ERZ change significantly just as the crest of the rift dips below sea level. As mentioned above, lavas erupted from the onshore ERZ form smooth, low-angle slopes, except where interrupted by faults. Gently dipping low-relief lava flows mantle the slopes. In contrast, volcanic features on the flanks of the Puna Ridge are common and include flow fields, pillow ridges, cones, and large semicircular flat-topped features (terraces) that have diameters of 1 km or more and sides several hundreds of meters high (Smith et al., 2002a). These terraces, many with pits on their tops, often appear to form staircases, one on top of the next.

On a finer scale, three basic flow morphologies are observed: pillow, lobate, and sheet flows (Figure 17). All three are found in each of seven camera sites at the Puna Ridge at water depths ranging between 660–3700 m (Smith et al., 2002a). GLORIA images (Holcomb et al., 1988) and ground-truth sampling and photography (Clague et al., 1988) indicate that sheet flows cover large areas north and southeast of the ridge base.

Kilauea's Southwest flank

At Kilauea's ERZ diking is accommodated by seaward slipping of the south flank on a basal décollement (Dieterich, 1988; Delaney et al., 1990; 1993) that is located 7-9 km beneath the surface (Owen et al., 1995). The north flank of the ERZ is relatively stable and does not appear to respond to diking events, presumably because it is buttressed by Mauna Loa Volcano and/or because seaward displacement of the south flank is favored by the pre-existing gravitational stress field imposed by Mauna Loa (Swanson et al., 1976). The average "spreading rate" of the ERZ is estimated to be $\sim$10 cm/yr, the rate that the south flank is slumping (Swanson et al., 1976). Between the SWRZ and ERZ of Kilauea are the Koae and Hilina Fault Systems. These fault systems are inferred to be associated with southward displacement of the south flank (e.g., Parfit and Peacock, 2001). Structurally the Hilina Fault System is comparable to the San Andrés Fault System between the South and Northeast rifts on El Hierro, Canary Islands. It is along this fault system that the Las Playas I and II slumps in the Las Playas Embayment became detached from the rest of the Northeast Rift.

Some workers have proposed that the southward displacement of the southeast flank of Kilauea is due to dike injection into the ERZ leading to compressive stresses in the rift zone flank that are relieved by seaward displacement (Duffield, 1975; Thurber and Gripp, 1988; Morgan et al., 2000). Others have inferred that the displacement of the southeast flank is due to gravitational spreading and forces exerted deep within the volcano by the accumulation of magma (Denlinger and Okubo, 1995). There is likely some balance reached between the horizontal stress generated by dike intrusion, the weight of the flank, and the friction on the basal décollement to produce the landslides and slumps that are observed.

Southwest Rift Zone

Kilauea's SWRZ extends 32 km from the summit caldera to the shoreline. It has been suggested that the SWRZ acts as a headwall for the south flank slump. The SWRZ is $<$1.5 km wide in its northern section. The rift zone changes trend from 230° to 195° $\sim$15–20 km from the summit caldera. The southern section of the SWRZ is dominated by the Great Crack, a 15 km long pre-1790 AD fissure and graben system, and the site of an 1823 eruption (Holcomb, 1987).

West of the upper end of the crack are two 30 m high normal fault scarps (Puu Nahaha Faults) mantled by lavas as young as 500 years old. South of the Nahaha Faults the rift zone widens to at least 5 km. According to Dzurisin et al. (1984) only 10 % of the magma supply to the summit of Kilauea from July 1956 to April 1983 was injected into the SWRZ, 55% was intruded into the ERZ, and 35% was erupted at the summit. The submarine extension of the SWRZ is poorly defined.

Mauna Loa

The Onshore and Offshore rift sections

Mauna Loa is an active volcano on the island of Hawaii. Mauna Loa rose above sea level about 300 Ka and covers the largest area of the island, about 5000 km^2 (Moore and Chadwick, 1995). At present Mauna Loa is erupting a greater portion of lava above sea level than on its submarine sections. Mauna Loa has two rift zones: the Southwest Rift Zone (SWRZ) and the Northeast Rift Zone (NERZ) (Figures 13 and 14). The 65 km long subaerial SWRZ extends another 35 km offshore (García et al., 1995). The submarine section of the SWRZ is called Ka Lae Ridge (Holcomb et al., 2000). The NERZ does not reach the shoreline, or has been covered by Kilauea lavas (e.g., Holcomb et al., 2000). Here we concentrate on the SWRZ.

The subaerial section of Mauna Loa's SWRZ is disrupted by a 40° change in trend near an elevation of 2400 m. This may be due to the interaction of the rift zone with the growing Kilauea edifice, which has inhibited the inflation of Mauna Loa (Lipman, 1980). Presumably, the presence of Kilauea prevents slumping of the southeast flank of Mauna Loa, causing the SWRZ to migrate westward, and leading to the bend in the middle section of the rift. South of the bend, the rift zone is defined by the fault line scarp Kahuku Pali. The fault has a relief of about 170 m subaerially and it increases dramatically to about 1800 m of relief offshore (Fornari et al., 1979). It extends offshore for about 50 km (Fornari, 1986). The fault cuts through the Ka Lae Ridge, which was created from the intrusion and eruption of dikes along the SWRZ. Submersible observations coupled with bathymetric data show that the top of the Ka Lae Ridge does not have any volcanic cones; this is unusual for submarine rift zones of Hawaii (Clague et al., 2000). In order to explain this observation and the characteristics of the magnetic dipole over the submarine ridge, García et al. (1995) have suggested that the scarp did not cut the ridge at its axis, but rather the scarp cut the eastern flank of the Ka Lae Ridge and its axis is offset to the west. Recently, Holcomb et al. (2000) suggested that Ka Lae Ridge is part of Hualalai Volcano rather than Mauna Loa, although this remains to be proven.

Kohala

The Offshore section of the Southeast Rift

The Kohala Volcano, at the northwest tip of Hawaii, completed its shield phase 245 Ka (Moore and Clague, 1992). The deeper sections of the submarine Hilo Ridge, extending from the east coast of Hawaii east of Mauna Kea, have recently been hypothesized to be the continuation of the southeast rift zone of Kohala (Holcomb et al., 2000; Kauahikaua et al., 2000). Previously, the entire Hilo Ridge had been interpreted as the extension of the east rift zone of Mauna Kea (e.g., Moore and Clague, 1992).

The volcanic morphology of the 60 km long Hilo Ridge varies along its axis. The proximal part of the ridge (out to about 20 km from the shoreline) is dominated by two submerged terraces. The leading edge of the shallow terrace is at a depth of about 400 m; the leading edge of the deeper terrace is at about 1100 m. The shallow terrace is well known and thought to be associated with Mauna Kea (Moore and Clague, 1992). The deeper terrace is thought to be related to a terrace at similar depths that extends around the northern section of Kohala Volcano and that we speculate to document the end of Kohala's shield phase at about 245 Ka.

If this correlation is correct then the segment of Hilo Ridge deeper than 1100 m is older than Mauna Kea (Holcomb et al., 2000). Such a correlation is supported by the isotopic composition of basalts from the terrace that are distinct from those from Mauna Kea, but are similar to those from Kohala. This conclusion also is supported by three-dimensional gravity modeling that shows a gravity high extending from Kohala, across the northeast slope of Mauna Kea to Hilo Ridge (Kauahikaua et al., 2000). The morphology of Hilo Ridge deeper than 1100 m is very similar to that of the Puna Ridge. Flat-topped volcanoes and terraces mark its crest and flanks (Clague et al., 2000). The average along-axis slope for the ridge deeper than 1100 m is 60 m/km. The flanks have slopes of ~200 m/km. Both of these values are similar to those at the Puna Ridge. It seems that both submarine ridges were built by the same means: dike injection and eruption in a narrow axial zone to build a distinctive volcanic ridge.

Discussion

Studies of the Hawaiian Islands have led to a clearer understanding of how oceanic islands evolve. These investigations have demonstrated that oceanic islands are constructed not only by summit eruptions from shallow magma reservoirs, but by dike intrusions and fissure eruptions along well-defined rift zones. Most oceanic island volcanoes have radial rift zones of varying lengths, some attaining lengths of over 100 km. Rift zones may develop along fractures resulting from the stress field generated by an upwelling plume (Wyss, 1980). Rift zone locations also are strongly influenced by the gravitational stress fields created within an individual growing volcano modified by the stress fields of neighboring volcanoes (Fiske and Jackson, 1972). Once a rift zone is initiated its development is encouraged by the concentration of tensional gravity stresses along the rift zone axis and by rift zone flanks spreading across the underlying oceanic crust (e.g., Borgia and Treves, 1992).

Several generalizations can be made about the morphologic similarities and differences between the volcanic rift zones of La Palma and El Hierro in the Canary Islands and those of the Island of Hawaii that may illuminate similarities and differences in magmatic processes between them. In general, the rift zones on La Palma and El Hierro are shorter (a few tens of kilometers in length) than those on Hawaii (ranging up to >100 km in length). One possible explanation for this is the composition of the magma. In Hawaii, the shield-phase lavas are tholeiitic and the post-erosional lavas are alkali basalts, whereas the shield and post erosional phase lavas in the Canary Islands are both alkali basalts (Carracedo et al., 2001). We infer that as alkali basalts are more viscous than tholeitic liquids, encourages building of the volcano upwards rather than outwards. An important consequence of this is that volcano flank slopes are steeper, which in turn leads to catastrophic collapses of the volcano. The dramatic embayments on La Palma and El Hierro are a result of such massive flank failures. Another possibility is that dikes are able to propagate further, for a number of reasons including the effects of pressure within the magma pressure and gravity acting downslope along the rift.

Many of the rift zones on Hawaii have well defined axial zones, both in the on- and offshore portions. Typically rift volcanism builds a ridge that has a relatively narrow (a few kilometers wide) crest characterized by structures associated with dike intrusion and eruption: faults, volcanic flows and edifices, and collapse pits. In contrast, volcanism along some of the rift zones we have considered in this paper on El Hierro and La Palma is more diffuse and often it is difficult to define the lava pathways that fed the scattered volcanic edifices. This is particularly true for the Northwest and Northeast Rifts of El Hierro (Figures 9 and 10).

Submarine rift zones of Hawaii show some evidence of widening at their distal ends, where the stress field within the rift may be insufficient to trap dikes within a narrow region. At these places, the axis of the rift zone is hard to define and volcanic lineations fan out (Lonsdale, 1989; Clague et al., 1995). The widening observed at the distal end of rift zones in Hawaii, may be a small-scale model for the fanning of the Northeast and Northwest Rifts on El Hierro. This fanning is so pronounced in these rifts, however, that they closely resemble deep-sea sedimentary submarine fans.

Another possibility is that rift zones fan out down slope because dike injection and volcanism has shifted laterally. Such lateral migration of volcanism may be an important process at the Northwest Rift of El Hierro. The more recent igneous activity along this rift is concentrated along its north side where it appears that lava has flowed onto an adjacent debris flow. In contrast, the south side of the rift has the appearance of a sediment apron, suggesting that either it has not been active recently or that its mode of emplacement differs from that on the north side.

Lénat et al. (1986) suggested that the same processes that form the narrower, ridge shaped submarine rift zones in Hawaii have constructed the wide submarine rift zones on Réunion Island. That is, dikes injected from the summit reservoir move along a narrow zone and erupt along linear vents. The difference in the morphology of the rift zones between those in Hawaii and those in Réunion may be explained, according to Lénat et al. (1986), by the more complex history of volcano building in Réunion than in the Hawaiian volcanoes. Lénat et al. suggested that the presence of an older volcano beneath a younger one in Réunion may lead to a more complex distribution of gravitational stresses within the volcanic edifice leading to a broader zone of dike injection. This may be the case on El Hierro as well where El Golfo Volcano has been constructed on top of El Tiñor Volcano.

In Hawaii, the stress field associated with a volcanic edifice may remain more or less constant throughout its history. If another volcano begins to form nearby, a different stress field is superimposed on the first that is related both to the new volcano and the older one. In La Palma and El Hierro, this scenario may be slightly different. A volcano with its associated stress field is established. That volcano is destroyed and another one is created over the remains of the older one with its stress field superimposed on the older one. The orientation and magnitude of the new stress field may not necessarily be the same as the previous one, and thus may produce the apparent widening of rift zones down slope. If this is the case, it is important to note that in some places the old and new stress fields must be similar to produce narrow, ridge-like rifts (e.g., the South Rift of El Hierro, and Cumbre Vieja on La Palma) in addition to the wide rifts.

Why are the volcanoes in the Canary Islands built one atop another rather than next to each other? We assume that this is related to slow absolute plate motion in the region. In the Canary Island region the rate of motion is on the order of 1.9 cm/year, whereas in the region of Hawaii the absolute plate motion is five times faster, on the order of 10 cm/year (Carracedo, 1999).

Magma supply may also play a role in the width of rift zones. As a result of a low rate of magma supply, the orientation at which dikes are injected into the country rock may become random and not concentrated along a narrow rift zone. The reservoir pressure created by magma-supply mechanisms may be incapable of overcoming the compressive stresses along a previously intruded rift. The magma is forced to find other avenues of intrusion and extrusion into the country rock. As this new route is then closed, it in turn forces the magma to find other pathways creating a dike swarm fanning out from the magma body. This process may be responsible for the widening of the Northeast and Northwest Rifts in El Hierro, but also in Mauna Loa, a volcano that lacks large rift-zone faults and fissures and contains numerous radial vents outside its two rift zones (Rubin, 1990).

As mentioned, submarine rift zones in La Palma and El Hierro have built narrow ridges comparable to those in Hawaii. The ridge-shaped rifts in La Palma and El Hierro, however, are both shorter and more irregular along-axis, displaying significant curvature in plan shape than is typical on the Island of Hawaii. In addition the rifts in the Canary ridges are often discontinuous along strike. These differences also may be related to magma supply rate. Apparently the magma supply rate on La Palma and El Hierro is too low to smooth out any topographic irregularities along the axis. For example, the South Rift offshore El Hierro is cut by a 400 m deep east-west saddle located at 27°30′ N. This suggests that dike injection and eruption may have varied through time. Gee et al. (2001) proposed that the section of the ridge south of the saddle must represent an early phase of rift zone construction. For comparison, the rifts associated with Nintoku Guyot in the Emperor Seamount Chain have uneven crests with major reversals of slope as a result of more than 50 m.y. of volcanic inactivity (Vogt and Smoot, 1984).

Another possibility to explain the along-axis irregularities is that they are a result of the juxtaposition of ridges associated with different volcanoes. The north-south trending rift in La Palma does appear to represent distinct eruption cycles from two different volcanoes superimposed on one another. The eruption cycles differ in age by hundreds of thousands of years. A similar condition exists along the Hilo Ridge where the section of the ridge below 1000 m is an extension of a rift zone from Kohala Volcano. The section shallower than 1000 m was buried by lavas extruded during the construction of Mauna Kea. In the case of the Hilo Ridge, however, eruptions were large enough to smooth out the morphologic transition between the two ridges.

Other considerations may contribute to shaping the morphology of rifts including large scale factors such as: (1) age of the oceanic crust that the volcanoes are built on, (2) orientation of the tectonic fabric of the surrounding seafloor, (3) pre-existing sediment thickness, and 4) proximity of a plate boundary/continental margin. All of these must play some role in controlling the construction and evolution of oceanic islands, but we do not think they are the most important in controlling the characteristics of individual volcanic rift zones. Instead, as discussed above we infer that rift zone morphology is controlled primarily by the rate and constancy of magma supply, the regional stress field that may have been influenced by older volcanoes, and the rate of absolute plate motion.

Conclusion

Rift zones on La Palma, El Hierro and Hawaii Island are distinct and similar in many of their topographic characteristics. The rift zones in Hawaii tend to consist mostly of linear ridges. In contrast, the rift zones on La Palma and El Hierro display various geometries ranging from linear ridges having smooth to irregular crests to structures displaying a fan-like morphology in plan view. Although the toes of some of the Hawaiian rifts do partially fan out, they do not to the extent displayed by the rifts offshore El Hierro. The pronounced fanning may be a reflection of lower reservoir pressure, which is insufficient to overcome the compressive stress in the previous pathway so that new pathways are created that will be randomly distributed away from the magma source. It may also be caused by lateral shifts in the location of the rift zone. Another significant factor that may influence rift morphology and fanning of the rift may be absolute plate motion. In the Canary Islands, because plate motion is so slow, the volcanoes are built nearly one atop of another, with the stress field of one structure nearly superimposed on the other. The interaction of the stress fields may cause both widening and narrowing of the rift zone.

Irregularities and curvature of the rift zones in plan view on La Palma and El Hierro may be a reflection of differences in the rate of magma production in the islands groups (86 km^3/years for Kilauea Volcano versus 0.40–50 km^3/1000 years for El Hierro). There appear to be insufficient volumes of lavas erupted to smooth out topographic irregularities in the Canary Islands volcanoes. The superposition of rifts from different volcanoes may also add to the topographic irregularities, especially if eruption rates are low. Finally, such factors as the thickness of sediment cover over oceanic basement, the positions of the volcanic islands relative to the continent-ocean boundary and composition of the discharged magma also may have played roles in creating the rift morphologies observed in the Canary Islands and Hawaii.

Acknowledgements

We are indebted to the officers and crew of the R/V Hespérides and Vizconde de Eza for their cooperation during the various cruises to the Canary Island. Financial for the different cruises to the Canary Islands came from the Oceanographic and Hydrographic Research of the Spanish Economic Exclusive Program. Thanks are also extended to P. Clift, E. de Reus, Donna Blackman and an anonymous referee who helped greatly improve the manuscript. D.K. Smith's work in this study was funded by National Science Foundation Grant OCE-9618226. Contribution number 11171 of the Woods Hole Oceanographic Institution.

References

Ancochea, E., Hernán, F., Cendrero, A., Cantagrel, J.M., Fúster, J.M., Ibarrola, E. and Coello, J., 1994, Constructive and destructive episodes in the building of a young oceanic island, La Palma, Canary Islands, and genesis of the Caldera de Taburiente, *J. Vol. Geotherm. Res.*, **60**: 243–262.

Anguita, F. and Hernan, F., 1975, A propogating fracture model versus a hotspot origin for the Canary Islands, *Earth Planet. Sci. Letts.*, **27**: 11–19.

Atwater, T. and Severinghaus, J., 1989, Tectonic maps of the northeast Pacific, in E.L.Winterer, D.M. Hussong, and R.W. Decker, R.W. (eds,), *The Eastern Pacific and Hawaii, The Geology of North America*, **N**, *Geol. Soc. America*, 15–20.

Borgia, A. and Treves, B., 1992, Volcanic plates overriding the oceanic crust: structure and dynamics of Hawaiian volcanoes, in: Parson, L.M., Murton, B.J. and Browning, P. (eds), *Geol. Soc. Lond. Spec. Publ.* **60**: 277–299.

Bravo, T., 1982, Formaciones geológicas en la isla de El Hierro, *Inst. Estud. Canarios Aulas de Cultura, Cabildo de Tenerife*, 85–99.

Canales, J.P. and Dañobeitia, J.J., 1998, The Canary Islands swell; a coherence analysis of bathymetry and gravity, *Geophys. J. Int.*, **132**: 479–488.

Carracedo, J.C., 1994, The Canary Islands: an example of structural control on the growth of large oceanic-island volcanoes, *J. Vol. Geoth. Res.* **60**: 225–241.

Carracedo, J.C., 1999, Growth, structure, instability and collapse of the Canarian volcanoes and comparison with Hawaiian volcanoes, *J. Vol. Geoth. Res.* **94**: 1–19.

Carracedo, J.C., 1996, A simple model for the génesis of large gravitational landslide hazards in the Canary Islands, in McGuire, W.J., Jones A.P., and Neuberg, J. (eds.), Volcano Instability on the Earth and Other Planets, *Geol. Soc. Lond. Spec. Publ.* **110**, 125–135.

Carracedo, J.C., Badiola, E.R., Guillou, H., de la Nuez, J. and Perez Torrado, F.J., 2001, Geology and Volcanology of La Palma and El Hierro, Western Canaries, *Estud. Geol.* **57**: 175–273.

Carracedo, J.C., Day, S.J., Guillou, H. and Gravestock, P., 1999a, Later stages of volcanic evolution of La Palma, Canary Islands: rift evolution, giant landslides, and the genesis of the Caldera de Taburiente, *Geol. Soc. Am. Bull.* **111**: 755–768.

Carracedo, J.C., Day, S.J., Gillou, H. and Pérez–Torrado, F.J., 1999b, Giant Quaternary landslides of La Palma and El Hierro, Canary Islands, *J. Vol. Geoth. Res.* **94**: 169–190.

Cendrero, A., 1970, Report on the International symposium on volcanology (I.A.V.C.E.I.) held in Spain (Canary Islands) in September 1968, *Earth Extraterr. Sci.* **1**: 79–88.

Clague, D.A. and Dalrymple, G.B., 1989, Tectonics, geochronology, and origin of the Hawaiian-Emperor volcanic chain, in E.L.Winterer, D.M.Hussong, and R.W. Decker, R.W. (eds,), *Eastern Pacific Ocean and Hawaii, The geology of North America*, **N**, *Geol. Soc. America*, 188–217.

Clague, D.A., Hon, K.A. Anderson, J.L., Chadwick, W.W. and Fox, C.G., 1994, Bathymetry of Puna Ridge, Kilauea Volcano, Hawaii, *U.S. Geol. Survey. Misc. Invest. Ser.* **Map MF–2237**.

Clague, D.A., Moore, J.G., Dixon, J.E. and Friesen, W.E., 1995, Petrology of submarine lavas from Kilauea's Puna Ridge, *J. Petrol.* **36**: 299–349.

Clague, D.A., Moore, J.G. and Reynolds, J.R., 2000, Formation of submarine flat-topped volcanic cones in Hawaii, *Bull. Vol.* **62**: 214–233.

Clague, D.A., Moore, J.G., Torresan, M.E., Holcomb, R.T. and Lipman, P.W., 1988, Shipboard report for Hawaii GLORIA ground-truth cruise11-88-HW, *U.S. Geol. Survey Open File Rep.* **88–0292**: 54 pp.

Clague, D.A., Uto, K., Satake, K. and Davis, A.S., 2002, Eruption style and high emplacement in the submarine north arch volcanic field, Hawaii, in E. Takahashi, P.W. Lipman, M.O. García, J. Naka and S. Aramaki, (eds.), *Hawaiian Volcanoes. Deep Water Perspectives, Am. Geophys. Union Monogr.* **128**: 65–84.

Coello, J., Cantagrel, J.M., Hernán, F., Fúster, J.M., Ibarrola, E., Ancochea, E., Casquet, C., Jamond, C., Diaz de Téran, J.R. and Cendredo, A., 1992, Evolution of the eastern volcanic ridge of the Canary Islands based on new K-Ar data, *J. Vol. Geoth. Res.* **53**: 251–274.

Day, S.J., Carracedo, J.C., Guillou, H. and Gravestock, P., 1999, Recent structural evolution of the Cumbre Vieja volcano, La Palma, Canary Islands: volcanic rift zone reconfiguration as a precursor to volcanic flank instablity, *J. Vol. Geoth. Res.* **94**: 135–167.

Decker, R.W., 1989, Geologic map of the state of Hawaii, in E.L.Winterer, D.M. Hussong, and R.W. Decker, R.W. (eds,), *The Eastern Pacific and Hawaii, The geology of North America*, textbfN, *Geol. Soc. America.*

Delaney, P.T., Fiske, R.S., Miklius, A., Okamura, A.T. and Sako, M., 1990, Deep magma body beneath the summit and rift zones of Kilauea Volcano. Hawaii, *Science* **247**: 1311–1316.

Delaney, P.T., Miklius, A., Arnadottir, T., Okamura, A.T. and Sako, M.K., 1993, Motion of Kilauea Volcano during sustained eruption from the Pu'u 'O'o and Kupaianaha vents, 1983–1991, *J. Geophys. Res.* **98**: 17,801–17,820.

Denlinger, R.P. and Okubo, P., 1995, Structure of the mobile south flank of Kilauea Volcano, Hawaii, *J. Geophys. Res.* **100**: 24,499–24,507.

Dieterich, J.H., 1988, Growth and persistence of Hawaiian volcanic rift zones, *J. Geophys. Res.* **93**: 4258–4270.

Duffield, W.A., 1975, Structure and origin of the Koae Fault System, Kilauea Volcano, Hawaii, *U.S. Geol. Survey Prof. Paper*, **856**, 12 pp.

Dzurisin, D., Koyanagi, R.Y. and English, T.T., 1984, Magma supply and storage at Kilauea Volcano, Hawaii, *J. Vol. Geoth. Res.* **21**: 177–206.

Elsworth, D. and Day, S.J., 1999, Flank collapse triggered by intrusion: the Canarian and Cape Verde archipelago, *J. Vol. Geoth. Res.* **94**: 323–340.

Emery, K.O. and Uchupi, E., 1984, The Geology of the Atlantic Ocean, *Springer-Verlag*, New York, 1050 pp.

Epp, D., Decker, R.W. and Okamura, A.T., 1983, Relation of summit deformation to East rift zone, eruptions of Kilauea Volcano, Hawaii, *Geophys. Res. Lett.* **10**: 493–496.

Feraud, G., Giannerini, G., Campredon, R. and Stillman, C.J., 1985, Geochronology of some Canarian dike swarms: contribution to the volcano-tectonic evolution of the archipielago, *J. Vol. Geoth. Res.* **25**: 29–52.

Fiske, R.S. and Jackson, E.D., 1972, Orientation and growth of Hawaiian volcanic rifts: the effect of regional structure and gravitational stresses, *Proc. R. Soc. London* **A329**: 299–326.

Fornari, D.J., 1986, The geomorphic and structural evolution of Hawaiian submarine rifts, *U.S. Geol. Survey Prof. Paper* **1350**: 125–132.

Fornari, D.J. Campbell, J.F., 1986, Submarine topography around the Hawaiian Islands, *U.S. Geol. Survey Prof. Paper* **1350**: 109–123.

Fornari, D.J., Malahoff, A. and Heezen, B.C., 1978, Volcanic structure of the crest of the Puna Ridge, Hawaii: geophysical implications of submarine volcanic terrain, *Geol. Soc. Am. Bull.* **89**: 605–610.

Fornari, D.J., Malahoff, A. and Heezen, B.C., 1979, Submarine extension of the southwest rift zone of Mauna Loa Volcano, Hawaii: visual observations from U.S. Navy deep submergence vehicle DSV Sea Cliff, *Geol. Soc. Am. Bull.* **90**: 435–443.

Funck, T. and Schmincke, H.-U., 1998, Growth and destruction of Gran Canaria deduced from seismic reflection and bathymetric data, *J. Geophys. Res.* **103**: 15,393–15,407.

Fuster, J.M., Hernán, F., Cendrero, A., Coello, J., Cantagrel, J.M., Ancochea, E. and Ibarrola, E., 1993, Geocronología de la Isla de El Hierro (Islas Canarias), *Bol. R. Soc. Hist. Nat. (Sec. Geol.)* **88**: 85–97.

García, M., Hulsebosch, T. and Rhodes, J., 1995, Olivine-rich submarine basalts from the southwest rift zone of Mauna Loa Volcano: Implications for magmatic processes and geochemical evolution, in J.M. Rhodes and J.P. Lockwood (Eds.), *Mauna Loa revealed. Structure, composition, history and Hazards*, *Am. Geophys. Union Mon.* **92**: 219–239.

Gee, M.J.R., Masson, D.G., Watts, A.B. and Mitchell, N.C., 2001, Offshore continuation of volcanic rift zones, El Hierro, Canary Islands, *J. Vol. Geoth. Res.* **105**: 107–119.

Gregg, T.P.K. and Smith, D.K., 2003. Volcanic investigations of Puna Ridge, Hawaii: relations of lava flow morphologies and underlying slopes. *J. Vol. Getherm. Res.* **126**: 63–77.

Guillou, H., Carracedo, J.C. and Day, S.J., 1998, Dating of the upper Pleistocene-Holocene volcanic activity of La Palma using unspiked K-Ar technique, *J. Vol. Geoth. Res.* **86**: 137–149.

Guillou, H., Carracedo, J.C., Pérez, Torrado, F. and Rodríguez Badiola, E., 1996, K-Ar ages and magmatic stratigraphy of a hotspot-induced, fast grown oceanic island: El Hierro, Canary Islands, *J. Vol. Geoth. Res.* **73**: 141–155.

Hausen, H., 1959, On the geology of Lanzarote, Graciosa and the Isletas (Canarian archipelago), *Comm. Phys. Math.* **23**: 1–116.

Hausen, H., 1969, Some contributions to the geology of La Palma, *Comm. Phys. Math.* **35**: 1–140.

Hausen, H., 1973, Outlines of the geology of Hierro (Canary Islands), *Comm. Phys. Math.* **43**: 65–148.

Hernández Pacheco, A., 1982, Sobre una posible erupción en 1793 in la isla de El Hierro (Canarias), *Estud. Geol.* **38**: 15–25.

Hernández Pacheco, A. and Fernandez Santin, S., 1975, The submarine formation of the Caldera de Taburiente in La Palma (Canary Islands) and their metasomatic processes, IAVCEI Symposium, Santaigo de Chile, *Bull. Vol. Spec. V.*

Hills, D.J., Morgan, J.K., Moore, G.F. and Leslie, S.C., 2002, Structural variability along the submarine south flank of Kilauea Volcano, Hawaii, from a multichannel seismic reflection survey, in Takahashi, E., Lipman, P.W., García, M.O. and Aramaki, S. (eds.), *Hawaiian volcanoes. Deep Water Perspective*, *Am. Geophys. Union Monogr.* **128**: 105–124.

Hoernle, K., Tilton, G. and Schmincke, H.-U., 1991, Sr-Nd-Pb isotopic evolution of Gran Canaria: evidence for shallow enriched mantle beneath the Canary Islands, *Earth Planet. Sci. Letts.* **106**: 44–63.

Holcomb, R.T., 1987, Eruptive history and long-term behavior of Kilauea volcano, in R.W. Decker, T.L.Wright, T.L. and P.H. Stauffer (eds.), *Volcanism in Hawaii*, *U.S. Geol. Survey Prof. Paper* **1350**: 261–350.

Holcomb, R.T., Moore, J.G., Lipman, P.W. and Belderson, R.H., 1988, Voluminous submarine lava flows from Hawaiian volcanoes, *Geology* **16**: 400–404.

Holcomb, R.T., Nelson, B.K., Reiners, P.W. and Sawyer, N.L., 2000, Overlapping volcanoes: the origin of Hilo Ridge, *Geology* **28**: 547–550.

Holcomb, R.T. and Searle, R.G., 1991, Large landslides from oceanic volcanoes, *Mar. Geotechnol.* **10**: 19–32.

Holik, J.S., Rabinowitz, P.D. and Austin, J.A., 1991, Effects of Canary hotspot volcanism on structure of oceanic crust off Morocco, *J. Geophys. Res.* **96**: 12,039–12,067.

Johnson, K.T.M., Reynolds, J., Vonderhaar, D., Smith, D.K. and Kong, L.S.L., 2002, Petrological systematics of submarine lavas from the Puna Ridge, Hawaii: implications for rift zone plumbing and magmatic processes, in E. Takahishi, P.W. Lipman, M.O. García, J. Naka and S. Aramaki (eds.), *Hawaiian volcanoes. Deep underwater perspectives*, *Am. Geophys. Union Monogr.* **128**: 143–160.

Kauahikaua, J., Hildenbrand, T. and Webring, M., 2000, Deep magmatic structures of Hawaiian volcanoes, imaged by three-dimensional gravity models, *Geology* **28**: 883–886.

Klein, F.W., Koyanagi, R.Y., Nakata, J.S. and Tanigawa, W.R., 1987, The seismicity of Kilauea's magma system, *U.S. Geol. Survey Prof. Paper* **1350**: 1019–1185.

Klügel, A., 1997, Ascent rates of magmas and genesis, transport, and reactions of mantle and crustal xenoliths of the 1949 eruption on La Palma (Canary Islands), *Ph.D. thesis*, Christians-Albrechts-Universität, Kiel, 209 pp.

Langenheim, V.A.M. and Clague, D.A., 1987, The Hawaiian-Emperor volcanic Chain: Part II. Stratigraphic framework of volcanic rocks of the Hawaiian Islands, *U.S. Geol. Survey Prof. Paper* **1350**: 55–84.

Le Bas, M.J., Rex, D.C. and Stillman, C.J., 1986, The early magmatic chronology of Fuerteventura, Canary Islands, *Geol. Mag.* **123**: 287–298.

Lénat, J.F., Vincent, P. and Bachèlery, P., 1986, The off-shore continuation of an active basaltic volcano; Piton de la Fournaise (Réunion Island, Indian Ocean); structural and geomorphological interpretation from sea beam mapping, *J. Vol. Geoth. Res.* **36**: 1–36.

Lipman, P.W., 1980, The Southwest Rift Zone of Mauna Loa: implications for structural evolution of Hawaiin volcanoes, *Am. J. Sci.* **280-A**: 752–776.

Lipman, P.W., Clague, D.A., Moore, J.G. and Holcomb, R.T., 1989, South Arch volcanic field-newly identified young lava flows on the sea floor south of the Hawaiian Ridge, *Geology* **17**: 611–614.

Lonsdale. P., 1989, A geomorphological reconnaissance of the submarine part of the East rift zone of Kilauea Volcano, Hawaii, *Bull. Vol.* **51**: 123–144.

Malahoff, A. and McCoy, F., 1967, The geologic structure of the Puna Ridge, Hawaii, *J. Geophys. Res.* **72**: 541–548.

Mangan, M.T., Heliker, C.C., Mattox, T.N., Kauahikaua, J.P. and Helz, R.T., 1995, Episode 49 of the Pu'u 'O'o – Kupaianaha eruption of Kilauea volcano: breakdown of a steady-state eruptive era, *Bull. Vol.* **57**: 127–135.

Martí, J., Mitjavila, J. and Araña, V.,1994, Stratigraphy, structure and geochronology of the Las Cañadas Caldera (Tenerife, Canary Islands), *Geol. Mag.* **131**: 715–727.

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

Masson, D.G., 1996, Catastrophic collapse of the volcanic island of Hierro 15 Ka ago and the history of landslides in the Canary Islands, *Geology* **24**: 231–234.

Masson, D.G., Watts, A.B., Gee, M.J.R., Urgeles, R., Mitchell, N.C., Le Bas, T.P. and Canals, M., 2002, Slope failures on the flanks of the Canary Islands, *Earth-Sci. Rev.* **57**: 1–35.

McDougall, I., 1979, Age of shield-building volcanism in Kauai and linear volcanism in Hawaiian Island chain, *Earth Planet. Sci. Letts.* **46**: 31–42.

Mehl, K.W. and Schmincke, H.-U., 1999, Structure and emplacement of the Pliocene Roque Nublo debris avalanche deposit, Gran Canaria, Spain, *J. Vol. Geoth. Res.* **94**: 105–134.

Mitchell, N.C., Masson, D.G., Watts, A.B., Gee, M.J.R. and Urgeles, R., 2002, The morphology of the flanks of volcanic ocean islands: A comparative study of the Canary and Hawaiian hotspot islands, *J. Vol. Geoth. Res.* **115**: 83–107.

Monterey Bay Aquirium research Institute (MBARI), 2000, Monterey Bay Aquarium Research Institute Hawaii multibeam survey, *Digital Series* **2**, Moss Landing CA.

Moore, J.G. and Campbell, J.F., 1987, Age of tilted reefs, Hawaii, *J. Geophys. Res.* **92**: 752–759.

Moore, J.G. and Chadwick, Jr., W.W., 1995, Offshore geology of Mauna Loa and adjacent areas, Hawaii, in Rhodes, J.M. and Lockwood, J.P. (eds.), *Mauna Loa Revealed. Structure, Composition, History and Hazards, Am. Geophys. Union Geophys. Mon.* **52**: 21–44.

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

Moore, J.G., Clague, D.A., Holcomb, R.T., Lipman, P.W., Normark, W.R. and Torresan, M.E., 1989, Prodigious submarine landslides on the Hawaiian Ridge, *J. Geophys. Res.* **94**: 17,465–17,484.

Moore, J.G. and Fiske, R.S., 1969, Volcanic substructure inferred from dredge samples and ocean-bottom photographs, Hawaii, *Geol. Soc. Am. Bull.* **80**: 1191–1201.

Moore, J.G. and Fornari, D.J., 1984, Drowned reefs as indicators of rate of subsidence of the island of Hawaii, *J. Geology* **92**: 752–759.

Moore, J.G., Phillips, R.L., Grigg, R.W., Peterson, D.W. and Swanson, D.A., 1973, Flow of lava into the sea, 1969–1971, Kilauea Volcano, Hawaii, *Geol. Soc. Am. Bull.* **84**: 537–546.

Moore, R.B. and Trusdell, F.A., 1991, Geologic map of the lower east rift zone of Kilauea Volcano, Hawaii, *U.S. Geol. Survey Misc. Invest. Ser.*, **Map I–2225.**

Morgan, W.J., 1970, Plate motions and deep mantle convection, in R. Shagan et al. (eds), *Studies in Earth and Space Sciences. Geol. Soc. America Memoir*, **132**: 7–22.

Morgan, W.J., 1972, Deep mantle convection plumes and plate motions. *Am. Assoc. Pet. Geol. Bull.* **56**: 203–213.

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

Moss, J.L., McGuire, W.J. and Page, D., 1999, Ground deformation monitoring of a potential landslide at La Palma, Canary Island, *J. Vol. Geoth. Res.* **94**: 251–260.

Owen, S., Segall, P., Freymueller, J., Miklius, A., Delinger, R., Arnadottir, T., Sako, M. and Burgmann, R., 1995, Rapid deformation of the south flanks of Kilauea Volcano, Hawaii, *Science* **267**: 1328–1332.

Parfitt, E.A., Gregg, T.K.P. and Smith, D.K., 2002, A comparison between subaerial and submarine eruptions at Kilauea Volcano, Hawaii, *J. Vol. Geoth. Res.* **113**: 213–242.

Parfitt, E.A. and Peacock, D.C.P., 2001, Faulting in the south flank of Kilauea Volcano, Hawaii, *J. Vol. Geoth. Res.* **106**: 265–284.

Peterson, D.W. and Moore, R.B., 1987, Geologic history and evolution of geologic concepts, island of Hawaii, *U. S. Geol. Survey Prof. Paper* **1350**, 149–189.

Pollard, D.D., Delaney, P.T., Duffield, W.A., Endo, E.T. and Okamura, A.T., 1983, Surface deformation in volcanic rift zones, *Tectonophysics* **94**: 541–584.

Price, I., 1980, Gravity tectonics on a passive margin.: Deep-Sea Drilling Project site 415 in the light of regional seismic data, in

Y. Lancelot, E.L. Winterer et al., *Init. Repts. of the Deep Sea Drill. Proj.* **50**: 759–771.

Ridley, W.I., 1971, The origin of some collapse structures in the Canary Islands, *Geol. Mag.* **108**: 477–484.

Rubin, A.M., 1990, A comparison of rift-zone tectonics in Iceland and Hawaii, *Bull. Vol.* **52**, 302–319.

Rubin, A.M. and Gillard, D., 1998, A reinterpretation of seismicity associated with the January 1983 dike intrusion at Kilauea Volcano, Hawaii, *J. Geophys. Res.* **103**:10,003–10,015.

Ryan, M.P., 1988, The mechanics and three-dimensional internal structure of active magmatic systems: Kilauea Volcano, Hawaii, *J. Geophys. Res.* **93**: 4213–4248.

Ryan, M.P., Koyanagi, R.Y. and Fiske, R.S., 1981, Modelling of the three dimensional structure of macroscopic magma transport systems: application to the Kilauea Volcano. Hawaii, *J. Geophys. Res.* **86**: 7111–7129.

Smith, D.K. and Cann, J.R., 1999, Constructing the upper crust of the Mid-Atlantic Ridge: a reinterpretation based on the Puna Ridge, Kilauea Volcano, *J. Geophys. Res.* **97**: 25,379–25,399.

Smith, D.K., Kong, L.S., Johnson, K.T.M. and Reynolds, J.R., 2002a, Volcanic morphology of the submarine Puna Ridge, Kilauea Volcano, in Takahashi, E., Lipman, P.W., García, M.O. and Aramaki, S. (eds.), *Hawaiian volcanoes. Deep Water Perspective*, *Am. Geophys. Union Monogr.* **128**: 125–142.

Smith, D.K., Tivey, M.A., Gregg, P. and Kong, L.S., 2001,, Magnetic anomalies along the submarine Puna Ridge, Kilauea Volcano, Hawaii, *J. Geophys. Res.* **106**: 16,047–16,060.

Smith, J.R., Satake, K., Morgan, J.K. and Lipman, P.W., 2002b, Submarine landslides and volcanic features on Kohala and Mauna Kea volcanoes, in Takahashi, E., Lipman, P.W., García, M.O. and Aramaki, S. (eds.), Hawaiian volcanoes. Deep Water Perspective, *Am. Geophys. Union Monogr.*, **128**: 11–28.

Staudigel, H. and Smincke, H.-U., 1984, The Pliocene seamount series of La Palma/Canary Islands, *J. Geophys. Res.* **89**: 11,190–11,215.

Stearns, H.T. and Macdonald, G.A., 1946, Geology and groundwater resources of the Island of Hawaii, *Hawaii Div. Hydro. Bull.* **9**: 363 pp.

Stillman, C.J., 1999, Giant Miocene landslides and evolution of Fuerteventura, Canary Islands, *J. Vol. Geoth. Res.* **94**: 89–104.

Swanson, D.A., Jackson, D.B., Koyanagi, R.Y. and Wright, T.L., 1976, The February 1969 rift eruption of Kilauea Volcano, Hawaii, *U.S. Geol. Survey Prof. Paper* **891**: 30 pp.

Thurber, C.H. and Gripp, A.E., 1988, Growth and persistence of Hawaiian rift zones, *J. Geophys. Res.* **93**: 4271–4278.

Tilling, R.I. and Dvorak, J.J., 1993, Anotomy of a basaltic volcano, *Nature* **363**: 125–133.

Troll, V.R., Walter, T.R. and Schmincke, H-U., 2002, Cyclic caldera collapse: piston or piecemeal subsidenced? Field experiment evidence, *Geology* **30**: 135–138.

Uchupi, E., Emery, K.O., Bowin, C.O. and Phillips, J.D., 1976, Continental margin off western Africa: Senegal to Portugal, *Am. Assoc. Petrol. Geol. Bull.* **60**: 809–878.

Urgeles, R., Canals, M., Baraza, J., Alonso, B. and Masson, D.G., 1997, The most recent megaslides on the Canary Islands: the El Golfo debris avalanche and the Canary debris flow; west of El Hierro Island, *J. Geophys. Res.* **102**: 20,305–20,323.

Urgeles, R., Masson, D.G., Canals, M., Watts, A.B. and Le Bas, T., 1999, Recurrent large-scale landsliding on the west flank of La Palma, Canary Islands, *J. Geophys. Res.* **104**: 24,331–25,348.

Vogt, P.R. and Smoot, N.C. 1984, The Geisha Guyots: multi-beam bathymetry and morphometric interpretation, *J. Geophys. Res.* **89**: 11,085–11,107.

Walker, G.P.L., 1988, Three Hawaiian calderas: an origin through loading by shallow intrusions? *J. Geophys. Res.* **93**: 14,773–14,784.

Ward, S.N. and Day, S., 2001, Cumbre Vieja volcano – potential collapse and tsunami at La Palma, Canary Islands, *Geophys. Res. Lett.* **28**: 3397–3400.

Watts, A.B., Peirce, C., Collier, J., Dalwood, R., Canales, J.P. and Henstock, T.J., 1997, A seismic study of lithospheric flexure in the vicinity of Tenerife, Canary Islands, *Earth Planet. Sci. Letts.* **146**: 431–447.

Wolfe, E.W., García, M.O., Jackson, D.B., Koyanagi, R.Y., Neal, C.A. and Okamura, A.T., 1987, The Pu'u 'O'o eruption of Kilauea volcano, episodes 1–20, January 3, 1983, to June 8, 1984, *U.S. Geol. Survey Prof. Paper* **1350**: 471–508.

Wyss, M., 1980, Hawaiian rifts and recent Icelandic volcanism: expressions of plume generated radial stress fields, *J. Geophys. Res.* **47**: 19–22.

Zazo, C., Goy, J.L., Hillaire-Marcel, C., Gillot, P.Y., Soler, V., Gozález, J.A., Dabrio, C.J. and Ghaleb, G., 2002, Raised marine sequences of Lanzarote and Fuerteventura revisited-a reappraisal of relative sea-level changes and vertical movements in the eastern Canary Islands during the Quaternary, *Quat. Rev.* **21**: 18–19: 2019–2046.