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Growth, structure, instability and collapse of Canarian volcanoes and comparisons with Hawaiian volcanoes

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Abstract

Recent onshore and offshore investigations in the Canarian archipelago, especially in the western islands of La Palma and El Hierro, have greatly improved the understanding of the genesis and evolution of these islands and allow interesting comparisons with other hotspot-induced oceanic island volcanoes, especially the Hawaiian archipelago. Genesis induced by hotspot activity and, consequently, shield and post-erosional stages of growth allow the definition of similar main stratigraphic units. The Canarian and Hawaiian volcanoes show common constructional and structural features, such as rift zones, progressive volcano instability and multiple gravitational collapses. However, the Canaries present some important geological differences from the prototypical hotspot volcanoes of the Hawaiian Islands, particularly the absence of significant subsidence; The Canarian volcanoes remain emergent until completely mass-wasted by gravitational collapses and erosion. Volcanic formations over 20 million years are observable in outcrop in the Canaries, including the seamount stages of growth in several of the islands. © 1999 Elsevier Science B.V. All rights reserved.

Keywords: Canary Islands; Hawaiian Islands; oceanic islands; comparative volcanism and structure

1. Introduction

Apart from the Hawaiian Islands, the Canaries are probably one of the most extensively studied groups of hotspot-generated oceanic islands in the world.

The comparison of the Hawaiian and Canarian volcanoes leads to some interesting clarifications. Each archipelago represents a distinctly characteristic hotspot-induced oceanic group: (1) the Hawaiian archipelago, the prototypical hotspot islands, with a readily distinguishable hotspot signature (a clear hotspot trace, mantle anomaly and swell), generated on a fast-moving plate (Wilson, 1963; Morgan, 1972; Clague and Dalrymple, 1987; Walker, 1990); and (2) the Canaries, where the hotspot signature is geochemically and geophysically less evident and the plate velocity is lower by about an order of magnitude (Filmer and Mcnutt, 1988; Hoernle and Schmincke, 1993; Canas et al., 1994; Watts, 1994).

There are several specific advantages in the Canaries for the study of the processes and patterns of development, the structure, instability, and collapse of island volcanoes that are difficult to investigate fully in other oceanic groups. Perhaps, the only

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similarly-advantaged group is the Cape Verde Islands; unfortunately, the geology and geochronology of these remote islands are relatively little known. The individual islands of the Hawaiian archipelago have a relatively short lifespan: high-rate subsidence (as quoted by Moore, 1987, most of the Hawaiian volcanoes have subsided 2–4 km since emergence) precludes the islands from continuing above sea level after only a few million years (Moore and Fornari, 1984; Moore, 1987; Moore and Thomas, 1988; Walker, 1990). Kauai, the oldest emerged island in the Hawaiian chain, is less than 6 Ma (McDougall, 1964, 1979). Thus, only the relatively young parts of the islands are exposed to observation. High rainfall gives rise to a dense vegetation screen and rapid weathering of rocks, making exposures relatively scarce. Thus, geological observations as well as radiometric dating and magnetic stratigraphy are difficult in many areas, increasing the difficulty of the complete reconstruction of these volcanoes (McDougall, 1963, 1964, 1969, 1971, 1979; McDougall and Tarling, 1963; McDougall and Ur-Rahman, 1972; McDougall and Swanson, 1972; Holcomb, 1980; Naughton et al., 1980; Clague and Dalrymple, 1987; Holcomb et al., 1997).

In contrast, there is no comparable subsidence in the Canaries, as discussed below. Volcanic formations over 20 Ma are observable in outcrop, including the seamount stages of growth in several of the islands, as discussed later. Since all the Canaries are still emergent above sea level, it is feasible to study in outcrop volcanic formations and structures over a wide range of ages and original emplacement depths, according to the age of the islands and the different stages of erosive mass-wasting. The vegetation cover is comparatively poor and the lavas have a high K content, favouring geochronological (K/Ar and palaeomagnetic reversals) determinations (Abdel-Monem et al., 1971, 1972; McDougall and Schmincke, 1976; Carracedo, 1979; Ancochea et al., 1990, 1994, 1996; Coello et al., 1992; Guillou et al., 1996, 1998; Carracedo et al., 1999-this volume). The high population and the absence of surface waters on the islands have necessitated the mining of groundwater by means of more than 3000 km of sub-horizontal tunnels (locally known as ''galerías''), which allow the observation of the internal structure of the volcanoes at almost any point and depth in the islands of La Palma and Tenerife and, to a lesser extent, El Hierro (Carracedo, 1994).

In this paper we review some important features of the growth, structure, progressive instability and flank collapse of the Canarian volcanoes and their comparison with other oceanic volcanoes, especially those of the Hawaiian Islands.

2. Growth of the Canarian volcanoes

2.1. Geochronology and stratigraphy

The first comprehensive stratigraphy of the Canaries was compiled by Fúster et al. (1968a,b,c,d) for the islands of Lanzarote, Fuerteventura, Gran Canaria and Tenerife. Numbered stratigraphic series (I to IV) were defined by the presence of a general erosional unconformity and interbedded fossil beaches. Absolute dating of the volcanic formations using radiometric (K/Ar) and palaeomagnetic techniques (Abdel-Monem et al., 1971, 1972; McDougall and Schmincke, 1976; Carracedo, 1979) made it clear that the stratigraphy defined was not applicable to all the Canaries.

The term ''Series'' used by Fúster et al. conformed to the stratigraphic code in use at the time. However, it does not conform to the present code (NACSN, 1983), which restricts the use of this term to geological units formed during the same time-span, with synchronous boundaries. Recently improved geochronological (radiometric and palaeomagnetic) data reinforced these ideas (Ancochea et al., 1990, 1994; Guillou et al., 1996, 1998; Carracedo et al., 1999-this volume) suggesting to find a more appropriate stratigraphic definition for the Canaries.

Fig. 1 presents a plot of the published K/Ar ages from lava flows of the Canarian and the Hawaiian islands. When compared, we can easily see that individual islands in both archipelagos show two main stages of growth separated by periods of volcanic repose (gaps). The combined use of palaeomagnetic reversals and radiometric dating has proved to be efficient in the Hawaiian Islands (Holcomb et al., 1997) and the Canaries (Abdel-Monem et al., 1971, 1972; Carracedo, 1979; Carracedo and Soler, 1995; Pérez Torrado et al., 1995; Guillou et al., 1996) in defining the presence of eruptive gaps or

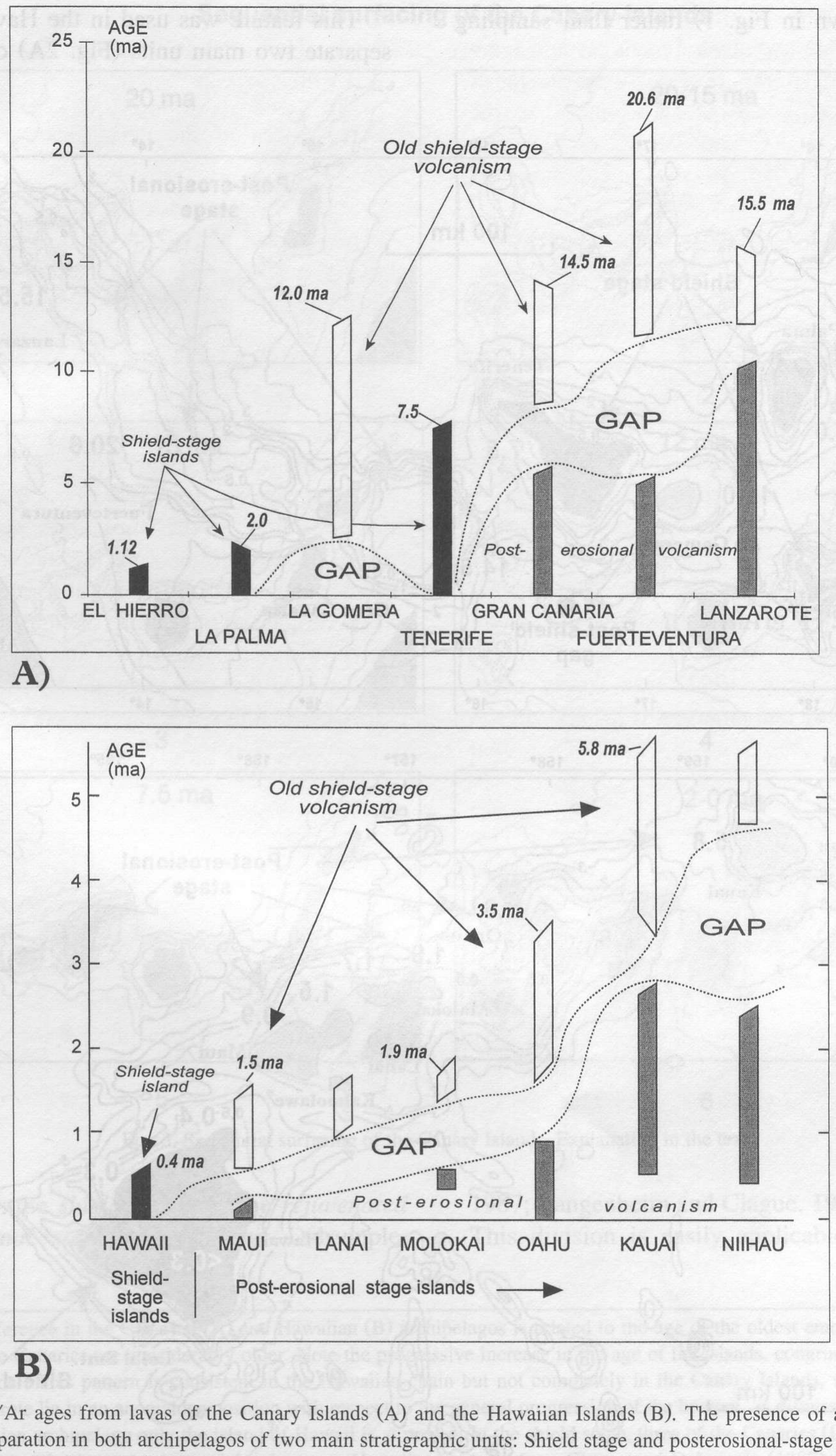
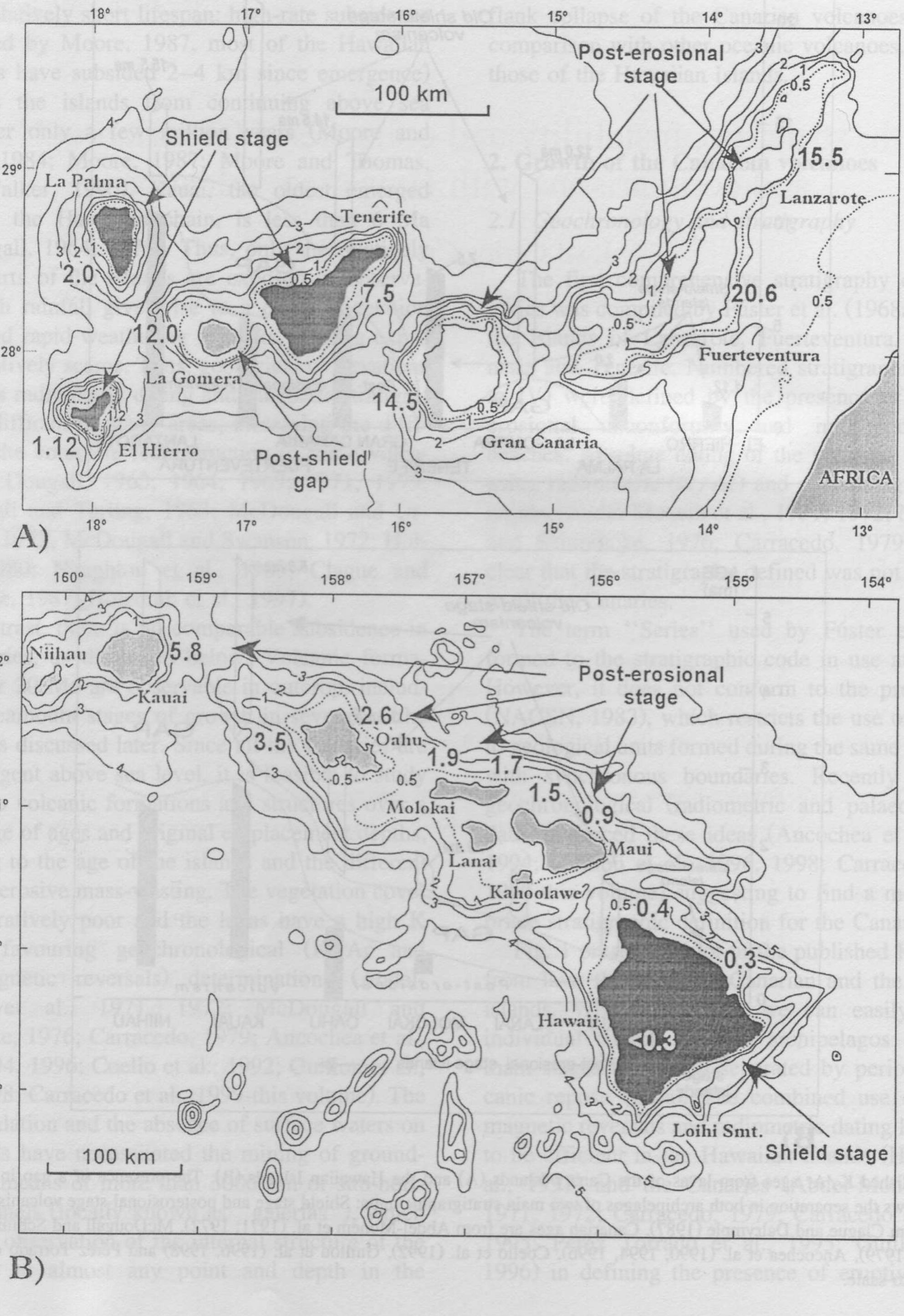


Fig. 1. Published K/Ar ages from lavas of the Canary Islands (A) and the Hawaiian Islands (B). The presence of a gap in the eruptive activity allows the separation in both archipelagos of two main stratigraphic units: Shield stage and posterosional-stage volcanism. Hawaiian ages are from Clague and Dalrymple (1987). Canarian ages are from Abdel-Monem et al. (1971; 1972), McDougall and Schmincke (1976), Carracedo (1979), Ancochea et al. (1990, 1994, 1996), Coello et al. (1992), Guillou et al. (1996, 1998) and Pérez Torrado et al. (1995). X-axis not to scale.

hiatuses (as shown in Fig. 1) rather than sampling gaps.

This feature was used in the Hawaiian Islands to separate two main units (Fig. 2A) of general use in



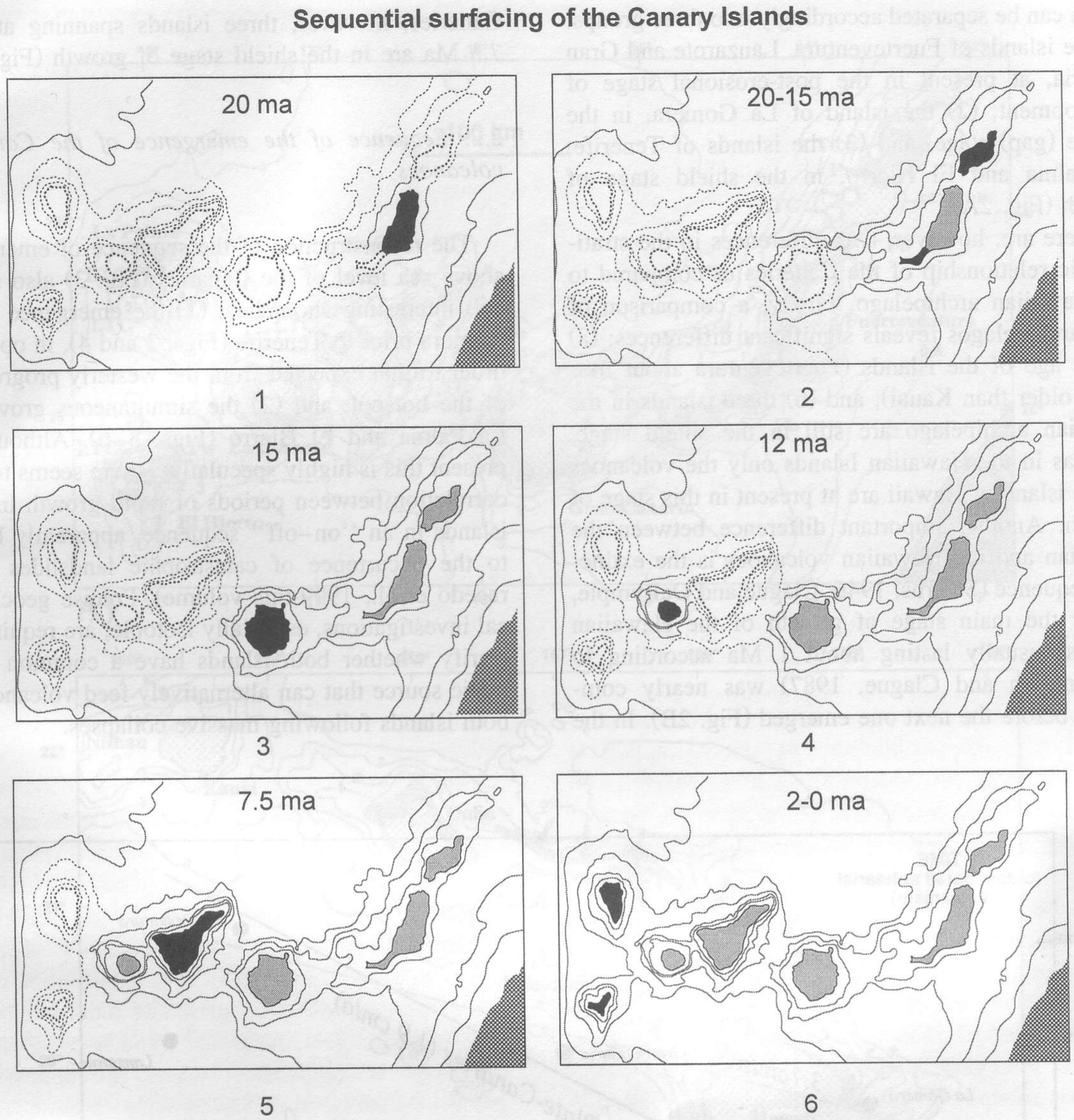


Fig. 3. Sequential surfacing of the Canary Islands. Explanation in the text.

the archipelago: the *shield stage* and the *rejuvenated or post-erosional stage* (Clague and Dalrymple, 1987; Langenheim and Clague, 1987; Walker, 1990). This division is easily applicable to the Canaries,

Fig. 2. A main difference in the Canarian (A) and Hawaiian (B) archipelagos is related to the age of the oldest emerged volcanism in each island, which in the Canaries are considerably older. Note the progressive increase in the age of the islands, congruent in both archipelagos with a hotspot model. This pattern is consistent in the Hawaiian chain but not completely in the Canary Islands, where the islands of La Gomera and Lanzarote lie in an anomalous position with respect to the general progression of the hotspot, as discussed in the text. Note that while in the Hawaiian archipelago only the island of Hawaii is at present in the shield stage, three of the Canaries (La Palma, El Hierro and Tenerife) are at present in the same stage of development. Hawaiian ages are from Clague and Dalrymple (1987). Canarian ages are from Abdel-Monem et al. (1971), McDougall and Schmincke (1976), Ancochea et al. (1990, 1994, 1996), Coello et al. (1992) and Guillou et al. (1996).

which can be separated accordingly into three groups: (1) the islands of Fuerteventura, Lanzarote and Gran Canaria, at present in the post-erosional stage of development; (2) the island of La Gomera, in the repose (gap) stage, and (3) the islands of Tenerife, La Palma and El Hierro, in the shield stage of growth (Fig. 2A).

There are, however, two differences in the stratigraphic relationship of the Canaries as compared to the Hawaiian archipelago. Firstly, a comparison of both archipelagos reveals significant differences: (a) in the age of the islands (Fuerteventura about four times older than Kauai); and (b) three islands in the Canarian archipelago are still in the shield stage, whereas in the Hawaiian Islands only the volcanoes on the island of Hawaii are at present in this stage of growth. Another important difference between the Canarian and the Hawaiian volcanoes is the extinction sequence (Stearns, 1946; Clague and Dalrymple, 1987): the main stage of growth of the Hawaiian Islands (usually lasting about 1 Ma according to Langenheim and Clague, 1987) was nearly completed before the next one emerged (Fig. 2B). In the Canaries, however, three islands spanning at least 7.5 Ma are in the shield stage of growth (Fig. 2A).

2.2. Sequence of the emergence of the Canarian volcanoes

The reconstruction of the sequence of emergence above sea level of the Canaries (Fig. 3) also shows two interesting anomalies: (1) the emergence of La Gomera prior to Tenerife (Figs. 3 and 4), in opposite order to that expected from the westerly progression of the hotspot; and (2) the simultaneous growth of La Palma and El Hierro (Figs. 3–6). Although at present this is highly speculative, there seems to be a correlation between periods of rapid growth in both islands in an ''on–off'' sequence, apparently linked to the occurrence of catastrophic landslides (Carracedo et al., 1999-this volume). Precise geochemical investigations, especially isotopic, are required to clarify whether both islands have a common magmatic source that can alternatively feed volcanoes in both islands following massive collapses.

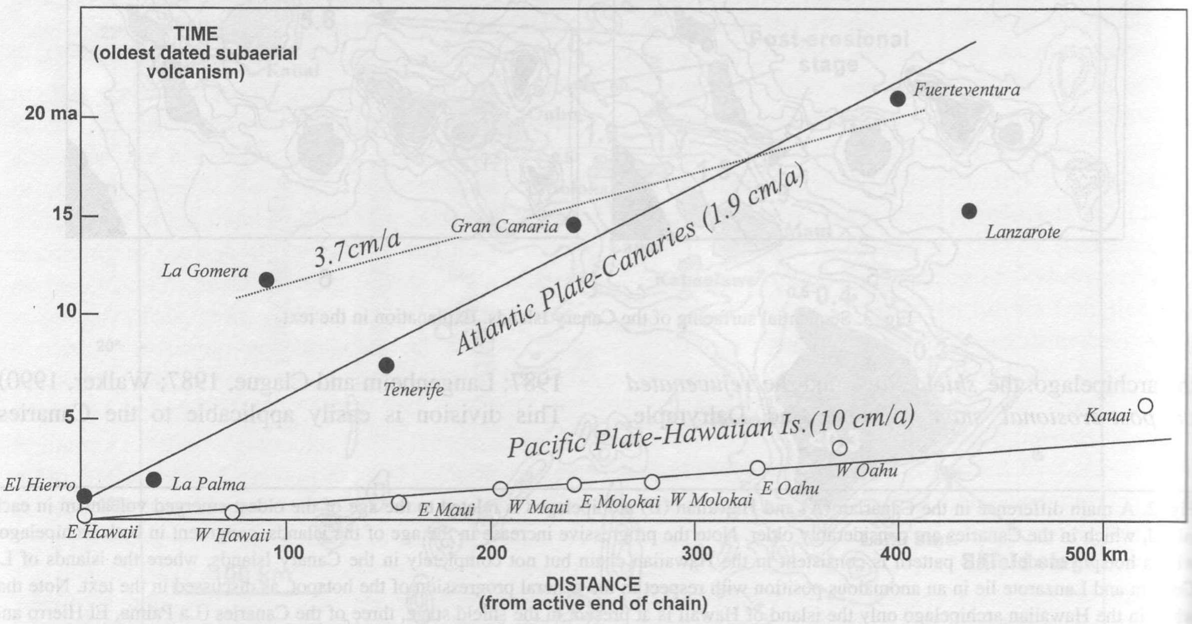


Fig. 4. Oldest subaerial age vs. distance plot for the Canarian and Hawaiian islands. See text for explanation. Hawaiian ages are from Clague and Dalrymple (1987). Canarian ages are from Abdel-Monem et al. (1971), McDougall and Schmincke, 1976, Ancochea et al. (1990, 1994, 1996), Coello et al. (1992) and Guillou et al. (1996).

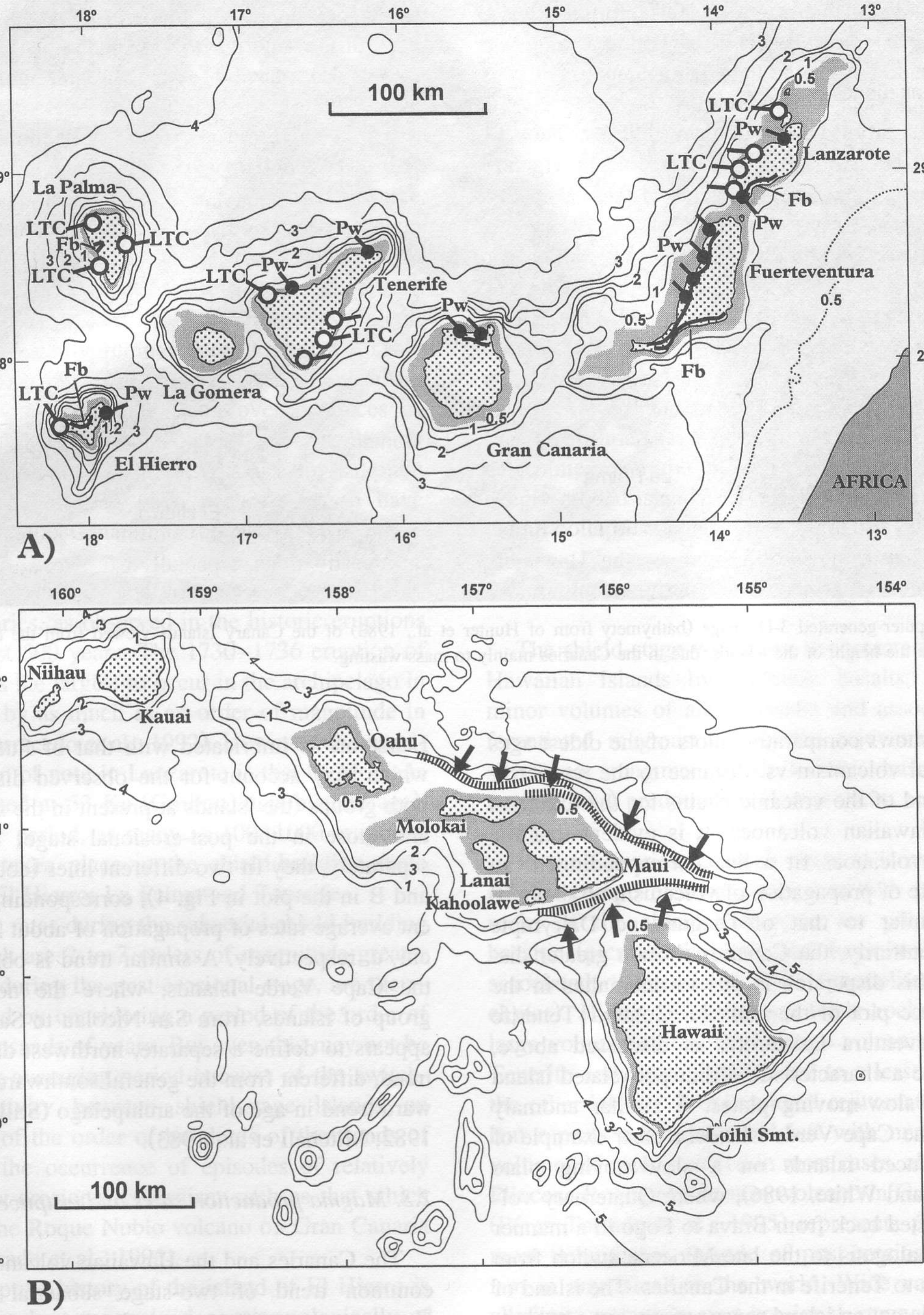


Fig. 5. (A) The Canaries have been very stable, at least in the last 15 Ma; subsidence is not significant, as shown by the abundant litoral cones (LTC in the figure), fossil beaches (Fb) and pillow lavas (Pw) located at about the present sea level. (B) Contrarily, subsidence is very active in the Hawaiian Islands, resulting in the presence of submerged reef terraces as shown in the figure (Moore, 1987). Subsidence finally ends the emerged life of the island in a relatively short time span, probably less than about 6–7 Ma, changing to atolls and late guyots.

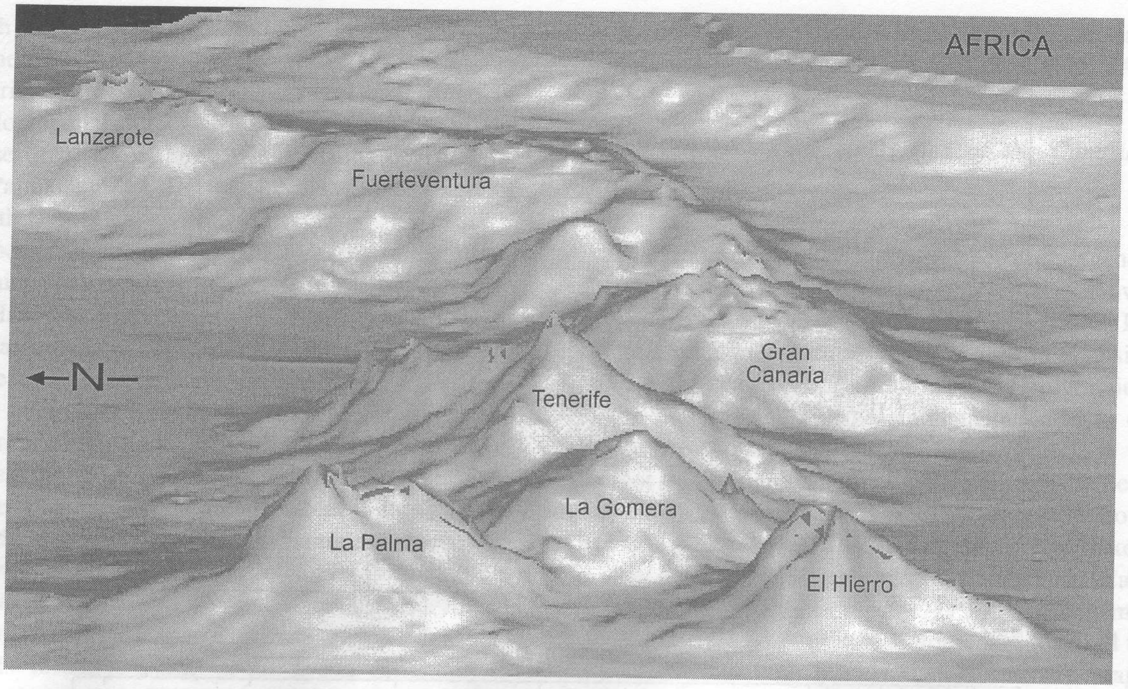


Fig. 6. Computer-generated 3-D image (bathymety from of Hunter et al., 1983) of the Canary Islands viewed from the east. Note the differences in the height of the islands, due in the Canaries mainly to mass-wasting.

Fig. 4 shows comparative plots of the oldest ages of subaerial volcanism vs. distance to the respective younger end of the volcanic chains for the Canarian and the Hawaiian volcanoes. It is evident that the Hawaiian volcanoes fit a line corresponding to an average rate of propagation of volcanism of about 10 cm/a, similar to that of Clague and Dalrymple (1987). Contrarily, the Canaries show a greater dispersion. This dispersion is due to anomalies in the age–distance plot of the pairs La Gomera–Tenerife and Fuerteventura–Lanzarote, as discussed above, and may be a characteristic of hotspot-related island groups on slow-moving plates. A similar anomaly occurs in the Cape Verde islands, a clear example of hotspot-induced islands on a slow-moving plate (Courtnay and White, 1986), where Quaternary volcanism shifted back from Brava to Fogo in a manner possibly analogous to the late Miocene switch from La Gomera to Tenerife in the Canaries. The island of Lanzarote is not an island sensu stricto, but a prolongation towards the NE of Fuerteventura: both islands are separated by the La Bocaina straight, less than 50 m deep (see Fig. 6). The volcanic construction of Lanzarote is imbricated with that of Fuerteventura, which may account for the observed dispersion. If both groups (the islands at present in the shield stage and those in the post-erosional stage) are plotted separately, they fit two different lines (dotted lines A and B in the plot in Fig. 4), corresponding to different average rates of propagation of about 2.0 and 3.7 cm/a, respectively. A similar trend is observable in the Cape Verde Islands, where the northwestern group of islands, from San Nicolau to Santo Antao, appears to define a separate, northwest directed age trend, different from the general southward and westward trend in age of the archipelago (Stillman et al., 1982; Mitchell et al., 1983).

2.3. Magma production rates and eruptive frequency

The Canaries and the Hawaiian volcanoes share a common trend of two-stage subaerial evolution. However, they differ greatly in the type of magmas involved and the rate in which they are produced and supplied, much greater in the latter. This is clearly reflected both in the much greater volumes of the

Hawaiian shields compared to the products of the post-erosional magmatism in each island and in the much higher frequency of eruptions during the shield-building stage of each volcano (Holcomb, 1987).

Magma production rates may be difficult or impossible to evaluate in the Canary Islands. The discontinuous character of volcanism, in which eruptive gaps, inherently difficult to date, may predominate over periods of activity, make true evaluation of magma production rates unreliable unless large time intervals are compared. Furthermore, giant lateral collapses repeatedly remove large fractions of the mass of an island, especially during the shield-building stage and redistribute them over distances of hundreds or even thousands of kilometres. Several megaturbidites deposited in the Madeira abyssal plain within the past 1 Ma have been shown to have originated in the Canarian Archipelago (Weaver et al., 1992).

Eruptive frequency and volume vary considerably in the Canaries, as observed in the historic eruptions over the last 500 years. The 1730–1736 eruption of Lanzarote is the largest to occur in the archipelago in this period by as much as an order of magnitude in volume (Carracedo et al., 1992). However, the previous eruption of note in Lanzarote is that of Montaña Corona, dated at 53 Ka (Guillou, unpublished age). In the same period, as many as 100–1000 eruptions may have taken place in the shield-building stage islands of El Hierro, La Palma and Tenerife.

Eruption rates during the subaerial shield-building stage, which are 2 to 3 orders of magnitude greater than those during the post-erosional stage, are therefore implied by considering a period of the order of tens of thousands of years. But even this may not be a sufficient averaging period because of the switching of activity between shield-stage islands on timescales of the order of hundreds of thousands of years and the occurrence of episodes of relatively intense post-erosional volcanism such as that which produced the Roque Nublo volcano on Gran Canaria (Pérez Torrado et al., 1995).

The eruptive history of the island of El Hierro is probably the best-constrained geochronologically of any of the Canary Islands. The uncomplicated development of the island, which is still in its juvenile stage of shield growth, together with the abundant and accurate K/Ar ages, and with magnetic stratigraphy (Guillou et al., 1996), allow the closest possible approach to the reconstruction of the entire emerged volcanic history of any of the Canaries. 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/ka. However, rates increase significantly if we take into consideration the three consecutive giant lateral collapses that affected the island, each clearly exceeding 100 km^3.

A similar evaluation of shield-stage magma production rates in the presently post-erosional islands is highly problematic. This is because it is impossible to evaluate the volume removed by lateral collapses (Stillman, 1999-this volume): it is difficult to determine even the number of collapses in these deeply-eroded islands, let alone the volumes of individual collapses.

2.4. Evolution of magmas

The shield-stage volcanism is characterised in the Hawaiian Islands by tholeiitic basalts, with late minor volumes of alkali basalts and associated differentiated magmas; during the rejuvenated stage, silica-poor magmas (alkali basalts, basanites and nephelinites) predominate (Langenheim and Clague, 1987).

In the Canaries, no such contrast is evident and magma compositions are varied in both stages. The rocks of the shield volcanism are predominantly basaltic (picrites, tholeiites and basanites) but with associated differentiated lavas (phonolites and trachytes). Highly differentiated felsic rocks occur in large volumes in the shield-stage volcanism of both Tenerife and Gran Canaria, and to a lesser extent in the other islands. Post-erosional rejuvenated volcanism repeats a similar trend but with much smaller volumes of rock involved in most cases, although the Pliocene Roque Nublo stratovolcano in Gran Canaria (Pérez Torrado et al., 1995) represents perhaps the most voluminous episode of post-erosional volcanism in any island in the world. Wide variations in alkalinity occur in post-erosional stage volcanism, sometimes within individual eruptions. The variation from basanites to alkali basalts seems to be a common feature in Holocene volcanic eruptions in the

Canaries but exceptional variations from basanites to alkali basalts and tholeiites in a single eruption have been observed in the 1730–1736 eruption of Lanzarote (Carracedo and Rodríguez Badiola, 1991; Carracedo et al., 1992). The latter is one of only two historic eruptions to have occurred in a post-erosional stage island in the Canaries (the other is the small eruption of 1824, also in Lanzarote).

A similar lack of contrast between shield-stage and post-erosional magma compositions may also be evident in other island groups on slow-moving plates. In the Cape Verde islands, magma compositions in all stages of activity are alkaline, and commonly extremely so. The only systematic trend identified by Davies et al. (1989) was spatial: they found that rocks in the northwestern islands (Sao Vicente–Santo Antao) were systematically more silica-undersaturated than those from the islands (Sal-Brava) which in their shield stage of activity were located above the inferred mantle plume head. The spatial and temporal patterns of compositional variation in hotspot-related island groups on slow-moving plates may therefore, in general, be more complex than in island groups on fast-moving plates such as the Hawaiian Islands.

2.5. Subsidence history

As already mentioned, all the Hawaiian islands subside and will eventually become seamounts. The amount, age and rate of subsidence have been derived from studies of submarine canyons, submerged coral reefs and coastal terrace (see Fig. 5) of lava deltas (Moore and Fornari, 1984; Moore, 1987; Moore and Campbell, 1987).

Evidence for the position of contemporary sea levels, in the form of marine abrasion platforms, littoral and beach sedimentary deposits, coastal volcanic deposits such as hyaloclastite-based lava deltas and Surtseyan tuff rings, and erosional palaeocliffs, is widespread in the Canary Islands (Fig. 5). It also invariably occurs close to present sea level, within the range of eustatic sea level change. Marine abrasion platforms up to several million years old and close to present-day sea level occur in the oldest islands of Fuerteventura and Lanzarote (Crofts, 1967; Lecointre et al., 1967; Fúster et al., 1968a,b; Meco and Stearns, 1981; Carracedo and Rodríguez Badiola, 1993; Meco et al., 1997). Even in the very young island of La Palma, still in the early phase of the shield-building stage, a beach shoreface sand deposit of about 0.5 Ma old is to be found at El Time, close to present sea level. Surtseyan tuff rings occur at present sea level in the coastlands of the majority of the Canaries (Fig. 5) as well as pillow lavas of ages ranging from the Miocene to present-day (Ibarrola et al., 1991; Carracedo and Rodríguez Badiola, 1993; Carracedo and Soler, 1995).

These observations are clearly incompatible with the model proposed by Watts and Masson (1996) that Tenerife has subsided by at least 2.5 km during the last 3–6 Ma. These authors base their model on some submarine features they observed in the swath bathymetry off the northern coast of Tenerife and they interpreted as subaerial gullies that subsequently subsided to form submarine canyons. An alternative explanation that does not require subsidence is that these submarine canyons form the underwater continuation of subaerial barrancos and have been carved by submarine erosive processes related to the discharge of the latter into the sea. Schmincke et al. (1997) have come to the conclusion, after analysis of extensive land and marine data, that the island of Gran Canaria has been remarkably stable with respect to sea level since about 14 Ma. These authors have identified nearly horizontal seismic reflectors of Late Miocene to recent age that can be traced to northern Tenerife, an observation clearly in conflict with the model proposed by Watts and Masson.

Therefore, it appears that the islands of the Canarian archipelago, located close to the African continental margin, are extremely stable and has undergone neither subsidence nor uplift after emergence. The lack of post-emergence subsidence is in very strong contrast to the rapid subsidence seen during the shield stage and later in the Hawaiian Islands (Moore, 1987; Ludwig et al., 1991; Moore and Chadwick, 1995), but is also a feature of the Cape Verde archipelago (Stillman et al., 1982; Courtnay and White, 1986).

The lack of post-emergence uplift, in contrast to the major uplift implied by the occurrence of the Seamount formations, is also of interest since it implies that prior to emergence large intrusive complexes grew within the volcanoes, whilst post-emergence, endogenous growth of the edifices was limited: this is consistent with the many geochemical and petrological data sets for subaerial volcanic suites in the Canary Islands which indicate that these suites are for the most part fed by magma reservoirs in the underlying oceanic crust and/or oceanic lithosphere with only the more evolved suites of rocks showing some evidence for the presence of shallow magma reservoirs (Hasteen et al., 1997; Kluegel et al., 1997, 1999-this volume; Mangas et al., 1997). The systematic occurrence of this pattern implies an explanation based on structural and mechanical differences between subaerial and submarine volcanic edifices.

The computer-generated image in Fig. 6 shows an ''empty ocean'' view of the Canarian archipelago from the east. Differences in height of the shield stage and post-erosional stage islands can be readily observed. However, this is not a constructional feature, since Fuerteventura and Gran Canaria may have reached similar or even greater elevations (Pérez Torrado et al., 1995; Stillman, 1999-this volume). In contrast to the Hawaiian and other oceanic island groups, this is not a consequence of subsidence, but of mass-wasting through erosion and catastrophic landsliding, as mentioned before (Carracedo et al., 1998, 1999-this volume; Stillman, 1999-this volume).

3. Main structural features of Canarian volcanoes

3.1. Rift zones

Rift zones that concentrate vents and dense dyke swarms have been described in the majority of oceanic volcanic islands (Wentworth and Macdonald, 1953; MacDonald, 1972; Fiske and Jackson, 1972; Walker, 1992, 1999-this volume). They may possibly be an invariable component of these islands, as noted by Walker (1992). Their typical three-armed rift zone geometry of oceanic island volcanoes, with angles of about 120° between them, was mentioned by Wentworth and Macdonald (1953), MacDonald (1972) and Carracedo (1994). The original regular geometry of rift zones is often lost during the evolution of the volcano, especially if buttressed among older volcanoes, as frequently happens in the very active Hawaiian volcanoes (Fig. 7B).

Rift zones can be observed in exceptional detail in the Canaries (Fig. 7A), by means of water tunnels or *galerías* (Carracedo, 1994). A simple model explaining the genesis of these features and their regular geometry was proposed by Carracedo (1994), based on previous mathematical work carried out by Luongo et al. (1991). Triple fracturing in a symmetrical pattern with angles of 120° would be the least-effort response to magma-induced vertical upward loading. As shown in Fig. 12B, these three-armed rifts at 120° were also detected using Bouger gravity anomalies in Tenerife (MacFarlane and Ridley, 1968) and La Réunion (Lesquer, 1990). Dyke swarms are clearly in relation with the manifestly regular geometry of the three spectacular erosive cirques of Piton des Neiges in La Réunion. Although difficult to observe, dykes generally were oriented parallel to the main axes of the corresponding cirques. A possible explanation for these erosive cirques is by a process of inversion of relief: erosion may have preferentially affected the softer pyroclastic facies associated with the concentrations of vents along the rifts. Anisotropy due to the dyke-swarms enhanced deepening of the gulches and active collapses at the steep walls enlarged them to become the present erosive cirques.

Apart from frequently losing their regular geometry through volcano evolution, rift zones are usually extensively dismantled during the post-shield-stage gap and the post-erosional stage due to the susceptibility of their pyroclastic rocks to erosion. As observed in the Hawaiian Islands, vents are no longer associated with the pre-existing rift zones during the rejuvenated stage (Langenheim and Clague, 1987) and this later stage of activity further covers and conceals these structures. However, detailed structural observations may help to identify the presence of these rift zones in other intraplate oceanic volcanoes.

4. Instability and collapse of Canarian volcanoes

Island volcanoes grow to a great height. Considering the edifice as a whole, they frequently rise several kilometres from the ocean floor. The western Canaries rise from about 6.5 km (El Hierro) to about 8.5 km (Tenerife). Gravitational stresses and dyke injections tend to progressively increase the mechanical instability of these volcanoes, especially in the most active shield-stage phases of growth.

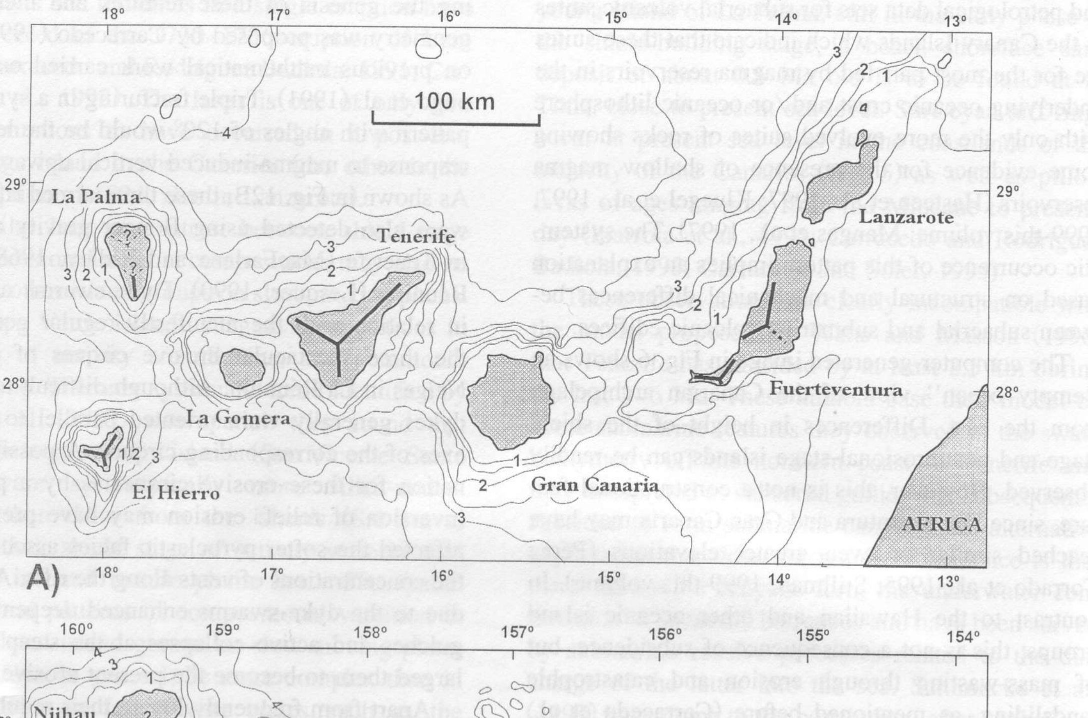


Fig. 7. Rift zones are a common feature of the Canaries and the Hawaiian Islands. In the former, rifts show a regular "Mercedes Benz" triple star configuration and are easily recognised in the shield-stage islands. This geometry, mentioned in the Hawaiian rifts by Wentworth and Macdonald (1953), is obscured in these islands probably by the active growth of the volcanoes.

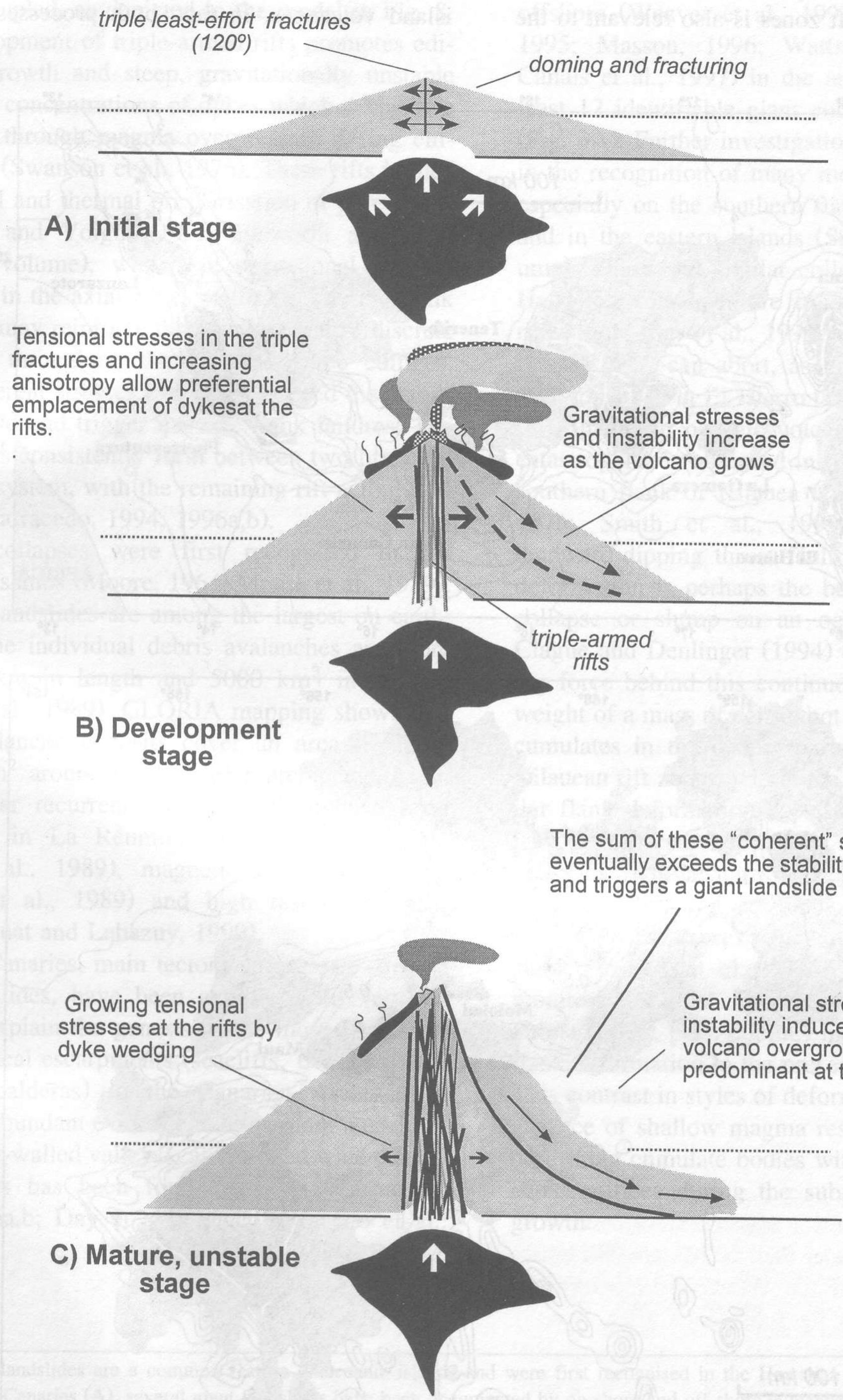
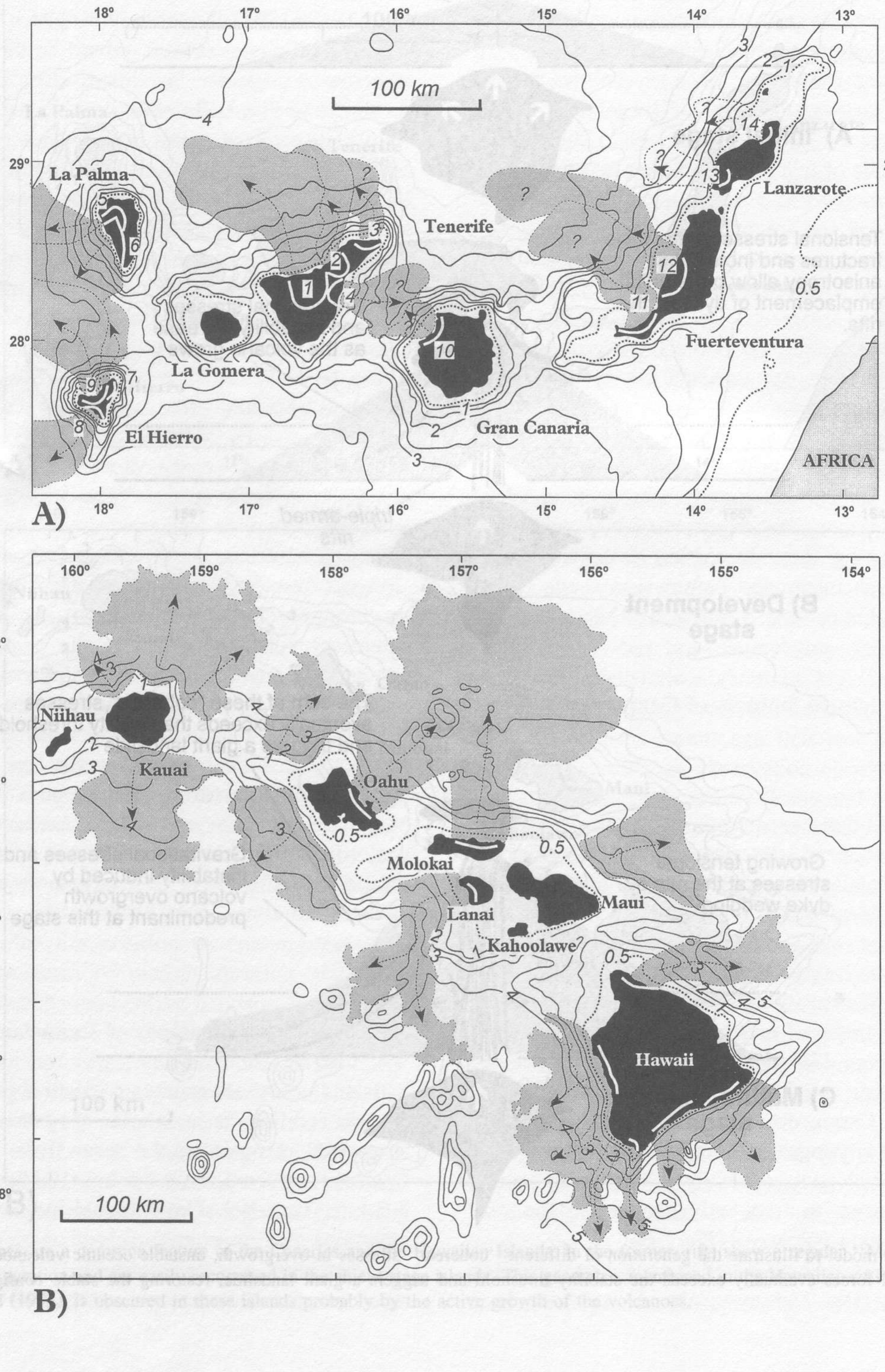


Fig. 8. Simple model to illustrate the generation of different ''coherent'' stresses in overgrowth, unstable oceanic volcanoes. The sum of these tensional forces eventually exceeds the stability threshold and triggers a giant landslide, restoring the stable configuration of the volcano.

The model proposed by Carracedo (1994) to explain the genesis of rift zones is also relevant to the progressive mechanical instability and collapse of island volcanoes. In fact, both processes (develop-



ment of triple-armed rifts and progressive instability) may be coupled, as depicted in the model in Fig. 8. The development of triple-armed rifts promotes edifice overgrowth and steep, gravitationally unstable flanks and concentrations of dykes which destabilise the flanks through magma overpressure during emplacement (Swanson et al., 1976). These rifts induce mechanical and thermal pressurisation of pore fluids (Elsworth and Voight, 1995; Elsworth and Day, 1999-this volume), while the extensional stresses developed in the axial zones of the rifts by the flank instability may reinforce the development of discrete rift zones throughout the growth of the edifices. These coherent stresses eventually exceed the stability threshold and trigger massive flank failures. The slide blocks consistently form between two branches of the rift system, with the remaining rift acting as a buttress (Carracedo, 1994, 1996a,b).

Giant collapses were first recognised in the Hawaiian Islands (Moore, 1964; Moore et al., 1989). Hawaiian landslides are among the largest on earth: some of the individual debris avalanches are more than 200 km in length and 5000 km^3 in volume (Moore et al., 1989). GLORIA mapping shows that debris avalanche deposits cover an area of about 100 000 km^2 around the Hawaiian archipelago (Fig. 9B). Similar recurrent large landslides have been recognised in La Réunion based on bathymetric (Lénat et al., 1989), magnetic and gravity data (Rousset et al., 1989) and high resolution sonar images (Lénat and Labazuy, 1990).

In the Canaries, main tectonic processes, such as giant landslides, have been avoided until very recently to explain the genesis of the more prominent morphological escarpments (seacliffs, U-shaped valleys and calderas) in the Canarian Archipelago. However, abundant evidence relating giant landslides and straight-walled valleys, calderas and wide coastal embayments has been found onshore (Carracedo, 1994, 1996a,b; Day et al., 1997; Carracedo et al., 1999-this volume; Stillman, 1999-this volume) and offshore (Weaver et al., 1992; Masson and Watts, 1995; Masson, 1996; Watts and Masson, 1996; Canals et al., 1997) in the archipelago, showing at least 12 identifiable giant collapses in the Canaries (Fig. 9A). Further investigations probably will result in the recognition of many more of these structures, especially on the southern flanks of the archipelago and in the eastern islands (Stillman, 1999-this volume). These and similar collapses on the flanks of Hawaiian volcanoes are frequently entirely submarine (McMurtry et al., 1999-this volume) and exceptionally they can abort, as did the San Andrés rift flank collapse in El Hierro (Day et al., 1997).

Collapse in oceanic volcanoes is not necessarily catastrophic, as observed in the slow collapse of the southern flank of Kilauea volcano (Swanson et al., 1976; Smith et al., 1999-this volume) on a landward-dipping thrust fault system. This ongoing deformation is perhaps the best example of a slow collapse or slump on an oceanic island volcano. Clague and Denlinger (1994) consider that the driving force behind this continuous deformation is the weight of a mass of dense, hot (and therefore plastic) cumulates in the deeper parts of the highly active Kilauean rift zones, within the volcanic edifice. Similar flank deformation does not appear to occur, at least in inter-eruptive periods, in the Canarian volcanoes: the edifices are essentially free of shallow seismicity between eruptions; there are no active, growing fault scarps comparable to the Hawaiian pali (Swanson et al., 1976); and, recent geodetic monitoring of the Cumbre Vieja volcano on La Palma (Moss, pers. comm.) indicates the absence of flank deformation in the present inter-eruptive stage. This contrast in styles of deformation may reflect the absence of shallow magma reservoirs or plastically-deforming cumulate bodies within the Canarian volcanic edifices during the subaerial shield stage of growth.

Fig. 9. Giant landslides are a common feature of oceanic islands and were first recognised in the Hawaiian archipelago (B) by Moore (1964). In the Canaries (A), several giant landslides have been documented by on-shore and off-shore evidence in the shield-stage islands. However, this must be a common feature in the juvenile stages of the remaining islands, where these landslides are more difficult to identify (Stillman, 1999-this volume). Giant landslides in the Hawaiian Islands are from Moore et al. (1989). Giant landslides in the Canaries are from Masson and Watts (1995), Watts and Masson (1996), Carracedo et al. (1999-this volume) and Stillman (1999-this volume).

5. Summary, discussion and conclusions

The Canary Islands have many important geological features in common with the Hawaiian Islands and many other oceanic volcanic groups. The main stages of growth of the Canarian volcanoes, from which general stratigraphic divisions can be defined, are comparable to other oceanic groups of the world.

The Canarian volcanoes also show common constructional and structural features, such as rift zones, progressive volcano instability and multiple gravitational collapses.

However, the Canaries, as well as the Cape Verde islands, present some important geological differences from the prototypical hotspot volcanoes of the Hawaiian Islands, possibly related to a comparatively low-activity hotspot and to a slow-moving plate.

The nature and evolution of magmas are quite different in both archipelagos: the main volcanic edifices in the Hawaiian archipelago are constructed by voluminous eruptions of tholeiitic basalts, while in the post-shield stage alkalic basalts and differentiates cover the main shield (MacDonald, 1972). In the Canaries, silica-poor magmas (alkalic basalts and differentiates) form the bulk of the islands, tholeiitic basalts being exceptional (Carracedo et al., 1992). Highly differentiated lavas (phonolites and trachytes) occur in large volumes in the central Canaries (Fúster, 1975).

Lower eruptive rates and possibly cooler magmas in the Canaries than in the Hawaiian Islands result in higher aspect ratio volcanoes. Slopes are generally steeper in the early shields which progress towards unstable configurations. Catastrophic flank failures in the Canaries have shallower slide planes and collapses have lower mass volumes than those of the Hawaiian volcanoes.

Perhaps one of the most important characteristics of the Canaries is the lack of significant subsidence at least over the last 15 Ma. This is relevant to explain why islands older than 20 Ma are still emergent, while islands in the Hawaiian-Emperor volcanic chain submerge by subsidence after about 7 Ma. A main characteristic of the Canaries is that their emergent phase will end after a very long time compared to the HI, and only because of mass-wasting processes (catastrophic in the early stages of development and erosion afterwards).

These features and favourable observational characteristics (lack of thick vegetation cover, very good rock exposures, access to the internal structure of volcanoes by means of water tunnels) made the Canaries an exceptionally advantageous location to observe the structural features and the evolutionary stages of oceanic volcanoes.

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