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Giant Miocene landslides and the evolution of Fuerteventura, Canary Islands

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Abstract

Fuerteventura is the oldest island in the Canaries archipelago; following the growth of a pre-Miocene seamount, its shield stage began 20.6 Ma ago, and ended in the mid-Miocene. There followed a very extended period of quiescence, then very minor Quaternary post-erosional volcanism. The shield stage produced volcanoes up to 3000 m above sea level which were rapidly eroded by 17.5 Ma. Both Fuerteventura and Lanzarote, the older eastern-most islands, are in a post-erosional stage and have subdued topography with elevations rarely rising above 400 m. The western islands, La Palma and El Hierro, together with Tenerife, are the youngest in the archipelago, and are still in their shield stage which began at most 7.5 Ma ago. They have a rugged topography with peaks rising to several thousand metres. Volcanic activity persists in these islands and in Lanzarote. The shield-stage construction of the volcanically active western islands demonstrates that recent eruptive activity has concentrated in volcanic centres aligned in well-defined rift zones in which major dyke emplacement has taken place, and that volcano flank collapse has occurred with giant landslides initiated on the rifts. In Fuerteventura, most of the shield-stage volcanic rocks have been removed; however, evidence has now been advanced to show that major dyke swarms were emplaced in the Oligocene and Miocene. These belonged to a line of intrusive complexes above which, by the early to mid-Miocene, volcanic peaks had been constructed. Three volcanic centres have been identified and in the central one, a peak as high as the present Mount Teide on Tenerife was rapidly denuded in less than 2 Ma. In the west of the island, this erosion has exposed a window of more than 300 km^2 of submarine volcanics and sediments belonging to the pre-shield phase seamount. The principal mechanism of erosion seems to have been the generation of giant landslides analogous to those seen in the younger islands. These appear to have removed some 3500 km^3 of lavas and volcaniclastics, stripping the western sectors of the Miocene volcanoes down to the pre-shield phase ‘‘Basal Complex’’, and transporting the mass of volcanic material into the Atlantic ocean as debris flows. Remnants of the shield volcanoes still exist in the eastern part of the island. Present day topographic variation indicates that the Miocene base level of erosion is still visible in a ‘‘Central Depression’’ which runs through the axis of the island, the eastern hills are composed of the Miocene shield-stage volcanics which still preserve a drainage pattern initiated on these peaks, and the ‘‘Basal Complex’’ in the west has a juvenile landscape suggesting incision on a post-Miocene domal uplift of the volcano core. Structural links between intra-island volcano-tectonic events and the South Atlas Fault zone are not thought to be likely.

Keywords: Canary Islands; Miocene; basal complex

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1. Introduction

1.1. Location and volcanological history

The Canary Islands are a group of volcanic islands off the northwestern African coast which have been generated above a thermal anomaly in the mantle. The islands were formed by multiple volcanic episodes (Schmincke, 1982; Carracedo, 1994, 1996) and each show different evolutionary histories. In general, the age of the earliest emergent shield formation decreases from east to west, but this does not indicate a regular westward migration of the thermal anomaly. Indeed, the most easterly, Lanzarote, still maintains a very high heat flow and eruption has taken place as recently as 1824 AD.

The subaerial volcanic history of the western islands Tenerife, La Palma, and El Hierro, is short, starting no earlier than 7.5 Ma (see Fig. 1) and activity has continued with only relatively brief periods of quiescence to the present (Carracedo, 1996). In the eastern islands, subaerial volcanicity started substantially earlier: Gran Canaria in 14.5 Ma, Lanzarote in 15.5 Ma, and Fuerteventura in 20.6 Ma. In all these islands, the shield phase was followed by a very extended period of volcanic quiescence and erosion, followed by very minor Quaternary post-erosional volcanism (Carracedo, 1996).

1.2. Processes causing collapse in volcanic islands

Active volcanoes are prone to the development of structural instability, and island volcanoes demonstrate this throughout their lives, the submarine pedestals being largely composed of pillow lavas and hyaloclastite, which rapidly becomes palagonitised and altered to weakly coherent clay minerals. Though laced together by a network of small sheet intrusions, nevertheless, the flanks are prone to oversteepening and spalling off as flows of water-saturated clastic material which generate proximal debris flows and distal turbidites.

Once emergent and sealed from the sea, the shield volcanoes build up, sometimes to great height, but all the while, the structure is being distended by the injection of dyke swarms and high-level plutons. As pointed out by McGuire (1996), instabilities develop, and flank collapse is triggered sometimes by volcano-seismic shock associated with eruption, sometimes by distension and increase in pore fluid pressure as a result of dyke intrusion, and sometimes by those associated with swelling of the flank due to high-level magma emplacement. Although the

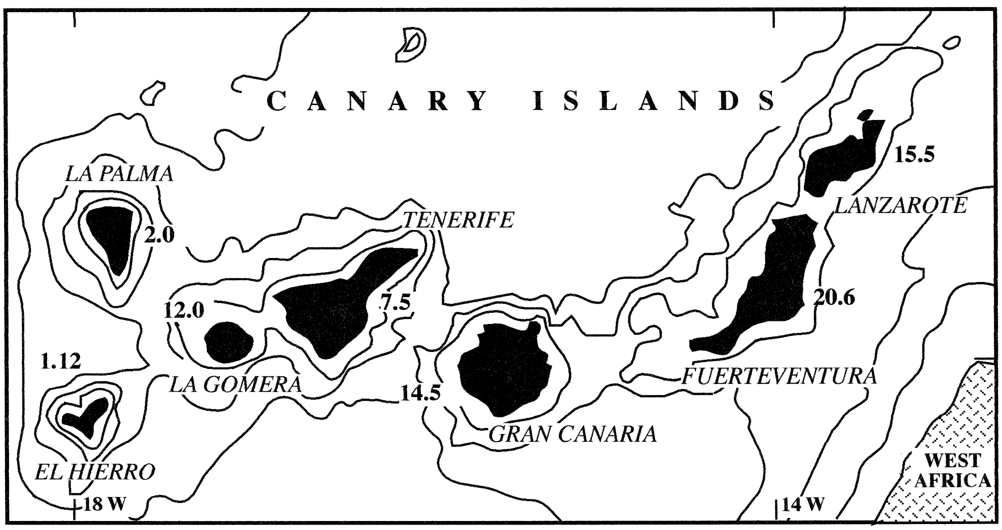

Fig. 1. Location map of Canary Islands. Numbers indicate ages (in Ma) of oldest shield-stage lava.

magma has a significant part to play, the single linking factor in both the development of instability and the initiation of structural failure is the influence of gravity. Not only does this directly affect the edifice stress regime so as to increase destabilisation as the edifice grows larger with time and factors such as oversteepening and overloading become more prevalent, but it also provides the energy for the post-failure transport of the detached material (McGuire, 1996). Since the observed failure of the northern flank of Mount St. Helens in the May 1980 eruption, such collapses and the resultant giant debris avalanches have been identified as a major process in the mass wasting of volcanoes, and deposits produced by collapse of the volcano flanks have been recognised around both continental volcanoes and oceanic island volcanoes such as the Canaries (Holcomb and Searle, 1991; Watts and Masson, 1995; Masson, 1996).

1.3. Contrasts in geomorphology and landform evolution

The volcanoes of the younger western islands are still high mountains; Mount Teide on Tenerife is over 4000 m. On these islands, relatively fast developments of volcano instability and often instantaneous mass wasting have been common features in the recorded volcanic history and can be readily traced in the stratigraphy (Carracedo et al., op. cit.). In particular, the existence of giant landslides is now indubitable. (Ancochea et al., 1990, 1994; Carracedo, 1994, 1996; Watts and Masson, 1995).

Recent eruptive activity has been concentrated in well-defined rift zones in which major dyke emplacement has taken place. Such rift zones are thought to play a key role in both construction and destruction of oceanic volcanoes and Carracedo (1996) believes that the forceful emplacement of the dykes could be responsible for incipient displacement of unstable, unbuttressed flanks of rifts. This view was given support by the 1949 eruption in La Palma which was accompanied by the formation of broadly curving faults, probably resulting from stresses that tended to collapse the steep west flank of the southern rift. Similar processes may have generated the huge calderas of Taburiente and Cumbre Nuevo; in effect, mass-wasting responses to the overgrowth of the rift, nearly 5 km above the sea floor (Carracedo, 1996).

The older eastern islands, Fuerteventura and Lanzarote, have, in contrast, rather subdued topography. There are no high volcanic mountains, indeed the highest points on Fuerteventura seldom exceed 400 m. Until now, it has been assumed by most workers that the landscape is a mature one which has been moulded by gradual erosion over the very long available time span of 10–12 million years. However, evidence is presented here which suggests that this may not be the case, and that the processes of construction and mass wasting operated in Fuerteventura in much the same fashion as on the western islands, but some 18 million years earlier.

It may be thought that the variation in elevation could also be due to differences in heat flow between the still active western islands and the extinct eastern ones. Some volcanic islands do subside as the thermal plume moves away and the crust beneath cools down. But this does not seem to be the case for the Canary Islands. Firstly, Lanzarote still exhibits a high heat flow. Secondly, though there is ample evidence of uplift during the lengthy periods of volcanism which built the islands, there is no geological or geomorphological evidence of subsequent sagging or downwarping of the crust. Furthermore, recent seismic wave studies (Canas et al., 1995) have indicated that such crustal sagging is unlikely as both the crust and the asthenosphere are strong, and any weakness due to the partial melting responsible for magma generation must have been at considerable depth.

1.4. Reasons for studying Fuerteventura

Fuerteventura presents a number of features which are unique within the Canaries archipelago. Firstly, it is the oldest island and on its western side are found extensive exposures of the submarine seamount stage of growth, with a stratigraphy extending from the Jurassic to late Oligocene/Miocene. No other island provides such exposure of the pre-shield stage construction. Secondly, its erosional history is far from clear. Its emergent shield stage began 20.6 Ma ago, and the main construction effectively ended 3.5 Ma later, in the mid-Miocene. This shield stage produced very substantial volcanoes rising up to 3000 m above sea level, which were rapidly eroded by 17.5 Ma and now only remnants are to be seen along the eastern side of the island (Ancochea et al., 1996).

The subdued topography seen today is close to that of the mid-Miocene erosional surface, (Ancochea et al., 1996), on which there has been emplaced a small number of mid-Miocene lava flows, followed much later by a small number of Quaternary cinder cones and lava flows of limited extent. The process by which this surface was produced is the subject of considerable speculation and on-going research. This paper puts forward the view that, on the basis of existing knowledge, it seems likely that the principal mechanism of erosion was the generation of giant landslides analogous to those seen in the younger islands.

2. Fuerteventura — the constructive phase, from the Palaeocene to Miocene

2.1. The seamount stage of growth

Much of Fuerteventura is covered by Miocene and younger volcanics, but uniquely in the archipelago, erosional processes have exposed a large window of more than 300 km^2 of the pre-shield stage edifice.

This complex previously known as the ''Basal Complex'' (Fuster et al., 1968; Stillman et al., 1975), comprises a submarine stage with pillow lavas and hyaloclastites, volcaniclastics and sheet intrusions, which passes up to subaerial lavas and pyroclastics. The layered rocks are intensely intruded by dyke swarms and plutons, many of which belong to the cores of subsequent Miocene volcanoes.

Within the complex, marine sediments from latest Early to Middle Jurassic are exposed; these are the oldest rocks seen in the archipelago and apparently formed the ocean floor on which the seamount was constructed (Steiner et al., 1998). These predate the initiation of plume volcanism, but the latest (Albian and Senonian) deposits appear to be associated with strong uplift (Robertson and Stillman, 1979a) which may have signalled the arrival of the plume head which was to power the subsequent magmatism of the Canary Islands. The first pillow lavas follow this uplift.

In the area of Basal Complex at present exposed to view, intense injection by sheet intrusions makes it impossible to construct a continuous stratigraphic log; for the submarine series, however, an incomplete chronology can be constructed from the large number of K–Ar and ^{39}Ar–^{40}Ar age determinations published in recent papers by Feraud et al. (1985), Le Bas et al. (1986), Coello et al. (1992), and Ancochea et al. (1996). These suggest that the earliest pillow lavas are probably Palaeocene to early Eocene and that the build-up continued to the early Miocene with repeated intrusion of dykes and small plutons. All of these lithologies are cut by intense NNE-trending dyke swarms. The earliest date available for subaerial volcanics in the Central Complex is 20.4 ± 0.4 Ma (Coello et al., 1992). The geochronological evidence thus indicates a submarine build-up from Palaeocene to late Oligocene times, with emergence before 20 Ma, in the late Oligocene to early Miocene.

More indirect evidence of the timing of emergence is given by some inter-lava sediments seen in coastal outcrops south of Los Molinos, in the north of the Basal Complex (see Fig. 2).

Just below the planar unconformity at the top of the sequence, late Oligocene to early Miocene interlava bioclastic and volcaniclastic sediments, dated on the basis of diagnostic benthonic foraminifera, represent the highest exposed levels of the submarine sequence, and were apparently deposited close to the submarine flanks of an emergent island (Robertson and Stillman, 1979b). The lower parts of these deposits are interpreted as comprising planktonic radiolarian micrites deposited in an open marine environment interbedded with calcarenites originated by mass flow on steep slopes. Benthonic foraminifera suggest redeposition from a shallow water environment. Above these are conglomeratic calcirudites laid down as a series of debris flows derived by disintegration of algal and coral reefs. Upwards, volcaniclastic sandstones become more abundant. In the conglomerates and sandstones are clasts of terrigenous volcaniclastics, lavas, and most significantly, fragments of hypabyssal and plutonic igneous

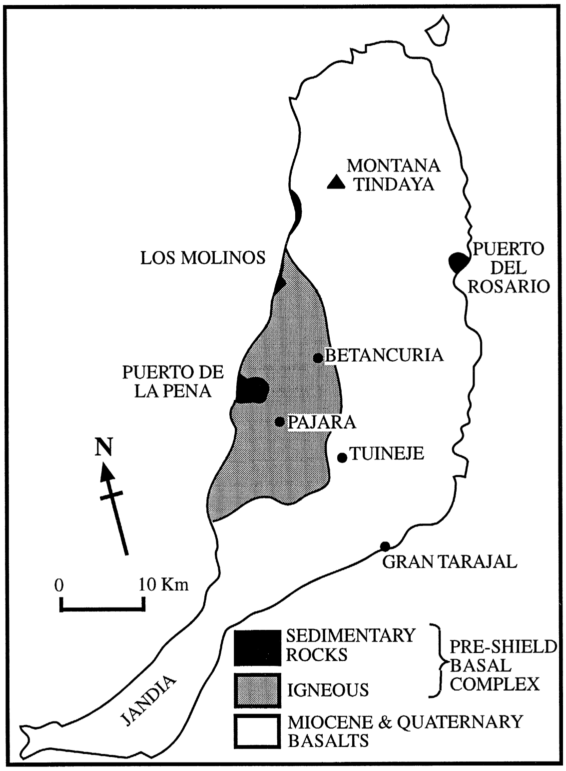

Fig. 2. Fuerteventura, with area of outcrop of the pre-shield Basal Complex (boundary approximate) and place names mentioned in text.

rocks which show that the source area was at this stage undergoing subaerial erosion which had unroofed some plutons (Robertson and Stillman, 1979b).

These shallow water marine deposits give clear evidence of erosion of an adjacent emergent landmass which provided terrigenous volcanic and intrusive igneous clasts.

2.2. The subaerial stage

Detailed study by Ancochea et al. (1991) and Ancochea et al. (1996), has shown that the subsequent subaerial volcanism constructed shield-stage volcanos which reached their maximum elevation in the early Miocene. The lavas and volcaniclastics, named the Older Basalt Series by Ancochea et al. (1996), are the upward continuation of the subaerial part of the Basal Complex, and these workers have identified three recognisably individual volcanic structures designated as the Northern, Central, and Southern Volcanic Complexes (NVC, CVC, and SVC) (Fig. 3), each of which has outward-dipping lavas, and radial dykes. They also show that the transition from seamount phase to subaerial Miocene volcanics varies in time from one centre to the other. In certain areas, this transition is marked by an unconformity; in others, it appears to be conformable. In many places, it is marked by deep weathering and erosion, though in at least one case, it is difficult to observe a significant break until later in the Miocene succession. The ages of these volcanic complexes differ slightly — the CVC is the oldest with the bulk of construction occurring between 20 and 18 Ma, followed by a significant pause, and a later smaller phase from 17.5 to 13 Ma; in the NVC, the main activity took place between 14 and 12 Ma, and in the SVC, between 16 and 14 Ma (Coello et al., 1992).

Whilst subsequent erosion has removed most of the Miocene volcanics from the western half of the island, sufficient remains in the east to permit reconstruction of the volcanic topography of the three complexes. For the CVC, the summit would have been above the gabbro and pyroxenite plutons seen today in the vicinity of Pajara, (Fig. 2) and have reached a height of 3000 m above sea level — an estimate which concurs very well with that suggested by the stable isotope data from the dykes. From the stratigraphy and ages of the plutons, the youngest of which — the Vega syenite/gabbro ring complex, the roof of which intrudes subaerial lavas — has an intrusion age of 20 Ma; this elevation was most probably reached not earlier than 20 Ma.

Two other structures shown by the bathymetry to be small seamounts off the southern tip of Fuerteventura, are also thought by these authors to be comparable volcanic centres.

The alignment of all these centres bears a strong resemblance to the ‘‘rift’’ alignment of volcanic centres quoted by Carracedo (1994; 1996) as being characteristic of the construction of the western Canaries. The pattern of dyke and high-level pluton emplacement would support this view. Whilst some of the dykes were clearly part of the seamount construction, it is apparent that many were emplaced later during a period of intense linear fissural injec-

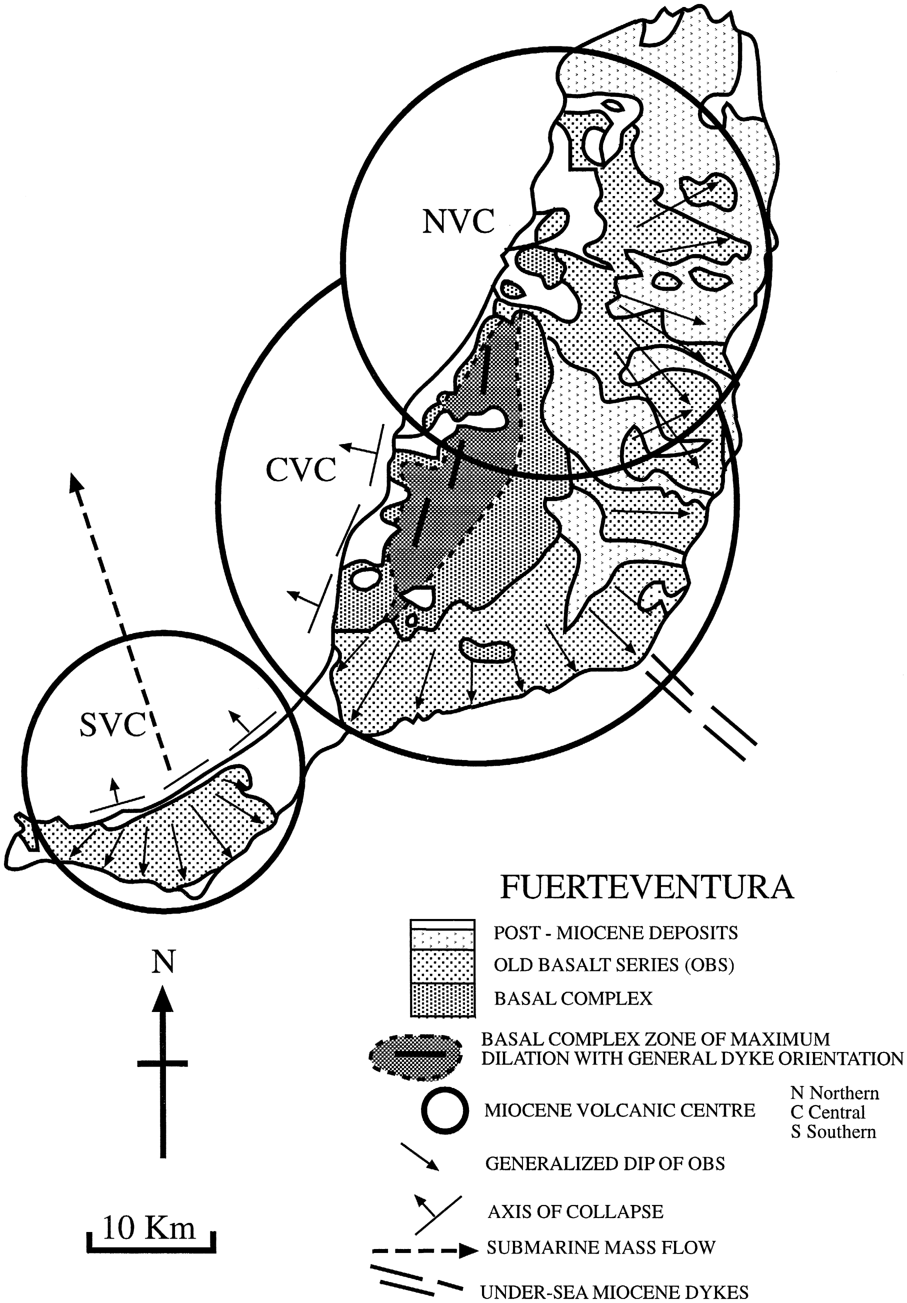

Fig. 3. Miocene volcanic centres after Ancochea et al. (1996, fig. 6). Within the Basal Complex, the area of maximum dyke intensity is outlined, with general dyke orientation shown.

tion which lasted into the early Miocene, producing complexes dominated by swarms of NNE-trending parallel dykes emplaced in en-echelon fissures, in the axial zones of which the total dilation exceeded 80% (Stillman, 1987). These intense swarms have many of the features of the coherent intrusive complexes described in Hawaii by Walker (1986; 1992) and are associated with small high-level plutons that appear to be co-magmatic with many of the dykes. Such dyke complexes, accepted as a common feature in the growth of many oceanic volcanoes, are well represented in the Canaries.

3. The end of the constructive phase, and the destruction of the CVC volcano

Almost the whole bulk of the CVC had been built by about 20 Ma, when the intensive constructional volcanicity which had been nearly continuous for almost 35 Ma came to an end. Following this, there appears to have been a pause in volcanic activity coincident with major erosional activity which stripped the shield basalts away, exposing and weathering the Basal Complex to produce a topographic surface which was apparently close to that seen today. Onto that surface, a small area of mid-Miocene basalts belonging to the Malindraga, Tamacite, and Tableros Formations (17.5–13.2 Ma) flowed, locally filling the palaeotopography and fossilising a relief excavated into the Basal Complex (Ancochea et al., 1996).

Evidence for the timing and mechanism of denudation comes both from the geomorphology and the geology. Whilst there is little evidence for the geomorphic history of much of the island from about 13 Ma to 5 Ma, there does appear to have been no volcanic construction during this period, and the preserved Miocene volcanics which cover almost half the island still display a drainage system of ravines which appear to have been initiated in the early Miocene when the high volcano peaks existed. Thus, the volume of material removed can be very roughly calculated from the estimated dimensions of the Miocene volcanoes (Ancochea et al., 1996). This gives a figure of the order of 3000 km^3. If erosion took place steadily over 20 Ma, it would require a rate of 150 km^3 per million years, or 150 m^3 per year; sufficient to denude the island without resort to mass movements. However, examination of the present topography and volcanic stratigraphy indicates that erosion has not proceeded at a steady rate from the mid-Miocene to the present.

3.1. Geomorphological evidence for the timing of volcano denudation

The topography of the CVC and NVC can be described in terms of three geomorphic regions, which reveal a considerable correspondence between geology and topography. From west to east, these are the Basal Complex dome, the Central Depression, and the Miocene lavas cuchillo and ravine terrain.

3.1.1. The Central Depression

The Central Depression holds the clue to geomorphic history from the mid-Miocene to recent times. It is an almost level plain with an overall height between 100 and 200 m, on which are found thick successions of recent sediments, calcrete, colluvium deposits, and Pliocene to Pleistocene volcanics — lava flows and small cinder cones. These deposits cover and largely obscure an eroded surface of both Basal Complex and Older Basaltic Series subcrop.

The Depression was developed after the early Miocene eruptions, during and subsequent to local deep weathering of Basal Complex lithologies, and before the deposition of the much less abundant later Miocene volcanics. A minimum age of around 18 Ma for its development may be deduced from the Tamacite Formation dated at 17.55 Ma and the Malindraga Formation, dated at 18.29 Ma (Ancochea et al., 1996); units of which are seen in the area of Las Huertos de Chilegua to have flowed onto the depression and filled a palaeotopography excavated into the Basal Complex (Ancochea et al., 1996). Its surface represents the depth to which the early Miocene 3000 m volcano was excavated.

3.1.2. The western dome

In the west is an elongated oval area of elevated ground (between 200 and 300 m) in the stream valleys, of which are exposed abundant outcrops of the seamount stage of the Basal Complex (see Fig. 4b). This area displays a geomorphologically immature landform developed on a low aspect domal surface which is heavily weathered and covered in calcrete, colluvium, and local quartz sand deposits identical to those in the Central Depression.

The age of the doming event is unknown, but the extremely juvenile drainage pattern with radial streams and interlocking spurs incised into a surface still covered with similar deposits to those found on the lower-lying level plain of the Central Depression, suggest localised and relatively recent uplift of what was the depression floor. It seems highly likely that prior to the uplift of the Dome, the low land surface of the Central Depression would have extended westward to the sea.

The overall stratigraphy in the CVC is characterised by an eastward progression of younger strata forming arcs around the deeper-seated pluton-dominated area which would have been directly beneath the peak of a Miocene volcano (Ancochea et al., 1996) and the plutons' exposure at today's surface elevation suggests uplift of a volcano core.

The uplift may have occurred as an isostatic rebound following the stripping of the volcanic superstructure, or alternatively, it may have been related to an increased mantle heat budget associated with the Pliocene Betancuria volcano (Coello et al., 1992). In either case, it may well have caused the uplift which has produced the raised beaches around the western shores of the island. In at least one outcrop, Pliocene beach deposits rest on an erosional surface cut across a weathered Basal Complex, and are overlain by c. 5 Ma basalts with pillowed bases and columnar tops which have run across the beach (see Coello et al., 1992). These have been subsequently raised more than 30 m above present sea level.

3.1.3. The Miocene lavas cuchillo and ravine terrain

Along the eastern edge of the Central Depression, the ground rises steeply to a ridge crest at around 300 m above sea level, from which the land surface slopes eastward to the sea, with a gradient of around 20°. This slope is dissected by a striking series of deep ravines and sharp-crested intervening ''cuchillos'', draining radially away from the island centre. Miocene eruptives are the only lithologies exposed here; the residual eastward-sloping surface appears to be that of the early Miocene volcanoes, and the drainage pattern was apparently initiated on the slopes of the volcanic peaks.

This terrain represents all that is left of the high early Miocene volcanoes, and the youngest of the remaining basalt successions constrains the maximum age of the denudation process. In the Central Complex, the land surface west of this terrain — that of the Central Depression and the dome — appears to be the product of a relatively short and intense period of erosion which decapitated the peaks and reduced the elevation to something close to today's surface, before the eruption of the Tamacite and Melindraga Formations (Ancochea et al., 1996). Thus, the period of erosion may well be constrained to a period of some 2 million years.

3.2. Geological evidence for the mechanism of volcano decapitation

Miocene Fuerteventura appears to have had many constructional features which are comparable with those in the much younger western Canaries. Carracedo (1996) comments that in the islands with Quaternary volcanic activity, the construction has been closely controlled by rift-type volcano-tectonic features. They show forceful injection of dykes associated with extension, and at erosional windows, coherent swarms (Walker, 1992) of feeder dykes can be observed with increasingly dense packing with depth towards the axis of the rifts. Miocene Fuerteventura fits this description with remarkable accuracy, and, as shown above, the shield volcanoes appear to have attained elevations similar to that of Teide in Tenerife.

The destruction was likewise relatively rapid; the Fuerteventura volcanic edifices were apparently reduced from 3000 m to around 200 m in a period of less than 2 million years. The processes by which this was achieved may be deduced both from the evidence available on the island, and by analogy with the younger western Canary Islands which Miocene Fuerteventura is now seen to closely resemble. The topography, in particular, the Central Depression, might suggest some type of rift-controlled tectonic subsidence. However, extensive tectonic foundering as a principal means of disposing of the western halves of the volcanoes (Ancochea et al., 1996) appears to be unlikely, since the rocks exposed beneath the Miocene erosional surface are not those of the upper parts of the shield volcanos, but those from varying levels in the seamount, including the pre-volcanic Jurassic to Cretaceous ocean floor sediments, which would have been below the whole subaerial edifices.

The upper Oligocene–lower Tertiary marine sediments of the Basal Complex do suggest that the excavation of the land surface at that time was sufficiently profound to expose some plutons, and the debris flows indicate slope instability. Flank failure would be a process analogous to that in other islands, and one plausible mechanism would seem to by flank failure and mass wasting, initiated, if not completed, by a series of giant landslides carrying the volcanic superstructure north-westward into the sea, in a manner analogous to the giant landslips on El Hierro, Tenerife, and La Palma (Carracedo, 1996).

Repeated or multiple slips could explain the denudation which cut back to expose the seamount, created the Central Depression, and retained the remains of the early Miocene lava slopes around the eastern side of the island, whilst the western flank was transported into the sea. This process would require the proximal disposal of vary large quantities of Miocene volcanics — possibly producing the bathymetric feature seen west of the CVC.

Since on Fuerteventura all the deposits involved have long been removed, an attempt to quantify this model requires a number of assumptions to be made. Firstly, the dimensions of the early Miocene volcano are taken as those suggested by Ancochea et al. (1996) — which correspond well as regards height with the estimates from stable isotope studies (Javoy et al., 1986). Secondly, the present geology and topography indicate that any such slippage must have been to the north or west, permitting the survival of the eastern arc of early Miocene lavas and

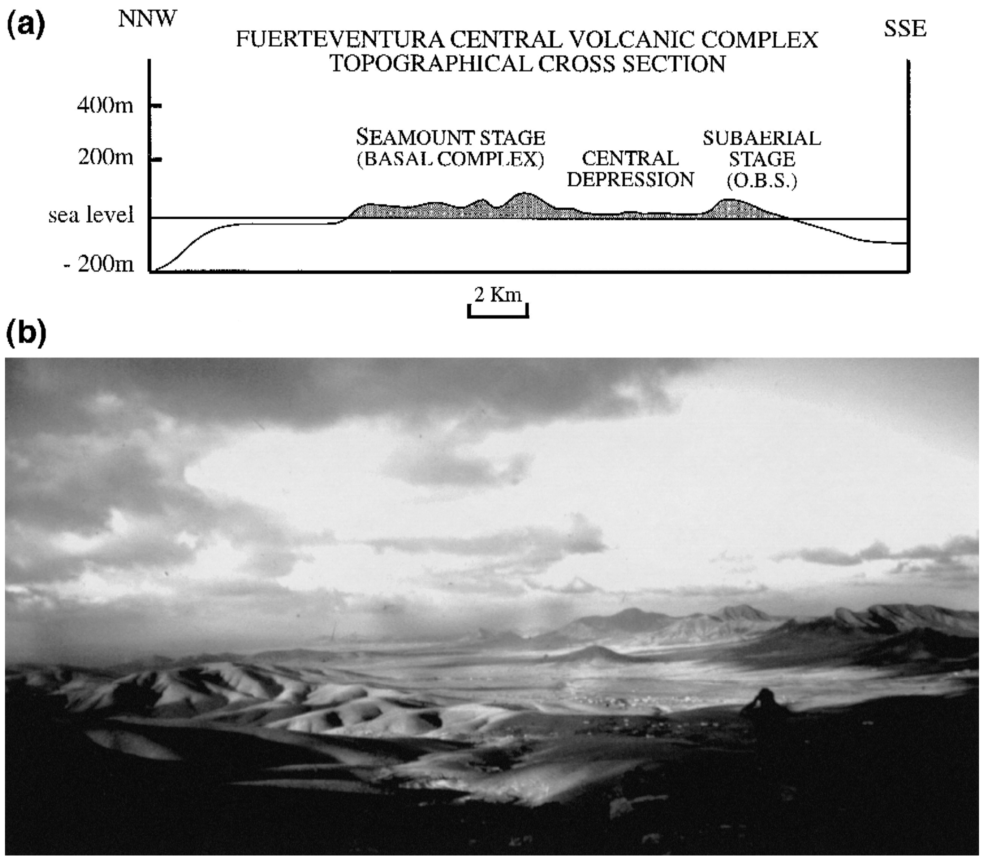

Fig. 4. (a) Topographic cross-section NNW–SSE across Fuerteventura, from the coast in the vicinity of Puerto de la Pena to the coast N of Gran Tarajal. (b) Panoramic view looking north towards Val de Santa Inez; left to right: Basal Complex, Central Depression, Miocene lavas.

pyroclastics. It is also postulated that the distension caused by the emplacement of dykes and plutons within the volcano superstructure (measured as up to 80% in the most intense areas) may have preferentially weakened northern and western slopes, producing a bulge like that seen on Mount St. Helens prior to the 1980 eruption, and facilitated flank failure as an initial giant landslip.

The model, shown in Fig. 5, is based on an estimated peak height for the CVC volcano of 3500 m above sea level and a slope angle of 12°. As a first approximation to quantify the process, a landslip model has been constructed using calculated σ^2 (confining pressure) stress trajectories for rock close to a steep slope, from which slide surfaces generated at an angle of 45° have been plotted (Allen, 1997; after Middleton and Wilcock, 1994, fig. 4.26). It can be seen that a single flank collapse so generated would not excavate the full area of the Basal Complex. However, the fact that this was an active volcano means that extra destabilising features may also be taken into account, in particular, the effects of intensive dyke intrusion (Elsworth and Voight, 1995, 1996). To the distension caused by dyke injection may be added the bulging and oversteepening of slope caused by the uprise of significant magma chambers, as occurred in the Mount St. Helens eruption of 1980. A further factor is the effect of destabilisation of the volcano slope by variations in the pore fluid pressure exerted by hydrothermal fluids associated with the emplacement of the dyke swarm (Day, 1996) — fluids for which there is ample evidence in the Fuerteventura dyke mineralogy. From modelling carried out on the 1949 eruption of the Cumbre Vieja volcano on La Palma (Elsworth and Day, 1997), one result is the generation of a steep upper segment of slip plane slope and a near-horizontal lower segment. Such a model fits well with that postulated by Carracedo (1996) for the western Canaries, and by Holcomb et al. (1997) for the collapse of Wailau Volcano, East Molokai, Hawaii. A slide geometry of this type would provide a mechanism for a wide zone of dislocation, and if continued magma uprise caused the volcano to distend and erupt several times during this period, then successive landslips may well have occurred (see Fig. 5). This situation would find a close analogy in the history of La Palma over the past 2 million years during which a succession of calderas have been generated along the alignment of the axial rift (Carracedo, 1996).

3.3. Evidence for mass-wasting

In the CVC and NVC, this evidence is scarce. An indication of the early erosional topography is given by the unconformity below the mid-Miocene Melindraga Formation in the area WSW of Tuineje (Fig. 2). This takes the form of a series of generally northward to northeastward facing steep arcuate

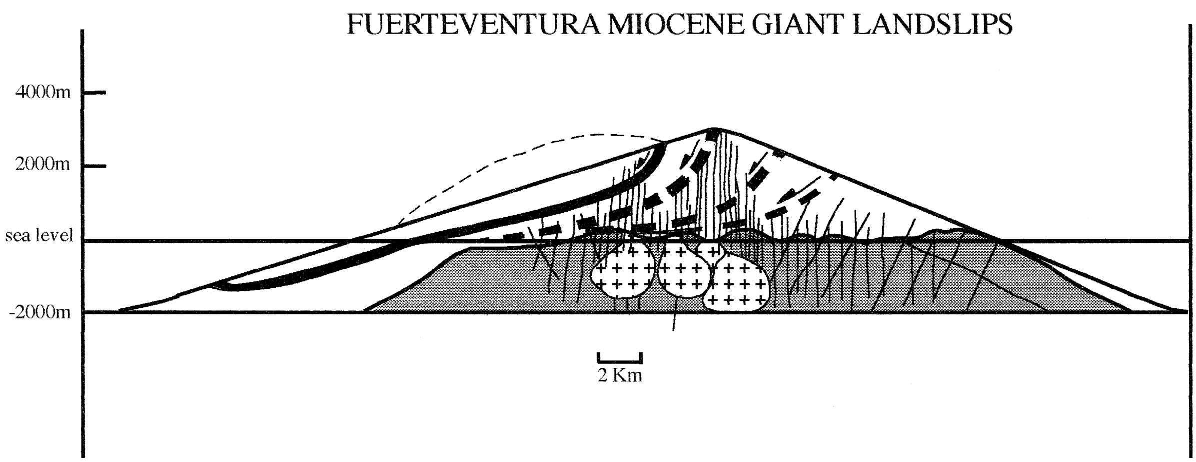

Fig. 5. Model cross-section for the collapse structures which exhumed the Fuerteventura Basal Complex. (Line of section as for Fig. 4a.) For model parameters, see Section 3.2. Solid heavy line — the calculated slide surface; dashed heavy lines — postulated successive slip planes. Dashed light line — the postulated flank distension caused by dyke emplacement and magma chamber uprise.

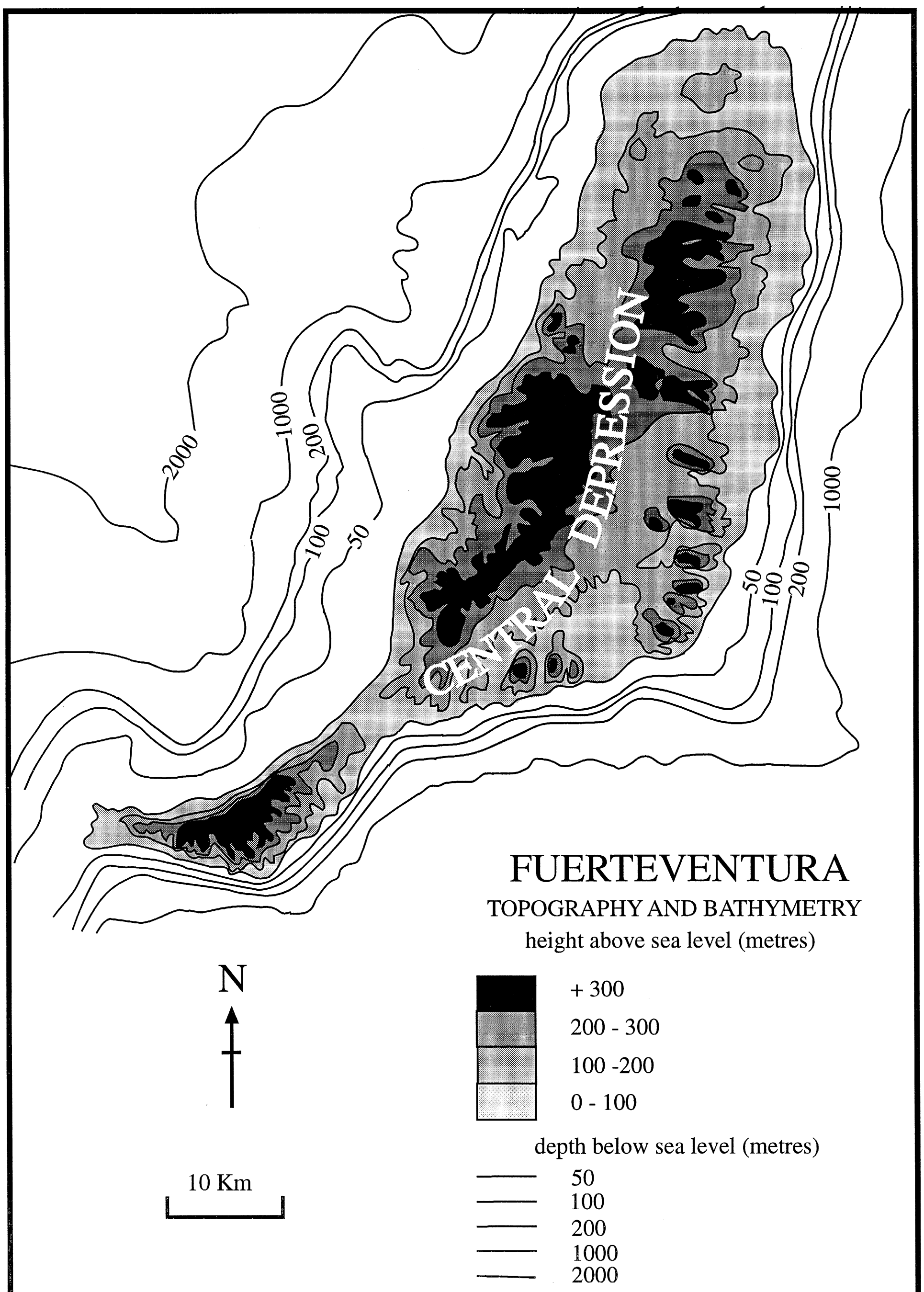

Fig. 6. Topographic map of Fuerteventura; on-land contours at 100 m; bathymetric contours at 50, 100, 200, 1000 and 2000 m depth below sea-level.

slopes which truncate dykes and bedding features in the underlying Basal Complex. The surfaces are weathered and some detrital clastic sediment is seen beneath 100–200 m of Melindraga lavas which fill the valleys and flow down to the Central Depression.

The only significant debris deposits so far identified are those of the youngest sedimentary deposits of the Basal Complex seen in coastal outcrops south of Los Molinos (see Section 2.1 and Fig. 2). These give evidence of early-Miocene deep erosion of an adjacent landmass which provided clasts of both volcanics and plutonics from the unroofing of an emergent volcano. The mechanisms of erosion may not be defined from these deposits, but their internal fabric, clast population, and fauna give evidence of mass-flow movement which transported the subaerial material to open marine waters. Clasts of similar material are reported in debris-flow deposits in a core taken from DSDP site 397 some 90 miles (145 km) SW of the island (Schmincke and Von Rad, 1979) dated at 17.2 and 16.5 Ma.

Ancochea et al. (1996) deduce that the volcanic complexes may well have been centred offshore to the west of the present coastline, and in the SVC, there are some significant offshore morphological features to support this view. To quote Ancochea et al. (1996) ''On the southern flank of this complex the submarine floor descends rapidly whereas on the northern coastal line the submarine slope is gentle. This would indicate that it (the complex) must have extended further north. The centre of submerged relief coincides with that derived from the morphology and the dyke systems, and supports the hypothesis of a large central volcano centred offshore . . . (off) the central part of the Jandia peninsula the arcuate shape of the isobaths may denote a northward slide causing a nearly 12 km wide depression. The Jandia escarpment probably originated as a consequence of that slide. The scarp may have retreated later through erosion''. The arcuate form of the scarp may possibly be controlled by the dominant dyke trends; if so, the age of the dykes (c. 15 Ma) would give a maximum age for this collapse.

The bathymetric contours in fact indicate a valley almost 12 km wide descending northward for 30–40 km from the shore to depths of over 2000 m (see Figs. 3 and 6). This may well be the route of a debris or mass flows derived from the slide.

The implication here is that major debris flows would have initiated in the region now offshore and could only be sought in the sea floor.

The rather sparse bathymetric detail west of the CVC and NVC is less readily interpreted. The sea surface slopes gently to the NNW for a distance of 5–10 km, then steepens rapidly to a depth of 1000 m, and Ancochea et al. (1996) suggest that the 100-m isobath may be an approximate reflection of the original shape of the volcano. A north-facing embayment in the steep slope down from the 100-m to the 1000-m isobath (Fig. 6) may mark a landslide similar to that off the SVC. Confirmatory evidence for such landslides must await the sampling of the offshore sediments to a depth at which early Tertiary deposits might be expected.

It is notable that the only bathymetric features which might give evidence of offshore structures are seen west of the island. Around the eastern coast, bathymetric contours closely parallel the shore and a steep and steady descent from the shoreline to 1000 m carries the unbroken Miocene basalt dip slope down to that depth. This regularity is broken only by a narrow eastward projection of shallow sea bed east of Gran Tarajal (Fig. 2) which Ancochea et al. (1996) have interpreted as marking the extension of a Miocene dyke swarm (Fig. 3). The contours offer no evidence for collapse, subsidence or uplift.

4. Fuerteventura and its geotectonic situation

The western Canary Islands are isolated volcanic structures for which solely volcanological and volcano-tectonic processes have been considered. However, Fuerteventura and Lanzarote together with the Concepcion Bank, are part of a well-defined ridge running roughly parallel to the African coastline. Hence, for these structures, consideration has been given to the possibility of constructional and destructional mechanisms resulting from crustal tectonism related to events on the continental margin. It has long been recognised that activity on the South Atlas fault zone was related to the orogenic development of the Atlas mountains, and a number of authors (Dillon and Sougy, 1974; Anguita and Hernan, 1975) have shown a correspondence of events off the Atlantic shore with Atlas orogenic pulses. Direct mechanical connection to continental plate motion on the South Atlas fault belt is not likely, since the innermost magnetic anomalies show no evidence of significant offset (Hayes and Rabinowitz, 1975), but stresses may be transferred into the oceanic plate, and localised by weaknesses inherited from the early-stage ocean opening processes.

4.1. Cretaceous pre-seamount deposition and deformation

There is evidence that the crust beneath the eastern islands was involved in tectonism before active volcanism began. The Fuerteventura Cretaceous sedimentary succession starts with hemi-pelagic sediments and turbidites deposited on a passive continental margin; the deposition being brought to an end by a period of uplift and submarine erosion. The Albian sediments which followed were pelagic marls and chalks, which contrast with the equivalent age organic-rich siltstones and black shales on the nearby continental shelf off the Spanish Sahara, a contrast which suggests strong uplift in the vicinity of Fuerteventura (Robertson and Stillman, 1979a).

There is also clear evidence of a subsequent strong deformation which affected the Albian and Senonian sediments of the Fuerteventura seamount; the more southerly outcrops all having ongoing directions consistent with an inverted limb of a major reclined NE-facing neutral fold. Minor folds are rare, but stereographic projections show the statistical fold axis plunging at a shallow angle to the SSE, and a cleavage-bedding intersection lineation plunging to the south. Such folding would be consistent with dextral movement within a shear zone oriented almost N–S (Robertson and Stillman, 1979a).

Both the uplift and deformation predate the eruption of the bulk of the seamount and indicate that the volcano was built on crust which had recently undergone deformation. It is possible that this deformation may be the result of transferred continental crustal stresses, but it is also possible that it may relate to the arrival of the Canaries plume head.

4.2. Tertiary tectonism in Fuerteventura

4.2.1. Extensional dyke swarm emplacement

Once magmas began to rise, strongly anisotropic stress would have been necessary to produce the extensional conditions under which the sheeted dyke swarms were emplaced (Stillman, 1987). These may well have been caused by crustal stretching above the plume head, with rifts oriented by weaknesses in the crust inherited from the early stage ocean-opening processes. However, the conclusion is not clear-cut, since the orientation of the dyke swarms parallels the eastern Canaries ridge, and thus may relate to whatever mechanism is invoked for the formation of this ridge.

4.2.2. Collapse tectonics

Two later events have also been attributed to tectonism by Ancochea et al. (1996) who suggest that the production of the Central Depression, the exposure of the Basal Complex above sea level, and possibly the disappearance of the western halves of the NVC and CVC, imply important Miocene tectonics.

The first event is recorded in the development of the CVC where they demonstrate a break between the earliest shield activity which produced the CVCI lavas, and a second phase of activity which produced the CVCII lavas. They suggest that the steeper dip and more fractured nature of the CVCI lavas compared to the CVCII lavas indicate that the break represents a tectonic event which would have involved an uplift of a block coincident with the area where the older rocks are exposed, and would be more than 22 Ma old. Another line of evidence supports movement at about that time. Uplift, deep erosion, and the generation of debris flows are clearly indicated by the late Oligocene — early Miocene sediments of Fuerteventura (Robertson and Stillman, 1979b), and a horst block mechanism was proposed by Robertson and Stillman (1979a).

However, by analogy with other Canary Islands (Carracedo, 1996), it now seems more likely that most of not all the uplift during the constructional phase could have been caused by abundant sheet and pluton injections into the seamount; the emplacement ages of which would correspond with the uplift proposed by Ancochea et al. (1996).

The second event took place after the CVCII was formed and before the eruption of the Melindraga and Tamacite Formations, which erupted when the Central Depression had already been created. The event must then have taken place between 20 and 19

Ma, the age of the youngest CVCII rocks, and 17.5 and 16.5, that of the Melindraga Formation. There are also possibly related unconformities or hiatuses in the SVC and NVC at this time. This event must have been the one responsible for the prime demolition of the early Miocene volcanic peaks, and eventually resulted in the production of the Central Depression.

Whether this event was due to failures of the volcano superstructure causing catastrophic collapse with giant landslides, or by tectonic faulting on a regional scale, is still a matter for debate. Earlier authors (e.g., Hausen, 1958) suggested that the depression was produced by faults with a NNE–SSW trend — presumably generating a graben structure. Later workers (e.g., Fuster et al., 1968) found no evidence of faulting and explained the depression in terms of erosion. If Ancochea et al. (1996) are correct in their supposition that the western sector of the CVC volcano has subsided, and forms the prominent bathymetric feature seen off the western coast, then faulting may be a related cause. The fault planes may owe their orientation to the sheeted dyke swarm, but this extensional fissure system in turn perhaps depend on the tectonics of the passive continental margin.

Carracedo (1994), has presented cogent arguments in favour of rift zones generated above magma-induced upwellings playing a key role in both dyke orientation during growth, and mass-wasting and destruction of the Canaries volcanoes. Ancochea et al. (1996) imply that a rift zone model would appear to be relevant to Fuerteventura, as they have suggested that the volcanic centres of Fuerteventura and Lanzarote, and the seamounts off the southern tip of Fuerteventura, are all controlled by such rift zones.

4.3. Is the rifting controlled by volcano-tectonic processes or by tectonism on the Continental crust?

Ancochea et al. (1996) imply that the rifting may not be volcano-tectonic, but rather be the product of regional tectonism, and that the origin of the eastern Canary Islands is linked to the tectonic evolution of the African Continent.

Given the proximity, and the orientation of the inner Canaries alignment, it might seem logical to propose a link between island and continent evolution, in particular with the South Atlas Fault. There are pulses in the lower Miocene and mid to upper Miocene of the region which would be coeval with events in Fuerteventura (Dillon and Sougy, 1974; Anguita and Hernan, 1975).

There are certain attractions to this view, already proposed by some authors (e.g., Anguita and Hernan, 1975; Robertson and Stillman, 1979a). Tectonic processes may then be called into play to explain the uplift which has brought the deeper regions of the seamount succession to a position above sea level. From the alignment and position of the eastern Canaries and the Concepcion Bank, it was tempting to propose a major NE–SW to ENE–WSW fault system running along the chain of volcanoes making up Fuerteventura and Lanzarote, raising horst blocks which exposed the sea bed and the roots of the volcanoes. Also the early uplift and later overturning of the Albian sediments may be related to deformation of the Moroccan continental margin with which is associated the Turonian or Senonian emplacement of major gravity slide sheets seen in seismic records north of Agadir canyon (Robertson and Stillman, 1979a).

Whilst there appear to be some points in favour of this interpretation, the most powerful argument against the influence of Continental tectonic processes remains the total absence of any geophysical evidence for linkages between the islands and between the eastern Canaries ridge and the mainland of Africa. Furthermore, if the elevation of the sea floor owes its origin to such tectonism, only those islands on the extension of the South Atlas fault belt should be affected. There is no evidence of any tectonic link between the eastern Canaries and the other islands of the archipelago; nevertheless, some parts of the submarine seamount stage are also exposed on La Palma and La Gomera. If these seamounts are uplifted as extensively as Fuerteventura, then the tectonic model would seem to present a case of special pleading.

4.4. Conclusion

It would seem most likely, on balance that the combination of magmatic and volcano/tectonic activity, followed by flank collapse due to major landslip, operated in the eastern islands in the same way as the western ones. This has been responsible for the growth and denudation seen on Fuerteventura, though taking place in the Miocene rather than the Quaternary. Fuerteventura's evolution is not significantly different to that of the western Canary Islands; it just happened about 15 million years earlier.

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