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Crustal structure, gravity anomalies and flexure of the lithosphere in the vicinity of the Canary Islands

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SUMMARY

Simple elastic plate models have been used to calculate the flexure of the lithosphere caused by volcanic loading at the Canary Islands and sediment loading at the Moroccan continental margin. By comparing the calculated flexure to observations based on seismic refraction and free-air gravity anomaly data, constraints have been placed on the long-term ($>10^6$ years) elastic thickness of the lithosphere, T_e. The best fit between the calculated and observed flexure in the vicinity of the Canary Islands is for $T_e = 20$ km. This value of T_e also explains seismic reflection data in regions that flank the island *provided* that the lithosphere underlying the Moroccan margin is sufficiently weak ($T_e < 5$ km) for sediment loading to contribute little to the island flexure. A backstripping study, in which gravity data are used to constrain the value of T_e, supports the suggestion that the lithosphere underlying the Moroccan margin, like its conjugate at the Baltimore Canyon Trough, is weak. Although sediment loading at the Moroccan margin appears therefore to have exerted little influence on the structure of the Canary Islands, there is evidence that island flexure may have influenced the stratigraphic development of the Moroccan margin, especially in the region of the 'slope anticline'. A T_e of 20 km is about 15 km *lower* than would be expected on the basis of the thermal age of the oceanic lithosphere that underlies the Canary Islands. Similar low values have been reported from oceanic islands in the Pacific 'superswell' region of French Polynesia where they have been attributed to re-heating of the lithosphere by one or more hotspots. A general E–W age progression of the volcanic rocks suggests that the Canary Islands were also generated by a hotspot. They lack, however, the topographic swell and gravity/geoid high which usually accompanies these features. One possibility is that the low T_e values are the result of a *pre-existing* weakness in the oceanic crust. A T_e of 20 km is large enough, however, for a significant part of the mantle to still be involved in the support of the island loads. A more likely explanation is thermal weakening by a hotspot which has been localized enough to reduce T_e but not to produce a swell or gravity/geoid high. Irrespective of its cause, the low T_e suggests that oceanic lithosphere does not necessarily progressively increase its strength with age, so that even 140 Ma old lithosphere is vulnerable and can, in some regions, be significantly weakened by later thermal and mechanical processes.

Key words: Canary Islands, crustal structure, gravity anomalies, lithosphere.

INTRODUCTION

The Canary Islands are among a number of oceanic volcanic islands that have formed in an intraplate setting away from the influence of active plate boundaries. The islands are Neogene in age (e.g. Staudigel & Schmincke 1984) and are located on Jurassic-age oceanic crust (e.g. Roest *et al.* 1992) in close proximity to the Moroccan continental margin. The apparent E–W age progression of the volcanic rocks has led to the suggestion (e.g. Morgan 1971) that the islands were generated by a hotspot in the underlying mantle.

Hotspots are considered to be the surface manifestations of deep mantle plumes. One of their characteristics is an abundance of volcanic activity (Morgan 1972) which persists for long periods of time. Another is their association with broad topographic swells (Crough 1983) and gravity and geoid anomaly 'highs' (Detrick & Crough 1978; Sleep 1990). Two classes of models have been invoked to explain swell uplift: 're-heating' (Detrick & Crough 1978; Menard & McNutt 1982) and 'dynamic' (Menard 1973; Parsons & Daly 1983; Richards, Hager & Sleep 1988) models. Both model types involve the transfer of heat from an ascending plume to the overlying plate by some form of conduction and convection. In the re-heating model, swells are produced by thermal expansion which is confined to the conducting portion of the lithosphere (the thermal boundary layer), while in the dynamic model there is a contribution to the uplift that is produced by vertical normal stresses applied to the base of the seismically defined lithosphere (the mechanical boundary layer) by convection. Unfortunately, it is difficult to use surface observations to distinguish between these models. Geoid anomaly data (Parsons & Daly 1983), for example, only indicate the depth of the greatest temperature differences—typically 60 km for swells—rather than the actual thickness of the thermal and mechanical boundary layer.

The re-heating model predicts a higher surface heat flow than the dynamic model so it should be possible to distinguish between them using subsidence history and heat-flow data. Some seamounts and oceanic islands have a subsidence history that follows what would be expected for the age of the lithosphere on which they were emplaced. Others, however, show a *greater* subsidence. Eniwetok and Bikini atolls in the central Pacific Ocean, for example, formed during the Eocene but, their subsidence history is more typical of 25 Ma old oceanic lithosphere than the 90 Ma old lithosphere on which they were emplaced (Detrick & Crough 1978). These observations lend support therefore to the re-heating model. Heat-flow data, however, are more difficult to account for. At Hawaii, the heat flow over the swell is not significantly different to that of the surrounding sea-floor (Von Herzen *et al.* 1989). Similarly, heat flow seems to be normal in both the present-day (i.e. French Polynesia) and former (i.e. Darwin Rise) 'superswell' regions of the central and west Pacific Ocean (Stein & Abbott 1991).

Oceanic flexure studies (e.g. Watts 1978) suggest that the long-term ($>10^6$ years) elastic thickness of the oceanic lithosphere, T_e, is a strong function of its thermal structure. According to the re-heating model, the geothermal gradient would be 're-set' by a hotspot (e.g. McNutt & Menard 1978) to that of younger lithosphere. If the plate is subsequently loaded it should therefore have a *lower* T_e than expected. The most commonly sited examples of this phenomena are seamounts and oceanic islands in the 'superswell' region of French Polynesia (McNutt & Fisher 1987) which are associated (Calmant 1987; Calmant & Cazenave 1986) with unusually low values of elastic thickness. The association of low T_e with hotspot swells appears to breakdown, however, away from the French Polynesia region. At Hawaii, T_e is similar (or even a little higher) to what would be expected for the thermal age of the underlying lithosphere (Watts & ten Brink 1989). This, together with the normal heat flow (Von Herzen *et al.* 1979), has led some to suggest (e.g. Watson & McKenzie 1991) that the Hawaiian swell may be dynamically supported.

Recent studies in French Polynesia (Filmer, McNutt & Wolfe 1993; Goodwillie & Watts 1993) raise some doubts about the influence that regional effects such as re-heating and mantle dynamics may have on T_e. These studies, which are based on better bathymetric and geoid data sets than were available to Calmant & Cazenave (1986), show that not all values of T_e in the Pacific superswell region are unusually low. Some values (e.g. Gambier, Marquesas) previously thought to be low, are in fact normal for the age of the lithosphere on which they were emplaced. More importantly, a normal T_e seamount (e.g. Gambier) can occur on the swell in close proximity to a low one (e.g. Pitcairn). These results suggest a high degree of 'individuality' between seamounts, and local rather than regional controls on T_e.

One explanation for a normal T_e at Hawaii is that the absolute motion of the Pacific plate has been fast enough for there not to have been sufficient time for the mechanical boundary layer (which includes the elastic thickness) to have been heated. Sleep (1990) has calculated, for example, that at least 10 Ma are required for the heating to produce a heat-flow anomaly at the surface if only conduction is important. It this is correct, the heat-flow anomaly at Hawaii, together with its effect on T_e, should have been 'swept' downstream of the present-day hotspot. A consequence of this is that at slow-moving plates, such as Africa, enough time has elapsed for extensive heating to have occurred locally. Indeed, a number of the seamounts and oceanic islands that formed on the African plate (e.g. Reunion, Great Meteor and Cape Verdes) appear to be associated with lower than expected values of T_e (e.g. Watts, Cochran & Selzer 1975; Menard & McNutt 1978; Courtney & White 1986).

The Canary Islands, which are also located on the African plate, may provide additional constraints on the relative influence on T_e of thermal re-heating and dynamic effects. Previous studies of these islands, however, give conflicting results. For example, Filmer & McNutt (1989), using geoid and 'SYNBAPS' data (Van Wyckhouse 1973), show that T_e is higher than would be predicted on the basis of plate age, while Canales (1993) and Danobeitia *et al.* (1994) using shipboard gravity and topography data suggest that it may be significantly lower. On the basis of the high T_e, Filmer & McNutt (1989) speculated that there was no hotspot beneath the Canary Islands. Canales (1993) and Danobeitia *et al.* (1994), however, have proposed that there *is* a hotspot. One problem with both these studies is that they are based on admittance techniques so that the T_e deduced is only an *average* over a broad region. Here we use a a 'forward modelling' approach in order to better constrain T_e at the Canary Islands and the Moroccan margin. As in previous studies, the paper is based on gravity and geoid data, but differs in that consideration is also given to the flexure associated with the islands and the margin and its effects on the subsidence and uplift history of the region. The main objectives of the paper are to better understand the long-term mechanical properties of old (i.e. >100 Ma) oceanic lithosphere, the uplift and subsidence history of the Canary Islands and Moroccan margin, and the thermal and mechanical effects that mantle plumes might have in perturbing the physical properties of slow-moving plates.

GEOLOGICAL AND GEOPHYSICAL SETTING

The Canary Islands are made up of volcanic lavas which erupted in a number of well-defined cycles. The oldest of the subaerial eruptions (>35 Ma) occur on Fuerteventura while the youngest are found on La Palma (1–2 Ma). Tenerife, which has the highest volcano in the group (Mt Tiede: elevation = 3.4 km), is generally considered to be of intermediate age (about 16 Ma), although there is no evidence that any rocks older than 6 Ma occur on the island (Schmincke, H.-U., private communication). Basal complexes that represent the earlier submarine phase of eruption have been mapped on Fuerteventura, La Gomera and La Palma (e.g. Hoernle & Schmincke 1983) and yield ages that range in age from 60–80 Ma on Fuerteventura to 3–4 Ma on La Palma.

Morgan (1971) was one of the first to suggest that the Canary Islands are a hotspot trace which formed as the African plate moved slowly over the mantle. The morphology of the islands, however, is complex and it is difficult to explain both the general E–W trend of the westernmost islands (i.e. Gran Canaria to La Palma) and the general NNE–SSW trend of the easternmost islands (i.e. Fuerteventura and Lanzarote) by a single hotspot trace such as predicted by current plate motion models (e.g. DeMets *et al.* 1990). Holik & Rabinowitz (1992) proposed, however, that a strong reflector in seismic reflection profile data beneath the Moroccan continental rise corresponds to the top of a thick volcanoclastic deposit and therefore that the hotspot trace extends from El Homera, through Tenerife and across the continental rise west of Fuerteventura and Lanzarote (Fig. 1) to the region of the prominent reflector.

It is known from volcanological studies that all the islands in the 'chain', except La Gomera, have been active during the past 5000 years. The hotspot model has therefore been questioned particularly by Anguita & Hernán (1975) who suggest a propagating fracture model instead. These authors point out that the NNE–SSW trend of the easternmost islands is similar to the Trans-Agdir Fault in Morocco while the westernmost islands have a similar E–W trend to central Atlantic Ocean fracture zones. A fault model may also help to explain the large amounts of uplift that have been observed in Fuerteventura and Lanzarote (>1.5 km) because continental strike-slip fault and oceanic fracture zone systems are known to be associated with large vertical movements. These observations by themselves do not,

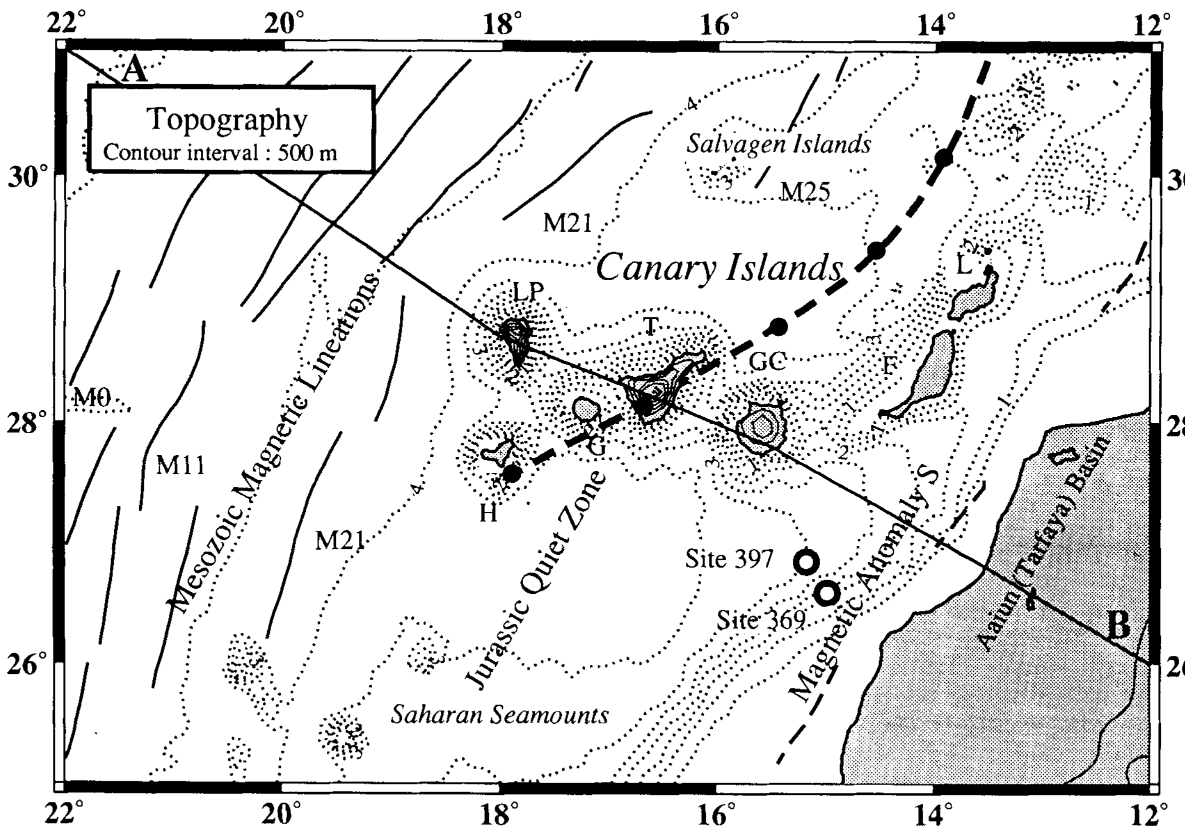

Figure 1. Location map of the Canary Islands region. Bathymetric contours (dotted lines) are shown at 500 m intervals. Thin solid lines show Mesozoic magnetic lineations M0 through M25 based on Hayes & Rabinowitz (1975) and Roest *et al.* (1992). Thin dashed line shows magnetic anomaly S. Thick dashed line shows the trace of the Canary Island hotspot according to Holik & Rabinowitz (1992). The solid circles show the predicted ages along the trace (in 10 Ma intervals) from zero age at a location just to the south of El Hierro (H) to 50 Ma to the east of the Salvagen Islands. The open circles show DSDP sites 369 (Lancelot *et al.* 1978) and 397 (von Rad *et al.* 1979). T = Tenerife, LP = La Palma, GC = Gran Canaria, F = Fuerteventura, L = Lanzarote and G = La Gomera. AB = Schematic geological cross-section shown in Fig. 2.

however, discriminate between the propagating fault and hotspot model. There is no evidence, for example, in the deep structure of the Moroccan margin for an offshore extension of the Trans-Agadir Fault. Furthermore, the hotspot model is also associated with uplift (e.g. Sleep 1990) because of thermal and dynamic effects associated with the underlying plume.

The Canary Islands formed on oceanic crust of Jurassic Magnetic Quiet Zone (JMQZ) age (e.g. Hayes & Rabinowitz 1975; Roest *et al.,* 1992). The boundary that separates the JMQZ from the Mesozoic magnetic lineations intersects the Canary Islands chain near La Gomera. La Palma and El Hierro seem to have developed on or close to M21 to M25, suggesting that these islands formed on crust between 150 and 156 Ma in age. Tenerife and Gran Canaria are both located within the JMQZ and so the age of the sea-floor underlying them cannot be determined. However, extrapolation of spreading rates between M0 and M21 suggests that the oceanic crust may be as old as 180 Ma beneath Tenerife and 190 Ma beneath Gran Canaria. The eastern boundary of the JMQZ is defined by a large-amplitude magnetic anomaly (S) which follows the shelf break and is considered by some as the boundary between oceanic and continental crust. If this is correct then it implies that Fuerteventura and Lanzarote formed on oceanic crust while sediments of the Moroccan margin formed on thinned (stretched?) continental crust.

There is no evidence from magnetic anomaly data for any significant offset in the JMQZ either side of the Canary Islands (Hayes & Rabinowitz 1975; Roest *et al.* 1992) 'lineament'. However, the existing magnetic anomaly data are not sufficient to determine whether any of the 'small offset' fracture zones mapped by Klitgord, Hutchinson & Schouten 1988 in the western Atlantic are present in the JMQZ of the Canary Islands region or not.

The oceanic setting of the Canary Islands is in accord with the available seismic refraction data (e.g. Bosshard & MacFarlane 1970; Banda *et al.* 1981). These studies show that the upper crust is characterized by *P*-wave velocities in the range of 2.8–4.8 km s^{-1}, which have been interpreted by Banda *et al.* (1981) as sedimentary and volcanic 'cover rocks'. Large variations in velocity exist in the underlying main crustal layer. For example, velocities increase from about 6.4 to 7.7 km s^{-1} beneath Tenerife. The high velocities resemble those of the Hawaiian ridge (e.g. Watts & ten Brink 1989), where they have been interpreted as an intrusive sill complex in the lower part of the oceanic crust. Below the main crustal layer, velocities increase to about 8.0 km s^{-1}. The estimated depth to Moho beneath Tenerife is about 15 km. Beneath Gran Canaria and Fuerteventura, however, there is evidence of a single velocity layer in the range of 6.1–6.3 km s^{-1} and 6.5–6.7 km s^{-1} respectively, which are more typical of 'layer 2' and 'layer 3' in the oceanic crust. Mantle velocities of 8.0 km s^{-1} were only encountered beneath Gran Canaria, but the 7.4 km s^{-1} velocity beneath Fuerteventura—originally interpreted by Banda *et al.* (1981) as mantle—could equally well represent oceanic crust which has been intruded by mantle-derived material (White 1993). The estimated depth to Moho beneath Gran Canaria based on these velocities is 17 km but beneath Fuerteventura it is indeterminate.

The Canary Islands are located in close proximity to the Moroccan continental margin (Fig. 1). Seismic reflection profile and commercial well data (e.g. Arthur *et al.* 1979) in the Aaiun (Tarfaya) Basin region suggest that the margin is underlain by substantial thicknesses (>10 km) of Mesozoic–Tertiary sediments. The earliest of these sediments (Triassic to early Jurassic?) consist of rapidly deposited terrigenous and evaporitic 'syn-rift' sediments which formed in narrow half-graben structures, along the margin. Overlying these sediments is a thick 'post-rift' sequence of late Jurassic to Tertiary sediments. The early Cretaceous consists of shallow-water carbonates which were deposited on a gently subsiding shelf (e.g. Arthur *et al.* 1979). The early Cretaceous is separated from overlying early Miocene units in Deep Sea Drilling Project (DSDP) sites 397 (Fig. 1; von Rad, Ryan *et al.* 1979) and 369 (Fig. 1; Lancelot, Seibold *et al.* 1978) by a major unconformity. The unconformity has been identified in seismic reflection profiles of the continental slope and rise where it has been interpreted (Arthur *et al.* 1979) as the result of corrosive bottom counter-currents which developed following the initiation of the Antarctica glaciation during the late Eocene/early Miocene. Elsewhere on the slope and shelf the early Cretaceous is overlain by a late Cretaceous deep-water sequence. The overlying Neogene comprises mainly of oozes, but includes several prominent ash layers which reflect the gradual emergence of the Canary Islands above sea-level during the early Miocene. The thickness of the late Cretaceous and Tertiary sediments exceeds 2 km, suggesting that a significant portion of the total sediment accumulation formed more than 80 Ma after continental breakup.

FLEXURE OF THE LITHOSPHERE

The Canary Islands are of importance to flexure studies since they—along with the Cape Verdes, the northern end of the Louisville Ridge, and some of the western Pacific seamounts—have the potential to provide information on the long-term mechanical properties of the oldest parts of the oceanic lithosphere. The main problem is that since T_e depends on the age of the lithosphere *at the time of loading,* volcanic loads need to be sufficiently long in duration (i.e. >1 Ma) for the stress relaxation effects to be minimized (Bodine, Steckler & Watts 1981) but short enough for the flexure to be representative of a single 'point' load, rather than the cumulative effects of a number of different loads.

Unfortunately, the Canary Islands have had a long loading history and this may complicate their use in flexure studies. On Fuerteventura, for example, the subaerial lavas are 0–25 Ma old (i.e. Miocene and younger) but the underlying submarine lavas may be as old as 60–80 Ma (i.e. late Cretaceous). If the age of the underlying oceanic lithosphere is 190 Ma, as implied by the extrapolation of oceanic spreading rates then these ages imply that the subaerial lavas loaded 165–190 Ma lithosphere while the submarine lavas loaded 110–130 Ma old lithosphere. The corresponding difference in T_e, however, is in the range 30–37 km, assuming a 450 °C controlling isotherm. Therefore, it is unlikely that T_e at Fuerteventura will vary by much more than a few kilometres because of its 'long' loading history. La Palma formed on still younger oceanic lithosphere (150–156 Ma) and so we might expect larger changes in T_e. However, the duration of eruption cycles on

La Palma is smaller than on Fuerteventura, ranging from 1–3 Ma (e.g. Hoernle & Schmincke 1993) for the subaerial to as much as 3–4 Ma for the submarine lavas. La Palma was therefore emplaced on the oceanic lithosphere *relatively rapidly,* making it an ideal load for the estimation of T_e. Less is known about the age of the basal complexes on Tenerife and Gran Canaria but the available evidence suggests that they are intermediate in age between La Plama and Fuerteventura.

One of the first accounts of flexure in the vicinity of the Canary Islands was by Filmer & McNutt (1989). These workers used a 3-D elastic plate model to compute the geoid effect of the load and flexure for different values of T_e. They then subtracted the calculated geoid from the satellite-derived geoid to obtain the so-called 'isostatic geoid anomaly'. Filmer & McNutt (1989) found that the value of T_e which gave the least correlation between the isostatic geoid anomaly and the topography was about 48 km, suggesting that this value most satisfactorily accounted for the observed geoid anomaly due to the islands and their compensation. A T_e of 48 km is significantly higher than would be expected by the depth to the 450 °C controlling isotherm which suggests a value closer to 35 km. Filmer & McNutt (1989) based their study, however, on the 5 × 5 min 'SYNBAPS' (Van Wyckhouse 1973) topography data set. This data set has been shown (e.g. Smith 1993) to be smoother and have less total variance than the Earth's actual topography. This may explain why the depth anomaly maps of Filmer & McNutt (1989; Fig. 2) show the summits of La Plama, Tenerife and Gran Canaria as only rising to 5000 m. 4000 m and 3550 m respectively *above* the mean depth of the sea-floor. As Fig. 2 shows, Tenerife rises some 6900 m above the mean depth of the sea-floor. Under-estimating the topography by such amounts reduces the positive load effect and so a smaller negative flexure effect (which is provided by a higher T_e) is needed in order to produce the same geoid anomaly. In fact, an island signature is seen in each of the isostatic anomaly maps of Filmer & McNutt 1989)—even for the one that corresponds to their 'best-fitting' case of $T_e = 48$ km.

The only other studies are the recent one by Canales (1993) and Danobeitia *et al.* (submitted) who estimated T_e using a gravitational admittance technique. To avoid problems with SYNBAPS these workers used a topography data base based on individual 'point' measurements. They also used shipboard gravity data since these data have a

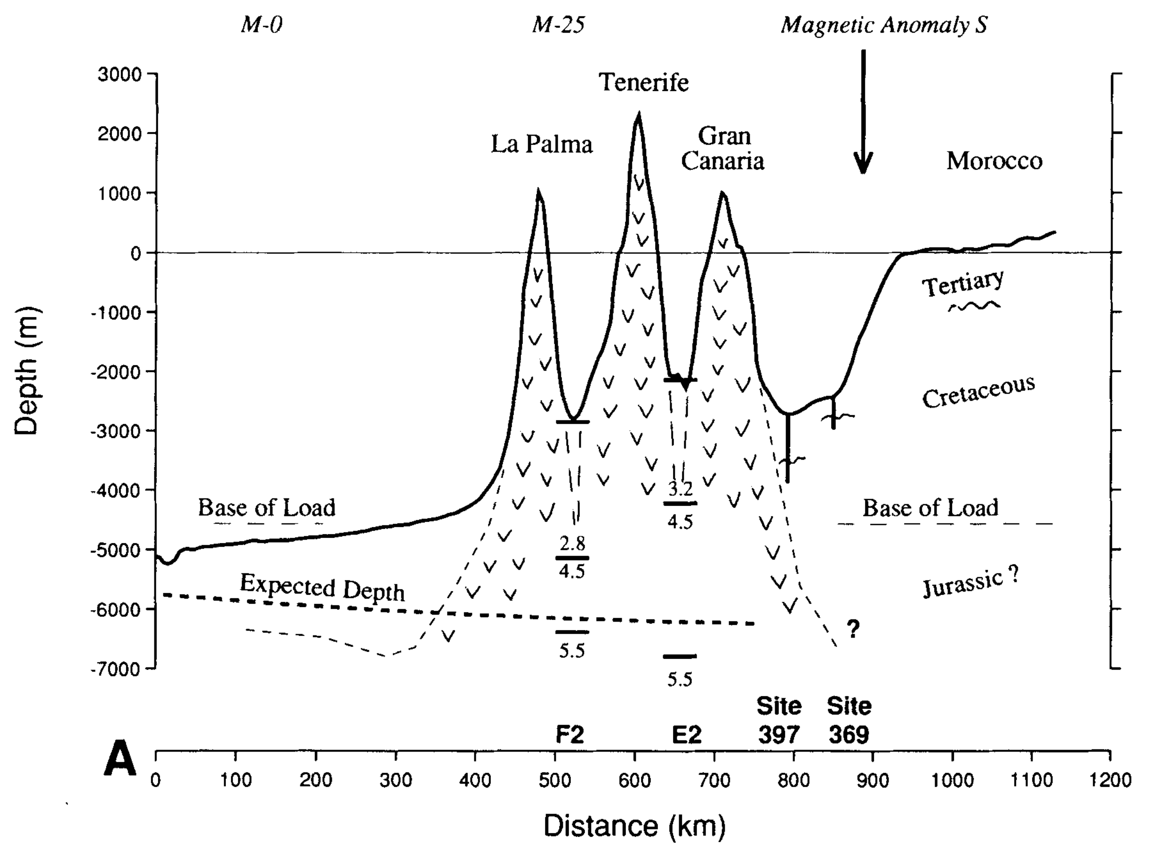

Figure 2. Schematic geological cross-section along profile AB (Fig. 1). The profile extends from the Moroccan continental margin, along the crest of the Canary Islands to the Madeira abyssal plain. The topography profile was constructed by sampling the 5 × 5 minute grid used to construct Fig. 1. The 'vvv' stippling shows the estimated extent ot the Canary Islands volcanic load. The age of sequences beneath the Moroccan margin is based on Arthur *et al.* (1979) and the thickness of sediments west of La Palma is based on Mezcua *et al.* (1991). The principal results from DSDP sites 397 and 369 have been projected onto the profile. Seismic refraction data at stations E2 and F2 are based on Bosshard & MacFarlane (1970). The heavy dashed line is the expected depth that sea-floor should be in the absence of sediment and volcanic loading and the thin dashed line is the depth (4.5 km) that was used in the flexure calculations to define the base of the surface loads that act on the lithosphere (depths shallower than the base are defined as downward-acting loads, depths deeper are upward loads).

better signal-to-noise ratio than the geoid at the short wavelengths that are of interest in flexure studies. Canales (1993) and Danobeitia *et al.* (submitted) showed that the 'best-fit' T_e to the theoretical admittance for a broad region which included the western islands (La Palma, Tenerife and Gran Canaria) and part of the Moroccan margin was in the range 15–20 km. These values, as they point out, are significantly less than those deduced by Filmer & McNutt (1989).

This study also uses a topography data set that is based on individual 'point' measurements. On land, a data set that was compiled by the University of Leeds as part of its African Gravity Project (AGP) was used. This data set includes all the available commercial, national survey and academic data. At sea, a data set obtained from the National Geophysical Data Centre (NGDC) in Boulder was used. This data set included more than 77 individual ship tracks. The land and sea point data were smoothed together over 2.5 × 2.5 min. 'squares' and then gridded at 5 × 5 min. using a minimum curvature technique (Wessel & Smith 1991). Of the total number of 2.5 × 2.5 minutes blocks available in the Canary Islands region, more than 65 per cent had one or more topography measurement indicating that the overall data coverage was good.

From the viewpoint of flexure, the topography of a volcanic load can be considered in two parts: a 'surface' load caused by the displacement of water and air by relatively dense volcanic rock, and an 'infill' load of volcanic and other material which displaces the water-filled flexure. Both loads combine to cause the lithosphere to flex downwards beneath the surface load and upwards in flanking regions. The surface load is defined here as all the topography (including topography above sea-level) *above* a mean water depth of 4500 m. This depth—referred to in Fig. 2 as the 'base of the load'—ensures that all the loads due to the displacement of water *and* air by volcanic material are included. It excludes, however, the loading effects of any 'buried' loads between 4500 m and 6000 m depth—the approximate depth the sea-floor should be on the basis of its age. Fig. 2 shows, however, that these loads have mainly displaced *pre-existing* Mesozoic sediments of the Moroccan continental slope and rise. The density contrast between the volcanic load and the sediments is small (when compared to that of the surface loads) and so their omission is unlikely to significantly affect the flexure. Finally, the infill load was assumed to fill the flexure up to a horizontal base depth of 4500 m. This assumption appears reasonable, especially since the Canary Islands, unlike Hawaii, are not associated with an unfilled moat.

As was pointed out earlier, the Canary Islands are located in close proximity to the Moroccan continental margin. Geological studies suggest the margin consists of a thickness of at least 10 km of Mesozoic–Tertiary sediments. The margin sediments, like the island volcanic rocks, act as a load on the lithosphere which should respond by flexure. In order to calculate the sediment contribution a similar base depth of 4500 m, as was used to define the volcanic loads, was assumed. This depth, which is reasonable in view of the tectonic subsidence that is likely to have occurred at this margin, implies that sediments of the Moroccan shelf, as well as the slope and upper rise, are included as surface loads. Unfortunately, it was not possible in the calculations to vary either the density of the sediment load or the T_e *at the same time* as varying the density and T_e of the volcanic load.

Figure 3 shows the flexure due to the topographic loads of the Canary Islands and the Moroccan margin for values of the elastic plate thickness, T_e, of 5, 20 and 35 km. Other parameters assumed in the calculations are summarized in the Appendix. The most striking feature of Fig. 3 is the NE–SW trending flexural depression associated with the Moroccan continental slope and shelf. The amplitude of flexure is seen to be largely independent of T_e, but its wavelength is controlled by T_e: the shortest wavelengths occurring for $T_e = 5$ km and the largest for $T_e = 35$ km. The flexural depression associated with the load of the Canary Islands is superimposed on the margin flexure (Fig. 3). The amplitude of the flexure depends, in this case, strongly on T_e. For example, the $T_e = 35$ km case, which corresponds to the expected value for the age of the underlying oceanic lithosphere, shows a small contribution (<1 km) whereas the $T_e = 5$ km case, which corresponds to a weaker than expected plate, shows a larger contribution (>10 km).

CRUSTAL STRUCTURE

Seismic reflection and refraction data have the potential to directly measure the flexure in the region of an oceanic island or seamount load. By comparing these data to predictions based on simple elastic plate models it should be possible to estimate T_e with a high degree of confidence. Unfortunately, with the possible exception of Hawaii, there have not been sufficient seismic data acquired in the region of oceanic islands and seamounts to test the flexure models and therefore it has proved difficult in practice to use these data as constraints.

Seismic refraction data are limited to the Canary Islands (Banda *et al.* 1981) themselves and so little information exists on the crustal and upper mantle structure beneath the flanking flexural moats and bulges. These data cannot therefore be used to constrain the flexure calculations in Fig. 3 directly since the amount of crustal thickening that has occurred between the oceanic crust of the JMQZ, for example, and the islands is not known. The crustal and upper mantle structure is probably best defined beneath Tenerife where two intersecting profiles (Fig. 4; Lines A and B of Banda *et al.* 1981) reveal the velocity structure of the main crustal layer and the depth to Moho. The Moho depth can be compared directly to the predictions of the flexure model *provided* some assumptions are made about the depth to the top of the crust and the thickness of the crust in the undeformed regions that flank the islands.

Figure 4 compares the predicted crustal structure based on the flexure model to the seismically constrained structure along profile CD (Fig. 3). This profile was chosen because it is subparallel to the local trend of the margin and therefore shows the maximum extent of flexure beneath the islands. The predicted crustal structure shows the flexure of both the top of the oceanic curst and the Moho. The depth to the top of the oceanic crust was obtained by adding 4.5 km (i.e. the depth to the base of the load) to the flexure while the depth to Moho was calculated by adding 7.5 km to the top of the oceanic crust since this is the approximate thickness of undeformed 140–145 Ma old ocean crust (e.g. White,

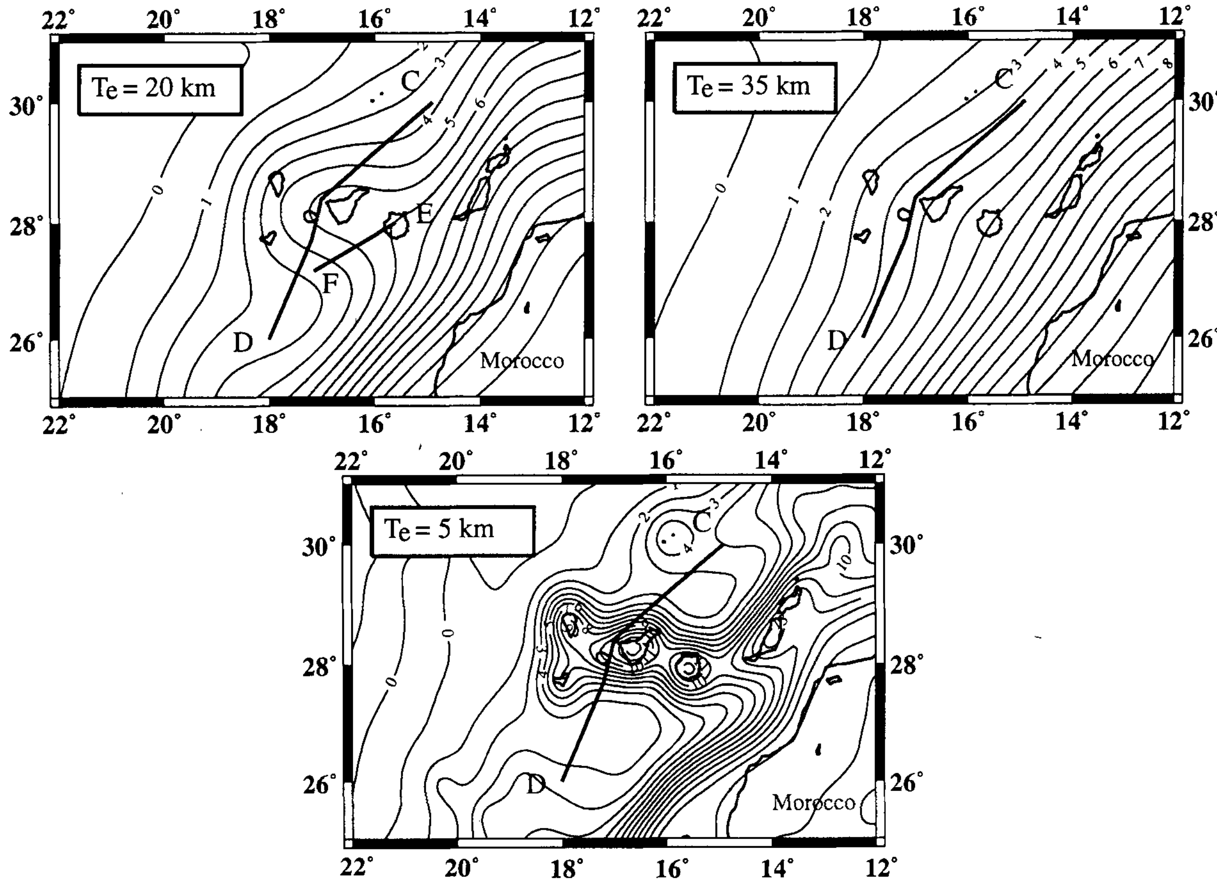

Figure 3. Flexure of the lithosphere due to the volcanic loads of the Canary Islands, the Salvagen Islands and the Saharan seamounts and the sediment loads of the Moroccan continental margin. The flexure shows the deformation of an arbitrary surface which prior to loading was horizontal. There is a strong dependence between the amount of flexure predicted in the vicinity of the Canary Islands and the value of T_e assumed: $T_e = 5$ km produces a large-amplitude flexure (>10 km) whereas $T_e = 35$ km produces little or no flexure. The NE–SW trend, which dominates each of the flexure solutions, is caused by sediment loading of the Moroccan margin. CD shows the location of the crustal cross-sections and free-air gravity anomaly profiles presented in Figs 4, 5 and 8.

McKenzie & O'Nions 1992). There is a good agreement between the predicted and seismically constrained Moho depth (Banda *et al.* 1981) beneath Tenerife for $T_e = 35$ km but poor agreement for $T_e = 5$ km. The $T_e = 35$ km case predicts, however, that the *total* thickness of the sediment and volcanic material that infills the flexure in the regions flanking the islands is about 4.5 km. This thickness is much larger, however, than the 1–2 km reported by Mezcua *et al.* (1991) on the GECO Line B west of La Palma. The main factor that controls the island infill thickness is sediment loading at the Moroccan margin. The margin influence is largest for $T_e = 35$ km because the strength of the lithosphere spreads the response to the sediment loads over a broad region and is smallest for $T_e = 5$ km since the lithosphere is too weak for the response to extend much beyond the margin itself. The predicted infill thickness for $T_e = 5$ km is in better agreement with the observed data than the $T_e = 35$ km case. We can reconcile the seismic refraction and reflection data therefore *if* it is assumed that the margin sediments loaded a weak lithosphere ($T_e < 5$ km) while the volcanic islands were emplaced on more rigid lithosphere (Fig. 5). The actual value of T_e required for the volcanic islands cannot be determined precisely. However, the calculations suggest that $T_e = 35$ km predicts too shallow a Moho while $T_e = 5$ km is too deep, thereby favouring an intermediate value of $T_e = 20$ km.

FREE-AIR GRAVITY ANOMALY

The free-air gravity anomaly at a volcanic island is sensitive to both the load and the amount of flexure. By comparing the amplitude and wavelength of the calculated gravity effect of flexure models to the observed gravity anomaly it is possible to use these data to better constrain the value of T_e.

The gravity anomaly data have been compiled from a number of sources. In land areas the data are based on a 2.5×2.5 min. smoothed free-air gravity anomaly data set that was compiled as part of the Leeds-based AGP. The data in sea areas are based on NGDC as well as other publicily available data. A total of 44 individual ship tracks were used in the compilation. Fig. 6 shows the distribution of the 2.5×2.5 min. smoothed 'squares' for both land and sea areas. These data were then interpolated onto a 5×5 min. grid using the same minimum curvature technique as was used for the topography.

The calculated gravity anomalies were obtained by combining the positive gravity effect of the topographic loads with the negative effect of the flexure. A range of

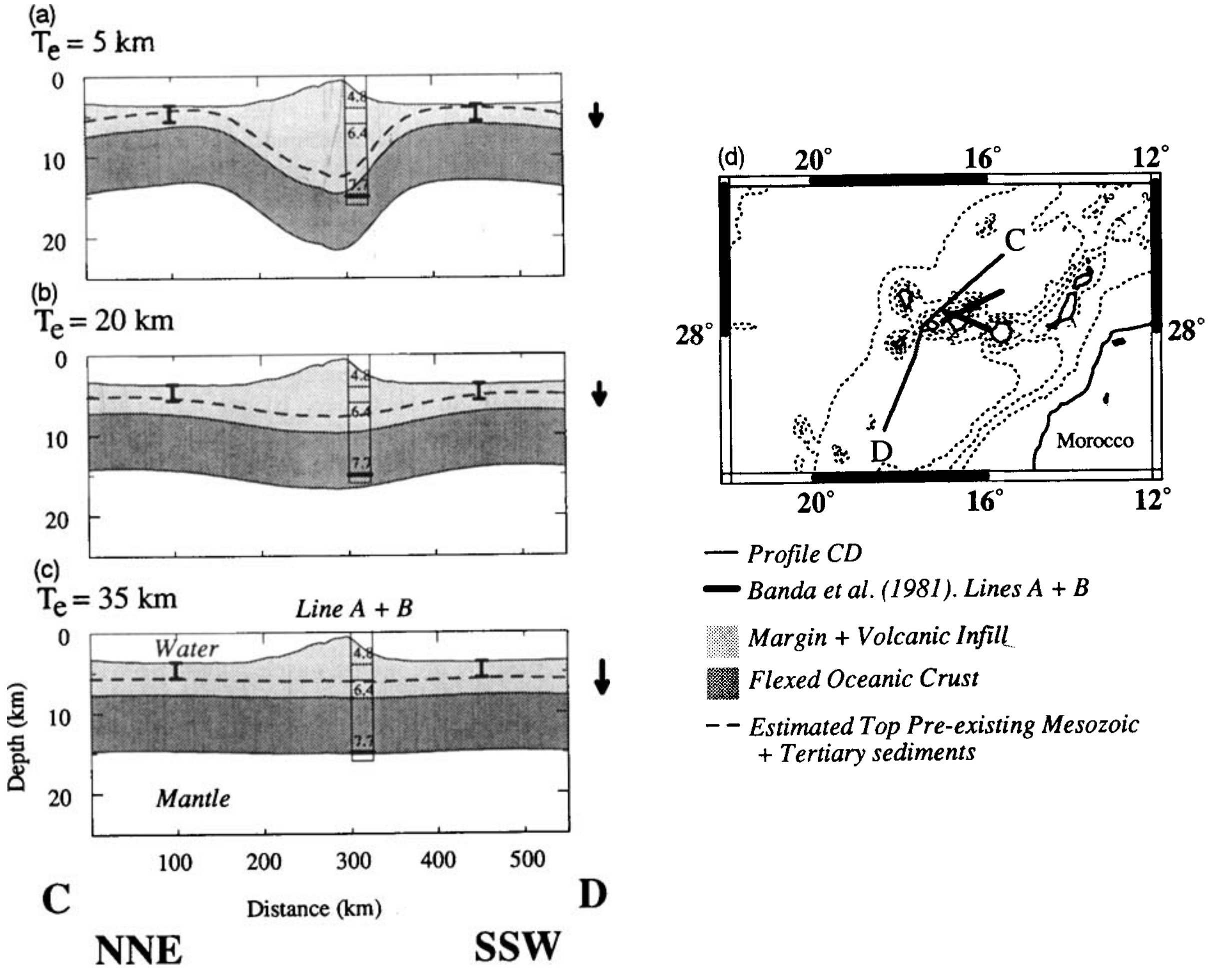

Figure 4. Comparison of the calculated flexure along Profile CD (Fig. 3) to the observed depth to Moho beneath Tenerife (heavy horizontal bar) according to Banda *et al.* (1981) and the thickness of sediments in regions flanking the islands (heavy vertical bar) based on Mezcua *et al.* (1991). The depth to the top of the oceanic crust was calculated by adding the flexure in Fig. 3 to the depth of the base of the load (4.5 km). The depth to the base of the crust (i.e. Moho) was then calculated by adding an additional 7.5 km (the average thickness of 145 Ma old oceanic lithosphere; White, McKenzie & O'Nions 1992) to the top of the flexed crust. The dark shaded region shows the extent of the flexed crust, while the light shaded region is the material which infills the flexure. The infill consists of two parts: a Neogene infill associated with the volcanic loads and a Mesozoic infill due to the margin sediment loads (heavy dashed line shows a 2 km stratigraphic 'marker' in the infill). The margin loads contribute to the overall thickness of the infill (shown by arrows) while the island loads determine the thickness variations that are seen along the profile. (a) $T_e = 5$ km, (b) $T_e = 20$ km, (c) $T_e = 35$ km and (d) location map showing profile CD and the locations of lines A + B.

densities for the load and infill were used. In calculating the effect of the flexure it was assumed that the mean unflexed thickness of the oceanic crust in the Canary Islands region was 8 km and that the density contrasts were concentrated at the top and bottom (i.e. the Moho) of the flexed oceanic crust.

The calculated and observed gravity anomalies were compared along profile CD which crosses the Canary Islands between Tenerife and Gomera (Fig. 3). The profile was chosen to be subparallel to the local trend of the margin so as to limit the contribution of the margin sediment loads to a broad, long-wavelength flexure. The gravity anomaly associated with the loading of the islands and their compensation will therefore be enhanced along profile CD and contributions to it from say the Moroccan margin free-air gravity 'edge effect' anomaly minimized. A disadvantage of profile CD is that it is not located along an actual ship track and so gravity anomaly data have to be interpolated on to it. However, as Fig. 6 shows, the data coverage along the profile is reasonably good, except in the northern part of the profile where there is a significant data gap (shaded region, Fig. 6).

Figure 7 compares the observed free-air gravity anomaly along profile CD to the calculated anomaly based on the flexure model with T_e in the range 5–35 km. The observed profile (heavy line) shows that the Canary Islands are associated with a large-amplitude gravity 'high' of about 200 mGal which is flanked by gravity 'lows' of about 40 mGal. The calculated profile which best explains both the amplitude and wavelength of the observed profile is based on $T_e = 20$ km. Higher values of T_e produce larger anomaly amplitudes over the islands than are observed, while smaller T_e values produce amplitudes that are too small. The flanking lows, however, appear to indicate a lower T_e. This is well seen on the southern part of the profile where the shape of the flanking low follows the $T_e = 15$ km profile quite well.

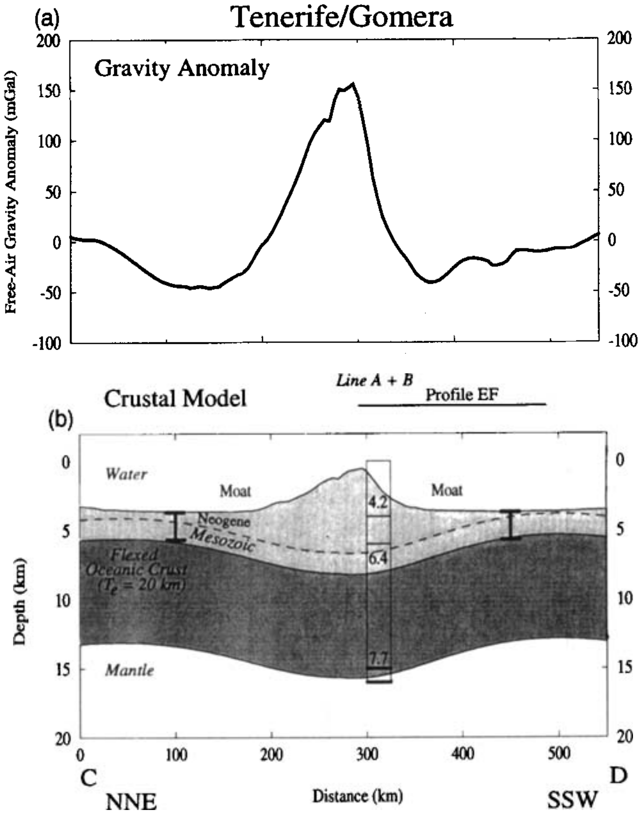

Figure 5. Comparison of the calculated crustal structure to seismic refraction and reflection data in the vicinity of the Canary Islands. The crustal structure is based on the flexure profile with $T_e = 20$ km. This value, however, overestimates the thickness of the infill in flanking regions which the available seismic reflection profile data indicate is about 2 km. The flexure profile has therefore been adjusted to reflect this infill thickness and in so doing it is implied that T_e of the Moroccan margin is about 5 km or less. The light dashed line shows the approximate position of the base of the infill, assuming that the 2 km thick infill in undeformed regions is made up of 0.5 km Neogene and 1.5 km Mesozoic. (a) Free-air gravity anomaly along profile CD, (b) 'best-fit' crustal model.

Unfortunately, the northernmost part of the observed profile is complicated because of a data gap (Fig. 7) and so it is not known whether the $T_e = 15$ km case fits this part of the profile as well in this region.

The root mean square (rms) difference between the observed and calculated gravity anomaly along profile CD is shown in Fig. 8 for a range of likely values of the load and infill densities from 2500 to 3000 kg m^{-3} and T_e from 0 to 40 km. The minimum rms value (A, in Fig. 8) is for $T_e = 20$ km and an infill and load density of 2800 kg m^{-3}. However, the trend of the minimum is such that the observed gravity could equally well be explained by either a lower T_e with a higher load and infill density or a higher T_e with a lower infill and load density. Although a number of parameter combinations could therefore explain the gravity anomaly data, limits can be imposed on T_e on the basis of the plausibility of the densities that are required. For example, Fig. 8 shows that the observed data cannot be explained by the T_e deduced by Filmer & McNutt (1989) *unless* load and infill densities of about 2200 kg m^{-3} are assumed. Although the load density on the Canary Islands is not actually known, deep drill hole data in Bermuda (Hyndman, Christensen & Drury 1979) suggest values of 2670 and 2800 kg m^{-3} for subaerial and submarine basaltic lavas respectively. Moreover, seismic refraction measurements (Watts *et al.* 1985) in the Hawaiian region suggest that moats that are infilled largely by material derived from the loads (e.g. debris flows and ash layers) could have densities as high as 2400 kg m^{-3}. A load and infill density of 2200 kg m^{-3} is therefore too low and so T_e must be less than was proposed by Filmer & McNutt (1989).

The calculated gravity anomaly based on the best-fit parameters (infill and load density of 2800 kg m^{-3} and $T_e = 20$ km) is compared to the observed free-air gravity anomaly in the Canary Islands region in Fig. 9. There is good agreement between the calculated and observed data along the crest of the island chain and over the flanking moats. This is well seen (Fig. 9c) in the residual anomaly (i.e. observed—calculated gravity anomaly) which shows that—with the exception of La Palma, where there is a paucity of land measurements (Fig. 6)—the gravity effect of the islands and its compensation has been almost entirely removed from the observed field. The main features remaining are a long-wavelength NE–SW trending positive anomaly to the west of the Canaries which probably reflects the free-air gravity anomaly 'high' that is associated with the slow-spreading mid-Atlantic ridge (e.g. Cochran & Talwani 1978), and a NE–SW trending belt of positive and negative anomalies to the east of Gran Canaria which probably indicates that the T_e deduced for the westernmost islands may not be applicable to either Fuerteventura and Lanzarote or the Moroccan Margin.

DISCUSSION

The comparison of seismic and free-air gravity anomaly data with calculations based on simple elastic models suggests that the best-fit elastic thickness of the lithosphere, T_e, beneath the Canary Islands is about 20 km. This estimate was derived from an analysis of a single topography and gravity anomaly profile between Gomera and Tenerife but is probably representative of all the islands between Gran Canaria and La Palma. The available age data suggest that the submarine basaltic lavas of Tenerife formed on oceanic lithosphere that was approximately 140–150 Ma old at the time of loading (this assumes an age of the underlying oceanic lithosphere of 180 Ma and the age of the submarine lavas is 30–40 Ma). As Fig. 10 shows, oceanic lithosphere of this age should have an elastic thickness of 35 km, assuming a 450 °C controlling isotherm. The best-fit T_e deduced in this paper is therefore some 15 km *lower* than would be expected on the basis of the thermal age of the lithosphere at the time of loading—a result that has implications for the thermal and mechanical evolution of the Canary Islands region.

The influence of the Canary Islands on the subsidence/uplift history of nearby oceanic islands/seamounts and the Moroccan continental margin

The T_e deduced at the Canary Islands implies that the flexural depression or moat extends outward from the flank of the islands to a distance of about 180 km (e.g. Fig. 5).

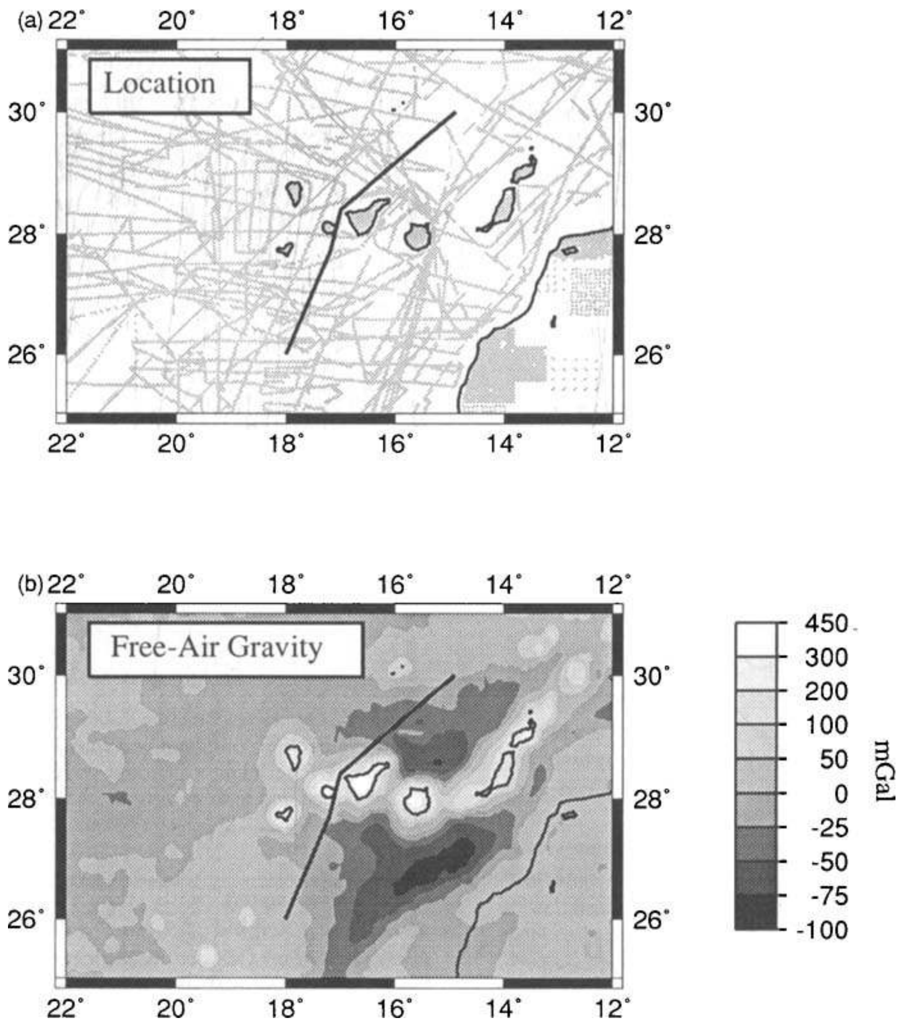

Figure 6. Free-air gravity anomaly data in the region of the Canary Islands. (a) Location of the 2.5 × 2.5 min. values used to construct the free-air gravity anomaly map in (b). (b) Free-air gravity anomaly map which has been shaded at 50 mGal intervals.

Beyond the depression there would be a bulge or arch which reaches its maximum amplitude (about 4 per cent of the maximum flexural depression) about 210 km from the islands. Although the T_e associated with the Canary Islands is smaller than expected on the basis of the thermal age of the lithosphere, it is large enough to have influenced the subsidence and uplift history of any other pre-existing geological features in the region.

Topography maps (e.g. Fig. 11) show that the Salvagen Islands and the Saharan seamounts are both presently located in the vicinity of the Canary Islands flexural arch. Furthermore, the predicted trace of the flexural 'node' appears to follow certain bathymetric contours (e.g. 3400 and 3700 m) in the vicinity of these features. Unfortunately, there are no reliable age dates reported from either the Salvagen Islands or the Saharan seamounts so it is not known whether these features pre-date the main shield-building phase in the Canary Islands or not. The main island in the Salvagens (Ilha Selvagem Grande), however, is basaltic and has a flat surface that is presently about 95 m above mean sea-level. This uplift is therefore in general agreement with the predicted uplift in the Salvagen Islands region, which for $T_e = 20$ km is about 110–140 m. Although one of the Saharan seamounts (Echo Bank) has a flat surface, they are now submerged and so their location may be unrelated to the bulge.

One of the more interesting consequences of Canary Islands flexure is the influence that it may have had on the stratigraphic development of the nearby Moroccan margin. The flexural node in Fig. 11 was estimated from the location of the near-zero free-air gravity anomaly contour and so is deflected to the north and south as the Moroccan margin is approached because of the influence of the margin free-air edge-effect anomaly. This is not to imply, however, that the islands would have little flexural effect on the margin. To the contrary, the location of the topographic loads of Gran Canaria, Fuerteventura and Lanzarote suggest that the flexural node should strike E–W *across* the local trend of the Moroccan margin and intersect the shelf break about 210 km from the centre of Gran Canaria. From there it would be expected to cross the coast in the vicinity of Cape Bojador and then follow a more NE–SW course, passing some 20 km inland of Cape Juby.

Figure 11 shows that the predicted trace of the Canary

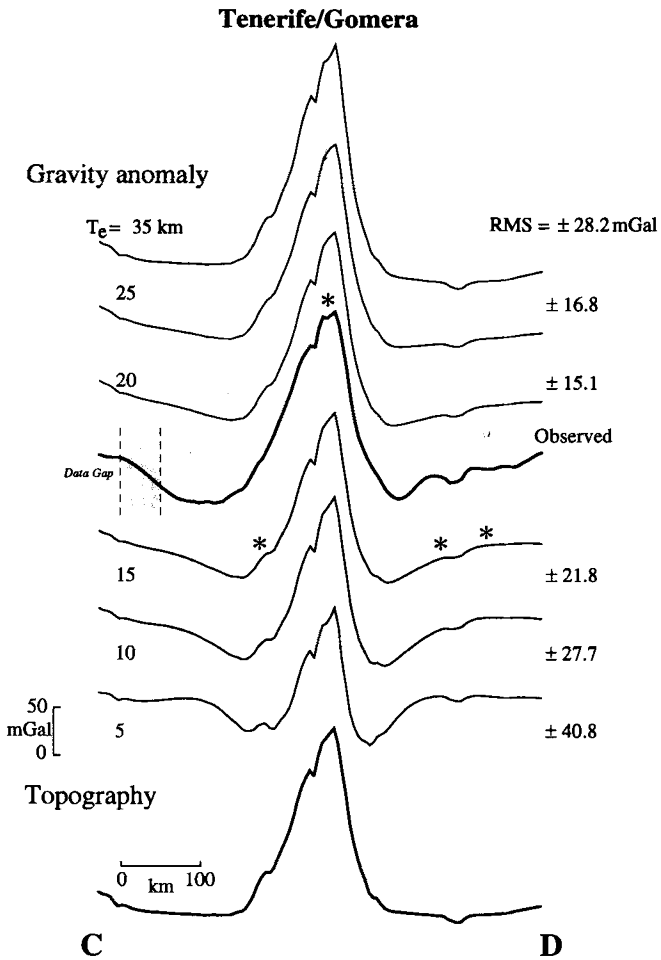

Figure 7. Comparison of the free-air gravity anomaly and calculated anomalies along profile CD (Fig. 3). The calculated anomalies are shown for T_e in the range 5–35 km, a mean thickness of the oceanic crust of 7.5 km, an assumed load and infill density of 2800 kg m^{-3} and a mantle density of 3330 kg m^{-3}. The numbers to the right of each calculated profile show the root mean square (rms) difference between observed and calculated anomalies. The asterisks indicate portions of the calculated profiles that are particularly well explained by the observed data. The shaded area is a data gap (Fig. 6a) along the northern part of the profile.

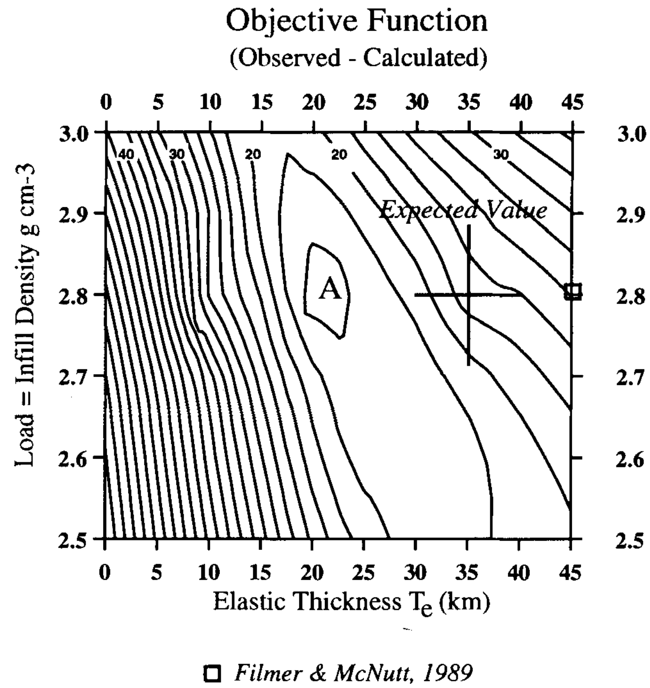

Figure 8. Objective function obtained by calculating the rms difference between observed and calculated gravity anomalies along profile CD. The minima (A) and therefore the best-fit parameters are for a load and infill density of 2800 kg m^{-3} and a T_e of 20 km.

Islands flexural node follows closely a pronounced anticlinal structure in seismic reflection profile data which Hinz *et al.* (1974) have referred to as the 'slope anticline'. Age data from DSDP site 397 suggest that the anticline involves early Cretaceous and older (Jurassic?) sediments. Moreover, the slope anticline has a similar trend to an erosional unconformity separating early Cretaceous and Neogene strata and it has been suggested that both features were caused by the same mechanism. The erosion is believed from biostratigraphic data to have occurred during a relatively rapid period of time, perhaps during the late Eocene to Oliogocene. This corresponds in age to some of the basal complexes on the islands and therefore the main submarine shield-building phase. Seibold & Hinz (1974), when discussing the unconformity, suggested that it was caused by uplift of the continental margin and subaerial or shallow-water erosion in response to to 'a pulse of tectonic activity associated with the Alpine orogeny'. As an alternative, Seibold & Hinz (1974) also suggested the possibility of an Oligocene/Miocene oceanic 'event' due to a lowering of sea-level, the initiation of Antarctica glaciation, or to changes in palaeobathymetry (e.g. the opening of the Norwegian–Greenland Sea) and its consequences on global ocean circulation patterns.

Recently, McGinnis *et al.* (1993) argued that a series of erosional events associated with a eustatic sea-level fall during the Oligocene may have modified the stratigraphy along a 1000 km long segment of the West African margin offshore Gabon, Congo and Angola. They suggest that the erosion caused the margin to flexurally rebound along much of its length and that flexure has influenced the stratigraphic 'architecture' of the margin, especially the pattern of coastal onlap and offlap that develops. The observation that offshore Morocco the slope anticline trends *across* the margin (rather than parallel to it) suggests, however, that while flexure may still be an important control, it is more likely to have been caused by local volcanic loading effects rather than by unloading of the margin due to a regional erosional event.

Figure 11 shows that the slope anticline crosses the shelf a few tens of kilometres to the north of the projected position of the Canary Island flexure node. The node is constrained mainly, however, by flexure around the westernmost islands of La Palma, El Hierro and Tenerife and so this observation might indicate that the easternmost islands of Fuerteventura and Lanzarote formed on lithosphere with a slightly lower T_e than the westernmost islands, perhaps because of their proximity to stretched continental crust. Tectonics in the form of flexure is therefore the preferred mechanism to

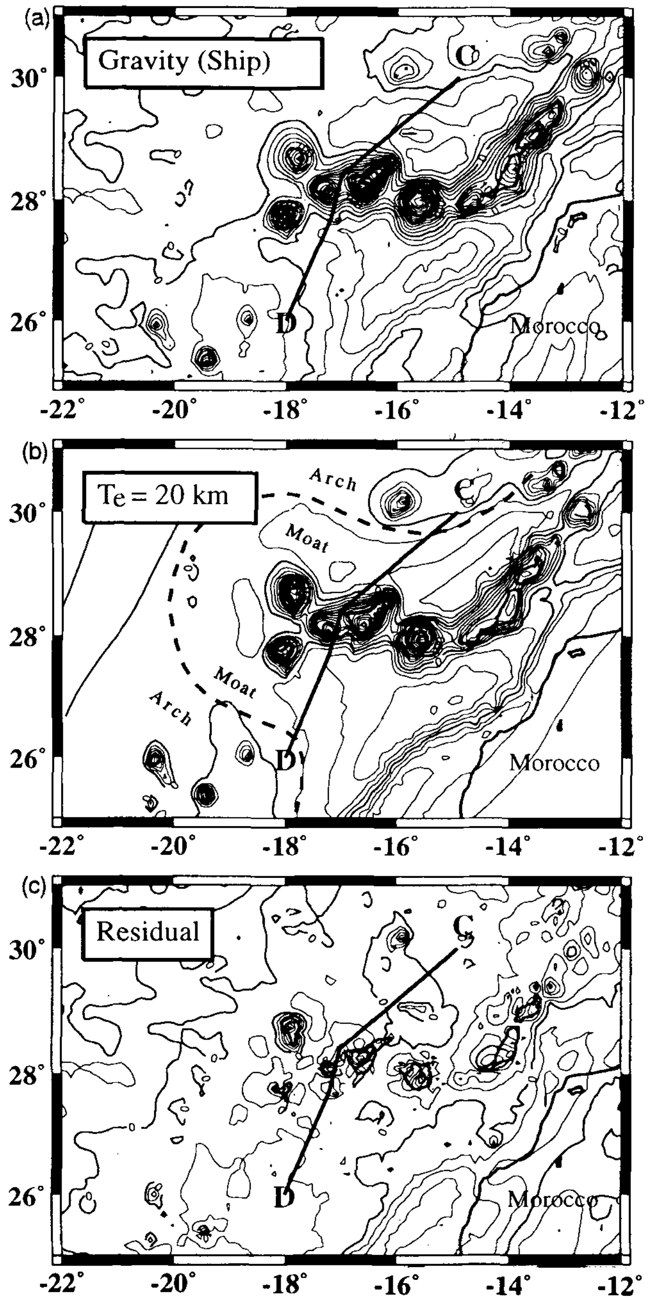

Figure 9. Comparison of observed and calculated gravity anomalies in the Canary Islands region. The contour interval is 25 mGal. The observed anomalies are based on Fig. 6 and the calculated anomalies are based on the the best-fit parameters in Fig. 8. (a) Free-air gravity anomaly. (b) Calculated gravity anomaly based on a load and infill density of 2800 kg m^{-3} and a T_e of 20 km. The heavy dashed line shows the approximate extent of the flexural node which separates the region of uplift (arch) and flexure (moat). (c) Residual anomaly obtained by subtracting the calculated gravity anomaly from the observed.

explain the slope anticline and, perhaps, the erosional unconformity. We cannot rule out, however, the effect of other local processes such as corrosive bottom counter-currents in accounting for the large-scale backcutting of the margin. Arthur *et al.* (1979), for example, have argued that there is no evidence for shallow-water fossils or sediments at site 369 as required by a tectonic model, and have proposed instead that bottom currents associated with a rapid deterioration of the climate during the Eocene and Oligocene is the best explanation for the feature. Indeed, the Saharan slide (e.g. Embley 1976) is documentation that large-scale mass wasting at the Cape Bojador margin is occurring at the present time and there is no reason to believe that such processes have not occurred in the past.

The long-term mechanical properties of stretched continental lithosphere

The relatively small thickness of the volcanic (or island-derived) and sediment (or margin-derived) infill in the regions that flank the Canary Islands suggest (e.g. Fig. 5) that the flexural effects of the Moroccan margin are limited spatially and that sediment loading there has occurred on lithosphere with a relatively small T_e. Although there are now a number of studies that support the suggestion that stretched continental lithosphere in rifted continental margins is weak (e.g. Watts 1988; Fowler & McKenzie, 1989), it has not yet been demonstrated that the lithosphere that underlies the Moroccan margin in the region of the Canary Islands is also weak.

The free-air gravity anomaly 'edge effect' at continental margins contains useful information (e.g. Watts 1988) on the long-term mechanical properties of the lithosphere. In general, margins associated with a large-amplitude long-wavelength edge effect are associated with high T_e while margins with a small-amplitude, short-wavelength edge effect indicate low T_e. Fig. 6 compares the gravity anomaly edge effect observed at the Moroccan margin to calculated profiles based on values of $T_e = 5$, 20 and 35 km. The observed gravity has been constructed along a profile (AB in Fig. 6) of a region where the depth to basement is reasonably well known from commercial well and seismic reflection profile data (e.g. Arthur *et al.* 1979). The calculated anomaly has been constructed assuming that the edge effect is made up of two main components: a 'rifting' anomaly and a 'sedimentation' anomaly. In the absence of seismic refraction data, the rifting anomaly was estimated by first backstripping the sediments along the profile and then calculating the gravity effect of the restored crustal structure. The sedimentation anomaly was then calculated by reloading the sediments on the restored crustal structure. The gravity anomaly edge effect is obtained by adding the rifting and sedimentation anomaly. Fig. 12 shows that there are discrepencies between the observed and predicted gravity anomalies. In the region of the edge-effect 'low' the discrepency is attributed to the influence of the Canary Islands flexural moat while that over the eastern part of the Tarfaya Basin is due to the paucity of observations in these regions (e.g. Fig. 6). The most important features of the calculated gravity, however, that are indicative of flexure are the amplitude and wavelength of the edge-effect 'high'. This feature is less likely to be influenced by the islands because profile AB intersects the shelf break in slope in the region of the flexural node where the gravity effect of the volcanic loads and their compensation approach zero. The best fit to the amplitude and wavelength of the edge effect high is for $T_e = 5$ km (Fig. 6) although it is possible that a lower T_e could explain the observations equally well. A

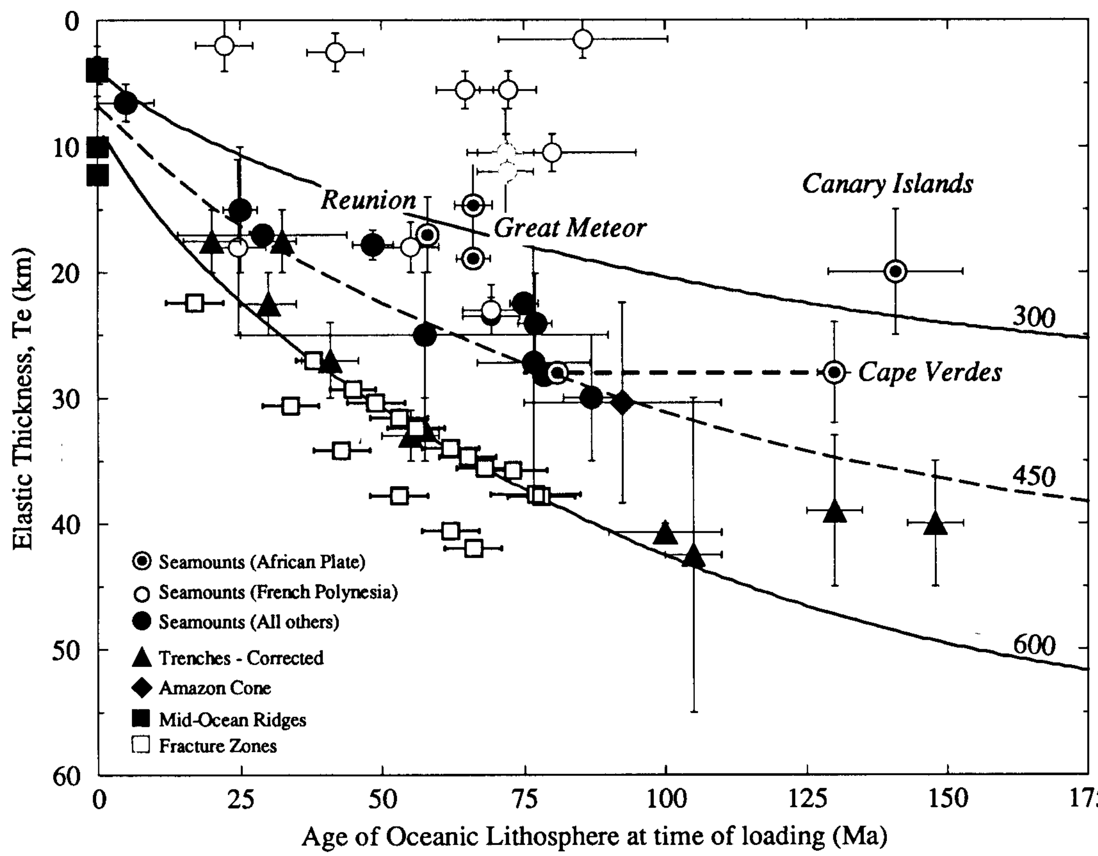

Figure 10. Plot of T_e against the age of the oceanic lithosphere at the time of loading. The main difference between this plot and a recent one by Watts (1992) is that (i) the results of Calmant & Cazenave (1986) in French Polynesia have been replaced by those of a more recent study by Goodwillie & Watts (1993) and Filmer *et al.* (1994) who used better altimetric and bathymetric data than was available to Calmant & Cazenave (1986) and (ii) the value deduced by Filmer & McNutt (1989) for the Canary Islands has been replaced by the value deduced in this paper. (Note: a 5 km and 5 Ma error has been assumed on the T_e and age respectively for the values deduced in this paper). The solid lines show the 300, 450 and 600 °C oceanic isotherms based on a cooling plant model. The shaded thick dashed line illustrates the possible re-setting of the lithosphere that has occurred at the Cape Verdes hotspot.

higher T_e, however, has to be ruled out because these values predict too large an amplitude for the edge-effect high.

The results in Fig. 12 indicate that the Moroccan margin has a low T_e and therefore can be considered, along with its conjugate margin at the Baltimore Canyon Trough (Watts 1988), as a mechanically weak margin. This result has important implications in estimating T_e at the Canary Islands but, it contributes little to our understanding of why stretched continental lithosphere might be weak and has remained so for long periods of time. The close association of the Canary Islands with the margin, however, may help to constrain the problem. The islands themselves are associated with a low T_e and it has been suggested by several workers that they were generated at a hotspot. One possibility therefore is that the low T_e at *both* the Canary Islands and the Moroccan margin is due to the influence of a mantle hotspot which because of African plate motions has spent a sufficient time under the region to allow heat to be conducted from the asthenosphere to the lithosphere thereby weakening it.

The problem that remains, of course, is in explaining the low T_e that has been reported from other margins such as the Baltimore Canyon Trough where there is little evidence of hotspot activity. There is evidence, however, in the form of high P-wave velocities in the lower part of the thinned continental crust and seaward-dipping reflector sequences (e.g. Holbrook & Kelemen 1993) that some magmatism has occurred at this margin. The problem is that it is difficult to use seismic data to distinguish between the thickness of the pre-existing crust and any magmatic material that has been added to it. Backstripping and gravity modelling studies (e.g. Watts 1988) suggest that only small amounts of magmatic material have been added to the margin. While we cannot therefore rule out that temperatures may have been elevated during rifting, it seems likely that other factors, such as the rheology of the continental lithosphere, must also play some role in controlling the low T_e observed at this margin.

The elastic thickness, T_e, and hotspot swells

The low T_e deduced at the Canary Islands is similar to the values that were reported by Calmant & Cazenave (1986) from seamounts and oceanic islands in the French Polynesia region of the central Pacific Ocean. This region is the site of a broad topographic swell, termed the Pacific 'superswell' by NcNutt & Fischer (1987), and it has been suggested (Calmant

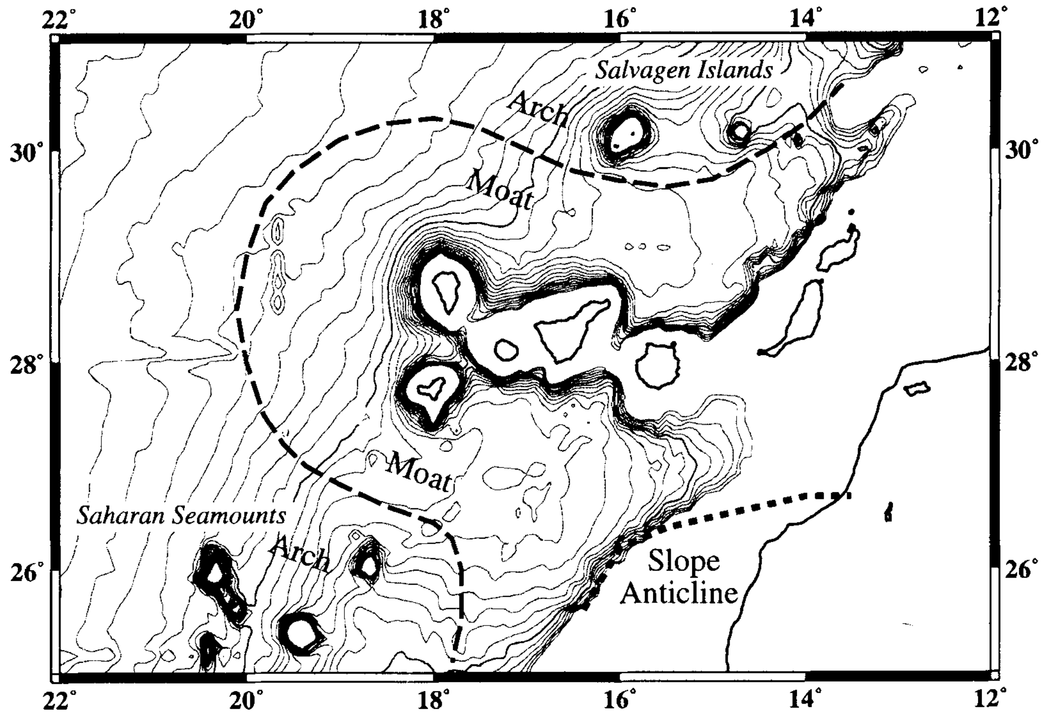

Figure 11. Comparison of the flexural node inferred from the calculated gravity anomaly based on $T_e = 20$ km (Fig. 9b) to bathymetry data in the Canary Islands region. Dashed line (long dashes) = position of the flexural node that separates the arch and moat regions. Dashed line (short dashes) = slope anticline (Arthur *et al.* 1979). Solid lines = bathymetric contours at 100 m intervals. For clarity only contours greater than depths of 2.5 km are shown.

& Cazenave 1986) that the low elastic thickness values there reflect the re-setting of the thermal age of the lithosphere by the same hot ascending plume which caused the superswell. This is apparently supported (e.g. NcNutt & Fischer 1987) by seismic studies which suggest anomalous low *P*-wave velocities at depth beneath the superswell region.

A difficulty with applying the re-heating model to the Canary Islands is that there is no evidence for a topographic swell in the region. Monnereau & Cazenave (1990) suggested that the islands were superimposed on a 1500 km wide and 1500 m high swell but these authors appear to have included the Moroccan continental rise as part of the swell. Fig. 13 shows that when topographic profiles are examined, there is no evidence for any regional decrease in depth as the Canary Islands are approached beyond that which would be expected for a normal continental rise. Moreover, the Canary Islands appear to lack the long-wavelength geoid anomaly high that characterizes mid-plate swells such as Hawaii, Bermuda and the Cape Verdes. This is well seen in Fig. 14 which shows the residual geoid anomaly (i.e. observed geoid minus the calculated geoid based on $T_e = 20$ km) in the Canary Islands region. Although there is a long-wavelength residual geoid anomaly high in the study area (Fig. 14), it is centred over the Saharan seamounts rather than the Canary Islands.

The problem remains, therefore, in explaining the low T_e values. One possibility is that the oceanic crust was already weakened *prior to* volcanic loading. Commercially acquired seismic reflection profiles (Mezcua *et al.* 1991; Banda *et al.* 1992) show that the lower oceanic crust to the west of the Canary Islands is characterized by planar reflectors which dip generally eastward, away from the mid-ocean ridge. These reflectors were interpreted by Banda *et al.* (1992) as extensional faults which formed at or near the ridge crest. If the faults extended to beneath the Canary Islands region then they would have been in the crust prior to volcanic loading.

One difficulty with the mechanical weakening hypothesis is that although T_e is low, it is large enough for a significant part of the mantle to still be involved in supporting the loads of the Canary Islands. Unlike its continental counterpart, the oceanic crust is thin, contains no 'weak zones' and plays little role in the support of surface loads (Bodine *et al.* 1981). The main support is provided instead by the strong underlying mantle which oceanic flexure studies (e.g. Fig. 10) suggest is as much as two to three times the thickness of the oceanic crust. It is difficult therefore to see how a faulted lower oceanic crust would significantly affect the strength of the >150 Ma old oceanic lithosphere that underlies the Canary Islands.

A more likely explanation is that *despite the lack of a swell* the lithosphere beneath the Canary Islands has been weakened by the thermal effects of an underlying plume. The main evidence for the association of low T_e values with a swell has come from studies in French Polynesia. Recent studies (Goodwillie & Watts 1993; Filmer *et al.* 1993) suggest, however, that normal values of T_e occur, alongside the low values, in French Polynesia (Fig. 10, open circles). Furthermore, Hawaii, which geochemically is one of the strongest plumes, has a T_e that is normal for its age. Thus, low T_e is not necessarily associated with a swell so swells may not always be associated with a low T_e.

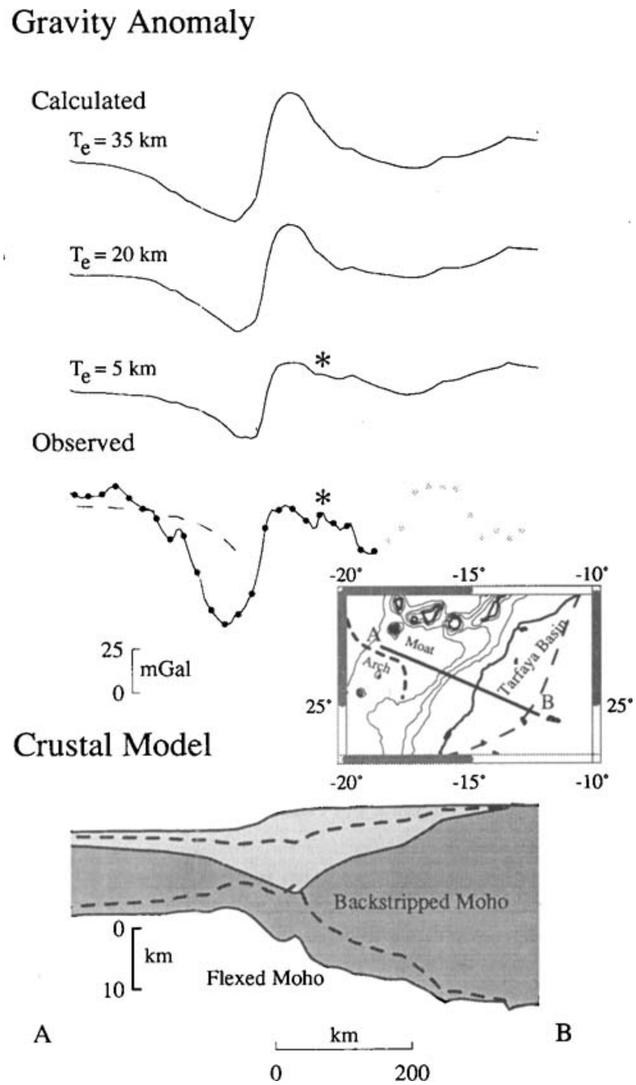

Figure 12. Comparison of the observed free-air gravity anomaly 'edge effect' at the Moroccan margin in the region of Cape Bojador to calculated anomalies based on $T_e = 5$, 20 and 35 km. The calculated gravity anomalies are based on Watts (1988). In this method the crust is first 'backstripped' and the 'rifting anomaly' is calculated by assuming that the restored crust is in isostatic equilibrium. The 'sedimentation anomaly' is then calculated by adding the positive effect of the sediments to the negative effect of the displacement of the crust and mantle by the sediments. The 'edge effect' anomaly is the sum of the two effects assuming that rifting and sediment loading are the only two processes that are operative during the development of the margin.

One explanation for the normal T_e values at Hawaii and French Polynesia is that the Pacific plate has been moving sufficiently fast with respect to the underlying mantle that there has not been enough time for heat to be conducted from the asthenosphere to the upper, elastic part of the lithosphere. Support for this view comes from heat-flow (Von Herzen *et al.* 1989), subsidence (Detrick & Crough 1978) and surface-wave (Woods *et al.* 1991) studies which suggest that the thermal structure beneath Hawaii is similar to what would be expected for the age of the lithosphere on which the islands formed. Thus, much of the temperature perturbation caused by the plume is stored in the lower part of the lithosphere and will be carried away 'downstream' (Von Herzen *et al* 1982) as the plate moves over the Hawaiian hotspot.

A consequence of these observations is that at slow-moving plates there should have been sufficient time for the lithosphere to be heated and weakened. The low T_e at the Canary Islands may therefore be the consequence of loading the African plate which is known to have moved at much smaller rates (about 20 mm yr^{-1}) in the hotspot frame than the Pacific plate. This is supported by flexure studies at other seamounts and oceanic islands on the African plate. These studies show, for example, that Cape Verdes, Great Meteor and Reunion are associated with T_e values some 10 km lower than expected for the age of the lithosphere on which they were emplaced (Fig. 10).

If the low values for seamounts and oceanic islands on the African plate are the result of re-heating of the lithosphere, then they can be used to estimate the re-set age and hence the height that the swell should be. The low T_e values at Great Meteor, Reunion and Cape Verdes imply, for example, that the thermal age of the lithosphere that underlies these features of 65, 60 and 125 Ma has been re-set to 25, 25 and 75 Ma respectively (Fig. 10). The height of the swell is obtained by subtracting the depth expected for the re-set age of the lithosphee from the depth expected for its thermal age. For example, if the 60 Ma old lithosphere at Reunion has been re-set to 25 Ma—as suggested by the T_e result in Fig. 10—then the Parsons & Sclater (1977) cooling plate model predicts that sea-floor depths should be 4250 m (rather than the 5169 m which would be expected for the thermal age) and the swell height would be 919 m. The observed swell height at Reunion (Monnereau & Cazenave 1990) is 1100 m, which is in reasonable agreement with the predicted height based on the re-heating model. A similar agreement is found at Great Meteor where the re-set age suggests a sea-floor depth and swell height of 4250 m and 1013 m respectively. The Cape Verdes is probably the best documented of the three swells on the African plate as heat flow as well as T_e results are available. The re-set age based on T_e is 80 Ma (Fig. 10) which implies a significantly smaller swell height (528 m) than is observed (2100 m; Monnereau & Cazenave 1990). This re-set age is in accord with heat-flow studies (Courtney & White 1986) which shows a 'high' of about 20 mW m^{-2} centred over the swell. This high implies a re-set age of about 60 Ma and a swell height of 820 m. Sleep (1992), who considered both the swell height and heat-flow estimates, estimated that somewhere between 700 and 1200 m of the total swell uplift of 2100 m could have been caused by sublithospheric effects. Thus, re-heating can only contribute a small amount to the Cape Verdes swell height.

There is still the problem though that despite a low T_e there is no swell associated with the Canary Islands hotspot. Swells, however, may result from *either* thermal expansion which is confined to the conducting portion of the lithosphere (thermal boundary layer) *or* by vertical normal stresses exerted to the base of the seismically defined lithosphere (mechanical boundary layer) by convection. The T_e data at Great Meteor and Reunion are supportive of the re-heating model. The data for Cape Verdes, however, which is better constrained by T_e and heat-flow data, suggest

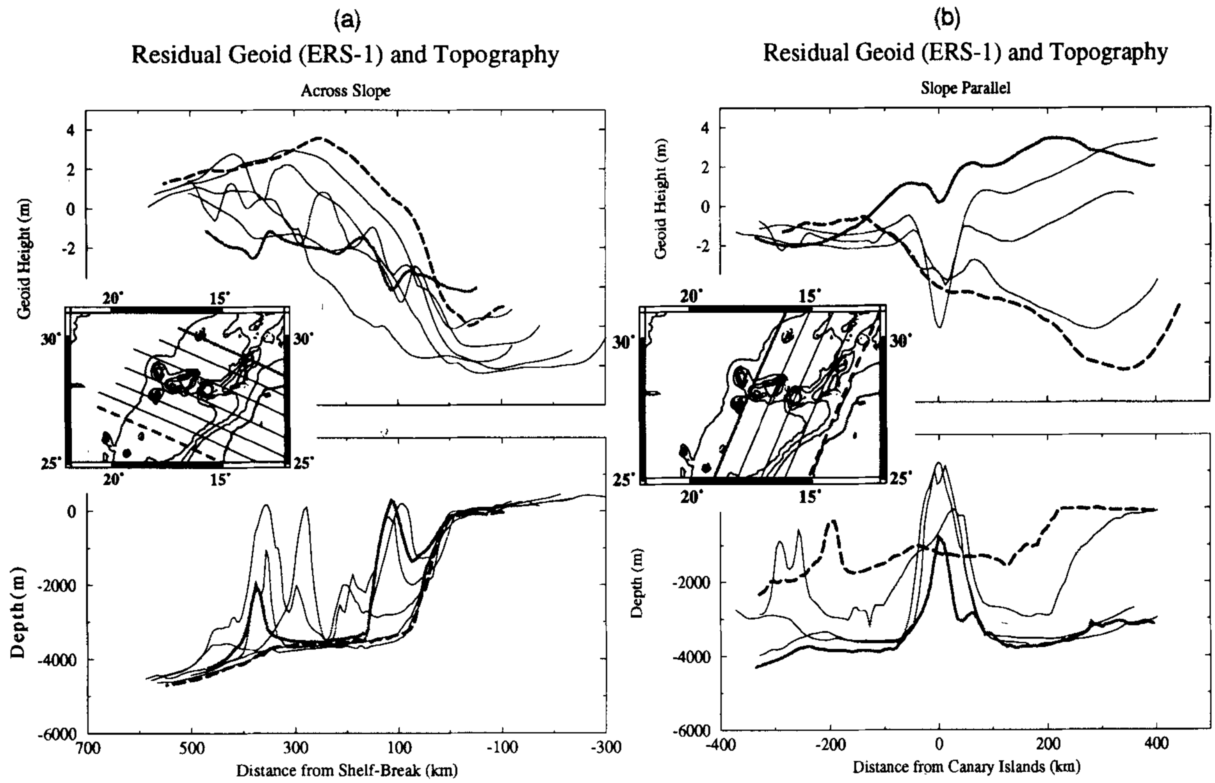

Figure 13. Geoid anomaly and topography profiles across the Canary Islands region. The profiles were constructed by interpolation from 5 × 5 min. grids of geoid anomaly and topography data. The geoid anomaly was obtained by subtracting the geoid derived from smoothed sea-surface height measurements obtained during the ERS-1 mission from a geoid complete to degree and order 36. The topography is based on the same data set that was used to construct the contour map in Fig. 1. (a) Profiles that extend across the continental slope off Morocco. (b) Profiles that extend parallel to the trend of the slope. Heavy solid and dashed lines are the outermost profile on each data set.

that the re-heating cannot explain the full heigh of swells so that other, possibly dynamic, effects must be involved (Courtney & White 1986; Sleep 1990; Sleep 1992).

If swells result in part from vertical loads that act on the base of the lithosphere due, for example, to convection then it would be expected that both their height and width might be influenced by the mechanical properties of the lithosphere. For example, recent seismic studies, (Zhang & Tanimoto 1992) support the view (Sleep 1990) that plumes rise through the mantle as relatively narrow conduits (about 300 km in diameter). When they impinge on the base of the lithosphere, however, they may spread out laterally (e.g. Hill 1991) over regions as wide as 1000–2000 km. At these widths the lithosphere would appear as a relatively weak structure in response to upward-directed loads so that the swell would more or less reflect the planform of the upwelling plume head. Other studies, however, suggest that the plume may be much narrower when it impinges. Anderson, Tanimoto & Zhang (1992), for example, have argued that the plume beneath the Canary Islands is narrow (about 200 km) and is only wide in an E–W direction because it has been 'deflected' in the direction of the African plate motion. With such a narrow width in one direction it is unlikely that a broad, high swell could form in response to an upward-directed load because the flexural strength of the lithosphere would prevent it.

A narrow deflected plume model has also been invoked by Hoernle & Schmincke (1993) in order to explain the geochemical diversity in both time and space or Canary Islands volcanism. According to their model, the plume consists of two components: 'blobs' of mantle material from great depth and entrained upper mantle material. The largest concentration of material from depth occurs above the centre of upwelling which is presently located between La Palma and Tenerife. Volcanic hiatuses reflect intervals when the cooler, entrained mantle enters the melting zone. As the plate moves over the conduit, fewer hot 'blobs' will reach the melting zone and the hiatuses will become longer and more frequent. However, because the plume is deflected beneath the island chain by the plate motions there is always the potential to generate lavas downstream of the upwelling conduit. Thus, eruptions can occur at similar times along the entire chain. Tholeiites, for example, can form as far away from La Palma as Fuerteventura and Lanzarote—provided that the individual 'blobs' are sufficiently large.

Irrespective of the origin of the low T_e, or the lack of a swell, the results in this paper imply a high degree of variability in the observations that characterize hotspots. Some hotspots have low T_e and high heat-flow values (e.g. Cape Verdes and Bermuda). Others (e.g. Hawaii and French Polynesia), however, have normal values. Many hotspots are associated with mid-plate swells (e.g. Hawaii, Bermuda, French Polynesia, Great Meteor, Reunion). Others have small (e.g. Easter) or no (e.g. Canary Islands) swells. It is unlikely that such variations in T_e, heat flow and

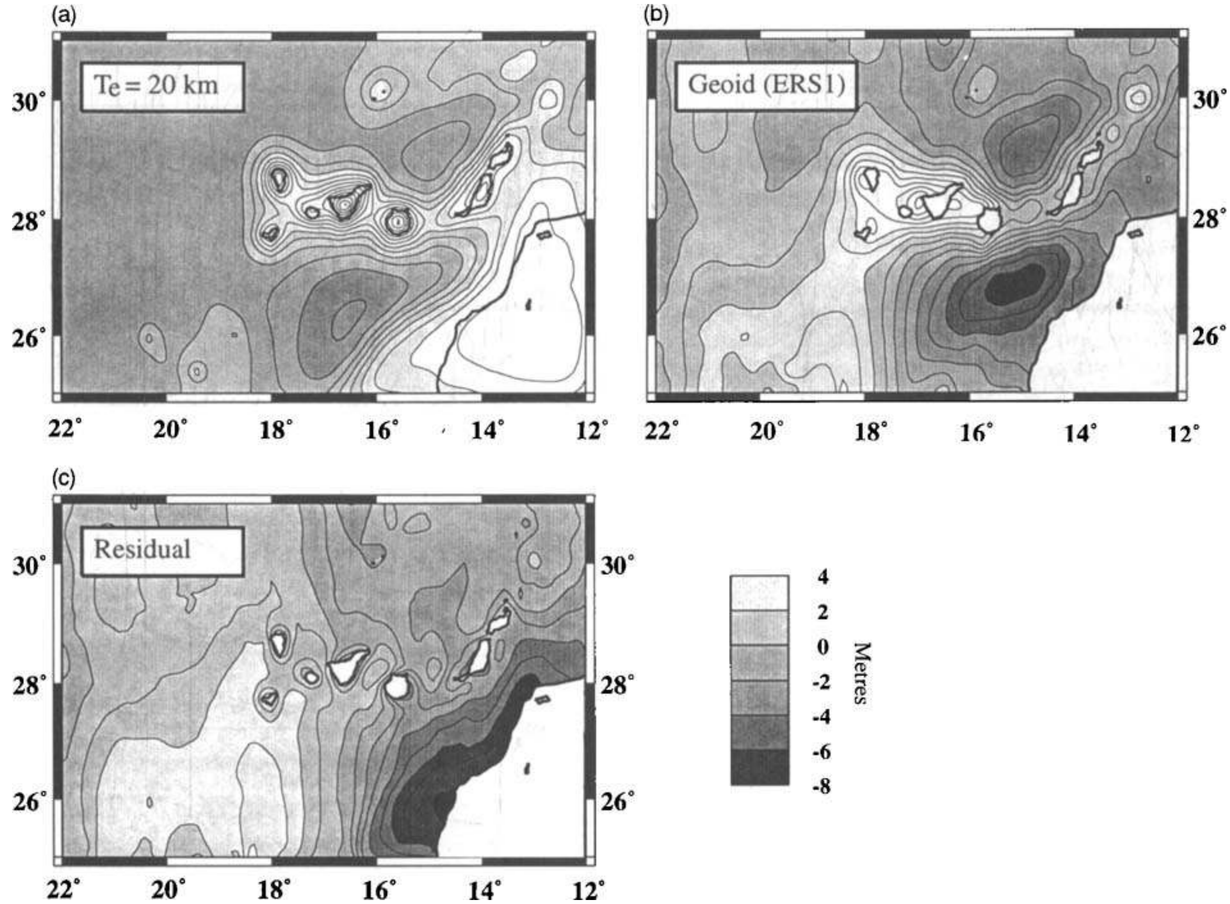

Figure 14. Comparison of observed and calculated geoid anomalies in the Canary Islands region. (a) Observed geoid anomaly. (b) Calculated geoid anomaly based on the best-fitting load and infill densities and T_e values deduced in Fig. 8. (c) Residual geoid obtained by subtracting the calculated geoid from the observed.

swell height can all be explained by differences in the heat lost by conduction or by dynamics effects of hot, low-density, ascending plumes. The lithosphere probably plays a major role and as a result makes it difficult to use surface observables such as elastic thickness, heat flow and depth to infer about the dynamics of mantle plumes. Variations in absolute plate motion and their long-term thermal and mechanical properties have been identified as two contributing factors but, others may exist. The result is a high degree of individuality in hotspot settings such that cases exist where there is little apparent relationship between the local anomalies of individual oceanic islands and seamounts and the regional topographic, seismic and geochemical anomalies that characterize these regions.

CONCLUSIONS

This study of crustal structure and gravity and geoid anomalies in the region of the Canary Island allows the following conclusions to be drawn:

(1) the comparison of seismic and free-air gravity anomaly data with calculations based on simple elastic models suggests that the elastic thickness of the lithosphere, T_e, that underlies the volcanic loads of the Canary Islands in the region of Tenerife and La Gomera is approximately 20 km.

(2) The models suggest that the top of the oceanic crust is flexed downwards by as much as 4 km beneath the Canary Islands and that the deformation extends laterally some 150–200 km from the islands. They also suggest that the flexural depression which flanks the islands is infilled by as much as 2.5 km of Teritary sediments and that the flexural bulges, which flank the depression, may have influenced the uplift history of the Salvagen Islands and the Saharan seamounts and, possibly, the 'slope anticline' in the Moroccan continental margin.

(3) The flexure caused by the Canary Islands is superimposed on a larger flexure which is caused by sediment loading at the nearby Moroccan continental margin. The margin flexure reaches its maximum values (>12 km) beneath the outer shelf but extends to the region of the Canary Islands itself. The value of T_e that best explains sediment thicknesses in the regions that flank the islands is 5 km, which is in accord with recent studies at stretched continental margins.

(4) The Canary Islands in the region of Tenerife and La Gomera were emplaced on 140–150 Ma old oceanic

lithosphere which has an expected T_e of about 35 km, assuming a 450 °C controlling isotherm. The value deduced in this paper is therefore some 15 km *less* than expected—suggesting that there has been a significant weakening of the oceanic lithosphere in the region.

(5) The weakening is most likely the result of thermal pertubations in the lithosphere which are caused by an underlying mantle plume. The Canary Islands differ, however, from other hotspots in the Atlantic (e.g. Cape Verdes, Bermuda) and Pacific (e.g. Hawaii, French Polynesia) oceans since they lack both a long-wavelength topographic swell and a gravity/geoid anomaly high.

(6) The lack of a topographic swell can be explained if the plume that underlies the Canary Islands is sufficiently narrow that the oceanic lithosphere appears as a relatively strong structure to loads that act on its base and as a result does not deform over a broad area. There may still be some uplift due to the plume, however, but it is probably concentrated locally within the region of the islands rather than over a broad region.

(7) There is a high degree of individuality at seamounts and oceanic islands such that no simple relationships exist between their T_e and the regional anomalies that characterize these features such as mid-plate topographic swells and gravity/geoid highs. The reason for this is not clear but the absolute plate motions and the long-term themal and mechanical properties of the lithosphere both appear to play a role.

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APPENDIX: SUMMARY OF PARAMETERS USED IN CALCULATIONS

Young's modulus = 100 GPa.

Poisson's ratio = 0.25.

Thickness of unflexed oceanic crust = 8 km.

Density of mantle = 3330 kg m^{-3}.

Density of unflexed oceanic crust = 2800 kg m^{-3}.

Density of seawater = 1030 kg m^{-3}.