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A seismic study of lithospheric flexure in the vicinity of Tenerife, Canary Islands

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Abstract

Seismic data have been used to determine the crustal and upper mantle structure of Tenerife, Canary Islands, a volcanic island of Tertiary age located on > 140 Ma oceanic crust. Reflection data show that oceanic basement dips gently towards the island, forming a flexural moat which is infilled by 2–3 km of well stratified material. The moat is characterised by a major angular unconformity, which we attribute to volcanic loading of pre-existing oceanic crust and overlying sediments and the subsequent infilling of the flexure by material that was derived, at least in part, from the islands. Refraction data show that the flexed oceanic crust has a mean thickness of 6.41 ± 0.42 km and upper and lower crustal velocities of 4.8–5.4 km s^{-1} and 6.7–7.3 km s^{-1} respectively. The flexure, which has been verified by gravity modelling, can be explained by a model in which Tenerife and adjacent islands have loaded a lithosphere with a long-term (> 10^6 yr) elastic thickness of approximately 20 km. Seismic and gravity data suggest that up to 1.5×10^5 km^3 of magmatic material has been added to the surface of the flexed oceanic crust which, assuming an age of 6–16 Ma for the shield building stage on Tenerife, implies a magma generation rate of about 0.006 to 0.02 km^3 a^{-1}. This rate is similar to estimates from other African oceanic islands (e.g., Reunion and Cape Verdes), but is significantly less than that which has been calculated at Hawaii. There is no evidence in either the seismic or gravity data that any significant amount of magmatic material has "underplated" the flexed oceanic crust. The crustal and upper mantle structure at Tenerife therefore differs from other oceanic islands such as Hawaii and Marquesas where > 4 km of underplated material have been reported.

Keywords: Tenerife; oceanic crust; mantle; volcanic arcs; lithosphere; flexure

1. Introduction

The Canary Islands are a volcanic chain of Tertiary age that developed in close proximity to the Moroccan passive continental margin. According to Morgan [1] and Holik and Rabinowitz [2], the islands were generated as the African plate moved slowly over a deep mantle hotspot that is now centred beneath El Hierro, the westernmost island in the

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chain. The Canary Islands do not show, however, either the simple age progression or linear topography that would be expected of a hotspot track [3].

Bathymetry data (Fig. 1a) show that the Canaries comprise seven main islands. The easternmost islands of Lanzarote and Fuerteventura are located on a NNE-trending submarine ridge that extends for some 250 km sub-parallel to the African coast. The westernmost islands of La Palma, El Hierro, La Gomera, Tenerife and Gran Canaria, however, are more isolated, being separated from each other by narrow channels. Magnetic lineation studies [4–7] suggest that the westernmost islands overlie oceanic crust of Jurassic age (i.e., > 140 Ma). Banda et al. [8] propose that the easternmost islands also overlie oceanic crust. However, the location of these islands near magnetic anomaly "S1" [6] suggests that they may have formed on or close to the extended continental crust of the Moroccan passive margin.

The Canary Islands rise some 7 km above the expected depth of Mesozoic age oceanic crust [9] and therefore represent a large load on the lithosphere which should deform under their weight. Parameters that are sensitive to the magnitude of the load and the deformation are the gravity and geoid anomaly. By comparing the observed gravity and geoid anomaly with predictions based on elastic plate models, Dañobeitia [10] and Filmer and McNutt [11] have shown that the elastic thickness of the lithosphere, T_e (which is determined by its long-term (i.e., > 10^6 yr) flexural rigidity) is in the range 18–48 km. Dañobeitia et al. [12], using a gravitational admittance technique, estimated a T_e of 23 km, which is some 15 km lower than would be

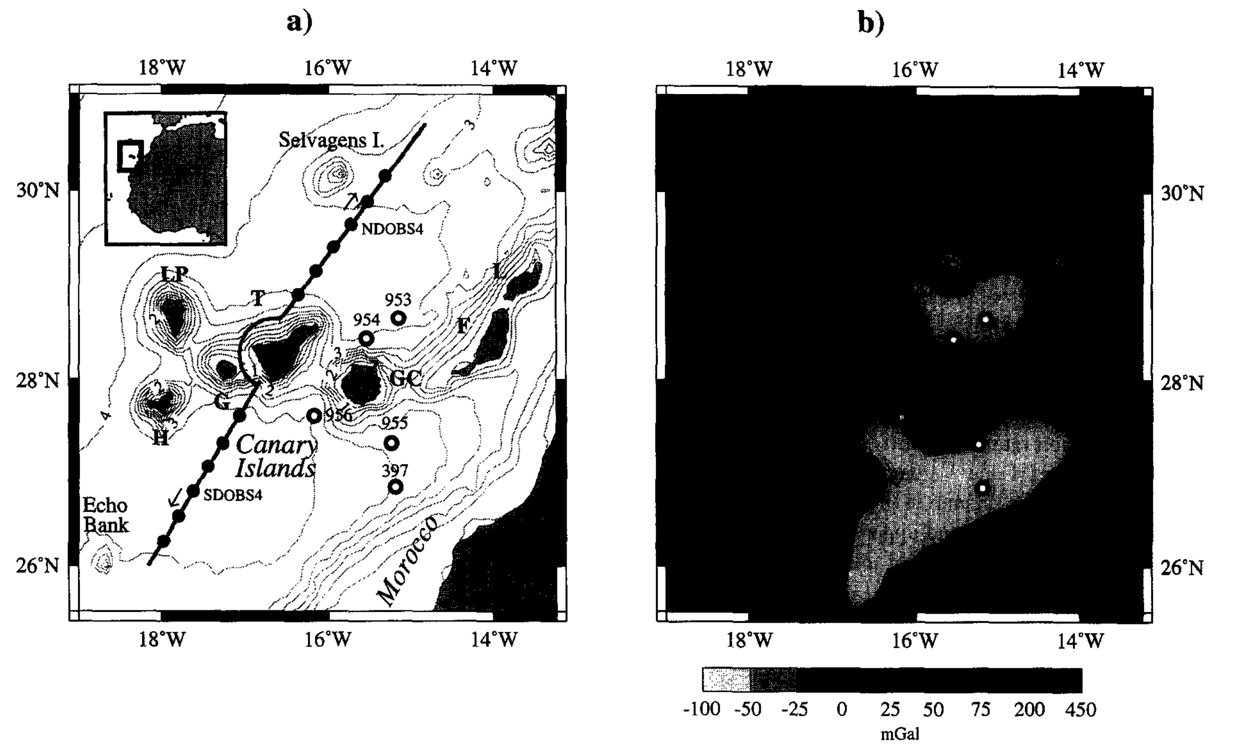

Fig. 1. Map showing the location of the Canary Islands transect. Thick lines show ship tracks of RRS *Charles Darwin* (cruise CD82) along which seismic reflection and refraction data were obtained. ● = location of the 12 DOBS deployments; ▲ = location of the land stations; partially filled circle = location of DSDP Site 397 [39] and ODP sites 953–956 [40]. (a) Bathymetry map based on National Geophysical Data Centre GEODAS data, supplemented by data acquired during CD82. Contour interval = 500 m. *L* = Lanzarote; *F* = Fuerteventura; *LP* = La Palma; *T* = Tenerife; *G* = La Gomera; *H* = El Hierro; *GC* = Gran Canaria. Small arrows show the region of the sub-seafloor where P_mP arrivals impinge on the Moho at NDOBS4 and SDOBS4 (Fig. 3). (b) Gravity anomaly map based on data compiled as part of the African Gravity Project [58] and data acquired during CD82. Free-air gravity anomalies at sea and Bouguer anomalies on land.

expected for the age of > 140 Ma oceanic lithosphere [13]. Dañobeitia et al. [12] speculated that the low value may be biased by the nearby Moroccan margin. By minimising the margin effect, they obtained a best fit T_e of 35 km. Watts [14], however, using a single NE–SW trending gravity anomaly profile which crossed the volcanic island chain between La Gomera and Tenerife, obtained a best fit T_e of 20 km, suggesting that, irrespective of margin strength, a low elastic thickness and, hence, a weak region, may exist beneath part of the chain.

Unlike other hotspot generated Atlantic oceanic islands (e.g., Bermuda and Cape Verdes), the Canary Islands do not appear to be associated with a large amplitude topographic swell. There is a regional shallowing as the islands are approached from the west [15] but, when corrections are applied for sediment thickness, the depth anomaly that results appears to be less than 800 m [16]. Furthermore, the islands lack a long-wavelength gravity and geoid anomaly high that, at Bermuda and Cape Verdes, correlates positively with the swell. The absence of a large amplitude swell at the Canaries is surprising, especially in view of the possible low elastic thickness values [12,14].

These conclusions concerning the thermal and mechanical properties of the lithosphere beneath the Canary Islands are based mainly on gravity and geoid modelling. The gravity, and especially the geoid, are influenced by all the masses in a region and it is possible that the gravitational effect of rifting, sedimentation, and erosion at the Moroccan continental margin may interfere with the anomaly associated with the Canary Islands volcanic load and its compensation.

One way to address this question is to use seismic techniques to determine the structure of the crust and upper mantle. Seismic experiments have now been carried out in the Pacific in the region of the Hawaiian [17–20], Marquesas [21] and Tuamotu [22] islands and in the Atlantic in the Madeira–Tore Rise region [23]. At Marquesas [21,24] and Hawaii [18,20] flexed oceanic crust is underlain by a high P-wave velocity (i.e., > 7.4 km s^{-1}) lower crustal body, up to 4 km thick and 100 km wide, which has been interpreted as underplated magmatic material. The cross-sectional area of the underplated material has been used [25–28] to estimate melt generation rates. In contrast to Hawaii and Marquesas, however, the Canary Islands are located on a slow-moving plate.

Previous seismic studies in the Canary Islands region have been limited to a few refraction [8,29,30] profiles along the axis of the islands and reflection [31–34] profiles of the island flanks and adjacent moats. In November–December 1993, we carried out a combined seismic reflection and refraction study of the crust and upper mantle structure in the vicinity of Tenerife as part of RRS *Charles Darwin* cruise CD82. The purpose of this paper is to present some of the initial results from the cruise, especially as they relate to: (1) the structure of Mesozoic age oceanic crust; (2) the elastic thickness of old oceanic lithosphere; and (3) the extent of magmatic underplating beneath oceanic islands.

2. The geology of Tenerife

Tenerife, the largest of the Canary Islands, is dominated by the twin strato-volcanic cones of Teide-Pico Viejo. The oldest rocks on the island are preserved in three massifs: Anaga, in the northeast, Teno, in the northwest, and Roque del Conde, in the south. The massifs have been dated [35] in the range 3.3–12.0 Ma using K–Ar techniques. About 1.8 Ma a large composite volcano, Las Cañadas, was constructed at the junction of the three older massifs. Between about 1.9 and 0.2 Ma the volcano experienced at least three major collapse events, which together formed the present day Las Cañadas caldera [36]. Subsequent eruptions within the caldera built the 3.7 km high peak of Pico de Teide. Volcanic activity has continued through to historical times with eruptions focused on a three-arm ''rift'' system that connects the caldera to the older massifs [37]. During the past 700 ka, [38] the north flank of Tenerife appears to have been subject to at least 4 major sub-aerial and submarine slides, which scalloped the coastline and initiated debris flows which extend laterally for distances > 100 km on the seafloor.

3. Seismic experiment

The seismic experiments were carried out along a 570 km long ''transect'', which extends from Echo Bank in the southwest, across the Canary Islands

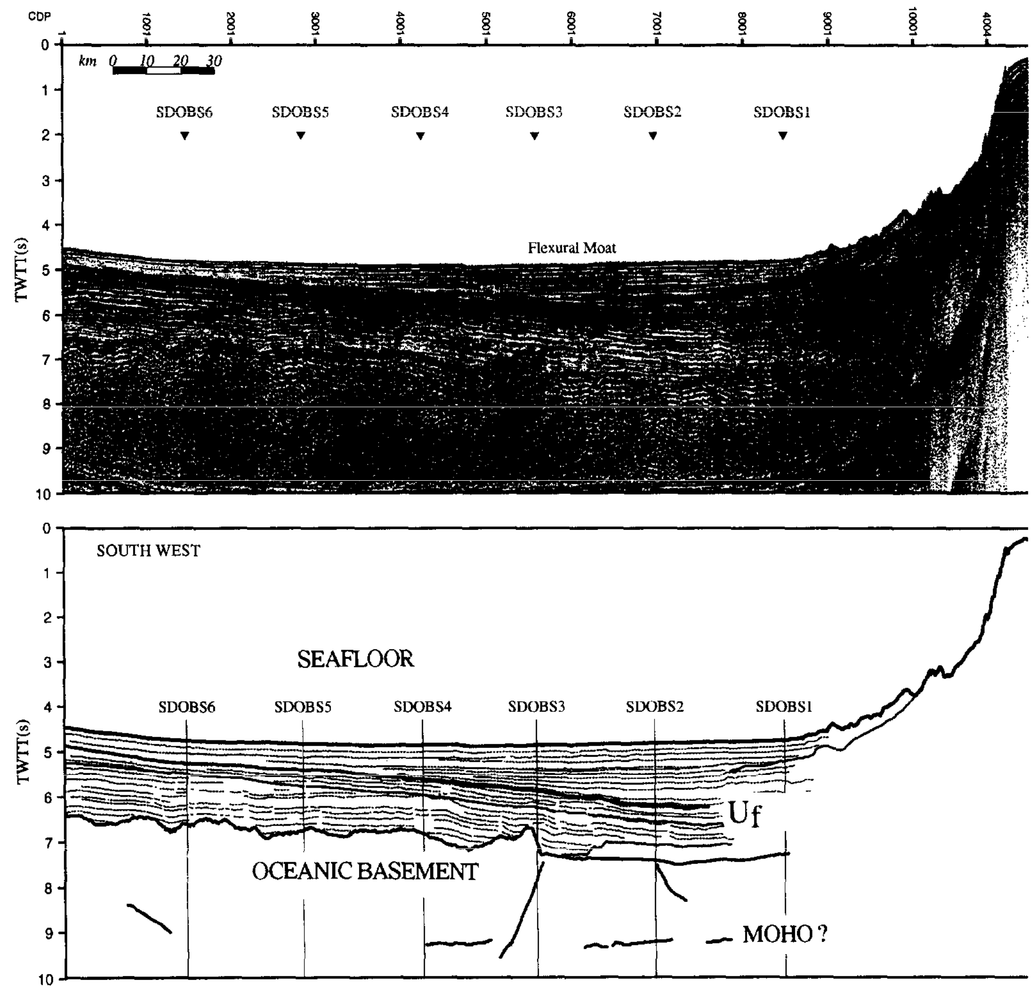

Fig. 2. Seismic reflection profile data obtained along the Canary Islands transect. Top: Filtered stack (band-pass filtered between 1–4 and 40–60 Hz) with velocity analysis every 160 CDPs (i.e., 4 km). The original 12-fold geometry has been increased to 48-fold by stacking CDP "supergathers" composed of 4 adjacent CDP gathers. ▼ = location of DOBS. Bottom: Line drawing showing interpretation of prominent seismic reflectors. Bold lines show the sea-floor, the U_f and U_{f1} unconformities and oceanic basement. Other reflectors are arbitrary and cannot be correlated either side of Tenerife.

chain between Tenerife and La Gomera, to the Selvagens Islands in the northwest (Fig. 1). The transect crossed a deepening of the seafloor, or moat, adjacent to the submarine flanks of Tenerife and a shallowing, or bulge, in flanking regions which previous gravity and geoid studies [12,14] have attributed to volcanic loading in the Canary Islands. The transect was oriented NE–SW, so as to parallel the local trend of the Moroccan margin shelf break and hence maximise the flexural loading effects of the islands whilst minimising the effects of the margin.

During the experiment, RRS *Charles Darwin* was equipped with a 2.4 km long streamer with 48 active channels; a 12-element airgun array; 6, four-component, University of Durham, Digital Ocean Bottom Seismometers (DOBS); and GPS navigation. Shot intervals and line locations were designed so that reflection data could be acquired coincident and contemporary with the refraction data. The *Charles Darwin*'s 4 compressors were able to deliver an air supply of 16.9×10^5 l $\min^{-1}$ at 13.8×10^6 Pa which, when combined with a hydrophone spacing of 50 m and a nominal survey speed of 8.9 km h^{-1},

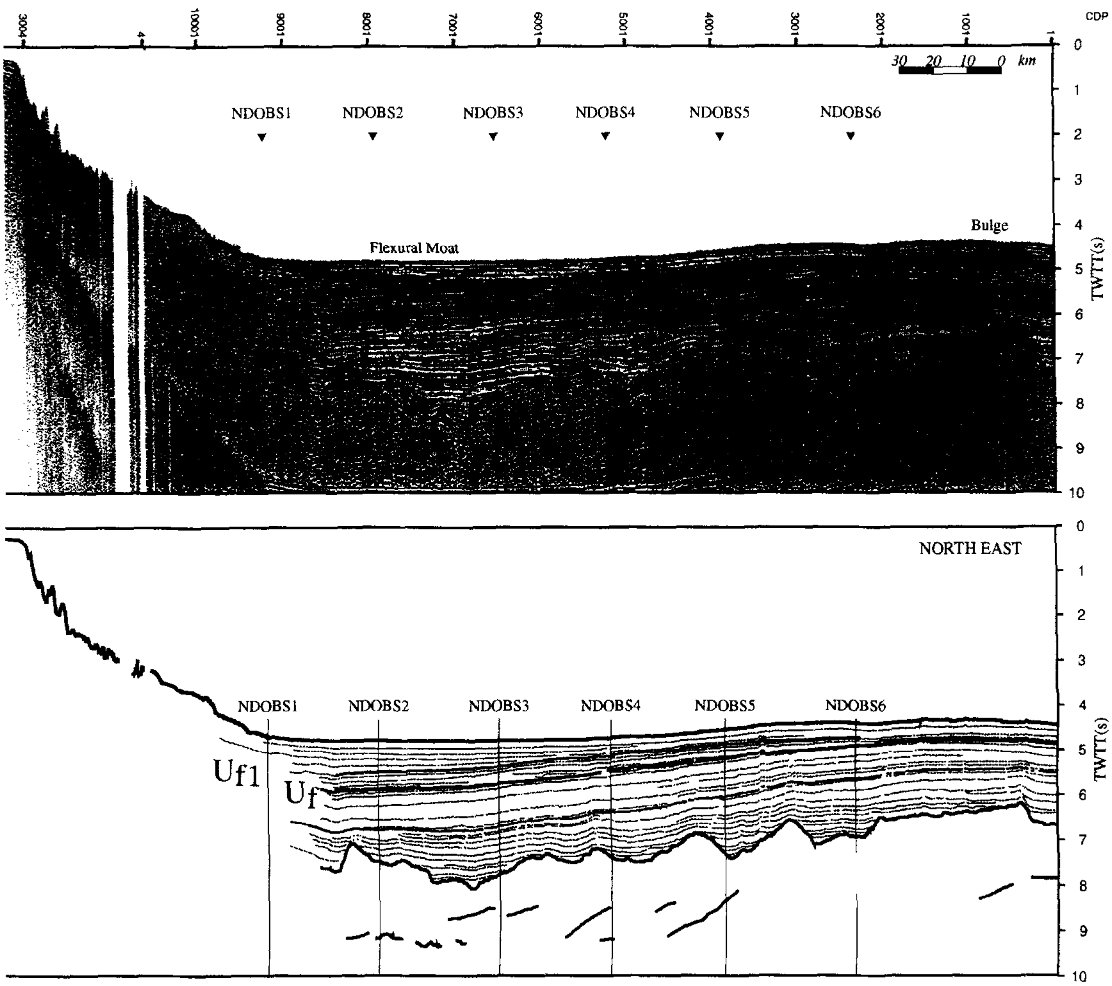

Fig. 2 (continued).

enabled a 74.8 l airgun array to be fired every 40 s. Seismic refraction data were acquired by each DOBS (deployed twice) and by 6 Institute of Earth Sciences, Barcelona, land recording stations on Tenerife.

The seismic data were processed at the Universities of Oxford and Durham using ProMax Version 6.0 software with the reflection data being de-multiplexed to SEG-Y format at sea. The main processing steps included band-pass filtering, f/k multiple suppression, deconvolution, time migration and time-variable gain filtering. Each DOBS recorded 4-channel (i.e., hydrophone and three geophone components) data which was processed into SEG-Y format using Durham "in house" software. Corrections were applied for statics using a water velocity of 1.51 km s^{-1}.

4. Seismic data

4.1. Reflection

Fig. 2 shows the normal incidence reflection profile data acquired along the transect. Oceanic basement is recognised as an irregular and rough series of reflectors which generally dip inwards towards the islands. In the bulge regions, oceanic crust is overlain by about 2.0 s Two-Way Travel Time (TWTT) of well stratified material. The TWTT increases towards the Canary Islands, reaching a maximum of about 3 s in the flexural moat that flanks Tenerife. Within the moat infill, reflector terminations define a major angular unconformity (U_f, Fig. 2).

We interpret the unconformity as the result of flexing of pre-existing > 140 Ma old oceanic crust and overlying sediments by the volcanic loads of Tenerife and adjacent islands and the subsequent infilling of the flexure by material that was derived, at least in part, from the islands. Estimates of the age of the submarine edifice of Tenerife range from 6 to 16 Ma [33,34], which suggests that the flexed sediments are Jurassic, Cretaceous, Paleogene and, possibly, early Neogene in age. There is evidence from the truncation of reflectors immediately below the unconformity, that some of the flexed sediments have been removed by erosion. Support for this suggestion comes from Deep Sea Drilling Project (DSDP) Site 397 [39] in Gran Canaria's southern flexural moat (Fig. 1), which recovered a section of Early Miocene clastic debris flows that rest unconformably on Early Cretaceous (Hauterevian) mudstones. The unconformity may extend [34] westward into Tenerife's moat.

Fig. 2 shows a pronounced asymmetry in U_f: the overlying sequences are thinner and the underlying sediments thicker to the north of Tenerife than to the south. The differences in thickness above the unconformity may be due to variations in sediment supply from the nearby Moroccan margin. Lanzarote and Fuerteventura, which are older than Tenerife, might, for example, have restricted the supply from the Moroccan slope during the late Neogene [33], the greater supply being to the south of the islands. Alternatively, the islands themselves could be the primary source of the infill material. This is supported by the provenance studies at ODP Site 953 in Gran Canaria's flexural moat [40] and sonar evidence [41,42] that land and submarine slides have shed large volumes of material from La Palma and Tenerife into their flanking moats. Irrespective of their source, it is likely that the main depocentres for the infill material have shifted as the pattern of flexure changes in response to different ages of loading along the Canary Islands chain. Unconformities *within* the moat may reflect these shifts in volcanic loading patterns. We speculate, for example, that the sediments that underlie unconformity U_{f1} in the moat north of Tenerife (Fig. 2) may have formed in Gran Canaria's flexural moat while sediments above this unconformity infilled the younger flexural depression that was created by the load of Tenerife.

The differences in thickness of the sediments *below* the unconformity may be due to differential compaction, with the sediments that underlie the moat infill south of the islands being more compacted than those to the north. However, this mechanism does not easily explain why the sediments below the unconformity maintain such a uniform thickness while the sediments above it thicken towards the islands. More likely, we believe, is that the northern part of the transect simply crosses a more distal part of the margin than the southern one.

The sediments below the unconformity are associated with a variable pattern of reflectors, including some quite bright ones which are continuous for distances of up to 50 km. There is evidence of several minor faults, some of which appear to extend up into the sediments that overly the unconformity. We see no evidence, however, of any strongly diffracting units within the sediments such as those Holik and Rabinowitz [2] identified to the northeast of the Selvagens Islands and interpreted as the result of the emplacement at crustal levels of melt generated by the Canary Islands plume.

It is difficult in the normal incidence data to determine the reflectivity of the flexed oceanic crust that underlies the Mesozoic and Tertiary sediments. We see no clear evidence of the planar reflectors that were imaged, for example, on isochron lines in the GRID [43] and OCEAN [44] survey areas, 250 km to the west and southwest of La Palma. This could be due to the greater sediment thickness in the Canary Islands region and the poor of penetration of these sediments by the *Charles Darwin* air-gun source. Alternatively, their absence might reflect [45] the general lack of reflectivity of the middle and lower oceanic crust of the African plate compared to that of the more reflective North American plate.

4.2. Wide-angle

Fig. 3 shows examples of the DOBS data that were acquired from the south and north side of

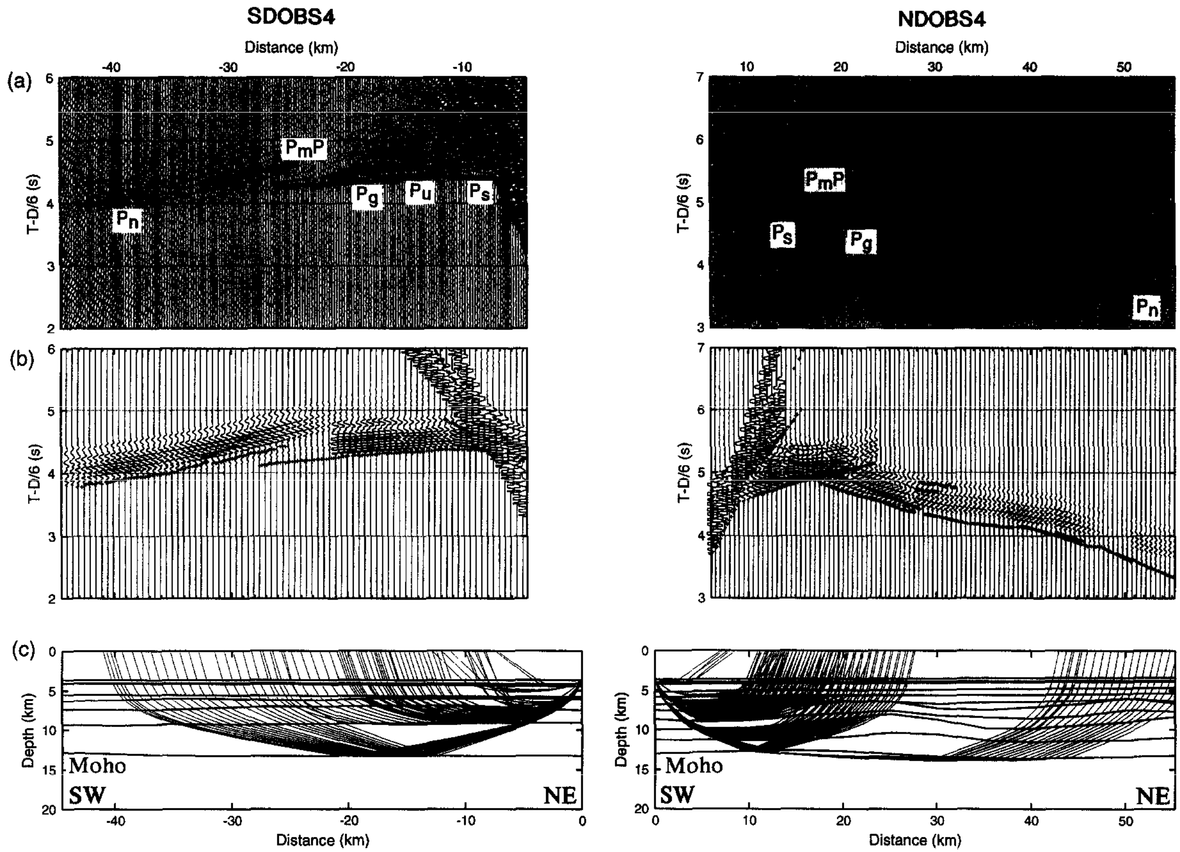

Fig. 3. Wide-angle seismic data from the southwest part of SDOBS4 and the northeast part of NDOBS4 located southwest and northeast of Tenerife (Fig. 1). (a) Vertical geophone component seismic data, range corrected and reduced at a velocity of 6 km s^{-1}. P_s = sedimentary layer diving rays; P_u = upper crustal diving rays; P_g = lower crustal diving rays; P_mP = Moho reflection and P_n = mantle diving rays. (b) Synthetic data showing the best-fit modelled arrivals. + = observed "picks" of prominent arrivals on the time–distance plot. The fit between observed and modelled arrivals is better than 50 msec. Note how the phase amplitude variation with offset has also been well modelled. (c) Ray-tracing of best-fit velocity model.

Tenerife (Fig. 1). The figure shows clear evidence of P_g, P_mP and P_n arrivals out to ranges of up to 58 km. The velocity distribution that best fits the observed arrivals was determined using a forward modelling method combined with RAYINVR of Zelt and Smith [46] which takes into account both travel times and amplitudes. The final velocity model was verified by comparing the calculated TWTT to prominent reflecting boundaries in the best fit model to the observed TWTT of prominent reflectors on the normal incidence profile. A close agreement (i.e., better than ±50 ms) was obtained for a range of prominent reflectors, indicating the general reliability of our modelled velocities.

The best-fitting velocity–depth model, determined by the combined modelling of arrivals at each DOBS, was used to determine the structure of the Tenerife region (Fig. 4q). Above sea-level, the island correlates with a region with velocities in the range of 3–5 km s^{-1}. There is evidence of a high velocity "core" to Tenerife with velocities in the range 5–6 km s^{-1}. These velocities are similar to the Hawaiian Ridge between Oahu and Kauai [18] but differ from the main island of Hawaii, where Zucca et al. [17] report velocities that exceed 7 km s^{-1}. The material infilling the moats that flank Tenerife is characterised by velocities in the range 2–5 km s^{-1}. Velocities within the infill appear to increase towards the island.

The uppermost oceanic crust is associated with a wide range of velocities (4.8–5.8 km s^{-1}) which we interpret as the result of the fracturing, porosity and weathering [47]. We see no evidence for any systematic lowering of velocities in the high curvature region between the flexural moat and bulge as was reported by Brocher and ten Brink [48] and attributed by them to flexing of the oceanic crust to the northeast and southwest of Oahu.

In general, the lower oceanic crust shows much less variation in velocity and smaller gradients than the upper crust. The onset to the lower crust is marked by an abrupt change from a higher gradient (0.4 s^{-1}) to a lower gradient layer (0.1 s^{-1}), frequently accompanied by a step in velocity to > 6.6 km s^{-1}. The velocity increases typically from about 6.7 km s^{-1} at the top of the lower crust to 7.3 km s^{-1} at its base. We see no evidence for any high velocity (i.e., > 7.4 km s^{-1}) lower crustal bodies of the type described by Watts et al. [18] and Caress et al. [49] beneath the Hawaiian and Marquesas Islands. These bodies are up to 4 km thick, 200 km in width and underlie both islands and their flanking moats. Unfortunately, the DOBS ray coverage in the Tenerife region is limited to the region beneath the flexural moats. DOBS data in these regions show clear evidence, however, of P_mP and P_n arrivals at ranges of less than 40 km. These rays have been correctly modelled in time and amplitude and do not require a high velocity lower crustal body to explain them.

The best evidence for the lack of a high velocity lower crustal body beneath Tenerife has come from modelling of the land recording station data. The data at the Vilaflor station show, for example, evidence of P_mP and P_n arrivals at ranges of 52–62 km (Fig. 5). These arrivals, which impinge on the Moho beneath the northern sub-aerial part of Tenerife, can be explained by a model with an oceanic crust which has a similar velocity structure to the flanking moat

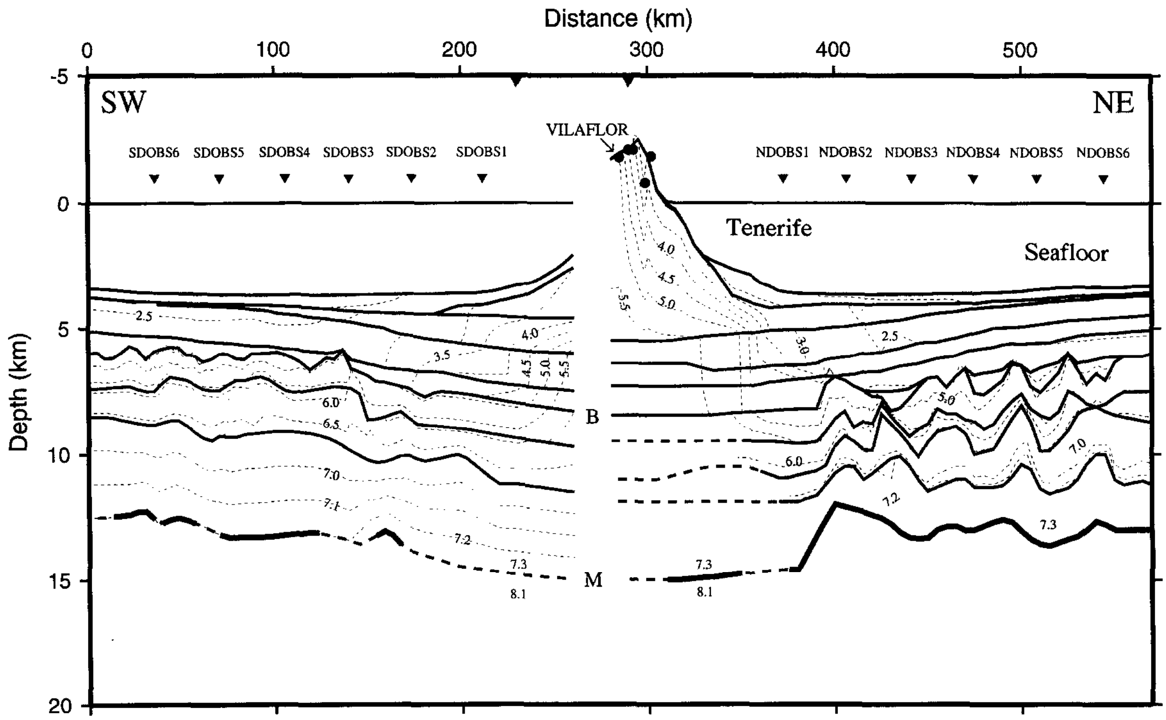

Fig. 4. Best-fit velocity model along the Canary Islands transect. Solid lines show prominent velocity discontinuities, which generate rays which are well constrained by the DOBS and land recording seismic data. Bold solid lines show region of the Moho well constrained by modelled P_mP arrivals. Bold dashed lines show interpolated discontinuities which are not constrained by the seismic data. Light dashed lines show velocity contours annotated in km s^{-1}. ▼ = location of the DOBS; ● = projected location of the land stations on Tenerife. Arrow indicates the Vilaflor station (Fig. 5). M = Moho. B = oceanic basement.

and bulge regions. The Moho depth implied by the travel times is 15 km, in accord with the earlier refraction study of Banda et al. [8]. The data cannot rule out, however, the possibility of a thin (i.e., < 1 km) high velocity lower crustal body or that a thick body may underlie the central and southern sub-aerial region of Tenerife.

Although there is a general symmetry to the crustal structure either side of Tenerife, there are differences. Oceanic crust is generally thinner to the north (6.13 ± 0.59 km) than to the south (6.74 ± 0.33 km) of the island. Such differences are surprising in view of the lack of evidence [4] for any large-offset fracture zone, and hence crustal thickness changes, in the Mesozoic magnetic lineation patterns west of the Canary Islands. There is also a greater variability in crustal thickness to the north than to the south of Tenerife. This is reflected in the roughness of the oceanic basement which, 80–240 km northeast of Tenerife, varies with a peak to trough amplitude of 500 m and a wavelength of 20–40 km. The origin of the "rough" zone is not known, but it may represent the "ridge and trough" topography of a cluster of small-offset fracture zones [50] within the Jurassic Quiet Zone.

The average thickness of the oceanic crust along the transect is less than that reported by Collier et al. [43] from the GRID survey area to the west of the

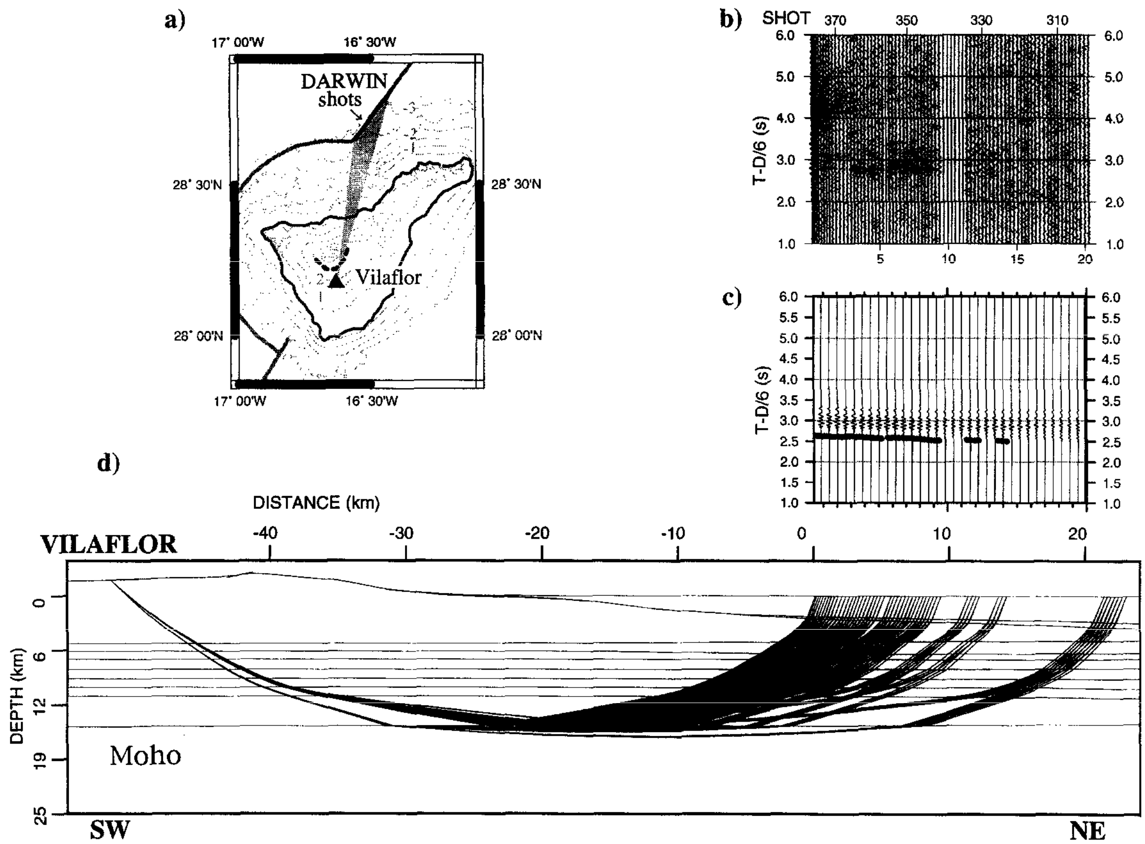

Fig. 5. Wide-angle seismic data from the Vilaflor land station on Tenerife. (a) Map showing the location of the Vilaflor station (▲) south of Las Cañadas caldera wall (bold dashed line) and *Charles Darwin*'s sea shots (bold solid line). Topographic contours are plotted at 0.5 km intervals [42]. Shaded region shows ray coverage. (b) Vertical geophone component seismic data, range corrected and reduced at a velocity of 6 km s^{-1}. The horizontal scale shows the distance (kilometres) from the nearest shot to Vilaflor. Note the decay in amplitude of the P_mP arrivals at distances > + 10 km (i.e., a shot-receiver range > 62 km). (c) Synthetic seismograms calculated from the final best-fit velocity model. Note the decay in the *predicted* P_mP amplitude at ranges > + 10 km. (d) Ray diagram of modelled phases. Note the region of the Moho beneath the northern sub-aerial flank of Tenerife constrained by the modelling.

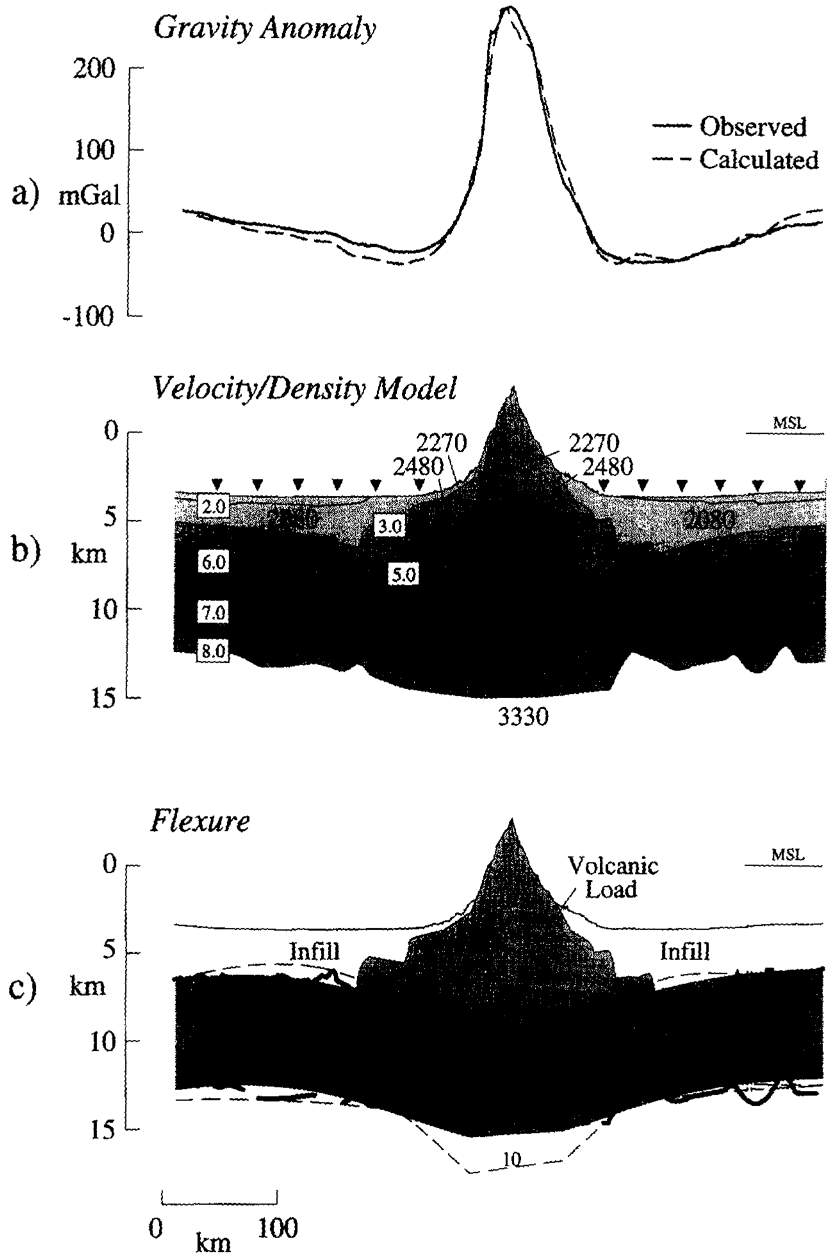

Fig. 6. Structure of the crust and upper mantle along the Canary Islands transect. Seismic and gravity anomaly data have been projected orthogonally onto a line with an origin at latitude 28.267° N, longitude 16.680° W and, azimuth 22°. (a) Comparison of the observed and calculated gravity anomalies. The "observed gravity" is a composite of the free-air gravity anomaly measured along the CD82 ship tracks and Bouguer anomalies measured on land [29]. The "calculated gravity", computed at sea-level from the final seismic model, was obtained by assigning a density to each velocity "layer" in Fig. 4 and summing their individual gravity effect using a 2-D line-integral technique. (b) Structure cross-section with P-wave velocity contours in km s^{-1}. Other numbers indicate densities in kg m^{-3} based on velocity–density relationships [47,53]. ▼ = DOBS locations. (c) Comparison of observed crustal structure with predictions of a 3-D elastic plate model for the deformation of the lithosphere with a T_e of 10, 20 and 35 km. The flexure calculations assume a "driving" load that is given by the topography above a depth of 4.5 km, a load and infill density of 2750 kg m^{-3}, a mantle density of 3330 kg m^{-3}, a mean thickness of oceanic crust of 6.41 km, a Young's Modulus of 100 GPa, and a Poisson's ratio of 0.25. Bold solid lines show only the seismically well constrained segments of oceanic basement and Moho. MSL = mean sea level.

Canary Islands but is in accord with the values reported by Holik et al. [2] and Hinz et al. [51] for the region to the north of the Selvagens Islands and seaward of the Moroccan margin. The thicknesses differ from the other limb of the spreading system, in the Blake Spur region off Florida. Morris et al. [52], for example, report an average thickness of 7.9 ± 0.60 km for ridge-parallel profiles in this region. This thickness, however, is based on widely spaced refraction profiles over oceanic crust that is some distance from the Blake Spur Fracture Zone. Within the influence of the fracture zone, crustal thicknesses are less and are more in the range 5.5–7.0 km, which overlap with the estimates from this study.

5. Gravity modelling

The velocity contours shown in Fig. 4 imply significant lateral density changes in the oceanic crust which should be reflected in the observed gravity anomaly. The seismic structure can therefore be verified by computing the gravity effect associated with these density changes and comparing it to observed anomalies.

There are two problems in using the seismic data to compute the gravity effect of the crustal and upper mantle structure. First, some relationship needs to be assumed between velocity and density. In this study, we assumed a Nafe and Drake [53] relationship for relatively low velocities associated with the Neogene and Mesozoic sediments and the mass wasting products on the flanks of the islands (i.e., velocities 1.9–4.5 km s^{-1}) and a Carlson and Raskin [47] relationship for the higher velocities associated with the core of the island and the underlying igneous oceanic crust (i.e., velocities > 4.5 km s^{-1}). The assumed velocity–density relationships yield densities of 1900–2400 kg m^{-3} for the sediments, 2700–2800 kg m^{-3} for the volcanic core and 2700–3000 kg m^{-3} for the oceanic crust. The velocity structure of the uppermost mantle is not well determined and so we *assumed* a uniform density of 3330 kg m^{-3} for the mantle.

The other problem concerns dimensionality. Unfortunately, the seismic experiment only involved a *single* transect of the island so it is not known how the velocity structure determined along the transect extends either side of the profile. We therefore assumed a 2-D approach which, as a general rule, is valid for features that are longer than about 4 times their width. Tenerife's edifice appears to be wide enough that a 2-D assumption is probably valid. However, the roughness of the oceanic basement to the north of the islands may not be 2-D and we estimate that errors of up to 5 mGal could arise in the calculations because of dimensionality.

Fig. 6a shows that, despite these problems, there is a close agreement between the amplitude and wavelength of the observed and calculated anomalies. The calculated profile shown is based *only* on the velocity model derived from the seismic data. This model (Fig. 4) included velocity discontinuities which were well constrained by the seismic data and others that were not. We made no attempt, however, to adjust the velocity model so as to fit the observed gravity anomaly more closely. The main discrepancy between observed and calculated gravity anomalies occurs to the south of the island, where the observed is generally higher than the calculated, reaching a maximum of about 10–15 mGal in the moat region. This relatively long-wavelength discrepancy may be due to the influence of the Cape Verde mid-plate swell to the southwest. Alternatively, it may reflect local differences in the density of the upper mantle to the north and south of the Canary Islands. For example, a model in which the upper mantle density is *lower* to the north than the south by about 120 kg m^{-3} will explain the discrepancy quite well.

6. Results

The close agreement between calculated and observed gravity anomalies suggest that the seismically constrained crustal and upper mantle structure (e.g., Fig. 4) can be used with some confidence in our aim to understand the geological development of the Canary Islands. In particular, the crustal and upper mantle structure has implications for the growth and decay of Tenerife, the thermal and mechanical structure of the underlying lithosphere, and the amount of magmatic material that has been added to the surface and base of the lithosphere.

6.1. Volcano growth

There is evidence from the land recording data that Tenerife has a high velocity (i.e., 5–6 km s^{-1}) central "core". The highest velocities appear to be displaced to the south of the strato-volcano of Pico de Teide, the highest peak on Tenerife. The asymmetry in velocity structure is also seen in the free-air gravity anomaly data, which show significantly steeper gradients over the south flank of the island than to the north. The high velocities correlate with the outcrop of Roque del Conde, one of three massifs which formed part of the old shield-building basalts on Tenerife. It is difficult, however, to use the land seismic data to draw any firm conclusions regarding the composition of the island.

Two 1 km deep wells on Bermuda and the Azores [54] provide some constraints on the relationship between seismic velocity and composition beneath oceanic islands. On Bermuda, for example, velocities in excess of 5.5 km s^{-1} correlate with submarine basalts, suggesting that the high velocity core to Tenerife represents the part of the island that formed during the submarine phase of oceanic island evolution [55]. The high velocity core on Tenerife is flanked by low velocities (3–5 km s^{-1}) which correlate with the strato-volcano of Pico de Teide and with Las Cañadas post-caldera flank eruptions. These basalts formed during the sub-aerial post-erosional phase of volcanic island evolution. On Bermuda, sub-aerial deposits have been largely removed by erosion but, on the Azores the sub-aerial basalts correlate with similar low velocities. Interestingly, the low velocities on Tenerife drape the high velocity core and extend to the submarine flanks of the island. This may represent material that was formed sub-aerially and has then subsided below sea level. Alternatively, it may represent mass wasting products that were derived during the later stages of volcano evolution.

As has already been pointed out [25–28], seismic data provide useful constraints on magma generation rates. Previous estimates [35] at Tenerife have been based only on volume of the sub-aerial rocks and therefore have not taken into account the volume of the submarine part of the island. Our seismic data only provide constraints, however, on the amount of magmatic material on a single cross-section of the island. If we assume that the transect crossed the centre of Tenerife's edifice and that the magmatic material is cone-shaped, with a base radius of about 100 km and a height of 10 km, then a volume of material of about 1.0×10^5 km^3 is implied. The corresponding magma generation rates, depending on whether Tenerife is no older than 6 Ma [33] or 16 Ma [34], are in the range of 0.006–0.02 km^3 a^{-1}.

The accuracy of the magma generation rates obtained from the seismic data is limited. They do not take account of any magmatic material that may have been added at the other islands (e.g., El Hierro and La Palma) at the same time as at Tenerife. They also ignore the fact that some of the material in the volume estimate may be of a sedimentary rather than magmatic origin. However, it is not considered likely that these competing effects will substantially alter the range of our rate estimates. The high end of these estimates is comparable [27] to Reunion (0.04 km^3 a^{-1}) and Cape Verdes (0.03 km^3 a^{-1}) which, like the Canary Islands, formed on the relatively old oceanic crust (65–140 Ma) of the slow moving African plate. The estimates are significantly *smaller*, however, than the rates that have been deduced from seismic data across the Hawaiian Ridge [25–28], which are in the range 0.14–0.18 km^3 a^{-1}.

This comparison between oceanic island magma generation rates indicate differences in the "productivity" of hotspots, with some producing more melt than others. It is well known that hotspots located beneath thin, young oceanic lithosphere at mid-oceanic ridges are capable of generating large melt volumes, as is evidenced by Iceland at the present day and by the Hess, Manihiki and Shatsky oceanic plateaus in the Pacific during the Mesozoic. As the lithosphere ages, its flexural rigidity increases [13] and hence it would be expected to become less vulnerable to melt penetration with age. However, at Hawaii, which formed on 80–90 Ma oceanic crust, the magma generation rates are almost as large as Iceland. Moreover, the age of the oceanic crust at Hawaii is within the age range of the African oceanic crust underlying the Reunion, Canary and Cape Verdes islands, yet apparently has a significantly higher melt generation rate than these island groups. White [27] has pointed out the importance of absolute plate motions, which are much higher for the Pacific (90 mm a^{-1}) than for the African plate.

Alternatively, it may be that the Hawaiian plume has simply been more vigorous than any of the plumes that developed beneath the African plate — due to reasons that are unconnected to absolute plate motions.

6.2. Lithospheric flexure

Previous studies based on gravity and geoid data give conflicting results for the elastic thickness of the lithosphere in the Canary Islands region. The most recent studies, for example [14,12], both of which attempt to isolate the loading effects of the nearby Moroccan passive margin, suggest T_e values that are in the range of 20 and 35 km. The value derived by Watts [14] is based on a single profile that crosses the island chain between Tenerife and La Gomera, whereas the value proposed by Dañobeitia et al. [12] is based on gravity data from a broad region that includes all of the Canary Islands, together with the Selvagens Islands and Madeira.

Elastic plate models [12,14] show that the different estimates of T_e imply different amounts of flexure beneath the Canary Islands which should be resolvable in seismic data. Beneath Tenerife, for example, a T_e of 20 km suggests that the oceanic crust should be flexed downwards by about 4 km while a T_e of 35 km, the expected value, predicts a small amount of flexure. A T_e of 10 km predicts > 6 km of flexure.

Unfortunately, the lack of arrivals which impinge on the Moho beneath Tenerife (Fig. 4) complicates the use of our seismic data to constrain T_e. A comparison of the seismically well constrained segments of oceanic basement and Moho to predictions of a 3-D elastic plate model with a load and infill density of 2700 kg m^{-3} and T_e of 10, 20 and 35 km shows that a T_e of 20 km accounts quite well for both the curvature of the oceanic basement beneath the flexural moat and the depth to Moho beneath the northern sub-aerial part of Tenerife. A T_e of 10 km predicts too great a curvature and depth to Moho, whereas a T_e of 35 km predicts too small values. The "DC level" of the calculated profile for T_e of 20 km has been adjusted vertically, however, so as to best fit the oceanic basement in bulge regions. Such an adjustment is justified because the flexural loading effects of the Moroccan margin, the magnitude of which are not known, would contribute essentially a DC shift over the length of the seismic transect. If a similar adjustment is also applied, however, to the T_e of the 35 km profile then Fig. 6c suggests that a closer fit would be obtained between the observed and calculated profiles. We cannot rule out the possibility, therefore, that the T_e is higher than 20 km, although values much in excess of 30 km would have difficulty in explaining the curvature of oceanic basement and the Moho, neither of which should not be significantly influenced by the margin.

The flexure calculations shown in Fig. 6 are based on a 3-D model and therefore ignore the effects of spatially varying load and infill densities and T_e. We may speculate, however, on how changes in these parameters might influence our results. Fig. 6b shows, for example, that an infill density of 2750 kg m^{-3} appears to be quite a reasonable assumption for the density of the load beneath Tenerife and for some distance beyond the flank of the island, but it breaks down in moat and the bulge regions. A lower infill density in these regions would decrease the load and produce less overall flexure which, in turn, would require even lower T_e values to explain the seismically constrained flexure. The influence of the sediment loads of the Moroccan margin on the Canary Islands are more difficult to quantify. However, if the stretched crust that underlies the Moroccan margin is weak (i.e., $T_e < 5$ km) — as gravity modelling at the conjugate margin offshore New Jersey, USA, suggests [56] — then there will be little mechanical coupling between the ocean and continent and, hence, only a small contribution of the margin flexure to Tenerife.

These inferences have been generally confirmed in a recent modelling study by Weddeling [57] who examined the effects of spatially varying load and infill densities and T_e on the gravity anomaly and the flexure. He found that, for a broad region around the Canary Islands, a good fit to gravity data could be obtained for infill density of 2800 kg m^{-3} beneath the islands, 2050 kg m^{-3} beneath the margin and 1030 kg m^{-3} elsewhere, accompanied by a T_e of 32 km beneath the islands and a low T_e (i.e., 1 km) beneath the Moroccan margin. Along the seismic transect, however, his model predicts an amplitude of flexure of only about 1 km — compared to 3.8 km for a model with $T_e = 20$ km beneath the islands.

The observed flexure based on seismic data is 2.5 km (e.g., Fig. 6), which suggests an intermediate value of T_e of about 26 km. These values need to be further examined by comparing the predicted curves to the actual curvature of oceanic basement, rather than to just its amplitude.

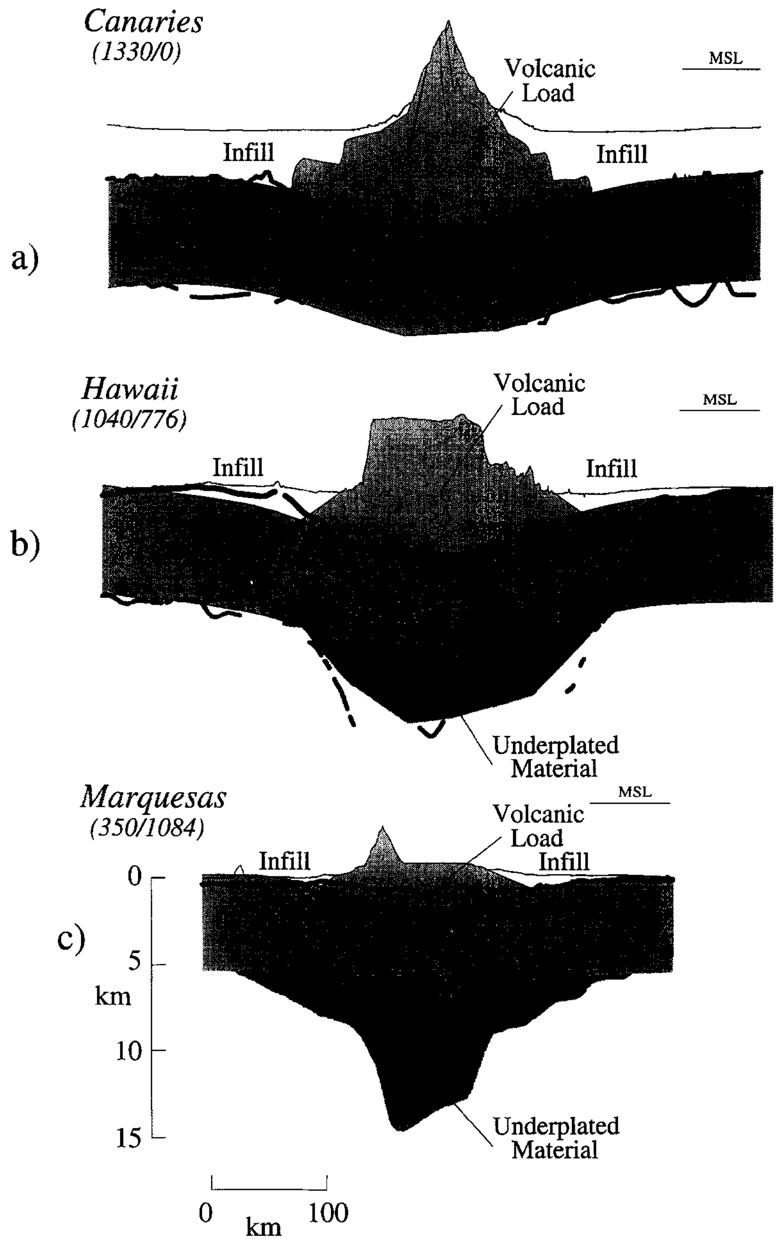

Fig. 7. Comparison of the crustal and upper mantle structure of Tenerife, Canary Islands, with other intra-plate oceanic islands. Numbers in italics indicate the cross-sectional area (in km^2) of volcanic material that has been added to the surface (first number) and base (second number) of the flexed oceanic crust. Grey shades show the flexed oceanic crust (medium) and the material that has been added to the top (light) and bottom (heavy) of the plate. Bold lines show depth to seismic reflectors 2, 3 and 4. (a) Tenerife, Canary Islands, based on this paper. (b) Oahu/Molokai, Hawaiian Ridge [18]. (c) Marquesas [21,24]. Note that, unlike Hawaii and the Marquesas, the flexed crust beneath Tenerife does not appear to be underplated.

We conclude that the new seismic data is consistent with a model in which the T_e of the lithosphere underlying Tenerife is about 20 km. Uncertainties in fitting the observed data and in the model assumptions suggest that the T_e might, however, be as high as 30 km. Assuming a range of 20–30 km, then T_e beneath Tenerife is some 5–15 km less than expected for the thermal age of the underlying lithosphere.

6.3. Magmatic underplating?

There is evidence that flexed oceanic crust beneath some intra-plate oceanic islands may be underplated by magmatic material (Fig. 7). The existence of underplating was first suggested for the Hawaiian Islands [18] on the basis of reflection data, which show a ''split'' Moho reflector, and refraction data which reveal unusually high (i.e., > 7.4 km s^{-1}) velocities in the lower part of the oceanic crust. The data suggest that the high velocity lower crustal body ''underplates'' a wide region of the flexed oceanic crust which extends from beneath Oahu and Molokai to the flanking moats. Unfortunately, the quality of the refraction data immediately beneath these islands is poor and evidence of underplating there has been disputed by Lindwall [19]. Caress et al. [21], however, has proposed that the Marquesas Islands are underlain by a 8 km thick, 300 km wide underplated body.

In contrast to Hawaii and Marquesas, the Canary Islands do not appear to be underplated. We point out, however, that P_n and P_mP arrivals at the land recording stations only impinge on a segment of the Moho beneath the northern sub-aerial flank of Tenerife. It is possible, therefore, that the central and southern sub-aerial flank of Tenerife is underplated, although our seismic data suggest that, even if it was present, it would be of much smaller lateral extent than beneath Hawaii and Marquesas. This conclusion is consistent with the gravity data which, as Fig. 6 clearly shows, can be adequately accounted for by the volcanic load and its flexural compensation and does *not* require a high velocity/high density body in the lower oceanic crust beneath the load in order to explain it.

The lack of underplating is important in regard to the problem of the origin of mid-plate swells. Both Hawaii and Marquesas are associated with a large topographic swell that rises some 1.5 km above normal sea-floor depths over distances that exceed 2000 km. As pointed out earlier, the amplitude of the Canary Islands swell does not exceed 0.8 km [16]. In addition, the Canary Islands do not appear to be associated with a long-wavelength gravity and geoid anomaly high; a characteristic feature of the large-amplitude swells. Thus, crustal underplating may play some role in the support of mid-plate swells. Further seismic experiments at other oceanic islands, such as the Cape Verdes, which loaded similar age sea-floor as the Canary Islands and have a mid-plate swell, may help resolve this problem in the future.

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