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Crustal structure and origin of the Cape Verde Rise

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ABSTRACT

The Cape Verde Islands are located on a mid-plate topographic swell and are thought to have formed above a deep mantle plume. Wide-angle seismic data have been used to determine the crustal and uppermost mantle structure along a ~440 km long transect of the archipelago. Modelling shows that 'normal' oceanic crust, ~7 km in thickness, exists between the islands and is gently flexed due to volcano loading. There is no direct evidence for high density bodies in the lower crust or for an anomalously low density upper mantle. The observed flexure and free-air gravity anomaly can be explained by volcano loading of a plate with an effective elastic thickness of 30 km and a load and infill density of 2600 kg m^{-3}. The origin of the Cape Verde swell is poorly understood. An elastic thickness of 30 km is expected for the ~125 Ma old oceanic lithosphere beneath the islands, suggesting that the observed height of the swell and the elevated heat flow cannot be attributed to thermal reheating of the lithosphere. The lack of evidence for high densities and velocities in the lower crust and low densities and velocities in the upper mantle, suggests that neither a crustal underplate or a depleted swell root are the cause of the shallower than expected bathymetry and that, instead, the swell is supported by dynamic uplift associated with the underlying plume.

1. Introduction

Oceanic volcanic islands provide information on both plate mechanics and mantle dynamics. By observing the deformation of the crust in the vicinity of submarine volcanic loads and comparing it to predictions based on elastic plate models, it has been shown that a relationship exists between the elastic effective thickness (Te) and the age of oceanic lithosphere at the time of loading (Watts, 1978). As the oceanic lithosphere ages it cools and becomes more rigid and so Te correspondingly increases. Flexure studies also provide a means by which to assess the role that processes such as thermal rejuvenation of the lithosphere and upward-acting dynamic loads on the base of the lithosphere may play in contributing to surface uplift.

Seismic techniques play a fundamental role in such studies as they provide a means by which the flexure may be measured from the deflection of sub-surface crustal interfaces in the vicinity of volcanic loads (e.g. Hawaii — Watts and ten Brink, 1989; Marquesas — Caress et al., 1995; Tenerife — Watts et al., 1996; Society Islands — Grevemeyer et al., 2001). Comparison of observations with predictions based on simple seismic-based flexure models allows estimation of the elastic thickness of the lithosphere. For example, seismic reflection data from around the island of Oahu have been used to determine a best-fit Te of 40 km to explain the observed deflection. However, this value is higher than would be expected for the 80 Ma lithosphere that underlies the island (Watts and ten Brink, 1989). Seismic refraction data across the island, on the other hand, identify a 4 km thick igneous body, with a *P*-wave velocity of 7.4–7.7 km s^{-1}, beneath the flexed oceanic crust. The authors suggest that this body is buoyant and so provides support for the overlying crustal load, making the lithosphere appear more rigid (hence older) than it actually is.

In this paper we present the results of a pilot study to estimate Te at the Cape Verde Islands using seismic refraction data acquired along an ~440 km NNE–SSW transect of the archipelago. The data are interpreted to create a structural velocity model of the crust and uppermost mantle from which the free-air gravity anomaly is calculated and compared to the observed to determine the density structure. Consequently, the combined seismic layer boundaries and densities are used to determine the Te of the oceanic lithosphere, the age of loading and if thermal rejuvenation has occurred. Using the results of this study, the origin of the Cape Verde Rise, the large bathymetric swell upon which the islands are superimposed, is also considered.

2. Regional setting

The Cape Verde archipelago is located in the Atlantic approximately 500 km west of Africa and 2000 km east of the Mid-Atlantic Ridge (Fig. 1), and comprises a horseshoe-shaped cluster of active and

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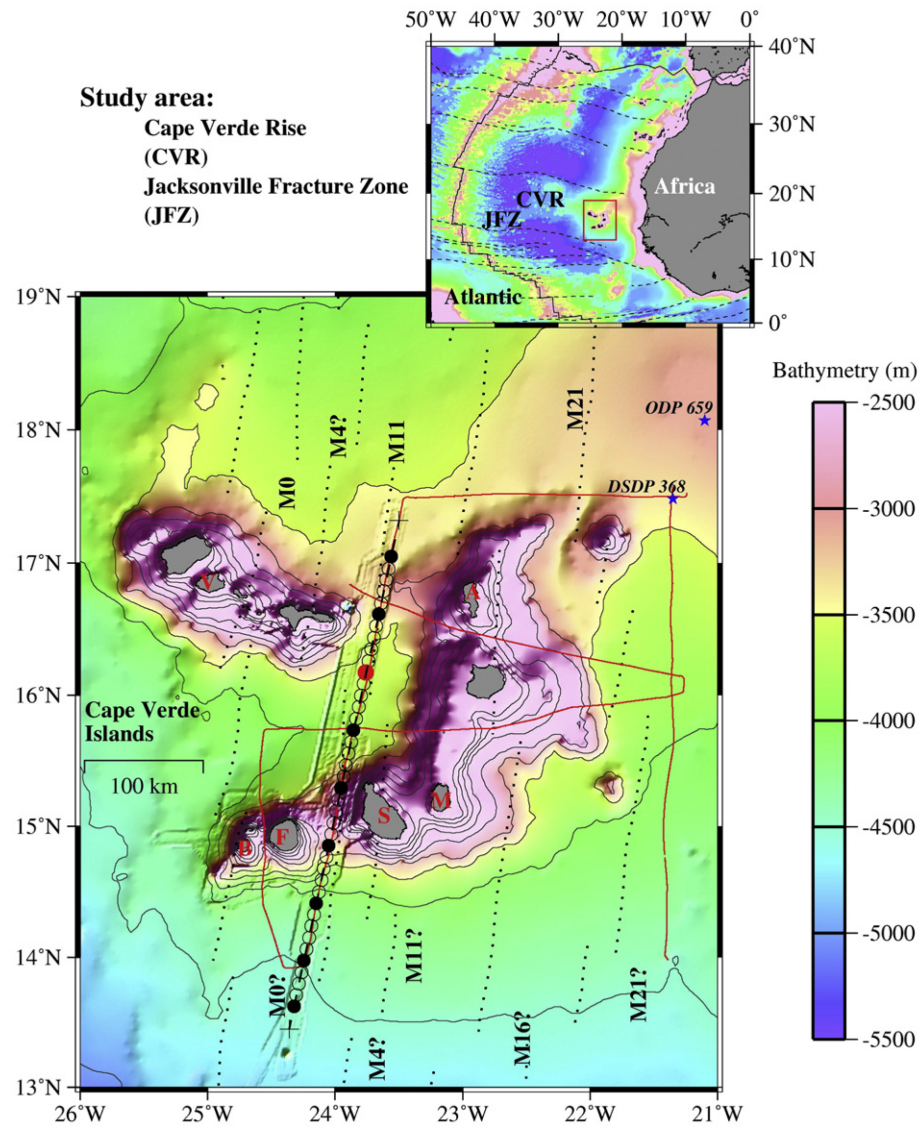

Fig. 1. Survey location. Bathymetry of the Cape Verde Rise compiled from swath bathymetry data acquired during the R/V Meteor cruise and the ship-track-derived GEBCO 1 min x 1 min grid. Individual islands are shaded grey and annotated B — Brava, F — Fogo, M — Maio, V — São Vicenté, A — Sal and S — Santiago. The locations of magnetic anomalies M0–M21 are marked by dotted lines and ODP and DSDP boreholes by blue stars. The RRS Charles Darwin seismic reflection profiles of Ali et al. (2003a,b) are shown by the thin red lines. The ~440 km wide-angle seismic refraction transect runs between the two black crosses and OBS are marked by open circles, with the nine instruments included in this study marked by black infilled circles. The red circle marks OBS 30 whose record section is shown in Fig. 3. The inset shows the location of the Cape Verde archipelago, Mid-Atlantic Ridge and fractures zones in the eastern Atlantic. The red box outlines the region shown in the main part of the figure.

inactive volcanic islands. Magnetic anomalies M0 to M21 have been identified in the vicinity of the islands suggesting that they have been emplaced on oceanic lithosphere ~125–150 Ma in age. The islands themselves range in age from ~8 Ma in the west, to as old as 20 Ma in the east.

Little is known about the geological setting of the islands. The oldest rocks exposed on the islands occur in central Maio and on the northern peninsula of Santiago, and are 128–131 Ma pillow lavas that formed at the Mid-Atlantic Ridge (Mitchell et al., 1983). In Maio, geological field observations suggest that these rocks have been tilted and uplifted about 4 km from the ocean floor to outcrop as a near vertical pipe, dipping steeply from a central intrusion complex (de Paepe et al., 1974; Stillman et al., 1982). Stillman et al. (1982) describe volcanic basement overlain conformably by a stratigraphically continuous pelagic carbonate succession that demonstrates a shallowing depositional environment from Upper Jurassic to Upper Cretaceous times.

The first phase of volcanism began in the Early Miocene (~20 Ma), reaching its peak activity in the Middle to Late Miocene (7–15 Ma) when the islands grew to their maximum size (Mitchell et al., 1983). Historically, volcanic activity has been restricted to the island of Fogo, which rises 7 km above the surrounding seafloor, and seismic activity to the nearby island of Brava (Heleno and Fonseca, 1999). Since the Late Miocene, erosion and mass wasting (e.g. Mitchell, 1998; Day et al., 1999) have deposited up to 2 km of volcanoclastic material in the flexural moats that flank the islands (Ali et al., 2003a).

The origin of the volcanism has been attributed to a hot spot: a characteristic feature of which is the association of active volcanism with a bathymetric swell (Morgan, 1971). The Cape Verde Rise is one of the largest swells in the world's oceans, rising some 2.2 km above the expected depth of Early Cretaceous-aged seafloor according to depth-age relationships (e.g. Parsons and Sclater, 1977), within a pseudo-circular region ~1200 km in diameter (Fig. 2). The swell is also associated with a geoid anomaly 'high' of ~8 m (Monnereau and Cazenave, 2000; Fig. 2) and elevated surface heat flow (Courtney and White, 1986). These observations, coupled with a lower than expected average Te of ~24 km (e.g. Young and Hill, 1986; McNutt, 1988;

Calmant et al., 1990), suggest that the lithosphere has been thermally rejuvenated by an underlying hot spot (e.g. Crough, 1978) with its apparent age reset to ~60 Ma. However, when considered in isolation, the swell height implies a thermal age of ~5 Ma (Courtney and White, 1986) from which McNutt (1988) and Sleep (1990) conclude that other upward-acting dynamic forces must contribute, such as buoyancy forces associated with magmatic underplating (Watts and ten Brink, 1989; Ali et al., 2003a) or hot spot related melting processes operating in the sub-lithospheric mantle (Lodge and Helffrich, 2006).

The first geophysical investigation of the Cape Verde Islands was conducted by Dash et al. (1976) who determined a crustal thickness of 16-17 km from 1-D slope-intercept analysis of seismic refraction data recorded by land stations installed on the islands of Sal, Santiago and São Vicenté. More recently Ali et al. (2003a) interpreted seismic reflection data to determine the stratigraphic "architecture" of sediments deposited within the archipelago. Subdividing the 1–2 km thickness of poorly to well-stratified infill material into four units, Ali et al. (2003a) demonstrate that Units I and II, deposited between the Late Cretaceous and Early Miocene, are thickest towards the east and thin westwards, suggesting that they reflect sediment loading and flexure at the West African continental margin prior to volcanic island loading. Units III and IV, however, are interpreted as Early Miocene to Recent deposits which thicken concentrically around the islands, suggesting that these units reflect flexural loading by the islands themselves and the subsequent infilling of their flexural moats. The most striking feature of the moat infill stratigraphy is that, while thickening toward the islands, sediment layers are up-arched and tilted away from the islands suggesting that the islands have been uplifted.

In September 2004, an ~440 km long, wide-angle, controlled-source seismic refraction transect across the Cape Verde Rise was carried out onboard R/V Meteor, using 40 ocean-bottom seismographs (OBSs) deployed at ~10 km intervals (Fig. 1). Shots were fired at 60 second intervals by a ~2000 in^3 (~32.7 l) water-gun array towed at ~9 knots, resulting in a shot spacing of ~190 m. Gravity, bathymetry and magnetic data were also acquired along this transect. The transect location was chosen to be coincident with RRS Charles Darwin reflection profile Line 3 from Ali et al. (2003a,b) to provide the best constraint initially on the sediment column, geometry of the basement surface, and the pattern of reflectors in the crust. For this pilot study, data from just nine of the deployed OBSs have been analysed with the goal of outlining the broad-scale crustal structure and the depth to and geometry of any flexed intra-crustal interfaces (e.g. the basement surface and Moho), to enable estimation of the extent of flexure associated with the island load and, consequently, Te to be constrained.

3. Seismic data and modeling

Fig. 3 shows an example of the interpreted record section for OBS 30 (see Fig. 1 for 1 location). One of the most striking features of the section is the lack of sediment (Ps) phases as prominent first arrivals at near-offsets, despite the sediment layer being at its thickest (~2 km) along the transect at this instrument location (Ali et al., 2003a). Phases travelling through the upper crust (Pu) are the first prominent arrival to emerge from the water wave (Ww), for example at an offset of 7 km to the SSW of the OBS. The lower crustal (Pg) phase is the next prominent arrival, extending from 10 to 35 km offset. The variation

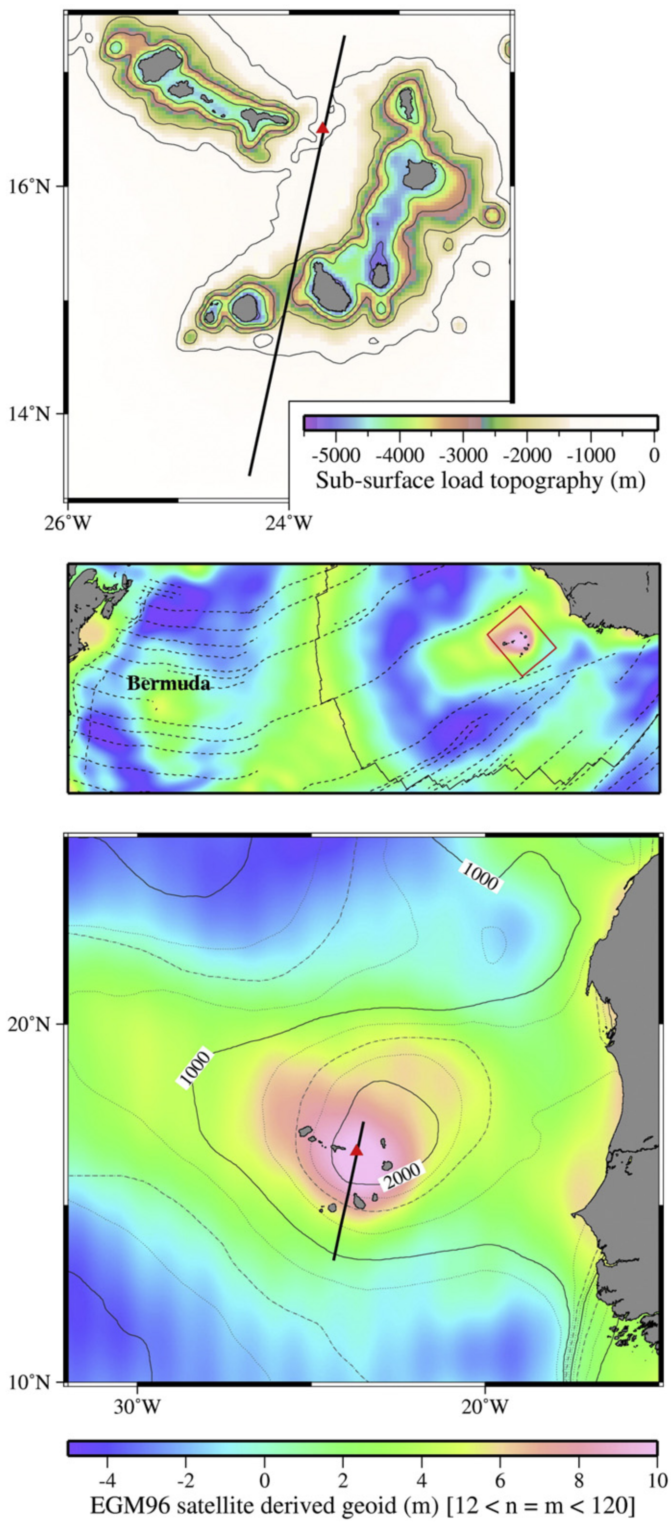

Fig. 2. Regional geoid and depth anomalies associated with the Cape Verde islands. Bottom: Geoid anomaly obtained by subtracting the EGM96 satellite-derived geoid (Lemoine et al., 1998) complete to degree and order 12 from the geoid complete to degree and order 120. The resulting anomaly, which reflects wavelengths in the range ~330 to ~3300 km, delineates well the high associated with the Cape Verde swell. The depth anomaly, which was obtained by filtering the difference between the observed bathymetry and depth of the seafloor expected for its age using a median filter with w=500 km, has been superimposed on the geoid anomaly, contoured at 250 m intervals. The seismic transect is marked as a solid line and the interpreted centre of the swell and geoid by a red triangle. Middle: The apparent conjugate relationship between the Cape Verde and Bermuda swells demonstrated by the geoid anomaly. The Mid-Atlantic Ridge and associated fractures zones are marked. The red box outlines the region shown in the main part of the Fig. 1. Top: The topography of the subsurface load required to explain the relationship between the free-air gravity anomaly and topography as a function of wavelength in the region of the Cape Verde islands (Ali et al., 2003a). Note that along the seismic transect, the thickness of the magmatic underplate comprising the subsurface load is close to the seismic resolution at this depth.

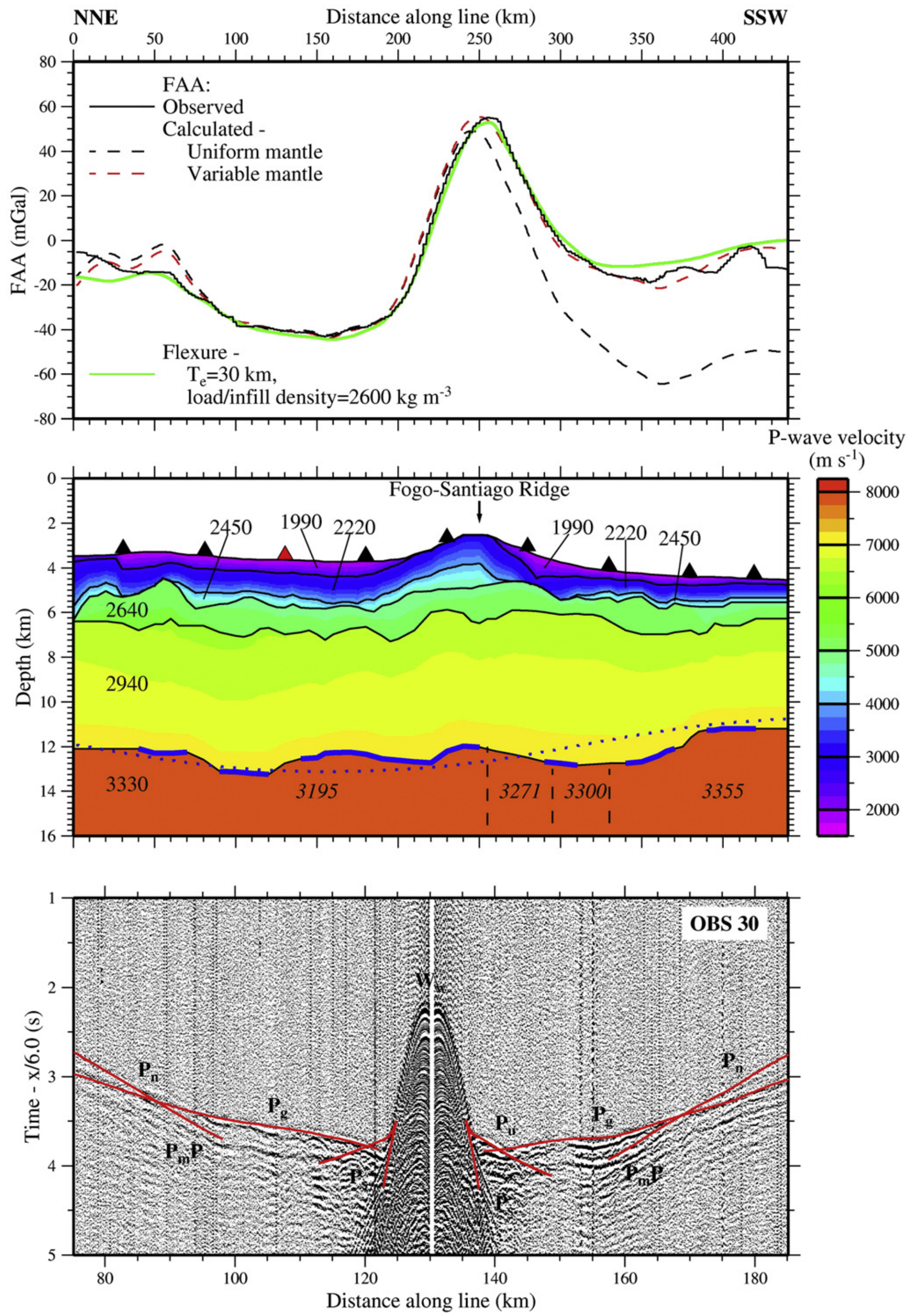

Fig. 3. Data modelling. Bottom: seismic refraction data recorded by OBS 30, reduced at 6 km s^{-1}. Interpreted phases are annotated. Red lines show the best-fit results of forward, ray-trace modelling for this instrument. Middle: preferred velocity-depth model with layer boundaries for both seismic and gravity modelling annotated as solid lines. OBS locations are marked by black triangles, with the red triangle showing the location of OBS 30. Solid blue lines show the seismic control on the Moho and the dotted blue lines show the predicted Moho depth and geometry for a Te of 30 km and a load/infill density of 2600 kg m^{-3}. Densities used for 2-D gravity modelling are also annotated in kg m^{-3}, with those in italics showing the density gradient in the mantle required to match the observed FAA to the SSW. Top: results of 2-D gravity and 3-D flexural modelling of the FAA.

in travel-time of this phase suggests corresponding topographic variation of the top basement surface and/or significant vertical or lateral velocity variation within the upper/lower crust with offset. Weak upper mantle phases (Pn) and associated Moho reflections (PmP), truncate Pg phases at an offset of ~35 km.

The wide-angle refraction dataset was interpreted and the travel-times of arrivals picked for forward modelling using RAYINVR (Zelt and Smith, 1992), using a method similar to that outlined in Watts et al. (1996) for Tenerife, to produce an outline 2-D velocity-depth model of the crust along the transect. The RMS misfit for each modelled OBS dataset is typically 0.040 s. Modelling resolution was assessed by sensitivity testing the model by adjusting layer velocities and interface depths until an unacceptable fit to within the errors resulted. A change in velocity of ±0.1 km s^{-1}, or in interface depth of ±0.15 km in the upper parts of the model, was sufficient to generate an unacceptable fit. At depth, the model is less well resolved and a velocity change of 0.35 km s^{-1} or a change in interface depth of ±0.6 km is required.

Fig. 3 shows the final velocity-depth model which comprises five distinct layers:

1) The uppermost layer is the water column with an average *P*-wave velocity of 1.50 km s^{-1} and depth 2.5–3.75 km. The depth to seabed in the model is constrained by swath bathymetry measurements made during seismic acquisition.
2) The next layer is interpreted as the sediment column, the mean thickness of which is 1.71 km. The velocity of the layer is typically 1.68–4.60 km s^{-1} with the exception of the Fogo Santiago Ridge (~250 km offset) where the velocity is higher at 3.00–4.80 km s^{-1}, most likely indicative of eruption and/or mass wasting from the adjacent islands.
3) Below, the next model layer has an average thickness of 1.16±0.37 km and a velocity of 4.90–5.20 km s^{-1}. This layer shows little variation along the length of the transect, except for its upper and lower boundaries which demonstrate significant topographic variation. The *P*-wave velocity and shallow velocity gradient of this layer suggest that it represents the upper, extrusive part of the oceanic crust — oceanic crustal layer 2 (White et al., 1992).
4) The next layer has a mean thickness of 5.95±0.45 km from 0 to 360 km along the transect. Between 360 and 440 km, however, the layer thins with an average thickness of 4.78±0.26 km. The *P*-wave velocity of this layer is 6.40–7.10 km s^{-1} along the entire length of the transect suggesting that this layer represents the lower, intrusive part of the crust — oceanic crustal layer 3 (White et al., 1992).
5) The final, deepest layer is interpreted as the upper mantle and has a velocity of 7.80 km s^{-1} at the Moho which is ~12 km below sea level. The Moho shows some topographic variation at the limits of resolution of the model at this depth, but is considered well-constrained by modelled PmP and Pn arrivals.

The crustal structure beneath the Cape Verde transect can be divided into two distinct zones of thick and thin crust. Between 0 and 360 km, the crust has a mean thickness of 7.16±0.48 km, comparing well with an expected thickness of 7.08±0.80 km for normal oceanic crust of similar age (White et al., 1992). There is also no evidence that crustal structure has been affected by fracture zones. This is compatible with the results of Ali et al. (2003b) who show little evidence from seismic reflection profile data for fracture zones along Line 3. From 360–440 km the crust is thinner, with a mean thickness of 5.72±0.34 km. The transition from thick to thin crust is abrupt, occurring over 40 km offset, and is accommodated by a thinning of the lower crust. To the south of 410 km model offset, crustal thickness is unconstrained. Consequently, it is not possible to determine how far the thin crust extends, nor whether it is associated with the Jacksonville Fracture Zone which intersects the transect at this point (Williams et al., 1999), or if it reflects thinning of the lithosphere as a consequence of long-term heating by an underlying mantle plume (Courtney and White, 1986), given that the Cape Verde islands are effectively stationary, sitting near the pole of rotation of the slow moving African plate (McNutt, 1988).

Modelling has provided no evidence of lateral variation in *P*-wave velocity immediately beneath the Moho. There is also no direct evidence for the crustal underplate proposed by Ali et al. (2003a) nor for a crust thickened to 22 km and underlain by a high velocity (8.1–8.6 km s^{-1}), low density body in the mantle identified by Lodge and Helffrich (2006) from the analysis of *P*–*S* mode converted phases recorded by a network of seismographs temporarily installed on the islands. However, in this study we only investigate the swell structure away from the edifices of the islands and the mantle is only sampled to a depth of ~2 km by Pn phases, and so an anomalously low density mantle below this may not be entirely ruled out.

4. Gravity modeling

Gravity modelling is a way of both independently assessing the uniqueness of the velocity-depth model and investigating the deep crustal and upper mantle structure least constrained by the seismic modelling. The method presented here is based upon the approach used by Watts et al. (1996) to model the gravity anomaly in the vicinity of Tenerife.

A simple density model was constructed from the preferred velocity-depth model with *P*-wave velocities in the sediment layer converted to density using the relationship of Nafe and Drake (1957) and crustal velocities using the relationship of Carlson and Raskin (1984). As there is significant *P*-wave velocity variation within the sediment layer, for gravity modelling it was divided into three layers using the 2.0 km s^{-1} and 4.0 km s^{-1} iso-velocity contours. The widely accepted densities of 3330 kg m^{-3} and 1030 kg m^{-3} were included initially for the mantle and water column respectively. The free-air gravity anomaly (FAA) was then calculated using a 2-D line integral technique. Fig. 3 shows the density model and a comparison between the calculated and observed FAA.

The short wavelength fit is good, suggesting that the sediments and upper crust of the velocity constrained density model approximate the true density structure. However, the long wavelength fit is poor, particularly to the SSW. To match the observed FAA, sensitivity testing demonstrates that the Moho would have to deepen by ~2 km between 0 and 330 km or shallow by ~0.6 km between 330 and 440 km. These values are in excess of the seismic model resolution on the Moho.

An alternative cause of the misfit may be due to our assumption of a uniform density mantle of 3330 kg m^{-3}. To fit the observed FAA best, lateral changes in mantle densities of 3300, 3195, 3271, 3300 and 3355 kg m^{-3} are required, from NNE to SSW along the transect (see Fig. 3). The resulting density depth model produces a much better FAA anomaly fit to within the error bounds and without any changes to the model interface geometries and depths derived from the seismic modelling. This density model implies an anomalously low density mantle to the north of the Fogo-Santiago Ridge, beneath the centre of the swell.

A third possibility is that the 2-D assumption is invalid since there are significant regional changes in bathymetry and flexural loading (Ali et al., 2003a) suggesting that crust and mantle structure is likely to vary either side of the transect. To test this possibility, we have used 3-D elastic plate models (e.g. Walcott, 1976) to compute the combined gravity anomaly associated with volcano loading and its flexural compensation and compared it to the observed FAA.

The volcano load was constructed using a combination of bathymetry measured during the R/V Meteor cruise and the ship-track-derived GEBCO 1 min x 1 min ft grid (see Fig. 1). We assumed in the flexure calculations a load and infill density in the range 1000–3000 kg m^{-3}, a density and thickness of the crust based on the 2-D density-depth model, and Te in the range 10 to 50 km. All other parameters, summarised in Table 1, were kept constant.

The best-fit between the calculated and observed FAA was achieved with a Te of 30 km and a load/infill density of 2600 kg m^{-3}. This value of

Table 1
Flexural modelling parameters

Parameter	Value
Young's modulus (E)	100 GPa
Poisson's ratio (σ)	0.25
Density of oceanic crustal layer 2	2640 kg m^{-3}
Thickness of oceanic crustal layer 2	1.0 km
Density of oceanic crustal layer 3	2940 kg m^{-3}
Thickness of oceanic crustal layer 3	6.0 km
Density of the mantle	3330 kg m^{-3}
Density of the water column	1030 kg m^{-3}

Te explains both the amplitude and wavelength of the observed FAA and the regional crustal and mantle structure and depth to seismic Moho (Fig. 3). Moreover, the Te that has been derived is in accord with the expected value, based on the age of the oceanic lithosphere that underlies the Cape Verde islands.

We note that the 3-D flexural model accounts for the observed gravity anomaly, even at the SSW end of the transect in the region where 2-D modelling implies lateral density variations in the sub-crustal mantle. We believe that this is because the 3-D model over-simplifies the lateral and vertical density structure within the crust and mantle, especially so for the infill density which will vary along the transect and is unlikely to equal the load density, being highest beneath the load and lowest in the island flanking moats.

The density of the volcanic load is, unfortunately, not well constrained by our seismic and 2-D gravity modelling. Watts et al. (1996) estimate a density in the range 2700–3000 kg m^{-3} for the volcanic core of Tenerife. However, the 2-D gravity modelling constrains the average density of the material that infills the flexural moat at 2200–2320 kg m^{-3}. Therefore, it seems likely that the load/infill density of 2600 kg m^{-3} constrained from 3-D gravity and flexure modelling reflects an average of the actual load and infill densities. This average value is consistent with that of Ali et al. (2003a) who found a best-fit load/infill density of 2700 kg m^{-3} for the region as a whole.

The existence of magmatic underplate, as suggested by Ali et al. (2003a), would predict a thicker crust and, hence, smaller FAA when compared to a flexure model. Our results, however, provide no evidence for magmatic underplating, either regionally or locally beneath individual islands or beneath the Fogo-Santiago Ridge where Ali et al. (2003a) predict up to 2 km thickness (Fig. 2). However, comparing our modelling results with Ali et al.'s (2003a) flexural predictions of the dimensions of the subsurface load and considering that the main surface loads are located more than 50 km away from the transect, the predicted thickness of underplate along the transect would be comparable to the lower limits of that resolvable within the data and modelling resolutions. Therefore we cannot rule out the possibility of a small, spatially limited, contribution of magmatic underplate to the crustal structure.

5. Discussion

Hot spot bathymetric swells, like the Cape Verde Rise, and their associated volcanism are generally viewed as the surface expressions of plumes ascending from the deep mantle. Three models have been proposed for their origin, namely: a) thermal reheating of the lithosphere (e.g. Detrick and Crough, 1978); b) underplating by less dense mantle residue (e.g. Phipps Morgan et al., 1995); and c) dynamic support by hot, upwelling asthenosphere (e.g. Sleep, 1995).

The results of this study allow discrimination between, and consideration of which of these models mostly likely explains the swell. Firstly, the seismic modelling shows that there is no clear evidence for a regional low density upper mantle beneath the Cape Verde transect although imaging is restricted to the top ~2 km only along the transect. However, 2-D gravity modelling seems to suggest a lateral density gradient may exist. Flexural modelling in 3-D, on the other hand, shows that the observed FAA can be matched with a simple model in which the mantle has a uniform density of 3330 kg m^{-3} and the infill and load both have a density of 2600 kg m^{-3}. The discrepancy between the FAA predicted by the 2-D gravity model and the observed, therefore, seems to be a result of 3-D effects rather than an anomalous mantle. The absence of a low density mantle also makes the existence of a depleted swell root, as suggested by Lodge and Helffrich (2006), unlikely.

The velocity-depth model shows that the oceanic crust that underlies the islands has a 'normal' thickness and structure when compared to that of the Atlantic as a whole (White et al., 1992), which suggests that the Cape Verde Rise was probably not formed at a hot spot influenced mid-ocean ridge in the Early Cretaceous. Flexure modelling can best match the observed FAA for a Te of 30 km with the predicted flexural Moho generally matching that of the seismic model (Fig. 3). A regional comparison of the predicted and observed FAA shows a good match assuming a Te of 30 km (Pim, 2006), except for directly beneath the islands, which Ali et al. (2003a) suggest may be caused by localised underplate. There is no evidence in our seismic model for underplate at the base of the crust, even beneath the Fogo-Santiago Ridge, although the transect does not pass directly through any island. A regional Te of 30 km is therefore a robust result and suggests little, if any, thermal rejuvenation of the lithosphere has taken place.

The remaining hypothesis is that the Cape Verde Rise is a result of dynamic uplift due to the hot, upwelling asthenosphere of an underlying mantle plume (Ali et al., 2003a). Although previous estimates of the compensation depth of the swell (e.g. Crough 1982; McNutt 1988) are shallower than would be expected for dynamic uplift, Ali (2002) used the best-fit slope of 1°×1° averages of the residual depth anomaly (observed minus predicted bathymetry based on age) and the FAA (32 mgal km^{-1}) to estimate a compensation depth of ~107 km, suggesting the swell may indeed be supported by a deep mantle plume beneath. The results of this pilot study do not discount this conclusion. The dynamic support provided by a mantle plume may comprise, either independently or in combination, isostatic uplift from the hot buoyant plume head interacting with the base of the lithosphere, or buoyancy forces associated with the plume conduit (Sleep, 1990). The modelling of the data undertaken this far, and presented in this paper, cannot distinguish between these.

Future work that combines seismic modelling of the seismic refraction data for all 40 OBS along the transect with gravity and flexure modelling will better constrain the geometry of the Moho and the velocity (and hence density) structure of the lower crust and uppermost mantle and, hence, help resolve any remaining uncertainties concerning the origin of the Cape Verde swell.

6. Conclusions

The main conclusions resulting from this study are as follows.

1) A sediment layer, 1–2 km thick and with a velocity of 1.68–4.95 km s^{-1} overlies a rugged basement.
2) The crust on which the islands have been superimposed is 'normal' oceanic crust, with a mean thickness of 7.16±0.48 km and with upper and lower crust velocities of 4.90-5.20 km s^{-1} and 6.40–7.10 km s^{-1} respectively.
3) Gravity modelling verifies the shallow crustal structure of the velocity-depth model, but fails to confirm the depth of the Moho and/or the density in the uppermost mantle most likely due to 3-D effects.
4) There is no direct evidence for underplate or an anomalously low density upper mantle from the observed velocity structure.
5) Flexure modelling produces a good fit to the observed FAA for a load/infill density of 2600 kg m^{-3} and a Te of 30 km.
6) A Te of 30 km is expected for the age of the oceanic lithosphere beneath the Cape Verde islands, suggesting that it has not been significantly thermally reheated and weakened by an underlying hot spot.
7) The Cape Verde Rise is, most likely, the result of other deep mantle processes such as dynamic uplift which not only has elevated the islands and tilted outwards their flanking flexural moats, but shallowed the seafloor by some 2.2 km more than would be expected for the thermal age of the plate.

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