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Small-scale crustal variability within an intraplate structure: the Crozet Bank (southern Indian Ocean)

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SUMMARY

The Crozet Bank, the easternmost region of the Crozet Plateau (austral Indian Ocean), is capped by two groups of islands which form the Crozet Archipelago (Terres Australes and Antarctiques Françaises). A N–S-trending 2 km deep basin, the Indivat Basin, extends between the two groups of islands and bisects the Crozet Bank into two separate domains. The interpretation of the KeOBS8 seismic refraction profile shot during the KeOBS/MD66 cruise (January–February 1991) over the eastern Crozet Plateau was carried out by ray tracing and the computation of synthetic seismograms. This interpretation leads to a rather complex seismic structure and outlines a pronounced eastward crustal thinning from 16.5 to 10 km at the expense of layer 2. The thinning of the crust is abrupt east of the Indivat Basin. Unlike below the Hawaiian Islands and the Marquesas Islands, no underplated mantle material underlies the crust of the Crozet Bank. Moreover, this deep seismic sounding has further revealed that a high-velocity feature outcropping the seabed 30 km east of Ile aux Cochons could be a major structural feature, deeply rooted within the lower crust. The occurrence of this feature (a submarine volcano?) is associated with a mantle dyke causing a steep shallowing of the lower crustal interfaces. Gravity modelling was performed along line KeOBS8, with a density structure deduced from the seismic section, to model free-air anomalies derived from altimetry data. This modelling confirms that the Indivat Basin, underlined by a gravity low running roughly N–S between the two groups of islands, is a major structural boundary. As the model generates medium-wavelength anomalies of adequate amplitude, it also confirms that the volcano, located west of the Indivat Basin, is a deeply rooted feature. The Crozet Bank clearly appears as a plume-affected structure, which may have originated from a deep thermal anomaly within the lithosphere. More recent volcanic episodes, related to a still active plume activity under the Crozet Bank, could have uplifted upper-mantle material and caused the emplacement of the newly discovered feature and of the western group of islands.

Key words: aseismic ridge, Crozet Islands platform, crustal structure, gravity anomalies, hotspots, seismic refraction.

THE SUBMARINE CROZET PLATEAU AND THE CROZET BANK

The submarine Crozet Plateau (southern Indian Ocean) is a composite E–W-trending, shallow bathymetric feature, which extends roughly between 43°E and 53°E and is 300 km across (Fig. 1). To the west this feature abuts the Southwest Indian Ridge. The Madagascar Ridge is the symmetrical structure to the Crozet Plateau with respect to the Southwest Indian Ridge. Both features overlap in pre-anomaly 24 reconstructions (Goslin 1981; Goslin & Patriat 1984). The western region of the submarine Crozet Plateau, the Del Cano Rise, has a rather smooth completely submerged topography, and culminates on average at a depth of roughly 1500 m below sea level.

The Crozet Bank (Fig. 2) sits at the eastern end of the Crozet Plateau on oceanic crust emplaced sometime before anomaly 31 at the axis of the Southeast Indian Ridge (McKenzie & Sclater 1971). The Crozet Bank is on average shallower than the Del Cano Rise and is steep-sided and capped by two separate groups of volcanic islands. Between

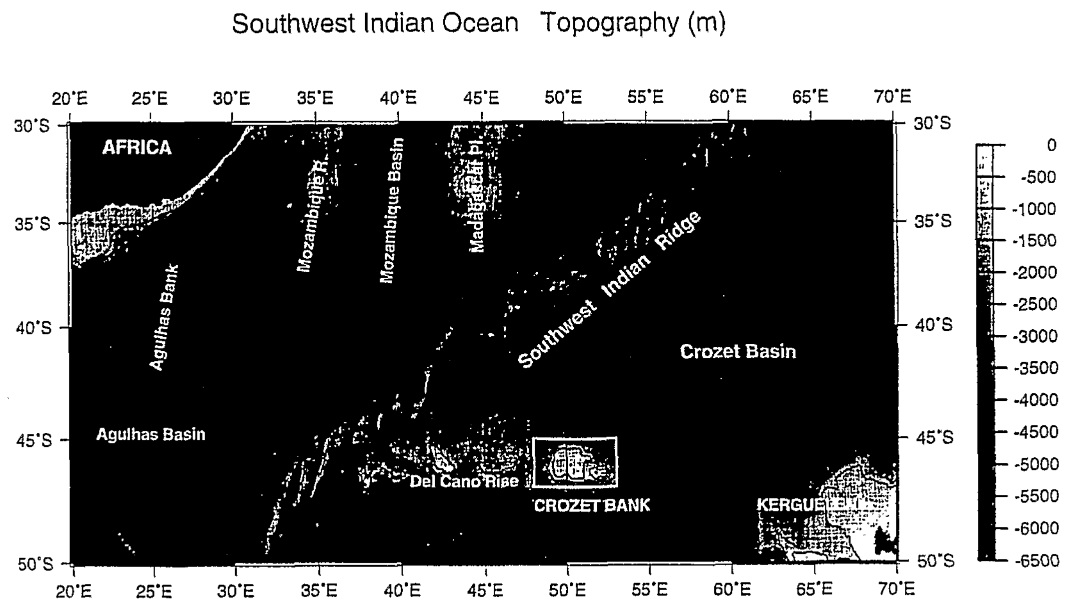

Figure 1. General topography of the Southwest Indian Ocean contoured from the ETOPO5 gridded data set. The Crozet Bank is outlined by a rectangular white frame, which delineates the bounds of Figs 2 and 7.

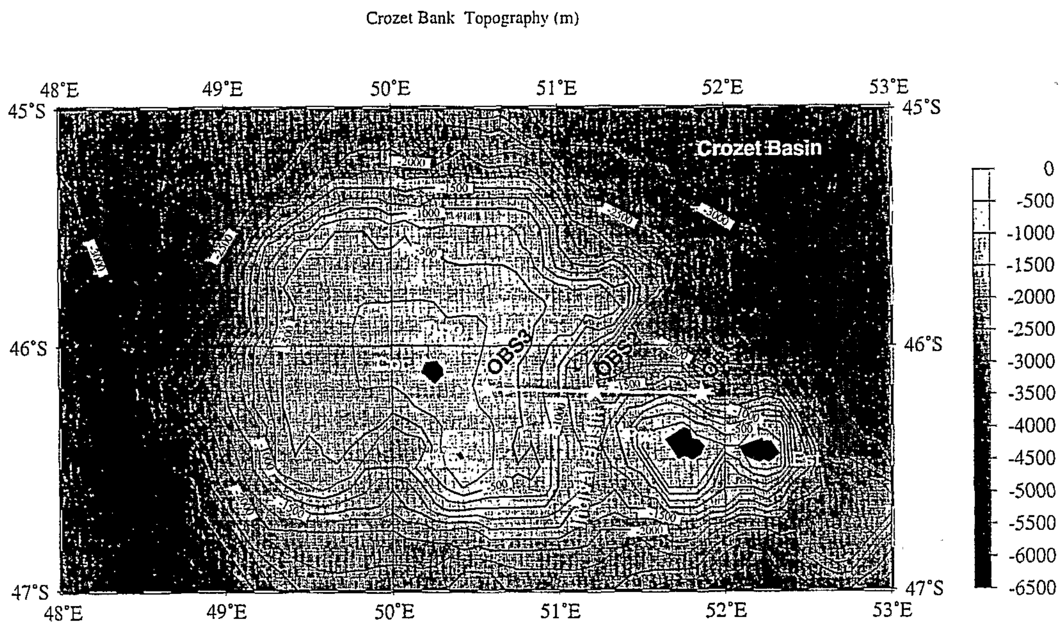

Figure 2. Simplified bathymetric map of the Crozet Plateau contoured from ETOPO5 data. The 250 m contours are shown as plain lines. The contours are labelled every 500 m. The locations of OBSs are indicated by white stars. The names of the islands have been abbreviated, from east to west: 'Est': Ile de l'Est; 'Pos': Ile de la Possession; 'Pin': Ile des Pingouins (Penguin Island); 'Coc': Ile aux Cochons (Hog Island); 'Apo': Iles des Apôtres (Apostles' Rocks). English island names are those quoted in the 'Challenger Reports'.

the two groups of islands is the Indivat Basin, which reaches depths of 2000 m.

The western group of islands comprises the small strato-volcano of Ile aux Cochons, culminating at 770 m above sea level, and the reefs of Iles des Pingouins and Iles des Apôtres. The eastern group, 120 km to the east, comprises Ile de la Possession (19 km long, 15 km across) and the easternmost arid Ile de l'Est (19 km long, 10 km across). Ile de la Possession, the largest island of the archipelago, rises over 1000 m above sea level, and Ile de l'Est culminates at 900 m.

There is no sedimentary material on the islands except some detrital Quaternary deposits (Bellair 1964). Ile de la Possession and Ile de l'Est are predominantly composed of alkaline basaltic rocks with fragmental deposits and lava flows (Philippi 1912; Reinisch 1912; Lacroix 1923, 1940; Tyrrell 1937; Dreux & Rémy 1963; Girod & Nougier 1971; Bellair 1964; Gunn *et al.* 1972; Chevallier 1981). The earliest volcanism occurred 8 Ma ago on Ile de la Possession, while Iles aux Cochons, Ile des Pingouins and Ile des Apôtres were formed by more recent volcanism, only 1 Ma ago (Chevallier & Nougier 1981;

Boudon & Nougier 1982; Lameyre & Nougier 1982). Recent volcanism in the Crozet Islands is confirmed by CO_2-rich hot springs.

Magnetic anomalies (McKenzie & Sclater 1971; Goslin 1981; Schlich 1982) and the short sonobuoy refraction line MD5–16 running over Del Cano Rise (Goslin *et al.* 1981) demonstrate that the Crozet submarine plateau is undoubtedly an oceanic feature. Magnetic mapping in the southern Crozet Basin (Goslin 1981; Schlich 1982) confirms that the E–W trend of the plateau is parallel to anomalies 31 and 33. This is consistent with the assumption that the plateau sits on crust that spread from the axis of the Southeast Indian Ridge. Symmetrical magnetic lineations have been observed in the Central Indian Basin, including the sequence of anomalies 33–31 (McKenzie & Sclater 1971). These results substantiate the volcanic foundation of the Crozet Bank, according to Goslin & Patriat (1984), at the time of anomaly 32 near a spreading centre.

The interpretation of profile MD5–16 shows that the oceanic crustal thickness is about 14–17 km under the Del Cano Rise, similar to that measured over the southern Madagascar Ridge (Goslin *et al.* 1981). Two single-channel reflection profiles GA3–22 and GA3–24 (Goslin 1981), orientated SE and NE, respectively, were shot downslope from Ile aux Cochons. The sediment cover reaches 1.2 s (twtt) downslope on profile GA3–24. Unfortunately, no seismic reflection data have been collected on the Crozet Bank.

Preliminary information on the deep structure of the Crozet Bank can be derived from sparse reliable gravity data collected by Rouillon 1963) from Ile de la Possession and Ile aux Cochons. The Bouguer anomaly in Ile de la Possession ranges from 206 to 235 mgal, but is only 95 mgal in Ile aux Cochons. These values are higher than those observed in the Kerguelen Archipelago, 45 mgal on average (Charvis 1984; Recq & Charvis 1986). The large difference between gravity anomalies in Ile de la Possession and Ile aux Cochons is the first indication of a major change in the structure of the crust from west to east across the Indivat Basin. Recent studies of volcanic plutonic complexes in Ile de l'Est led Zhou *et al.* (1995) to suggest a continental influence, a situation comparable to the South Madagascar Upper Cretaceous tholeiites. However, in the last 9 Ma, magmatic evolution has occurred in an entirely oceanic environment, producing a major alkaline series and resulting in the build-up of the Crozet Archipelago (Zhou *et al.* 1995).

Moreover, the isostatic responses of the lithosphere to the load of the Del Cano Rise and to that of the Crozet Bank are strikingly different (Goslin & Diament 1987). According to these authors, Del Cano Rise and the Crozet Bank, although presently adjacent, were constructed in different ages by distinct processes. The Del Cano Rise was formed on a young lithosphere near the active Southwest Indian Ridge, whereas the Crozet Bank was generated by a genuine intraplate process and overloaded an older and cooler lithosphere.

In 1984, the MDFLUX-INDIVAT heat-flow survey (Courtney & Recq 1986) was carried out on the R/V Marion Dufresne. The profile was run northwest from the deep Crozet oceanic basin towards the centre of the Crozet Bank. The positions of the stations were chosen to lie on crust of the same age (67 Ma from the identification of the magnetic anomaly) in order to lessen the influence of lithospheric cooling along the profile. Heat flow increases significantly from a predicted value of 58 mW m^{-2} for a 67 Ma plate (Parsons & Sclater 1977) to higher values, 96 mW m^{-2} and even 165 mW m^{-2} at a less reliable station on the island platform. This suggests a thermal origin for the uplift of the Crozet Bank, a theory substantiated by the high geoid anomaly, +9 m (Courtney & Recq 1986). Neither a mere crustal thickening with Airy local compensation, nor an increase of the plate's elastic thickness (Goslin & Diament 1987) can account for the observed geoid/topography ratio (G/T). Indeed, the geoid anomaly, 1.7 m, and the G/T, 0.8 m km^{-1} computed for an Airy-type isostatic response do not match the values of 9 m and 4.9 m km^{-1}, respectively, measured over the Crozet Bank. The ratio of 4.9 m km^{-1} is similar to that seen over other oceanic hotspot swells (Courtney & White 1986; Detrick *et al.* 1986). Courtney & Recq (1986) postulated that a convective plume would cause dynamic uplift of the Crozet Bank and would cause the low density of the mantle material below the platform.

THE KeOBS8 REFRACTION LINE ON THE CROZET BANK

Refraction line KeOBS8 (Fig. 2) was shot in early 1991 during the MD66/KeOBS cruise to elucidate the structure of the crust and uppermost mantle of the Crozet Bank. This experiment was conducted on R/V 'Marion Dufresne'. This 111 km long profile (Fig. 2) runs eastwards from the vicinity of Ile aux Cochons, where the water depth is about 0.4 km, towards the northern slope of the Crozet Basin, passing over the 2000 m deep Indivat Basin (Vanney *et al.* 1986). KeOBS8 is the only long-range refraction profile that has been carried out over the Crozet Plateau, and thus the sole experiment providing information on the deep structure of the eastern portion of the Plateau. KeOBS8 was not formerly planned in the MD66/KeOBS cruise. This additional profile was carried out on the way back to Ile de la Réunion. Time was running out. Only three OBSs from ORSTOM were deployed for this experiment without any loss, one at either end and one in the middle of the profile, in the Indivat Basin (Figs 2 and 3). Excellent weather conditions enabled us to collect high-quality seismic data.

The seismic source consisted of an untuned array of eight 16-litre air guns with a shot spacing of $\approx$300 m. GPS provided continuous and precise ship positioning. The locations of OBSs on the sea bottom were computed relative to the ship using the traveltime and polarity of the direct water wave (Nakamura *et al.* 1987). Further technical information on MD66-KeOBS is reported in Operto (1995), Charvis *et al.* (1995, 1998) and Operto & Charvis (1995, 1996).

MODELLING THE SEISMIC STRUCTURE OF THE CROZET BANK (FIG. 3)

Seismic data were first processed at the ORSTOM laboratory at Villefranche-sur-mer near Nice (France). Ray tracing and synthetic seismograms were then computed in Brest at the Université de Bretagne Occidentale using the RAYAMP program (Spence *et al.* 1984). The interpretation of the profile is based primarily upon first arrivals. Reflection phases are not well identified on seismic record sections. For convenience, the profile running westwards from OBS1 to OBS3 is labelled P13; the profile running eastwards from OBS3 to OBS1, P31; from OBS2 to OBS1, P21; from OBS2 to OBS3, P23.

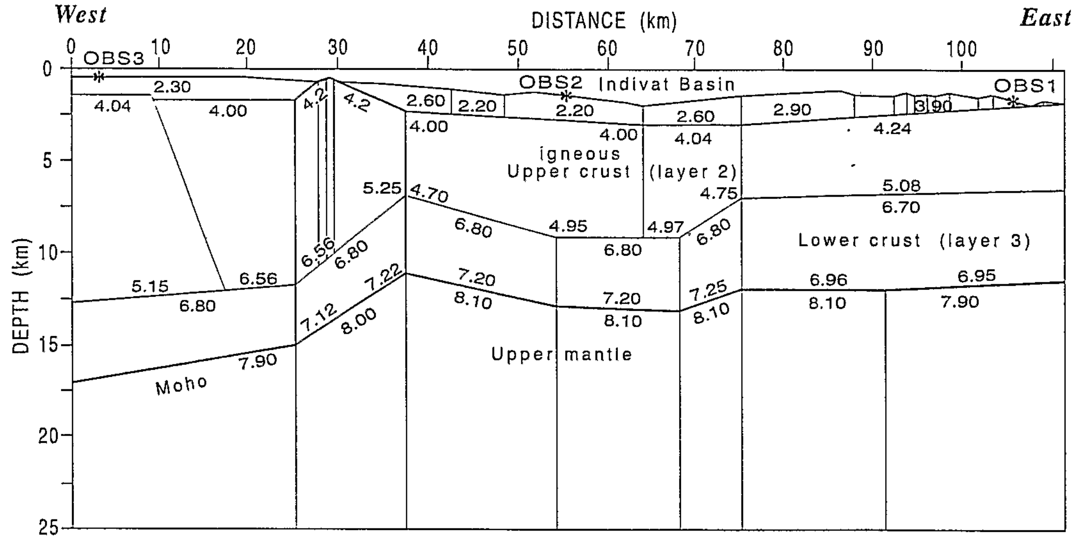

Figure 3. Best-fit P-wave velocity model of the Crozet Bank from the interpretation of seismic profiles P13, P31, P23, P32, P12 and P21.

The detrital Quaternary deposits

As mentioned above, no reflection seismic profile has been carried out on the Crozet Bank itself. Modelling the shallow structure using ray tracing leads us to assume the occurrence of a layer between Iles aux Cochons and Ile de la Possession just below the seabed with a velocity increasing downwards from 1.8 to 4.0 km s^{-1}.

East of Ile aux Cochons, in the vicinity of OBS3, the depth of the seabed gently increases from 0.4 to 0.6 km. This smooth slope is discontinued further east by an apparent small-scale edifice at 0.48 km below sea level, culminating at 0.12 km above the seabed.

From observations on the islands, it is inferred that the seabed material is composed of detrital deposits or volcanic ash. The thickness of this layer varies greatly from 0 to 1500 m in the Indivat Basin. On the eastern slope of the Crozet Bank, the detrital deposits are very thin or non-existent. Bottom water circulation is intense near the archipelago, eroding the sediments and hampering thick deposition of detrital debris. Many attempts to perform heat-flow measurements on this platform during the MDFLUX-INDIVAT cruise were unsuccessful (Courtney & Recq 1986). The probe could not penetrate into the hard and gravelly seabed, as shown by coarse grains of pyroxene-rich basalt embedded in the probe during its ascent. The rather low P velocity (1.8–2.4 km s^{-1}) within the seabed west of the Indivat Basin is a consequence of the high porosity of this thin material, which facilitates the penetration of sea water.

On the smooth eastern flank of the Indivat Basin, shallow velocities are higher (2.4–3.2 km s^{-1}) than those within detrital deposits of the western flank. Further west of the eastern flank, the depth of the seabed increases slowly to 1.6 km at the end of the profile on the slope of the Crozet Bank towards the Crozet Basin. The eastward transition from smooth to rugged seabed is abrupt. Peaks observed eastwards on the seabed are probably outcropping volcanic edifices. The velocity within the uppermost layer is quite high, 3.6–4.0 km s^{-1}, compared to that observed westwards, 2.0–2.8 km s^{-1}.

The upper igneous crust

It seems reasonable to anticipate that the igneous crust below the Crozet Bank is structurally complex, with strong lateral inhomogeneities. To the west, the Crozet Bank abuts the thick oceanic crust of Del Cano Rise (Goslin *et al.* 1981), while the eastern end of KeOBS8 approaches the oceanic deep Crozet Basin.

Just below the hard and gravelly seabed, velocities of 4.0–4.2 km s^{-1} are observed all along the profile and could be ascribed to altered basalts as observed over the Kerguelen Plateau (Charvis *et al.* 1995; Operto & Charvis 1996). Eastwards, the top of this layer lies at very shallow depths, just below the outcropping volcanic dykes. Velocities observed in the upper igneous crust fall within the range of those observed in oceanic layer 2 (White *et al.* 1992).

There is a progressive transition from uppermost layer 2 velocities to higher velocities ascribed to the lowermost layer 2. In the western section of the profile adjacent to Del Cano Rise, the seismic model (Fig. 3) exhibits a large increase in the thickness of layer 2 and displays a slight oblique velocity discontinuity within this layer. To the west of this discontinuity, velocities within layer 2 increase from 4.04 to 5.16 km s^{-1} at a depth of 12.7 km. Westwards, on the western flank of the submarine high-velocity edifice, probably of volcanic origin, the velocity gradient within layer 2 is higher, 0.25 s^{-1}. Velocities increase from 4.0 to 6.50–6.60 km s^{-1} at a depth of 11.8 km.

From the eastern flank of this edifice to the Indivat Basin, velocities range from 4.15 to 6.52–6.59 km s^{-1} between depths of 1.73 and 12 km. Just below the Indivat Basin, the thickness of the igneous crust increases from 4.7 to 6.2 km and sharply decreases further eastwards to 4.1 km below the eastern flank of the Indivat Basin.

The lower crust and the crust–mantle interface

Ray tracing and synthetic seismograms show that, below layer 2, velocities range from 6.8 to 7.2–7.3 km s^{-1}. Velocities reach 7.2 km s^{-1} below the volcanic feature. Compressional velocities of 7.0–7.4 km s^{-1} should be attributed to the crust (White *et al.* 1992; Brandsdottir *et al.* 1997).

The most striking feature is the eastward sharp uplift of the lower crust to 7 km below the eastern flank of the high-velocity feature. East of the Indivat Basin, the thinning of layer 2 causes an uplift of the layer 2–layer 3 interface from a depth of 10 km to 7 km, while layer 3 thickens eastwards from 4 km to 5.5 km. Upper-mantle velocities below the Crozet Bank (7.9–8.1 km s^{-1}) are slightly lower than those observed below deep ocean basins (8.2–8.3 km s^{-1}).

The KeOBS8 profile runs between two oceanic features which differ greatly in age and internal structures: the Del Cano Rise and the Crozet Basin. Thick (17 km) oceanic crust underlies the Del Cano Rise (Goslin *et al.* 1981). Below the Crozet Basin, a 4.4–6 km thick 'normal' oceanic crust overlies a 10.5 km deep Moho (Francis & Raitt 1967). As previously noticed for layer 2, the crust–mantle interface rises rapidly from 13 to 12 km below the eastern flank of the Indivat Basin. It is thus reasonable to infer that the Moho shallows eastwards below the Crozet Bank. A preliminary inspection of seismic record sections gives a hint of the eastward thinning of the crust. Unlike P23 (Fig. 4), P21 exhibits first arrivals at mantle velocities at ranges as short as 30 km (Fig. 4). Modelling the crust confirms that the Moho rises unevenly from 17 km near Del Cano Rise to 11.5 km at the eastern end of the profile towards the Crozet Basin (Fig. 3), while the crust thins from 16.5 to 10.5 km.

The submarine shallow high-velocity edifice

The seismic experiment on the Crozet Bank reveals the presence of an unexpected large feature, 12 km across, which contributes a slight signature to the seabed topography. This feature is located near Iles aux Cochons, at a distance of 27 km from the western end of the profile. Seismic record sections P13 (Fig. 5) and P23 (Fig. 4) exhibit a large traveltime anomaly, but this is less visible on P31 (Fig. 6). First-arrival P_n waves, travelling in the lower crust and the upper mantle, arrive early on P13 (Fig. 5), indicating that this feature is rooted down to the upper mantle.

Unlike Mont Ross in the Kerguelen Islands (Recq *et al.* 1994), ray tracing and synthetic seismograms do not exhibit reflections within this submarine edifice. The transition to lower-crust velocities is therefore progressive. Within this feature, velocities range from 4.2 to 6.8 km s^{-1} in the lowermost layer 2, higher than those observed elsewhere below the Crozet Bank. A high vertical velocity gradient within the upper crust (0.25 s^{-1}) is required to upturn seismic rays within this feature.

Modelling the crustal structure with a sharp shallowing of both the layer 2–layer 3 interface (from 11.8 to 7 km) and the Moho (from 15 to 11.2 km) below this edifice fits the observed data properly, thus substantiating the involvement of the lower crust and mantle in the build-up of this feature, resulting in a locally shallower mantle.

INTERPRETATION

The upper crust

Velocities within the upper-crustal material below the detrital deposits, 4.0–4.5 km s^{-1} (Fig. 3), are rather low to be ascribed to the standard oceanic layer 2 of deep oceanic basins. These velocities are encountered in basalts altered by hydrothermal circulation, in basalts interbedded with sediments deposited between successive lava flows (Munschy & Schlich 1987; Charvis *et al.* 1993; Charvis *et al.* 1995) or in low-density lava flows (e.g. altered and/or highly porous volcanic rocks), as previously postulated by Recq & Charvis (1986) for the shallow crust of the Kerguelen Islands. Sediments deposited between successive lava flows act like thin low-velocity layers within the upper igneous crust. Multiple reflections and refractions within these layers may generate amplitude decrease at short distances on seismic records. The uppermost section of the igneous crust, where velocities range from 4.0 to 4.5 km s^{-1}, is much thicker (2–4 km) below the Crozet Bank than below the Kerguelen Plateau (1–2 km), according to Charvis *et al.* (1995), and below Iceland (Bjarnason *et al.* 1993). Such postulated frequent volcanic episodes, resulting in large melt generation and occurring in a relatively short time, could be a consequence of recent activity of the Crozet hotspot. The ages of the volcanic edifices range, as previously mentioned, from 9 Ma at Ile de l'Est to 1 Ma at Ile aux Cochons.

High velocities and vertical gradients in the igneous crust of the volcano could be assigned to massive intrusion of deeper material by a large outcropping dyke post-dating the build-up of the Crozet Bank. A similar process probably occurred in the Kerguelen Islands, where the age of Mont Ross is only 1 Ma, compared to 39 Ma ascribed to the oldest material of the archipelago (Giret 1983; Giret *et al.* 1988).

Close to Del Cano Rise, the large thickness of oceanic layer 2, which reaches 11 km if altered basalts are aggregated in this layer, represents most of the increase in the total crustal thickness. Such a thick layer 2 was previously noticed below the Kerguelen Islands (Recq *et al.* 1983; Recq & Charvis 1986; Recq *et al.* 1990; Charvis *et al.* 1995; Operto & Charvis 1995). Conversely, north of Ile de la Possession and Ile de l'Est, the crust thins, the thickness of layer 2 being only 5 km, inclusive of altered basalts. There, the crust is similar to the Northern Kerguelen Plateau or Iceland, with a rather thin layer 2 and a thick layer 3. The thickness of layer 2 beneath the northern and southern Kerguelen Plateau is only 5 km (Bjarnason *et al.* 1993; Charvis *et al.* 1995; Operto & Charvis 1995). The Indivat Basin must therefore play a key role as a boundary between two types of oceanic structures. The KeOBS8 experiment definitely shows that the two groups of islands of the Crozet Archipelago sit on quite different crust: the western island group (Ile aux Cochons, Ile des Apôtres and Ile des Pingouins) appears to be linked to the Del Cano Rise, whereas the eastern group (Ile de la Possession and Ile de l'Est) appears to have originated from an intraplate hotspot.

Seismic velocities and petrology of the lower crust under the Crozet Bank

Velocities of 6.8–6.9 km s^{-1} at the top of layer 3 under the Crozet Bank (Fig. 3) might be regarded as too low to be assigned to material produced by a mantle plume. However, the downward velocity increase to 7.2–7.3 km s^{-1} found at the base of layer 3 is compatible with material generated by such a process (White & McKenzie 1989). Velocities are higher than those observed in the lower part of oceanic crust generated at spreading centres away from the influence of mantle plume (White & McKenzie 1989). These velocities, of 7.2–7.3 km s^{-1},

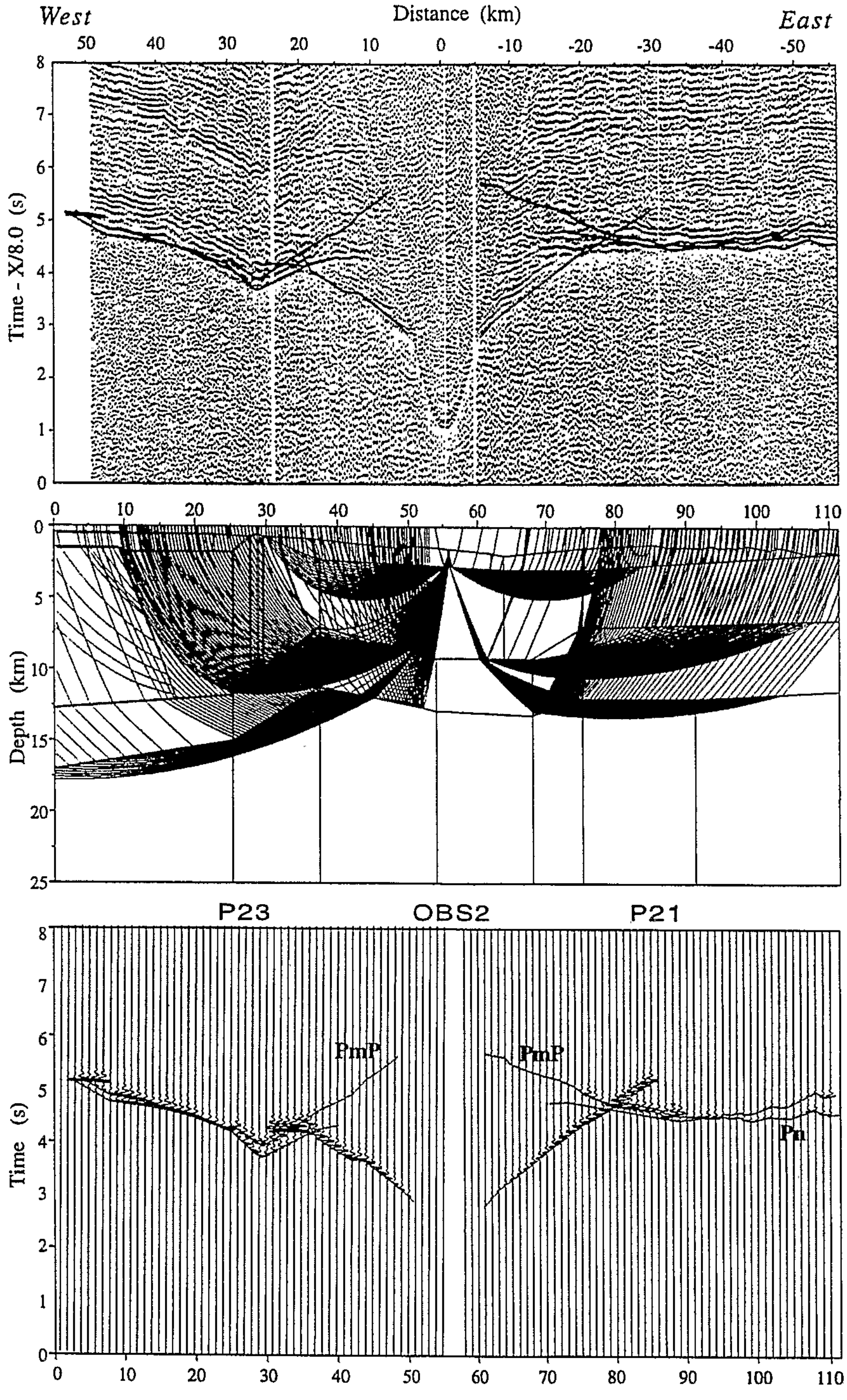

Figure 4. Record sections, ray-tracing models and synthetic seismogram along profile P23, running westwards from OBS2 to OBS3, and along profile P21, running eastwards from OBS2 to OBS1; data recorded on OBS2.

are rarely observed below deep-sea oceanic basins (Christensen & Salisbury 1975; White *et al.* 1992). The melt generated by adiabatic decompression may have a velocity of 7.2 km s^{-1} if large quantities of hot picritic melt containing an average of 16 per cent MgO are produced (McKenzie & Bickle 1988; White & McKenzie 1989). The increase of MgO in the melt is a consequence of the elevated mantle temperature in the plume, leading to contamination of the lower crust by mantle material in the vicinity of a hot spot (McKenzie & Bickle 1988).

Velocities as high as 7.1–7.3 km s^{-1} were observed at the base of layer 3, below the high-velocity feature and, generally, west of the Indivat Basin. Similar velocities are encountered below the Enderby Basin, within the area of influence of the Kerguelen hotspot (Charvis & Operto 1998) and at the base of layer 3 below the Kerguelen–Heard Plateau (Charvis *et al.* 1995). The seismic model (Fig. 3) does not exhibit any obvious trace of underplated material below layer 3. This result is similar to that of Dalwood *et al.* (1997) under the island of Tenerife (Canary Archipelago) and differs from the Pacific

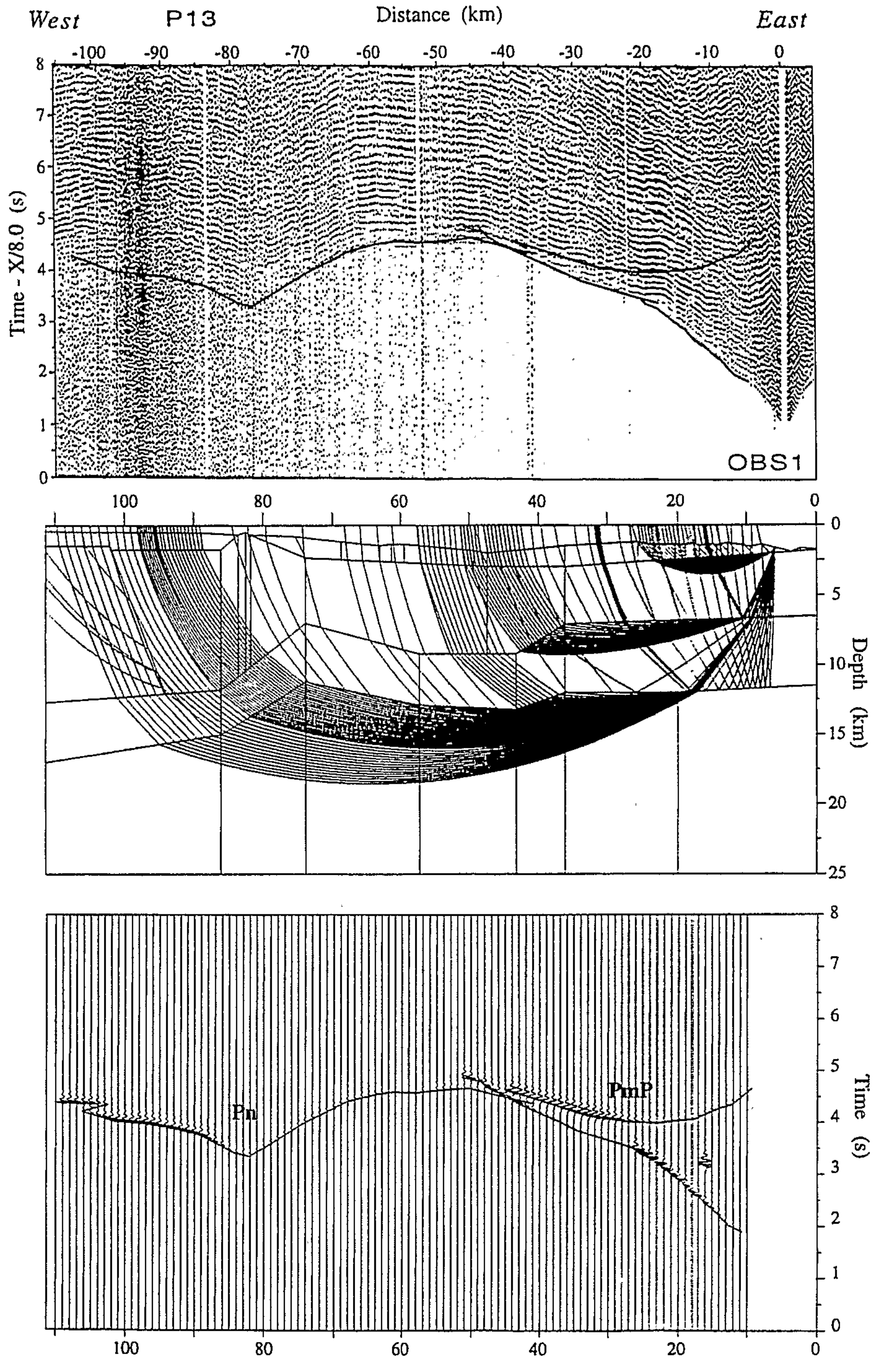

Figure 5. Record section, ray-tracing model and synthetic seismograms along profile P13 running westwards from OBS1 to OBS3; data recorded on OBS1. In Figs 4, 5 and 6, the computed hodochrons are superimposed on synthetic seismograms and the reduced velocity is 8 km s^{-1}.

islands, such as Hawaii, where thicknesses of underplated material are in excess of 4 km below the lower crust (Watts *et al.* 1985; Watts & ten Brink 1989), and the Marquesas Islands (Caress *et al.* 1995).

According to White *et al.* (1992), the normal oceanic crustal structure, calculated from a compilation of profiles modelled with synthetic seismograms, is made up from layer 2 (velocity 2.5–6.6 km s^{-1}; thickness 2.11 ± 0.55 km); layer 3 (velocity 6.6–7.6 km s^{-1}; thickness 4.97 ± 0.90 km) overlying the mantle (velocity >7.6 km s^{-1}). The total igneous crust is thus 7.08 ± 0.78 km thick on average. Undoubtedly, the crustal thickness below the Crozet Bank is far thicker, 10–12 km on average, consistent with plume-affected structures. In this latter case, the mean igneous crustal thickness, according to White *et al.* (1992), is 10.7 ± 1.6 km. Further results on Iceland (Bjarnason *et al.* 1993) also fall within this range.

The slightly low velocities within the upper mantle support the occurrence of a low-density mantle, as previously suggested

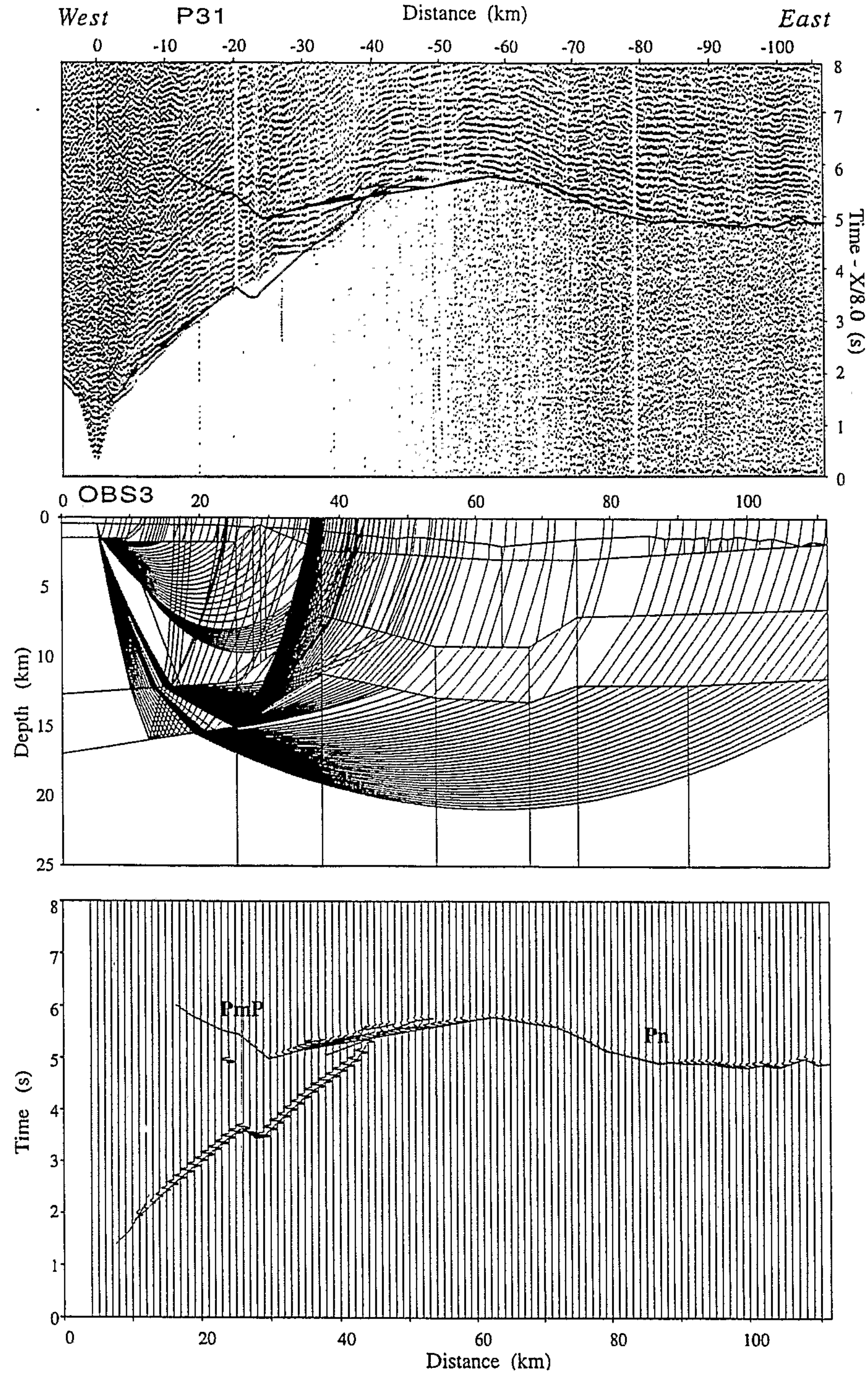

Figure 6. Record section, ray-tracing model and synthetic seismograms along profile P31 running eastwards from OBS3 to OBS1; data recorded on OBS3.

by Courtney & Recq (1986) from the interpretation of heat-flow data, and by Marks & Sandwell 1991) from analysis of geoid versus topography (G/T) for oceanic plateaux.

FREE-AIR GRAVITY MODELLING OVER THE CROZET BANK

A subgrid of the free-air anomaly values over the Crozet Bank (Fig. 7) was extracted from the global free-air anomaly grid derived by Smith & Sandwell (1995) from GEOSAT and ERS-1 altimeter measurements. The main structural units, described above, stand out in Fig. 7. The Indivat Basin is underlined by a relative gravity low running roughly N–S between the two groups of islands. Moreover, the extent of the relative gravity high near OBS3, as defined for example by the 80 mgal contour, confirms that the high-velocity structure is deeply rooted.

The contribution of the crustal structure to the free-air anomaly is computed from the seismic model derived above (Talwani & Ewing 1959). A 2-D gravity model is preferred

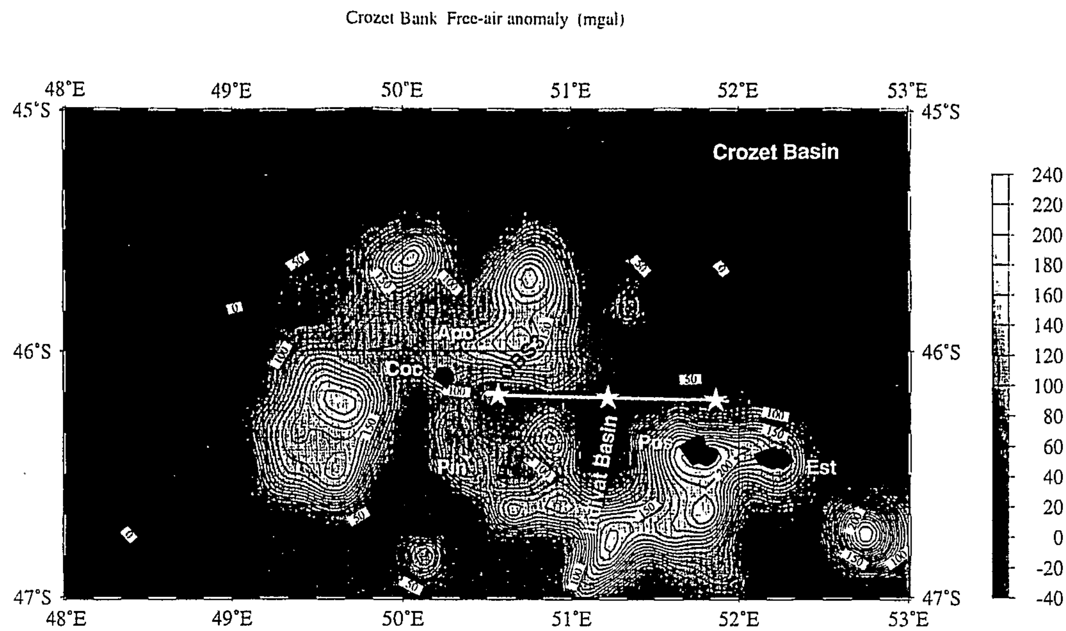

Figure 7. Free-air gravity anomaly contoured from the gridded data set derived by Smith & Sandwell (1995) from altimetry data. Anomalies are contoured every 10 mgal. The 50 mgal contours are labelled.

(Fig. 8), as the interpretation of seismic data only provides control on the crustal structure below the OBS profile. The density structure is therefore assumed to be cylindrical in a direction perpendicular to the refraction profile (for example made of prisms of infinite length in a direction normal to the profile azimuth). Densities are inferred from seismic velocities using the Ludwig *et al.* (1970) curve for the sediments and layer 2A and the relation proposed by Carlson & Herrick 1990) below layer 2A (Fig. 8). Synthetic 'observed' free-air anomaly values are interpolated along the refraction line from the subgrid shown in Fig. 7.

The results of gravity modelling are displayed in Fig. 9. Consistent with expectations, the seismic crustal model generates gravity anomalies of shorter wavelengths than the satellite-derived data. It achieves a good fit to the observed free-air gravity values at longer wavelengths, especially in the case of the observed relative gravity low centred on the Indivat Basin (Fig. 9). The standard deviation of the difference between the observed and modelled free-air anomaly profiles is about 7.7 mgal. We believe that combining a 3-D seafloor topography and a 2-D crustal density structure would have run the risk of introducing unnecessary complexities in the interpretation (as density is only constrained under the OBS profile itself). Indeed, the overall good quality of the fit between the observed and modelled free-air anomaly over most of the profile shows a minor influence of 3-D effects and therefore supports a

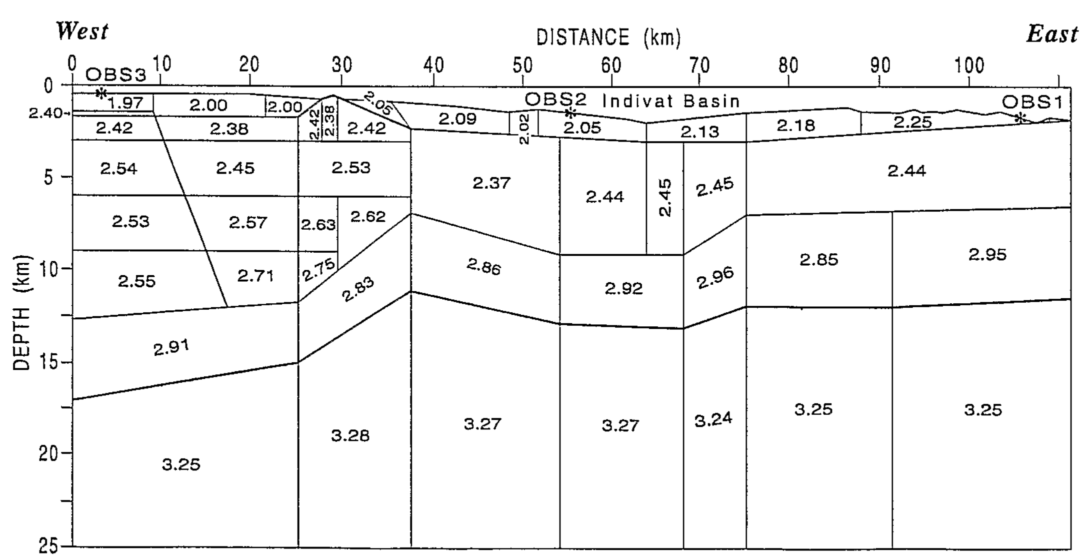

Figure 8. Density model used for 2-D gravity modelling along the KeOBS8 profile. Seismic velocities in Fig. 3 are converted into densities using the Ludwig *et al.* (1970) curve for the sediments and layer 2A and the relation proposed by Carlson & Herrick (1990) below layer 2A.

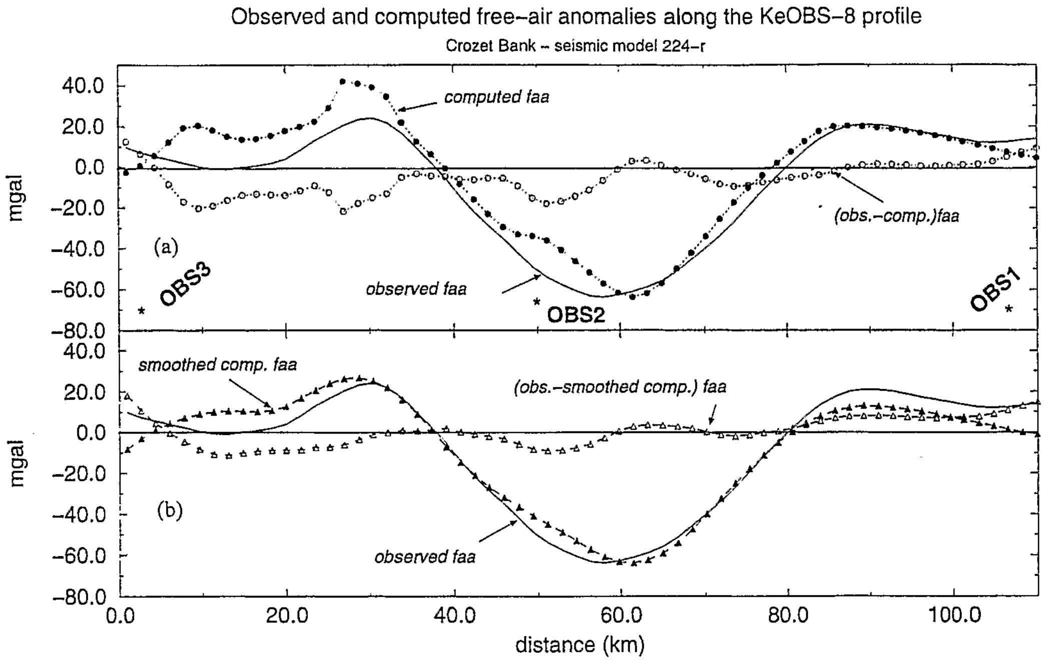

Figure 9. 2-D free-air gravity model along the KeOBS-8 seismic profile. (a) The observed free-air anomaly profile, shown as a thin continuous line, has been extracted from the grid computed by Smith & Sandwell (1995) from satellite altimeter measurements. Black dots indicate the free-air values (in mgal) produced by the crustal model shown in Fig. 3. Open circles show the difference between observed and computed figures. (b) Same as (a), but the computed profile is bandpass filtered to comply with the reliable wavelengths of the satellite-derived profile. Black triangles indicate the filtered free-air values (in mgal) produced by the model. Open circles show the difference between observed and computed figures, after filtering of the latter.

general cylindrical structure of the Indivat Basin at depth. This further confirms that this N–S-trending feature is a major structural boundary (Fig. 9a).

Some of the misfit could come from the shorter wavelengths, which, as mentioned above, are not resolved by satellite observations. Fig. 9(b) shows the computed free-air anomalies after a moving-window low-pass filtering was performed in order to compare them to altimetry-derived data, whose amplitudes are strongly degraded for smaller wavelengths. The standard deviation between model anomalies after filtering has been performed and observed anomalies is slightly lowered (6.9 mgal). The fit between computed and observed anomalies can be considered to be quite good for most of the profile sections.

The fit between the model and the observations is poorer along some portions of the profile, especially over the high-velocity structure. Part of this misfit could be due to the 3-D geometry of density anomalies of probable limited horizontal radius, related to the high-velocity structure. This can be clearly seen when looking at the better fit between the 'smoothed computed faa' and the observed gravity profile in Fig. 9(b). This fit is notably improved at shorter wavelengths as compared to the fit shown in Fig. 9(a), an indication that the 2-D model probably does not neglect large-scale 3-D crustal density heterogeneities under the Crozet Bank. Finally, the somewhat poorer fit is restricted to the westerly section of the profile, where the cylindrical model more strongly departs from the more 3-D geometry of the observed data (Fig. 7). In this section, we consider that introducing a 3-D seafloor topography in the density model would have resulted in little improvement to the computed free-air anomaly, as no 3-D bathymetric features stand out in the vicinity of the profile's western end (Fig. 2), while 3-D free-air anomalies appear in Fig. 7 in the same area.

CONCLUSIONS

Although only three OBSs were used for this experiment, the interpretation of a 110 km long refraction line, supplemented by the analysis of 2-D gravity modelling, has produced the following first results on the crustal structure of the Crozet Bank.

(1) Crustal seismic velocities below the Crozet Bank are quite similar to those observed below the Kerguelen–Heard Plateau (Charvis *et al.* 1995). However, the relative thicknesses of crustal layers differ considerably under these two features. The upper part of layer 2, up to 4 km thick, is made up of highly altered basalts, as shown by dredgings on the slope of the Crozet Bank during the cruise MDFLUX-INDIVAT. Under the Kerguelen–Heard Plateau, this layer is much thinner (1–2 km according to Recq *et al.* 1983; Charvis 1984; Recq & Charvis 1986; Recq *et al.* 1990; Charvis *et al.* 1995).

(2) Our experiment confirms that the two different provinces constituting the Crozet Plateau, the Del Cano Rise and the Crozet Bank differ in structure and probably, therefore, in origin. Below the Del Cano Rise, the 17 km thick crust, which was emplaced near the active Southwest Indian Ridge (Goslin *et al.* 1981) complies with the local Airy equilibrium of this structure (Recq & Goslin 1981; Goslin & Diament 1987). The Crozet Bank, covered by 0–2 km of water, is underlain by a crust whose thickness varies from 16.5 to 10 km. Crustal velocities correspond to densities which preclude an Airy-type isostatic response of the crust and lithosphere to the load of

the Crozet Bank. Our interpretation thus corroborates Courtney & Recq's (1986) hypothesis of a dynamic support for this feature, provided by an upper-mantle thermal anomaly. The Crozet Bank is thus a truly intraplate feature, recently emplaced by a mantle hotspot over old oceanic lithosphere.

(3) The Indivat Basin corresponds to a major crustal discontinuity within the Crozet Bank itself: to the west, towards the Del Cano Rise, a thick crust with higher upper-crust velocities is found. According to Chevallier & Nougier (1981), intense recent magmatism has lead to the build-up of the Ile aux Cochons and, probably, of the submarine high-velocity feature which was revealed by our experiment. This intense magmatism could have contributed to the enhanced thickness of upper layer 2 abutting the Del Cano Rise by interlayering volcanic material within sediments. Conversely, the limited volume of recent (less than 0.5 Ma) magmatism on the Ile de l'Est (Chevallier & Nougier 1981) correlates well with the thinner layer 2 at the eastern end of the profile.

The correlation between the gravity anomalies computed along a 2-D model profile and the observed values further confirms that the Indivat Basin is the structural boundary which separates the Crozet Bank from the adjacent Del Cano Rise.

(4) The submarine high-velocity structure, revealed by the KeOBS8 line, is deep-rooted into the crust and associated with a mantle 'dyke'. Speculation about a volcano is reasonable, even if it is not formally required. This intrusive feature exerts a major influence on the structure of the Crozet Bank. Our experiment demonstrates a marked elevation of the Moho beneath the volcano. A similar structure is observed below the Krafla Volcano in Iceland (Staples *et al.* 1997; Brandsdottir *et al.* 1997). Unlike the Hawaiian and Marquesas Islands, underplated mantle material is not found below the lower crust of the Crozet Bank, especially below the submarine volcano. The uplift of the lower crust below the western flank of the Indivat Basin is probably related to this intrusion, while the general eastward thinning of the crust is probably a consequence of the general dynamic uplift of the Crozet Bank.

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