

Lithospheric response to volcanic loading in the Southern Cook Islands

Kurt Lambeck

Research School of Earth Sciences, Australian National University, Canberra, A.C.T. 2600 (Australia)

Received September 10, 1980

Revised version accepted July 9, 1981

The response of the ocean lithosphere to volcanic island loading has been investigated for the Southern Cook Islands of Rarotonga, Mangaia and the Aitutaki-Mauke group using (1) estimates of the change—with distance from the load—of uplift of islands located on the flexure-created arch, (2) geoid height perturbations as observed with the GEOS 3 and SEASAT radar altimeters, and (3) bathymetric evidence for the moat and arch about the island loads. The effective flexural rigidity for Rarotonga is found to be higher than for the older and nearby islands of Mangaia and Aitutaki. This, together with an earlier result for Tahiti [1], suggests that stress relaxation of the lithosphere has occurred and that this layer behaves as a viscoelastic material with an effective viscosity of about $5-10 \times 10^{25}$ poise. The observations for the flexural rigidity do not indicate any dependence on the age of the lithosphere at the time the loading occurred. This suggests that there are no significant differences in lithospheric properties for plates in the age range of $6-7 \times 10^7$ years. Altimeter passes to the east of Mauke indicate that there may be an uncharted submarine extension of the Aitutaki-Mauke chain.

1. Introduction

The response of the ocean crust and lithosphere to surface loading is of considerable interest in understanding the long-term rheology of these layers which in turn has significant consequences on the boundary conditions appropriate for mantle convection and plate tectonic models. Loading by seamounts is potentially of great interest because seamounts of widely different ages occur sufficiently frequently to anticipate that the dependence of the isostatic response on age of the load, on the age of the lithosphere and on the magnitude of the stress differences may ultimately be unscrambled [2,3].

There remain however considerable uncertainties in the models of this response and the available observational data are still not very satisfactory. These observations include gravity or geoid height anomalies over the seamount, the moat and

arch geometry about the load, and past or present vertical motions of the load itself or of nearby islands. On their own, any one observation does not constrain the response model particularly well but taken together they appear to provide some useful constraints. For example, I have previously used geoid height anomalies, the arch-moat geometry, and change in island uplift with distance from the load, to determine the ocean lithospheric response to the Tahiti volcanic load [1]. A similar approach is used here for the neighbouring Southern Cook Islands.

2. The Southern Cook Islands

The Southern Cook Islands comprise a group of islands extending over about 3° in longitude, from Aitutaki in the west to Mauke in the east. To the south lie the islands of Rarotonga and Mangaia.

Wood and Hay [4] have given a detailed geological description of the group. The tectonic and volcanic history of the islands is complex and does not fit obviously into the pattern of other Pacific island

chains, whereby the islands are formed over a stationary hot spot as the lithosphere moves over the mantle [5].

The islands in the group (Fig. 1) consist of

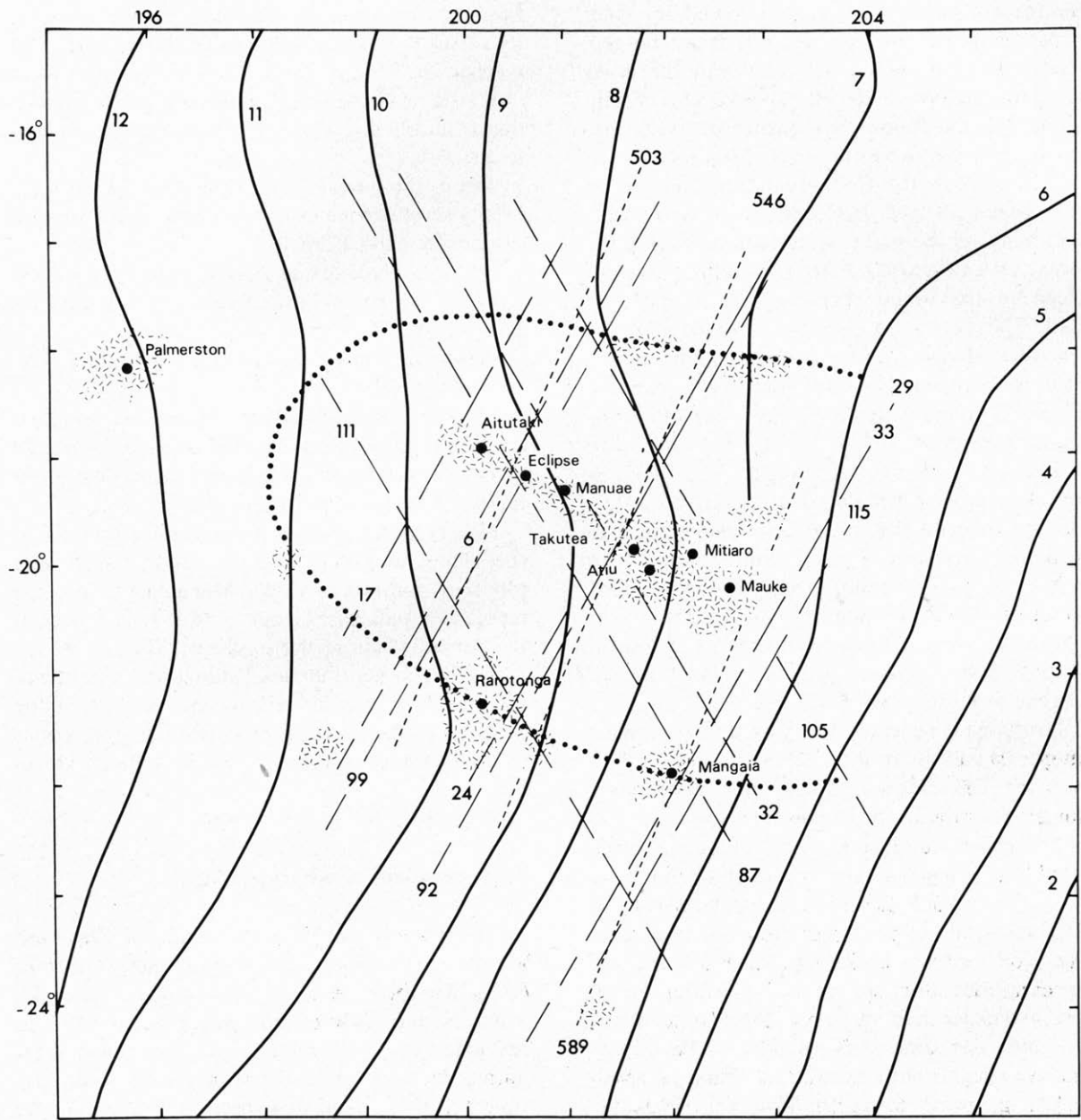


Fig. 1. Location map of the Southern Cook Islands. Bathymetry above 2400 fathoms is shaded. The location of the arch suggested by Summerhayes [10] is indicated by the dotted line. The geoid heights based on the GEOS 3 altimeter data are shown by solid lines and are in metres. Subsattelite tracks of passes near the islands are shown by dashed lines. The track numbers as referred to in the text are given in parenthesis.

several types. Rarotonga is a high mountainous volcanic island; Mangaia, Mauke, Mitiaro and Ataiu are raised coral islands with well-defined makatea and volcanic cores; Manuae is an atoll, Aitutaki is a near atoll with a volcanic core and Takutea is a sand-cay on a coral formation. Their history is varied and complex but from the geological survey of Wood and Hay and the K-Ar dating of Dalrymple et al. [6] and Turner and Jarrard [7], the following sequence of events appear to have occurred since early Miocene.

Volcanism in the Cook islands is thought to have begun in the Paleocene with submarine mountains being formed up to sealevel during the Eocene and Oligocene. After this initial period of volcanism the islands became extinct and subsided. The oldest documented volcanism in the area is for Mangaia where alkali basalts have been dated at between 21.4 and 16.6 Ma ago [6,7]. Volcanism on the Aitutaki-Mauke group of islands ranged from prior to 11.6 Ma ago for Mitiaro to 5.1 Ma ago for Mauke. The oldest lava flows on Aitutaki predate 6.9 Ma ago but this island differs from the other in that it has also been subject to much more recent volcanism, from about 1.8 to 0.7 Ma ago [6,7]. The more recent flows, however, consist of olivine nephelinites and do not appear to have been very extensive. The nearby islands of Manuae and Takutea give no indications of Miocene or more recent volcanism.

Rarotonga is generally recognized to be a much younger island than the others, having both a distinctly different morphology and persistently younger K-Ar dates for its volcanic flows. According to Turner and Jarrard, most of the subaerial part of Rarotonga was constructed of alkali basalts prior to about 3.5–1.8 Ma ago, and this agrees with Dalrymple et al's observation of two main periods of near sea level eruption at 1.8 and 1.2 Ma ago, eruptions that appear to have been much more extensive than those on Aitutaki at about this time. The time of formation of Rarotonga coincides roughly with the uplift of Mangaia, about 2–3 Ma ago, while the uplifts of the other makatea islands to the north are believed to be post-Miocene and pre-Pleistocene. Following McNutt and Menard [8] this suggests that the formation of Rarotonga may have been responsible for the up-

lift of the surrounding islands although McNutt and Menard spoil their argument by considering that the observed uplifts are also due to the additional loads of Aitutaki (whose main period of volcanism significantly predates the uplift) and of Manuae (for which no evidence of volcanic activity exists at all since at least the Miocene [4]). The mechanism for this uplift is via the flexure theory of elastic or viscoelastic lithospheric plates, overlying a fluid-like medium, and subjected to surface loads.

From the above brief summary of tectonic events several conclusions relevant to the present discussion, can be drawn:

(1) The only volcanic load that can have caused uplift of the islands in the Southern Cook group is Rarotonga.

(2) The makatea elevations may provide a measure of the uplift.

(3) The atoll and makatea islands are considerably older than Rarotonga and as such the nature of compensation of these island loads may differ from that of the Rarotonga load.

(4) The K-Ar dates of the volcanic eruptions on the islands do not fit into the hot spot or mantle plume hypothesis and only Mangaia fits into the recognized pattern by supposing that it is part of the Austral chain to the southeast [7].

(5) The rather complex loading and subsidence histories of the older islands may not make the flexural models very appropriate representations of the isostatic response of the lithosphere to loading.

3. Observations of differential uplift

The islands of Mangaia, Atiu, Mitiaro and Mauke have makatea that vary in elevation from 70 to 10 m above sea level. The islands of Aitutaki, Manuae and Takutea do not possess elevated makatea although the first island does have several marine terraces up to 120 m above sea level. Presumably this island has been uplifted like the makatea islands but the subsequent volcanism penetrated and destroyed the uplifted reef of which only a few traces remain in the form of coral limestone fragments in the basalt flows. Manuae

and Takutea are low islands (10 m and 6 m maximum elevation above sea level respectively) neither of which show evidence of any volcanism although they probably rest on volcanic foundations at depth. Presumably these islands are identical in origin to the others in this group (apart from the late stage of volcanism seen on Aitutaki). The absence of elevated limestone reefs may be a consequence of their small size so that they have eroded away, or it may be a consequence of these islands having originally been sufficiently far below sea level for significant reefs not to have formed prior to uplift. Table 1 summarizes (1) the observations of uplift with respect to present sea level, and (2) the differential uplift used in this paper to constrain the flexure models.

In the presence of erosion, sea level variations and regional vertical motions, it is impossible to establish the absolute uplift of islands. I have earlier argued that the appropriate observable to test the flexure theory is not the uplift but the differential island uplift or the change in uplift with distance from the load [1]. This measure is independent of sea level changes and from regional, non-flexure-associated, uplifts although it does assume that erosion rates have been the same for all islands. The question as to what is the best geological indicator of total uplift remains unresolved. McNutt and Menard [8] suggest that the island elevation provides a good indicator but their results are quite inconsistent with the evidence [4,9]. For Mangaia, McNutt and Menard have adopted the height of the makatea above sealevel as a measure of uplift. They give an uplift for Mitiaro of 27 m whereas the maximum island

elevation is only about 12 m and does not exceed that of the makatea. their 30 m uplift for Mauke resembles the maximum island elevation but the makatea uplift is only 18 m at most, while that of Atiu is between 20 and 30 m. In the following I have adopted the makatea elevations as a measure of relative uplift, assuming that the amounts of erosion have been roughly similar for all the makatea islands. Wood and Hay pointed out that for these islands a correlation exists between the subaerial surface area and makatea elevation, with the larger islands possessing the higher coral platforms. On the other hand the same correlation can be seen between the reef uplift and the distance from Rarotonga (see Table 1).

4. Regional bathymetry

Bathymetry maps for the area are given by Summerhayes [10] and by Mammerickx et al. [11]. The former paper includes detailed bathymetric surveys of all the islands in the group and the maps have also been published by the New Zealand Institute of Oceanography. The Cook Islands lie in a relatively shallow part of the Pacific Basin with oceanic depths rarely exceeding 5 km but to the north and south the depths gradually increase to about 5.5 km. Summerhayes suggested that the islands are surrounded by moat and arch structures. In particular, around the Aitutaki-Mauke group an arch is suggested at a distance varying between 100 and 160 km to the north and at about 190 km to the south (Fig. 1). Rarotonga and Mangaia lie on this proposed arch. The moat

TABLE 1

Summary of maximum island and makatea elevations for those islands that may have been uplifted by the Rarotonga load. The differential uplifts dw/dr are given at the distances r midway between successive uplifted islands

Island	Distance from Rarotonga (km)	Maximum elevation * (km)	Elevation of makatea * (m)	r (km)	dw/dr ($\times 10^{-3}$)	Mean radius of subaerial surface (km *)	McNutt and Menard's uplift (m)
Mangaia	205	169	70			4.0	70
Atiu	222	70	20-30	214	0.19-0.24	3.1	20
Mitiaro	264	12	12	234	0.03-0.08	2.8	27
Mauke	281	30	18	273	0.02	2.5	30

* From Wood and Hay [4].

depth is estimated to be about 150 m. Summerhayes also suggests that there are moat-like depressions about Rarotonga and Mangaia but the chart of Mammerickx et al. does not confirm this, probably because the resolution of the bathymetry is inadequate for this and the moat depths are likely to be only of the order of 100 m or less.

Much of the ocean floor around the islands consists of smooth archipelagic aprons made up of consolidated sediments and basalts. The aprons extend up to 60 km from the islands, sloping imperceptibly into the deep ocean floor. Summerhayes has suggested that large-scale slumping of the flanks of the seamounts is the source of the debris which ultimately forms the bulk of the aprons.

5. Seismic data

Seismic studies of the crustal structure in the immediate vicinity of the Southern Cook Islands appear to be limited to the work by Hochstein [12] who reported seismic velocities of 4.05–4.35 km s⁻¹ down to a depth of 1.65 km beneath both Aitutaki and Rarotonga. Below this depth the velocity increases to about 6 km s⁻¹ although the slopes of Aitutaki were assumed to be of low-velocity (≈ 2.1 km s⁻¹) material of up to 1.5 km thickness. Hochstein's velocity model of Aitutaki is indicative of low-density material ($\rho \approx 2.35$ g cm⁻³) for the upper 1.65 km of the island overlying material of density of about 2.5–2.7 g cm⁻³ at a depth of about 9 km below sealevel. A density of 2.0 g cm⁻³ is appropriate for the flanks of the island where the velocities correspond to those of the ocean layer 1. The densities for the upper parts of the island agree with those deduced by Robertson [13] from gravity although the latter finds evidence for a higher-density material (≈ 2.85 g cm⁻³) central volcanic core. Rarotonga, being a more recent island that has been subject to extensive volcanism could be expected to have a higher average density immediately below the surface than Aitutaki but Robertson [14] concludes that it does not differ greatly from that of the other Cook islands.

The velocity model of the crust to the south of

Aitutaki comprises an ocean layer 2, of velocity 5.7 km s⁻¹ and thickness of about 5 km, although these depths and velocities are poorly constrained with the available observations. The mantle velocity was found to be 7.7 km s⁻¹ with the Moho dipping below the island by about 1.6 km where the mantle velocities were observed to be 8.0–8.2 km s⁻¹. The interface of layers 2 and 3 was not seen below the islands and the velocity model immediately below the island remains speculative. The abnormally low mantle velocities suggest that what is seen is not the upper mantle but the high-velocity "basal layer" of the crust reported in some models by Christensen and Salisbury [15] and Woollard [16]. Woollard, in his survey of crustal refraction results for the Pacific Basin, gives two sections just to the north of the Southern Cook Islands (see his fig. 5), sections that indicate a layer 2 of 1.5 km thickness with velocity in the range 5.1–5.8 km s⁻¹ and a layer 3 of about 5 km thickness with velocity in the range of 6.5–6.8 km s⁻¹. Woollard's mantle velocities are "normal" at about 8.2 km s⁻¹. These results are very similar to the average ocean crust model of Raitt [17]. The adopted crustal model is (1) an ocean layer 2 of 2.5 km thickness and with density 2.5 g cm⁻³ and (2) an ocean layer 3 of thickness 5.0 km and density 2.9 g cm⁻³. The mantle density is taken as 3.35 g cm⁻³.

6. Geoid height anomalies

Radar altimeter measurements have been made over the Cook Islands with both the GEOS 3 and SEASAT spacecraft. The former data has been used to compute the geometric form of the mean sea surface for a region bounded by 190° and 210° east longitudes and by 10° and 30° south latitudes. This surface, assumed here to correspond to the geoid, is dominated by the southeast to north-west slope culminating in the geoid height over the Fiji plateau. Superimposed upon this trend are considerably shorter wavelength variations and it is these that are of interest in the present study. Model calculations show that the wavelength of the sought anomalies over the islands do not exceed a few hundred kilometres. The long-

wavelength information has been removed by low-pass filtering the regional geoid, based on all GEOS 3 altimeter data, such that anomalies of wavelength less than 300 km are removed. This geoid is then used as the reference with respect to which the observed GEOS 3 and SEASAT profiles are reduced. Fig. 1 illustrates the filtered geoid for the area about the Southern Cook Islands. Also illustrated are those ground tracks of GEOS 3 and SEASAT altimeter data that potentially contain information on the regional compensation, passes that lie within a few tens of kilometres from the centre of the islands.

The precision of the GEOS 3 high-intensity mode altimeter data is taken to be of the order of 20 cm [18]. The precision of the SEASAT altimeter data is taken to be 10 cm. Most, but not all, of the GEOS data has been processed at the Wallops Flight Center and tests have indicated that their processing yields essentially the same result if a spline function is fitted to the raw data with a standard deviation of 10 cm. A further comparison of GEOS 3 and SEASAT passes for nearly similar ground track segments confirms the estimate of 20 cm (cf. profiles 6, 503 and profiles 3, 589 of Fig. 2). All height measurements taken at points where the islands fall within the radar footprint have been rejected in the subsequent analysis.

For Mangaia, three descending passes (numbers 3, 29 and 589) pass within 10 km of the island. The GEOS pass 3 and SEASAT pass 589 give almost identical results in the vicinity of Mangaia (Fig. 2). These passes also contain a smaller anomaly due to Mauke. A fourth pass, at 25 km from Mangaia contains a much reduced anomaly. The ascending pass 87 lies 8 km from the load centre. Pass 32, at about 60 km from Mangaia contains no noticeable signal. There is a suggestion in all of the passes near Mangaia that the anomaly to the north is more negative, by about 20–30 cm, than at the same distance to the south (see passes 3, 29 and 589 in particular). Only a small part of this, about 5–10 cm, can be attributed to flexure under Mauke since, at these distances and for nominal flexural rigidities, the expected geoid anomaly is negative.

Several satellite passes lie near Aitutaki, descending passes 6 and 17 lie on either side of the

island while SEASAT pass 503 gives a very similar result to GEOS pass 6. Pass 99, about midway between Aitutaki and Manuae, lies over the small Eclipse seamount. Ascending pass 87 contains the most pronounced signal.

The closest pass to Rarotonga is 111 which lies at about 20 km from the centre of the load. The nearly orthogonal pass 24 is at a minimum distance of about 25 km. At these distances the expected anomaly is about 50% of what it would be over the centre of the island. Passes 92 and 111 have gaps in the record at rather crucial points near closest approach. The two descending passes 24 and 92 exhibit a behaviour similar to the Mangaia results in that the negative flank anomalies are more pronounced to the north of the island than to the south. Pass 111 indicates a small anomaly 70 km to the southwest of Rarotonga and this corresponds to a small bathymetric high on the map of Mammerickx et al. [11]. Pass 24 indicates an asymmetry about the point of closest approach. This is indicative of a small bathymetric high about 40 km south of Rarotonga but this is not seen in the maps of Mammerickx et al. [11] or of Summerhayes [10]. The altimeter data is generally unsatisfactory for Rarotonga.

Other passes lie near Mauke (3, 29, 105, 115, 589) and near Takutea (24, 32, 92, 546). Of the former group, pass 105 contains a small anomaly about 65 km southeast of Mauke which does not have a bathymetric counterpart in the above cited charts. Evidence for this geoid anomaly is also seen in passes 3, 29, 115, and 589 in that the maximum anomalies in these passes do not occur at the point of closest approach to Mauke. Taken together these altimeter results point to an uncharted seamount, at 20°45'S, 203°0'E, of a magnitude that is comparable to Eclipse seamount (see Lambeck and Coleman [19] for a discussion of this and other anomalies in the Cook Island region). This suspected seamount illustrates well the uncertainties in the bathymetry of the region in areas beyond those reported by Summerhayes.

While the relative heights are well known, the absolute heights above the reference surface may be in error due mainly to very long-wavelength orbital uncertainties. Such differences have been accounted for in the geoid computations by impos-

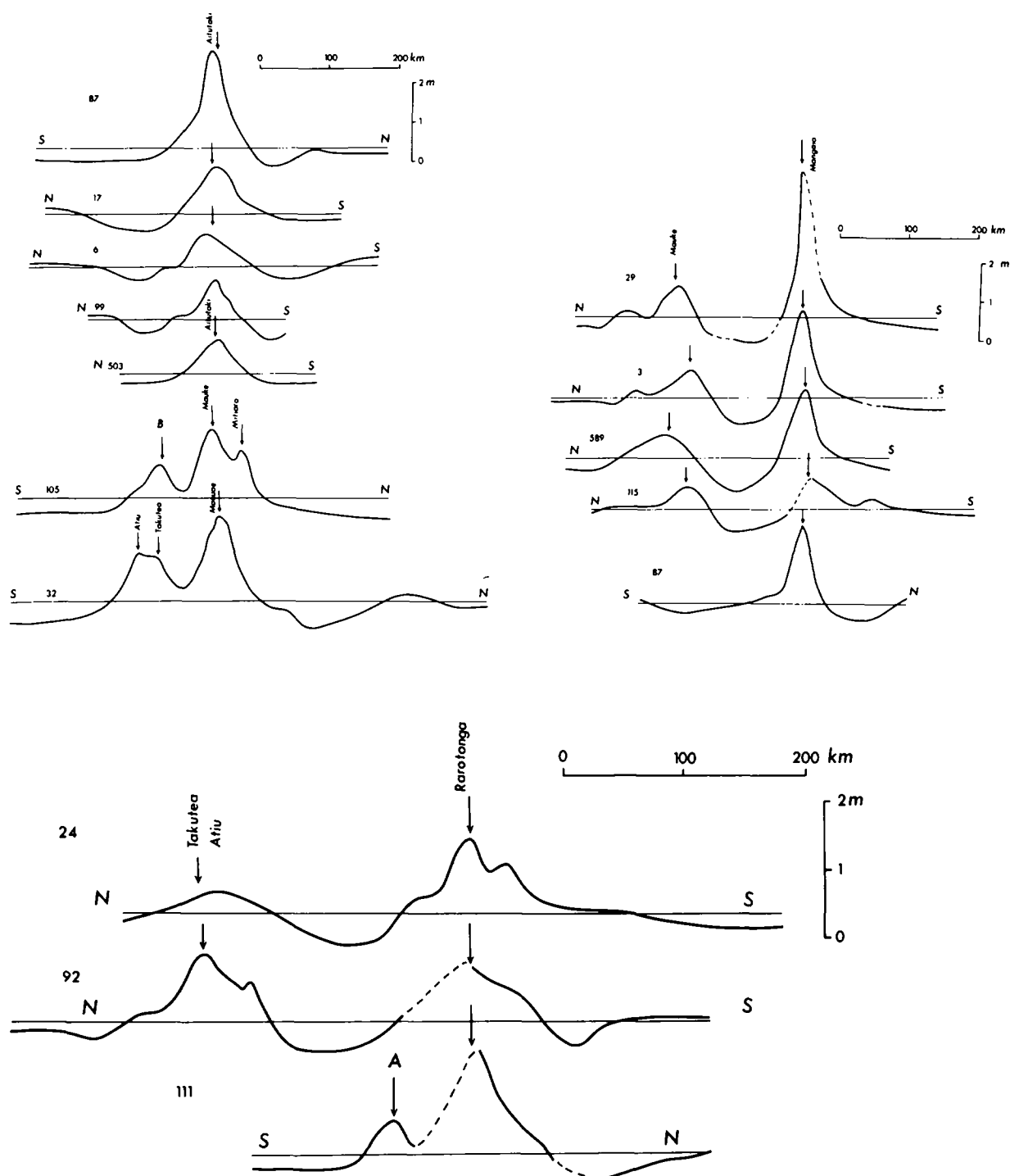


Fig. 2. Residual geoid height profiles near the island loads. The heights are with respect to a band-pass filtered geoid such as to remove all wavelengths greater than about 300 km. The residual profiles have been further low-pass filtered to reduce the measurement noise by fitting a spline function to the data with a 10-cm standard deviation.

ing the condition that at all cross-over points of altimeter profiles the height residual must be identical for the ascending and descending passes, and any discrepancies have been distributed by a least-squares algorithm [22]. These discrepancies have not, however, been subsequently applied to the individual profiles illustrated in Fig. 2 and the absolute heights and any tilts of these passes are unknown. Hence, to reduce all profiles about a given seamount to the same origin, the individual profiles have been translated in height relative to each other so as to give a mean zero height in the distance range of 100–200 km from the centre of the load. The computed profiles have been matched with the observed profile by requiring that the average discrepancy is zero over the above distance range.

For isolated seamounts the several passes containing a geoid anomaly have been combined into a single record by plotting the geoid heights as a function of distance from the centre of the load. Where a pass lies close to several islands, such as pass 32, the contributions from all loads have been computed and compared with the single pass.

The accuracy of these compound observed profiles is a function of several factors including measurement noise and any departures of the load from axial symmetry. The latter tend to increase with distance from the load axis. These errors have been estimated from the scatter of points making up each composite profile and the uncertainty estimate is given for each such profile (e.g. Fig. 4).

7. Lithospheric flexure model

The model used here for the flexure of the lithosphere due to a volcanic load has been discussed in detail by Lambeck and Nakiboglu [1–3]. The same formulation, in terms of an effective flexural rigidity as used in the last of these papers, is also used here. The loads are represented by a number of superimposed co-axial discs, a core whose density contrast with the remainder of the load is taken as 0.5 g cm^{-3} and a variable degree of fill-in of the flexure-created moat. The moat, formed by the downward deflection $w(r)$ of the crust subjected to the seamount load, is assumed

to have been filled in to a depth $\beta w(r)$ that is proportional to the total deflection.

The load geometries have been determined from the detailed bathymetry surveys reported by Summerhayes [10] and the axes of symmetry of the loads have been fitted so as to give a best-fit of the model to the bathymetry. The assumption of axial symmetry, using co-axial disc loads whose total volume equals that of the observed volume, is adequate for the deflection calculations since the islands generally exhibit good symmetry and because the deflections are relatively insensitive to the detail in the load. The contribution to the geoid anomaly of the deflected surfaces of density contrast is also adequately represented by the symmetric loads. Less satisfactory is the contribution of the above-seafloor load to the geoid anomaly since this mass is quite close to the ocean surface. For the more important loads considered here, Mangaia, Aitutaki and Rarotonga, the assumption of axial symmetry does not appear to be very important if profiles at different azimuths are combined in the manner discussed above. Departures from symmetry in these cases do not introduce errors that exceed the measurement noise, “bathymetric” noise and the “geological” noise of lateral density variations in the upper oceanic crustal layers.

The unknowns in the flexure model include the average density of the load, the fill-in parameter β and the rheology of the lithosphere, at present conveniently summarized by the apparent or effective flexural rigidity D . The dependence on these parameters of the moat and arch geometry, of the differential uplift and of the perturbations in gravity and geoid height have been discussed elsewhere [1,3]. Briefly, the effect of increasing the degree of fill-in of the moat is to increase the distance to the arch, to increase the magnitude of the arch uplift, to decrease the moat depth, and to decrease the gravity perturbation and the geoid height anomaly. An increase in load density increases the moat depth, the uplift of the arch and the magnitude of the gravity and geoid anomalies while the wavelengths of the moat and geoid perturbation are only marginally dependent on the density. Increasing the flexural rigidity results in an increase in the distance to the moat and in the magnitude of the

geoid anomaly, but the maximum uplift of the arch is decreased. Following the seismic and gravity evidence we adopt $\rho = 2.5 \text{ g cm}^{-3}$ and 3.0 g cm^{-3} for the central core. From the bathymetric evidence any moat will be shallow and we adopt $\beta = 0.75$. This results in a moat depth that is generally less than 100 m for the range of D found to be appropriate here. Also it is noted that for these loads the geoid heights are not very dependent on β once the degree of fill-in exceeds about 75%. Hence for all subsequent models and unless otherwise stated, $\rho = 2.5 \text{ g cm}^{-3}$ and $\beta = 0.75$. The remaining parameter to be solved for is the effective flexural rigidity. The adoption of ρ and β values is justified here because we will be seeking a dependence of D on age of the load and, in the first approximation, any error in ρ or β will affect dD/dt only minimally.

For each island, the geoid height anomalies have been computed for a range of flexural rigidities and compared with the observed profile to obtain a best estimate of D . Because the observed profiles have been subject to some smoothing, the same filtering was also applied to the model results. The principal consequence of this is to reduce the maximum amplitude over the centre and to increase the anomaly over the immediate flanks of the load by about 5–10%. The crustal model discussed in section 5 has been used to compute the geoid anomalies.

As indicated in the introduction, the effective flexural rigidity may be expected to depend on several quantities. The stress-magnitude effect is unlikely to be important here since all island models result in very comparable stress-difference estimates. The age of the crust is also unimportant here since the variation in age is small. Hence the most likely dependence will be that on the age of the load if stress relaxation is important. From Lambeck and Nakiboglu [3], an approximate relation between the effective flexural rigidity D_{eff} and effective viscosity η of the lithosphere is given by:

$$\frac{D_{\text{eff}}}{D_{\text{el}}} = \frac{\alpha^4 e^{-\alpha^4 t / \tau(1 + \alpha^4)}}{(1 + \alpha^4) - e^{-\alpha^4 t / \tau(1 + \alpha^4)}} \quad (1)$$

where D_{el} is the elastic flexural rigidity for time $t = 0$, $\tau = \eta / \mu$ is the relaxation constant, μ being

the rigidity, and:

$$\alpha = \frac{A}{\pi} \left[\frac{(\rho_s - \rho)g}{D} \right]^{1/4}$$

where A is the radius of the load at its base, ρ_s is the density of the substratum, ρ = density of load and g is gravity. D_{el} is an unknown quantity unless the response of very young loads can be evaluated. However, this is not important here since the quantity of interest is the change in D_{eff} with time (i.e. with age of load). That is:

$$[D_{\text{eff}}(t - t_1) - D_{\text{eff}}(t_1)] / D_{\text{eff}}(t_1) \quad (2)$$

where $D_{\text{eff}}(t_1)$ is the effective flexural rigidity for the youngest load of age t_1 .

8. Results

Rarotonga. Fig. 3 illustrates the change in uplift dw/dr with distance across the peripheral bulge for a range of parameter values that result in an approximate agreement with the observations. Only approximate limits on the three unknown parameters ρ , β , D can be established from the uplift data. Values of $D \simeq 2 \times 10^{29}$ dyne-cm are excluded since the maximum dw/dr would occur at distances well beyond the observed position. Values of D much less than 10^{29} dyne-cm are also excluded since they result in an arch that lies much closer to the islands than suggested by the data. High load densities $\rho \gtrsim 2.8 \text{ g cm}^{-3}$ also have to be excluded since they require a low degree of fill-in, incompatible with the absence of a well-developed and deep-moat about Rarotonga. For $\rho = 2.5 \text{ g cm}^{-3}$, as suggested by the seismic data, acceptable solutions are $D = 10^{29}$ dyne-cm with $\beta = 0.85$ – 1.0 or slightly higher D with smaller β e.g. $D = 1.5 \times 10^{29}$ dyne-cm if $\beta = 0.75$.

The value of $D = 1$ – 2×10^{29} dyne-cm is close to that found by McNutt and Menard [8] but this agreement is more fortuitous than real since (1) they assumed that sealevel remained constant, (2) their estimates of uplift are inconsistent with observations, and (3) they consider that Aitutaki and Manuae contributed to the uplift.

The composite geoid anomaly profile based on the three satellite passes 24, 92 and 111 (Figs. 1

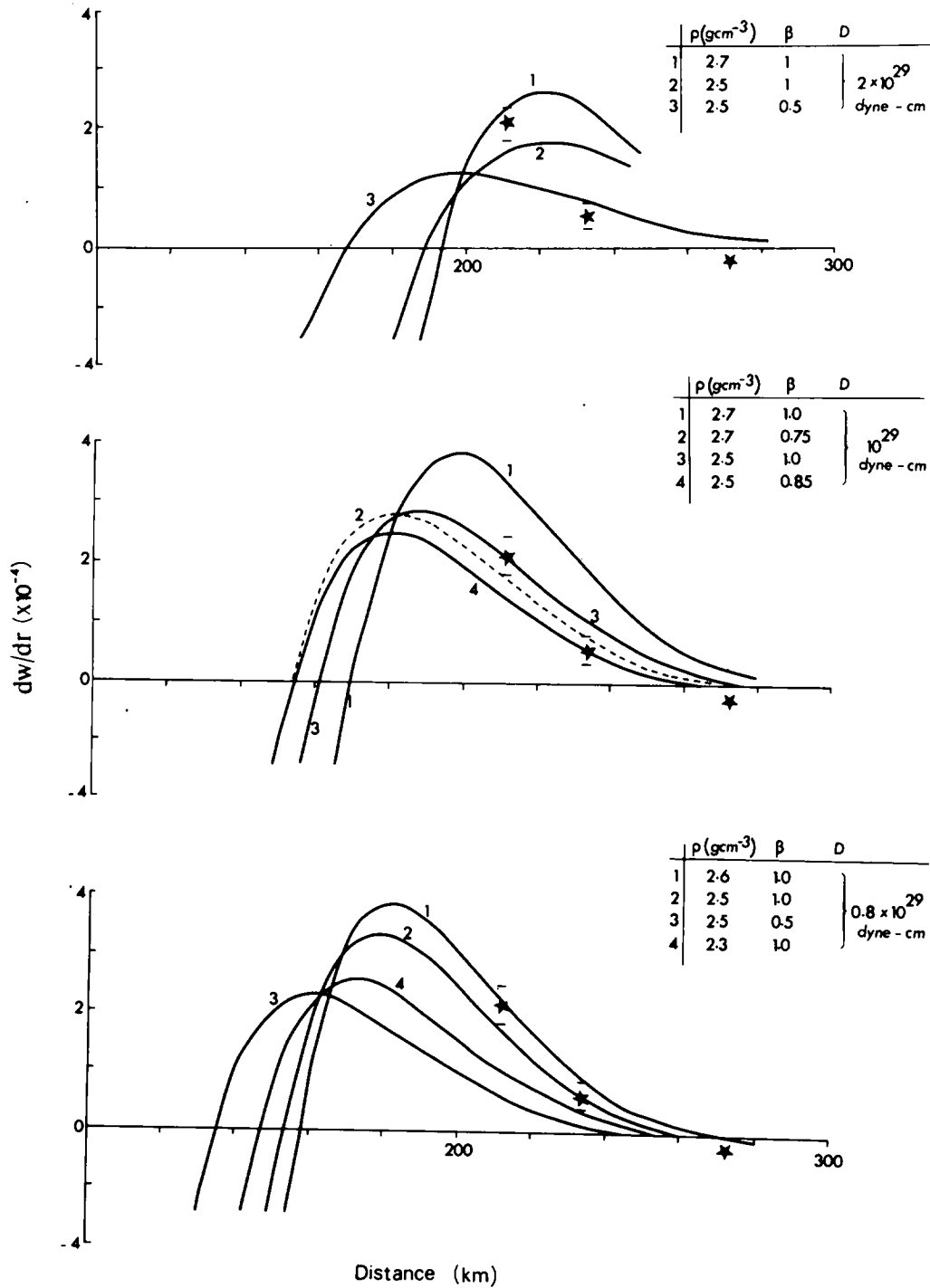
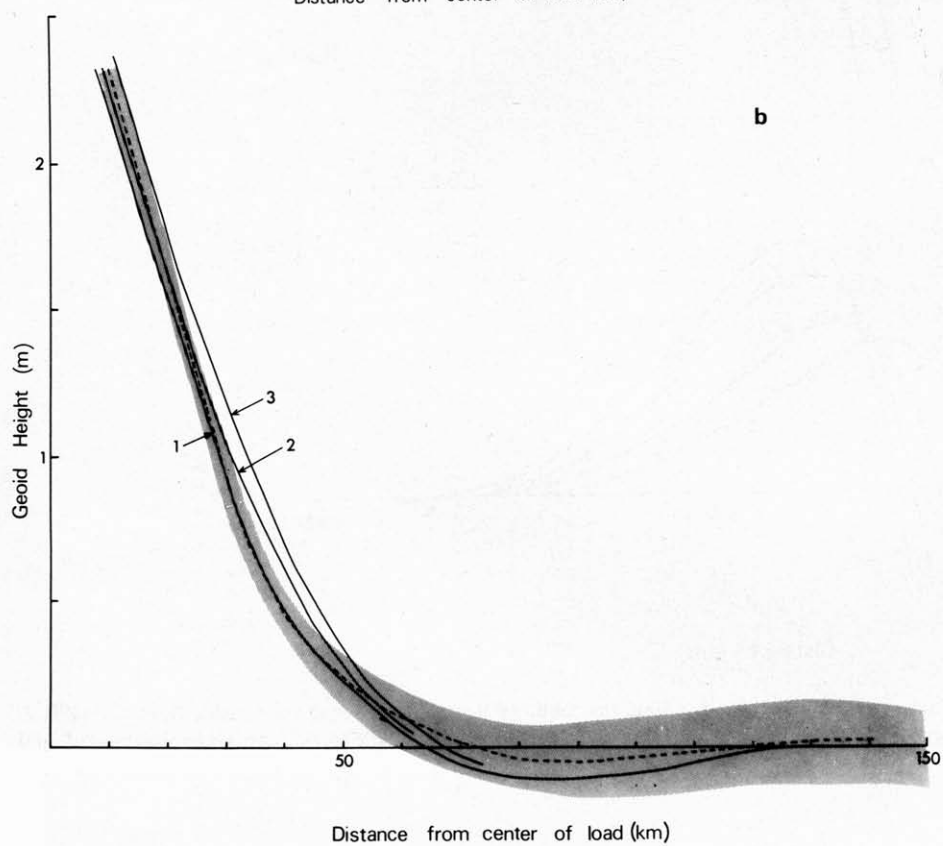
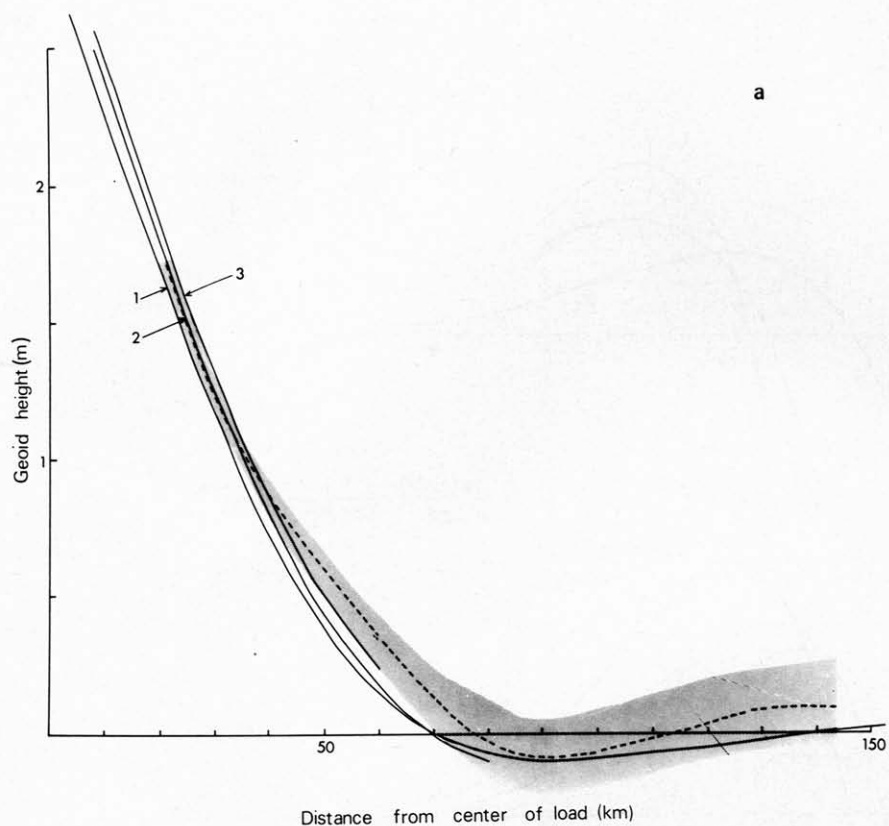


Fig. 3. Differential uplift dw/dr as a function of distance from the centre of the Rarotonga load for variable flexural rigidity D , density ρ and fill-in β . The observed values (see Table 1) are indicated by the star symbols and where different interpretations of uplift are possible the range of permissible values are shown by the error bars.



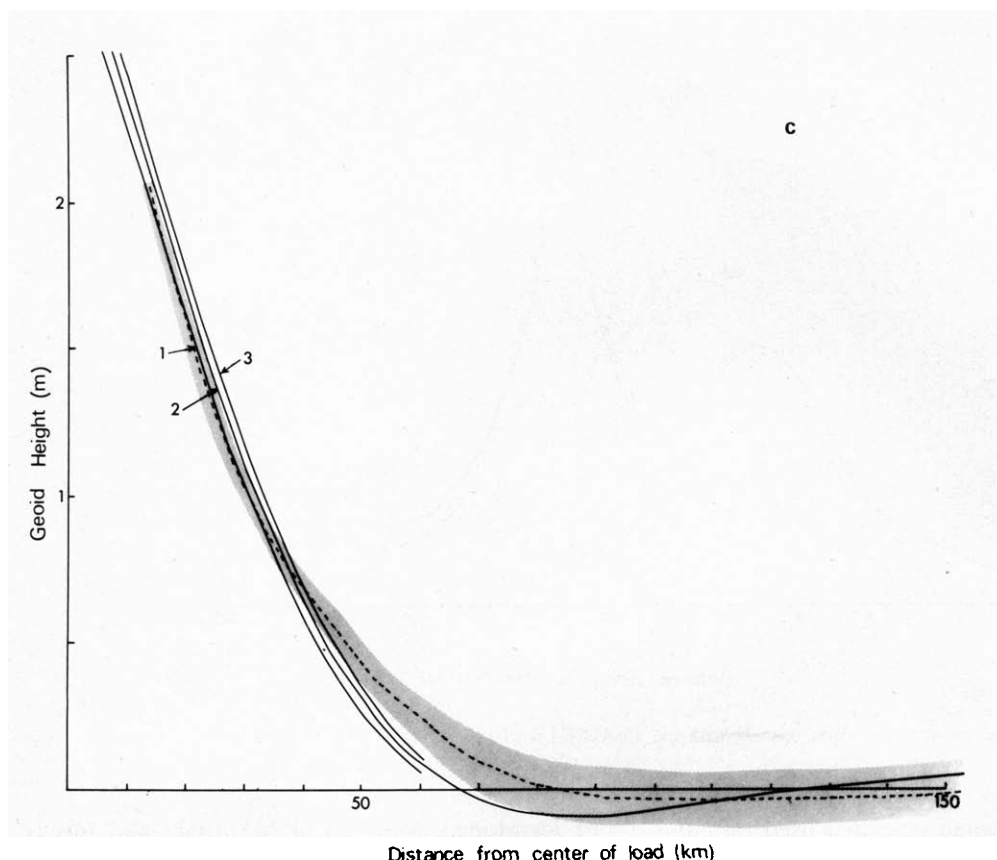


Fig. 4. The composite geoid anomaly (dashed line) and error limits (shaded area) compared with theoretical profiles for $\beta=0.75$, $\rho=2.5$ and variable D (in units of 10^{29} dyne-cm): (a) Rarotonga, (b) Mangaia, (c) Aitutaki.

and 2) is illustrated in Fig. 4a. The sections of the profiles to the north of Rarotonga have been corrected for the small contributions, less than 10 cm, coming from the Aitutaki-Mauke flexure. This composite profile is not very satisfactory since the point of nearest approach to the centre of the load is only about 25 km and at this distance the anomaly is already much attenuated compared with the anomaly that would be observed over the centre of the load. For $\rho = 2.5 \text{ g cm}^{-3}$ and $\beta = 0.75$ the comparison of model and observed profiles indicates that $10^{29} < D < 3 \times 10^{29}$ dyne-cm. As noted previously [1], the geoid anomalies are less sensitive to D than is the differential uplift of the outer arch. If β is increased to 1 then $D \approx 2 \times 10^{29}$ dyne-cm. Together with uplift data, $D \approx 1.5\text{--}2 \times 10^{29}$ dyne-cm for the standard ρ and β values.

Mangaia. The composite profile is illustrated in Fig. 4b and is based on the four GEOS passes (3, 29, 87, 115) and the SEASAT pass 589. The location of these passes, as well as the quality of the data, leads to a more satisfactory solution for D than was the case for Rarotonga.

Values of D greater than 2×10^{29} dyne-cm are excluded unless the load density is less than 2.3 g cm^{-3} while for standard ρ and β , $D \approx 0.5 \times 10^{29}$ dyne-cm. If $\beta = 1$ then, because the geoid anomaly decreases in amplitude with increasing β , the agreement is marginally better.

Aitutaki. The flexure model for Aitutaki is constrained by the GEOS passes, 6, 17, 87 and SEASAT pass 503 (Fig. 4c). An initial solution was first obtained using only the profiles 17 and

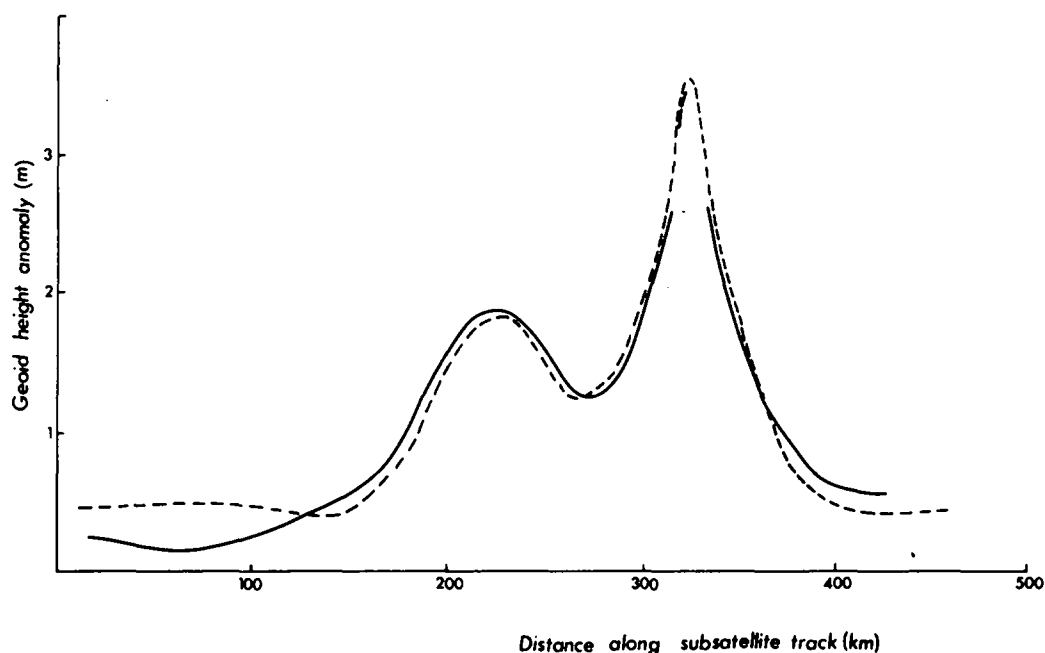


Fig. 5. Comparison of the geoid anomaly along pass 32 with the theoretical model based on $D = 1 \times 10^{29}$ dyne-cm.

87. The resulting D value was then used to compute corrections, mainly to passes 17 and 503, resulting from the Eclipse seamount and Manuae. The composite profile in Fig. 6 is the mean of the four corrected passes. Satisfactory agreement between model and observation results if $D \approx 1 \times 10^{29}$ dyne-cm. Such a value also describes well the geoid anomalies in the other passes near Manuae, Atiu and Takutea. This is particularly illustrated by profile 32 (Fig. 5) and the compensation mechanisms for these islands appears to be very similar to that of Aitutaki.

9. Discussion

Table 2 summarizes the results for the effective flexural rigidities all based on $\rho = 2.5 \text{ g cm}^{-3}$ and $\beta = 0.75$. All flexure models require relatively low D of $0.5\text{--}2 \times 10^{29}$ dyne-cm. This implies that, for $\mu \approx 3.2 \times 10^{11}$ cgs, the effective elastic thickness is in the range 8–13 km for Upper Cretaceous ocean lithosphere. The maximum deflections of the crust under the loads are of the order 1 km for the loads

of Rarotonga, Mangaia and Aitutaki and this is not an insignificant fraction of the elastic thickness. Hence the simple plate theory used here can be considerably in error, not so much for the estimates of deflection and thereby geoid anomaly, but mainly for estimates of the stress differences. In any event, while gravity and geoid heights are not particularly good indicators of deformation they are even less satisfactory indicators of the stress state and the actual stresses may be much less than predicted by the simple plate theory [2].

The analysis of the regional isostasy for Tahiti, using the same methods for analysing the uplift

TABLE 2

Summary of flexural rigidities and ages of loads

Island	D (10^{29} dyne-cm)	Age of load (10^6 years)
Rarotonga	1–2	35–1.8
Mangaia	<1	21–17
Aitutaki	≈ 1	7
Tahiti (Lambeck [1])	3	1.5–0.5

and geoid data as used in this paper, led to an estimate of the flexural rigidity of about 3×10^{29} dyne-cm for the same ρ and β values as adopted here. The difference in lithospheric ages for Tahiti and Rarotonga is of the order $10\text{--}15 \times 10^6$ years if seafloor spreading has occurred at an average rate of $6\text{--}7 \text{ cm yr}^{-1}$ in the intervening interval. This represents about 15–20% of the present age of the seafloor near Tahiti ($\approx 70 \times 10^6$ years). Thermal evolution calculations of the lithosphere would argue for a cooler and hence stronger plate under the islands whose lithospheric age at the time of loading is older. That is, the lithosphere under Rarotonga should have a higher D than under the Society Islands. That this is not seen suggests that the thermal evolution of the lithosphere is not important at times greater than about 6×10^7 years. Loads on younger lithospheres need to be analysed to determine at what point this does become significant.

Together, the Tahiti and Southern Cook results indicates a dependence of the effective flexural rigidity on the age of the load. The age of the Tahiti complex ranges from 1.5×10^6 years for the island of Moorea to about 0.4×10^6 years for the Taiarapu Peninsula. The result for Rarotonga, with an age span of $1.8\text{--}3.5 \times 10^6$ years yields $D = 1\text{--}2 \times 10^{29}$ dyne-cm. The difference in ages is small but may still be sufficient for significant stress relaxation to have occurred under the older load. The oldest major lava flows on Aitutaki, exceed 6.9×10^6 years and if these reflect the age of the load as a whole then the estimated D would be smaller than that for Rarotonga. That for Mangaia, whose age is about 20×10^6 years, would be smaller still. This is exactly the trend observed (Table 2) indicating that viscoelastic stress relaxation is not an insignificant factor in interpreting effective flexural rigidities. This is contrary to the conclusions reached by Watts [20] due to his use of inappropriate viscoelastic solutions [3].

If the viscoelastic or elastic-viscoelastic models are appropriate instead of the elastic models, then the effective flexural rigidities are dependent on the wavelength of the load, shorter-wavelength loads resulting in higher D_{eff} than longer-wavelength loads [3,21]. This also follows from equation (1). Thus the larger Tahiti load could be

expected to yield a lower D_{eff} than Rarotonga but this is not likely to be important for the relatively young loads considered here. Equation (1) is evaluated in Fig. 6 for three relaxation times and for two wavelengths, one corresponding to the smaller loads that characterize the Cook Islands and one corresponding to Tahiti. If $\tau \approx 10^7\text{--}10^8$ then the wavelength effect is unimportant for loads whose ages are less than a few million years. With equation (2) the resulting time constant is in the range $5\text{--}10 \times 10^6$ years. This corresponds to an effective viscosity for the lithosphere of the order $0.5\text{--}1 \times 10^{26}$ poise.

These results, while preliminary in that they must await confirmation from analyses of other seamounts, do indicate that the response of the oceanic lithosphere to seamount loading is a function of several parameters and that, by a careful choice of the loads, these factors can be separated. Specifically, stress relaxation in the lithosphere makes a significant contribution to the effective flexural rigidity. The results also indicate that the

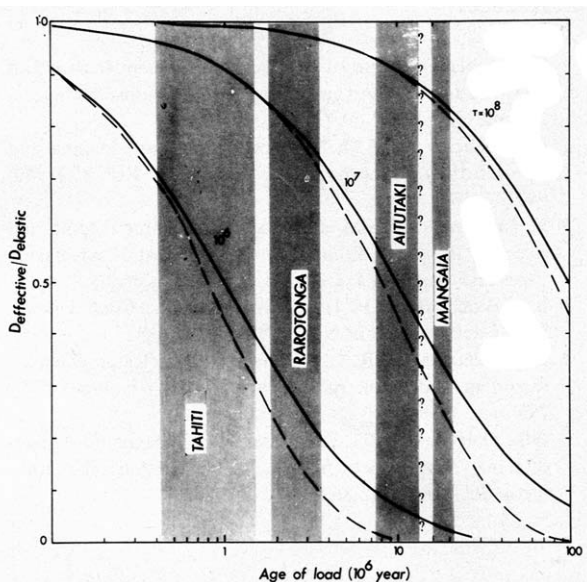


Fig. 6. Ratio of the effective flexural rigidity to the elastic value for a viscoelastic lithosphere as a function of duration of loading and for three viscous time constants ($\tau = 10^6, 10^7$ and 10^8 years). The solid lines are for "small" seamounts (typical of the Southern Cook loads) and the dashed lines are for "large" seamounts (typical of Tahiti). The ages of the island loads discussed in the text are indicated.

most satisfactory estimates of the effective flexural rigidity follow from observations of geoid height or gravity perturbations, from island uplift and from bathymetric evidence. Geoid height perturbations alone do not suffice in establishing the three principal flexural parameters D , ρ and β and, in fact, are often more sensitive to ρ and β than the rheologically more interesting parameter D .

Acknowledgements

The geoid computations have been carried out by R. Coleman and B. Hirsch using GEOS 3 altimeter data supplied by the National Aeronautics and Space Administration. R. Coleman also analysed the GEOS 3 and SEASAT altimeter profiles and P. Hopgood carried out the elastic flexure calculations. S.M. Nakiboglu contributed greatly to the considerations of viscoelastic models.

References

- 1 K. Lambeck, Flexure of the ocean lithosphere from island uplift, bathymetry and geoid height observations: the Society Islands, *Geophys. J.* 66 (in press).
- 2 K. Lambeck and S.M. Nakiboglu, Seamount loading and stress in the ocean lithosphere, *J. Geophys. Res.* 85 (1980) 6403–6418.
- 3 K. Lambeck and S.M. Nakiboglu, Seamount loading and stress in the ocean lithosphere, 2. viscoelastic and elastic-viscoelastic models, *J. Geophys. Res.* 86 (in press).
- 4 B.L. Wood and R.F. Hay, Geology of the Cook Islands, *N.Z. Geol. Surv. Bull. N.S.* 82 (1970) 103 pp.
- 5 I. McDougall and R.A. Duncan, Linear volcanic chains—recording plate motions?, *Tectonophysics* 63 (1980) 275–295.
- 6 G.B. Dalrymple, R.D. Jarrard and D.A. Clague, K-Ar ages of some volcanic rocks from the Cook and Austral Islands, *Geol. Soc. Am. Bull.* 86 (1975) 1463–1467.
- 7 D.L. Turner and R.D. Jarrard, K-Ar dating of the South Cook Islands: a test of the hot-spot hypothesis, *Hawaii Symp. on Intraplate Volcanism and Submarine Volcanism, Abstract Volume* (1979) 129.
- 8 McNutt and H.W. Menard, Lithospheric flexure and uplift atolls, *J. Geophys. Res.* 83 (1978) 1206–1212.
- 9 R.D. Jarrard and D.L. Turner, Comment. *J. Geophys. Res.* 84 (1979) 5691–5694.
- 10 C.P. Summerhayes, Bathymetry and topographic lineation in the Cook Islands, *N.Z. J. Geol. Geophys.* 10 (1967) 1382–1399.
- 11 J. Mammerickx, R.N. Anderson, H.W. Menard and S.M. Smith, Morphology and tectonic evolution of the East-Central Pacific, *Geol. Soc. Am. Bull.* 86 (1975) 111–118.
- 12 M.P. Hochstein, Seismic measurements in the Cook Islands, Southwest Pacific Ocean, *N.Z. J. Geol. Geophys.* 10 (1967) 1499–1526.
- 13 E.I. Robertson, Additional gravity surveys in the Cook Islands, *N.Z. J. Geol. Geophys.* 13 (1970) 184–198.
- 14 E.I. Robertson, Gravity survey in the Cook Islands, *N.Z. J. Geol. Geophys.* 10 (1967) 1484–1498.
- 15 N.I. Christensen and M.H. Salisbury, Structure and constitution of the lower oceanic crust, *Rev. Geophys. Space Phys.* 13 (1975) 57–86.
- 16 G.P. Woollard, The interrelationships of crustal and upper mantle parameter values in the Pacific, *Rev. Geophys. Space Phys.* 13 (1975) 87–137.
- 17 R.W. Raitt, The crustal rocks, in: *The Sea*, 3 M.N. Hill, ed. (John Wiley, New York, N.Y., 1963) 85–102.
- 18 R. Coleman, On the recovery of ocean dynamic information from satellite altimetry, *Mar. Geod.* 4 (1980) 351–386.
- 19 K. Lambeck and R. Coleman, On the verification of bathymetric charts from satellite altimeter data in the region of the Cook Islands, (submitted to *N.Z. J. Geol. Geophys.*).
- 20 A.B. Watts, An analysis of isostasy in the world's ocean, 1. Hawaiian-Emperor seamount chain, *J. Geophys. Res.* 83 (1978) 5989–6004.
- 21 C. Beaumont, The evolution of sedimentary basins on a viscoelastic lithosphere: theory and examples, *Geophys. J. R. Astron. Soc.* 55 (1978) 471–497.
- 22 R.S. Mather, R. Coleman, C. Rizos and B. Hirsch, A preliminary analysis of GEOS-3 altimeter data in the Tasman and Coral Seas, *Unisurv. G26*, School of surveying, University of New South Wales, Sydney, N.S.W. (1977) 27–46.