

Viscosity structure of the oceanic lithosphere inferred from the differential late Quaternary sea-level variations for the southern Cook Islands

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SUMMARY

Late Quaternary sea-level variations for the southern Cook Islands such as Rarotonga and Mangaia provide information on the time-dependent crustal movement due to viscoelastic arching in response to loading by the Pleistocene volcanic island of Rarotonga. The lithospheric responses to both external and internal loads have been investigated to estimate the viscosity of the lower part of the lithosphere and to examine the initial stage of swell formation. Detailed observations of sea-level variations for the past 125 kyr indicate that the crustal uplift for Mangaia is greater than 10 m, while Rarotonga was apparently stable for this period. The following geophysical implications for the lithospheric rheology and loading model are derived from these observations. The observed differential crustal movement implies that the viscous relaxation associated with this volcanic loading is still proceeding in the lithosphere. The layer supporting stresses has therefore been migrating with time from weaker lower zones into the stronger upper zones for a lithosphere with a depth-dependent viscosity structure. This fact provides an important constraint on the viscosity of the lower part of lithosphere. The observation that Rarotonga has been apparently stable for this period is indicative of a local buoyant internal load in the upper mantle. This load may be related to small-scale and secondary convection in the asthenosphere. Surface uplift due to an internal load is therefore required to cancel the subsidence by volcanic loading. This problem has been examined for two simplified background density models. One is a model in which the density of the lithosphere is equal to that of the asthenosphere. For this model, very large mass anomalies which are 10 times larger than the external load are required beneath the lithosphere in order to explain the observed differential crustal movement of the islands. For an earth model for which the density of the lithosphere is greater than that of the asthenosphere, which is possible for mature oceanic lithosphere, the observed differential crustal movement is explained for an internal-load model with density anomalies of less than 20 kg m^{-3} . The volume of the internal load is at most twice the volume of the external load. A high-viscosity layer with an effective viscosity of 10^{24} Pa s and with a thickness of greater than 60 km is required beneath the top elastic layer with a thickness of 10–15 km. The thickness of thermal lithosphere estimated by the plate age of this region is approximately 80–90 km, regardless of the age–thickness relationship adopted. It is therefore suggested that the major part of the thermal lithosphere is composed of a viscoelastic layer with an effective viscosity of 10^{24} Pa s and with a relaxation time of 1 Myr.

Key words: lithospheric viscosity, oceanic lithosphere, Quaternary sea-level

INTRODUCTION

The rheological structure of the oceanic lithosphere is very important in order to understand quantitatively the mantle dynamics associated with geological phenomena at several

timescales. In fact, it has been proved that lithospheric rheology plays an essential role in modelling realistic mantle convection that satisfies surface expression and seismic tomographic observations especially of plate margins (Zhong & Gurnis 1995). Its viscosity distribution, however, has not yet been elucidated

because the inferred lithospheric viscosity depends not only on plate age but also on the details of load. For example, an effective thickness of oceanic lithosphere will be dependent on the duration of external and/or internal loading. If the lithospheric viscosity structure is assumed to be depth-dependent, then its thickness is defined by the thickness of layer supporting the stresses caused by the load with characteristic time and spatial scales (Walcott 1970; Lambeck 1981a,b; Bodine, Steckler & Watts 1983; Watts & Ribe 1984). This means that the layer supporting stresses will migrate with time from weaker lower zones into the stronger upper zones for lithosphere with a depth-dependent viscosity structure. It is, however, difficult to observe the surface crustal movement associated with this time-dependent process of viscous relaxation within the lithosphere. In most cases, we can observe the final state, that the stresses related to the loading are being supported by the upper part of lithosphere effectively working as an elastic body.

Seismic estimates of lithospheric thickness, short-term elastic thickness, will generally give relatively large values, and its thickness (l) increases with the square root of the age of oceanic lithosphere (t), i.e., $l \propto \sqrt{t}$ (Yoshii 1973; Leeds, Knopoff & Kausel 1974; Parsons & Sclater 1977). This is approximately identical to the thickness of the thermal boundary layer defined by the 1100°C isotherm, and its thickness l is roughly $10\sqrt{t}$ (km), where the plate age t is in units of Myr (million years) (e.g. Turcotte & Schubert 1982). Its thickness can also be evaluated from the sea-level variations due to the glacial rebound for a timescale of 10^3 to 10^4 years. In fact, the related problem of crustal tilting at continental margins, which has a typical wavelength of less than 1000 km, has the potential to provide information on the upper-mantle rheological structure (Nakada & Lambeck 1987, 1989). More sufficient observations are, however, required to solve the trade-off between lithospheric thickness and upper-mantle viscosity.

Estimates inferred from topographic and gravity measurements of the flexure of islands, trenches, and mountain belts with characteristic loading timescales of greater than 10^6 years will usually lead to lower values of 10 to 30 km associated with a temperature of 300–450°C (Watts & Daly 1981; Calmant & Cazenave 1986; Calmant 1987; Goodwillie & Watts 1993; Filmer, McNutt & Wolfe 1993; Watts 1994). This thickness will correspond to the effective elastic thickness of the upper part of lithosphere because the stress relaxation time of the lower part of lithosphere is shorter than the characteristic loading timescale. For the value of 10–30 km estimated by these methods, the thickness of the lower part of lithosphere amounts to 80 km, for example, for a 100 Myr old oceanic plate. It is only suggested that its Maxwell relaxation time is less than 10^7 years, and it may be difficult to evaluate the viscosity structure of the lower part of lithosphere from these data (e.g. Bodine *et al.* 1983).

Several studies have, however, revealed the nature of oceanic viscoelastic lithosphere based on the outer topographic rise near trenches. For example, Ida (1984) evaluated the effective viscosity to be 10^{24} Pa s or less using a model of viscoelastic lithosphere, and De Bremaecker (1977) estimated its viscosity to be about 10^{23} Pa s based on a viscous plate model. They noted, however, that the bathymetric data alone cannot uniquely determine the relaxation time or effective viscosity. Thus their results for effective viscosity have been obtained so as not to make the horizontal force abnormally large. Considering that the lower layer occupies the major part of

the lithosphere in the framework of mantle convection, determination of its viscosity is also extremely important in modelling mantle convection with realistic subduction slabs (Zhong & Gurnis 1995). In fact, it has recently been possible to obtain high-resolution seismic tomographic images for the subducting slabs, which provide important constraints on the modelling of mantle convection (van der Hilst *et al.* 1991; Fukao *et al.* 1992; van der Hilst 1995).

In this paper we examine the crustal movement during the past 1.5 Myr for the southern Cook Islands. These observations were first used by McNutt & Menard (1978), and also analysed by Lambeck (1981a). The work by McNutt & Menard (1978) is a pioneering work on the evaluation of effective elastic thickness of oceanic lithosphere. They estimated its thickness by assuming that the observed uplift for Mangaia, Mauke, Atiu and Mitiaro was caused by viscoelastic arching in response to the volcanic loading of Rarotonga about 1.5 Myr ago. They used the observations for total lithospheric deformation for uplifting islands such as Mangaia, and estimated the elastic thickness to be 10–15 km for the lithosphere around this region. It is, however, possible to use the time-dependent crustal displacement inferred from the sea-level variations during the past 6 kyr (thousand years) and at about 125 kyr BP (before present) of the last interglacial stage. These observations therefore provide information on the time-dependent viscous relaxation within the lithosphere. In evaluating lithospheric viscosity based on the scenario assumed by McNutt & Menard (1978), we should examine the observations of crustal movement for both Mangaia and Rarotonga. If the crustal movement for the uplifting islands were solely caused by the time-dependent lithospheric response due to the loading of Rarotonga, then significant subsidence about 10 times greater than that for Mangaia would be expected at Rarotonga. There is, however, no evidence supporting such significant subsidence at Rarotonga, at least during the past 125 kyr.

Internal loads in the upper mantle as well as external loads may therefore be required in order to explain the time-dependent differential crustal movement inferred from late Pleistocene and Holocene sea-level observations. Thus, lithospheric thinning (erosion) by secondary asthenospheric convection associated with mantle plume and its related crustal uplift (e.g. Yuen & Fleitout 1985; Olson *et al.* 1988; Monnereau, Rabinowicz & Arquís 1993) may be required to explain the differential crustal movement in this region. Comparison between observations and predictions here may provide constraints on the formation of a swell like the Hawaiian swell. In particular, it may be possible to obtain information for the transient processes of the lithospheric erosion corresponding to the squeezing of the ductile part of the lithosphere numerically simulated by Monnereau *et al.* (1993).

OBSERVATIONAL EVIDENCE FOR THE CRUSTAL UPLIFT

McNutt & Menard (1978) have attributed uplifts of the southern Cook Islands to lithospheric arching as a consequence of nearby growth of recent volcanoes such as Rarotonga. In fact, Mangaia, Atiu, Mauke and Mitiaro are surrounded by a raised coral reef (makatea), and uplifted atoll (Jarrard & Turner 1979). The ages for Mangaia, Atiu, Mauke and Mitiaro are respectively 16–21, 7–8, 5 and >12 Ma, and that for Rarotonga is 1.2–1.8 Ma (Dalrymple, Jarrard & Clague 1975;

Jarrard & Turner 1979). A very extensive coral reef and a single minimum age of >6.5 Ma suggest that the bulk of the volcano for Aitutaki is similar in age to Atiu and Mauke, although most dated volcanism is 0.8–1.8 Ma (Jarrard & Turner 1979). The uplifting of Mangaia, Atiu, Mauke and Mitiaro may therefore be explained by the lithospheric arching associated with volcanic loading of Rarotonga on the 87 ± 5 Myr old plate (Calmant 1987).

McNutt & Menard assumed the total uplift of Mangaia by volcanic loading to be equal to the present altitude of the makatea surface at 70 m above sea level. Jarrard & Turner (1979), however, concluded that the uplift values are greater than the present elevation of 169 m for Mangaia by taking account of the considerable erosion of makatea. Thus the altitude depicted in Fig. 1 may correspond to the total uplift for these islands (Jarrard & Turner 1979). The other problem in accurately estimating the uplift caused by the loading of Rarotonga is that the history of eustatic sea-level change during the Plio–Pleistocene is not accurately known, even if the timing of uplift for these islands were accurately known (Jarrard & Turner 1979). Thus it may be difficult to obtain

reliable information for the total crustal uplift related to the loading of Rarotonga.

It is, however, possible to obtain time-dependent information for the sea-level variations during the past 150 kyr. If the uplift is attributed to the lithospheric response to the volcanic loading of Rarotonga (McNutt & Menard 1978; Lambeck 1981a), then the sea-level variations for this period will be able to provide information on the relaxation processes within the lithosphere. Time-dependent crustal movement during the past 150 kyr will not be expected, for example, if the Maxwell relaxation time for viscoelastic lithosphere is significantly shorter or longer than the loading period of about 1.5 Myr for Rarotonga. We next summarize the Holocene and late Pleistocene sea-level observations of these islands, particularly for Mangaia and Rarotonga, because detailed sea-level observations have recently been made for these islands (Yonekura *et al.* 1984, 1988; Stoddard, Spencer & Scoffin 1985; Berryman 1987; Spencer, Stoddard & Woodroffe 1987; Woodroffe *et al.* 1990, 1991). Detailed geological descriptions of the Cook Island group were made by Wood & Hay (1970).

Sea-level variations during the past 20 kyr are ascribed to

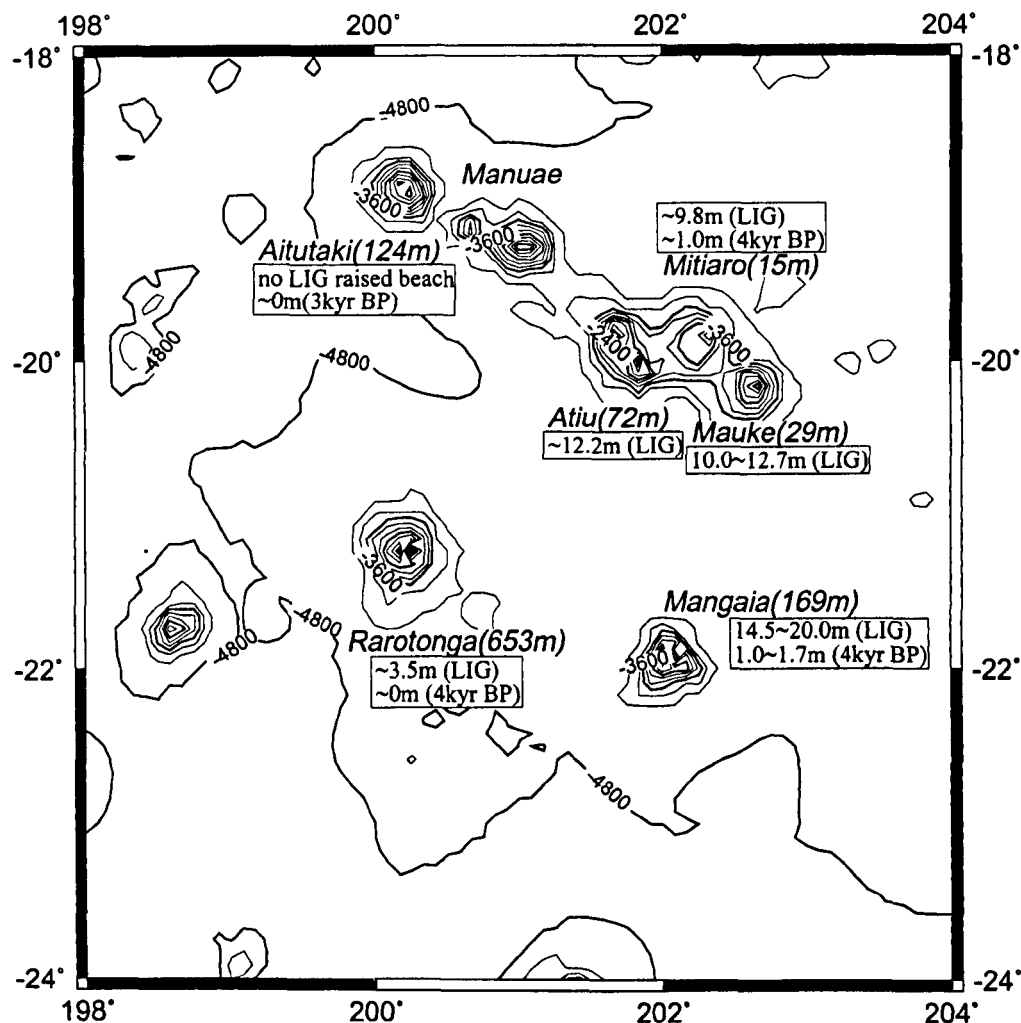


Figure 1. Location map of the southern Cook Islands and the magnitudes of observed relative sea-level at the last-interglacial phase (LIG) of about 125 kyr BP and at about 4 kyr BP. Bathymetry (depth in metres) is based on the ETOPO5 topography data set available from the National Geophysical Data Center in Boulder, Colorado. The numbers in parentheses represent the height of the islands.

both tectonic movement and sea-level change associated with the melting of the last great ice sheets (glacial rebound). The sea-level variations caused by glacial rebound should be the same for all the Cook islands, regardless of the viscosity model or ice model (e.g. Nakada & Lambeck 1989). Thus no differential crustal movement is predicted for these islands. On the basis of glacial rebound models, Nakada & Lambeck (1989) predicted that sea-level increases monotonically until the present in the Cook Islands. The observations are, however, indicative of differential crustal movement for these islands. The sea-level curve at Rarotonga is consistent with the model predictions, and the sea level during the past 4 kyr is nearly the same as the present value (Yonekura *et al.* 1984; Woodroffe *et al.* 1990, 1991). Evidence of late Holocene emergence has not been found on Aitutaki or Rarotonga either, where the sea level reached its present level about 3 kyr BP (Berryman 1987). Observations at Mangaia (Yonekura *et al.* 1988), however, show a sea-level high-stand of about 1.0–1.7 m at 4–5 kyr BP (see Fig. 1). A sea-level high-stand of about 1 m at 4 kyr BP has been observed for Mitiaro (Woodroffe *et al.* 1991). Mauke also appears to have emerged, though by how much is uncertain, and an erosional bench and notch occur on Atiu (Woodroffe *et al.* 1990). Thus Holocene emergence has been widespread for the islands of Mangaia, Mitiaro, Atiu and Mauke. These sea-level observations therefore indicate a differential crustal movement of 1 m for this period.

In a pioneering study on the application of uranium-series disequilibrium dating (U-series dating) of coral on oceanic islands, Veeh (1966) demonstrated that limestones of last interglacial age at about 125 kyr BP reach several metres above present sea-level on a number of mid-plate islands. Last interglacial (LIG) reefs identified by U-series dating reach 12.2 m on Atiu, 9.8 m on Mitiaro, 10–12.7 m on Mauke and 14.5–20 m on Mangaia (Woodroffe *et al.* 1991). Aitutaki, however, lacks not only Tertiary limestones but also LIG raised reef (Stoddart *et al.* 1985; Spencer *et al.* 1987). The Pleistocene island of Rarotonga lacks Tertiary makatea, but an old reef remnant with corals in the position of growth is apparent, reaching 3.5 m above the present reef flat near Ngatangia on the east coast (Wood & Hay 1970). Schofield (1970) reported two radiocarbon ages from this limestone, of 28 200 and >43 400 yr BP. These ages are approaching the limit of radiocarbon dating, and it has been suggested that the limestone is most likely to be of LIG age (Berryman 1987). Observed data for these islands are shown in Fig. 1. The elevations of the raised reefs are minimum values, as some erosion may have occurred since the formation of individual reefs (Jarrard & Turner 1979). These observations indicate the differential crustal movement between Rarotonga and Mangaia, Mauke, Atiu and Mitiaro surrounding Rarotonga since the last interglacial.

In late Quaternary sea-level and coastal tectonic studies, the last interglacial is generally cited as having a sea level of +6 m relative to present (e.g. Broecker *et al.* 1968; Bloom *et al.* 1974). There are, however, indications that the glacial isostatic component is important for the LIG sea level on the Earth because of its viscous mantle (Lambeck & Nakada 1992). In particular, evaluation of this component is essential on an earth model with a lower-mantle viscosity of greater than 10^{22} Pa s (Lambeck & Nakada 1992). Thus the generally accepted assumption that the sea level 125 kyr ago was effectively 5 m higher than the present level at sites in tectonically stable areas

cannot be supported. In fact, Lambeck & Nakada (1992) indicated that the Caribbean sea-level observations for the last interglacial phase (Chen *et al.* 1991) can be reasonably explained by the glacial rebound sea-level variations on a viscoelastic earth model. According to these results, the observed altitudes of LIG phase for the Cook Islands are 3 m higher than the present level. Accurate dates by U-series dating have not been obtained for the raised reef with an altitude of about 3.5 m on Rarotonga, but the observed altitudes of sea level during the past 125 kyr for Rarotonga, youngest island in the southern Cook Islands, seem to be dominantly attributed to the sea-level variations caused by the glacial isostatic adjustment. On the other hand, Mangaia, Mauke, Atiu and Mitiaro have been uplifting during this period by another mechanism.

The uplift for islands such as Mangaia has generally been attributed to the viscoelastic lithospheric response to the volcanic loading of Rarotonga about 1.5 Myr ago (McNutt & Menard 1978; Lambeck 1981a). An unanswered question is why the islands such as Mangaia surrounding the young oceanic island of Rarotonga are significantly uplifting while Rarotonga is apparently stable. If their scenario is correct, then the magnitude of subsidence of Rarotonga should be 10 times larger than that for the uplift of Mangaia. Internal loads in the upper mantle as well as external loads may therefore be required in order to explain the time-dependent differential crustal movement inferred from late Pleistocene and Holocene sea-level observations. We model this by considering simplified external and internal loads. We also assume that the uplift on Mangaia is close to the maximum for the region. This assumption is supported by the last interglacial sea levels on the other islands. Most of the discussion is restricted to fitting the crustal movement for Mangaia and Rarotonga.

MODEL DESCRIPTION

Geophysical observations in French Polynesia, such as depth anomalies, low velocities for Love waves and a broad low in the geoid, can be explained as manifestations of a superplume rising through a low-viscosity zone in the upper mantle (McNutt & Judge 1990). In fact, oceanic flexure studies from bathymetric and geoid data indicate that the effective elastic thickness in this region is unusually low compared with the value for the North Pacific region (Calmant & Cazenave 1986). The low elastic thickness in this region has been attributed to the regional lithospheric reheating associated with plume rising (McNutt & Judge 1990). Recent studies in French Polynesia, however, suggest that the local thermal anomaly in the lithosphere and the mantle dynamics beneath each seamount are the main factors controlling the lithospheric thickness (Stein & Abbott 1991; Goodwille & Watts 1993; Filmer *et al.* 1993). These studies, which are based on better bathymetric and geoid data sets than were available to Calmant & Cazenave (1986), show that the effective elastic thickness from flexure analyses for seamounts in the Pacific superswell region is not necessarily low. Thus flexure studies on the oceanic lithosphere from bathymetric and geoid data suggest that the lithospheric response for loading is controlled by the thermal perturbation in the lithosphere and/or local and small-scale convection in the asthenosphere beneath the lithosphere.

The thermal anomalies related to an internal buoyant source were first proposed by Crough (1978) to explain the forma-

tion of hotspot swells and the rapid subsidence of reheated lithosphere after moving off the heat source. Several recent numerical and laboratory studies, however, indicate that thermomechanical erosion of the lithosphere by secondary asthenospheric convection associated with a mantle plume seems to be a main factor in the formation of hotspot swells such as the Hawaiian swell (Yuen & Fleitout, 1985; Olson *et al.* 1988; Robinson & Parsons 1988; Ceuleneer *et al.* 1988; Monnereau & Cazenave 1990; Monnereau *et al.* 1993; Davies 1994; Sleep 1994). Thus it may be reasonable to consider local internal loads in the upper mantle, as is the case here. Differential crustal movement in this region may then provide constraints on the initial stage of the lithospheric erosion leading to crustal uplift. The convective process in the asthenosphere is related to the squeezing of the ductile part of the oceanic lithosphere numerically simulated by Yuen & Fleitout (1985) and Monnereau *et al.* (1993).

Differential crustal movement inferred from the sea-level observations for Mangaia and Rarotonga is indicative of relatively local internal loads in the upper mantle. Volcanic loading of Rarotonga corresponds to external loads, and underplating of relatively low-density material at the base of lithosphere (e.g. Crough 1978; Fleitout & Moriceau 1991) and small-scale convection in the asthenosphere (e.g. Yuen & Fleitout 1985; Monnereau *et al.* 1993; Davies 1994) work as internal loads. Thus an internal load, by which the surface subsidence due to volcanic loading is cancelled for the central region of the external load, is likely to be required in the shallow upper mantle. In fact, Moriceau, Christensen & Fleitout (1991) and Monnereau *et al.* (1993) have indicated that geoid and topography above swells can be explained by hot material at a depth of the order of 100 or 150 km and that the deep part of the plume itself contributes very little to the surface signal. Also, large igneous intrusions have been seismically observed at the base of the crust beneath the Marquesas Islands (Wolfe, McNutt & Detrick 1994; Caress *et al.* 1995). These extensive intrusions at the base of the crust significantly affect the surface gravities and topography, but are not important for the time-dependent crustal movement examined here if the crust essentially behaves as an elastic body (Fleitout & Moriceau 1991; Nakada 1994). We therefore ignore the intrusive internal loads at the base of the elastic top layer for the model used here. Thus the gravity and geoid anomalies associated with the loads adopted here may not be consistent with the gravity observations. Future studies that include these effects will be required when seismic information for this region is available.

Fig. 2 depicts the configuration of the model used here. A Maxwell viscoelastic body is adopted for mathematical simplicity, and for an extensive examination of the responses of arbitrary earth models to arbitrary loads. A viscoelastic body can be easily applied to the deformation problem of a viscoelastic lithosphere with both external and internal loads. In the steady state, however, the viscoelastic body with pre-stress is treated as a viscous fluid (Wolf 1991; Wu 1992), since pre-stress will decay in a viscous body and can be ignored once viscous strains are greater than elastic strains. A normal-mode method is adopted for evaluating the responses of the lithosphere and upper mantle to internal and external loads with cylindrical geometry. Thus the response to a unit load at an arbitrary depth is obtained by imposing the physical conditions of continuity of displacement and stress and of a stress-free

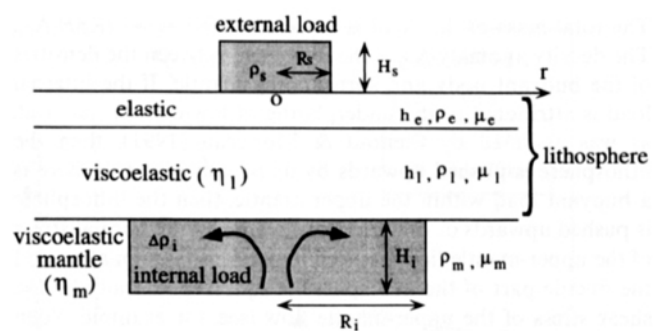


Figure 2. Model configuration used to examine the relationship between the earth and load models and the lithospheric response. The lithosphere is composed of an upper elastic layer and a viscoelastic layer with viscosity η_l and density ρ_l . The mantle beneath the lithosphere is treated as a viscoelastic body, and its density is represented by ρ_m . The external load corresponding to volcanic loading has a cylindrical geometry with radius R_s , height H_s and density ρ_s . A steady cylindrical internal load with radius R_i , height H_i and density anomaly $\Delta\rho_l$ is assumed. In this study, we examine lithospheric responses to external and internal loads for two models with background densities of $\rho_l = \rho_m$ and $\rho_l > \rho_m$.

condition at the surface. At the depth of the internal load, we impose the condition of a jump of normal stress corresponding to the internal load (density anomaly). This formulation is the same for an external loading problem such as glacial rebound if we consider the shearing and normal loads at the surface to be mathematically equivalent to the internal load (Nakada 1994). In the calculation, the internal source layer is divided into thin sheets of 10 km thickness. The response to each 10 km sheet is calculated by imposing boundary conditions at the depth of the sheet, and the total response is evaluated by adding the response due to each internal load.

The lithosphere is divided into two parts. The upper part is treated as an elastic layer with density ρ_e , thickness h_e , rigidity μ_e and infinite Newtonian viscosity, corresponding to the effective elastic layer estimated from bathymetric and geoid data (Calmant & Cazenave 1986; Goodwille & Watts 1993). The lower part is assumed to be Maxwell viscoelastic with Newtonian viscosity η_l , density ρ_l , thickness h_l and rigidity μ_l . Fleitout & Moriceau (1991) adopted a viscous lower layer with depth-dependent viscosity overlain by an elastic lid for examining the topography and geoid anomalies due to density heterogeneities at the base of the thermal lithosphere. As they suggested, the viscoelastic model adopted here is sufficiently realistic for the discussion of stress relaxation in the lithosphere. The mantle beneath the lithosphere is treated as a viscoelastic body with viscosity η_m , density ρ_m and rigidity μ_m . In some models examined here, a low-viscosity asthenospheric layer with viscosity η_a , density ρ_a and rigidity μ_a is assumed beneath the lithosphere. In this study, the rigidities of the elastic layer (μ_e) and the other layers are assumed to be 10^{10} and 10^{11} Pa respectively. Densities of the elastic layer (ρ_e) and mantle (ρ_m) are assumed to be 3000 and 3300 kg m⁻³ respectively. The following results do not essentially change, even if we adopt other values for these parameters.

The external load has a cylindrical geometry with radius R_s , height H_s and density ρ_s , and its total mass (m_s) is $m_s = 4\pi R_s^2 H_s \rho_s$. A steady internal load with cylindrical geometry is also assumed. Its radius, height and density anomaly relative to the ambient upper mantle are R_i , H_i and $\Delta\rho_l$, respectively.

The total mass of this load is represented by $m_i = 4\pi R_i^2 H_i \Delta\rho_i$. The density anomaly $\Delta\rho_i$ is the difference between the densities of the buoyant body and surrounding mantle. If the internal load is attributed to the underplating of low-density material, as was assumed by Fleitout & Moriceau (1991), then the lithosphere is pushed upwards by its negative load. If there is a buoyant load within the upper mantle, then the lithosphere is pushed upwards or dragged downwards by the normal stress of the upper-mantle flow induced by its density anomalies, and the ductile part of the lithosphere is squeezed laterally by the shear stress of the upper-mantle flow (see, for example, Yuen & Fleitout 1985; Monnereau *et al.* 1993; Nakada 1994; Davies 1994). In this latter process, the lower part of the lithosphere is replaced by hot asthenosphere, leading to thinning of the lithosphere. Both loads therefore contribute to the surface crustal movement. In the following, we examine the lithospheric responses for two models with background densities of (i) $\rho_i = \rho_m$ and (ii) $\rho_i > \rho_m$. Case (ii) will be possible as oceanic lithosphere cools, and it may be efficient to push up the lithosphere by a Rayleigh–Taylor instability, as advocated by Neugebauer (1983) and Mareschal (1983) to explain the uplift history associated with rifting.

MODEL RESULTS

Model for the case of $\rho_i = \rho_m$

A cylindrical external load with a radius of 50 km, height of 3000 m and density of 3000 kg m^{-3} is adopted as the volcanic loading of Rarotonga. The effective density acting as an additional load is 2000 kg m^{-3} , because the density of water is 1000 kg m^{-3} . The total mass of this external (surface) load (m_e) is $7.07 \times 10^{16} \text{ kg}$, and its effective mass as a load (m_e^*) is therefore $4.71 \times 10^{16} \text{ kg}$. We assume that this load was instantaneously applied to the lithosphere 1.5 Myr ago. In the following calculations, we set 1.5 Myr ago to time zero ($t=0$). The time it was loaded is an average age from the observations, and the load size is average.

As was discussed in the section on sea-level observations (Fig. 1), there has been differential crustal movement for Rarotonga and Mangaia during the past 125 kyr. Thus crustal uplift for this period amounts to more than 10 m for Mangaia, while there is no crustal movement for Rarotonga. The uplift for Mangaia continued into the Holocene, as can be inferred from the Holocene sea-level observations. These observations therefore indicate that stress relaxation is still taking place in the lower part of the lithosphere, indicating that it may be possible to evaluate its viscosity structure. In this study we assume that the uplift on Mangaia is close to the maximum, which is supported by the LIG sea-level observations. We mainly use the crustal movement for Mangaia and Rarotonga, and examine the crustal response around Mangaia by imposing the condition for the crustal movement of Rarotonga that the subsidence during the past 125 kyr due to the external (volcanic) loading is perfectly cancelled by the uplift due to the internal load.

Fig. 3 illustrates the crustal displacement during the past 125 kyr by showing the following: (i) surface subsidence caused by the external load; (ii) surface uplift caused by the internal load; and (iii) total crustal displacement caused by both loads. The density anomaly for the internal load is determined so as to satisfy the condition that there should be no net crustal

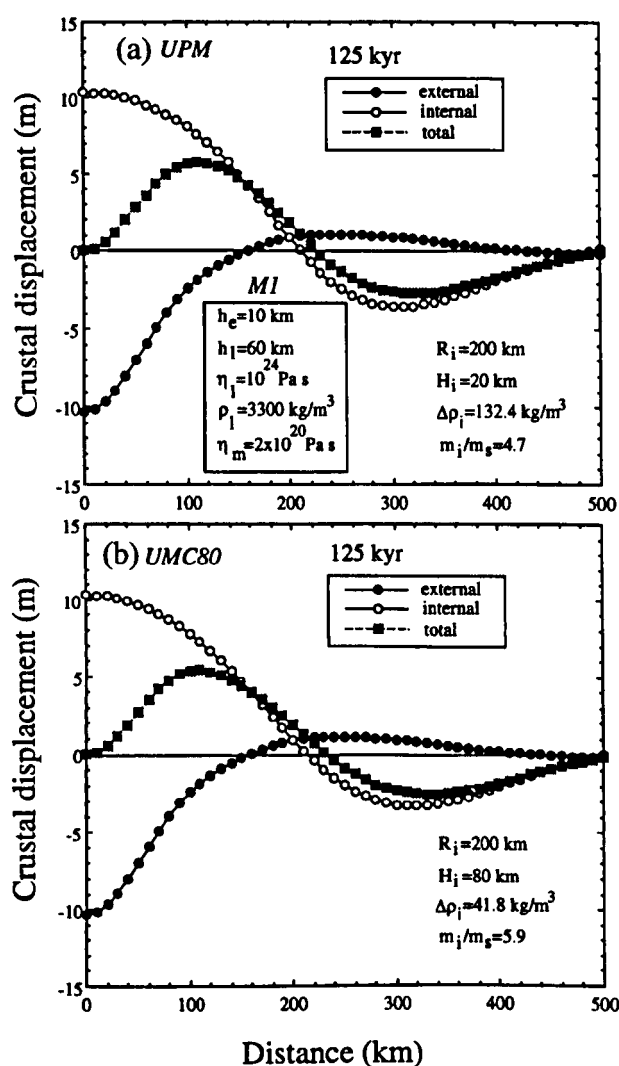


Figure 3. Crustal displacement during the past 125 kyr due to external and internal loads, and both loads. Model parameters describing earth and load models are shown in the figure. The earth model used here is M1 (see Table 1). The internal load for (a) is the underplating model (UPM) with internal load just below the lithosphere. (b) is a model result for the internal-load model UMC80 in which there is an internal load in the asthenosphere beneath the lithosphere.

movement at the centre later than 125 kyr BP. The thickness of the top elastic layer is 10 km for the model used for this figure. The thickness, viscosity and density of the lower part of lithosphere (LPL) are, respectively, 60 km, 10^{24} Pa s and 3300 kg m^{-3} . The viscosity beneath the lithosphere is assumed to be $2 \times 10^{20} \text{ Pa s}$. This earth model is referred to as M1. The magnitude of the crustal displacement in the central region due to the external load is about 10 times larger than that for the bulge region. For example, its values at the centre and $r = 250 \text{ km}$ are about -10 and 1.5 m , respectively. This is true regardless of the earth and external load models.

Two types of model are adopted for internal loads. In one model the density anomaly exists just below the lithosphere (underplating model). In this case, referred to as UPM, the lithosphere is uplifted by its positive buoyancy (upward) force. This internal-load model is the same as that adopted by Fleitout & Moriceau (1991), and is similar to the lithospheric

reheating model of Crough (1978). Fig. 3(a) illustrates the crustal displacement for a cylindrical underplating load with radius of 200 km. If we quantify this internal load as a cylindrical load with a thickness of 20 km, its density anomaly $\Delta\rho_i$ and relative mass ratio defined by m_i/m_e are 132.4 kg m^{-3} and 4.7, respectively. The other model has an internal load within the uppermost mantle. In this case, the lithosphere is pushed upwards by the normal stress of mantle flow induced by this density anomaly, referred to here as UMC. This internal-load model may correspond to the convective upwelling or secondary convection model in the asthenosphere used to explain the formation of swells (e.g. Yuen & Fleitout 1985; Robinson & Parsons 1988; Olson *et al.* 1988; Monnereau *et al.* 1993). Fig. 3(b) depicts the relative displacement by a load model with a cylindrical internal load of radius 200 km and a thickness of 80 km (UMC80). The density anomaly $\Delta\rho_i$ and relative mass ratio m_i/m_e are 41.8 kg m^{-3} and 5.9, respectively. The internal load for UMC80 ($H_i=80 \text{ km}$ in Fig. 2) is located within the upper mantle in the depth range $h_e + h_1 \leq z \leq h_e + h_1 + 80$ (see Fig. 2). Then total responses are evaluated by adding the response for each source layer of a thin sheet with 10 km thickness.

For these load models, we assume that the centre of the internal load is identical to that for the external load. If the radius of the internal load is moderately larger than that for the external load, then this assumption will not significantly alter the results examined here. For a load model with different centre positions, the magnitude of the internal load will increase. This is because we need additional internal load to get the same magnitude of uplift as that for the subsidence at the load centre of the external load if we shift the centre of the internal load from that of the external load (see Fig. 3). In this study, we use this assumption for model calculations in order to estimate the minimum value of the internal load.

The relationship between the thickness of LPL (h_l), the magnitude of the internal source and the lithospheric response is illustrated in Fig. 4 for the underplating load model (UPM). The viscosity of the LPL (η_l) is assumed to be 10^{24} Pa s , and open and solid circles represent the results for models with a top elastic layer thickness (h_e) of 10 and 20 km, respectively. Fig. 4(a) shows the maximum relative displacement (h_{max}) during the past 125 kyr, and Fig. 4(b) depicts the magnitudes of the density anomaly and relative mass anomaly (m_i/m_e). The density anomaly is obtained by assuming an internal underplating load with a cylindrical geometry of 200 km radius and 20 km thickness. The distance of maximum relative displacement is about 100–150 km (see Fig. 3). Although this distance depends on the earth model, we ignore this dependence in the present trade-off study. In this paper, we only show the relative displacement during the past 125 kyr, because its value during the past 5 kyr is about 5 per cent of the value during the past 125 kyr. For example, its value during the past 5 kyr is about 0.5 m, compared with the value of 10 m during the past 125 kyr. Sensitivity to LPL viscosity is examined for an earth model with $h_l=60 \text{ km}$ (Table 1). The maximum relative displacement increases with decreasing thickness of the LPL. Then, an increase of the density anomaly (relative mass anomaly) is required because the surface subsidence due to a given external load increases as the thickness of the LPL decreases. The maximum relative displacement significantly decreases for models with a top elastic layer of 20 km thickness (Fig. 4), and for models with a high-viscosity LPL (Table 1). For example,

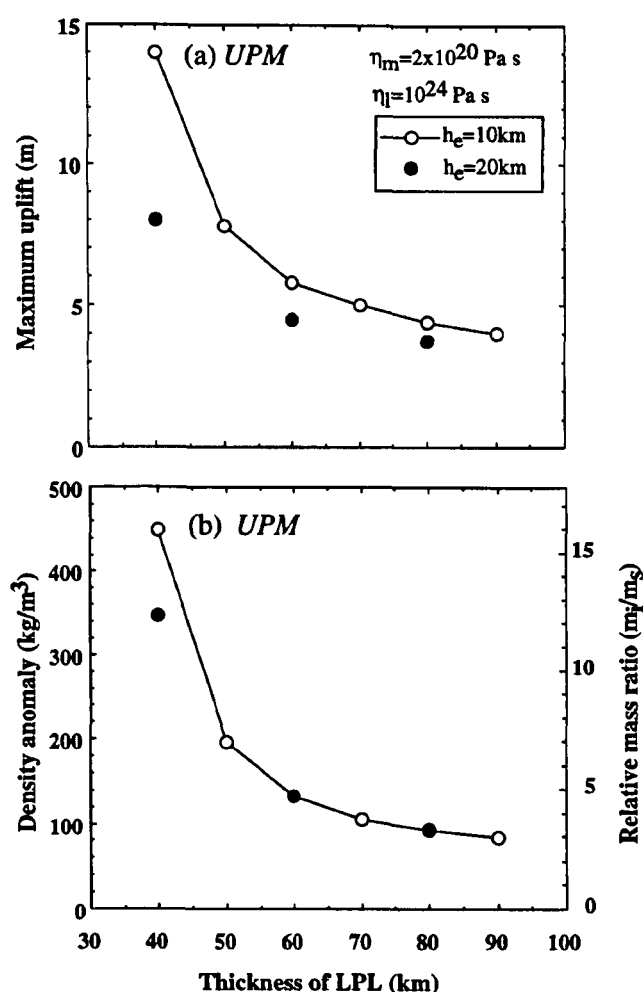


Figure 4. Relationship between earth model, magnitude of internal source, and lithospheric response for the underplating load model (UPM). Model parameters for the earth models are shown in (a). Viscosity of the upper mantle (η_m) is $2 \times 10^{20} \text{ Pa s}$ and that for the LPL layer (η_l) is 10^{24} Pa s . Open and solid circles represent model results for a top elastic layer thickness (h_e) of 10 and 20 km, respectively. (a) Maximum relative displacement during the past 125 kyr at a distance of about 100–150 km, and (b) magnitude of the density anomaly and relative mass ratio (m_i/m_e) of the internal load (m_i) to external load (m_e).

its value for a model with a viscosity of $5 \times 10^{24} \text{ Pa s}$ is only 1.4 m during the past 125 kyr. For a model with a viscosity of $5 \times 10^{23} \text{ Pa s}$, it amounts to 12 m, but the density anomalies are about 360 kg m^{-3} .

Fig. 5 shows the results for the upper-mantle convection model UMC80. The symbol notation in Fig. 5 is the same as that for Fig. 4. Mass anomalies that are larger than those for the UPM model are required in these models because the normal stress of mantle flow induced by internal sources becomes weaker with increasing source depth (Moriceau *et al.* 1991; Nakada 1994). This is also true for the sensitivity of the surface response to LPL viscosity (Table 1). Model results for earth models with a low-viscosity asthenosphere of 200 km thickness are also shown in Table 1. A low-viscosity asthenosphere with temperature- and pressure-dependent viscosity has been considered to be essential to the formation of a swell in the local secondary convection model leading to lithospheric

Table 1. Results for the earth models with $\rho_l = \rho_m$.

η_u (Pa s)	η_a (Pa s)	h_a (km)	η_l (Pa s)	h_l (km)	H_l (km)	Internal load model	$h_{\max}$ (m)	$\Delta\rho_l$ (kg/m ³)	m_l/m_s
2x10 ²⁰ (model M1)			1x10 ²⁴	60	20	UPM	5.8	133.6	4.8
LPL viscosity (η_l) dependence (UPM)									
2x10 ²⁰			5x10 ²³	60	20	UPM	12.0	368.9	12.9
2x10 ²⁰			2x10 ²⁴	60	20	UPM	3.1	80.0	3.0
2x10 ²⁰			5x10 ²⁴	60	20	UPM	1.4	58.9	2.1
LPL viscosity (η_l) dependence (UMC)									
2x10 ²⁰			5x10 ²³	60	80	UMC80	10.0	121.7	17.4
2x10 ²⁰			2x10 ²⁴	60	80	UMC80	3.0	24.6	3.4
2x10 ²⁰			5x10 ²⁴	60	80	UMC80	1.3	17.7	2.5
asthenospheric viscosity (η_a) dependence (UMC)									
2x10 ²⁰	2x10 ¹⁹	200	1x10 ²⁴	60	80	UMC80	5.8	50.4	7.1
1x10 ²¹	2x10 ²⁰	200	1x10 ²⁴	60	80	UMC80	5.4	46.7	6.6

η_u : upper mantle viscosity, η_a : asthenospheric viscosity, h_a : asthenospheric thickness, η_l : LPL viscosity, h_l : LPL thickness, H_l : thickness of internal load, $h_{\max}$: maximum uplift during past 125 kyr,

$\Delta\rho_l$: density anomaly of internal load, m_l/m_s : relative mass ratio

erosion (e.g. Yuen & Fleitout 1985; Olson *et al.* 1988; Monnereau *et al.* 1993). The predicted maximum uplifts for these earth models are, however, identical to those for the uniform viscosity model M1. For an earth model without a density contrast between the lithosphere and asthenosphere, therefore, the asthenospheric low-viscosity layer is not effective in the formation of swell and crustal uplift. An intrinsic instability of denser lithosphere relative to asthenosphere seems to be important in the formation of swell and crustal uplift. In fact, we can expect rapid crustal uplift for an earth model with a decrease in density from the lithosphere to asthenosphere.

Fig. 6 depicts the relative displacement and relative mass ratio for models with a depth-dependent viscosity structure in the lower part of the lithosphere (LPL), where the thickness of the LPL is assumed to be 60 km. The viscosity of the upper half of the LPL is assumed to be 10²⁴ Pa s, and the sensitivity to the viscosity of the lower half of the LPL is examined (see earth model in Fig. 6a). Both relative displacement and relative mass ratio increase with decreasing viscosity of the lower half of the LPL. The relative mass ratio is at best 20 for the models with relative displacement larger than 10 m. The ratio of relative displacement to relative mass ratio (internal density anomaly) decreases as the viscosity of the lower half of the LPL decreases. This is a consequence of the weaker transmission of vertical stress through a less viscous medium (Fleitout & Moriceau 1991).

For the models discussed above, the load radius of the internal load is assumed to be 200 km. The sensitivity to load radius is shown in Fig. 7 for earth model M1. The relationship between relative displacement and relative mass ratio is depicted for models with radii of 100, 200 and 300 km. Both relative displacement and relative mass ratio increase with increasing radius of internal load. Larger mass anomalies are required for internal-load models with larger radii.

For an earth model with the same density for lithosphere and asthenosphere ($\rho_l = \rho_m$), density anomalies and relative

mass ratios greater than 100 kg m⁻³ and 10, respectively, are required in order to explain the relative displacement of greater than 10 m during the past 125 kyr. The density anomaly is attributed to lateral temperature variations in the shallow upper mantle. Kaula (1983) inferred temperature variations of 150–200 K from the plate velocities and heat-flow data. These estimates are consistent with thermal anomalies obtained by numerical simulation (Fleitout & Yuen 1984; Monnereau *et al.* 1993). The density anomaly corresponding to a thermal anomaly of 200 K is 20 kg m⁻³ for a coefficient of thermal expansion of $\alpha = 3 \times 10^{-5}$ K⁻¹. The load model with $\rho_l = \rho_m$ examined in this section is therefore unrealistic, and it is very difficult to explain the observed relative displacement for these oceanic islands by the model with $\rho_l = \rho_m$.

Model for the case of $\rho_l > \rho_m$

While it may be possible to consider another mechanism to explain the observed differential crustal movement for Rarotonga and Mangaia, we investigate this problem by considering an earth model with $\rho_l > \rho_m$. In fact, the part of the lithosphere that is denser than the asthenosphere will increase as the oceanic lithosphere cools with time. Dorman (1969) and Press (1970) indicated a density reversal between the lithosphere and asthenosphere for the oceanic upper mantle based mainly on surface-wave observations. A more detailed study was carried out by Yoshii, Kono & Ito (1976), who used the ocean topography, gravity, explosion seismology and surface-wave observations to examine the structure of oceanic lithosphere. They found a good correlation between the residual gravity anomaly (mantle gravity anomaly) and lithospheric thickness by assuming a constant density contrast of about 100 kg m⁻³ between the lithosphere (average vertically) and asthenosphere. Recently, Stunff & Ricard (1995) suggested that the density contrast is 50–60 kg m⁻³ by using the lithospheric

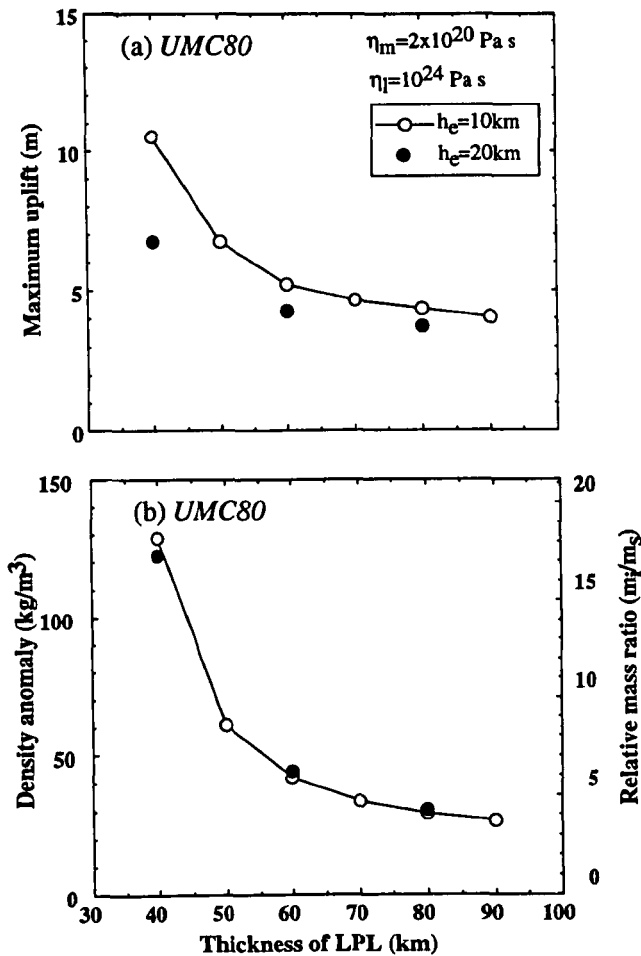


Figure 5. Relationship between earth model, magnitude of internal source, and lithospheric response for the internal load model UMC80. The viscosity of the upper mantle (η_m) is 2×10^{20} Pa s and that for the LPL layer (η_l) is 10^{24} Pa s. The symbol notation is the same as in Fig. 4. (a) Maximum relative displacement during the past 125 kyr, and (b) magnitude of the density anomaly and relative mass ratio (m_i/m_s).

eric mass anomalies and topography data. In this paper, we therefore adopt a density contrast of $50\text{--}100\text{ kg m}^{-3}$.

Fig. 8 depicts the relative displacement during the past 125 kyr by external and internal loads, and the total displacement by both loads. The assumed internal load is the underplating model (UPM) with cylindrical geometry, and the parameters for the earth model used here are the same as those in Fig. 3(a), except for the density structure. The density of the LPL is 3400 kg m^{-3} and its thickness is 60 km. This model is referred to here as N1. For the load model with a radius of 200 km and an equivalent thickness of 20 km, the density anomaly and relative mass ratio are 133.2 kg m^{-3} and 4.7, respectively. These values are nearly the same as those for the model with $\rho_l = \rho_m$ (see Fig. 3). Thus the density reversal between the LPL and asthenosphere does not affect the lithospheric response for the case of an underplating load. This suggests that the perturbation leading to the Rayleigh–Taylor instability may not be effectively excited by the underplating loads exerting only normal stress on the viscous lower part of lithosphere. Therefore we examine the lithospheric response to

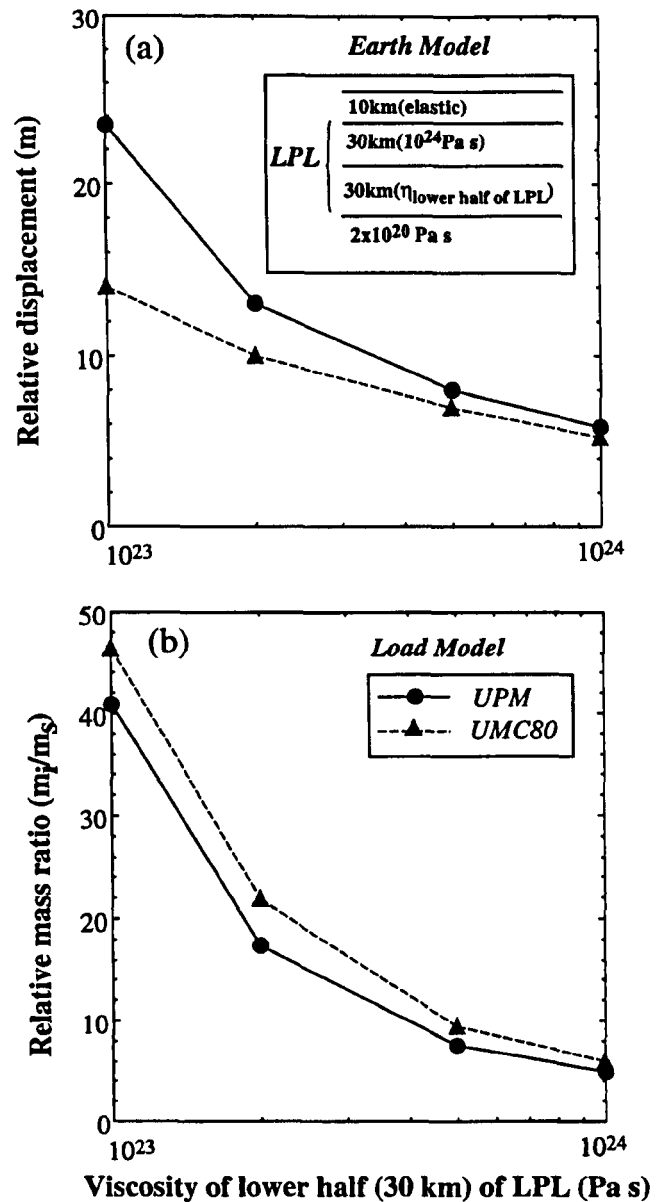


Figure 6. Relative displacement during the past 125 kyr (a) and relative mass ratio (b) for earth models with a depth-dependent viscosity structure for the LPL. Model parameters are shown in (a).

mantle flow induced by the internal load within the upper mantle (UMC internal-load model).

For load models UMC60, UMC80 and UMC100, the maximum relative displacement at a distance of about 200 km, the density anomaly of the internal load and the relative mass ratio are illustrated in Fig. 9 as a function of the thickness of the LPL. The thickness of the top elastic layer is 10 km, and the viscosity of the upper mantle is 2×10^{20} Pa s with a density of 3300 kg m^{-3} . The density and viscosity of the LPL are assumed to be 3400 kg m^{-3} and 10^{24} Pa s, respectively. It is remarkable that a relative displacement of greater than 10 m can be predicted for an internal load with a relative mass ratio of less than 2 (density anomaly of smaller than 20 kg m^{-3}) in the model with $\rho_l > \rho_m$. The thermal anomaly expected from this density anomaly is less than 200 K, and is comparable to the estimate by Kaula (1983). Fig. 9 also indicates that this

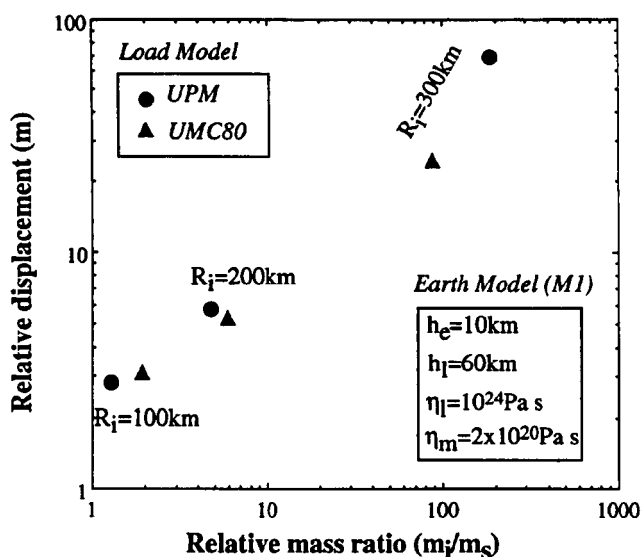


Figure 7. Relative displacement during the past 125 kyr and relative mass ratio for the internal-load models with radii of 100, 200 and 300 km. The earth model used for this figure is M1.

instability occurs in response to the convective flow for internal-load models with relatively deeper sources and in the earth models with LPL thicknesses of greater than 60 km.

Responses for the earth model with an LPL density of 3350 kg m^{-3} are also shown in Table 2, where 60 and 80 km thicknesses are adopted for the LPL and UMC80 is adopted as the load model. The relative displacement is efficiently increased by increasing the density of the LPL. For example, the relative displacement and relative mass ratio are respectively 10.1 m and 2.1 for the earth model with an LPL thickness of 60 km and an LPL density of 3400 kg m^{-3} . In the model with 60 km and 3350 kg m^{-3} for these parameters, the relative displacement is 7.5 m and the relative mass ratio is 4.1. For the earth model where the density of the LPL is depth-dependent the relationship between the relative displacement during the past 125 kyr and the relative mass ratio is shown in Fig. 10. The thickness of the top elastic layer is 10 km, and the density of the upper mantle beneath the LPL is 3300 kg m^{-3} . The LPL is divided into two layers, with an upper layer of thickness h_{lu} and a lower layer of thickness h_{ll} . The notation attached to each symbol is (h_{lu}, h_{ll}) . The density is 3400 kg m^{-3} for the upper layer of the LPL, and that for the lower layer is 3350 kg m^{-3} . The open squares and triangles indicate the results for the models with LPL thicknesses of 60 and 80 km, respectively. The relative mass ratio increases with increasing thickness of the lower layer of LPL with a density of 3350 kg m^{-3} . With the condition that the internal density anomaly should be smaller than 20 kg m^{-3} , the layer with density 3400 kg m^{-3} must be thicker than half of the LPL.

Results for models with LPL viscosities of 5×10^{23} , 1×10^{24} and $2 \times 10^{24} \text{ Pa s}$ are presented in Table 2. The LPL thickness is assumed to be 70 km for this model calculation. While relative displacement increases with decreasing LPL viscosity from $2 \times 10^{24} \text{ Pa s}$ to $5 \times 10^{23} \text{ Pa s}$, the relative mass ratio increases significantly. The sensitivity of the lithospheric response to a low-viscosity asthenosphere is examined by using the earth model with an LPL thickness of 60 km, a viscosity of 10^{24} Pa s and a density of 3400 kg m^{-3} (Table 2). A relative

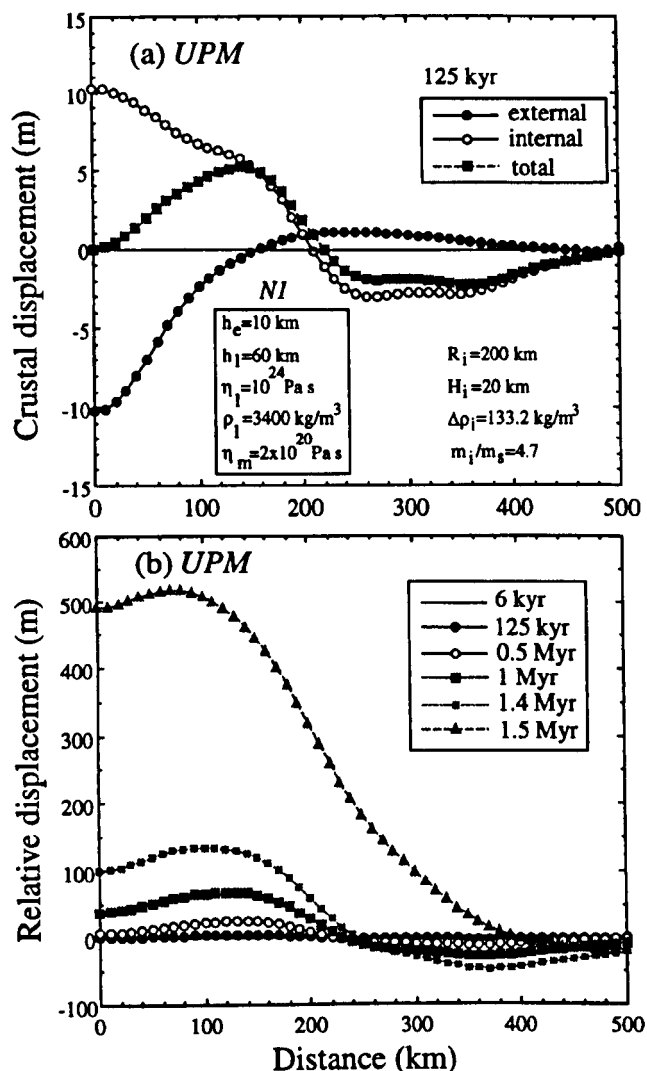


Figure 8. (a) Crustal displacement during the past 125 kyr due to external and internal loads, and both loads, and (b) relative displacement by both loads for the periods shown in this figure. Density of the LPL is 3400 kg m^{-3} , and its thickness is 60 km. The earth model used here is N1 (see Table 2). The density of the upper mantle beneath the LPL is 3300 kg m^{-3} . UPM is the internal load model.

displacement of greater than 10 m is obtained for load models with a density anomaly of less than 12 kg m^{-3} for this earth model. For example, the relative displacement and density anomaly are 12.3 m and 11.8 kg m^{-3} , respectively, for a model with an asthenospheric viscosity of $1 \times 10^{20} \text{ Pa s}$ and a thickness of 100 km. Thus the observed relative displacement of greater than 10 m can be explained by an earth model with an LPL thickness of greater than 60 km and a viscosity of about 10^{24} Pa s . This is the main conclusion of this study. It is, however, true that a low-viscosity asthenosphere effectively reduces the internal mass anomaly, as was indicated by Yuen & Fleitout (1985), Olson *et al.* (1988) and Monnereau *et al.* (1993).

The relative displacements for several earth models are illustrated in Fig. 11 (see also Fig. 12) in order to examine the effect of the thickness of the top elastic layer. UMC80 is adopted as an internal-load model. The thickness, viscosity and density for the LPL are respectively assumed to be 60 km,

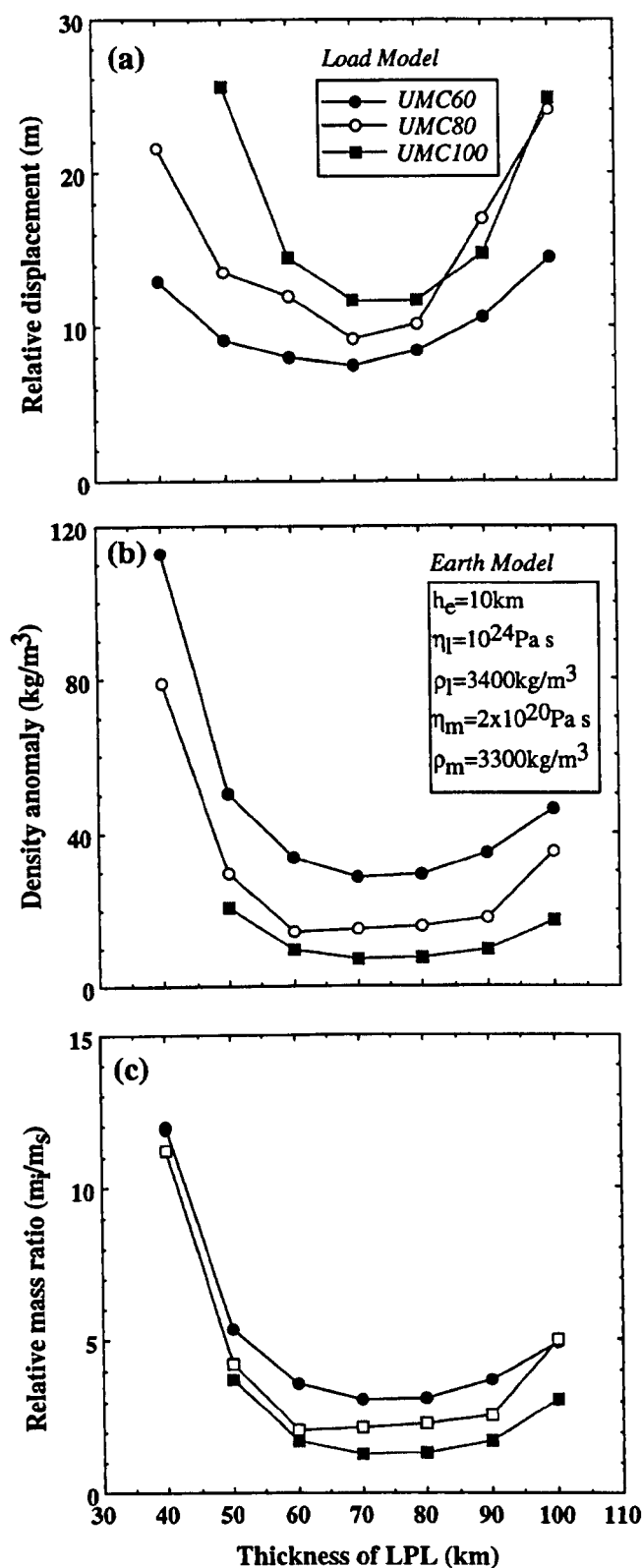


Figure 9. Maximum relative displacement at a distance of about 200 km (a), and density anomalies of internal load and relative mass ratio (b) for UMC load models. Model parameters for the earth and load models are shown in (a) and (b). It is remarkable that a relative displacement of greater than 10 m is predicted for internal loads with a relative mass ratio of less than 2 (density anomaly of smaller than 20 kg m^{-3}) in the model with $\rho_l > \rho_m$ (see Figs 4 and 5).

10^{24} Pa s and 3400 kg m^{-3} . These figures suggest that there is a trade-off between the thickness of the top elastic layer and asthenospheric viscosity. The earth model with a 15 km thick elastic layer and an asthenospheric viscosity of $5 \times 10^{19} \text{ Pa s}$ is also an acceptable model satisfying the prescribed conditions for the relative displacement and density anomaly. As is shown in Figs 11 and 12, the uplift observations for Mangaia, Mauke, Atiu and Mitiaro are reasonably explained for these earth models (see Fig. 1). The magnitude of subsidence at a distance of about 100 km generally decreases as the thickness of the top elastic layer increases, as is seen in these predictions. This subsidence, if observable, is a diagnostic point to verify the model examined here. The subsidence at a distance of 240 km may explain the geomorphological evidence of no raised last-interglacial coral reefs for Aitutaki (see Figs 11 and 12). In fact, the distance between Rarotonga and Aitutaki is slightly longer than that between Rarotonga and Mangaia, although this may depend on whether the centres of the internal and external loads coincide.

To conclude this section, we discuss the relationship between surface deformation and the deformation at the base of lithosphere, using the results shown in Fig. 12. For these model calculations, UMC80 is adopted as an internal-load model. The earth model is N1 with an LPL thickness of 60 km and a viscosity of 10^{24} Pa s . Fig. 12(a) illustrates each component for present-day surface deformation by external and internal loads, and by both loads. The total crustal displacement during the past 1.5 Myr is about 75 m at the central region of load, and about 200 m at a distance of 160–180 km. This corresponds to the swell associated with the external load and lithosphere–small-scale convection interaction. Thus an arch 200 m above the regional seafloor is expected for the area of Mangaia if the model examined here is valid. The relative displacement during the past 125 kyr at the surface is illustrated in Fig. 12(b), and that at the base of the lithosphere is shown in Fig. 12(c). Deformations at 1 and 0.5 Myr BP and at present at the base of lithosphere are depicted in Fig. 12(d). The deformations at both boundaries are correlative. Surface uplift therefore corresponds to the upward deformation of the bottom boundary of lithosphere. This indicates that the surface elastic plate is dragged by the viscous flow of the LPL associated with the Rayleigh–Taylor instability. This is consistent with the model prediction that the magnitude of surface deformation becomes smaller in an earth model with a top elastic layer of 20 km, which has a greater flexural rigidity than a model with an elastic thickness of 10 km (see Fig. 11). The normal-mode method used here as an analytical method is, however, only valid for small perturbations in the various layer boundaries, and is deemed to be applicable for an amplitude (ζ) to wavelength (λ) ratio of $\zeta/\lambda < 0.1$ (Ramberg 1981). For the model illustrated in Fig. 12(d), $\zeta/\lambda \sim 0.08$. In order to discuss these problems in detail, we therefore need finite-amplitude analyses.

DISCUSSION AND CONCLUSION

In this paper, we have evaluated the viscosity structure of the oceanic lithosphere by using the late Quaternary differential crustal movement for the southern Cook Islands. In previous studies, McNutt & Menard (1978) and Lambeck (1981a) estimated the effective elastic thickness of the lithosphere to be 10–15 km on the assumption that the observed total uplift

Table 2. Results for the earth models with $\rho_l > \rho_m$ for internal load model UMC80.

η_u (Pa s)	η_a (Pa s)	h_a (km)	η_l (Pa s)	ρ_l (kg/m ³)	h_l (km)	$h_{\max}$ (m)	$\Delta\rho_i$ (kg/m ³)	m_i/m_s
LPL density (ρ_l) dependence								
2×10^{20}			1×10^{24}	3350	60	7.5	28.8	4.1
2×10^{20} (model N1)			1×10^{24}	3400	60	10.1	14.6	2.1
2×10^{20}			1×10^{24}	3350	80	6.0	23.8	3.4
2×10^{20}			1×10^{24}	3400	80	10.2	16.0	2.3
LPL viscosity (η_l) dependence								
2×10^{20}			5×10^{23}	3400	70	16.2	26.0	3.8
2×10^{20}			1×10^{24}	3400	70	9.2	15.1	2.2
2×10^{20}			2×10^{24}	3400	70	5.2	8.5	1.7
asthenospheric viscosity (η_a) dependence								
2×10^{20}	5×10^{19}	100	1×10^{24}	3400	60	13.8	6.9	1.0
2×10^{20}	1×10^{20}	100	1×10^{24}	3400	60	12.3	11.8	1.7

η_u : upper mantle viscosity, η_a : asthenospheric viscosity, h_a : asthenospheric thickness, η_l : LPL viscosity,

ρ_l : LPL density, h_l : LPL thickness, $h_{\max}$: maximum uplift during past 125 kyr,

$\Delta\rho_i$: density anomaly of internal load, m_i/m_s : relative mass ratio

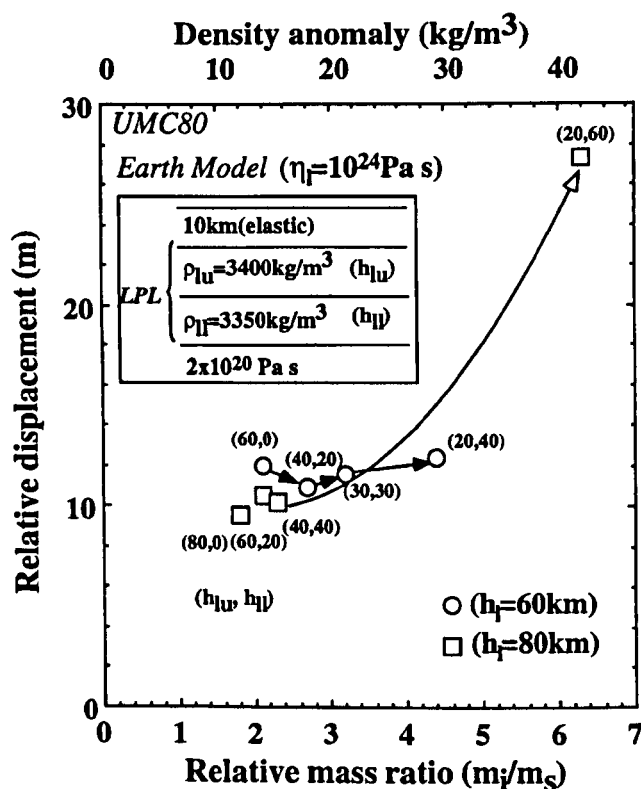


Figure 10. Relationship between the relative displacement during the past 125 kyr and relative mass ratio (density anomaly) for the earth model with a depth-dependent density structure for the LPL. Open circles and squares represent model results for LPL thicknesses (h_l) of 60 and 80 km, respectively. The LPL with a viscosity of 10^{24} Pa s is divided into two layers. The density of the upper layer of LPL is 3400 kg m^{-3} and that of the lower layer is 3350 kg m^{-3} .

for islands such as Mangaia is attributed to the viscoelastic arching by the volcanic loading of Rarotonga about 1.5 Myr ago. Pleistocene and Holocene sea-level observations support the idea that the stress relaxation associated with this volcanic loading is still proceeding in the lithosphere (e.g. Woodroffe *et al.* 1991). Time-dependent crustal movement related to the viscous relaxation within the lithosphere therefore provides constraints on the viscosity structure for the lower part of the lithosphere.

The observed crustal uplift for the past 125 kyr is 14.5–20 m for Mangaia, and that for Rarotonga is 3.5 m. Crustal uplift of approximately 3 m is due to glacial isostatic adjustment (Lambeck & Nakada 1992). These studies imply that Mangaia has been uplifting by another mechanism, and Rarotonga has apparently been stable for this period. This tendency is confirmed by observations of the crustal movement during the past 6 kyr. If the observed uplift for Mangaia, Mauke, Atiu and Mitiaro is attributed to the scenario described by McNutt & Menard (1978), then the magnitude of crustal subsidence for Rarotonga should be greater than 100 m for the past 125 kyr and greater than 5 m for the past 6 kyr. There is, however, no evidence supporting such significant subsidence for Rarotonga. An internal buoyant load, by which surface subsidence due to volcanic load is cancelled for the central region of the external load, is required in the upper mantle.

Two types of internal-load model are examined in this study. One is a model with density anomalies just below the lithosphere (underplating model), leading to surface uplift by its buoyancy force. Fleitout & Moriceau (1991) adopted this model to examine the lithospheric response to the thinning of the lithosphere proposed by Crough (1978). The other model examined here has an internal load within the uppermost mantle, which corresponds to the small-scale and secondary convection in the asthenosphere leading to thermomechanical lithospheric erosion. In particular, Yuen & Fleitout (1985), Olson *et al.* (1988) and Monnereau *et al.* (1993) find the

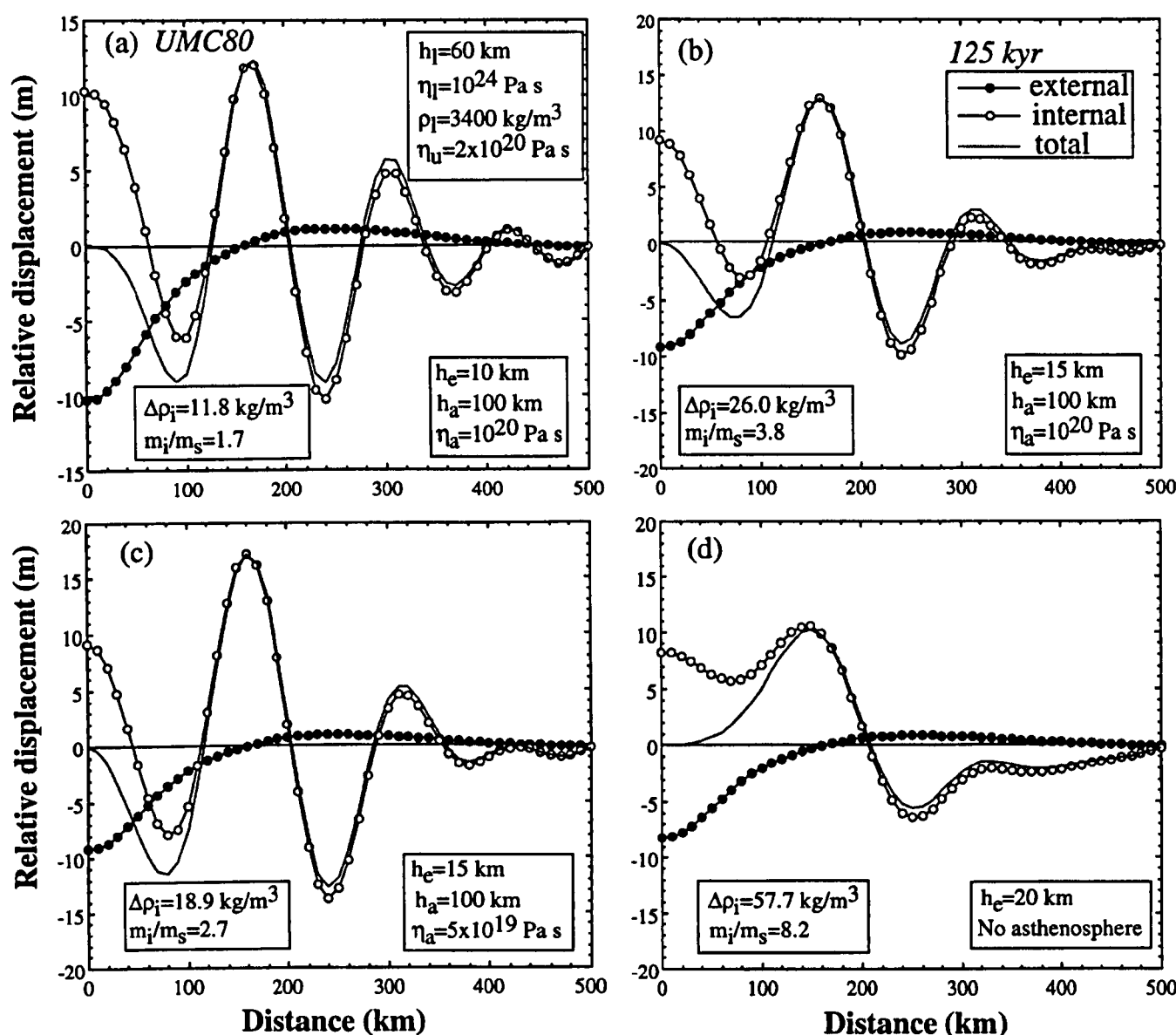


Figure 11. Relative displacement during the past 125 kyr due to external and internal loads, and both loads. The internal-load model is UMC80. Model parameters for earth and load models are shown in each figure.

erosion process to be rather slow unless the viscosity under the lithosphere is very low. They propose a low-viscosity asthenosphere due to a temperature- and pressure-dependent rheology. Such a model makes it possible to produce the rapid timescale of the swell uplift of the order of 10 Myr required for the formation of the Hawaiian swell (e.g. Yuen & Fleitout 1985). Numerical evaluation in this study, however, suggests that surface uplift due to lithospheric erosion is mainly determined by the intrinsic instability of a denser thermal lithosphere overlying a less dense asthenosphere, and by the viscosity structure of the lithosphere. This instability will increase with distance from the ridge because the thickness of the oceanic lithosphere increases with the square root of the age of the seafloor. Secondary convection confined to the low-viscosity asthenosphere may play a role in effectively eroding the lower part of lithosphere.

A very large density anomaly, which is significantly larger than that inferred from the lateral temperature variations of 150–200 K (e.g. Kaula 1983), is required regardless of internal-load model when the density of oceanic lithosphere (ρ_l) is the same as that for the layer beneath the lithosphere (ρ_m). This is due to the lack of intrinsic instability for such a model of oceanic lithosphere. The observed differential crustal movement for the southern Cook Islands is explained for the earth model with $\rho_l > \rho_m$ and for local convection in the upper mantle induced by density anomalies of less than 20 kg m^{-3} . The density anomaly is efficiently reduced for an earth model with low-viscosity asthenosphere. This means that intrinsic instability of mature lithosphere leading to lithospheric erosion is excited by the convective flow confined to the low-viscosity asthenosphere. The rheological structure of the oceanic lithosphere evaluated here is as follows. A high-viscosity layer with

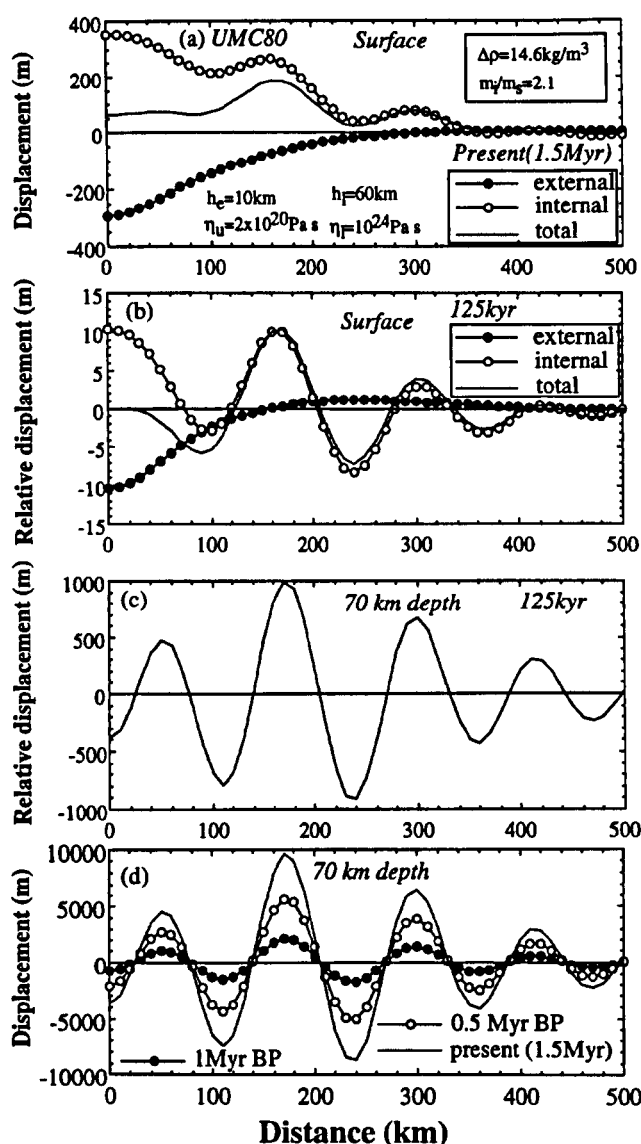


Figure 12. Relationship between surface deformation and the deformation at the base of lithosphere due to internal and external loads. Model parameters for the earth model (model N1) are shown in (a), and the internal-load model is UMC80. (a) Surface deformations at present due to external and internal loads, and both loads, (b) relative displacement during the past 125 kyr, (c) relative displacement at the base of lithosphere during the past 125 kyr, and (d) lithospheric bottom deformations at present, 0.5 Myr BP, and 1 Myr BP.

an effective viscosity of about 10^{24} Pa s and with a thickness of greater than 60 km is required beneath the top elastic layer of 10–15 km thickness. If the viscosity of the LPL is greater than 10^{24} Pa s, then very small crustal uplift is predicted for Mangaia, and vice versa. Thus Pleistocene crustal movement for Mangaia and Rarotonga is indicative of an LPL viscosity of about 10^{24} Pa s. This is the main conclusion of this study. The age of seafloor around this region is about 80 Ma (Calmant 1987), and the thickness of thermal lithosphere is approximately 80–90 km, regardless of its definition (Richardson *et al.* 1995). This suggests that its major part is composed of a viscoelastic layer with an effective viscosity of 10^{24} Pa s and

with a relaxation time of about 1 Myr. Our estimate of the relaxation time is consistent with the value inferred from the outer topographic rise (e.g. Ida 1984). The swell height for the model proposed here is 100–200 m during the past 1.5 Myr. This rapid uplift is comparable to that for the Hawaiian swell estimated by Yuen & Fleitout (1985), Olson *et al.* (1988) and Monnereau *et al.* (1993), although its spatial scale is significantly larger than that obtained here.

Pleistocene sea-level observations using U-series dating have been obtained for Oahu and Hawaii islands. Hawaii and Oahu correspond to Rarotonga and Mangaia, respectively, in our loading model. In fact, a series of submerged coral reefs off northwestern Hawaii is found, which have been subsiding at a roughly uniform rate of 2.6 ± 0.4 mm yr⁻¹ for the past 475 kyr (Ludwig *et al.* 1991). Thus the submerged coral reefs from -430 m correspond to the last interglacial reefs. On the other hand, Muhs & Szabo (1994) reported that last-interglacial marine deposits determined by U-series dating, which include corals in growth position, occur at elevations of 8.5–12.5 m above the present level. Tectonic uplift for Oahu is therefore estimated to be 5–10 m because the sea-level variation due to glacial rebound is about 3 m (Lambeck & Nakada 1992). Tectonic subsidence for Hawaii is 430 m because of the negligible magnitude of the glacial rebound component. These observations are indicative of significant differential crustal movement for these islands. Pleistocene crustal movement for the area around Hawaii is very different from that for Rarotonga.

The difference may be explained as follows. Two characteristic bathymetric topographies have been observed for the Hawaiian swell (e.g. Crough 1978). One is a topographic relief of the Hawaiian swell with a lateral scale of 1500 km, which may have attained steady-state. The other characteristic feature is a moat and arch system with a lateral scale of about 500 km responsible for the lithospheric deformation by the volcanic load. These bathymetric features and crustal movement during the past 150 kyr may indicate that stress relaxation due to the external load is still proceeding in the lithosphere. In particular, the large ratio of subsidence to uplift is indicative of stress relaxation in a relatively thin lithosphere. The relaxation time is 1 Myr for the LPL with the viscosity of 10^{24} Pa s estimated here. Thus recent crustal movement, which is related to the arch and moat system with a representative wavelength of 500 km, may be attributed to the stress relaxation of the upper part of the lithosphere. Then the ratio of subsidence to uplift will increase as the thickness of the lithosphere acting as a viscoelastic layer becomes thinner. Studies including observations of bathymetry, gravity and crustal movement with accurate dating data will be required for evaluating the rheological structure of the oceanic lithosphere and for examining the mechanism of swell formation.

In most cases, we can observe the final state where the load is being supported by the top elastic layer of the lithosphere. This layer may be a minor part of the thermal lithosphere in the framework of mantle convection. Thus the determination of the viscosity for the lower part of the oceanic lithosphere is very important in modelling mantle convection with realistic subducting slabs satisfying surface expression and seismic tomographic observations of plate margins. Late Quaternary sea-level variations in this region provide an important constraint on the rheological structure of the thermal lithosphere. It is also stressed that the crustal and upper-mantle structures

in this region are required in order to put an independent constraint on the rheology of the thermal lithosphere.

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