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Sea Level Change From Mid Holocene to Recent Time: An Australian Example with Global Implications

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Observed relative sea-level change reflects changes in ocean volume, glacio-hydro-isostasy, vertical tectonics and redistribution of water within ocean basins by climatological and oceanographic factors. Together these factors produce a complex spatial and temporal sea-level signal. For the tectonically stable Australian margin, geological evidence indicates that sea-levels at 7000-6000 years ago were between 0 and 3 m above present level, due primarily to glacio-hydro-isostatic effects of the last deglaciation. The spatial variability of this signal determines the mantle response to the surface loading and leads to an effective lithospheric thickness of 75-90 km and an effective upper mantle viscosity of $(1.5-2.5)\times10^{20}$ Pa s. Compared with results for other regions this is indicative of regional variation in upper-mantle response. Also, ocean volumes continued to increase after 7000 years ago by enough to raise global mean sea level by about 3 m. Much of this increase occurred between 7000 and 3000 years ago. Because of the spatial variability in mantle response, isostatic corrections to tide-gauge records of recent change should be based on regional model-parameters rather than on global parameters. The two longest records from the Australian margin give an isostatically corrected rate of regional sea-level rise of 1.40 $\pm$0.25 mm/year. Comparisons of this rate with rates from other regions indicates that the spatial variability in secular sea-level is likely to be significant, with estimates of regional rates ranging from about 1 mm/year to 2 mm/year. These rates of secular change cannot have persisted further back in time than a few hundred years without becoming detectable in high-resolution geological and archaeological indicators of sea-level change.

1. INTRODUCTION

Changes in sea level are usually expressed as a change in the level of the sea with respect to land. It is therefore a relative measure that is indicative of movement of the land, changes in ocean volume or, in most cases, of both. As such, sea-level change, both today and in the past, exhibits a complex spatial and temporal pattern that reflects tectonic, isostatic and climate contributions. Future change, likewise, will be geographically variable. Separation of the contributing factors becomes important when examining present change or when predicting future trends from incomplete observations. How much of an observed signal is of local relevance only because of tectonic contamination? How much of the signal is of global relevance, indicative of changes in ocean volume? Under some very special circumstances it becomes pos-

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sible to separate these contributions, but the present observational data base is mostly inadequate for this and reliance on numerical model results or on indirect indicators of change is usually required.

This paper examines this question of separability, using geological and tide gauge sea-level evidence from the relatively stable Australian margin. In particular, it examines the separability of glacio-hydro-isostatic contributions to sea-level from changes in ocean volume over the past 7000 years. (Unless otherwise indicated ages are expressed in calendar years. Palaeo sea-level data will be expressed mostly with respect to the radiocarbon ^{14}C time scale. Any conversion to calendar years is based on calibration by *Stuiver et al.*[1998] and*Bard et al.*[1998]. For the interval in question here, and within dating uncertainties, the difference between the two time scales is effectively linear and either is adequate for modelling, provided that all time-dependent functions (sea-level data, ice retreat history and viscosity) are defined in the same system.)

Evidence abounds that sea levels have changed by a few meters over the past few thousand years even though most of the melting of the large ice sheets was complete by about 7000 years ago. This evidence is in the form of raised shorelines, fossil corals above their current growth position, or submerged in-situ tree stumps and other normally sub-aereal vegetation or fauna. Such observations may be indicative of vertical tectonic movement of the land, of on-going changes in ocean volume, or of the on-going isostatic response to the last deglaciation of the high-latitude ice sheets. If the interest in sea-level data is the isostatic rebound or climate signal, then the tectonic component needs to be eliminated. If the interest is in estimating tectonic rates of vertical movement then the rebound and climate contributions must be removed. The importance of tectonic contributions can sometimes be assessed from the geological evidence: is there an independent record of crustal deformation that serves as warning that sea-level data from the locality may be contaminated by a tectonic signal? Can the position of the Last Interglacial (LIg) shorelines be identified and used as an indicator of tectonics? Globally LIg shorelines occur within a few metres of present sea level and if they are found to be well-elevated or submerged they points to tectonic uplift having occurred during the last 125000 or so years.

The margins of the old continents, away from tectonic plate boundaries and away from the former ice sheets, provide some of the best information on recent sea level change largely free of tectonic contributions. Along the Australian margin, for example, there is substantial evidence that sea levels have been up to three meters higher in the past few thousand years and that sea level has been dropping quasi-uniformly since that time (see, for example, papers in *Hopley* [1983]). Two results are illustrated in Figure 1. The first result, based on the past sea level recorded within chenier ridges, is from the Gulf of Carpentaria [*Chappell et al.*, 1982] and indicates that here sea level has fallen by about 2.5 m during the past 5000 (radiocarbon or ^{14}C) years. The second result, from near-shore Orpheus Island in northern Queensland, is based on fossil in-situ corals that have been stranded by a retreating sea and here sea levels have fallen by only about 1 m over a similar period (*Chappell et al.* [1983] and unpublished ANU data). In both instances the evidence points to sea levels having peaked at about 5000 to 6000 ^{14}C years ago and that before this time levels had been rising quite rapidly. This earlier rise can be seen in Figure 2 which summarises age-height data of coral samples from many localities on the Great Barrier Reef and the upper envelope represents a first approximation to the regional sea-level curve for the past 10000 ^{14}C years. Elsewhere along the Australian margin the high-stand occurs up to 3m elevation or higher [*Burne*, 1982]. In Tasmania the mid-Holocene high-stands appears to be absent. The amplitudes also vary across the shelf, reaching their maxima for coastal sites or localities at the head of narrow gulfs, such as Spencer Gulf (see Figure 1 for locations), and decreasing with distance from shore. In northern Queensland, for example, high-stands of about 1.5 m occur at coastal sites such as Yule Point [*Chappell et al.*, 1983], less for the near-shore islands such as Orpheus, and absent from the outer reef about 100 km offshore. At the outer reef corals at the mean-low-water tide level are all much younger than 5000-6000 ^{14}C years [*Hopley*, 1982]. While the precise details of the highstand morphology remains to be established for most locations, a general observation is that the greater its amplitude the more sharply will it be defined and the earlier the occurrence of the maximum level (see *Lambeck and Nakada* [1990] for a discussion of the Australian margin data then available).

Elsewhere, different patterns emerge. For tectonically stable islands in the Pacific the high-stands are usually low, less than about 1 m, and tend to occur later than along the continental margins (e.g. *Pirazzoli and Montaggioni* [1988] and *Moriwaki et al.* [2000]) Figure 3. The interpretation of these results is often complicated by the possibility that some vertical land motion may also occur that is not associated with glacio-hydro isostasy. In particular, Pacific island sites near the tectonic-plate margins should be avoided if the search is for climate or isostatic signals. But even within the

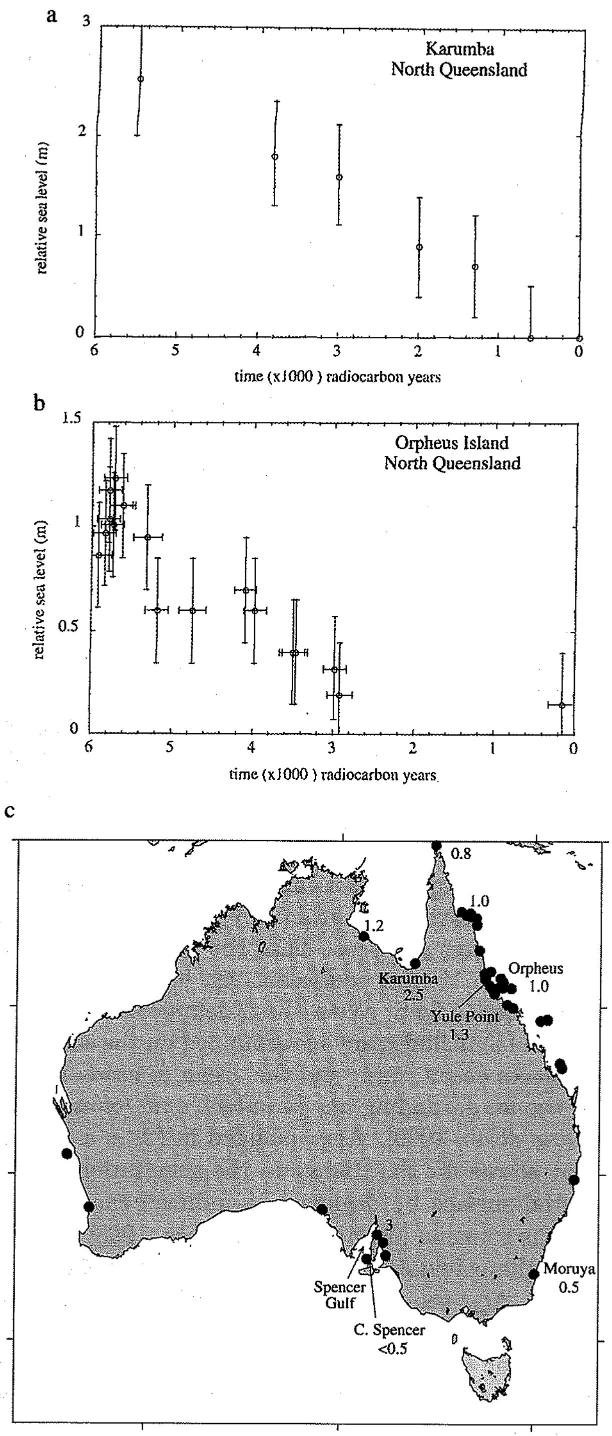

Figure 1. Observed sea level change at two localities in northern Queensland, Australia, for Late Holocene time. The time scale is in ^{14}C years. (a) Near Karumba in the Gulf of Carpentaria [*Chappell et al.*, 1982] and (b) near-shore Orpheus Island (*Chappell et al.*, [1983], and unpublished ANU data). (c) Map showing site locations where late-Holocene highstands have been observed. Numbers indicate the maximum elevations (in metres) observed for some of the sites.

plate interior some vertical movements may result from nearby volcanic loading of the crust, such as some of the islands in the Cook, Society, and Tuamotu groups, or from the motion of the plate over mantle hot spots.

2. THE STRUCTURE OF THE FAR FIELD MID HOLOCENE HIGH STAND.

The small high-stands that develop in mid-Holocene time at locations far from former ice margins - at far-field sites - are largely a result of the planetary response to the last glacial cycle and as such these signals provide significant information on the mantle rheology as well as on any late-Holocene changes in ocean volume. Far from the former ice margins, the first order change in sea level due to the melting of a land-based ice volume V_i is given by the ice-volume-equivalent sea-level function $\Delta\zeta_e(t)$ defined as

$$\Delta\zeta_e(t) = -\frac{\rho_i}{\rho_o}\int_{t_p}^{t}\frac{1}{A_o(t')}\frac{dV_i}{dt'}dt' \qquad (1)$$

where A_o(t) is the ocean surface area and ρ_i, ρ_o are the average densities of ice and ocean respectively. Superimposed on this is the isostatic crustal response to the time-dependent changes in the ice and water loads as mass is exchanged between the ice sheets and the oceans. Thus, in the absence of tectonic contributions,

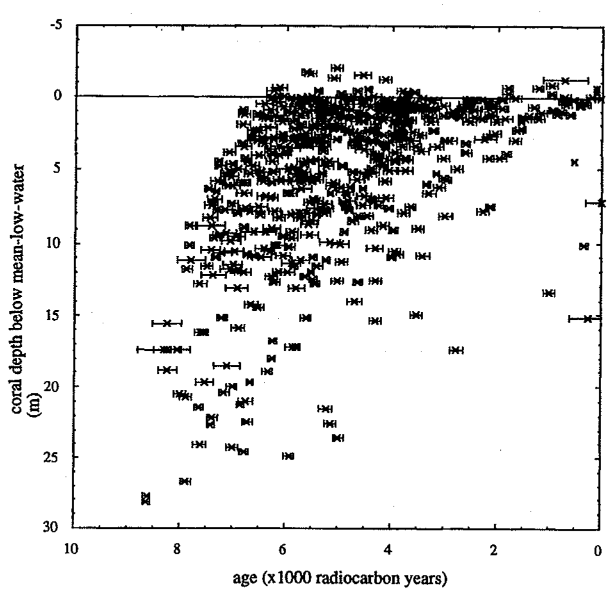

Figure 2. Age-height relationship of corals sampled from reef cores in the Great Barrier Reef region with ages less than 10000 years. The time scale is in ^{14}C years. The height datum is mean-low water. (Data compiled by D. Hopley and D. Zwartz.)

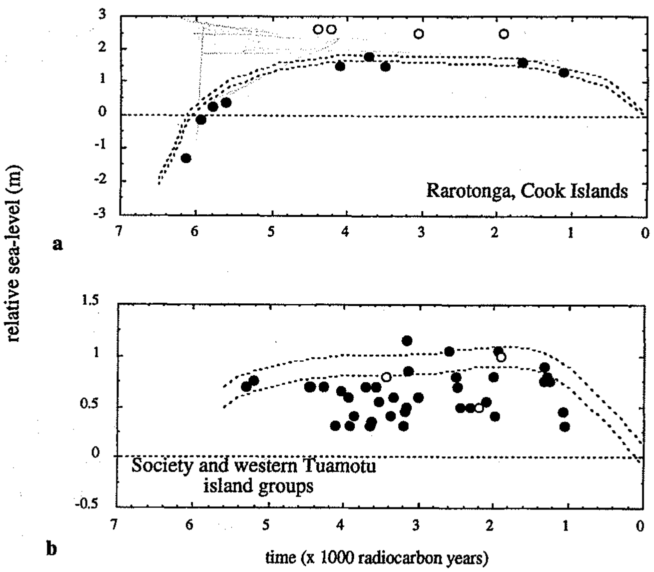

Figure 3. Observed mid- to late-Holocene sea levels in the South pacific. (a) Raratonga, Southern Cook Islands. Solid circles are from microatolls and lagoon and swale deposits and open circles are from sandridge deposits. The estimated mean sea level is shown by the dashed lines (from *Moriwaki et al.*, 2000). (b) Society and western Tuamotu Island groups. Solid circles denote estimates of mean sea level and open circles denote lower limits. The data and the approximate sea-level curve for the region (dashed lines) are from *Pirazzoli and Montaggioni* [1988].

the relative sea-level change can be written schematically as (see *Farrell and Clark* [1976]; *Nakada and Lambeck* [1987]; *Mitrovica and Peltier* [1991]; *Johnston* [1993]; *Milne and Mitrovica* [1998]; *Milne et al.* [1999] for various aspects of this equation)

$$\begin{aligned}\Delta\zeta_{rsl}(\varphi,t) &= \Delta\zeta_e(\varphi,t) + \Delta\zeta_I(\varphi,t)\\ &= \Delta\zeta_e(t) + \Delta\zeta_i(\varphi,t) + \Delta\zeta_w(\varphi,t) + \Delta\zeta_\Omega(\varphi,t) \qquad (2)\end{aligned}$$

where $\Delta\zeta_{rsl}(\varphi,t)$ is the height of the palaeo sea surface relative to present sea level and is a function of position φ and time t. The $\Delta\zeta_I = \Delta\zeta_i(\varphi,t) + \Delta\zeta_w(\varphi,t)$ is the total isostatic response contribution to the sea-level change, including contributions from the deformation of the earths surface and from the changes in gravitational potential. The $\Delta\zeta_i$ and $\Delta\zeta_w$ are the components of $\Delta\zeta_I$ representing the glacio- and hydro-isostatic load effects respectively. In formulating these terms ocean-ice mass is conserved and the ocean surface remains a gravitational equipotential surface throughout [*Farrell and Clarke*, 1976; *Nakada and Lambeck*, 1987]. (That is, these terms include the average over the oceans of the response functions to loading and have been included throughout the three generations of models described by *Nakiboglu et al.* [1983], *Nakada and Lambeck* [1987] and *Johnston* [1993, 1995].) The formulation and programming of the earth-response and sea-level equations has been independently established through these three generations of models and each successive program has been checked against its predecessor. The earth response functions have also been checked against independent code by G. Kaufmann whose formulation of the inversion of the Laplace transformed variables into the time domain was found to be the more robust one on time scales of 10^5 years and longer. Preliminary tests between independent codes developed by M. Nakada and J. X. Mitrovica also indicate agreement of results within the range of expected differences resulting from the use of different ice- and earth-models.

The water depth or terrain elevation at time t and location φ, expressed relative to coeval sea level, is

$$h(\varphi,t) = h(\varphi,t_0) - \Delta\zeta_{rsl}(\varphi,t) \qquad (3)$$

where $h(\varphi,t_0)$ is the present-day (t_0) bathymetry or topography at φ. The isostatic terms in (2) are functions of the earth rheology. In addition, the glacio-isostatic contribution is a function of the ice mass through time and the hydro-isostatic contribution is a function of the spatial and temporal distribution of the water load. The ocean surface area A_0 in (1) is a function of time because of (i) the advance or retreat of shorelines as the relative position of land and sea is modified and (ii) the retreat or advance of grounded ice over shallow continental shelves and seas. Thus the relationship between ice volume V_i and equivalent sea level $\Delta\zeta_e$ is weakly model dependent. With these definitions, the ice volume in (1) includes any ice grounded on the shelves that displaces ocean water and the ocean function is defined by the ice grounding line [*Lambeck and Johnston*, 1998; *Milne et al.*, 1999]. Also included in (2) is a term $\Delta\zeta_\Omega$ that allows for the change in the gravitational equipotential surface by deglaciation-induced changes in the Earths rotation [*Milne and Mitrovica*, 1998].

In the calculation of the isostatic term $\Delta\zeta_I(\varphi,t) = \Delta\zeta_i + \Delta\zeta_w$ the two load contributions are not wholly decoupled as implied by the second equality in the schematic representation (2). Near the ice sheet margin, for example, the gravitational attraction of the ice pulls the ocean water up and increases the water load in the neighbourhood of the ice margins. Also, during the ice loading stage, a broad swell forms beyond the ice margins which, when it occurs in an ocean environment, displaces ocean water and will, ignoring the other contributions, cause sea-level to rise and hence result in an increase in the water load elsewhere. When the ice sheet melts the bulge gradually subsides and the sea levels would fall if other contributions are again ignored. These effects are included in the numerical evaluations of the total isostatic term $\Delta\zeta_I$.

Well beyond the former ice margins $\Delta\zeta_i$ is not negligible but varies relatively slowly with position, independently of whether the site lies at or away from a continental margin. This glacio-isostatic term is the sum of the changes in gravitational attraction from the ice and earth deformation as well as the rebound of the crust peripheral to the ice load. The combined effect is a broad zone, out to distances of 3000 km or more from the ice margin, where the Late Holocene glacio-isostatic contribution is negative. Thus, in the absence of the other contributions, sea levels would be rising throughout late- and post-glacial time (see examples in *Lambeck* [1995; 1996]). Beyond this broad zone the glacio-isostatic signal is again positive but of smaller amplitude.

In contrast to the glacio-isostatic term, the water load term $\Delta\zeta_w$ exhibits a much greater variability because of its strong dependence on the spatial distribution of the added melt water. Its geographic form follows the outline of the coastline, appropriately filtered by the elastic or high-viscosity response of the lithosphere. The main reason for this signal is the subsidence of the sea floor under the increased water load which, through the elastic behaviour of the lithosphere, pulls the coastline down partly. This is accentuated by any uplift of the interior of the continent by mantle flow from ocean to continental mantle. A smaller contribution is from the change in gravitational attraction between the water and land as sea level rises.

In the Australian region the glacio-isostatic term is mainly positive in postglacial times and the two contributions combine to produce a fall in relative sea level for the postglacial period as is illustrated in Figure 4 for Queensland and South Australian localities. It is this and the water-load effect that produces the mid-Holocene high-stand at these sites, the peak high-stand indicating when melting ceased. The results in Figure 4 explain qualitatively the spatial variability observed: At Karumba and Orpheus the glacio-isostatic terms are very similar but the hydro-isostatic terms are substantially different because of the different geometries of the nearby water loads: for Karumba on the Gulf of Carpentaria, the Pacific load lies some 500 km to the east and is effectively an inland one such that the hydro-isostatic effect is considerably magnified. The nearer Gulf of Carpentaria water load, distributed through about 90° azimuth only, further magnifies the effect. For the second site, Orpheus Island on the Pacific coast, the high-stand amplitude is smaller because the coast moves to a greater extent with the ocean floor than it does at Karumba. Likewise, the predicted high-stand amplitude at the head of Spencer Gulf (Port Augusta) exceeds that at the entrance (Cape Spencer) because the former site is effectively 300 km inland.

3. LATE HOLOCENE SEA LEVELS: RHEOLOGY AND OCEAN VOLUME CHANGES.

Because the far-field glacio-isostatic contributions vary only slowly with position, the differential high-stands for near-by sites such as Karumba and Orpheus, or the two ends of Spencer Gulf, are not strongly dependent on details of the ice loads. The differential amplitudes are, however, earth-model dependent as explored by *Lambeck and Nakada* [1990]. Relatively thin lithospheres

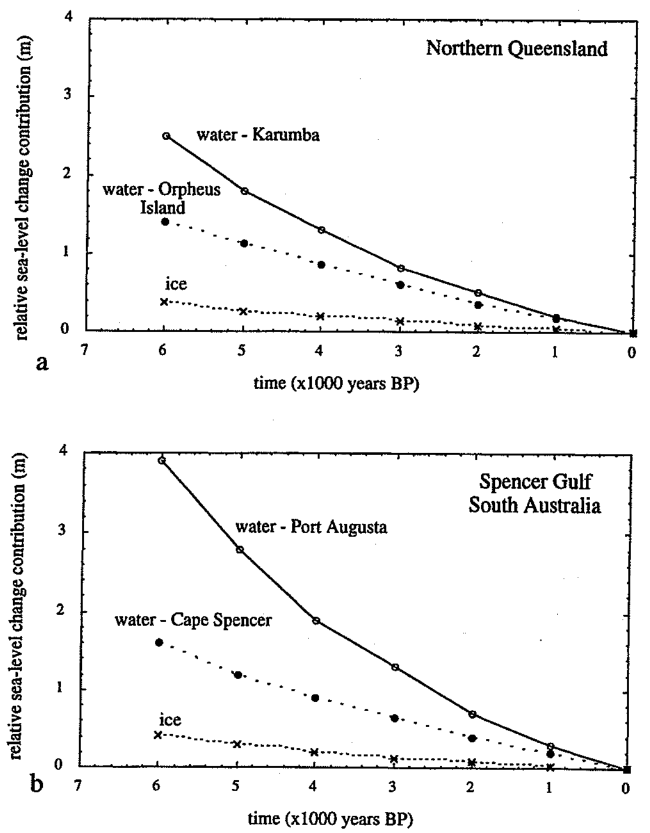

Figure 4. Predicted first-order isostatic contributions to relative sea-level change for the two northern Queensland localities illustrated in Figure 1 and for two localities in South Australia: Port Augusta at the head of the Gulf and Cape Spencer at the entrance to the Gulf. The curves labelled water refer to the hydro-isostatic contribution and the curves labelled ice refers to the glacio-isostatic contribution. This latter contribution is nearly identical for each of the two pairs of sites. (Adapted from *Lambeck and Nakada* [1990].)

and low viscosity upper mantles tend towards enhanced highstand amplitudes and strong gradients of this amplitude across the shelf, whereas thick lithospheres and high viscosity upper mantles reduce both the high-stand amplitude and its spatial gradients. These amplitudes, however, are not a function of rheology alone. They also depend on the melting history of the ice sheets. In the examples illustrated in Figure 4 all melting is assumed to have ceased at 6000 ^{14}C years ago but there is no *a-priori* reason to suppose this. For example, mountain glaciers in the last century have added between 20-30 mm to global sea level rise [*Zuo and Oerlemans*, 1997; *Dyurgerov and Meier*, 1997; *Gregory and Oerlemans*, 1998] and if this had been constant for the past 6000 years the total rise would have been between 1 and 2m. Potentially more important is the ongoing melting of Antarctic ice after cessation of melting of the northern ice sheets at about 6000 ^{14}C years ago. This ice sheet was much larger during the Last Glacial Maximum [*Denton and Hughes*, 1981] and the West Antarctic part, in particular, may have lost up to two thirds of its volume in the subsequent interval [*Bindschadler*, 1998]. In the Antarctic deglaciation model of *Huybrechts* [1994] the melting is largely driven by the rising sea levels that follow from the northern hemisphere deglaciation and much of the Antarctic decay is predicted to have occurred in Holocene time right up to the present. *Conway et al.* [1999] also suggest that the retreat of the West Antarctic ice is still ongoing such that it may contribute up to 0.9 mm/yr to the present sea-level rise [*Bindschadler*, 1998]. Some evidence for recent retreat is seen in new data from Mary Byrd Land where marginal moraines dated at a few thousand years BP occur up to 400 m above the present ice surface (J. O. Stone, private communication).

There is, therefore, no *a-priori* reason for assuming that all melting ceased at 6000 ^{14}C years ago and in interpreting the high-stand amplitudes the possibility of there being also a contribution from a change in ocean volume must be entertained [*Nakada and Lambeck*, 1988]. Thus, in general, the schematic observation equation relating observed sea levels $\Delta\zeta_o$ to the model-dependent predicted values $\Delta\zeta_p$ should be written as

$$\Delta\zeta_o + \epsilon_o = \Delta\zeta_p = \Delta\zeta_e + \delta\zeta_e + \Delta\zeta_i + \Delta\zeta_w \qquad (4)$$

where ϵ_o is the observation error and $\Delta\zeta_e$ is a corrective term to the nominal equivalent sea level function $\Delta\zeta_e$ based on the ice sheets used to evaluate the earth-rheology dependent isostatic components $\Delta\zeta_i$ and $\Delta\zeta_\Omega$. (The rotation term $\Delta\zeta_\Omega$ is ignored in this equation but not in the analysis.) (Equation (4) assumes that the $\delta\zeta_e$ is small such that its neglect in the evaluation of the isostatic terms is negligible. Typically $(|\ \Delta\zeta_i\ | + |\ \Delta\zeta_w\ |)/|\ \Delta\zeta_e\ | \approx (10-15)\%$ so that an error of 10% in $\Delta\zeta_e$ introduces errors of only $1-1.5\%$ in $\Delta\zeta_p$. If $\Delta\zeta_e$ is larger an iterative solution may be required.)

That a solution for both $\Delta\zeta_e$ and the rheological parameters contained in the functions $\Delta\zeta_i$, $\Delta\zeta_w$ is possible can be illustrated as follows. Consider a three-layer mantle model defined by a lithosphere of effective thickness H_l and effective viscosities (η_{um}, η_{lm}) for the lower and upper mantle with the boundary between the two mantle zones at 670 km depth. Consider also observations of the highstand amplitudes $\Delta\zeta_o^{(1)}$ and $\Delta\zeta_o^{(2)}$ at two nearby locations. The difference $\Delta\zeta_o^{(1)} - \Delta\zeta_o^{(2)} \approx \Delta\zeta_w^{(1)} - \Delta\zeta_w^{(2)}$ is independent of the equivalent sea-level function unless the starting ice models from which $\Delta\zeta_w^{(1)}$ and $\Delta\zeta_w^{(2)}$ are evaluated are grossly in error. Thus this differential highstand amplitude is mostly a function of rheology. This dependence is illustrated in Figure 5 for selected pairs of sites as a function of η_{um} and η_{lm} with a lithosphere of H_l= 50 km. The highstands are illustrated here for 6000 years BP (before present) only and superimposed upon the predicted differential values are the observational constraints for the upper and lower limits to the highstands. Shaded areas denote that part of the solution space that is consistent with observational data. For any one pair of sites the response parameters are not well constrained but, because the rheological dependence varies with coastline geometry, the solution based on all the data is well constrained for at least η_{um} (see the last panel of Figure 5). In this case, with an *a-priori* value of $H_l = 50$ km, η_{um}= $(2-3)\times10^{20}$ Pa s and η_{lm}= 10^{22} Pa s although the latter is not well constrained. If a lithospheric thickness of 100 km is assumed then the solution space is distinctly different Figure 6 but also less satisfactory: If the Cape Melville data is excluded two solutions for viscosity are possible but neither satisfy some of the other observational evidence that is not included in constructing Figure 6. In particular, the thicker lithospheric value leads to smaller than observed tilt gradients across the Queensland shelf.

Considerable trade-off occurs between the three earth-model parameters and, in particular if an *a-priori* value for lithospheric thickness is assumed, the resulting parameters for the viscosities are only as reliable as the initial choice. There are no relevant *a-priori* estimates for lithospheric thickness appropriate for the loading frequencies and magnitudes of the glacial cycles. Tectonic estimates, corresponding to long load cycles, result in thinner effective lithospheres because the load time-

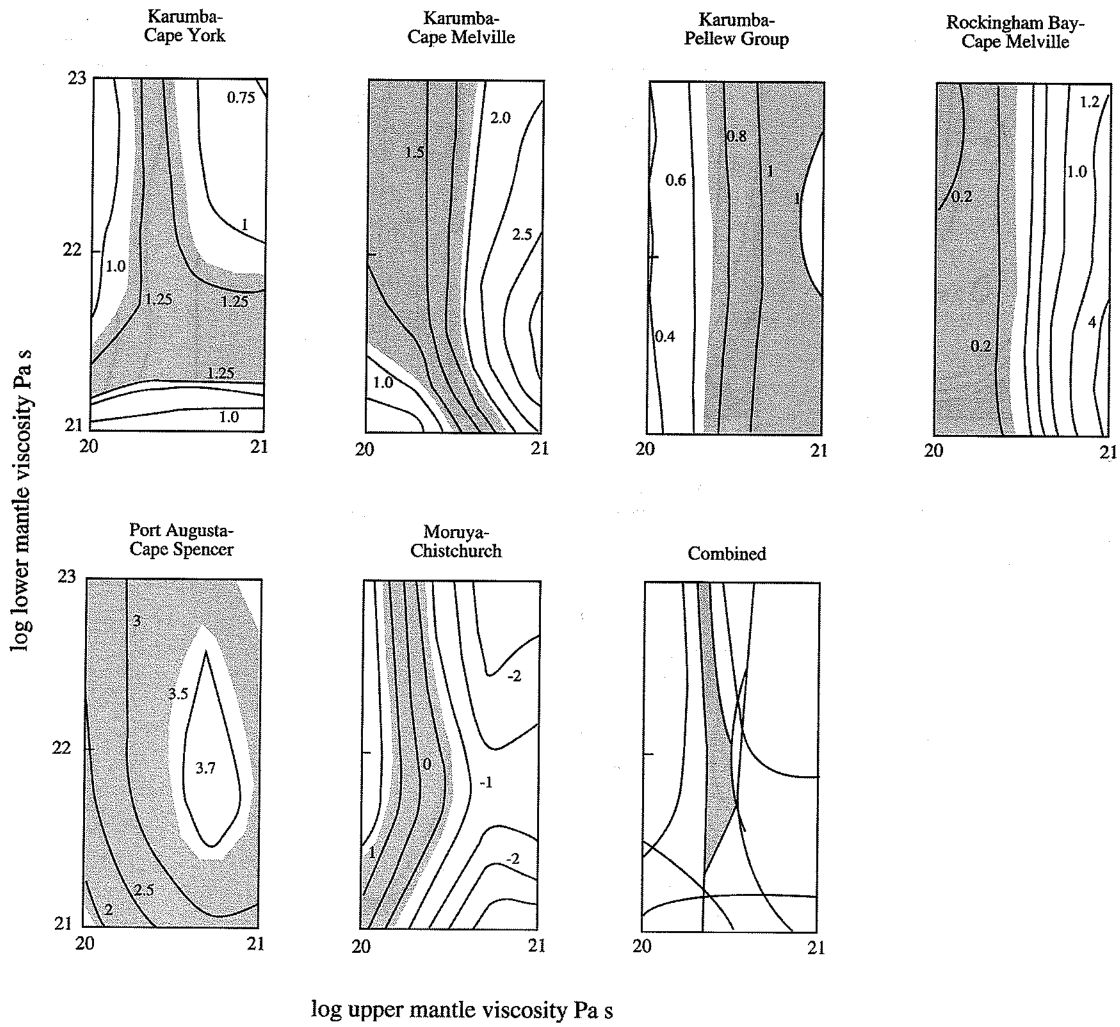

Figure 5. Predicted differential amplitudes for the mid-Holocene highstands at pairs of Australian sites as a function of upper η_{um} and lower η_{lm} mantle viscosities for lithospheric thickness $H_l = 50$ km. The shaded areas correspond to the η_{um} - η_{lm} solution space in which the predicted values lie within the observed range. (Adapted from *Lambeck and Nakada* [1990]).

constant becomes more comparable to the relaxation times for the lower lithosphere and the load stresses migrate with time into the colder parts of the layer. Seismic estimates, corresponding to very high frequency load cycles, give higher estimates for H_l and one of the reasons for attempting to estimate lithospheric thickness from the sea-level data is to explore this dependence of H_l on the load cycle. Hence, when solving for rheological parameters from rebound data, it is not acceptable to assume an *a-priori* value for H_l and this parameter must be considered as an unknown, along with the mantle viscosity parameters.

Once the rheological parameters are determined from the differential observations, the correction term for the ice-volume equivalent sea-level follows directly from (4). Some representative results for Karumba and Orpheus Island are shown in Figure 7(a, b) in which the predicted highstands, based on the above mantle solution and zero change in ocean volume after 6000 years ago, are compared with the corresponding observed values. The former are consistently higher than the observed highstands but the difference remains nearly constant from one location to another. It provides, therefore, a measure of the increase in ocean volume ($\delta\zeta_e$) that has occurred in the past 6000 ^{14}C years. The solution for $\delta\zeta_e$ based on all the Late Holocene data is shown in Figure 7(c) and points to the same general conclusion, that ocean volumes have continued to increase be-

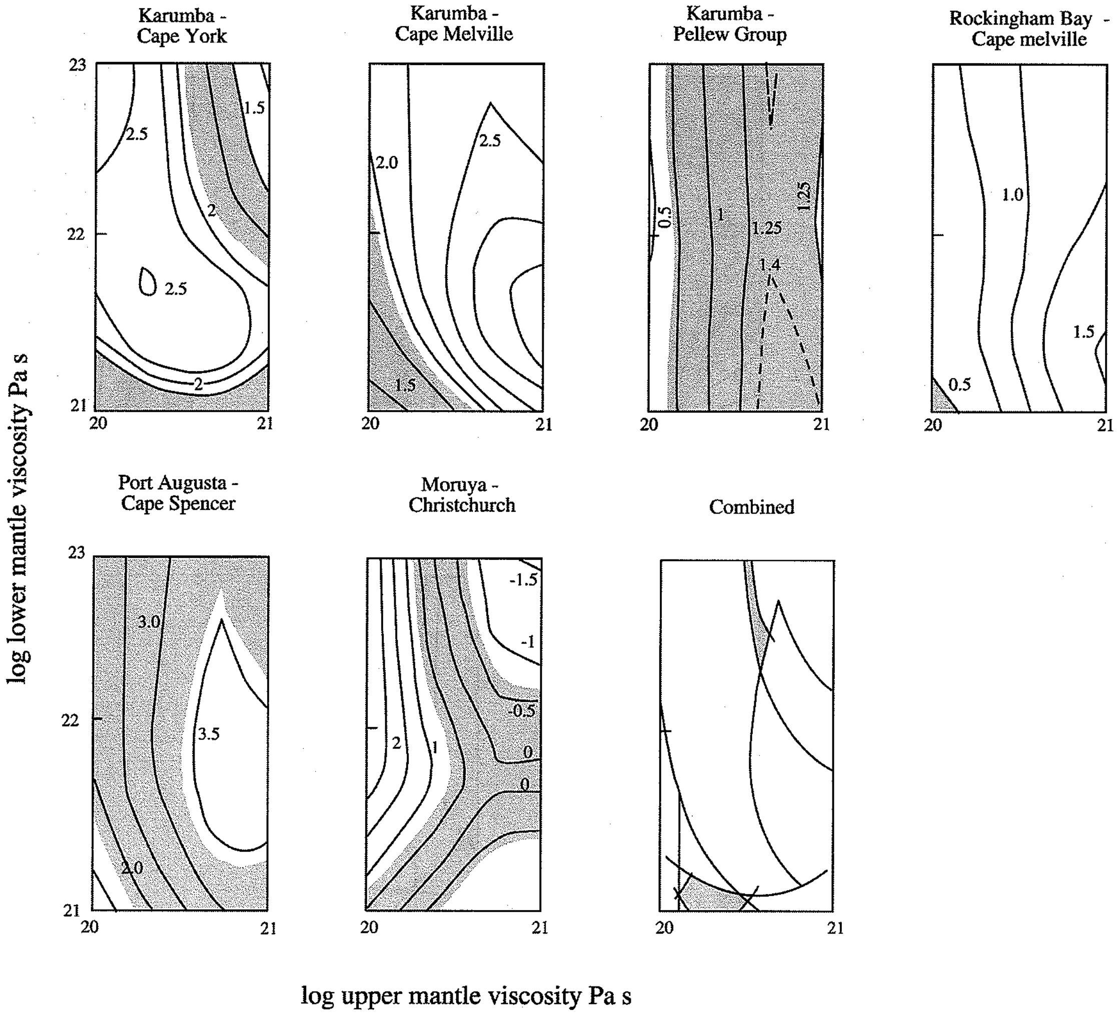

Figure 6. Same as Figure 4 but for H_l = 100 km. No single solution is found in this case that satisfies all the observational data used and the combined solution illustrated does not include the Cape Melville data.

tween about 6000 and 3000 ^{14}C years BP. The function $\delta\zeta_e$ will be earth-model dependent but this dependence is small for models within that part of the parameter space in the neighbourhood of the least variance. Also, iterative solutions in which Late Holocene melting is explicitly introduced into the ice models lead to the same results for earth-model parameters and the solutions appear to converge (c.f. examples in *Lambeck et al.* [1996]).

The results illustrated in Figures 5 and 6 are from *Lambeck and Nakada* [1990] and are based on a first-order solution of the sea-level equation in which the time dependence of the shorelines was treated in an approximate manner only. Also, the solution was based on a small subset of the total data available for the last 6000 ^{14}C years. A more comprehensive search through the model space using the complete formulation for the isostatic terms in equation (4) and a larger data base (about 90 sea level estimates with $t \leq 6000$ ^{14}C years) has therefore been carried out in which the model-parameter search has been limited to 3 mantle layers of which the lower mantle viscosity has been constrained to 10^{22} Pa s, consistent with the previously found value as well as that found for other localities (see Section 5). For a point in the earth-model parameter space, sea-level values are predicted for the time and location of each observed data point, using the ice models with zero melting after 6000 ^{14}C years BP. The corrective term $\delta\zeta_e$ is estimated along with the corresponding root mean square value of the misfit between

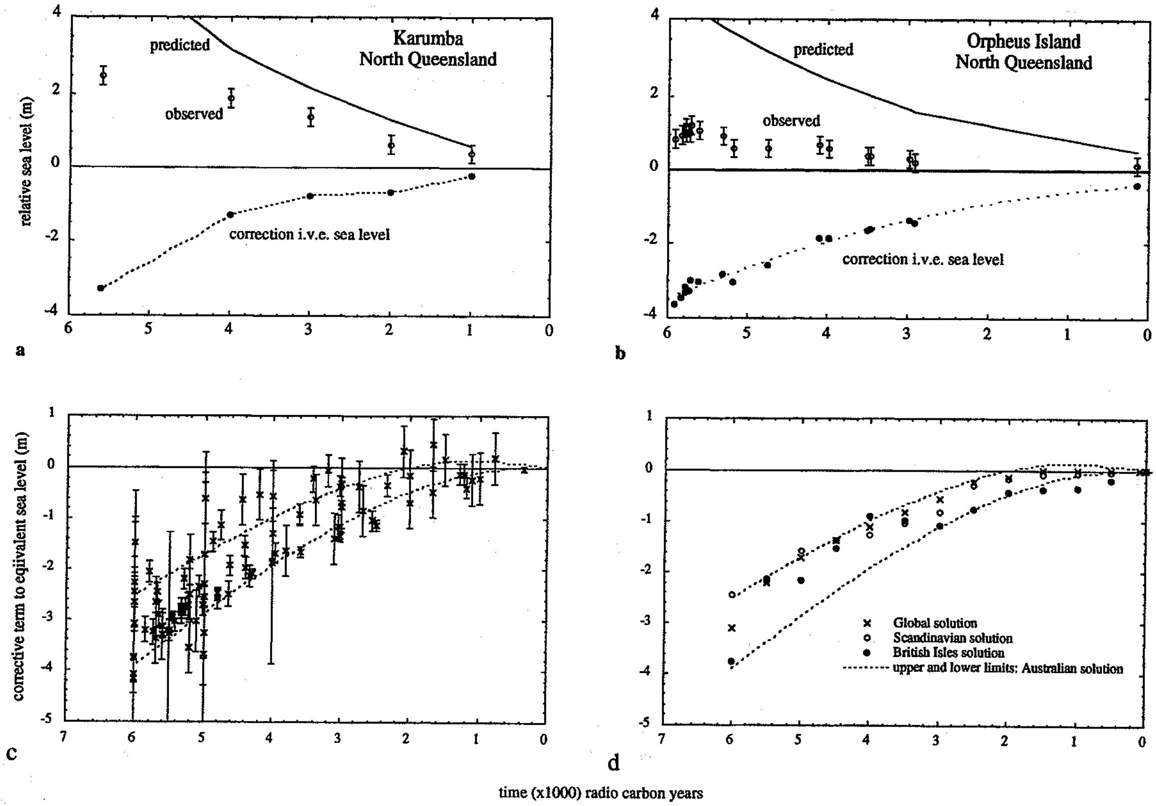

Figure 7. Comparison of predicted and observed late Holocene sea levels based on the earth-model parameters inferred from the differential highstand analysis and on the assumption that no increase in ocean volume occurred after 6000 ^{14}C years BP. The differences correspond to the corrective term $\Delta\zeta_e$ to the ice-volume equivalent (i.v.e) sea level. (a) and (b) illustrate two specific examples. (c) gives the solution for all of the Australian data discussed below. (d) corresponds to similar solutions from other regions. The global solution is from *Fleming et al.* [1998], the Scandinavian solution is from *Lambeck et al.* [1998b] and the British Isles solution is from *Lambeck et al.* [1996].

observed and predicted values. The model parameter space is then searched for the minimum value for this measure of model fit Figure 8. The minimum value found is about 0.5 m, equal to the average accuracy of the observational data, for H_l = 70-100 km and η_{um}= $(1.5-2.5)\times 10^{20}$. Of note is the correlation that occurs between the two earth-model parameters, thick values for H_l leading to higher values for η_{um}, a correlation previously noted in analyses of sea-level data when the evidence is mainly from the post-glacial period [*Lambeck*, 1993a; *Lambeck et al.*, 1996]. The solution for the corrective term to the ice-volume equivalent sealevel is illustrated in Figure 7c and is similar to that previously found by *Lambeck and Nakada* [1990]. Within the observational uncertainties, this function is not strongly dependent on earth-model parameters in the neighbourhood of the optimum solution.

Figure 9 illustrates the predicted mid-Holocene sea levels around the Australian margin based on the above rheology and ice-volume equivalent sea level solution. The predicted coastal sea levels are mostly consistent with the observed evidence (c.f. Figure 1c) as should be the case because this observational evidence was used to constrain the solution for model parameters. Levels at 6000 ^{14}C years ago were generally above present with the exception for Tasmania and some other southern margin localities. The ice sheet models used in these predictions contain a substantial component of melt water from Antarctica and Tasmania lies sufficiently close to this former ice load for the glacio-isostatic signal there to be negative and comparable to the magnitude of the hydro-isostatic contribution. The model predicts well the observed gradients in highstand amplitude observed along the deeply indented gulfs of south Australia, as well as across the North Queensland Shelf and along the shores of the Gulf of Carpentaria. In some other localities, such as the southwest corner of Western Australia, quite rapid changes in spatial vari-

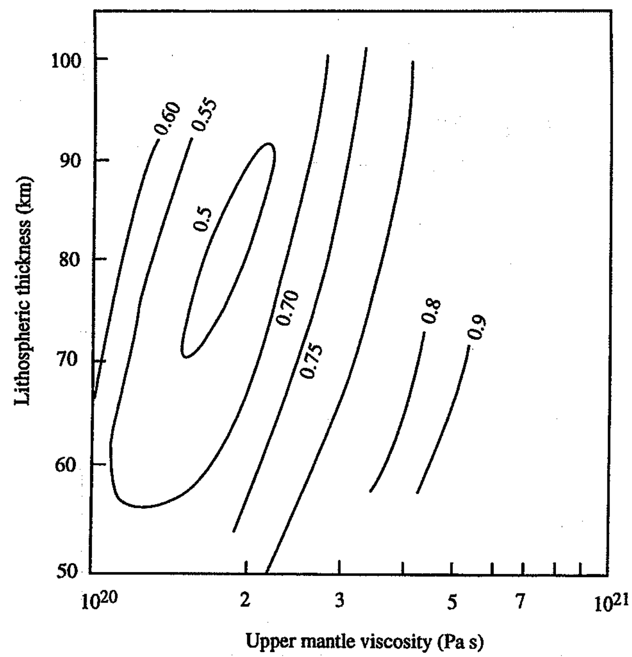

Figure 8. Solution space (H_l, η_{um}) for the Australian late Holocene sea level data for η_{um} = 10^{22} Pa s. Contours represent the rms misfit of the model predictions to observed values.

ability of the highstand amplitude are also predicted, reflecting the variability of the hydro-isostatic signal.

4. RECENT SEA LEVEL CHANGE IN THE AUSTRALIAN REGION.

The Australian margin is tectonically quite stable and the observed Holocene sea-level data is compatible with the predictions of glacio-hydro isostatic rebound models to within observational uncertainties. The geological evidence does not require, therefore, the introduction of a tectonic component. In particular, the gradients in the mid-Holocene highstand observed in South Australia do not require a tectonic origin, being wholly explicable in terms of glacio-hydro-isostasy. The apparent absence of the Last Interglacial shoreline in northern Australia may be indicative of some subsidence [*Chappell*, 1987] but it may also reflect a lack of preservation of former shoreline features. Corals from beach rock at about 5 m elevation on Camp Island, North Queensland, give Late Pleistocene ages [*Hopley*, 1971], as do surface outcrops of reefs on Digby and Marble Islands of the central Queensland coast [*Kleypas*, 1996]. In parts of Western Australia the elevations of the older interglacial surfaces point to the possible occurrence of uplift but at rates below 0.1 mm/year [*Lambeck*, 1987]. Furthermore, the Australian margin is characterised by minimal Quaternary sediment loading such that subsidence due to sediment loads or sediment compaction is also small. In Tasmania, in contrast, the Last Interglacial shoreline appears to be well elevated [*Murray-Wallace*, 1990;*Murray-Wallace and Belperio*, 1991] and is indicative of tectonic uplift at a rate of about 0.2 mm/year. Thus, with the possible exception of Tasmania, most of the Australian mainland margin provides a good platform for measurement of recent sea-

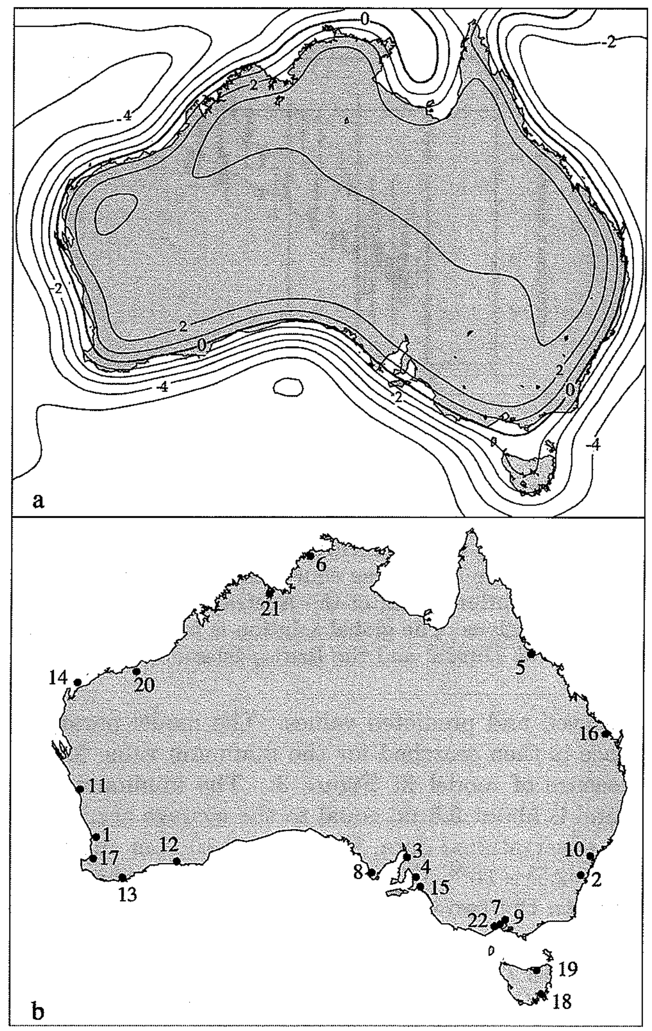

Figure 9. (a) Predicted amplitude of sea-levels around the Australian margin at 6000 ^{14}C years ago based on the glacio-hydro-isostatic rebound parameters and corrective term to the ice-volume equivalent sea-level function derived from the late Holocene sea-level data, Contours are in metres at 1 m intervals. The zero contour is identified by the thicker line. (b) Location of tide-gauge sites with record lengths in excess of 25 years. Numbers refer to the sites listed in Table 1.

Table 1. Summary of observed secular sea-level rates from Australian tide-gauge sites with records longer than 25 years [*Mitchell et al.*, 2000]. Also given are the glacio-hydro-isostatic corrections (Iso. corrn) based on the earth-response function discussed above, and the isostatically corrected (Corr. rate) results. Site locations are shown in Figure 9b.

Site	Tide gauge	Record length (year)	Obs. rate (mm/year)	Iso. corrn (mm/year)	Corr. rate (mm/year)
1	Fremantle WA	90.6	1.38	-0.27	1.65
2	Fort Denison NSW	81.8	0.86	-0.30	1.16
3	Port Pirie SA	63.2	-0.19	-0.46	0.27
4	Port Adelaide SA	55.1	2.07	-0.41	2.48
5	Townsville Qld	38.3	1.12	-0.47	1.59
6	Darwin NT	34.9	-0.02	-0.42	0.40
7	Point Lonsdale Vic	34.4	-0.63	-0.28	-0.35
8	Port Lincoln SA	32.3	0.63	-0.35	0.98
9	Williamstown Vic	31.8	0.26	-0.30	0.56
10	Newcastle NSW	31.6	1.18	-0.30	1.98
11	Geraldton WA	31.5	-0.95	-0.28	-0.67
12	Esperance WA	31.2	-0.45	-0.26	-0.19
13	Albany WA	31.2	-0.86	-0.16	-0.70
14	Thevenard Is WA	31.0	0.02	-0.25	0.27
15	Victor Harbour SA	30.8	0.47	-0.37	0.84
16	Bundaberg Qld	30.2	-0.03	-0.33	0.30
17	Bunbury WA	30.2	0.04	-0.21	0.25
18	Hobart Tas	29.3	0.58	-0.04	0.62
19	George TownTas	28.8	0.30	-0.14	0.44
20	Port Hedland WA	27.7	-1.32	-0.34	-0.98
21	Wyndham WA	26.4	-0.59	-0.45	-0.14
22	Geelong Vic	25.0	0.97	-0.28	1.25

level change free of tectonic factors in excess of about 0.1 mm/year.

With low background tectonic signals and provided that any tide-gauge site is free of local sediment compaction and subsidence, as is not the case for some of the harbour sites, observations from the Australian margin should provide indicative results of recent sea-level change. Such results become particularly important because there are few long records from the southern hemisphere and estimates of global change tend to be very much dominated by mid- to high-latitude records. Table 1 summarises recent estimates of secular sea-level change from a number of tide gauge sites with records longer than 25 years (see Figure 9b for locations). In two instances, Sydney (Fort Denison) and Fremantle, the records are longer than 80 years [*Mitchell et al.*, 2000]. The Sydney record has been carefully compiled, verified and analysed by numerous investigators over the past two decades (e.g. *Hamon*, 1987). The site itself is on rock, remote from harbour installations, adjacent to relatively deep water and unaffected by fresh-water riverine input. It constitutes the best record of sea-level change from the eastern Australian margin and one of the better southern hemisphere records. The Fremantle record is also from a well maintained and documented tide gauge although the levels are weather affected such that long-term trends are noisier than for most other Australian sites. However, with a 90 year record, this enhanced noise should not impact in a major way on the estimates of long term trends. The rates at these two sites do differ significantly, with the Sydney rate being substantially smaller than the Fremantle rate. Surprisingly, neither of these records are included in recent evaluations of global sea-level change [*Douglas*, 1997; *Peltier*, 2000], the Sydney record being discarded because it was found to be inconsistent with results from the Newcastle tide gauges about 100 km north of Sydney. Records are available from three gauges in Newcastle harbour and the result, of about 30 years duration from Newcastle III - the best of the three - is included in Table 1. This result is about 40% higher than the Sydney result but it is also much less satisfactory because of its shorter record length, its marked perturbation by river discharge and the harbours location on sediments of the river floodplain. Two records from Port Adelaide give consistent results but both sites are potentially perturbed by known sediment subsidence of the harbour environment. The nearby Port Pirie record of 63 year

Table 2. Estimates for earth response parameters (effective lithospheric thickness (H_l, and effective viscosities η_{um}), η_{lm} for the lower and upper mantle) from three different regional solution. The Australian results are based on the analysis of the late Holocene data. The three Scandinavian solutions are from geological evidence since Lateglacial time [*Lambeck et al.*, 1998b] , from tide-gauge analyses [*Lambeck et al.*, 1988a] , and from the Baltic Ice Lake shoreline gradients and the timing of the Baltic lake stages [*Lambeck*, 1999]. The British Isles solution is based on Lateglacial and Postglacial geological data [*Lambeck et al.*, 1996]. Lower mantle estimates in brackets are assumed for these particular solutions.

Solution	H_l (km)	η_{um} (10^{20} Pa s)	η_{lm} (10^{22} Pa s)
Australia	75-90	1.5-2.5	(1)
Scandinavia-1	65-85	3-4	0.6-1.3
Scandinavia-2	80-100	4-5	(1)
Scandinavia-3	40-80	2.5-4.5	0.7-1.5
British Isles	65-70	4-5	0.7-1.3

duration, in contrast, gives a distinctly different result: of an apparent fall in sea level.

Considerable variation occurs in the observed rate of sea-level change, reflecting variability of oceanographic, climatological and geological factors, possibly superimposed on issues related to maintaining tide gauges over very long time intervals. Part of the variability may be a consequence of the glacio-hydro-isostatic contributions. These effects are included in Table 1 for the solution discussed above and in the presence of only these effects the predictions are for falling sea levels at the mainland tide gauges at rates from 0.16 to 0.46 mm/year. (These predictions are based on the assumption of constant ocean volume for the duration of the tide gauge records.) For the Hobart site the predicted rate is much less than for the mainland localities, only 0.04 mm/year, consistent with the previously discussed Late Holocene evidence of an apparent absence of mid-Holocene highstands.

The isostatically corrected values given in Table 1, representing the sea-level change in the absence of the glacio-hydro-isostatic contributions, exhibit a similar spatial variability as the observed values and other, more important, factors must contribute to the observed variability. A number of the Western Australian sites, but not Fremantle, indicate a falling sea level over the past three decades. Tectonic causes, operating over time scales of thousands of years and longer, can be ruled out as an explanation. At Geraldton, for example, this would lead to about 4 m of uplift over 6000 years so that the mid-Holocene highstand would be predicted to occur at elevations of about 5-6 m. Also, the Last Interglacial shoreline would be about 80 m above present sea level whereas it actually lies at only 3 m elevation, where it is expected to occur in the absence of tectonic movement [*Stirling et al.*, 1998]. If only the two long records from Sydney and Fremantle are accepted then the average corrected sea-level rise is 1.4 mm/year but such averaging may mask information on causes of a spatially variable sea-level response.

5. SOME GLOBAL IMPLICATIONS OF THE AUSTRALIAN SEA LEVEL RESULTS

Upper mantle viscosity is laterally variable. The Australian rheology solution is compared with similar solutions for other regions in Table 2. (All solutions relate to the ^{14}C time scale and the viscosities are in Pa (^{14}C)s.) The Scandinavian results are based on three different types of data and yield concordant estimates for the effective rheological parameters. For both these and the British Isles solutions the parameter search was conducted through a much larger parameter space than for the Australian solution, including two orders of magnitude for the lower-mantle viscosity and some ice-sheet parameters. For the British Isles, mantle models with a greater degree of layering were also explored but only the three-layered models are compared here to facilitate the comparisons between regional solutions. A wholly independent analysis of the European data, independent in the numerical modelling of the earth response and sea-level change as well as in inversion technique, has led to comparable results [*Kaufmann and Lambeck*, 2000]. In all three regional solutions the lithosphere is considered as an unknown parameter and the corrective ice-volume equivalent sea-level term $\Delta\zeta_e$, as well as some ice-sheet scaling parameters in the northern hemisphere solutions, are included.

Of note when comparing the European and Australian results is that the estimates for effective lithospheric thickness are comparable but that the upper-mantle viscosity appears to be lower for the Australian region than for northern Europe. Because of the partial parameter search conducted, and because only the post 6000 year BP data is used (the earlier data contribute to the separation of the H_l and η_{um} parameters), the Australian result must be considered as preliminary but this difference in η_{um} is nevertheless suggestive of the occurrence of lateral variation in mantle viscosity, as already noted by *Nakada and Lambeck* [1991]. Models of upper-mantle structure from inversion of seismic shear wave or attenuation data [e.g. *Romanowicz*, 1998] indicate that considerable lateral variation in structure occurs and that corresponding lateral variations in mantle temperature and hence viscosity can also be expected. Thus the viscosity results are not qualitatively inconsistent with global seismic evidence of higher values beneath shields and stable continents than beneath oceans. In the absence of global solutions for a laterally variable mantle response to surface loading (but see *Kaufmann et al.* [1997]; *Tromp and Mitrovica* [2000]) the strategy adopted here is to estimate viscosities that are representative of the average regional conditions rather than to carry out a global analysis. At a minimum this has the merit of establishing whether lateral variations in mantle viscosity are likely to be significant.

Ocean volume has increased since mid-Holocene time. Solutions for the corrective term $\Delta\zeta_e$ are not limited to far-field locations and the same procedure of solving equation (4) for both rheological and ocean volume parameters (and ice-sheet parameters) has been used in the rebound analyses of formerly glaciated regions. (In these solutions the sea-level variation is effectively separated into a spatially variable part and a spatially invariable but time dependent part and the former constrains the viscosity and ice distribution and the latter constrains $\Delta\zeta_e$.) The corrective term to the ice-volume equivalent sea-level function has been found to be persistently negative immediately after 6000 ^{14}C years BP (Figure 7d). Together, the evidence points to there having been an increase in ocean volume of about 3m equivalent sea-level rise, with the major part of it having occurred in the interval of about 6000-3000 years BP and that any change subsequent to about 2000-3000 years BP and before the recent rise recorded by tide gauges, was at an average rate of only about 0.1-0.2 mm/year (Figure 7d, see also *Lambeck and Bard* [2000]). The source of this increase cannot be identified from these analyses but a likely major contribution is late melting of the West Antarctic ice sheet as discussed briefly above. The consistency of the results for the various regions indicates that the signal is a global one and not an artifact of analysis procedures; of trade-off between an inadequate modelled regional isostatic response of a nearby ice load and $\Delta\zeta_e$ for example. Also, within observational errors there is no indication of major rapid or abrupt changes in sea level over the last 6000 years by more than about 50 cm at any one time.

Holocene highstands observed at Pacific islands tend to be small in amplitude and to occur later than 6000 ^{14}C years ago [e.g. *Pirazzoli and Montaggioni*, 1988; *Moriwaki et al.*, 2000] (Figure 3). As illustrated schematically in Figure 10, this is a consequence of on-going ocean-volume increase in Late Holocene time. The upper panel illustrates isostatic components for different sites, with the largest values being representative of areas near former ice margins where the glacio-isostatic contribution dominates. The middle panel illustrates the predicted sea levels for the case where all melting ceased 6000 years ago (the ice-volume equivalent sea-level curve is denoted by esl). In this case the predicted mid-Holocene sea-level highstands occur at the time of cessation of melting for all cases. The lower panel is for on-going melting in late-Holocene time. When the isostatic effect is substantial, as in curve 1, this dominates the total sea-level signal and a well-defined highstand develops at the time of cessation of melting. But when the isostatic term is smaller the highstand occurs later and is broader (e.g. curve 3). This predicted behaviour is consistent with the observation that small-island Pacific highstands occur later than at continental margins. It is also consistent with observations from formerly glaciated northern Britain that the timing of the highstand is earliest where the rebound is greatest [*Lambeck*, 1993b]. This characteristic of the highstand cannot be reproduced by modifications of the rheological parameters and the observations are indicative of late-Holocene increases in ocean volume irrespective of the choice of mantle viscosity or lithospheric thickness.

Present-day sea level is spatially variable, even when corrected for glacio-hydro-isostasy. The two long tide-gauge records from Australia indicate a secular rise in relative sea level of 0.9 -1.4 mm/year which, when corrected for the isostatic effect of opposite sign, increases to about 1.2 - 1.6 mm/year respectively. These values are lower than globally averaged estimates of 1.8 - 2.0 mm/year sometimes quoted [e.g. *Peltier and Jiang*,

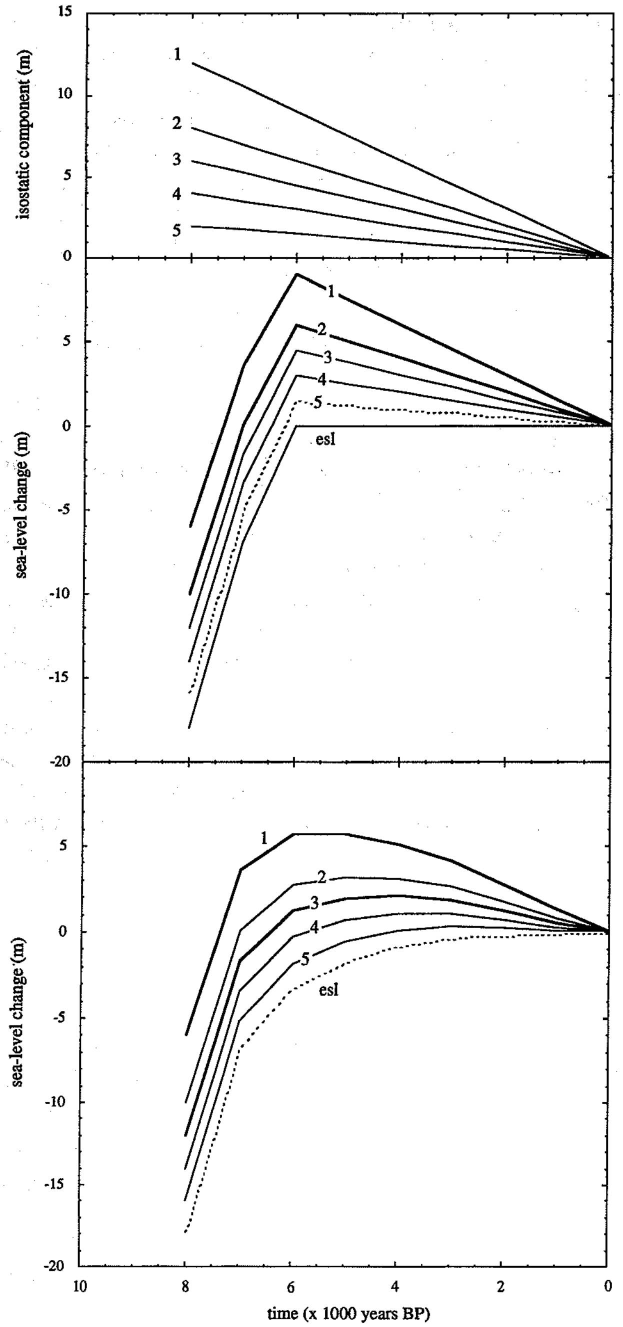

Figure 10. Schematic illustration of the shape of the Late Holocene sea-level curve in the far-field areas in the presence of ocean volume increase. The upper panel illustrates the isostatic components of different magnitudes; the middle panel illustrates the predicted sea levels for the case where all melting ceased at 6000 years ago; the lower panel is for the case of on-going melting in late-Holocene time.

1997; *Douglas*, 1997] but similar to some of the lower values found from regional analyses (e.g. *Lambeck et al.* [1998a]; *Woodworth et al.* [1999]). Part of the differences may lie in analysis methods used, in lengths of records analysed, or in the corrections applied for tectonic or glacio-hydro-isostatic effects. But it may also be indicative of lateral variation in present-day change due to long term regional changes in climatological and oceanographic factors, including ocean temperatures, salinity, currents and surface-wind stress. Table 3 summarises some results of recent analyses. Of interest is that while the three global solutions are based on very similar isostatic corrections the result based on the far-field analysis by *Davis and Mitrovica* [1996] is less than that resulting from the other two analyses [*Peltier and Jiang*, 1997; *Douglas*, 1997] that incorporate tide gauge data from sites close to or within the former ice margins.

The three global results in Table 3 are all based on the assumption that in calculating the isostatic effects the mantle response is laterally uniform. Furthermore, all three solutions set the lithospheric thickness at an *a-priori* value of 120-125 km and this and the associated viscosity profile may not be appropriate for all locations (c.f. Table 2). This limitation is overcome in the regional solutions included in Table 3 in which the rebound parameters are estimated from the local sea-level evidence and the model becomes an effective interpolator for present-day contributions within the region covered by the analysis. For the Scandinavian solution three approaches have been used to correct the observed tide-gauge for the isostatic effects [*Lambeck et al.*,1998a]. One is to use the geological data and rebound model to estimate the long-term trend and to subtract this from the tide-gauge result (solution SCAN-1, Table 3). In this sense, the rebound model is used as an interpolation device to give a best-fitting sea-level function to the palaeo sea-level indicators and any lack of separation of the model parameters that describe this function (viscosity, lithospheric thickness, the ice-volume equivalent sea-level term, ice model parameters) will not influence greatly the estimation of the present isostatic effects. Another approach is to estimate that part of the Earth's surface where the isostatic effects are zero or very small. Around an ice sheet, for example, there is a zone where the predicted isostatic effect is zero, or nearly so, for a broad range of model parameters and in the case of Scandinavia, a substantial number of quality tide-gauge records lies within this zone (solution SCAN-2). For Scandinavia a third solution is possible, namely to use the observed tilting of some of the large inland lakes [*Ekman*, 1996] to estimate the earth-response parameters and to then use these parameters to correct the marine tide gauges for isostatic effects (solution SCAN-3). All three approaches give consistent results that point to a secular sea level rise in northern Europe of about 1.1 ±0.2 mm/year for the past 100 years.

Table 3. Summary of recent estimates of secular sea level change. All results have been corrected for the glacio-hydro-isostatic contributions. The Australian result is the average of the two longest records from Table 1. The three Scandinavian results are discussed in text. The error bars on the *Woodworth et al.* [1999] solution is from *Shennan and Woodworth* [1992].

Solution/author	Region	Rate mm/year
Australia	regional	1.40 ± 0.25
SCAN-1	regional - Scandinavia	1.1 ±0.2
SCAN-2	regional Scandinavia	1.0 ±0.25
SCAN-3	regional - Scandinavia	1.01 ±0.2
Woodworth et al., 1999	regional British Isles	1.0 ±0.15
Davis and Mitrovica, 1996	regional - N. America E. coast	1.5 ±0.3
Peltier and Jiang, 1977	regional - N. America E. coast	2.0 ±0.6
Mitrovica and Davis, 1995	global (far-field only)	1.4 ±0.4
Peltier and Jiang, 1997	global	1.8 ± 0.6
Douglas, 1997	global	1.8 ±0.1

The solution by *Woodworth et al.* [1999] corrects for the isostatic effects, as well as any other long-term, millennial-scale contributions by differencing the geological and tide gauge records (see also *Shennan and Woodworth* [1992]). This solution is also at the lower end of the estimates. Of the three similar regional solutions for the North American east coast, two give higher values than the third (see Table 3) but the differences are probably statistically insignificant. But the result for all three analyses is significantly higher than that found for the other side of the Atlantic Ocean and this may relate to spatial variability in ocean warming and thermal expansion (e.g. *Levitus et al.* [2000]). The comparisons of estimates for the three regions indicate that the regional variability in secular sea level change, of oceanographic and climatological origin, may be of the order ±0.5 mm/year superimposed on a globally averaged value of about 1.5 mm/year.

Other than thermal expansion, other factors that contribute to spatial variability in long-term sea-level change include shifts in position or in strength of ocean boundary currents. Can variability in the Cape Leeuwin Current or in the East Australian Current, for example, explain the differences in results for the east and west coasts of Australia? Another factor is that when water is added into the oceans from mountain glaciers or from an exchange with ground and surface waters, the redistribution of the added water is not uniformly distributed because of changes in gravitational attraction between land and sea and because of the sea-floor deformation under changing loads. Such variability has been demonstrated for the mountain glacier contribution [*Nakiboglu and Lambeck*, 1991] but will also result from any addition of surface or ground water into the oceans.

What the tide-gauge results in Table 3 suggest, and what our understanding of surface loading problems dictates, is that there may be substantial spatial variability in secular sea-level change, even when the tide-gauge data has been corrected for the glacio-hydro-isostatic effects. Thus the averaging of geographically poorly distributed tide gauges may not lead to a very meaningful estimate of the global average. (This was one of the reasons for the analysis by *Nakiboglu and Lambeck* [1991] in which the tide gauge records were expressed as a low-degree spherical harmonic expansion the zero degree term of which provides the estimate of the globally averaged sea-level rise. Such an analysis has the advantages that it is not necessary to correct for the isostatic effects and that it may provide information on some of the long-wavelength spatial variability of the secular change.)

Irrespective of the detail of this spatial variability, the results do suggest that a globally averaged sea-level rise of about 1.5±0.5 mm/year has occurred for the duration of the tide gauge records, or about 100 years [*Church et al.*, 2001]. Tide gauge records for earlier periods are limited but some of the very long records do indicate an acceleration in mean sea level during the nineteenth century [*Woodworth*, 1999; *Ekman*, 1999; *Maul and Martin*, 1993]. If the secular trend noted for the past century was not a recent development then it would become visible in the geological or archaeological records of sea level change as early as 500 - 1000 years ago with shorelines at 75 - 150 cm above present. High resolution records indicate that little change in sea level has occurred, over and above that which can be attributed to the isostatic factors, during the past 2000 years (Figure 7 and also *Lambeck and Bard* [2000]; *Morhange et al.* [2001]; *Sivan et al.* [2001]) such that the present-day rise must indeed

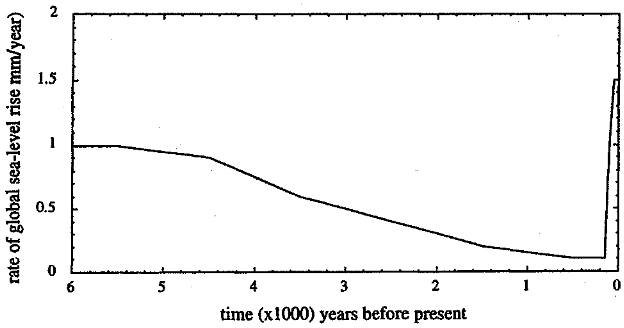

Figure 11. Inferred rates of sea-level change for past 6000 years.

have been a relatively recent phenomenon. The rates of sea-level rise for the past 7000 years are illustrated in Figure 11, based on the average of the results illustrated in Figure 7d and the additional results from the above references. The result for the past century is the global best estimate of *Church et al.* [2001] and this rise is assumed to have been initiated about 100-150 years ago, consistent with the evidence from the longest tide-gauge records.

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