POSTGLACIAL SEA LEVELS IN THE PACIFIC: IMPLICATIONS WITH RESPECT TO DEGLACIATION REGIME AND LOCAL TECTONICS

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ABSTRACT


Observations of Holocene sea level fluctuations contain information on the melting history of the major past ice sheets, on the viscosity of the mantle, and on vertical motions of tectonic origin. Sea level changes at eleven sites in the Southwest Pacific region have been examined with the following results. (1) The ICE-1 model of Peltier and Andrews (1976) for Arctic deglaciation is inadequate to explain all sea levels. Two alternative modifications of this model are: (a) a forward shift in the deglaciation history with the onset of major melting starting at about 14,000 yrs ago and ending at 6000 yrs ago, (b) the ICE-1 model for the Arctic deglaciation from 17,000 to 7000 yrs B.P. plus a significant amount of Antarctic melt-water. (2) The effective mantle viscosity is in the range $10^{21} - 10^{22}$ Pa s although the sea level results for the locations studied are not strongly dependent on viscosity. (3) Vertical motions of tectonic origin can be separated from the hydro-isostatic sea level fluctuations if these motions exceed about 0.5 mm yr$^{-1}$. Only two locations provide clear evidence for vertical motion consistent with independent geological observations: the Huon Peninsula of New Guinea is undergoing uplift of between 1.2 and 1.7 mm yr$^{-1}$ while the Hawaiian island of Oahu is subsiding at a rate of about 1-2 mm yr$^{-1}$. Stations along the southeastern and eastern coast of Australia are suggestive of small uplift of perhaps 0.5-1.0 mm yr$^{-1}$. The sea levels in New Zealand and the Malacca Straits are also found to be in agreement with other observations that point to regional tectonic stability.

INTRODUCTION

The study of past sea levels is important for understanding the glacial–interglacial cycles and global fluctuations in climate. A detailed knowledge of the temporal and spatial history of the ice sheets and the concomitant changes in global sea levels also provides information on the rheological properties of the mantle at intermediate time scales of $\sim 10^4$ yrs since the solid earth deforms under the changing ice-water load. Quantitative studies of the past sea levels (see for example Peltier and Andrews, 1976; Peltier et al., 1978; Peltier, 1981) are restricted to the late Pleistocene and Holocene because (1) there are more numerous and more accurate observations pertaining to this epoch than for the earlier periods, (2) the mechanical memory of
the earth is relatively short for long wavelength loads and the recent observations carry almost no signatures from the glacial cycles preceding the Late Wisconsonian. Another important, but largely neglected, product of studies of past sealevels is the possibility of detecting local tectonic motions relative to the changing sealevel. Such information extracted from the sealevel data covering periods of $\sim 10^4$ yrs supplement present-day observations of crustal motion and may contribute to understanding local tectonic problems.

The theoretical problem of sealevel changes on a deformable earth has been investigated by Farrell and Clark (1976) who developed an integral equation, the solution of which yields changes in sealevel due to accretion or disintegration of ice at arbitrary points on the earth. Their results demonstrated that the self-gravitational effects of ice and water contribute significantly to sealevel and that sealevel changes are far from uniform over the oceans. Since the total mass and the gravitational energy of ice plus water are conserved at all times, the regional fluctuations of sealevel contain information on mass, location and the melting history of continental and grounded marginal ice sheets which have subsequently disintegrated with time. Marine ice sheets grounded in shallow waters may have been a significant additional source of melt-water since the Late Wisconsinian ice sheets included substantial volumes of grounded and floating ice in the Kara, Barents, Ross and Weddell seas (Hughes et al., 1981; Stuiver et al., 1981).

With such unknowns as the past ice volumes and their melting histories, the mantle rheology and hydro-isostatic deformation, as well as local tectonic movements, the separation of the relevant parameters incorporated within the sealevel observations appears to be an ill-posed problem. However, some separation is possible by (1) identifying the dominant parameter at any given time and location, and by (2) incorporating other evidence such as isochrone maps, isostatic uplift, gravity anomalies and ice mechanics principles into the solutions.

The most comprehensive global studies of sealevel changes are by Peltier and Andrews (1976), Peltier et al. (1978) and Peltier (1981) who have incorporated all observables in a rigorous theory based on a linear viscoelastic rheology for the earth. The primary aim of these studies has been to determine the viscosity of the earth and to refine the Arctic deglaciation regime obtained initially from isochrone maps and ice mechanics considerations. Studies on regional scales are necessary for complementing this global approach by setting further constraints on one or more of the parameters involved depending on the dominant parameter for a particular region. Regional studies of the sealevels in North America have been pursued by Clark (1980) and Quinlan and Beaumont (1981). One of the objectives of these studies was to constrain the deglaciation regime of the Laurentide ice sheets. However, the region studied in these works is the near-field where the hydro-isostatic deformation is as dominant a contribution to sealevel changes as is the melt-water itself. Studies of this kind provide information on the fine detail of the deglaciation regime once the viscosity of the earth and the main characteristics of deglaciation have been
accurately established. Alternatively, if the ice history is known, analysis of near-field
sealevels is most appropriate for determining the viscosity of the earth. The inter-
mediate region, within several thousand kilometres of the melting centres, is less
sensitive to the fine details of the ice load. However, the deformational components
may still be complicated by arch-moat structures ringing the deglaciated area and by
a possible migration of this structure in time. Such structure and temporal behaviour
is a function of the viscosity profile in the mantle. For a mantle of nearly uniform
viscosity this structure will be less apparent and will gradually move away from the
deglaciation centre while diminishing in magnitude. However, if the mantle viscosity
increases rapidly with depth the arch-moat structure will be more pronounced,
subsiding in time while remaining more or less stationary in its location. Therefore
the analysis of sealevels in the intermediate region is important in establishing the
viscosity profile within the mantle. The study of this intermediate region is also
important in deciding whether the rheology of the mantle is linear or non-linear
since a linear rheological model will tend to predict a viscosity increasing with depth
if the actual rheology is non-linear (Cathles, 1980).

The third characteristic region with respect to sealevel signatures is the region
beyond the near- and intermediate-fields. Sealevels in this far-field region vary
smoothly both in location and in time, and they are influenced mainly by the added
melt-water and its self-gravitational effects. If deglaciation was confined to the
Arctic, then the Southern Hemisphere will be the far-field. On the other hand if both
Arctic and Antarctic deglaciations have synchronously contributed comparable
amounts of melt-water to the world oceans the far-field characteristics will be
confined to equatorial regions.

The trend in the Northern Hemisphere is well established by Clark (1980) but a
similar regional investigation of the sealevels in southerly latitudes has not yet been
made. Such a study may contribute to global investigations by constraining the
Antarctic contributions to the Late Wisconsinian sealevels.

Peltier and Andrews (1976) have developed a detailed model of the deglaciation
regime for the Laurentide, Cordilleran and Fennoscandia ice sheets—their ICE-1
model—on the basis of isochrone maps and ice mechanics considerations. In the
follow-up work (Peltier et al., 1978; Peltier, 1981) the ICE-1 model was tested
against the observed global sealevel changes and it was shown that ICE-2 can
explain the near and intermediate sealevels if the mantle has a uniform viscosity of
$\sim 10^{22}$ P. However, Peltier noted that the far-field sealevels require that the melting
history is shifted forward in time by about 2000 yrs. This shifted time history
(ICE-2) violates isochrones in North America and requires reduced ice volumes if
the mantle viscosity is around $10^{22}$ P. This led Peltier et al. (1978) to establish a new
melting history (ICE-3) by shifting forward the melting history of ICE-1 and by
reducing the ice volumes in accordance with the constraints imposed by the
near-field sealevels. But Peltier (1981) has shown that ICE-3 does not have enough
water to explain the higher observed sealevels in both intermediate- and far-field
locations.
In the present paper we restrict the study to the discussion of Pacific sealevels and we take the ICE-1 model of Peltier and Andrews (1976) as our starting ice model for the Northern Hemisphere ice sheets. We develop a semi-analytical solution for the sealevel equation on a deformable earth and utilize this to modify the ICE-1 model in concordance with the observed Pacific sealevels. Since the deformational component to sealevel fluctuation is relatively less important in this region than in the near-field, we adopt only an approximate model of the earth and represent it as a quasi-homogeneous body whose response at each wavelength is determined by the elastic Love numbers, effective rigidity and effective viscosity (Nakiboglu and Lambeck, 1981). This simple rheological model enables us to readily test various deglaciation regimes and mantle rheologies. We also attempt to separate crustal motion from sealevel observations in tectonically active areas like Oahu and Papua-New Guinea.

THEORY OF SEALEVEL CHANGES DUE TO DEGLACIATION

Farrell and Clark (1976) have obtained the governing integral equation for the change in sealevel with respect to shoreline resulting from a mass exchange between the oceans and glaciers. Farrell and Clark have taken into consideration the condition that the static sea surface remains an equipotential at all times and that the total mass of the ice plus water is conserved. The integral equation is:

\[ \tilde{\zeta}(p) = \delta_1 \left[ F(p, q) * \tilde{\zeta}(q) - \langle F(p, q) * \tilde{\zeta}(q) \rangle \right] + \delta_n \left[ F(p, q) * \tilde{\zeta}(q) - \langle F(p, q) * \tilde{\zeta}(q) \rangle \right] - \frac{M_i}{A_0 \delta_n} \]  

where \( p \) and \( q \) are the fixed and moving points respectively, \( \tilde{\zeta} \) and \( \zeta \), are the changes in sealevel and ice height respectively, \( M_i \) is the change in the total mass of the ice, positive for accretion and negative for melting, \( \delta_1 \) and \( \delta_n \) are the densities of ice and water, \( A_0 \) is the area of the oceans, \( \langle \cdots \rangle \) denotes the convolution integral over either the ice or the water areas depending on the integrant, \( \langle \cdots \rangle \) denotes the mean value over the oceans, \( F(p, q) \) is the kernel (Green's) function defined as:

\[ F(p, q) = \begin{cases} \frac{3}{4\pi\delta_e} \frac{1}{2 \sin \psi_{pq}/2} & \text{for the rigid earth} \\ \frac{3}{4\pi\delta_e} \left[ \frac{1}{2 \sin \psi_{pq}/2} + \sum_{n=0}^{\infty} (k_n - h_n) P_n(\cos \psi_{pq}) \right] & \text{for the elastic earth} \end{cases} \]

In the expressions for the kernel, \( \delta_e \) is the mean density of the earth, \( k_n \) and \( h_n \) are the elastic load Love numbers, \( \psi_{pq} \) is the geocentric angle between \( p \) and \( q \), and \( P_n \) is the Legendre polynomial. The integral equation (1a) is solved using an iterative technique based on the Liouville–Neumann series. In decreasing order of magnitude the causes of change in sealevel in a far-field location are (1) the addition of
melt-water into the oceans, (2) self attraction of the added water together with the diminishing attraction of melting ice, and (3) yielding of the crust under the added waterload and reduced ice load.

Let the change in height of an ice mass at location $(\theta, \lambda)$ at time $t$ be:

$$\xi(\theta, \lambda; t) = h(\theta, \lambda)T_k(t)$$

where $T_k(t)$ and $h$ are the representative load history and melting rate at colatitude $\theta$ and longitude $\lambda$.

For uniform melting we have:

$$T_k(t) = t_0 \left[ -\frac{t}{t_0} H(t_0 - t) + H(t - t_0) \right]$$

where the time origin $t = 0$ is taken to be the start of deglaciation, $t_0$ is the epoch of the end of deglaciation, and $H$ is the Heaviside unit step function.

Neglecting the gravitational and deformational effects we obtain the zero order approximation for the sealevel change from eq. 1 as:

$$\xi_e = -\frac{M_i}{A_0 \delta_e}$$

or:

$$\xi_e = -\frac{\delta_i}{4\pi a_{00} \delta_w} \int_{\Omega_i} hI(t) \, d\Omega$$

where $\Omega_i$ is the area of the ice, $a_{00}$ is the zero degree harmonic coefficient of the ocean function, and $d\Omega = \sin \theta \, d\theta \, d\lambda$ is the area element on a unit sphere. Equation 4 is nothing but the eustatic change in sealevel. A better approximation for $\xi$ can now be obtained by substituting $\xi_e$ into the right-hand side of eq. 1 and, again neglecting the deformational effects, the change in sealevel on the rigid earth becomes:

$$\xi_R = \xi_e(1 + C_1 g_1) + C_2 g_2$$

where:

$$C_1 = \frac{3 \delta_w}{4\pi \delta_e}, \quad C_2 = \frac{3 \delta_i}{4\pi \delta_e}$$

$$g_1(p) = \int_{\Omega_0} \frac{d\Omega_p}{2 \sin \psi_p/2} - \left\langle \int_{\Omega_0} \frac{d\Omega_p}{2 \sin \psi_p/2} \right\rangle$$

$$g_2(p) = \int_{\Omega_i} \frac{\xi d\Omega_q}{2 \sin \psi_q/2} - \left\langle \int_{\Omega_i} \frac{\xi d\Omega_q}{2 \sin \psi_q/2} \right\rangle$$

and:

$$\Omega_0 \equiv 4\pi a_{00}$$ is the ocean area.
Any further iteration of eq. 1 for the rigid earth solution is unnecessary because additional terms resulting from such an iteration will be of \( O(C_i^2) \approx 2 \cdot 10^{-3} \). That is, the contribution of these additional terms on sealevel is about 20 cm in 100 m and negligible in view of the observational accuracies of sealevel and ice history. The next step in this iterative kernel solution is to waive the assumption of a rigid earth and substitute eq. 5 into the right-hand side of eq. 1. This gives the sealevel changes on an elastic earth using the load Love numbers. Assuming that the earth is a linear (Maxwellian) viscoelastic body, the elastic equations can be viewed, by virtue of the correspondence principle (Peltier, 1974), as being those for the viscoelastic earth in the Laplace transform domain. The Love numbers can then be replaced by the Love operators (Nakiboglu and Lambeck, 1981) and the functions of time are replaced by their Laplace transforms.

Substituting the Laplace transform of eq. 5 into the right-hand side of the integral equation, neglecting terms of \( O(C_i^2) \) and taking the inverse Laplace transform, yields the approximate solution as:

\[
\zeta(\theta, \lambda, t) = \zeta_R(\theta, \lambda, t) + C_2 \sum_{n} \left[ \int_{\Omega} h_\theta_n P_n(\cos \psi) \, d\Omega \right. \\
- \left( \int_{\Omega} h_{\theta n} P_n(\cos \psi) \, d\Omega \right) + C_l \int_{\Omega} \theta_n P_n(\cos \psi) \, d\Omega \\
- \left( \int_{\Omega} \theta_n P_n(\cos \psi) \, d\Omega \right) \tag{6}
\]

where:

\[
\theta_n = L^{-1} \left[ (\hat{k}_n - \hat{h}_n) \hat{T} \right]
\]

where \( L^{-1} \) is the inverse Laplace transform operator, \( - \) indicates the Laplace transform of a quantity, \( \hat{k}_n \) and \( \hat{h}_n \) are the load Love operators.

Equation 6 should yield the same results as the ones obtained by Peltier et al. (1978) provided one determines the Love operators \( \hat{k}_n \) and \( \hat{h}_n \) for the real earth with layered mantle structure and a liquid outer core. However, as we confine ourselves to the study of the far-fields, an approximate description of the hydro-isostatic adjustment suffices. For a quasi-homogeneous, incompressible Maxwell earth, the Love operators are particularly simple, being:

\[
\begin{align*}
\hat{k}_n &= \frac{s + \tau^{-1}}{s - k_n/\tau} k_n, \quad n \geq 2, \\
\hat{h}_n &= h_n \frac{s + \tau^{-1}}{s - h_n/\tau}
\end{align*}
\tag{7}
\]

Here \( \tau \) is the relaxation time constant defined as the ratio of the effective viscosity to the effective rigidity of the earth and \( s \) is the Laplace transform variable with the unit of frequency. While the operators given in eq. 7 only approximate when applied to the earth, they represent rather well the response of the viscoelastic layered earth at long-wavelengths (Nakiboglu and Lambeck, 1981). Therefore eq. 6 is applicable at
far-field locations where higher degree components of ice load have relatively small effects on sealevels. Around the deglaciated areas a more rigorous formulation of the Love operator (eq. 7), such as the one given by Peltier and Andrews (1976) and Peltier et al. (1978), may be necessary.

That part of the sealevel change induced by deformation (the last two terms in eq. 6) can be computed on the assumption of one single melting history prevailing for a given ice mass, an assumption that is justified by noting that deglaciation started almost simultaneously everywhere in the northern ice sheets and that it was complete by about 7000–5000 yrs B.P. Errors in the deformational component of sealevel changes will therefore be small, of the order of several metres, even though some parts of a given ice sheet might have had a shorter time-span of melting. The average time histories of the Laurentide and Fennoscandian ice sheets implied by the ICE-1 model are given in Fig. 1. In the subsequent solutions the ice loads are defined in a $10^0 \times 10^0$ grid and the ocean geometry is defined by a spherical harmonic expansion up to degree 18. Hence the resolution of our solutions is inadequate near the ice loads but provides an adequate accuracy for the far-field regions.

Fig. 1. Average rate of deglaciation of the Laurentide (1), Fennoscandian (2), and combined (3) ice sheets according to the ICE-1 model of Peltier and Andrews (1976). Curve 4 is the linear approximation of ICE-1 used in deformation calculations.
SEALEVEL CHANGES IN THE PACIFIC: INFERENCES ON DEGLACIATION REGIME

Sealevel data at eleven locations in the Pacific region are discussed in the appendix. The selected locations, ranging in order of decreasing colatitude, from New Zealand ($\theta = 133.5^\circ$, $\lambda = 172.7^\circ$) to eastern China ($\theta = 57.3^\circ$, $\lambda = 126.6^\circ$) sample relatively well the equatorial, temperate and southerly regions in the Pacific. The relative sealevel curves in these localities should therefore give a reasonable indication of the principal characteristics of deglaciation. However, some of these localities are tectonically active, like Oahu and New Guinea, and some caution is required in interpreting the sealevel curves in these areas.

The observations of past sealevels are subject to several error sources. In particular, attention needs to be given to the relation between the sealevel indicator and the datum level, to the dating of the geomorphic features defining past sealevels and to post-depositional sedimentary processes. Several indicators have been used to reconstruct the post-glacial sealevel curves. Some, such as coral reefs, indicate that sealevel was above the sampled level at a given time, while others, such as peats indicate that sealevel was below the sampled level. Others such as mangroves, beach or chenier ridges are direct indicators of the tidal zone which may vary by several metres along any given coast. Errors of this nature are unlikely to exceed a few metres for sealevel changes over the past 14,000 years.

All dates are based on radiocarbon techniques. Counting errors will not exceed 1% of the age estimate of the sample but other factors, such as the aging effect of the $^{14}$C distribution in ocean water or isotope fractionation effect, could, if not corrected for, introduce errors ranging from 5 to 10% of the sample age. Significant deviations from the true age of the sample may also result from the diagenetic recrystallization of marine carbonates (Chappell and Polach, 1972) or from contamination of organic material with more recent material (Polach et al., 1981).

Perturbations from post-depositional sedimentary processes are difficult to evaluate but are potentially significant. Transported materials could be interpreted as being "in-situ", erosion of chenier and beach ridges, compaction of peats, are obvious examples of potential misinterpretations.

In a first solution the ICE-1 model of Peltier and Andrews is used to predict the theoretical sealevel changes at the eleven locations using three values for the effective viscosity; (a) $10^{22}$ P, as suggested by Peltier, (b) $5 \times 10^{21}$ P and (c) $10^{21}$ P (Fig. 2). For most of the stations the choice of viscosity is unimportant and only for stations where Late Holocene highs are predicted by this model do the sealevel curves show any marked dependence on viscosity (e.g., the Gulf of Carpentaria site, East China, or Oahu). The maximum predicted difference between the sealevel curves for $10^{21}$ and $10^{22}$ P are found for the East China sites where this effect approaches 7 m.

Agreement between the observed and predicted sealevels is unsatisfactory for all sites. In particular, for localities with reasonable data back to about 17,000 yrs B.P. the observed sealevel curve is shifted forwards in time relative to the predicted
Fig. 2. Observed and predicted sealevels for eleven sites in the Pacific: New Zealand (1), Melbourne (2), Moruya (3), SE Queensland (4), Gulf of Carpentaria (5), North Barrier Reef (6), Huon Peninsula (7), Malacca Straits (8), Micronesia (9), East China (10) and Oahu (11). The observed values are discussed in the appendix. Curves 1 are the predicted sealevels based on the ICE-1 (17,000–6000 yrs B.P.) model and curves 2 are based on the time-shifted ICE-1 (14,000–5000 yrs B.P.) model. The effective viscosity is (a) $10^{22}$ P, (b) $5 \cdot 10^{21}$ P and (c) $10^{21}$ P. (kybp: $10^3$ yrs B.P.)
curves and the former indicate steeper slopes. Also the models predict Holocene highs that are generally greater than observed. Either the ICE-1 model is inadequate or all sites, with the exception of New Guinea, are subject to substantial subsidence. The latter is unlikely and the comparisons point to a revision being required of the ICE-1 model as already noted by Peltier (1981). The Early Holocene slopes of the predicted curves can be steepened by (a) significantly increasing the thickness of ICE-1, (b) assuming a delayed start for deglaciation, at 14,000 yrs B.P. instead of 17,000 yrs B.P., or by (c) introducing new melt sources. The first option leads to near-field sealevel variations that are in contradiction with observations as well as with the gravity data (Peltier, 1981). The second option is closely related to the first
Fig. 2 (continued).
since an increase in ice height requires that the time of initiating of melting must be pushed backwards in order to give similar near-field sea level curves. Peltier revised ICE-1 by reducing the ice load and by bringing the melting interval forward but this did not improve the fit to the near-field data.

It appears that the sea level curves will match the observed slopes only if the start of deglaciation is brought forward substantially, by at least 3000 years (see the New Zealand result in Fig. 2). Even then we require additional melting sources to explain the steep pre-Holocene slopes observed at Moruya, the Malacca Straits and New Zealand. Any further forward shift in time history may require substantial reduction in ice load and an accompanying increase in the earth's viscosity in order to match the near-field data.

Fig. 2 also illustrates the predicted sea level curve for a deglaciation interval from 14,000–5000 yrs B.P. Such an interval can be justified on several grounds. Firstly, $^{14}$C dates of morains, associated with equatorial mountain glaciers, indicate that the ice retreat began at about 14,000–5000 yrs B.P. (Flenley and Morley, 1978). These dates may provide the earliest indications of post-glacial climate changes because equatorial mountain glaciers wax and wane rapidly in response to global temperature changes (Hope and Peterson, 1975). Secondly, according to Andrews and Barry (1978), the advance of the Cordilleran ice sheet also culminated close to 14,000 yrs B.P. More importantly, these authors conclude that evidence for an 18,000 yrs B.P. glacial maximum is lacking in the Northern Hemisphere and that glacial advances occurred as late as 8000 yrs B.P. In addition, estimates for the end of deglaciation also vary considerably with a range of 7000–5000 yrs B.P. (cf., the isochrone maps of Bryson et al., 1969; Prest, 1969; Fillon, 1972).

With this time shifted ice load—referred to as ICE-1 (14,000–5000 yrs B.P.)—the agreement between observed and predicted sea levels is now much improved, particularly for sites such as New Zealand and the Malacca Straits where sea level observations go back the furthest into time. The high Holocene levels previously predicted for the Gulf of Carpentaria, North Barrier Reef and elsewhere are now also much reduced and in greater harmony with the observations. The disagreement between observations and theory is accentuated for the New Guinea sites but here tectonic contributions are most important. The Malacca results, from an area of known tectonic stability (Batchelor, 1979) are suggestive of a further delay in the start of major deglaciation.

An alternative to these ice models is to introduce additional sources of ice. For example, the ICE-1 results for locations of apparent tectonic stability such as Moruya, New Zealand or the Malacca Straits, imply an additional source of melt-water, with a volume equivalent to about 20–25 m of eustatic change, or an additional $(7.0–8.8) \times 10^6$ km$^3$ of melt-water. This sets a lower limit of further deglaciated ice of $(7.8–9.8) \times 10^6$ km$^3$, excluding floating ice. The only potential sources of melt-water not included in ICE-1 are the eastern Siberian and the Antarctic ice sheets. An eastern Siberian ice sheet based on shallow waters and
joining the Laurentide ice sheet to the Kara Sea can provide the required melt-water if it were about 2 km thick over an area of roughly 2000 km². There is some evidence for a substantial East-Siberian ice sheet, comprising the Kara, Putorana, Uralian and Tallya ice masses, with a total volume of $3.3 \cdot 10^6$ km³ (Barashnikova et al., 1980). The melting of such an ice sheet, not included in the ICE-1 model, would have contributed about 10 m to the eustatic sealevels. The second, and possibly more substantial, source of melt-water is the ice sheets at the shallow continental margins of the Ross and Weddell seas in Antarctica for which Hughes et al. (1981) estimate that the deglaciated ice contributes about $9.8 \cdot 10^6$ km³. This is exactly the additional amount required to explain the Pacific sealevels although this good agreement is probably fortuitous in view of the paucity of Antarctic deglaciation data.

Because detailed information on the Antarctic ice history is still lacking we have developed a simple deglaciation regime by taking the difference between the maximum ice cover estimate of Stuiver et al. (1981) and the present ice volume. This difference has been distributed into eleven $10^0 \times 10^0$ areas around the Weddell and Ross seas shelves which are believed to represent the main areas of deglaciation. We adopt a single linear melting history for this load as in eq. 3. The net contribution of Antarctic ice to the eustatic sealevel from this model is 24.5 m in agreement with the estimate of Lingle and Clark (1979).

Figure 3 illustrates the results for the combined ICE-1 and Antarctic loads with glaciation occurring in the interval 17,000–6000 yrs B.P. In these models the southern latitude stations are now at intermediate-field locations and one sees a greater dependence of the sealevel curve on mantle viscosity. For New Zealand, for example, the maximum difference at any epochs for viscosities of $10^{21}$ and $10^{22}$ P is now 10 m whereas for the ICE-1 (17,000–6000 yrs B.P.) model this difference never reaches 1 m. For the combined ICE-1 Antarctic model the higher viscosity ($10^{22}$ P) mantles produce high Holocene levels at many of the stations (e.g. 7 m at Moruya, 9 m in the Gulf of Carpentaria and 16 m on Oahu) which are not observed. Thus, if the Antarctic deglaciation is accepted, the mantle viscosity is of the order $10^{21}$ P.

Of the locations for which reasonably reliable sealevels are available for the pre-end-of-melting phase, the time shifted ICE-1 (14,000–5000 yrs B.P.) and the Arctic plus Antarctic (17,000–6000 yrs B.P.) model fit—on average—the data equally well. This is certainly so for New Zealand for which significant Antarctic deglaciation is ruled out if the Arctic deglaciation is moved forward. For the Moruya site all predicted sealevels fail to satisfy the observed levels although the ICE-1 (14,000–5000 yrs B.P.) model gives closest agreement. Additional Antarctic melt-water will improve the agreement here by bringing the sealevel curve forward in time. Alternatively the end of glaciation can be delayed to give the same effect. For southeast Queensland the comparison would suggest either less melt-water from Antarctica and the ICE-1 (17,000–6000 yrs B.P.) model or the ICE-1 model shifted forward by a smaller amount than before. For the Malacca Straits ICE-1 (14,000–5000 yrs B.P.) and Arctic and Antarctic (17,000–6000 yrs B.P) fit the data
Fig. 3. Same as Fig. 2 but with curves 2 corresponding to the Arctic plus Antarctic (17,000–6000 yrs B.P.) model.
about equally well, although better agreement can be found by moving the Antarctic melting forward and increasing the mantle viscosity.

For the remaining stations for which the sealevel records do not go back beyond about 7000 yrs B.P. the time shifted ICE-1 model gives good results for Melbourne, North Barrier Reef, Caroline Islands and the East China coast. In particular this model represents well the observed Holocene highs if the viscosity is of the order \((5-10) \cdot 10^{21}\) P. The Gulf of Carpentaria results would also point to such a viscosity range for this ice-melt model. If an Antarctic contribution is added the agreement is generally satisfactory provided that this contribution is also brought forward. In all of these cases viscosities in the range \((5-10) \cdot 10^{21}\) P are preferred.

From these comparisons it is clear that there is considerable freedom in selecting
parameters defining the deglaciation regime. It is unlikely therefore that one set of parameters will satisfy all observations, even if they were error free, because of the contributions from local tectonics (e.g. uplift at the New Guinea sites, subsidence of Oahu). However, two classes of models generally satisfy the observations: (1) deglaciation in the Northern Hemisphere only with the major onset of deglaciation starting at 14,000 yrs B.P. and ending at 5000 yrs B.P., or (2) Northern Hemisphere deglaciation from 17,000–6000 yrs B.P. with a significant Antarctic contribution at the same time. In both instances the observations point to an average mantle viscosity in the range \((5-10) \cdot 10^{21}\) P.

VERTICAL CRUSTAL MOTION

The sealevel records also contain local tectonic signatures superimposed on the effects of added melt-water and concomitant hydro-isostatic deformation of the earth. Thus the characteristic trend of sealevel observations depends not only on the nature of the hydro-isostatic deformation but also on whether the location is uplifting or subsiding with respect to the centre of the earth. For example, in the far-field regions covering much of the equatorial Pacific, the hydro-isostatic adjustments, when viewed with respect to sealevel, appear as a gradual subsidence of the ocean floor. Island stations in this region should not, therefore, exhibit Holocene sealevels that are higher than the present level unless the island is also undergoing tectonic uplift. In mid-southern latitudes the hydro-isostatic uplift due to arctic ice-melt alone consists of a small uplift after all melt-water has been added. Thus Late Holocene levels will—in the absence of local subsidence—be higher than present sealevels.

Assuming that the deglaciation history and the hydro-isostatic response is known, to within several metres, we can use the discrepancies between the observed and predicted sealevels as a measure of local tectonic motions with an accuracy of the order \(\pm 0.5\) mm yr\(^{-1}\). Table I summarizes the results of a regression analysis of these discrepancies for each of the eleven stations. Three quantities have been computed; (1) the correlation coefficient of the discrepancies between the observed and predicted sealevels, (2) the estimated crustal motion \(\xi\) (negative \(\xi\) indicates uplift) and (3) the time \(t_0\) at which the tectonic component of uplift is zero. If there is ongoing crustal motion that is a linear function of time over the past \(10^4\) yrs, then the regression analysis should ideally yield \(t_0 = 0\). The regression analysis has been carried out for two models; (a) the Arctic and Antarctic melting from 17,000 to 6000 yrs B.P. and (b) Arctic melting only from 14,000 to 5000 yrs B.P. Both models are for \(\eta = 10^{21}\) P although, as previously discussed, these results are not very dependent on mantle viscosity. If uplift or subsidence is equally likely (this need not be so for in a tectonically inactive planet the predominant vertical motion will be subsidence due to stress relaxation and thermal contraction, with erosion resulting in either subsidence or uplift) the mean of the correlation coefficients, uplift rates and \(t_0\) should all
TABLE I

Regression analysis of the discrepancies between the observed and predicted sealevel changes for a mantle viscosity of 10^{21} P and two ice models: (a) ICE-1 (14,000–5000 yrs B.P.) and (b) Arctic and Antarctic (17,000–6000 yrs B.P.)

<table>
<thead>
<tr>
<th>Sites No. *</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5 **</th>
<th>6</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Correlation</td>
<td>-0.27</td>
<td>-0.23</td>
<td>0.08</td>
<td>-0.19</td>
<td>0.35</td>
<td>0.23</td>
</tr>
<tr>
<td>b. Coefficient</td>
<td>+0.23</td>
<td>-0.48</td>
<td>-0.27</td>
<td>-0.92</td>
<td>0.23</td>
<td>-0.54</td>
</tr>
<tr>
<td>a. Crustal motion rate $\dot{\xi}$</td>
<td>-0.4</td>
<td>-0.5</td>
<td>0.3</td>
<td>-0.2</td>
<td>0.6</td>
<td>0.1</td>
</tr>
<tr>
<td>b.</td>
<td>+0.4</td>
<td>-1.6</td>
<td>-1.1</td>
<td>-0.7</td>
<td>0.4</td>
<td>-0.3</td>
</tr>
<tr>
<td>a. Time at which motion is zero $t_0$</td>
<td>-5.8</td>
<td>-12.8</td>
<td>18.5</td>
<td>-1.7</td>
<td>-2.8</td>
<td>-7.6</td>
</tr>
<tr>
<td>b.</td>
<td>-0.7</td>
<td>-6.1</td>
<td>-8.7</td>
<td>-1.4</td>
<td>-3.5</td>
<td>0.1</td>
</tr>
</tbody>
</table>

* For the description of the Pacific sites see Appendix.
** Excluding Arafura Sea data.

be zero. The observed means and their standard deviations are given in Table I based on all sites except for the East China Sea locality. Model “a” gives the smallest absolute values for these three quantities suggesting that the Arctic–Antarctic ice model fits the data better but the results differ only marginally from those based on the time shifted ICE-1 model.

Three locations, New Guinea, Hawaii and the East China coast, give a significant degree of correlation for both models (a) and (b) and which yield rates of vertical motion that are greater than about 0.5 mm yr^{-1}. Of these, the third station yields a $t_0$ that is significantly different from zero due to all models predicting Early Holocene sealevels that are significantly less than the observed ones (see Figs. 2 and 3). If the early sealevel observations are correct this would require that the deduced high uplift rates ceased by about 8000 yrs B.P. The East China Sea levels are based on dredged samples from presently submerged geomorphic units and as such these results are less reliable than those based on either cored samples or on samples from emerged units.

The Huon Peninsula, New Guinea, is subject to significant tectonic emergence. Chappell (1974) estimated a rate of uplift of 1.8 mm yr^{-1} by examining uplifted reefs over a time span in excess of 10^5 yrs. His estimate is in good agreement with ours, based on Holocene levels only, indicating that a linear tectonic motion may be applicable for the Peninsula on time scales ranging from 10^4 to 10^5 yrs.

The Oahu, Hawaii, results point to a subsidence of the volcanic island of between 1 and 2 mm yr^{-1} for at least the last 8000 yrs. These results do not support the
suggestion by Easton and Olson (1976) of Holocene uplift. Neither do they support
the suggestion by Ku et al. (1974) that the island has been tectonically stable since
the last interglacial at 120,000 yrs B.P. These results do confirm Stearn’s (1974)
interpretation of drowned reefs at depths of 137 m and greater being the conse-
quence of island subsidence.

Data for the remaining stations generally point to insignificant vertical motion.
The New Zealand result is in keeping with the observations of Wellman (1979) that
the Christchurch area of the South Island has been relatively stable in recent times.
The stations of southeast and eastern Australia (Melbourne, Moruya and southeast
Queensland) are generally suggestive of uplifting when taken together. This is in
accordance with geological evidence although the individual rates of uplift deduced
from the sealevel data are not significant and generally greater than the geological
estimates. The Gulf of Carpentaria sites suggest a general subsidence but again the
rates are insignificant.

The results for the southern Malacca Straits are in agreement with the geological
and geophysical evidence for a tectonically stable region at least since the Miocene
(Stauffer, 1973; Batchelor, 1979). The Micronesia results are for several islands, each
of which may have its own component of vertical motion due to, for example, stress
relaxation of the volcanic loads, to thermal contraction of the lithosphere or to
variations in stress distributions along the nearby subduction zone. Thus while the
suggestion of subsidence seen in the sealevel results may be anticipated, the actual
subsidence values are not significant.
CONCLUSION

Observations of sealevel fluctuations in the Southwest Pacific region point to a need to modify the ice history model of Peltier and Andrews (1976) in which all melting was confined to the Arctic, with the melting ending at 7000 yrs B.P. Sealevel observations in the far-field alone are inadequate to determine the ice load but either the end of Arctic melting has to be brought forward by about 1000 yrs or a substantial amount of melting must have taken place in Antarctica in the period 17,000 to 7000 yrs B.P. The far-field sealevels are generally insensitive to the average mantle viscosity within the range $10^{21}$–$10^{22}$ P. These viscosities should be considered as lower limits since the effects of an elastic lithosphere have been ignored. On the other hand, Peltier et al.'s (1978) values for the viscosity are probably upper limits since they assume an excessively thick (100 km) elastic lithosphere. The above estimate compares favourably with $(2-6) \cdot 10^{21}$ P found by Nakiboglu and Lambeck (1980) from the secular polar motion and acceleration of the Earth's rotation.

The discrepancies between the observed and predicted sealevels provide estimates of vertical motion if these motions exceed about 0.5 mm yr$^{-1}$. Only two locations provide clear evidence for vertical motion: The Huon Peninsula is undergoing uplift of between 1.2 and 1.7 mm yr$^{-1}$ in accordance with Chappell's (1974) estimate of 1.8 mm yr$^{-1}$ based on evidence over a time interval of $10^3$ yrs; the Hawaiian island of Oahu is undergoing subsidence at a rate of about 1–2 mm yr$^{-1}$, in general agreement with the notion that stress relaxation is taking place. The stations along the southeastern and east coast of Australia are suggestive of small uplift of perhaps 0.5–1.0 mm yr$^{-1}$. Areas of known stability such as the New Zealand and southern Malacca Straits are sites for which we also find insignificant uplift results.

ACKNOWLEDGEMENTS

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REFERENCES


APPENDIX

Site 1: New Zealand

The data come from two locations. The longer record is based on estuarine and offshore samples from a relatively stable area near Christchurch on the South Island (43°33'S, 172°40'E). The point dated at 9400 ± 120 yrs B.P. is considered unsatisfactory by Schofield (1964) and Suggate (1968) due to some difficulty in relating the sealevel indicator to the datum level and to post-depositional processes. The data for the last 4000 yrs comes from a chenier sequence at Firth of Thames; North Island (36°55'S, 174°45'E). This area is also thought to be tectonically stable at least during the late Pleistocene (Schofield, 1960).

References

Sites 2 and 3. Melbourne (Victoria) and Moruya (New South Wales), Australia

These data indicate considerable scatter which may arise from (1) variations in the tidal range between the various localities, (2) stratigraphic compaction, (3) use of different indicators of past sealevel. The age estimates have not been corrected for the 14C reservoir effect (Thorn and Chappell, 1975) which may mean that these ages are 200-500 yrs too old.

References

Site 4. Southeast Queensland, Australia

This sealevel curve is compiled from several sources and covers an area from 22°22'S, 149°47'E to 27°33'S, 153°08'E. The region is believed to be relatively stable.

References
Flood, P.G., et al., 1979 Qld Gov. Min. 1, 80: 444-447

Site 5. Gulf of Carpentaria, Australia

Most data are from slightly emerged beach and chenier ridges from the prograded southeast and east plains along the Gulf. These are indicative of the tidal zone during the Late Holocene. The three points prior to 14,000 yrs B.P. are from submerged beach rock (wood in fossil soil and shell material) in the Arafura Sea (9°07'S, 133°52'E and 10°20'S, 130°46'E respectively).
References

Site 6. Northern Great Barrier Reef, Australia

The Late Holocene sealevels are based on geomorphic features such as microatolls, sand and shingle cays, and ramped rocks and cover an area between 15°52′S, 145°35′E and 14°18′S, 144°52′E.

References

Site 7. Huon Peninsula, New Guinea

The data are from in-situ corals and giant clams along a seacliff section of the post-glacial barrier reef. This formation belongs to the lower sequence of raised coral reef complexes extending from the last interglacial period (133,000–120,000 yrs B.P.) to 6000–8000 yrs B.P.

References

Site 8. Southern Malacca Straits, Malay Peninsula

Paleo sealevel evidence comes from the paludal stratigraphy of peats and mangrove swamps along the coast between Port Dickson (2°31′N, 101°48′E) to Singapore (1°20′N, 103°50′E). Secondary effects such as compaction on the depth and contamination of the 14C ages are believed to be small.

References

Site 9. Micronesia: East Caroline and Marshall Islands

Data from the East Caroline Islands are based on paludal stratigraphy while evidence from the Marshall Islands is from reef material associated with coral-reef crest facies. All depths have been converted to low-water level in accordance with the datum level used throughout.

References
Site 10. East China Coast

The evidence for the relative sealevels comes from three data sets:

1. Drowned geomorphic features from the East China Sea (shell ridges, sand beaches, marshes, estuary and delta deposits—29°17'N, 126°33'E to 38°31'N, 123°05'E). These record the early eustatic submergence of the shallow shelf from about 15,000 to 9000 yrs B.P. The samples have come from dredging rather than coring and there may be substantial uncertainties resulting from post-depositional processes.

2. Raised Late Holocene beach deposits along the East China Sea Coast (29°46'N, 120°53'E to 33°01'N, 120°43'E).

3. Raised shell ridges from the Gulf of Bo Hai (37°17'N, 119°24'E to 39°45'N, 117°16'E) for the last 6000 yrs. Datum levels are unknown for these sites and details on the 14C dating are lacking.

References

Site 11. Oahu, Hawaii

Sealevels are based on shallow cores taken through a fringing reef crest at Hanauma Bay, Oahu.

References