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Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records

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Abstract

A major discrepancy between the Late Quaternary sea level changes derived from raised coral reef terraces at the Huon Peninsula in Papua New Guinea and from oxygen isotopes in deep sea cores is resolved. The two methods agree closely from 120 ka to 80 ka and from 20 ka to 0 ka (ka = 1000 yr before present), but between 70 and 30 ka the isotopic sea levels are 20–40 m lower than the Huon Peninsula sea levels derived in earlier studies. New, high precision U-series age measurements and revised stratigraphic data for Huon Peninsula terraces aged between 30 and 70 ka now give similar sea levels to those based on deep sea oxygen isotope data planktonic and benthic $\delta^{18}O$ data. Using the sea level and deep sea isotopic data, oxygen isotope ratios are calculated for the northern continental ice sheets through the last glacial cycle and are consistent with results from Greenland ice cores. The record of ice volume changes through the last glacial cycle now appears to be reasonably complete.

Keywords: Quaternary; Huon Peninsula; U-234/Th-230; O-18/O-16; sea-level changes

1. Introduction

Sea level variations over the last 140 ka (ka = 1000 years before present) derived from raised coral terraces at Huon Peninsula (HP) in Papua New Guinea, dated by $^{230}Th/^{234}U$, are better defined than from any other land-based site, and generally march in step with $\delta^{18}O$ variations in deep ocean cores [1]. Furthermore, correlation of the HP sea levels with deep sea core records is supported by oxygen isotope data from giant clams (*Tridacna gigas*) that are preserved in the coral terraces [2]. Sea levels deduced from HP for oxygen isotope stages 5a (83 ka) and 5c (104 ka) are supported by results from other places, including Barbados [3], Timor [4] and Haiti [5]. HP terraces dated around 60 and 40 ka have age equivalents at Kikai–Jima [6] and Vanuatu

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[7]. The evidence supports the conclusion of Chappell and Shackleton [1], that the Late Quaternary sea levels deduced from HP represent a global pattern. Because the Huon Peninsula is situated near the equator and has no continental shelf, the hydro-isostatic and global gravitational effects that modify sea level changes [8] are expected to have had a small effect.

Oxygen isotope variations in deep sea cores reflect changes in both the isotopic composition of seawater and its temperature. Shackleton [9] subtracted the temperature effect from the high resolution benthic $\delta^{18}O$ record obtained from core V19-30, and proposed that the reduced isotopic record represents sea level changes. To do this, the temperature correction was approximated by taking smoothed differences between the benthic isotopic record of V19-30 and planktic data from equatorial core RC17-177, and subtracting these from the V19-30 record. This approach assumes that average sea surface temperature changes recorded in the planktic forams at the site of RC17-177 were negligible during the last 140 ka, and also assumes that the residual variation of $\delta^{18}O$ in deep ocean water, after allowance for temperature effects, was driven by changes in ice volume and therefore reflects sea level changes.

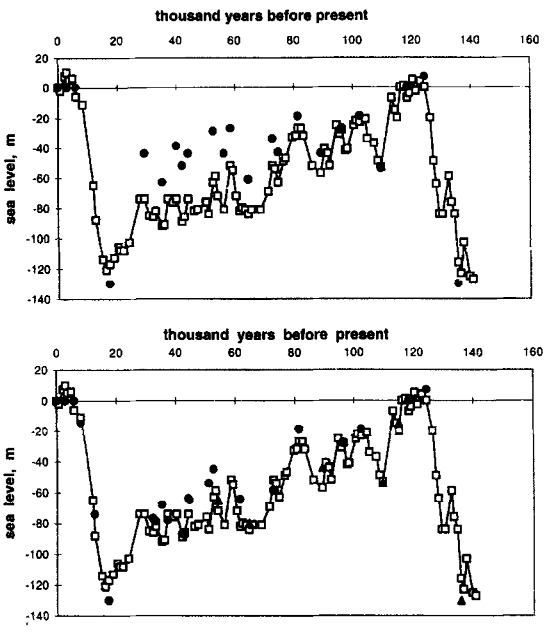

Fig. 1. Upper: Previous estimates of sea levels for the last 140 ka. ● = sea levels deduced from coral terraces at Huon Peninsula [1]; □ = estimates based on combined benthic and planktic deep sea core $\delta^{18}O$ data [9]. Lower: ● = new results between 30 and 75 ka (this paper) and previous results (0–20 ka and 75–145 ka) from HP; ▲ = undated sea levels based on lowstand deposits in the raised reef and fan-delta tracts at HP (cf. [13,25]); □ = isotopic sea levels as above.

The isotopic sea levels agree quite closely with HP results for the intervals 80–140 ka and 0–20 ka but disagree significantly in the interval from 30 to 70 ka, where the isotopic sea levels are 20–40 m lower than the HP results (Fig. 1). The differences mean that either the isotopic sea levels or the HP sea levels are wrong, or that both are wrong, between 30 and 70 ka. It is important to resolve this problem because changes in sea level, equated with ice volume, through the last glacial cycle are used to test climatic models that incorporate ice sheets [10,11].

Sea level estimates based on raised coral terraces are sensitive to the measured ages of the terraces, because the product of terrace age and uplift rate is subtracted from observed terrace height to give the associated sea level (see below). We have redated the Huon terraces that were previously dated as between 30 and 62 ka and, according to our results, all previous age estimates were too young. Hence, the recalculated sea levels between 30 and 70 ka are lower than estimated by Chappell and Shackleton [1] and lie closer to the isotopic sea levels. In this paper, we report the new age measurements and the recalculated sea levels. We also re-examine the temperature effect upon $\delta^{18}O$ data from core V19-30 and our results support the conclusion [1,9] that the mean deep ocean water temperature was 1.5–2.0°C cooler during the last glacial period than in the interglacials. We also show that the sea level and isotopic data are consistent with reasonable isotopic values for northern ice sheets during the last glaciation.

2. Field survey and sampling

We chose two field transects, referred to as KANZ (1 km north west of Kanzarua village) and BOBO (Bobongara headland, also known as Fortification Point), as the best sites for reconstructing late Quaternary sea levels (locations: Fig. 2). The transects intersect well-formed flights of coral terraces extending from the Holocene (reef tract I) to the Last Interglacial (reef tract VII). Terraces are not disrupted by faults or landslides at these sites and exposures of underlying coral reefs are good. These transects were carefully resurveyed in 1988 and 1992 and some differences from previous data were discovered. Coral samples collected from the BOBO and KANZ transects were precisely positioned on the 1988–1992 surveys. Coral reef biofacies and lithofacies were recorded in detail and the palaeoenvironment and maximum range of water depth in which each sample grew was identified from the facies [12]. Stratigraphy and sample sites are shown in Fig. 3. We follow the Roman numeral terrace nomenclature used at HP by Chappell [13], with alphabetical subscripts to denote subdivisions of each reef complex (Fig. 3 and Table 1). Additionally, three samples were collected from reef VI at the Kwambu section (location, Fig. 2; summary descriptions given by [3,13]), which we resurveyed.

3. $^{230}Th/^{234}U$ age measurements

Samples were dated by both alpha spectrometry and thermal ionisation mass spectrometry. Precisely located coral samples, in growth position, that show no visible evidence of diagenesis or secondary infilling were selected. To be acceptable, samples had to be 98–100% aragonite before cleaning, with a U concentration of 2–4 ppm. Furthermore, after U-series measurements, only those samples were accepted that had a calculated initial $^{234}U/^{238}U$ ratio of 1.130–1.150, which brackets the modern sea water value of 1.144 ± 0.007 [14]. Samples were examined microscopically and were cleaned with a dentist's drill, where necessary. Results from sample

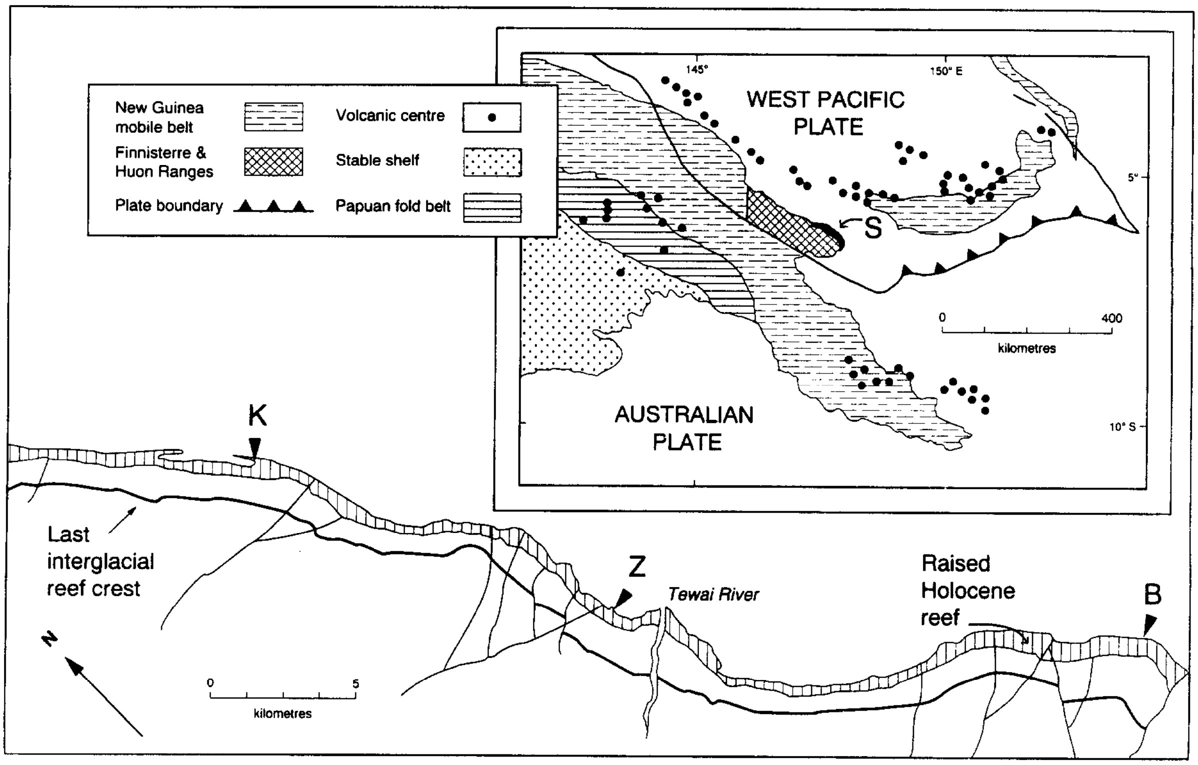

Fig. 2. Location of study area at Huon Peninsula, Papua New Guinea (S in inset), and locations of detailed transects across coral terraces: B = BOBO; Z = KANZ; K = Kwambu. Mean Late Quaternary uplift rates at these transects are: 3.3 m ka^{-1} at BOBO, 2.5 m ka^{-1} at KANZ and 1.9 m ka^{-1} at Kwambu. Thin lines oblique to coast are traces of faults and margins of very large landslides.

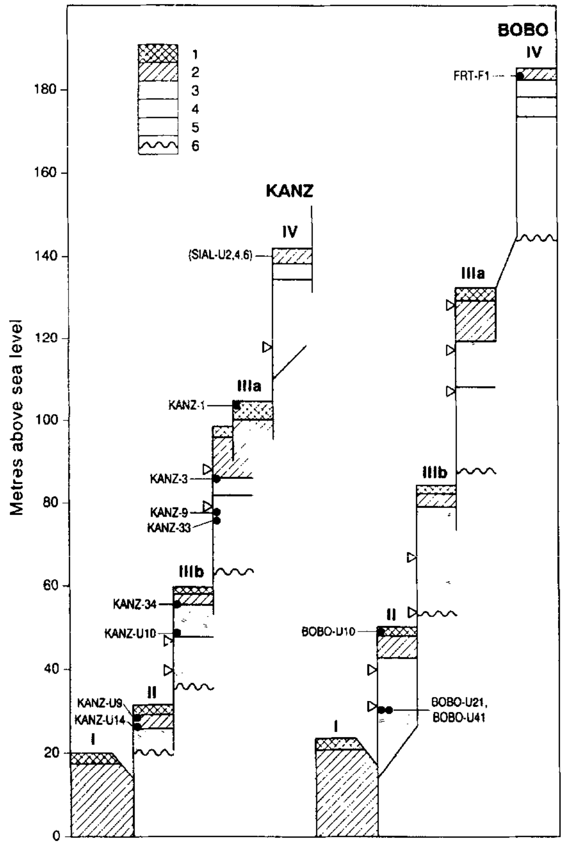

Fig. 3. Summary stratigraphy of major coral reefs associated with terraces I–IV at BOBO and KANZ transects, and locations of samples in Table 1. Key: *1* = reef platform; *2* = reef crest and buttress; *3* = upper reef slope; *4* = middle and lower reef slope; *5* = forereef and deeper water detrital limestone; *6* = base of exposure. Small triangles indicate positions of minor erosional terraces that are primarily erosional. Samples SIAL-U2, 4 and 6 are transferred laterally from the Reef IV crest at the surveyed Kwambu transect (see text).

sites shown in Fig. 3 satisfy the above criteria and are listed in Table 1.

Dating of the samples by $^{230}Th/^{234}U$ was undertaken jointly by A. Omura at Kanazawa University, by alpha spectrometry, and T. Esat and M. McCulloch at the Australian National University, by thermal ion mass spectrometry (TIMS). Descriptions of methods at the Kanazawa laboratory are published in Japanese and details in English can be supplied by A. Omura.

For TIMS dating, procedures for chemical separation of U and Th from corals were similar to those described by Edwards et al. [15]. The ^{233}U and ^{229}Th tracers were calibrated using a uraninite sample in secular equilibrium. Thorium isotope ratios were determined using multiple Faraday cups and charge-collection TIMS [16]. In this method the usual high-value resistor at the feed-back loop of an electrometer is replaced by a capacitor, which accumulates the charge collected in a Faraday cup, resulting in a time-varying voltage proportional to the ion current. Precision and stability of the system are significantly better than for electron multipliers. Uranium isotopes were measured with a combination of Faraday cups and an electron multiplier. Esat [16] and Stirling et

Table 1
New $^{230}Th/^{234}U$ ages and sea level estimates from Huon reefs I–IV

Sample	Height (m)	Reef Number	Water Depth	$\delta^{234}U(T)$ (‰)♣	Age (ky)♦	Sea level♠ (m, MSL) max	min
BOBO-U21	30	IIc	2-5	144 ± 9	33.4 ±0.6	84	73
BOBO-U24	30	IIc	2-5	142 ± 10	33.0 ±0.5	82	72
BOBO-U10*	49	IIa	0-2	142 ± 2	37.8 ±0.3	79	73
KANZ-U14*	26	IIa	2-5	139 ± 2	34.8 ±0.3	72	64
KANZ-U9	28	IIa	1-3	152 ± 8	41.8 ±0.6	92	82
KANZ-U9*	28	IIa	1-3	137 ± 2	42.2 ±0.3	91	84
KANZ-U10	49	$IIIc_1$	5-15	150 ± 10	43.9 ±0.7	73	55
KANZ-34	56	IIIb	2-5	132 ± 9	44.5 ±0.7	71	59
KANZ-9*	78	$IIIa_1$	2-10	137 ± 2	54.6 ±0.7	56	59
KANZ-33	77	$IIIa_1$	2-10	143 ± 9	45.8 ±0.7	53	37
KANZ-4	86	$IIIa_m$	2-5	131± 9	51.2 ±0.8	60	48
KANZ-3	96	$IIIa_u$	1-3	136± 9	51.8 ±0.8	53	41
KANZ-1*	105	$IIIa_u$	0-2	139± 2	61.4 ±0.6	70	59
FRT-F1	183	IV	0-5	147± 14	72.8 ±2.2	72	45
SIAL-U2*	-	IV	0-5	138± 2	64.9 ±1.7	**	
SIAL-U4*	-	IV	0-5	142± 2	69.9 ±1.1	**	
SIAL-U6*	-	IV	0-5	150± 2	71.9 ±0.5	**	

* Ages measured by TIMS; other ages measured by alpha spectrometry. ** Ages for SIAL-U4 and SIAL-U6 grouped with FRT-F1 (see text). ♣ $\delta^{234}U = \{[(^{234}U/^{238}U)/(^{234}U/^{238}U)_{eq}]-1\} \times 10^3$. $(^{234}U/^{238}U)_{eq}$ is the atomic ratio at secular equilibrium and is equal to $\lambda_{238}/\lambda_{234} = 5.472 \times 10^{-5}$ where λ_{238} and λ_{234} are the decay constants for ^{238}U and ^{234}U, respectively. $\delta^{234}U(0)$ is the measured value, the initial value is given by $\delta^{234}U(T) = \delta^{234}U(0)e^{\lambda_{234}T}$, where T is the age in years. Acceptable samples have $\delta^{234}U(T)$ values overlapping the range $149 \pm 10‰$. ♦ $^{230}Th/^{234}U$ ages are calculated iteratively using $^{230}Th/^{238}U = (\delta^{234}U(0)/1000)(\lambda_{230}/(\lambda_{230}-\lambda_{234}))(1-e^{(\lambda_{234}-\lambda_{230})T}) - e^{-\lambda_{230}T}) + 1$ where T is the age in years; λ_{230} is the decay constant for ^{230}Th. $\lambda_{238} = 1.551 \times 10^{-10}\ y^{-1}$, $\lambda_{234} = 2.835 \times 10^{-6}\ y^{-1}$ and $\lambda_{230} = 9.195 \times 10^{-6}\ y^{-1}$. ♠ Sea level at the time the sample grew is estimated from uplift rate, sample height above present sea level and paleo water depth at which the sample grew. The sea level range includes uncertainties in these parameters (see text).

al. [17] give full details of the measurement techniques.

4. Calculation of sea levels from coral terraces

The sea level (S) at time t, relative to present sea level, for each dated coral listed in Table 1 was calculated as follows:

$$S = (H + z) - Ut \quad \text{with}\, U = (H^* - S^*)/t^* \qquad (1)$$

where t = age of a coral sample from height H above sea level, corrected for the palaeo water depth, z, in which the dated sample grew, and U is the tectonic uplift rate at the site. U is calculated from the height H^* of a reference terrace of age t^* at the same terrace transect, which formed when sea level was S^*. The reference terrace was the crest of the reef that formed during the climax of the Last Interglacial (equivalent to oxygen isotope stage 5e in deep sea cores). We used a value of $S^* = 5 \pm 2$ m, on the basis of data from raised reefs at tectonically stable sites in many parts of the world [4].

Eq. (1) assumes that uplift rate (U) has been constant at a given terrace transect. U varies along the coast at HP [1,3], and U also varies on a 1000 year time scale because the uplift process at HP is dominated by metre-scale coseismic uplift events [18]. However, the assumption that U is constant over time scales exceeding several thousand years at any given transect is supported by close agreement at each site between the uplift rate since the Last Interglacial and the average Holocene uplift rate [18]. Furthermore, this assumption is supported because similar values of S are found for each terrace at different sites with different uplift rates, within measurement errors (cf. [1,3]).

The value of U is affected by uncertainties in the Last Interglacial reference sea level (S^*) and its timing (t^*). S^* may have been a few metres lower than our adopted value of 5 ± 2 m [17] but the effect of this uncertainty is negligible because the height (H^*) of the Last Interglacial reference terrace (crest of Huon reef VII) is very much larger than the uncertainty in S^* ($H^* = 320$ m at KANZ and 403 m at BOBO). The age of the crest is taken as 122 ± 4 ka, on the basis of dating of the Last Interglacial high sea level at many sites around the world including HP [17,19]. Accounting for all uncertainties, U for KANZ and BOBO are 2.8 ± 0.1 and 3.3 ± 0.1 m/ka, respectively.

The precision of a sea level estimate also depends on the value of z and t in Eq. (1). In previous reports, the palaeo water depth z was not considered because sea level estimates were based on heights of terrace surfaces and not on the heights (H) of the actual corals that were dated, which were usually collected from exposures significantly lower than the surface of the associated terrace. This procedure is erroneous; a 52 m drillhole in the Holocene terrace at HP proved to be 13 ka at the base and 7 ka at the top [20], which shows that sample ages from below a terrace surface do not represent its time of emergence. Hence, in this paper, a sea level is calculated for each dated sample. In each case, the value of z was estimated by using the coral community and reef zones described by Pandolfi and Chappell [12]. As regards precision of the age data (t in Eq. (1)), age errors for the results in Table 1 are substantially smaller than for previously reported dates from HP and initial $^{234}U/^{238}U$ ratios are within the narrow limits of acceptability.

5. Results

Paired samples from each of three reefs were used to test age reproducibility. Samples in each pair were collected within 10 m of each other (see Table 1), and the ages within each pair are statistically identical; for example, BOBO-U21 and BOBO-U24 in reef II; KANZ-U10 and KANZ-34 in reef IIIb; KANZ-3 and KANZ-4 in reef IIIa. Ages measured by alpha counting and by TIMS for the same sample (KANZ-U9) gave the same result. Furthermore, samples from each reef complex gave younger ages than samples from the next higher and stratigraphically older reef, at both KANZ and BOBO. These results give us confidence in our sample screening and dating procedures. The only disagreement is between KANZ-33 and KANZ-9, which were collected within 1 m of each other from the same reef ($IIIa_1$). We cannot reject either in terms of dating criteria but KANZ-33 appears to be anomalously younger than other samples from reef IIIa. We note that, within a given reef complex, samples from a position which is lower and seawards of one that is higher and further inland are not necessarily expected to be older, because each reef can preserve a downstepping, regressive structure at its front; this has been demonstrated by ^{14}C dating of the Holocene reef complex [21]. A regressive structure is implied by age relations between BOBO-U21, BOBO-U24 and BOBO-U10 in the reef II complex at the BOBO transect, although this was not defined by field exposures, which are restricted by vegetation and lack of deep gullies in the reef II complex.

Calculated sea levels (Table 1) are plotted against isotopic sea levels [9] in Fig. 4. Trapezoidal boxes show sea level uncertainties based on U-series age errors together with the upper and lower limits for U and for z. For each box, the vertical edges represent the combined uncertainties in U and z; box width represents the age error, and the sloping edges represent uplift rates. The oxygen isotope sea level estimates also have error terms. The deep sea core chronology follows the orbitally tuned scheme which has error estimates of up to ± 5 ka, varying through time [22]. For the isotopic sea level values, an error of ± 7 m is adopted, because isotopic sea levels for the last 6 ka show a range of about 14 m, although sea level itself was almost constant during this interval.

The new sea level results are significantly lower than previous values in the range 30–70 ka [1], and all results except for KANZ-33 are close to the isotopic sea levels. Our new estimates are lower principally because the measured ages of reefs II–IIIa are greater than previously reported, which makes for lower sea levels according to Eq. (1). Our dates from reef II are 3–12 ka older than the previous age estimate [23]; from IIIb they are 2–4 ka older and from IIIa they are 7–17 ka yr older than the previous values [1]. Furthermore, the structure of reef IIIa is more complex than was previously recorded [13]. The main IIIa terrace (Fig. 3), which is the largest in the sequence between reefs I and VII at KANZ and BOBO, is composite and appears to have formed in two stages. The younger stage is exposed in cliffs of regressive terraces below IIIa; dated samples near the crest (KANZ-3, KANZ-4) give an age of 51–52 ka. KANZ-9, from lower in the IIIa frontal section, is a little older (54 ka). KANZ-33, sampled close to KANZ-9, gave an age of 45.8 ± 0.7 ka, which is not

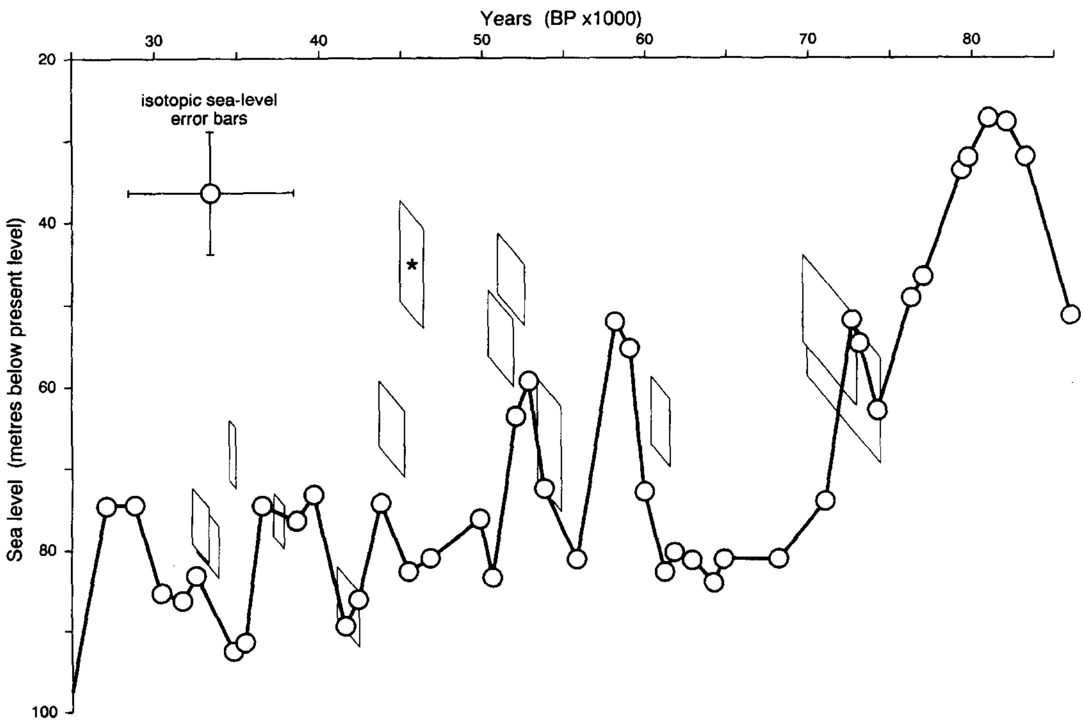

Fig. 4. Sea level estimates for samples listed in Table 1 (trapezoidal boxes; see text for explanation of error bands), and isotopic sea level estimates from Shackleton [2] (circles). Result for KANZ-33 (*) conflicts with other samples from the same reef and may be inaccurate.

consistent with the others in this group, as noted previously. The crest of the older stage is traced by a low terrace on the broad surface of IIIa at KANZ and may be disconformable with the outer part of the IIIa reef; sample KANZ-1 indicates that the older stage culminated at about 61 ka.

Four new results from reef IV are about 10 ka older than the age of 59–62 ka previously assigned to this reef. One new date (FRT-F1, 72.8 ± 2.2 ka) is from reef IV on the BOBO section; previous ESR and TIMS ages from reef IV are consistent with this date [24]. We failed to find samples suitable for dating at reef IV at KANZ and the other new dates are from the crest of reef IV at the Kwambu/Sialum section that has been described previously [3,13], which we resurveyed. Results from FRT-F1, SIAL-U4 and SIAL-U6 agree within errors and give a mean age of 71.3 ± 1.6 ka. SIAL-U2 (64.3 ± 1.7 ka) is significantly younger than the others from reef IV and is not included in this average. An age of 71.3 ± 1.6 ka is assigned to the crest of reef IV and the recalculated sea level for IV is plotted in Fig. 4. The new dates from IV conflict with the previous age of 61 ka that was based on four dates from the KANZ section [3]. The previous 'reef IV' ages are the same as our result for the older stage of IIIa and we suspect that the relevant samples, collected in 1971 before the first accurate survey at KANZ (carried out in 1973), may have come from the older part of IIIa.

We regard the results presented here as more reliable than previous sea levels from HP, for the interval 30–70 ka; they are based on detailed stratigraphic sections, carefully resurveyed, the $^{230}Th/^{234}U$ ages are more precise than previously obtained, and the method of sea-level calculation has been improved. Central values of the error boxes in Fig. 4 are replotted in Fig. 1(lower), for comparison with the previous HP sea levels of Chappell and Shackleton [1] in the upper part of Fig. 1. HP sea levels at 8.5 and 12.5 ka (from [20]) and previously dated sea levels prior to 80 ka (from [1]) also are plotted in Fig. 1. Five undated points, derived from stratigraphic relationships [13,25] are also plotted. Clearly, HP and isotopic sea levels in Fig. 1(lower) agree better than those in Fig. 1(upper), throughout the last glacial cycle. To simplify the direct comparison of 'new' and 'old' sea levels, we have plotted HP sea levels as points in Fig. 1, and not as the transgressive–regressive fluctuations that were shown in early accounts [1,3].

6. Implications for deep sea temperatures

Shackleton [9] derived isotopic sea levels by taking smoothed $\delta^{18}O$ differences between the benthic core V19-30 and planktic core RC17-177, and subtracting these differences from the V19-30 record. As explained earlier, this procedure assumes that average sea surface temperature changes were negligible at the RC17-177 site. Support for this assumption in seen in the agreement between the isotopic sea levels and the revised sea levels from HP presented here (Figs. 1 and 4); this, in turn, lends support to the results of CLIMAP [26], which indicate that glacial–interglacial SST changes were small over much of the tropical ocean.

Shackleton's procedure implies that deep ocean temperature variations at the site of V19-30 are represented by smoothed $\delta^{18}O$ differences between cores V19-30 and RC17-177. Deep ocean temperatures can also be derived from combined sea level and benthic $\delta^{18}O$ data; estimates based on previous sea level values from HP suggested that deep ocean temperatures at V19-30 were 1.5–2.5°C cooler during the last glacial period (isotope stages 2–5d) than during the Holocene or Last Interglacial [1].

We have recalculated deep ocean temperature variations through the last glacial cycle, using the new HP sea levels together with $\delta^{18}O$ data from V19-30. Relative to present-day temperature and isotopic values (T_o and $\delta^{18}O_o$), the temperature change ($T_o - T_i$) for the ith data point in the paired sea level/$\delta^{18}O$ time series is given by:

$$T_o - T_i = (\delta^{18}O_i - \delta^{18}O_o - K \cdot S_i)/F \qquad (2)$$

where S_i and $\delta^{18}O_i$ are the sea level and $\delta^{18}O$ values at i. K represents the change of mean ocean $\delta^{18}O$ for a given change of sea level and F is the temperature-dependent slope of oxygen isotope fractionation for calcium carbonate ($-0.23‰\ °C^{-1}$ [1]). Present sea level is taken as zero and S is negative for sea levels below zero.

To determine K, we regressed HP sea levels against corresponding $\delta^{18}O$ from V19-30 through the last glacial period, using the new HP sea levels

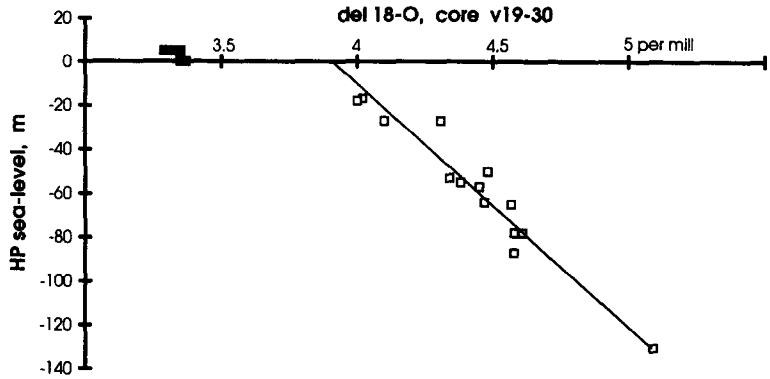

Fig. 5. Regression of HP sea levels versus benthic $\delta^{18}O$ (from core V19-30) during the last glacial cycle (□). Sea level data from Fig. 4 and Table 1 were used for the interval 30–70 ka and other sea levels are from dated points in Fig. 1(lower), except for two points during post-glacial sea level rise at 8.5 and 12.5 ka. Sea level and isotopic data for the present and for the last interglacial, shown by solid boxes, were not used in the regression.

(Table 1) in the time range 30–70 ka and the previous HP sea levels [1] for 18 ka and 80–110 ka (Fig. 5). The data closely fit a straight line with $K = 0.009‰\ m^{-1}$ with $R^2 = 0.93$, which is a better fit than for the previous HP sea levels ($R^2 = 0.8$). The straight line implies that, within uncertainties, K and $(T_0 - T_i)$ are constant from 18 to 110 ka. The intercept at $S = 0$ is $\delta^{18}O = 3.93$, compared with $\delta^{18}O_0 = 3.42$ which is the mean $\delta^{18}O$ for the last 6000 years in core V19-30 (Fig. 5). Hence, from 18 to 110 ka, the deep water temperature was cooler than at present by $(3.42–3.93)/0.23 = -1.8°C$.

This result reinforces the previous conclusion [1,9], that deep ocean temperatures were 1.5–2.0°C cooler than present during the last glacial cycle. This idea has been criticised on the grounds that it implies unrealistically rapid ocean turnover [27]. Hence, as a second means of checking our regression of sea level versus $\delta^{18}O$, the oxygen isotopic composition of last glacial ice sheets was calculated as follows. Let $\delta^{18}O_{ice}$ be the mean isotopic composition of the ice which equates to a sea level fall of S metres, associated with a shift $\Delta\delta^{18}O$ of the average ocean isotopic composition. To a first approximation, the dependence of $\delta^{18}O_{ice}$ on $\Delta\delta^{18}O$ and S is:

$$\delta^{18}O_{ice} = -\Delta\delta^{18}O \cdot (H - S)/S \quad (3)$$

where H is mean ocean depth (3790 m). Eq. (3) assumes that the mean isotopic composition of glacial-age ice was constant for all values of S and that ocean volume is proportional to depth. As $\Delta\delta^{18}O/S = K$ (in Eq. (2)) and $H \gg S$, our value of $K = 0.009‰\ m^{-1}$ gives $\delta^{18}O_{ice} = -34‰$. By comparison, if we assume that deep ocean temperatures were constantly the same as today throughout the last glacial cycle and use the $\delta^{18}O$ data from V19-30, together with HP sea levels, glacial-age values of $\delta^{18}O_{ice}$ calculated by Eq. (3) range from $-50‰$ to $-120‰$. The average isotopic value of Greenland ice that formed during the last glacial period, which is a reasonable guide to the composition of the northern continental Pleistocene ice sheets, is about -35 to $-40‰$ [28,29]. Hence, the conclusion that the deep ocean was 1.8°C cooler during the glacial period is consistent with reasonable values for $\delta^{18}O_{ice}$.

7. Conclusions

Our new results from Huon Peninsula reefs which formed 30–70 ka (reef complexes II–IV) are better constrained stratigraphically and the seventeen new $^{230}Th/^{234}U$ ages are more precise than the previous results. Except for one pair (KANZ-33 and SIAL-U2), ages of sample pairs agreed closely; statistically identical results were obtained by TIMS and alpha counting, and the ages are stratigraphically consistent.

Sea levels calculated from the dated samples, corrected for tectonic uplift and the water depth in which each sample grew, are close to isotopic sea levels derived by Shackleton [9]. Most of the dated samples from reefs II to IV represent reefs that formed close to relative sea level maxima (highstands). We have no dates from lowstand zones of the raised reef but previously reported stratigraphy indicates fluctuations in relative sea level during the formation of each reef complex [13,25]. We expect that lowstand reef and deltaic deposits, which are stratigraphically sandwiched between the dated coral samples, correspond to isotopic sea level minima.

Sea levels derived from Huon Peninsula and from oxygen isotopes now appear to be in agreement throughout the last glacial cycle. The two approaches are complementary; time series from high resolution isotopic records are more continuous than the Huon Peninsula record, while the new U-series dating provides more precise chronology than is available for deep sea cores.

Finally, the new sea level results further reinforce the previous inference [1,9] that deep ocean temperatures at the site of core V19-30 were 1.5–2.0°C cooler than present, through the last glacial cycle. The isotopic difference ($\Delta\delta^{18}O$) between the glacial maximum and the last 6 ka in V19-30 is 1.7‰; Fig. 5 indicates that about 0.5‰ represents the temperature effect and 1.2‰ represents the ice volume effect. Other benthic cores show a similar glacial–interglacial isotopic difference of $\Delta\delta^{18}O = 1.7‰$ [9], which implies that glacial-age cooling of the deep ocean was rather widespread. Provided that the spatial variation of $\delta^{18}O$ in the surface ocean during glacial times was similar to the present, it also follows that glacial surface water temperatures were probably similar to present temperatures at sites where planktic values of $\Delta\delta^{18}O$ (glacial–interglacial) are about 1.2‰. *[CL]*

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