Quaternary Science Reviews, Vol. 6, pp. 183–190, 1987.

0277-3791/87 $0.00 + .50

OXYGEN ISOTOPES, ICE VOLUME AND SEA LEVEL

N.J. Shackleton

Godwin Laboratory for Quaternary Research, Free School Lane, Cambridge CB2 3RS, U.K.

A careful comparison is made between the most detailed records of sea level over the last glacial cycle, and two high-quality oxygen isotope records. One is a high-resolution benthonic record that contains superb detail but proves to record temperature change as well as ice volume; the other is a planktonic record from the west equatorial Pacific where the temperature effect may be minimal but where high resolution is not available. A combined record is generated which may be a better approximation to ice volume than was previously available.

This approach cannot yet be applied to the whole Pleistocene. However, comparison of glacial extremes suggests that glacial extremes of stages 12 and 16 significantly exceeded the last glacial maximum as regards ice volume and hence as regards sea level lowering. Interglacial stages 7, 13, 15, 17 and 19 did not attain Holocene oxygen isotope values; possibly the sea did not reach its present level. It is unlikely that sea level was glacio-eustatically higher than present by more than a few metres during any interglacial of the past 2.5 million years.

INTRODUCTION

It is generally agreed that to a first approximation, the oxygen isotope record that is recovered by analysing foraminifera in deep-sea sediment cores gives a history of global continental ice volume and hence of the glacio-eustatic component of sea-level change. However, it is also widely understood that this is only a first approximation; the purpose of this review is to explore the possibility of making a closer approximation, and then to investigate the long records available.

One source of uncertainty has been well reviewed by Mix and Ruddiman (1984): the average isotopic composition of the former ice sheets must have varied with their size and their latitudinal position. Thus ocean isotopic composition was never a linear function of ice volume and hence was never a linear function of sea level. Mix and Ruddiman show how a growing ice sheet is less isotopically light than a steady-state ice sheet, so that ocean oxygen isotope values might continue to become more positive when the ice sheets were already stabilized at maximum extent.

A more serious complication is that no deep-sea sediment core preserves a perfect record of the history of the isotopic composition of the ocean. One reason is that whatever foraminifera are chosen for isotopic analysis, variations in the temperature of the ocean water in which they lived have to be considered. Another is the generality that no geological record is perfect; in particular, bioturbation is a virtually universal source of degradation for deep-sea records.

It will be recalled that when Emiliani (1955) first obtained oxygen isotope records derived from the analysis of planktonic foraminifera from the Caribbean and the tropical Atlantic Ocean, he presented the data scaled for sea surface temperature. The most convincing evidence that this model was incorrect came from the analysis of benthonic foraminifera, where the isotope values for glacial levels would imply temperatures well below freezing point if presented in an exactly analogous manner. Thus the first publication of a fully detailed benthic record by Ninkovitch and Shackleton (1975) led to the view that this would be the best source for an ideal ice volume record, since the temperature component might be regarded as negligible. Shackleton (1977) suggested that the observed oxygen isotopic difference between Stage 1 and Stage 2 in high-resolution benthic records of about 1.6‰ implied a sea level lowering of around 160 m, consistent with the larger estimates of Denton and Hughes (1981) for ice volumes at glacial maximum time.

A serious difficulty arises in comparing the oxygen isotope record of such cores with the sea level record derived from dated marine terraces. It seems well established from studies on Barbados (Mesolella *et al.*, 1969), New Guinea (Bloom *et al.*, 1974), Haiti (Dodge *et al.*, 1983) and elsewhere that during substages 5a and 5c at respectively 82 ka and 104 ka BP the sea was only around 20 m lower than its position during 5e at 124 ka BP. On the other hand equivalent points in the detailed benthic oxygen isotope sequences are generally about 0.7‰ more positive than the 5e extreme, implying of the order 70 m sea level difference.

Chappell and Shackleton (1986) made a detailed comparison between the best available oxygen isotope record from the deep Pacific and the sea level record inferred from an analysis of the uplifted sequence on the Huon Peninsula, New Guinea. Figure 1 shows their comparison. It is obvious that a serious disagreement exists, and they concluded that a step-like cooling in the deep water occurred about 115 ka BP that was reversed about 11 ka BP. Thus the simple interpretation of the benthic oxygen isotope record as an ideal ice volume or sea level record is no longer tenable.

An alternative strategy for interpreting the marine

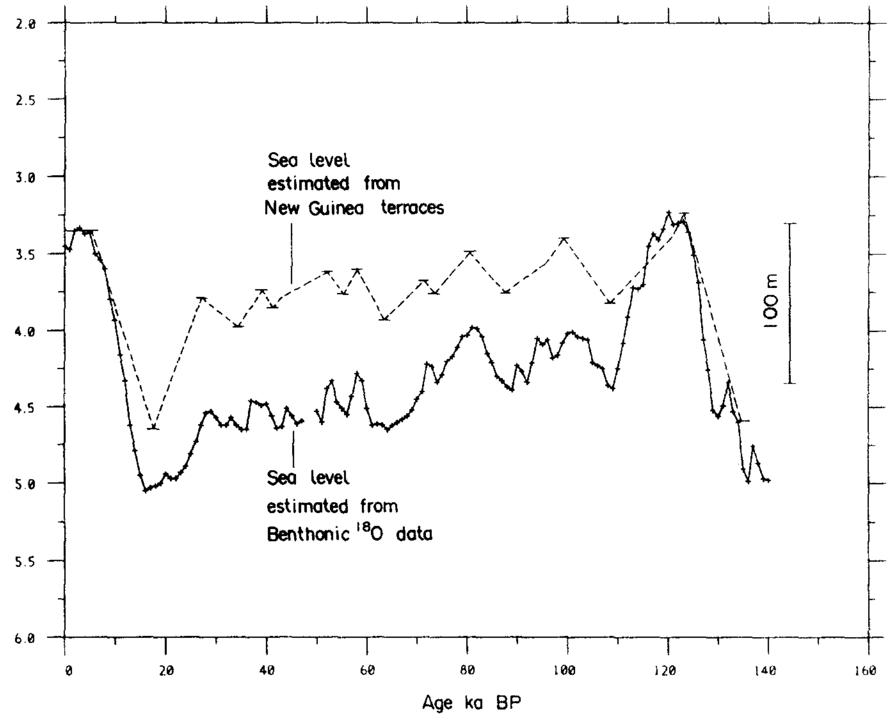

FIG. 1. Benthonic oxygen isotope record of East Pacific core V19-30 for the past 140 ka and sea level record derived from New Guinea (after Chappell and Shackleton, 1986). ——— Sea level estimated from benthonic ^{18}O data. - - - - Sea level estimated from New Guinea data.

oxygen isotope record was outlined by Matthews and Poore (1980). They postulated the existence of regions of the sea surface (especially the subtropical gyres) whose temperature is invariant (citing reasoning in Newell *et al.* (1978)). Their argument was largely directed towards a re-evaluation of the whole Cenozoic oxygen isotope record, but was also applied to the Pleistocene; they suggested that to assume a constant temperature for the deep ocean is unreasonable and that it is easier to justify the assumption of a constant temperature for the sea surface at low latitudes. Crowley and Matthews (1983) applied this approach in analysing a core from the central gyre of the North Atlantic, suggesting that the glacial–interglacial range of 1.4‰ in the core represented the true glacial–interglacial range attributable to ice volume. However, their record shows the same serious discrepancy as the benthic records when the Stage 5 section is examined; the values corresponding to substages 5a and 5c are over 0.7‰ positive relative to the 5e level.

When Shackleton and Opdyke (1973) compared their planktonic oxygen isotope record from the West Pacific with the best sea level data then available, they found quite good agreement. This perhaps suggests that the model proposed by Matthews and Poore (1980) is appropriate, but that the West Pacific is a more stable area than the North Atlantic gyre chosen by Crowley and Matthews (1984). Figure 2 shows a similar comparison but using newer and more detailed data from another core from the West Pacific, RC17-177. Although the gross agreement is better than in Fig. 1, it is clear that the details of sea level rise and fall are not recorded. This disagreement is especially prominent within Stage 5, where the New Guinea data imply oscillations of the order 50 m while the oxygen isotope data scarcely preserve any significant changes between 110 ka and 80 ka BP.

One possible explanation would be that the inferred sea level record from New Guinea is incorrect and that there were no glacio-eustatic fluctuations within Stage 5. However, here the data of Labeyrie *et al.* (1987) from the Norwegian Sea are critical. Today deep-water forms in the Norwegian Sea at a temperature of −1°C, close to the minimum possible (which is about −1.5°C). Labeyrie *et al.* (1987) show oxygen isotope values for benthic foraminifera that lived on the sea floor in the Norwegian Sea within Stage 5, and particularly in substage 5d, that are more than 0.5‰ more positive than specimens living today or than those that were present during substage 5e or 5c. Thus the ocean was certainly more isotopically positive during 5d than during 5e and 5c, consistent with the New Guinea record.

A more likely explanation for the discrepancy indicated by Fig. 2 is that the details within stage 5 in core RC17-177 have been obscured by bioturbation. A re-examination of Fig. 1 shows that between substage 5d and 5a the pattern recorded in core V19-30 is similar to that reconstructed from the Huon Peninsula. In order to use the deep-sea oxygen isotope record as a monitor of sea level, we need the detail preserved in this core. In fact Chappell and Shackleton (1986) deduced that once the deep waters cooled down at the end of substage 5e the temperature probably remained

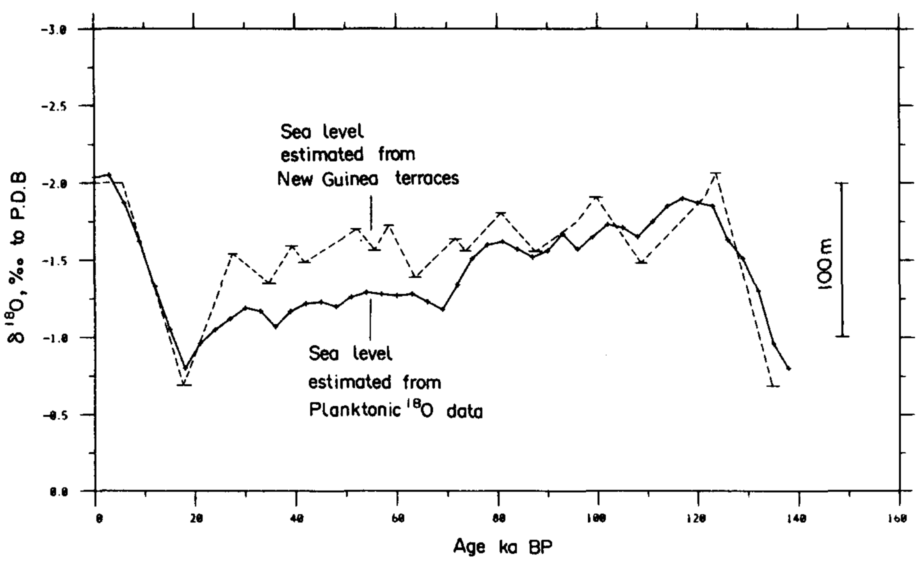

FIG. 2. Planktonic oxygen isotope record of West Pacific core RC17-177 for the past 140 ka and sea level record derived from New Guinea. ——— Sea level estimated from benthonic ^{18}O data. - - - - Sea level estimated from New Guinea data.

more or less constant until about 11 ka BP when they warmed to present temperatures. This postulated temperature record may be explored by examining the temperature difference between West Pacific surface waters (at constant temperature by the hypothesis of Matthews and Poore, 1980) and East Pacific deep water.

Figure 3 shows the result of subtracting the oxygen isotope value interpolated at 1 ka intervals in core RC17-177, from the time-equivalent value in core V19-30. Of course if it were the case that RC17-177 preserved a perfect record of ocean isotopic composition, then the result of this subtraction would be to yield an accurate record of changing temperature in the deep-water sampled by the benthic foraminifera in core V19-30. The effect of utilising the bioturbated and therefore smoothed record from core RC17-177 instead of a perfectly detailed record, is to enhance artificially the temperature variability implied by Fig. 3. Thus a more nearly correct record of deep-water temperature variation might be derived by digitally smoothing the record of Fig. 3. A smoothed version is shown in Fig. 4.

Figures 3 and 4 could have been scaled as temperature records for East Pacific deep water at the site of core V19-30, based on the hypothesis of Matthew and Poore (1980) that West Pacific surface temperature variability at the site of RC17-177 has been negligible. However, our purpose here is to evaluate the sea level record, for which purpose we only need to appreciate that if Fig. 4 is the more accurate representation of the deep-water temperature record, it follows that a more accurate sea level record may be derived by subtracting this temperature component from the original V19-30 data: the result of this exercise is shown in Fig. 5. As expected, this shows better agreement with the sea level record than either of the original data sets did. Thus Fig. 5 may be taken as a second approximation to the global sea level record.

In principal this approach will be extendable through the entire Pleistocene. However, before this is done it is important to avoid any long-term biasses in the resulting record. Our objective so far has been to obtain a continuous record based on oxygen isotope analyses that matches the details of sea level history for the past 130 ka better than any individual data sets have done.

A LONGER PERSPECTIVE: THE ENTIRE BRUNHES

A serious limit on the approach outlined above is that we do not understand exactly what controls the oxygen isotopic composition of an assemblage of planktonic foraminifera. Among the individuals comprising each sample analysed there is a range of season of life, depth in the water column inhabited and degree of post-depositional dissolution. Thus there is not a clearly defined relationship between the isotopic composition of the assemblage and the constant sea surface temperature postulated by the Matthews and Poore (1980) model, and there are reasons to suspect that the relationship may not remain constant. In contrast, many different genera of benthonic foraminifera are well behaved in this respect. If we wish to compare the sea levels of the interglacials, or of the glacials, over the last million years, we should look carefully at a wide selection of the available data both for benthonic and planktonic species. Table 1 lists cores from which this comparison is made.

Table 2 compares oxygen isotope values for interglacial extremes in a selection of well documented oxygen isotope records. Each extreme is then compared with the Stage 1 (Holocene) value in the same core (or the 5e value if stage 1 is unavailable) and the isotopic difference is given underneath the observed isotope value. Several important conclusions may be

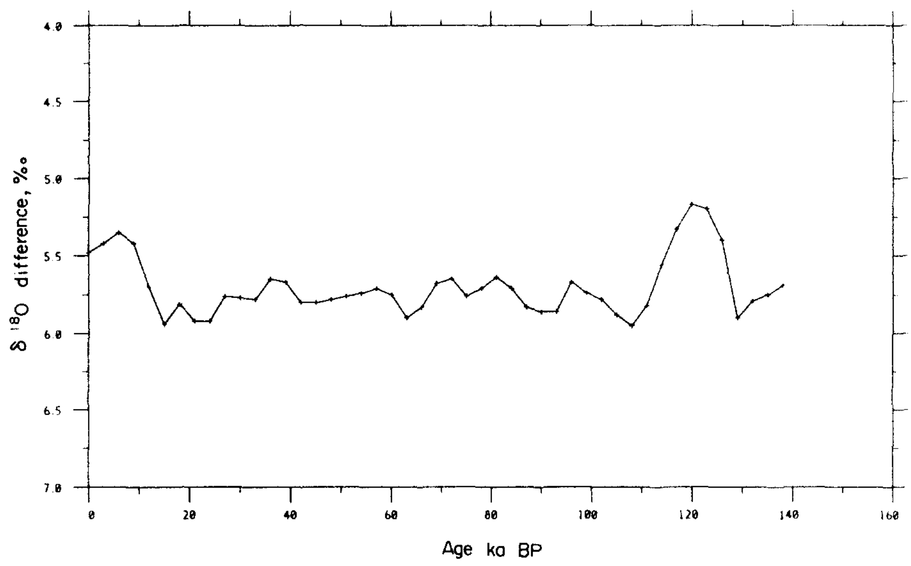

FIG. 3. Isotopic difference between V19-30 and RC17-177 records, here interpreted as resulting from changing deep water temperature at the site of core V19-30.

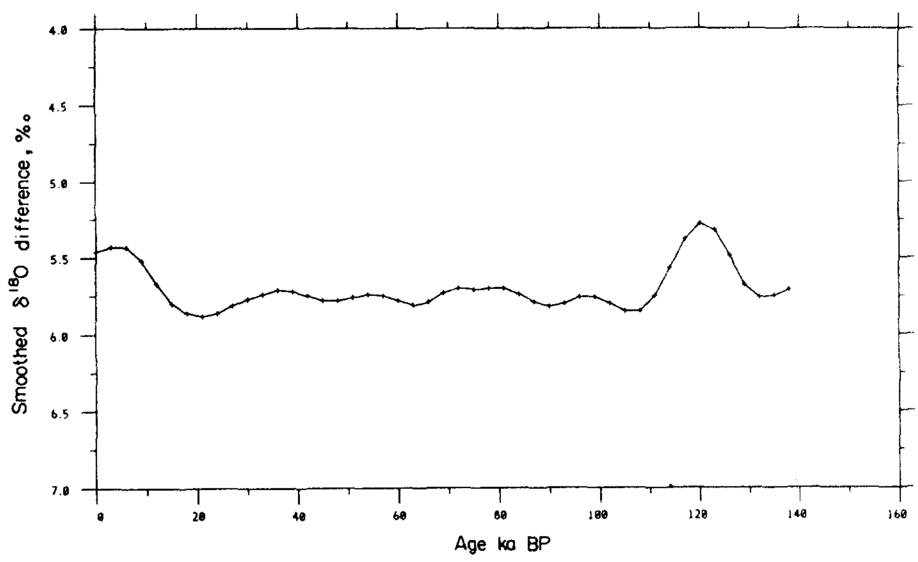

FIG. 4. Isotopic difference between V19-30 and RC17-177, smoothed.

drawn from this table. First, in the benthonic records (Table 2a) no interglacial is much isotopically lighter (i.e. more negative) than stage 1. The extremes of 1, 5e, 9 and 11 are all so similar that one should pay close attention to the raw data before reaching any conclusions; an isolated one-peak measurement in one core should not be given the same weight as a detailed suite in another. At present I do not believe that we can confidently state that any one of these four interglacials reached a significantly different level than any other. However, it may be important for our understanding of the Pleistocene record that these four interglacial peaks are significantly isotopically lighter than Stages 7, 13, 15, 17 or 19. Only with Stage 23 in the Jaramillo Normal event at about 0.9 Ma BP is there another interglacial (Stage 23, well recovered in DSDP552A) which is isotopically similar to the Holocene. During these remaining interglacials either some northern hemisphere ice must have remained, or ocean deep waters must have been colder than they are today.

By analogy with our analysis of the past 130 ka, we should be able to distinguish these alternatives by examining the planktonic data (Table 2b). Although the planktonic values are more scattered, the values for interglacial Stages 7, 13, 15, 17 and 19 are indeed systematically more positive than the extreme values for Stages 1, 5, 9 and 11. This strongly suggests that there was indeed excess ice remaining during these

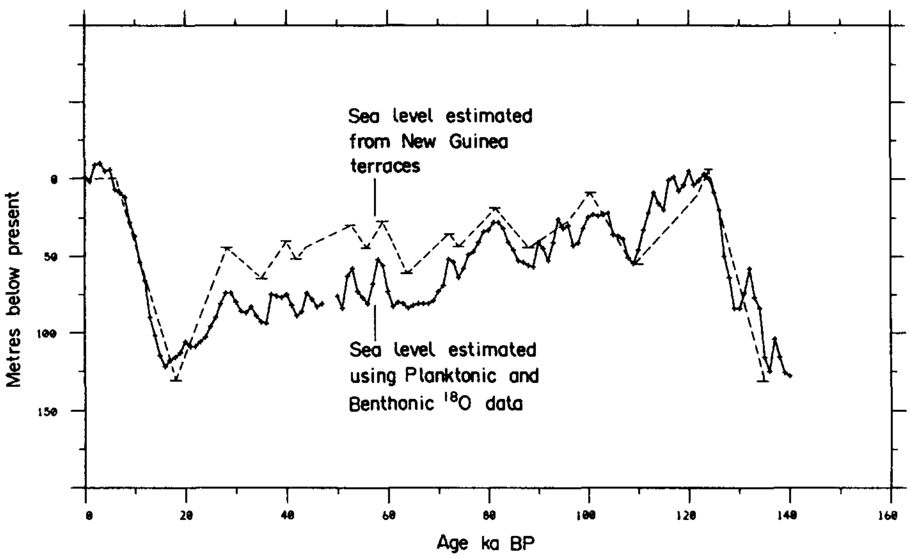

FIG. 5. Sea level history derived by subtracting the record shown in Fig. 4 from that shown in Fig. 1 (see text) and compared with the record derived from New Guinea. ——— Sea level estimated from benthonic ^{18}O data. - - - - Sea level estimated from New Guinea data.

TABLE 1. Location of cores discussed

Core	Latitude	Longitude	Water depth	Source
V19-28	02°22'S	84°39'W	2720	1
V19-29	03°35'S	83°56'W	2673	1
V19-30	03°23'S	83°21'W	3091	2,3
V22-174	10°04'S	12°49'W	2630	4
V28-238	01°01'N	160°29'E	3120	5
RC13-228	22°20'S	11°12'E	3204	6
RC13-229	25°30'S	11°18'E	4191	6
RC17-177	1°45.3'N	159°26.9'E	2600	9
DSDP552A	56°03'N	23°14'W	2311	7
M13519	05°39'N	19°51'W	2862	8
V21-146	34°41'N	163°02'E	3968	9

(1) Ninkovich and Shackleton 1975; (2) Shackleton *et al.*, 1983; (3) Shackleton and Pisias, 1985; (4) Shackleton, 1977; (5) Shackleton and Opdyke, 1973; (6) Morley and Shackleton, 1984, only *Uvigerina* data used; (7) Shackleton and Hall, 1984; (8) Sarnthein *et al.*, 1984; (9) Shackleton, *unpubl.*

interglacials. This in turn suggests that on slowly uplifting coastlines there should be a marked gap between the Stage 11 shoreline and the much older Stage 23 shoreline, while at least the relics of Stages 9 and 5e should be present below the Stage 11 line.

It is worth making a further point regarding Stage 11. This stage is easily identified in the deep sea because it immediately follows two widely used stratigraphic markers (the extinction of *Stylatractus universus*, documented by Hays and Shackleton, 1976, and the extinction of *Pseudoemiliania lacunosa* documented by Thierstein *et al.*, 1977). In the North Pacific the study of Sachs (1973) suggested that this was substantially the warmest interglacial in the last million years. In DSDP Site 552A in the North Atlantic, Stage 11 is represented by the thickest section of nannofossil ooze with the least ice-rafted contribution of any of the interglacials in the last 2.5 Ma (Zimmerman *et al.*, 1984). Moreover it immediately followed Stage 12, an exceptionally large glaciation. Although stage 16 may have been as extensive a glaciation, there was not the same direct transition to an interglacial extreme as was observed at the Stage 12–11 boundary. Although the oxygen isotope data does not indicate where the extra ice was located, it seems inherently likely that the very extensive Elster glaciation occurred during either of the extremes of Stage 16 or Stage 12. However, at sites such as Marks Tey (Turner, 1970), interglacial deposits fill the presumed subglacial channels of this glaciation. This situation seems to fit better the Stage 12–11 transition than the 16–15 transition. Sarnthein and Stremme (1986) have also concluded that the Holsteinian Interglacial should be correlated with Stage 11.

Turning now to the glacial extremes, Table 3 compares each glacial extreme in each core, with the Stage 1 value in the same core. Again both the observed value and below, the isotopic difference is given. Amongst the benthonic data sets (Table 3a) there is a very good measure of agreement. Without doubt Stages 12 and 16 were more extreme than Stage 2, Stage 6 perhaps marginally more extreme. Stage 10 was perhaps marginally less extreme than Stage 2 while Stages 4, 8, 14 and 18 were significantly less important. Amongst the planktonic data sets (Table 3b) no consistent pattern emerges. This is not surprising, since temperature variations must have played a part for many of the cores. The two West Pacific cores V28-238 and RC17-177, perhaps less affected by temperature, are consistent with the benthonic data sets in indicating Stages 12 and 16 as the extreme glacials but the differences between the two records are too great for much weight to be placed on them.

One way to compare the possible ice volume or sea level extremes of each glacial would be to assume that the benthonic data sets are the most reliable, and that

TABLE 2a. Comparison of Interglacial extremes of Stages 5, 7, 9, 11 with Stage 1 in benthonic data from cores shown. Above: observed value, below: difference from Stage 1 value

	Stage									
Core	1	5a	5e	7	9	11	13	15	17	19
V19-28	3.15	3.81	3.40	3.36	3.16	3.16	3.64*			
		0.66	0.25	0.21	0.01	0.01	0.49			
V19-29	3.21	3.90	3.19	3.47						
		0.69	−0.02	0.26						
V19-30	3.27	3.97	3.18	3.61	3.26					
		0.70	−0.09	0.34	−0.01					
V21-146	3.38	3.51	3.21	3.61	3.46	3.00	3.09?			
		0.13	−0.17	0.23	0.08	−0.38	−0.29?			
RC13-228	3.29	3.72		3.47						
		0.43		0.16						
RC13-229	3.11		3.28	3.57		2.95	3.32			
			0.17	0.46		−0.16	0.21			
DSDP552A			3.19	3.59	3.27	3.27	3.90	3.67	3.63	3.71
				0.40	0.08	0.08	0.71	0.48	0.44	0.52
M13519	3.26	4.09	3.60	3.59	3.36	3.33	3.81	3.64	3.70	3.65
		0.73	0.34	0.33	0.10	0.07	0.55	0.38	0.44	0.39

*Determined in core V19-25.

TABLE 2b. Comparison of Interglacial extremes of Stages 5, 7, 9, 11 with Stage 1 in planktonic data from cores shown. Above: observed value, below: difference from Stage 1 value

	Stage									
Core	1	5a	5e	7	9	11	13	15	17	19
V22-174	−0.92		−0.95	−0.90	−0.98	−0.82	−0.72	−0.80	−0.72	−0.74
			−0.03	0.02	−0.06	0.10	0.20	0.12	0.20	0.18
V28-238	−1.97	−1.82	−2.10	−1.78	−1.98	−1.96	−1.80	−1.86	−1.35	−1.68
		0.15	−0.13	0.19	−0.01	0.01	0.17	0.11	0.62	0.25
RC17-177	−2.11		−1.94	−1.72	−1.62	−1.86	−1.46	−1.69	−1.50	−1.50
			0.17	0.39	0.49	0.25	0.65	0.42	0.61	0.61
M13519	−1.52	−0.92	−1.77	−1.56	−1.71	−1.36	−1.37	−1.62	−1.37	−1.24
		0.60	−0.25	−0.04	−0.19	0.16	0.15	−0.10	0.15	0.28
RC11-120	1.87	2.37	1.79	2.13						
		0.50	−0.08	0.36						
E49-18			1.98	2.25	2.01	2.14				
				0.27	0.03	0.16				

TABLE 3a. Comparison of glacial extremes in benthonic data of Stages 2, 4, 6, 8, 10, 12, 14, 16 with Stage 1 in cores shown. Above: observed value, below: difference from Stage 1 value

	Stage								
Core	2	4	6	8	10	12	14	16	18
V19-28	4.92	4.69	4.94	4.74	4.85	5.04			
	1.77	1.34	1.79	1.59	1.70	1.89			
V19-29	4.93	4.74	4.90	4.74					
	1.72	1.53	1.69	1.53					
V19-30	5.03	4.67	5.17	4.85					
	1.76	1.40	1.90	1.58					
V21-146	5.11	4.62	5.20	4.69	5.21	5.23	4.41		
	1.73	1.24	1.82	1.31	1.83	1.85	1.03		
Rc13-229	4.96		4.86	4.74	5.06	5.28		5.13	4.65
	1.85		1.75	1.63	1.95	2.17		2.02	1.54
DSDP552A	4.73	4.56	5.18	4.55	4.63	4.96	4.77	5.05	4.71
	1.54	1.37	1.99	1.36	1.44	1.77	1.58	1.86	1.52
M13519	5.25	4.52	5.10	4.80	4.87	5.29	4.73	5.33	4.62
	1.99	1.26	1.84	1.54	1.61	2.03	1.47	2.07	1.37
Mean	1.76	1.36	1.82	1.51	1.71	1.94	1.36	1.98	1.48
Mean[1]	1.26	0.86	1.32	1.01	1.21	1.44	0.86	1.48	0.98
Scaled[2]	100	68	105	80	96	114	68	117	78

[1] After subtracting 0.5 temperature contribution.
[2] Taking Stage 2 ice volume = 100.

TABLE 3b. Comparison of glacial extremes in planktonic data of Stages 2, 4, 6, 8, 10, 12, 14, 16 with Stage 1 in cores shown. Above: observed value; below: difference from Stage 1 value

	Stage								
Core	2	4	6	8	10	12	14	16	18
V22-174	0.81		0.58	0.31	0.29	0.26	0.20	0.66	0.43
	1.73		1.50	1.23	1.21	1.18	1.12	1.58	1.35
V28-238	−0.96	−1.12	−0.73	−0.68	−0.70	−0.64	−1.04	−0.58	−0.71
	1.01	0.85	1.24	1.29	1.27	1.33	0.93	1.39	1.26
RC17-177	−0.76		−0.74	−0.85	−0.50	−0.31	−0.79	−0.43	−0.50
	1.35		1.37	1.26	1.61	1.80	1.32	1.68	1.61
M13519	0.45	0.07	0.06	−0.03	0.26	0.06	−0.17	−0.22	−0.19
	1.97	1.59	1.58	1.49	1.78	1.58	1.35	1.30	1.33
RC11-120	3.49		3.54						
	1.62		1.67						
E49-18			3.77	3.39	3.44	3.57			
			1.79	1.41	1.46	1.59			

the degree of cooling in the deep-water was identical in each glacial. Such a conclusion would be consistent with our earlier discussion, following Chappell and Shackleton (1986), which indicated that deep waters remained at the same temperature between 115 ka and 11 ka BP. Table 3a shows mean values for the extreme deviation of each glacial stage. Below is the mean after subtracting a fixed 0.5‰ contribution for the temperature effect. Finally these values are shown scaled relative to the Stage 2 value. This indicates that ice volume during Stages 12 and 16 was about 15% greater than at the last glacial maximum.

ACKNOWLEDGEMENTS

This work was supported by NERC grant GR3/3606. I thank Mike Hall for his operation of the mass spectrometer, and Julie Cartlidge for laboratory assistance. I am especially grateful to Ann Wintle for carefully reviewing a draft of this manuscript and significantly improving its clarity.

REFERENCES

Bloom, A.L., Broecker, W.S., Chappell, J.M.A., Matthews, R.K. and Mesolella, K.J. (1974). Quaternary sea level fluctuations on a tectonic coast: new $^{230}Th/^{234}U$ dates from the Huon Peninsula, New Guinea. *Quaternary Research*, **4**, 185–205.

Chappell, J. and Shackleton, N.J. (1986). Oxygen isotopes and sea level. *Nature*, **324**, 137–140.

Crowley, T.J. and Matthews, R.K. (1983). Isotope-plankton comparisons in a Late Quaternary core with a stable temperature history. *Geology*, **11**, 275–278.

Denton, G.H. and Hughes, T.J. (1981). *The Last Great Ice Sheets*. John Wiley, New York.

Dodge, R.E., Fairbanks, R.G., Benninger, L.K. and Maurrasse, F. (1983). Pleistocene sea levels from raised coral reefs of Haiti. *Science*, **219**, 1423–1425.

Emiliani, C. (1955). Pleistocene temperatures. *Journal of Geology*, **63**, 538–578.

Hays, J.D. and Shackleton, N.J. (1976). Globally synchronous extinction of the radiolarian *Stylatractus Universus*. *Geology*, **4**, 649–652.

Labeyrie, L., Duplessy, J.C. and Blanc, P.L. (1987). Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years. *Nature*, **327**, 477–482.

Matthews, R.K. and Poore, R.Z. (1980). Tertiary ^{18}O record and glacio-eustatic sea-level fluctuations. *Geology*, **8**, 501–504.

Mesolella, K.J., Matthews, R.K., Broecker, W.S. and Thurber, D.L. (1969). The astronomical theory of climatic change: Barbados data. *Journal of Geology*, **77**, 250–274.

Mix, A.C. and Ruddiman, W.R. (1984). Oxygen-isotope analyses and Pleistocene ice volumes. *Quaternary Research*, **21**, 1–20.

Morley, J.J. and Shackleton, N.J. (1984). *In*: Berger, A., Imbrie, J., Hays, J., Kukla, G. and Saltzman, B. (eds), *Milankovitch and Climate*, pp. 467–480. D. Reidel, Hingham, Mass.

Newell, R.E., Navato, A.R. and Hsuing, J. (1978). Long-term global sea surface temperature fluctuations and their possible influence on atmospheric CO_2 concentrations. *Journal of Pure and Applied Geophysics*, **116**, 351–371.

Ninkovitch, D. and Shackleton, N.J. (1975). Distribution, stratigraphic position and age of ash layer "L", in the Panama Basin region. *Earth and Planetary Science Letters*, **27**, 20–34.

Sarnthein, M., Erlenkeuser, H., von Grafenstein, R. and Schroeder, C. (1984). Stable-isotope stratigraphy for the last 750,000 years: "Meteor" core 13519 from the eastern equatorial Atlantic. *Meteor Forsch.-Ergebnisse*, **C38**, 9–24.

Sarnthein, M. and Stremme, H.E. (1986). The Holstein interglaciation: time-stratigraphic position and correlation to stable-isotope stratigraphy in deep-sea sediments. *Quaternary Research*, **26**, 283–298.

Sachs, H.M. (1973). Late Pleistocene history of the North Pacific: evidence from a quantitative study of radiolaria in core V21-173. *Quaternary Research*, **3**, 89–98.

Shackleton, N.J. (1977). The oxygen isotope stratigraphic record of the late Pleistocene. *Philosophical Transactions of the Royal Society*, **280**, 169–179.

Shackleton, N.J. and Hall, M.A. (1984). Oxygen and carbon isotope stratigraphy of Deep Sea Drilling Project Hole 552A: Plio-Pleistocene glacial history. D.G. Roberts, D. Schnitker *et al. Initial Reports of the Deep Sea Drilling Project*, **81**, 599–609. U.S. Govt. Printing Office, Washington.

Shackleton, N.J., Imbrie, J. and Hall, M.A. (1983). Oxygen and carbon isotope record of East Pacific core V19-30: implications for the formation of deep water in the late Pleistocene North Atlantic. *Earth and Planetary Science Letters*, **65**, 233–244.

Shackleton, N.J. and Opdyke, N.D. (1973). Oxygen isotope and palaeomagnetic stratigraphy of equatorial Pacific core V28-238: oxygen isotope temperatures and ice volumes on a 10^5 year and a 10^6 year scale. *Quaternary Research*, **3**, 39–55.

Shackleton, N.J. and Pisias, N.G. (1985). Atmospheric carbon dioxide, orbital forcing and climate. *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present*, Geophysical Monograph, **32**, 303–317.

Thierstein, H.R., Geitzenauer, K.R., Molfino, B. and Shackleton, N.J. (1977). Global synchroneity of Late Quaternary coccolith datums: validation by oxygen isotopes. *Geology*, **5**, 400–404.

Turner, C. (1970). The Middle Pleistocene deposits at Marks Tey, Essex. *Philosophical Transactions of the Royal Society of London*, **B257**, 373–440.

Zimmerman, H.B., Shackleton, N.J., Backman, J., Baldauf, J.G., Kaltenbach, A.J. and Morton, A.C. (1984). History of Plio-Pleistocene climate in the northeastern Atlantic, Deep Sea Drilling Project Hole 552A. *Initial Reports of the Deep Sea Drilling Project*, **81**, 861–875.