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The Oxygen Isotope Stratigraphic Record of the Late Pleistocene [and Discussion]

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The oxygen isotope stratigraphic record of the Late Pleistocene†

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Oxygen isotope measurements have been made in foraminifera from over 60 deep-sea sediment cores. Taken together with the oxygen isotope measurements published by Emiliani from Caribbean and Equatorial Atlantic cores, this comprises a unique body of stratigraphic data covering most of the important areas of calcareous sediment over the whole world ocean.

The oxygen isotopic composition of foraminifera from cores of Late Pleistocene sediment varies in a similar manner in nearly all areas; the variations reflect changes in the oxygen isotopic composition of the ocean. The oceans are mixed in about 1 ka so that ocean isotopic changes, resulting from fluctuations in the quantity of ice stored on the continents, must have occurred almost synchronously in all regions. Thus the oxygen isotope record provides an excellent means of stratigraphic correlation.

Cores accumulated at rates of over about 5 cm/ka provide records of oxygen isotopic composition change that are almost unaffected by post-depositional mixing of the sediment. Thus they preserve a detailed record of the advance and retreat of the ice masses in the northern hemisphere, and provide a unique source of information for the study of ice-sheet dynamics.

Introduction

Over the past few years the writer has made oxygen isotope measurements in over 60 deep-sea sediment cores widely distributed over the oceans. Figure 1 shows the distribution of those cores analysed with the intention of studying the last glacial stage (see also table 1). The oxygen isotope data are given in tables 2–65.‡

Although it is not possible to discuss all the data in a single paper, the results constitute a unique synthesis of the stratigraphy of the Late Quaternary of the ocean floor, and as such deserve to be made available.

There are three aspects of the data that will be discussed here. First, a major objective of oxygen isotope studies is to obtain an accurate record of the changing oxygen isotopic composition of the oceans as a function of time. For this purpose it is evident that some cores are more suitable than others. Second, the recognition of this record (albeit in a modified form) in each core permits that core to be correlated with another or with a synthetic 'ideal' sequence. Third, differences between individual records provide valuable insight into sea-floor processes.

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‡ Tables 2–65 appear in the accompanying microfiche.

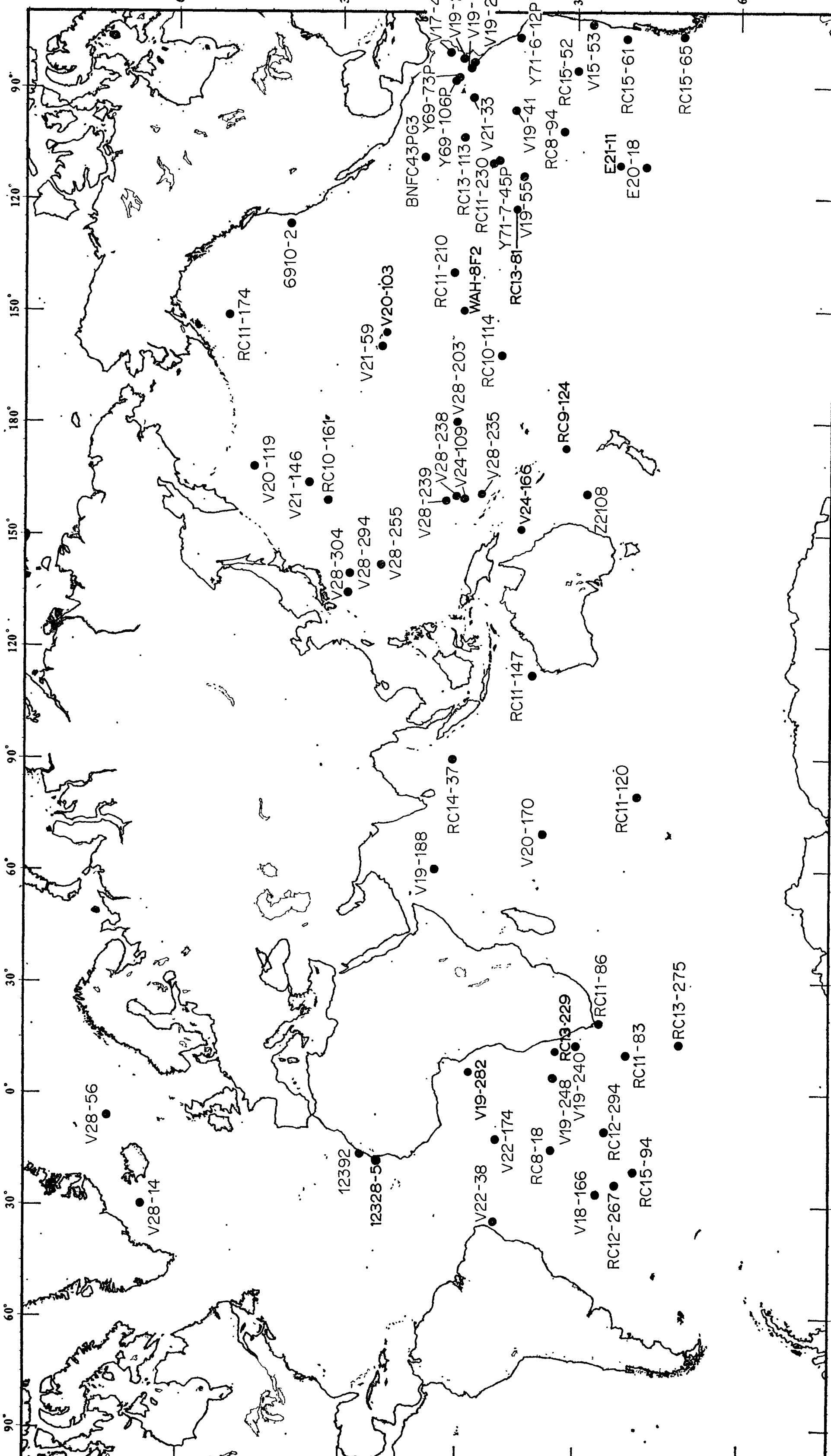

FIGURE 1. Distribution of the cores from which oxygen isotope data are presented (see also table 1).

Table 1. Location of cores

core	latitude	longitude	depth	table
North Atlantic				
V28-14	64° 47′ N	29° 34′ W	1855	2
V28-56	68° 02′ N	06° 07′ W	2914	3
12328-5	21° 08.7′ N	18° 34.4′ W	2812	4
South Atlantic				
RC8-18	24° 04′ S	15° 07′ W	3977	5
RC11-86	35° 47′ S	18° 27′ E	2829	6
RC12-267	38° 41′ S	25° 47′ W	4144	7
RC12-294	37° 15.6′ S	10° 05.8′ W	3308	8
RC13-229	25° 29.6′ S	11° 18.6′ E	4191	9
V19-240	30° 35′ S	13° 17′ E	3103	10
V19-248	24° 34′ S	04° 50′ E	3321	11
V19-282	02° 45′ S	04° 35′ E	4356	12
V22-38	09° 33′ S	34° 15′ W	3797	13
V22-174	10° 04′ S	12° 49′ W	2630	14
Antarctic				
RC11-83	41° 36.0′ S	09° 43.0′ E	4718	15
RC11-120	43° 31.0′ S	79° 52.0′ E	3913	16
RC13-275	50° 43.0′ S	13° 26.0′ E	1984	17
RC15-94	42° 53.9′ S	20° 51.3′ W	3762	18
Indian				
V19-188	06° 52′ N	60° 40′ E	3356	19
V20-170	21° 48′ S	69° 14′ E	2479	20
RC11-147	19° 03.5′ S	112° 45′ E	1953	21
RC14-37	01° 28′ N	90° 10′ E	2226	22
Pacific				
6910-2	41° 16′ N	126° 24′ W	2743	23
BNFC 43 PG3	10° 29.5′ N	109° 01.7′ W	2874	24
E20-18	44° 33′ S	111° 20′ W	2867	25
E21-11	39° 58′ S	112° 09′ W	2798	26
RC8-94	27° 17′ S	102° 05′ W	3074	27
RC9-124	28° 44.7′ S	172° 35.6′ E	2540	28
RC10-114	11° 11′ S	162° 55′ W	2791	29
RC10-161	33° 05′ N	158° 00′ E	3587	30
RC11-174	52° 34.6′ N	151° 21′ W	1618	31
RC11-210	01° 49′ N	140° 03′ W	4420	32
RC11-230	08° 48′ S	110° 48′ W	3259	33
RC13-81	19° 01.1′ S	124° 13.5′ W	3751	34
RC13-113	01° 39′ S	103° 38′ W	2195	35
RC15-52	29° 14.3′ S	85° 59.1′ W	3952	36
RC15-61	45° 36.6′ S	77° 12.1′ W	3771	37
RC15-65	53° 04′ S	78° 57.1′ W	3200	38
V15-53	32° 27′ S	73° 40′ W	3915	39
V17-42	03° 32′ N	81° 11′ W	1814	40
V19-27	00° 28′ S	82° 04′ W	1373	41
V19-28	10° 22′ S	84° 39′ W	2720	42
V19-29	03° 35′ S	83° 13′ W	3157	43
V19-41	14° 06′ S	96° 12′ W	3248	44
V19-55	17° 00′ S	114° 11′ W	3177	45
V20-103	33° 59′ N	177° 50′ W	3442	46
V20-119	47° 57′ N	168° 47′ E	2736	47
V21-33	03° 48′ S	92° 05′ W	3726	48
V21-59	20° 55′ N	158° 06′ W	2992	49
V21-146	37° 41′ N	163° 02′ E	3968	50
V24-109	00° 26′ N	158° 48′ E	2367	51
V24-166	16° 31′ S	150° 47′ E	781	52
V28-203	00° 57′ N	179° 25′ W	3243	53

TABLE 1 (*cont.*)

V28-235	05° 27′ S	160° 29′ E	1746	54
V28-238	01° 01′ N	160° 29′ E	3120	55
V28-239	03° 15′ N	159° 11′ E	3490	56
V28-255	20° 06′ N	142° 27′ E	3261	57
V28-294	28° 26′ N	139° 58′ E	2308	58
V28-304	28° 32′ N	134° 08′ E	2942	59
WAH-8F2	00° 01.5′ N	147° 59′ W	4308	60
Y69-73P	01° 27′ N	87° 56′ W	2707	61
Y69-106P	02° 59′ N	86° 33′ W	2870	62
Y71-6-12	16° 26.6′ S	77° 33.8′ W	2734	63
Y71-7-45P	11° 04.7′ S	110° 62′ W	3096	64
Z-2108	33° 23′ S	161° 37′ W	1448	65

Notes: Cores are curated in the following institutions from which routine lithological descriptions of the cores are available.

prefix	V, RC	Lamont–Doherty Geological Observatory, Palisades, New York, U.S.A.
(no prefix)	12392, 12328-5	University of Kiel, West Germany
(no prefix) / prefix	6910-2 / Y	Oregon State University, School of Oceanography, Corvallis, Oregon, U.S.A.
prefix	BNFC, WAH	Scripps Institute of Oceanography, La Jolla, California, U.S.A.
prefix	E	Florida State University, Talahassee, Florida, U.S.A.
prefix	Z	New Zealand Oceanographic Institute, Wellington, New Zealand

PREVIOUS WORK

The most important body of Late Pleistocene oxygen isotope data other than that contained here has been obtained by Cesare Emiliani (1955*a*, *b*, 1958, 1964, 1966, 1972). The cores that he analysed cover the Caribbean (several cores), the Equatorial Atlantic, and one core each in the Eastern Mediterranean and the North Atlantic. Additional work has been done by others in the Mediterranean, the Gulf of Mexico and the Red Sea; these areas present special problems because of their relative isolation from the ocean water masses, and will not be discussed further here. Duplessy, Chenouard & Reyss (1974) analysed an Equatorial Atlantic core and Duplessy, Chenouard & Vila (1975) analysed one from the Norwegian Sea, which is another area which presents some unusual problems. In the Indian Ocean, Oba (1969) made oxygen isotope measurements in two cores from south of Ceylon, and more recently M. Sommer (Brown University) and J.-C. Duplessy (Gif-sur-Yvette, France) have analysed several cores as a part of the CLIMAP project.

In addition to the measurements reported here the writer has recently published detailed data for core 12392, illustrated in figure 2 (Shackleton 1977). A few other cores have been analysed but are not included here because work on them is not complete.

A careful programme of inter-laboratory standardization operates between the Cambridge laboratory and that of Dr J.-C. Duplessy at Gif-sur-Yvette, and has been extended to that of Dr M. Sommer at Brown University. This calibration in terms of the P.D.B. reference standard is probably concordant with that used by Emiliani while working at Chicago (Shackleton 1974). All measurements reported here are in terms of that standard using the δ notation:

$$\delta = 10^3\left(\frac{R_{\text{sample}}}{R_{\text{standard}}} - 1\right),$$

where R is ^{18}O–^{16}O ratio (see also Craig 1957).

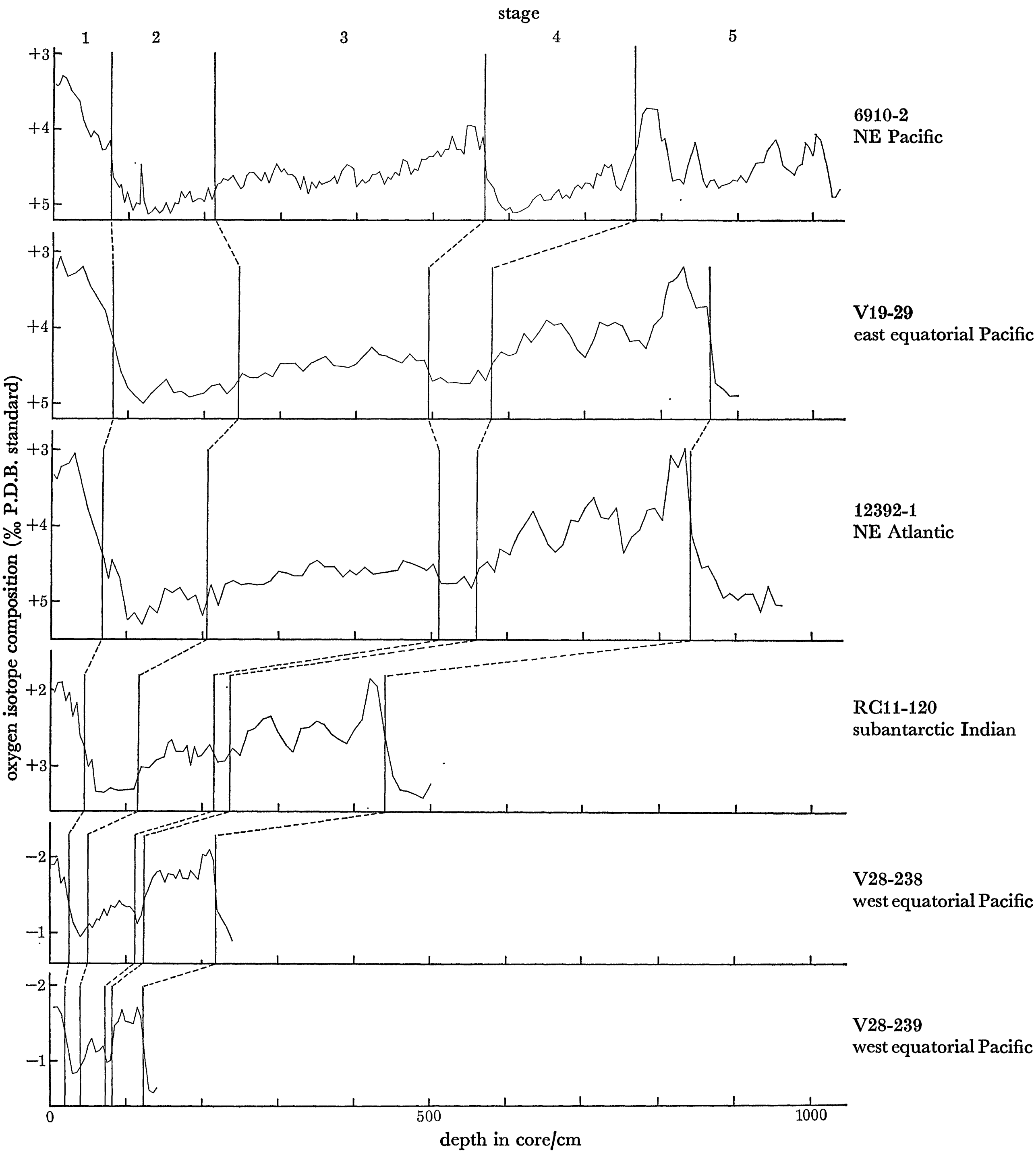

FIGURE 2. Oxygen isotope records in six cores from different areas characterizing sediment accumulation rates from about 10 cm/ka (top) to about 1 cm/ka (bottom). Core 12392-1 was taken at 23° 08.4′ N 17° 44.7′ W in 2066 m water depth; the record illustrated is based on analyses of several species of benthonic foraminifera published by Shackleton (1977). The locations of the remainder are given in table 1, and the data in tables 23, 43, 16, 55 and 56. Stages after Emiliani (1955) and Shackleton & Opdyke (1973).

Experimental methods

The data in tables 2–65 were obtained in the Cambridge laboratory between 1972 and, 1976. Until late 1973 measurements were made with a modified MS3 mass spectrometer, described by Shackleton (1965). In 1973 a VG Micromass 602C mass spectrometer was obtained with a grant from the Natural Environment Research Council. The methods used for sample handling and release of carbon dioxide for mass spectrometric analysis were described by Shackleton & Opdyke (1973) and Shackleton (1974).

Analyses are referred to the P.D.B. standard (Epstein, Buchsbaum, Lowenstam & Urey 1953) by analysing standard carbonates in the same way as the foraminiferal samples. About 10 % of all analyses made are of standard carbonates. If analyses of a homogeneous carbonate taken in aliquots of *ca.* 0.3 mg, are performed over long periods (years), the standard deviation of the measurements is about ± 0.07 per mille (‰). On the other hand, the variance among replicate analyses of planktonic foraminifera from a sediment sample is greater. A typical sample may be about 12 specimens, which may have lived at different depths in the water column, at different seasons of the year, and in different centuries (or even different millennia in a slowly accumulating sediment). To illustrate this variance, multiple analyses of *Globigerinoides sacculifer* are shown for core V28-238 (for which the mean values were given by Shackleton & Opdyke 1973).

At those times when ocean isotopic composition remained stable for millennia, replicate analyses of benthonic foraminifera are not subject to the same sources of variation as the planktonic foraminifera. Core RC12-229 shows this clearly. At times of rapid ocean isotopic change, such as the end of the last glacial (sample at about 50 cm), variance is very high, but over the interval from about 24 to 14 ka B.P., when ice sheets were fairly stable near their maximum extent, replicate analyses agree rather closely (samples from 70 to 90 cm).

Glacio-eustatic sea-level change and the ocean oxygen isotope record

The data in tables 2–65, and in particular those from cores in which benthonic foraminifera were analysed and which had a high rate of deposition (e.g. V19-27, V19-28, V19-29, 6910-2, 12392-1, RC12-267), show that ocean deep waters must have been about 1.65‰ enriched in ^{18}O at the maximum of the last glacial as compared with today. This could be an over-estimate of the ocean isotopic change, to the extent that a glacial drop in deep-water temperatures could have contributed to the measured isotopic change in the foraminifera. It seems unlikely, however, that any significant drop actually occurred. A reduction in ocean deep-water temperatures would require a substantial increase in the flux of Antarctic Bottom Water (AABW). There does not appear to be any direct evidence that this occurred, and oceanographic arguments (Weyl 1968) predict a diminution rather than an increase in the AABW flux during glacial periods.

An average oxygen isotopic composition of -30‰ for the excess ice stored during glacials has been estimated by Dansgaard & Tauber (1969) and by Shackleton (1967). Taking this figure, we can calculate that the ocean would have been enriched in ^{18}O by about 0.1‰ for every 10 m drop in sea level (neglecting isostatic effects). Hence, an ^{18}O enrichment of 1.65‰ would represent an ice excess equivalent to a sea level lowering of about 165 m. Although this figure is somewhat greater than most current estimates (for example, those of Donn, Farrand & Ewing 1962), recent work by members of the CLIMAP project working at the University of Maine suggests that the larger estimate is probably correct (Denton 1976, personal communication). If so, this greatly strengthens the case for regarding the oxygen isotope record as an ideal monitor of glacio-eustatic sea-level change, subject only to the reservations discussed below concerning ocean mixing times and bioturbation.

Discussion: the 'ideal' record of ocean isotopic change

In figure 2 six oxygen isotope records are illustrated with the same depth and isotopic difference scales. These clearly illustrate the extent to which the details of the record become lost as a result of bioturbation in slowly accumulating sediments. A comparison of the records in core V19-29 (Ninkovich & Shackleton 1975) from the south side of the Panama Basin, and Meteor core 12392-1 (Shackleton 1977) from off the coast of Morocco, shows a high degree of similarity in two areas of rather rapid sedimentation (average about 7 cm/ka). The similarity is enhanced by the fact that in both areas the high sedimentation rate is the result of wind-driven coastal upwelling; more intense glacial winds have given rise to a higher accumulation rate in glacial than in interglacial times in both areas.

Core 6910-2 (Oregon State University; discussed by Heusser, Shackleton, Moore & Balsam 1975) shows an even higher accumulation rate and perhaps less variability in rate. The isotopic peak at about 550 cm is believed to represent the base of stage 4, equivalent to about 470 cm in core V19-29. If this is correct, its age would be about 60 ka; it probably represents the deglacial event marked by the stand of sea level dated in Barbados (James, Mountjoy & Omura 1971), Japan (Konishi, Omura & Nakamichi 1974) and New Guinea (Bloom *et al.* 1974), at which time the sea is estimated to have reached within 28 m of its present level (Bloom *et al.* 1974). If the isotopic event at 550 cm in core 6910-2 is correctly identified, its duration may be estimated as about 1 ka on the basis of a uniform sedimentation rate. Since the interglacial accumulation rate for core V19-29 is about 5 cm/ka, an event as short as this would probably not be recorded in that sediment.

Bloom *et al.* (1974) have inferred sea level stands at −38 m at a date near 40 ka ago, and at −41 m about 28 ka ago. These events do not appear to be recorded in core 6910-2 and may have been of even shorter duration than the 60 ka event. If so, this is unfortunate, in that the mixing time of the ocean may be as much as 1 ka (Gordon 1975), so that events which lasted only a few centuries will have given rise to transient isotopic differences between water masses and will not have been recorded uniformly even in the most detailed sedimentary records.

Kennett & Shackleton (1975) have already documented large transient isotopic changes in the sediments of the Gulf of Mexico, associated with the flow of glacial meltwater into that sea at a fast rate compared with the rate of ocean mixing. Thus the interpretation of a record of ocean isotopic composition with a resolution below about 1 ka will require a careful consideration of ocean mixing, particularly since the ocean mixing rate will itself have been affected by the transient changes in North Atlantic surface salinity associated with glacial advance and retreat.

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Discussion: stratigraphic correlation

As was discussed by Shackleton & Opdyke (1973, 1976), the oxygen isotope record in deep-sea sediments provides a tool for stratigraphic correlation that is generally limited only by the stratigraphic resolution of the sediments themselves. The reconstruction of the distribution of ocean surface temperatures at 18 ka B.P. undertaken by the CLIMAP project (CLIMAP 1976) used the oxygen isotope record as a primary means of long-distance and inter-oceanic correlation. Within each geographically limited area many other stratigraphic tools were available. The cores discussed in this paper form the basis for this stratigraphic framework.

In areas with a high rate of sediment accumulation, several samples, taken from the interval representing *ca.* 24–14 ka B.P., typically yield oxygen isotope values that are statistically indistinguishable (for example, from 100 to 200 cm in core V19-29, figure 2), because over this interval the major ice sheets apparently remained relatively stable. In this case the mean sedimentation rate must be used to judge the sample most likely to represent 18 ka B.P. In core V19-29 the 150 cm sample probably represents a real, though brief, interstadial event, possibly the Erie Interstade (Mörner & Dreimanis 1973).

In areas of lower accumulation rate the most isotopically positive value may or may not be statistically different from its neighbours; in either case it is the sample likely to contain the highest proportion of glacial-maximum material. This statement recognizes the fact that if it were possible to separate from that sample only the specimens that lived (for example) 18 ka ago, they would be found to be isotopically heavier than the intermingled specimens that may be several thousand years older or younger. Thus the selection of the 29 cm level in core V28-239 (figure 2) as representing 18 ka attempts a best choice in an area in which any sediment sample must represent the imperfectly homogenized record of events covering several thousand years.

A few of the cores analysed (e.g. V28-294, V15-94, RC11-174) yield isotopic values for the top of the core, which indicate it is of glacial age rather than Holocene. Others (e.g. V20-103, V24-166) yielded an isotopic record that was not recognized in terms of the pattern that emerges from the majority of cores. This may be a function of the inefficiency of a burrowing animal as a homogenizer of sediment (i.e. an isolated, visible burrow in the sediment may contain material of quite different age and character from its surroundings), or it may be an indication that an unconformity is present. In the latter case other biostratigraphic tools are available for assessing the approximate age of the sediment. Thus oxygen isotope stratigraphy is a valuable means of checking that a core does in fact record events of the Late Pleistocene.

Discussion: sedimentation processes

There is considerable interest in the study of the process of bioturbation as it affects the record in deep-sea sediment (Berger & Heath 1968; Guinasso & Schink 1975). In the past the effect has been studied in detail mainly by examining the vertical distribution in the sediment of material deposited over a geologically brief interval (for example, a volcanic ash horizon). Another approach is to study the decay profile of radiocarbon or other radioisotopes (Peng, Broecker, Kipphut & Shackleton 1977). A simple approach, using the oxygen isotope data, is to examine the extent to which the isotopic change between the most recent sediment and that representing the last glacial maximum has been reduced by mixing. As discussed above, the ocean water masses probably underwent an isotopic change of about 1.65‰, yet the range observed in sediments accumulated at about 1 cm/ka is only a little over half the figure (e.g. core V28-239 in figure 2). This approach was used by Peng *et al.* (1977).

A second approach to the sediment mixing problem using oxygen isotope data is to examine the range of isotopic composition among individual benthonic foraminiferal tests in a single sediment sample, which (as discussed above) is a function of the amount of time averaged in that piece of sediment.

Since the oxygen isotope records may be expressed as a set of continuous records representing isotopic composition as a function of time, a third approach is to compare these records either at selected points (Shackleton & Opdyke 1976, table 3) or statistically, using all the data. The oxygen isotope records of core RC11-120 (illustrated in figure 2) and of another core nearby, both show spectral energy at a frequency corresponding to a period of about 20 ka (Hays, Imbrie & Shackleton 1976), while that of core V28-238 (also illustrated in figure 2) shows insignificant energy at this frequency (unpublished study by J. Imbrie). This suggests that in core V28-238 (accumulation rate 1.7 cm/ka) the signal with frequency 1 cycle per 20 ka (i.e. wavelength in the sediment 34 cm) has been attenuated by bioturbation to within the stratigraphic noise level. This is illustrated by the fact that the five substages in stage 5 cannot be seen in this latter core (figure 2).

Conclusions

In view of the fact that sediment accumulation rates below 2 cm/ka characterize a great proportion of the ocean area, the oxygen isotope record in most sediment cores is not an accurate representation of the changes in oxygen isotopic composition in the ocean water masses. For the same reason, the temporal resolution that may be obtained by the use of the oxygen isotope record as a tool for stratigraphic correlation is limited, and is probably inversely proportional to the accumulation rate. However, several regions are already known in which accumulation rates are sufficiently high that it is apparently possible to obtain oxygen isotope records whose time resolution is less than the mixing time for ocean water masses. In these areas there is the potential for the extraction of a great deal more information than has been obtained to date. Close attention to analytical accuracy, to optimal sampling frequency and to inter-core comparisons both over small and inter-oceanic distances, will provide unique insights on two important processes. The first is the dynamic behaviour of large and perhaps unstable ice sheets. The second is the mixing of the oceans under the flux boundary conditions associated with the removal of water by evaporation from, and the addition of meltwater to, the ocean surface layer. The importance of calibration accuracy must be particularly stressed. From the analytical point of view it may become possible to measure horizontal and vertical isotopic gradients in the deep waters through much of Late Pleistocene time with an accuracy of ±0.03‰ or less, equivalent to a palaeotemperature accuracy of about 0.1 °C; it is not at present possible to approach this accuracy in inter-laboratory comparison.

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I am deeply indebted to a large number of my friends and colleagues in CLIMAP for the free exchange of exciting ideas which this project has engendered, and in particular I would like to thank T. C. Moore and J. D. Hays for giving me such a geographically wide perspective on palaeo-oceanographic events in the Late Pleistocene, through their responsibility for the Pacific and Antarctic interests in CLIMAP. This work would not have been possible, and certainly would not have been undertaken in Britain, but for the support and encouragement of Sir Harry Godwin, F.R.S., and later of Richard West, F.R.S., and the Sub-department of Quaternary Research in the University of Cambridge.

I am especially grateful for the opportunity to work closely with my CLIMAP colleagues, and to have immediate access to the enormous library of deep-sea sediment cores at the Lamont-Doherty Geological Observatory, which was provided by the receipt of the Senior Visiting Research Fellowship at that institution in 1974–5.

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Discussion

H. H. Lamb (*Climatic Research Unit, University of East Anglia, Norwich NR4 7TJ*). Can one yet compare the land and ocean oxygen isotope records and say that the times when the ice sheets were forming or growing most rapidly were marked by much less depression of temperature than later developed when the ice cover was most extensive?

N. J. Shackleton. At present, we know very little about sea surface temperatures at the time of inception, or of most rapid growth, of the Northern hemisphere ice sheets.

B. C. Worssam (*Institute of Geological Sciences, 5 Princes Gate, London SW7 1QN*). The succession of warm and cold peaks shown by isotope stage 5 seemed to have a bearing on the question of the definition of the base of the Devensian. Are substages 5b and 5d likely to have been of such intensity and duration as to have allowed ice sheets to invade the British Isles?

J. M. Hodgson (*Soil Survey of England and Wales, Rothamsted Experimental Station, Harpenden, Herts*). The marked climatic fluctuations in stage 5 shown by Dr Shackleton prompt me to ask if there is any evidence from his work or from that of Professor McIntyre suggesting that there could be sea ice in the English Channel at this period.

N. J. Shackleton. So long as the base of the Devensian Stage remains defined by a till, I believe that it is unlikely that it should be correlated with any part of stage 5. However, climate during substages 5b and 5d was certainly of glacial character; at these times the size of the

Laurentide ice sheet was probably sufficient to have lowered sea level to a point where the English Channel was no longer marine.

A. R. Lord (*Department of Geology, University College London, Gower Street, WC1E 6BT*). The factors affecting the precision of ^{14}C dating have been discussed earlier today by Professor Shotton for material of generally terrestrial origin. Would Dr Shackleton assess the relative reliability of ^{14}C dates based on material from deep-sea cores, for example in relation to diagenetic change in the foraminiferid tests used, as this is clearly crucial when comparing continental events with those in the ocean basins.

F. W. Shotton, F.R.S. (*Dept. of Geological Sciences, University of Birmingham, Birmingham B15 2TT*)) In core V19-29, which has rather more rapid sedimentation rates than most, Dr Shackleton shows three well-marked peaks in stage 5 and suggests that the earliest and warmest, 5e, might mark the Ipswichian. If that is so, the subsequent record accords excellently with the principal interstadials of the Devensian, putting 5c as Amersfoort, 5a as Chelford and the steep change from stage 4 to 3 being the very rapid amelioration which ushered in the Upton Warren Interstadial. An age of about 125000 years had been suggested for the centre of the Ipswichian (Eemian). If this were so, one would have to think of dates of the order of 105–110000 for 5c, 80000 for 5a and 65000 for the 4/3 junction, and these figures disagreed seriously with the radiocarbon dates for the land-based interstadials. While the 'enriched' ^{14}C dates for Amersfoort (68000) and Brørup/Chelford (64000/60000) might still have to be regarded as only minima, there was no reason to question the validity of the several Upton Warren Interstadial dates around 43000.

What methods have been used to put dates on the ocean cores and do they have a degree of uncertainty sufficient to reconcile them with ^{14}C dates as discrepant as those cited above?

N. J. Shackleton. The accuracy of ^{14}C age determinations in marine sediments from the end of the last glacial is limited by bioturbation and related processes, rather than by diagenesis, while for older samples laboratory contamination is a more serious hazard than geological diagenesis. Fortunately, the chronology available is based on several other techniques as well as on ^{14}C. It is extremely unlikely that these techniques could all be so far in error that the true age of the stage 4–3 boundary could be as young as 43000 B.P.

D. J. Schove (*St David's College, Beckenham, Kent*)

Temperature turning points

Temperature curves show characteristic turning points which need to be fixed independently in ^{14}C, lake-core, ice-core and deep-sea chronology.

The writer would be interested to know how far the climatic turning points can be identified in ice-cores and deep-sea cores. It is clear from Dr Shackleton's paper that the ^{14}C chronology is proving too short when compared with K-Ar dating, and it would be helpful to know how the magnitude of the discrepancy varies. If the cold period hitherto dated 53000 B.P. is really 70000 B.P. should we stretch the ^{14}C scale in the Devensian not merely by the 3 % needed for the right half-life but by some 30 % throughout?

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Chronology of the Bölling (Zone Ib)

Minor temperature fluctuations, like the major changes demonstrated by Professor McIntyre and Dr Shackleton, are in the same sense on both sides of the Atlantic. This has been true in the last 200 years.

Varve counts on both sides of the Atlantic have thus been teleconnected using cross-correlation of non-random features. Curves of glacial recession, derived separately for the U.S.A. and for the Baltic and based on the isochrones of bottom varves, show parallel trends when superimposed at the positions proposed (see D. J. Schove (1971), Varve teleconnection across the Baltic, *Geografiska Annaler, Stockholm* **53**, 214–234).

In the Bölling (Pollen Zone Ib in northern Europe) for instance the abnormal warmth implied by Dr Coope's beetles is matched not only by a phase of rapid recession in southern Sweden but also by a synchronous recession in the U.S.A. In Vermont the rate of 335 m/year was described by Antevs (cf. his varves numbered 6950/7125) as 'the fastest rate observed in New England (cf. D. J. Schove (1969), A varve teleconnection project, *VIIIe Congres INQUA, Paris*, p. 932). In Europe the Alleröd recession of Zone II was slower if more prolonged; in the U.S.A. there is again a synchronous slower but more prolonged recession corresponding to varves 7280/8050.

The warmth of Bölling summers in the British Isles has been stressed at this conference. We should, nevertheless, distinguish between the Bölling in the new broad sense of 13000/12000 B.P. and in the narrow sense of the hot phase which lasted less than two centuries (12400/12200 B.P.) and was, surprisingly, punctuated in S. Sweden by a brief glacial readvance! In pollen diagrams the Bölling has often been included wrongly with the Alleröd in an extended but too broad sense of Zone II.

From the analysis of varves in SE Sweden an important result was announced by Tarling & Noel (in progress) at the Meeting of European Geological Societies in September 1975 that 'the Laschamp geomagnetic event can be dated as lasting 200 years with a reversal lasting for 30 years', that is between 12077 and 12103 varve-dates B.P. On the Nilsson scale this makes the reversal 10127/10153 N which I date 12400 B.P. ± 100 (cf. 1971, *Geografiska Annaler*, **53**, 230, Table 5). This geomagnetic event would have hindered modulation of the cosmic ray flux and this would

(*a*) have increased the ^{14}C content in the atmosphere, so making some radiocarbon dates that ought to relate to the mid-Bölling appear to be much too young;
(*b*) be partly responsible for the abnormal warmth of the Late Bölling.

L. H. N. COOPER (2 *Queens Gate Villas, Lipson, Plymouth*). There are two distinct oceanic boundaries of interest during the advance and retreat of Quaternary ice. One is the southern limit of polar water becoming precisely defined by the work of the CLIMAP group and reaching as far south as 43° N on the Iberian shores. This is confirmed by evidence on British Admiralty charts. An oceanic area of seasonal ice separates this polar front from the second boundary, i.e. the edge of the permanent ice. North of this permanent ice edge the parameters which control climate cease to be maritime, equable and moist, but tend towards continentality with wider diurnal and seasonal ranges of temperature, lower absolute humidity and precipitation largely limited to dry snow which may readily drift.

The evidence available for defining this limit of permanent ice is much like that drawn from facial features which enable us to identify people but is hard to put down in words. The extreme southern limit seems to be set along the western and southern edges of the Celtic Sea, retaining permanent ice on the bed of the English Channel and Celtic Sea down to 48° N lat. but continuing as a belt of shelf ice down the Biscayan coast of France. Kellaway and his associates (1975) seem to be alone in sharing this view.

The descriptive terms applied to present-day air masses break down during glacial maxima unless arbitrarily the permanent oceanic ice is treated without reservation as 'continental'. Even so redefinition is needed. Air masses of continental polar and continental arctic origin approached from directions which are now considered as sources of maritime polar air.

Though today maritime tropical and maritime polar air masses and the occluded weather systems to which they may give rise contribute most of our precipitation, this is mostly as rain. Such snowfall as they bring, though sometimes heavy, is usually 'wet' and 'warm'. Severe weather in the British Isles arises from continental polar and continental arctic air streams approaching from between southeast and northeast. The snow deposited as small crystals is 'dry' and 'cold' and readily drifts. The moisture to be precipitated here as snow has largely been acquired by very cold and dry Siberian air masses during passage over the present-day open waters of the North Sea. Air masses approaching over an ice filled North Sea would have arrived much colder and with less moisture and have been able to deposit less snow. Climate would have been sunny and arid.

When conditions at the end of an interglacial deteriorated, oceanic permanent ice would have crossed the Greenland–Scotland Ridge into the North Atlantic, properly so called. On my view at its maximum the permanent ice edge reached to 48° N lat., though there may have been an embayment in the ice edge west of the Celtic Sea and over the deeper water of the Porcupine Forebight. At this ice edge westerly winds would have tended to produce upwelling not of cooler water as is commonplace today but of warmer more saline water. Under such conditions a (pseudo-) continental air mass from the ice-bound northwestern Atlantic would have blown over the Porcupine Forebight to have its moisture content increased as over the North Sea today. A decision as to whether and when this embayment in the permanent oceanic ice occurred is of importance for understanding Quaternary climates in the British Isles.

Thus as the permanent oceanic ice moved south there would have been suppression, probably complete, of maritime tropical air masses and then replacement of maritime polar by continental polar and continental arctic air masses. The greatest volume of snowfall was likely when maritime polar air from the southwest was frequently occluded over continental polar air from west or northwest. But as the ice front advanced southwards prevalence of much continental arctic air should have led to long periods of very cold sunny weather with infrequent falls of 'cold' and 'dry' drifting snow. This leads to difficulties in defining the terms 'periglacial' and 'glacial'.

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