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Oxygen-Isotope and Paleomagnetic Stratigraphy of Pacific Core V28-239 Late Pliocene to Latest Pleistocene

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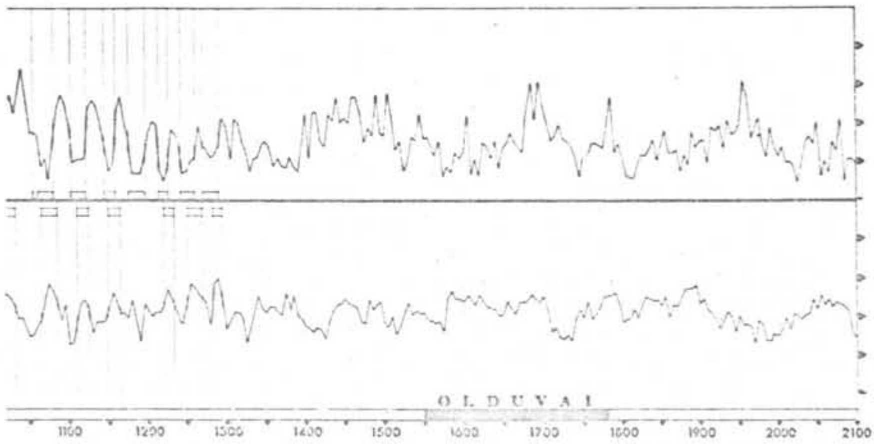
ABSTRACT

V28-239 core from cruise 28 of R/V *Vema* preserves a detailed oxygen-isotope and paleomagnetic record for all of the Pleistocene Epoch. The entire 21-m-long core has been analyzed at 5-cm intervals. Glacial stage 22, above the Jaramillo magnetic event, may represent the first major Northern Hemisphere continental glaciation of middle Pleistocene character. Prior to this, higher frequency glacial events extend to near the level of the Olduvai magnetic event. Glacial events of less regular frequency extend to the bottom of the core, which represents late Pliocene time. Fluctuations in carbonate dissolution intensity occur throughout the core with a similar frequency to the oxygen-isotope fluctuations.



Hays and others (1969); stages in the oxygen-isotope record are numbered after Emiliani (1955, 1966) and Shackleton and Opdyke (1973).

Sediment samples were disaggregated in distilled water; foraminifers were selected for analysis from the >180-μm fraction after sieving and ultrasonic cleaning. Sample pretreatment and chemical processing were identical to those used for core V28-238 (Shackleton and Opdyke, 1973). Isotope analysis was performed in a new V.G. Micromass 602C mass spectrometer. Analyses are referred to the PDB standard (Epstein and others, 1951, 1953) using a value of +0.29‰ for the Emiliani B-1 standard (Shackleton, 1974). This calibration is accurate to better than ±0.05‰. Analytical results in Shackleton and Opdyke (1973) were referred to the B-1 standard and must be corrected by +0.29‰ before comparison with the data presented in this paper.

A single analysis has been made at each level in the core. For each analysis, 15 specimens of *Globigerinoides sacculifer* were selected (in the lower part of the core *G. fistulosus* was used in some samples, three samples contained insufficient specimens for analysis, and a few contained less than 15). Analytical precision is estimated to be ± 0.05‰, the standard deviation for 100 analyses of a standard carbonate performed during the first six months of instrument operation. However, the uncertainty in analysis of a single sample from the sediment is ±0.11‰ (Shackleton and Opdyke, 1973). Isotopic variability among the specimens and analytical precision are combined in this figure. Analytical results are given in Table 1. Figure 1 shows the percentage by weight retained on the 180-μm sieve for each sample, the oxygen-isotope record, and the paleomagnetic record.

Character of the Isotopic Record

Jaramillo Magnetic Event to Present. Figure 1 suggests that the oxygen-isotope record may be divided into three episodes of differing character. The upper part, all of which is represented in core V28-238 (Fig. 2) as well as core V28-239, contains glacial stages at approximately 100,000-yr intervals. Apparently, the isotopic composition of the ocean changed by almost the same extent in every glaciation during this interval. The rather large variability among glacial extreme isotopic values in core V28-239 is an artifact of sedimentation processes. This is evident from the fact that the extreme isotopic values in successive glaciations are both less variable and more distant from the Holocene value in cores with higher accumulation rates. In core V28-239 (1.0 cm/10^3 yr) the extreme isotopic values in glacial stages 2 and 6 to 22 differ from the Holocene value by 1.22 ± 0.24‰. In core V28-238 (1.7 cm/10^3 yr) the same ten glacial extreme values differ from the Holocene value by 1.04 ± 0.14‰. In core V19-28 (4.0 cm/10^3 yr) the last five glacial extreme values differ from the Holocene value by 1.62 ± 0.11‰. (Ninkovich and Shackleton, 1975).

Figure 1 shows cyclic changes in the percentage of sediment that is greater than 180 μm as well as in oxygen-isotope composition. Thompson and Saito (1974) documented correlative cyclic variations in dissolution intensity in cores V28-238, V28-239, and RC11-210. The latter core is in the region where Hays and others (1969) defined dissolution zones on the basis of changing carbonate percentage. Thus, we may confidently ascribe the observed variations in coarse-fraction percentage to changes in dissolution intensity and assign them to zones according to the definition of Hays and others (1969). Figure 1 shows these dissolution zones. The delay between the climatic change recorded in the oxygen-isotope record and the dissolution change, noted by Luz and Shackleton (1975) and by Ninkovich and Shackleton (1975), is preserved throughout the sequence.

Characteristically, the transition from glacial to interglacial extreme occurred very rapidly (Broecker and van Donk, 1970); indeed, 12,000 yr ago deglaciation took place so fast that its record in sediment cores is almost invariably determined by the sediment-mixing depth rather than by the actual rate of change in the isotopic composition of the ocean (at least 0.3‰/10^3 yr). Glacial stages 2, 6, 10, 12, 20, and 22 terminated in this manner. Stages 4, 8, and 14 probably did

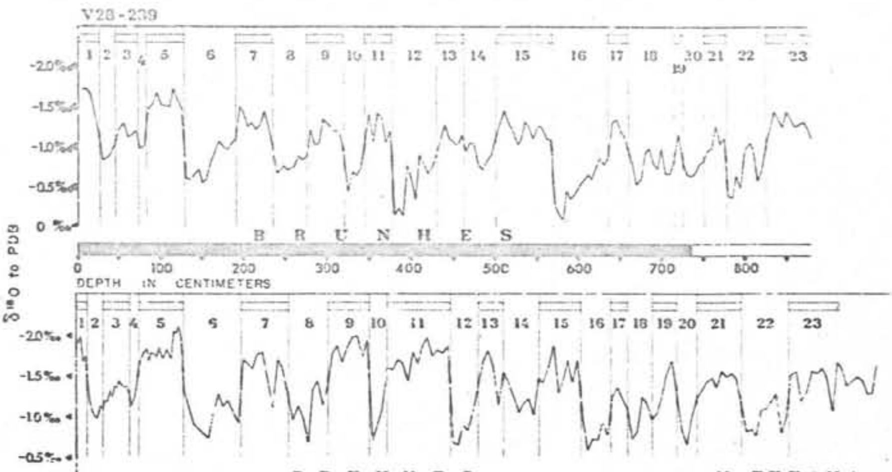


Figure 2. Oxygen-isotope and paleomagnetic record in upper 880 cm of cores V28-239 (above) and V28-238 (below).

V16-205 seems to contain even greater changes in accumulation rate than V28-239. The rate between the Jaramillo and the top of the Brunhes in core V16-205 is reported to be 0.25 cm/10^3 yr, although the average rate through the entire core is 0.55 cm/10^3 yr. It is to be hoped that interoceanic correlations for the Matuyama epoch will become more reliable with the analysis of more cores in both the oceans.

ISOTOPE STRATIGRAPHY AND ITS LIMITATIONS

Oxygen-Isotope Stages: Terminology

Emiliani (1955, 1966) used numbers 1 to 16 to designate stages that he recognized in oxygen-isotope records he obtained in sediment cores from the Caribbean Sea and Atlantic Ocean. We (Shackleton and Opdyke, 1973) recognized 22 stages in core V28-238, the first 16 coinciding with those used by Emiliani. As a step toward formalizing this nomenclature, stage boundaries were defined by the depth at which they were located in core V28-238 (Fig. 2). For core V28-239, 23 stages are shown in Figure 1; the depths of stage boundaries, placed by correlation with core V28-238, are given in Table 2.

Before considering the extension of this terminology, it is important to consider the assumptions on which use of the oxygen-isotope record as a stratigraphic tool is based and the limitations of its usefulness. It is universally agreed that at the

TABLE 2. STAGE BOUNDARIES IN CORE V28-239 AS DETERMINED BY CORRELATION WITH CORE V28-238

Boundary	Depth in core* (cm)	Age† (B.P.)	Termination§
1-2	25	13,000	I
2-3	45	32,000	
3-4	72	64,000	
4-5	82	75,000	
5-6	127	128,000	II
6-7	190	195,000	
7-8	235	251,000	III
8-9	275	297,000	
9-10	320	347,000	IV
10-11	345	367,000	
11-12	377	440,000	V
12-13	430	472,000	
13-14	462	502,000	
14-15	500	542,000	
15-16	567	592,000	VI
16-17	635	627,000	
17-18	660	647,000	
18-19	715	688,000	
19-20	725		
20-21	750		
21-22	777		
22-23	825		

*Determined by correlation with core V28-238.

†Ages are those estimated by Shackleton and Opdyke (1973) by linear interpolation in core V28-238 using a rate of 1.7 cm/10^3 yr.

§Terminations from Broecker and van Donk (1969). They defined terminations on the basis of their interpretation of the saw-toothed character of the oxygen-isotope record. Owing to a possible hiatus in core V12-122, it appears that the event labeled termination VI by them is the stage 16-15 boundary.

depth (Berger and Heath, 1968; Ruddiman and Glover, 1972), and the extremes analyzed must therefore approximate the extremes present in the core. The mean amplitude of isotopic fluctuations is less in core V28-239 than in V28-238, probably because the accumulation rate is lower in comparison with the mixing depth.

We have argued that the observed peak-to-peak amplitude of oxygen-isotope changes in core V28-238 was reduced by mixing (Shackleton and Opdyke, 1973). Thus, the full range of isotopic variation is attenuated in the sediment in core V28-238 and even more so in core V28-239. However, both cores preserve sufficient record that successive stages can be unambiguously recognized.

Carbonate Dissolution and the Oxygen-Isotope Record

Core V28-239 was taken at a depth of 3,490 m, compared to 3,120 m for core V28-238. This accounts for more intense dissolution occurring in core V28-239. Savin and Douglas (1973) pointed out that increasing dissolution not only progressively removes the more solution-susceptible species (often those that lived in shallower water), but it also selectively removes from the population of a single species those members that lived closer to the surface. Thus, the fossil population that has suffered more dissolution registers a lower isotopic temperature as a consequence of that dissolution.

TABLE 3. STAGE-BY-STAGE ISOTOPE EXTREMES FOR CORES V28-239 AND V28-238

Stage	Interval	V28-239 A (δ, ‰)	V28-239 B (range)	V28-238 C (δ, ‰)	V28-238 D (range)	Between-core Difference (A − C)*
1		−1.72		−1.93		0.26
	1-2		0.88		1.01	
2		−0.84		−0.97		0.13
	2-3		0.46		0.47	
3		−1.30		−1.44		0.14
	3-4		0.32		0.59	
4		−0.98		−0.85		−0.13
	4-5		0.74		1.26	
5		−1.72		−2.11		0.39
	5-6		1.16		1.37	
6		−0.56		−0.74		0.18
	6-7		0.94		1.05	
7		−1.50		−1.79		0.29
	7-8		0.82		1.10	
8		−0.68		−0.69		0.01
	8-9		0.67		1.30	
9		−1.35		−1.99		0.64
	9-10		0.90		1.28	
10		−0.45		−0.71		0.26
	10-11		0.97		1.26	
11		−1.42		−1.97		0.55
	11-12		1.28		1.32	
12		−0.14		−0.65		0.51
	12-13		1.14		1.16	
13		−1.28		−1.81		0.53
	13-14		0.56		0.76	
14		−0.72		−1.05		0.33
	14-15		0.73		0.82	
15		−1.45		−1.87		0.42
	15-16		1.35		1.28	
16		−0.10		−0.59		0.49
	16-17		1.25		0.77	
17		−1.35		−1.36		0.01
	17-18		0.82		0.64	
18		−0.53		−0.72		0.19
	18-19		0.63		0.97	
19		−1.16		−1.69		0.53
	19-20		0.52		1.03	
20		−0.64		−0.66		0.02
	20-21		0.63		0.87	
21		−1.27		−1.53		0.26
	21-22		0.90		0.76	
22		−0.37		−0.77		0.40

Note: Column A, extreme oxygen-isotopic composition in each stage in core V28-239, from Table 1. Column B, isotopic difference between adjacent stages in core V28-239. Mean 0.84 ± 0.28. Column C, extreme oxygen-isotopic composition in each stage in core V28-238, from Shackleton and Opdyke (1973, Table 1), corrected to PDB standard. Column D, isotopic difference between adjacent stages in core V28-238. Mean 1.00 ± 0.27

*Difference between the extreme reached in cores V28-239 and V28-238 for each stage. Mean 0.29 ± 0.21.

TABLE 4. THICKNESS OF CLIMATIC CYCLES IN A SUITE OF CARIBBEAN CORES AND IN PACIFIC CORES V28-238 AND V28-239

Stages	Cores						
	P6304-4	P6304-7	P6304-8	P6304-9	Mean P6304 suite	V28-238	V28-239
1	25	30	30	30	29	22	25
2-3	130	145	150	165	148	88	47
4-5	175	185	170	145	169	110	53
6-7	240	275	280	240	259	210	110
8-9	160	195	210	200	191	165	85
10-11	145	180	190	180	174	160	57
12-13	155	150	. .	130	144	105	85
14-15	85	80?	. .	200	200*	155	105
16-17						95	93

Note: Data for Caribbean cores from Emiliani (1955, 1972); for Pacific cores from Shackleton and Opdyke (1973); for core V28-239 from this paper.

*Data for stages 14-15 from P6304 suite is inconsistent. Value for 6304-9 has been adopted, because it is more consistent with the Pacific cores (Figs. 3, 4).

performed on the basis of an assumed uniform accumulation rate may be in error even if the extrapolation is based on average accumulation rates from numerous cores.

Effect of Varying Surface Temperature

In the Caribbean, Emiliani (1966) has shown that the oxygen-isotope composition of *G. sacculifer* in recent sediment implies deposition at or near surface temperature. Vincent and Shackleton (1975) have shown that this is also true in the Indian Ocean. While this situation holds, changes in surface temperature during Pleistocene time, if present, should affect the isotopic composition of *G. sacculifer* populations in Pleistocene sediment. Hence, it is generally assumed that changes in surface temperature may be estimated by subtracting the component that is ascribed to glacially induced changes in ocean isotopic composition from the total record of oxygen-isotopic change (Imbrie and others, 1973).

In the Pacific a different situation prevails: The oxygen-isotope composition of *G. sacculifer* corresponds to a temperature several degrees below sea-surface temperature in many core-top samples (Savin and Douglas, 1973). To what extent this figure is an indication of a difference in the depth distribution of calcification in the species and to what extent it is a function of selective dissolution of the individuals from shallower depths in the water column (Savin and Douglas, 1973) remain to be evaluated. However, we (Shackleton and Opdyke, 1973) argued that the close similarity between the isotopic records of *G. sacculifer* and benthic species in core V28-238 implies that both records depict the history of the isotopic composition of the ocean, and that changes in surface temperature, temperature-depth structure, depth distribution of *G. sacculifer*, and selective dissolution all play minor roles. Discrepancies between the planktonic and benthic records in core V28-238 were ascribed by us (Shackleton and Opdyke, 1973) to the effects of postdepositional sediment mixing by burrowing organisms rather than to the factors mentioned above.

Long-Term Trends in the Oxygen-Isotope Record

There is no general agreement regarding any long-term trends in climate during Pleistocene time; trends that emerge from studies based on the present-day ecological

the climatic record. The intensity of dissolution rose not at the boundary between stages 6 and 5 (termination II of Broecker and van Donk, 1970) but a few thousand years later. We now show that this relationship has held through the past 1.5 m.y.

Figure 1 indicates the boundaries of the dissolution zones, numbered according to the scheme of Hays and others (1969), and their relation to the oxygen-isotope stages. The dissolution zones are not manifested as carbonate fluctuations, and the carbonate content is high throughout the core (Thompson, 1976). However, Thompson and Saito (1974) have shown that change in dissolution intensity is the dominant factor in determining the downcore changes in foraminiferal faunal composition in this area.

Figure 1 clearly indicates that Hays and others (1969) were correct in their assertion that changes in carbonate content in eastern equatorial Pacific sediments could be correlated with the Northern Hemisphere climatic record and with oxygen-isotope records from the Caribbean. However, use of carbonate cycles as a precise stratigraphic tool may be misleading. Figure 1 shows that the base of each dissolution zone in the sediment is not found at the same position as the glacial-to-interglacial isotopic transition, but rather some 5 to 20 cm above. This represents a delay of a few thousand years between the climatic change and its effect on bottom-water chemistry in the equatorial Pacific. This delay may not be constant from one latitude to another or from one climatic cycle to another.

CHRONOLOGY

The record of changes in the oxygen-isotope composition of the world oceans may be readily used as a stratigraphic tool in Pleistocene deep-sea sediments of all oceans (the Atlantic and Caribbean, Emiliani, 1955; the Arctic, van Donk and Mathieu, 1969; the Indian Ocean, Oba, 1969; the Pacific Ocean, Shackleton and Opdyke, 1973; the sub-Antarctic regions, Hays and others, this volume). Moreover, since the primary mechanism giving rise to these changes is the growth and retreat of continental ice sheets in the Northern Hemisphere, the record is of considerable

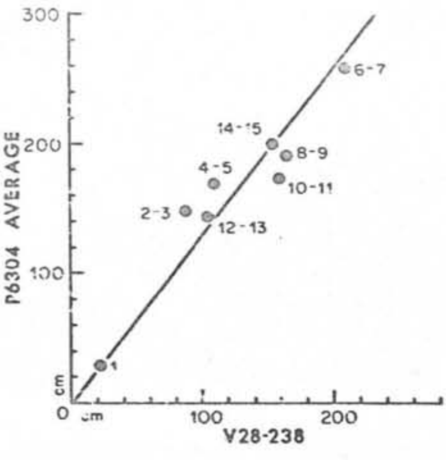


Figure 5. Comparison of thickness of climatic cycles (odd- and succeeding even-numbered stages) in core V28-233 and in suite of cores P6304-4, P6304-7, P6304-8, and P6304-9 from Emiliani (1966, 1972). Data from Table 4.

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