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Unlocking the Ice House: Oligocene-Miocene Oxygen Isotopes, Eustasy, and Margin Erosion

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Oxygen isotope records and glaciomarine sediments indicate at least an intermittent presence of large continental ice sheets on Antarctica since the earliest Oligocene (circa 35 Ma). The growth and decay of ice sheets during the Oligocene to modern "ice house world" caused glacioeustatic sea level changes. The early Eocene was an ice-free "greenhouse world," but it is not clear if ice sheets existed during the middle to late Eocene "doubt house world." Benthic foraminiferal $\delta^{18}O$ records place limits on the history of glaciation, suggesting the presence of ice sheets at least intermittently since the earliest Oligocene. The best indicator of ice growth is a coeval increase in global benthic and western equatorial planktonic $\delta^{18}O$ records. Although planktonic isotope records from the western equatorial regions are limited, subtropical planktonic foraminifera may also record such ice volume changes. It is difficult to apply these established principles to the Cenozoic $\delta^{18}O$ record because of the lack of adequate data and problems in stratigraphic correlations that obscure isotope events. We improved Oligocene to Miocene correlations of $\delta^{18}O$ records and erected eight oxygen isotope zones (Oi1-Oi2, Mi1-Mi6). Benthic foraminiferal $\delta^{18}O$ increases which are associated with the bases of Zones Oi1 (circa 35.8 Ma), Oi2 (circa 32.5 Ma), and Mi1 (circa 23.5 Ma) can be linked with $\delta^{18}O$ increases in subtropical planktonic foraminifera and with intervals of glacial sedimentation on or near Antarctica. Our new correlations of middle Miocene benthic and western equatorial planktonic $\delta^{18}O$ records show remarkable agreement in timing and amplitude. We interpret benthic-planktonic covariance to reflect substantial ice volume increases near the bases of Zones Mi2 (circa 16.1 Ma), Mi3 (circa 13.6 Ma), and possibly Mi5 (circa 11.3 Ma). Possible glacioeustatic lowerings are associated with the $\delta^{18}O$ increases which culminated with the bases of Zone Mi4 (circa 12.6 Ma) and Mi6 (circa 9.6 Ma), although low-latitude planktonic $\delta^{18}O$ records are required to test this. These inferred glacioeustatic lowerings can be linked to seismic and rock disconformities. For example, we link 12 Oligocene-early late Miocene inferred glacioeustatic lowerings with 12 of the sequence boundaries (= inferred eustatic lowerings) of Haq et al. (1987).

Background

We have recently summarized Cenozoic oxygen isotope data and discussed relationships between $\delta^{18}O$, sea level, and erosion on passive continental margins [*Miller et al.*, 1987]. Our benthic foraminiferal $\delta^{18}O$ synthesis suggests that large continental ice sheets existed on Antarctica at least intermittently beginning in the Oligocene (circa 35 Ma). The Oligocene to modern "ice house world" [*Fischer*, 1984] saw the waxing and waning of these ice sheets numerous times, resulting in glacioeustatic sea level changes of up to 100 m [*Miller et al.*, 1987]. Relationships between glacioeustatic changes and deposition on continental margins can be established for the ice house world. Middle to late Eocene $\delta^{18}O$ records are equivocal as to the presence of ice and therefore causes of eustatic change inferred from seismic and sequence stratigraphy [e.g., *Vail et al.*, 1977; *Haq et al.*, 1987] are not clear in this "doubt house world."

The challenge of estimating the ice volume record from the $\delta^{18}O$ signal is a controversial task. Three obstacles hinder unlocking the ice volume signal from the Cenozoic foraminiferal $\delta^{18}O$ record.

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First, the $\delta^{18}O$ value recorded by foraminifera is a function of the temperature and the sea water $\delta^{18}O$ value (δ_w) where the organism lived. For example, a foraminiferal $\delta^{18}O$ increase may reflect either a drop in water temperature, a global δ_w increase due to continental ice growth, or a local δ_w increase due to enhanced evaporation or reduced precipitation. Glacioeustatic fluctuations cause global δ_w changes which are recorded by benthic and planktonic foraminifera, although these global changes may be overprinted by temperature or local δ_w changes. Various strategies have been applied successfully using Pleistocene $\delta^{18}O$ records to estimate the timing and magnitude of glacioeustatic changes (see below). We apply these Pleistocene strategies to the pre-Pleistocene $\delta^{18}O$ record, although we note that pre-Pleistocene sea level magnitudes can only be roughly estimated. We previously suggested that pre-Pleistocene sea level magnitudes could be estimated only within a factor of 2 [*Miller et al.*, 1987]. Although magnitudes are uncertain, the timing of global $\delta^{18}O$ variations provides a valuable record of glacioeustatic changes which may be compared with other proxies for sea level change.

Second, the Cenozoic $\delta^{18}O$ record is still surprisingly poorly known. No one location contains a pristine record of any pre-Pliocene epoch, let alone a record of the entire Cenozoic. Our understanding of Cenozoic isotope changes is founded on splicing [e.g., *Prentice and Matthews*, 1988] or stacking [e.g., *Shackleton et al.*, 1984; *Miller et al.*, 1987] of $\delta^{18}O$ records from different locations. Most isotope data published in the 1970s were based on analyses of mixed species or species whose $\delta^{18}O$ composition relative to seawater $\delta^{18}O$ changed through time and therefore are not useful. Recent advances in drilling and mass spectrometer technology helped to delineate changes on the 10^6-year scale, but adequate data are lacking even on this scale in many intervals. In addition, higher-frequency (10^4-10^5 yr) $\delta^{18}O$ fluctuations dominated the Pliocene-Pleistocene record [e.g., *Imbrie et al.*, 1984; *Ruddiman et al.*, 1986]; similar changes apparently occurred during the Oligocene-Miocene [*Shackleton*, 1982; *Poore and Matthews*, 1984; *Pisias et al.*, 1985; *Mead et al.*, 1986]. Undersampling of such high-frequency $\delta^{18}O$ variations may result in signal aliasing possibly resulting in a spurious low-frequency signal [*Pisias and Mix*, 1988].

Third, correlation problems have obscured global $\delta^{18}O$ signals. Although several large (~1.0 ‰) global $\delta^{18}O$ changes have been recognized (e.g., the earliest Oligocene and middle Miocene increases; *Savin et al.* [1975]; *Shackleton and Kennett* [1975]; *Kennett and Shackleton* [1976]), smaller (~0.5 ‰) changes have eluded general recognition partially as a result of correlation problems. To address problems correlating isotope records, we previously stacked and smoothed isotope records from different locations; this removes periods shorter than 1 m.y. [*Miller et al.*, 1987].

Miller et al.'s (1987) oxygen isotope synthesis is appropriate to view overall changes in Cenozoic $\delta^{18}O$ records and to make general comparisons with other sea level records (e.g., *Vail et al.*, [1977]; see examples by *Miller et al.*, [1987]). However, finer scale (e.g., 0.5-2.0 m.y.) variations must be examined in individual and relatively complete $\delta^{18}O$ time series. If more than one location is used for a given interval, stratigraphic correlations must be precise. Within the past few years, resolution of Cenozoic sea level events inferred from passive continental margin studies has been improved (e.g, *Baum* [1986]; *Haq et al.* [1987]; *Miller et al.* [1990*a*;] among others). The overview of *Haq et al.* [1987] suggests that sea level events occurred approximately every 1.5 m.y. To address correlation problems of the stable isotope records, we established first-order correlations between stable isotopes and magnetostratigraphy (Figures 1-4) and integrated these with biostratigraphy. In addition, we use oxygen isotope stratigraphy to fine tune our stratigraphic correlations [e.g., *Miller et al.*, 1989]. In this paper, we examine Oligocene to Miocene $\delta^{18}O$ records at this resolution (0.5-2 m.y. scale) to compare them with passive margin records which have a similar resolution [e.g., *Haq et al.*, 1987]. We identify eight Oligocene-Miocene $\delta^{18}O$ increases in this paper and use them in the definition of isotope zones (Zones Oi1-Oi2, Mi1-Mi6). In addition, a poorly defined $\delta^{18}O$ increase apparently occurred during the "middle" Oligocene ("Oi2a"), while *Wright and Miller* [1990] note three additional Miocene $\delta^{18}O$ increases (events associated with their Zones Mi1a, Mi1b, and Mi7). These 12 Oligocene-early late Miocene $\delta^{18}O$ increases may be correlated with 12 inferred eustatic lowerings of *Haq et al.* [1987].

METHODS AND PRINCIPLES

Data Collection

New benthic foraminiferal oxygen and carbon isotope data were collected from Deep Sea Drilling Project (DSDP) Sites 529 (Oligocene; 28°55.83'S, 02°46.08'E; 3035 m present depth; 2400 to 2650 m paleodepth; Figures 1 and 3), 563 (the Miocene section only; 37°46.24'N, 37°20.61'W, 3754 m present depth; 3600-3200 m paleodepth; Figures 1 and 4), and 608 (Miocene; 42°50.21'N, 23°05.25'W; 3526 m present depth; 3500-3100 m paleodepth; Figures 1 and 4). The Oligocene-Miocene section at Site 563 and the Miocene section at Site 608 previously were analyzed for benthic foraminiferal stable isotopes at approximately one sample per

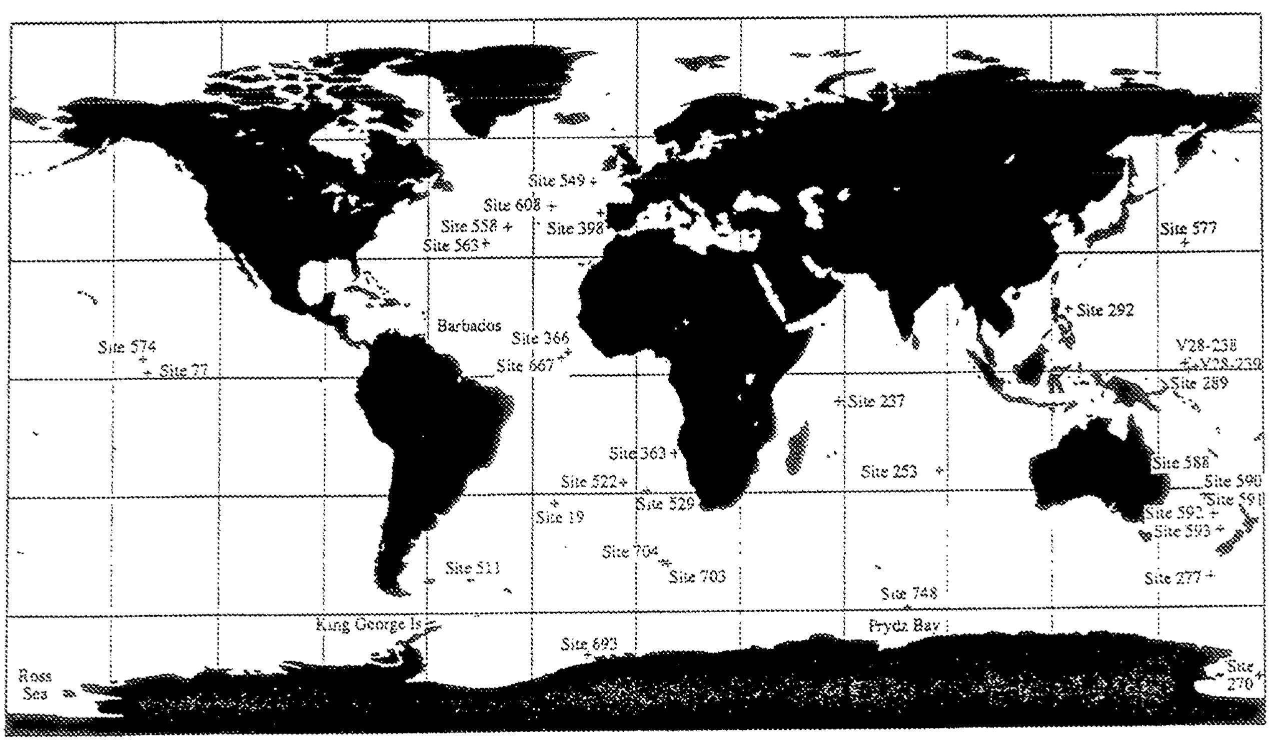

Fig. 1. Global location map showing sites discussed here.

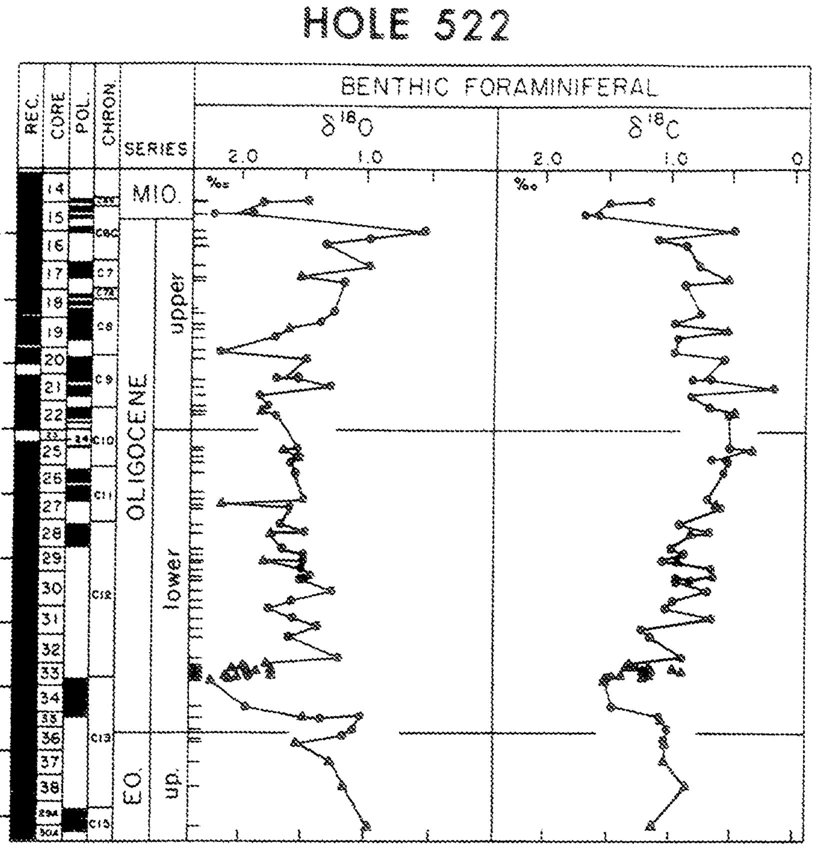

Fig. 2. Site 522 stable isotope record after *Miller et al.* [1988].

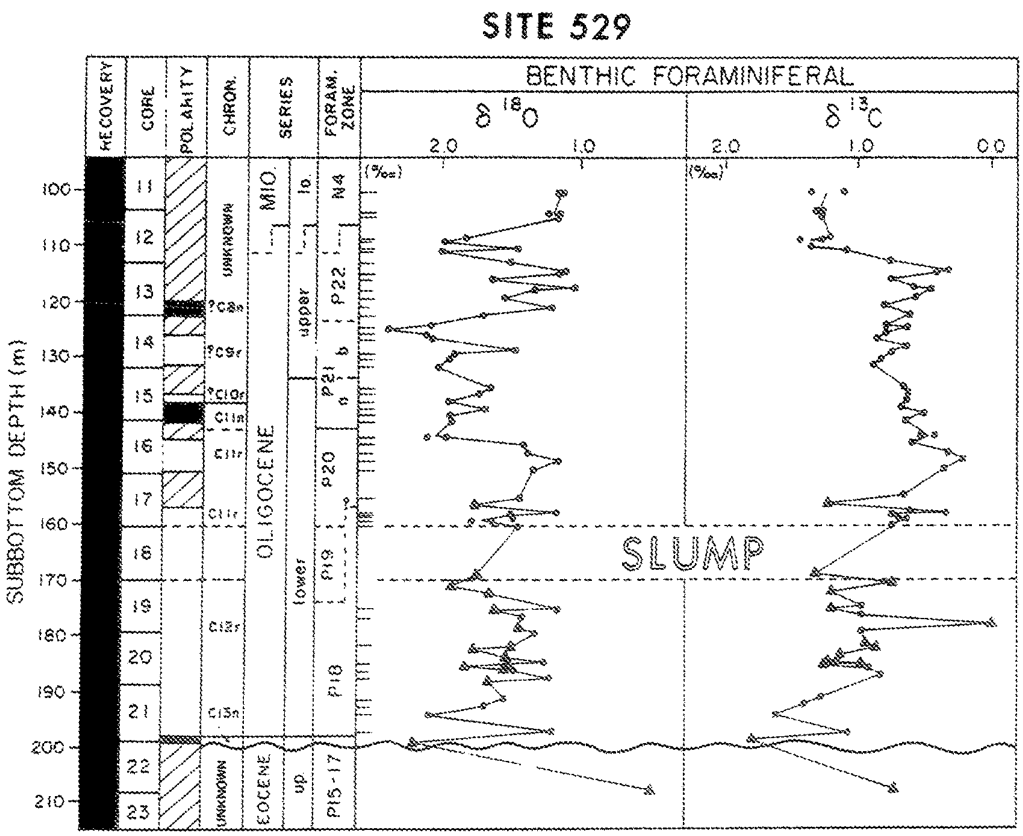

Fig. 3. Site 529 Oligocene isotope record. All data were derived from *Cibicidoides* spp. Magnetobiostratigraphy after K. G. Miller, D. V. Kent, and A. N. Brower [unpublished manuscript, 1990].

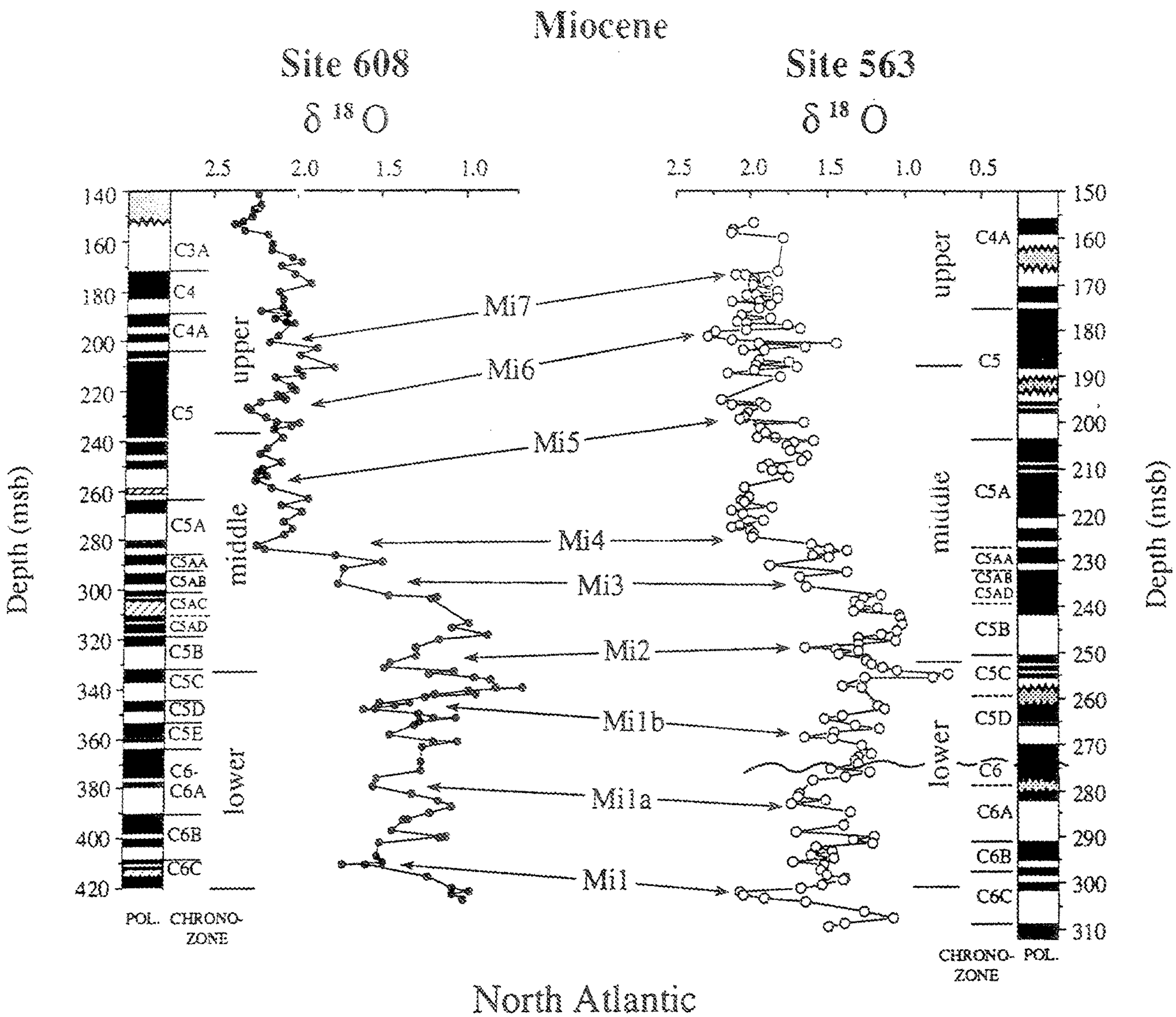

Fig. 4. Miocene oxygen isotope stratigraphy and magnetostratigraphy for eastern North Atlantic Site 608 and western North Atlantic Site 563. Magnetochronology of Site 563 is after *Miller et al.* [1985*b*], while that of Site 608 is after *Clement and Robinson* [1986] and *Miller et al.* [1990*b*]. Miocene Isotope Zone Mi1 is defined at Site 522; Zones Mi2 through Mi6 are defined at Site 608 and the correlations to the Site 563 records are shown. Zones Mi1a, Mi1b, and Mi7 are after *Wright and Miller* [1990].

2.5 m, yielding a resolution of ~0.3-0.4 m.y. [*Miller and Fairbanks*, 1985; *Miller et al.*, 1986]. As part of a Ph.D. thesis on Miocene stable isotopes and deepwater circulation, J. D. Wright increased the coverage of the benthic foraminiferal stable isotope records to one sample per 1.5 m at Site 608 (~0.15 m.y. resolution) and 1 sample per m at Site 563 (~0.1 m.y. resolution)(Figure 4). These data will be tabulated elsewhere [*Wright et al.*, 1990; J.D. Wright, et al., unpublished manuscript, 1990]. One sample for approximately every 2 m was analyzed for the Oligocene section at Site 529, yielding a resolution of 0.25 m.y. (Figure 3; Table 1).

Samples examined for benthic foraminiferal isotopic analyses were washed with sodium metaphosphate and/or hydrogen peroxide (3% solution) in tap water through a 63-μm sieve and air dried. Benthic foraminifera were ultrasonically cleaned for 5-10 s and roasted at 370°C in a vacuum. We analyzed samples of the benthic foraminiferal taxon *Cibicidoides* spp. Studies showed that this taxon accurately records $\delta^{13}C$ variations in seawater and is lower than $\delta^{18}O$ equilibrium by about 0.64 ‰ [e.g., *Shackleton and Opdyke*, 1973; *Graham et al.*, 1981]. The $CaCO_3$ was analyzed at Lamont-Doherty Geological Observatory by either a Carousel-48 automatic carbonate preparation device attached to a Finnigan MAT 251 or by a manual carousel on a VG Micromass 903E. Replicate samples from Site 529 yielded mean $\delta^{18}O$ differences of 0.103 ‰ and mean $\delta^{13}C$ differences of 0.113 ‰ (Table 1; n = 4), while those from Site 608 yielded differences of 0.079 ‰ and 0.055 ‰ (*Wright et al.* [1990]; J. D. Wright et al., unpublished manuscript, 1990); n = 10), respectively.

Age estimates were based on biostratigraphic and magnetostratigraphic correlations. The time scale of *Berggren et al.* [1985] was used for the ages of the biostratigraphic and magnetostratigraphic boundaries, and the ages of our samples were established by linearly interpolating between datum levels. The correlations at Sites 522 and 608 are based on magnetochronology [*Tauxe et al.*, 1983; *Clement and Robinson*, 1986]; (age models by *Miller et al.* [1988, 1990*b*]). The chronologies at Sites 529 and 563 are based on

TABLE 1. Oxygen and Carbon Isotope Data, *Cibicidoides* spp., Site 529

Sample Interval, cm	Depth, msb	Age, Ma	$\delta^{18}O$	$\delta^{13}C$
11-4, 120-124	99.70	22.38	1.14	1.11
11-4, 120-124	99.70	22.38	1.17	1.36
11-7, 21-25	103.21	23.14	1.24	1.32
11-7, 21-25	103.21	23.14	1.16	1.28
12-1, 66-70	104.16	23.34	1.50	1.28
12-3, 96-100	107.96	24.16	1.83	1.22
12-4, 50-54	108.50	24.28	1.98	1.27
12-4, 50-54	108.50	24.28	1.98	1.44
12-5, 20-24	109.70	24.54	1.47	1.37
12-5, 83-87	110.33	24.67	2.01	1.09
12-6, 120-124	112.20	25.08	1.50	0.76
13-1, 91-95	113.91	25.44	1.10	0.31
13-1, 143-146	114.43	25.56	1.15	0.41
13-2, 93-97	115.43	25.77	1.63	0.76
13-3, 90-94	116.90	26.09	1.04	0.59
13-3, 143-146	117.43	26.20	1.33	0.45
13-4, 121-125	118.71	26.48	1.54	0.56
13-5, 120-124	120.20	26.80	1.21	0.80
13-6, 140-144	121.90	27.16	1.70	0.61
14-1, 138-142	123.88	27.59	2.09	0.80
14-2, 36-40	124.36	27.69	2.41	0.64
14-2, 118-120	125.18	27.87	2.13	0.79
14-3, 63-67	126.13	28.08	2.07	0.86
14-4, 84-88	127.84	28.44	1.48	0.64
14-5, 23-27	128.73	28.64	1.91	0.76
14-5, 145-148	129.95	28.90	1.95	0.83
14-6, 111-115	131.11	29.15	2.05	0.90
15-3, 2-5	135.02	29.99	1.64	0.67
15-3, 100-104	136.00	30.20	1.74	0.63
15-4, 95-98	137.45	30.51	1.95	0.64
15-5, 67-71	138.67	30.78	1.70	0.68
15-6, 22-25	139.72	31.00	1.93	0.50
15-7, 7-10	141.07	31.29	1.92	0.65
16-2, 92-95	143.92	31.77	2.11	0.42
16-2, 92-95	143.92	31.77	1.97	0.53
16-3, 55-59	145.05	31.90	1.42	0.59
16-4, 77-80	146.77	32.10	1.38	0.32
16-5, 59-62	148.09	32.25	1.15	0.22
16-6, 80-85	149.80	32.45	1.34	0.36
17-3, 81-85	154.81	33.03	1.44	0.66
17-4, 71*	156.21	33.20	1.79	1.24
17-5, 47-51	157.47	33.34	1.17	0.61
17-5, 76-79	157.76	33.38	1.50	0.36
17-5, 113-118	158.13	33.42	1.49	0.75
17-6, 47-52	158.97	33.52	1.64	0.66
17-6, 47-52	158.97	33.52	1.80	0.71
17 cc	160.54	33.70	1.46	0.74
19-1, 73*	170.73	33.89	1.94	0.74
19-2, 73*	175.23	34.16	1.62	1.20
19-4, 10-14	174.60	34.12	1.16	0.98
19-5, 33-37	176.33	34.20	1.42	0.99
19-6, 73*	178.23	34.34	1.44	0.01
19-7, 32-35	179.32	34.35	1.32	0.98
20-2, 50*	181.50	34.54	1.49	0.95
20-2, 115*	182.15	34.57	1.77	0.86
20-3, 63*	183.63	34.66	1.54	1.12
20-4, 49-54	184.49	34.59	1.26	1.15
20-4, 87*	184.87	34.74	1.52	1.22
*20-4, 113	185.13	34.75	1.84	0.98
*20-5, 7	185.57	34.78	1.55	1.27
20-5, 42-46	185.92	34.62	1.47	0.92
20-6, 12-17	187.12	34.72	1.22	0.84
21-2, 53-57	191.03	34.91	1.55	1.27
21-3, 22-26	192.22	34.96	1.69	1.41
21-4, 19-23	193.69	35.04	2.10	1.62
21-6, 31-35	196.81	35.18	1.19?	1.09
*22-1, 50	199.00	35.58	2.22	1.81
*22-7, 38	207.88	38.00	0.47	0.73

*Data from *Shackleton et al.* [1984].

integrated magnetobiostratigraphy [*Miller et al.*, 1985*b*; K.G. Miller et al., unpublished data, 1990]; in addition, oxygen isotope stratigraphy was used to "tune" the Site 563 and 529 records to the more complete Sites 522 and 608 records.

Stable Isotopes and Ice Volume

Marine paleoclimate reconstructions rely heavily on foraminiferal $\delta^{18}O$ records which provide a proxy for ocean temperature, local salinity, and global ice volume changes. Foraminiferal $\delta^{18}O$ records do not provide unique interpretations, since their shells incorporate all three variables. All foraminiferal isotope studies make certain assumptions to isolate the ice volume signal contained in foraminiferal carbonate [cf., *Savin et al.*, 1975; *Shackleton and Kennett*, 1975; *Miller et al.*, 1987; *Prentice and Matthews*, 1988]. To place limits on foraminiferal $\delta^{18}O$ interpretations, we apply Pleistocene strategies [*Shackleton and Opdyke*, 1973, 1976; *Fairbanks and Matthews*, 1978], use realistic assumptions of possible temperature and salinity distributions (*Miller et al.* [1987], and this study), and compare the oxygen isotope record with independent climate monitors.

Pleistocene strategies. Late Pleistocene $\delta^{18}O$ records constrain ice volume (glacioeustatic) changes. The calibration for late Pleistocene sea level and $\delta^{18}O$ has been well established (0.11 ‰/10 m)[*Fairbanks and Matthews*, 1978] and the orbital (Milankovitch) forcing mechanism has been confirmed [*Hays et al.*, 1976]. Although previous estimates of the last glacial lowstand (18 Ka) have ranged from about 80 to 165 m below present, recent drilling of reefs offshore Barbados documented 121±5 m of glacioeustatic lowering at 18 Ka. This indicates that approximately 1.3 ‰ of the foraminiferal $\delta^{18}O$ signal from 18 Ka to present is attributable to global seawater changes [*Fairbanks*, 1989]. Oxygen isotope values in deepwater benthic foraminifera were 1.75 ‰ higher at 18 Ka than they are today, suggesting that global deep waters were cooler by approximately 2°C. Because of changing deepwater temperatures, benthic foraminiferal $\delta^{18}O$ records alone do not provide a precise indicator of the magnitude of Pleistocene glacioeustatic changes.

Concomitant changes in benthic and low-latitude (nonupwelling) planktonic $\delta^{18}O$ records have been used as a Pleistocene ice volume indicator [*Shackleton and Opdyke*, 1973, 1976]. Further studies validate this technique for the Pleistocene-present. Covariance between benthic and western equatorial Pacific planktonic foraminiferal $\delta^{18}O$ records was 1.2-1.3 ‰ from the last glacial maximum to present [*Shackleton and Opdyke*, 1973]. This predicts a sea level rise of 109-118 m [*Fairbanks and Matthews*, 1978] similar to the 121±5 m established by *Fairbanks* [1989]. We previously used covariance between benthic and low-latitude planktonic $\delta^{18}O$ signals as a proxy for Cenozoic ice volume changes [*Miller et al.*, 1987]. In contrast, *Savin* [1977] interpreted an early Oligocene $\delta^{18}O$ increase in both benthic and low-latitude planktonic foraminifera as a synchronous cooling of deep water and surface water, since the Oligocene-early Miocene was assumed to be ice free [*Savin et al.*, 1975; *Shackleton and Kennett*, 1975]. Based on the evidence discussed below, we assume that large ice sheets existed during parts of the Oligocene-early Miocene. If such ice sheets existed, covariance between benthic and low-latitude $\delta^{18}O$ signals provides the strongest evidence for a global change in δ_w resulting from ice volume variations.

The use of benthic-planktonic covariance does not require that either deepwater or surface water temperatures remain constant. However, surface water regions with large seasonal changes (e.g., upwelling, high-latitude regions) should be avoided in these comparisons. The western equatorial regions are the most thermally stable surface waters on an annual basis [*Ravelo et al.*, 1990]. These regions provide the best opportunity to obtain planktonic foraminiferal $\delta^{18}O$ time series which are less subject to temperature effects than other surface water regions [*Matthews and Poore*, 1980; *Prentice and Matthews*, 1988]. *Crowley and Matthews* [1983] suggested that planktonic $\delta^{18}O$ records from subtropical gyres also may record primarily an ice volume signal, although this is less certain than for the western equatorial regions.

All previous attempts at estimating tropical surface water $\delta^{18}O$ values have suffered from inadequate data, particularly from the western equatorial region (Figure 5). Low-latitude planktonic foraminiferal $\delta^{18}O$ time series are poor for the pre-Pliocene (Figure 5). No appropriate data are available from circa 55-37 Ma (middle to late Eocene). The only Oligocene data are from central equatorial Pacific Site 77 [*Keigwin and Keller*, 1984] which today has a substantial upwelling signal (i.e., the present-day temperatures average 24°C)(Figure 5). Burial depths for the lower Miocene at western equatorial Pacific Site 289 (Figure 5) exceed 500 m, and the oxygen isotope signal may be altered [*Elderfield et al.*, 1982].

Prentice and Matthews [1988] compiled planktonic foraminiferal $\delta^{18}O$ records for the Cenozoic, attempting to characterize average equatorial sea surface values. Because tropical nonupwelling records are scarce, they used middle latitude and tropical upwelling planktonic $\delta^{18}O$ records, calculated paleotemperature gradients, and projected their values to expected equatorial values. They assumed that sea surface temperatures (SST) in the equatorial regions were

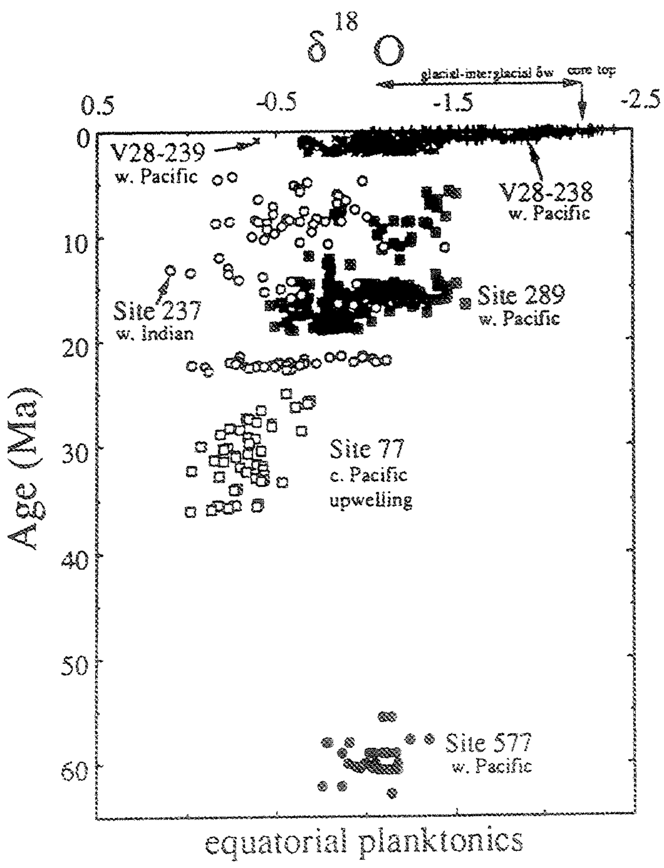

Fig. 5. Tertiary equatorial planktonic $\delta^{18}O$ records. Data after the following authors: Site 77, *Keigwin and Keller* [1984]; Site 237, *Vincent et al.* [1985]; Site 289, *Shackleton* [1982] and *Savin et al.* [1985]; Site 577, *Shackleton et al.* [1985]; V28-238, *Shackleton and Opdyke* [1973]; V28-239, *Shackleton and Opdyke* [1976]. Modern core top value is for V28-238; the glacial-interglacial δ_w change (1.3 ‰) is after *Fairbanks* [1989].

constant (28°C) throughout the Cenozoic. However, several studies suggested that western equatorial regions were cooler in the past [*Savin et al.*, 1985; *Shackleton and Boersma*, 1981; *Mix and Ruddiman*, 1984]. Thus, their projections may not be a valid means of estimating equatorial $\delta^{18}O$ records.

This discussion underscores our view that the present planktonic $\delta^{18}O$ data are inadequate. We need additional data from western equatorial locations (e.g., Ontong-Java Plateau (Leg 133), Ceara Rise, Mascarene Plateau (Leg 115)) to monitor the history of low-latitude nonupwelling regions. Nevertheless, available sections show that western equatorial regions faithfully record ice volume fluctuations. The lower-middle Miocene $\delta^{18}O$ record at Site 237 (Figures 1 and 5) provides a western equatorial Indian Ocean surface water monitor (9° S paleolatitude; *Sclater et al.* [1985]). Although the seasonally reversing monsoon potentially affects this site, the uplift of the Himalayas was not sufficient to cause intense monsoonal circulation changes until the late Neogene [*Wright and Thunell*, 1988; *Ruddiman and Kutzbach*, 1989]. Site 237 provides the best available western equatorial record suitable for monitoring early to middle Miocene ice volume changes (Figure 5). In the absence of suitable western equatorial records (e.g., the entire Oligocene), we used subtropical planktonic $\delta^{18}O$ records to document benthic-planktonic $\delta^{18}O$ covariance suggestive of ice volume changes [*Miller and Fairbanks*, 1985; *Miller et al.*, 1987].

Temperature-salinity distributions. High $\delta^{18}O$ values occurred in deep-sea benthic foraminifera (> 1.8 ‰ in *Cibicidoides* spp.) during portions of the Oligocene-Miocene (Figures 2-4). We attribute these high $\delta^{18}O$ values to deep water produced at high latitudes (see below). To assume ice-free conditions, production of deep water at high latitudes, and $\delta^{18}O$ values in *Cibicidoides* of greater than 1.8 ‰ requires that deep water and high-latitude surface water were colder than they are at present (~2°C). These cold high-latitude temperatures are incompatible with an ice-free world [*Miller and Fairbanks*, 1983, 1985; *Miller et al.*, 1987]. We suggest that benthic foraminiferal values greater than 1.8‰ (in *Cibicidoides*) are indicative of an ice house world. However, it is equivocal if ice sheets existed during intervals when benthic foraminiferal values were less than 1.8‰ [*Miller et al.*, 1987].

Alternatively, various studies suggested that Cenozoic deep waters may have been warmer and saltier, originating in low-latitude evaporative basins or continental shelves [e.g., *Brass et al.*, 1982; *Prentice and Matthews*, 1988; *Woodruff and Savin*, 1989]. Fractionation associated with higher salinities would result in higher deepwater $\delta^{18}O$ values. We do not dispute the possible production of Warm Saline Deep Water (WSDW) in low latitudes, although it was not the dominant mode of deepwater formation during the Oligocene to Miocene. First, there is geological evidence (seismic stratigraphic, rock stratigraphic, benthic foraminiferal, stable isotope) for the production of high-latitude deep water during the Oligocene to Miocene (*Kennett*, [1977]; *Miller and Tucholke*, [1983]; *Miller and Fairbanks*, [1983; 1985]; among others). Second, to assume production of warm saline deep water, an ice-free world, and $\delta^{18}O$ in *Cibicidoides* greater than 1.8 ‰ requires unrealistic deepwater salinities.

Salinities required for given temperatures and $\delta^{18}O$ values can be quantified (J. D. Wright et al., unpublished manuscript, 1990). To compute the $\delta^{18}O$ value of an ice-free world, we subtracted 0.9‰ from the modern *Cibicidoides* value (2.7‰)[*Shackleton and Kennett*, 1975]. The first-order $\delta^{18}O$/temperature relationship is ~0.23‰/1°C [*O'Neil et al.*, 1969]. We applied two $\delta^{18}O$/salinity relationships to estimate the effects of WSDW production on salinity; these relationships are derived from modern surface waters from the open ocean (0.6‰/1‰; *Broecker* [1986]) and the modern Red Sea (0.3‰/1‰; *Craig* [1966]). We calculated oxygen isotope-T-S isopleths using the $\delta^{18}O$/temperature relationship and each of the $\delta^{18}O$/salinity relationships. These isopleths represent deepwater temperature and corresponding salinity combinations required to produce an observed $\delta^{18}O$ value. Warm saline deepwater production in a subtropical basin should yield temperatures of 10° to 20°C versus modern deepwater temperature of approximately 2°C. For example, the assumption of constant 28°C equatorial SST and the compilation of *Prentice and Matthews* [1988] requires deepwater temperatures of 10°-18°C between 7 and 65 Ma (their fig. 3). Modern deepwater temperatures produced in evaporative basins such as the Mediterranean exceed 13°C [*Miller et al.*, 1970]. Assuming temperatures of 10° to 20°C, an ice-free world, and $\delta^{18}O$ greater than 1.8 ‰ (in *Cibicidoides*), requires that deepwater salinities were higher than modern salinities by 3‰ to 7‰ using the modern open ocean slope and 6‰ to 14‰ using the Red Sea slope. These salinity changes do not include global changes in salinity such as occurred during the Messinian. These global salt transfers do not affect the oceanic $\delta^{18}O$ values.

Salt must be conserved within the ocean at a given time. As a result, the production of WSDW transfers salt and ^{18}O into the deep ocean, while it stores fresh water in the surface ocean. The deep ocean reservoir is at least 3-4 times larger than the surface ocean reservoir [e.g., *Broecker and Peng*, 1987]; therefore, small changes in deepwater salinity will result in large surface-ocean $\delta^{18}O$ and salinity variations. For example, a 3‰ increase in deepwater salinities must result in at least a 9-12‰ decrease in average surface ocean salinities, with some surface ocean regions experiencing greater salinity changes. Such salinity changes may be unreasonable, because they would result in surface ocean salinities which are below the tolerance limits of many marine taxa.

This reasoning does not apply to isolated or restricted production of WSDW, but to intervals when the entire deep ocean was filled with high $\delta^{18}O$ values (> 1.8 ‰ in *Cibicidoides*). Because of conservation of salt, these high values cannot be explained by the filling the deepwater reservoir with low-latitude saline waters without very large surface ocean salinity changes. Thus we conclude that the high $\delta^{18}O$ values (> 1.8 ‰ in *Cibicidoides*) which occurred during parts of the Oligocene to early Miocene require that this was an ice house world.

Comparison with other proxies. We use covariance between benthic and low-latitude planktonic $\delta^{18}O$ records as our primary tool for recognizing Oligocene-Miocene ice volume changes. We discuss below the match between these inferred glacioeustatic fluctuations and other records of sea level and ice volume changes. We conclude that our inferred glacioeustatic history is consistent with the record of glacially derived sediments found on and near Antarctica. In addition, we note that there is a good correspondence between inferred glacioeustatic changes and the inferred sea level record of *Haq et al.* [1987]. These comparisons are discussed in the results.

Stable Isotope Stratigraphy

Pioneering stable isotope studies noted that large (> 1.0 ‰) benthic foraminiferal $\delta^{18}O$ increases occurred near the Eocene/Oligocene boundary and in middle Miocene [*Savin et al.*, 1975; *Shackleton and Kennett*, 1975]. An early question was whether these increases were synchronous from location to location. For example, *Kennett and Shackleton* [1976] reported a major benthic foraminiferal $\delta^{18}O$ increase in the lowermost Oligocene at high-latitude Site 277. In contrast, *Savin et al.* [1975] reported a large (> 1.0 ‰) benthic foraminiferal $\delta^{18}O$ increase in the upper Eocene at low-latitude DSDP sites. Further study has documented that these increases are biostratigraphically synchronous (i.e., within a 0.5- to

1.0- m.y. resolution), and they are correlated with earliest Oligocene Chron C13n (*Keigwin* [1980]; *Corliss et al.* [1984], *Oberhansli et al.* [1984], *Miller et al.* [1985*a*, 1988], among others). Similarly, the middle Miocene $\delta^{18}O$ increase is a synchronous event in every ocean (*Savin et al.* [1981], *Woodruff et al.* [1981], *Miller and Fairbanks* [1983, 1985], *Vincent et al.* [1985], among others).

Other $\delta^{18}O$ variations, both large (> 1.0 ‰) and small (~0.5 ‰), occurred during the Cenozoic on the 10^6-year scale. These variations provide potential for stratigraphic correlations. As we document below, at least eight Oligocene-Miocene benthic foraminiferal $\delta^{18}O$ increases are synchronous among locations within the resolution of biostratigraphy; seven of these are synchronous between at least two locations on the basis of magnetostratigraphy (e.g., Figure 4). This establishes the chronostratigraphic significance of these events, and we follow procedures of *Hedberg* [1976] in erecting oxygen isotope zones.

We erect eight zones for the Oligocene to Miocene, and *Wright and Miller* [1990] erect two additional early Miocene zones and one late Miocene zone. The base of each zone is defined on the maximum $\delta^{18}O$ value, and an alphanumeric name given to the zone numbering from the base of the Oligocene (Oi zones) and the base of the Miocene (Mi zones). For example, (1) the base of Zone Oi1 (Oligocene Isotope 1) is defined as the maximum benthic foraminiferal $\delta^{18}O$ value at Site 522 at 133.13 m subbottom (msb; Figure 2, Table 2); (2) the base of Zone Oi2 is defined as the maximum benthic foraminiferal $\delta^{18}O$ value at 143.92 msb at Site 529 (Figure 3, Table 3); (3) the base of Zone M1 is defined as the maximum $\delta^{18}O$ value at 56.93 msb at Site 522 (Figure 2, Table 4). The overlying zones for the Miocene, Mi2-Mi6, are defined at Site 608 (Figure 4, Tables 5-9); *Wright and Miller* [1990] defined Zones Mi1a, Mi1b, and Mi7 at Sites 563 and 608 and corroborated their significance at Site 747. This nomenclature varies from Plio-Pleistocene stable isotope "stages" or "events" (more properly zones; *Hedberg* [1976]) which are defined on the inflection points of the increases and decreases and are numbered from the top (present) downward. Our nomenclature is consistent with Cenozoic biozones which are numbered from the base upward.

At least two attempts have been made to designate formally Miocene isotope zones. *Loutit et al.* [1983] used carbon isotope variations in benthic foraminifera at Site 289 to subdivide the Miocene. However, the global significance of their isotope variations has not been established. *Vincent et al.* [1985] used oxygen isotopes, carbon isotopes, and calcium carbonate fluctuations to delineate markers A through H for the early to middle Miocene, and established that many of these were synchronous on the basis of biostratigraphy among Indian and Pacific locations. Their G event, the only one based on a $\delta^{18}O$ increase, is equivalent to the $\delta^{18}O$ increase associated with the base of Zone Mi3.

Ordinal pattern matching techniques (e.g., magnetostratigraphy, oxygen isotope stratigraphy) cannot be used readily to detect gaps in the record. There are cases when the Cenozoic $\delta^{18}O$ record has sufficient character that hiatuses can be recognized, but generally hiatuses are the bane of ordinal correlation tools such as oxygen isotopes. Using isotope correlations alone may result in the misidentification of an absent zone and spurious correlations. However, when properly integrated with biostratigraphy, oxygen isotope stratigraphy is a powerful correlation tool.

RESULTS AND DISCUSSION

Glacial Sediments Support the Oligocene Ice House World

Before 1980, isotope studies assumed substantially ice-free conditions prior to the middle Miocene [*Savin et al.*, 1975; *Shackleton and Kennett*, 1975]. *Matthews and Poore* [1980] attacked this assumption, noting that it was reasonable to interpret the earliest Oligocene $\delta^{18}O$ increase in both benthic and planktonic foraminifera (= base of Zone Oi1) as reflecting ice growth. Previously, the $\delta^{18}O$ record was hindered by poor Oligocene sections and analyses of mixed species which obscured evidence for Oligocene ice sheets. Drilling in the early 1980s yielded appropriate Oligocene sections, and high $\delta^{18}O$ values (> 1.8 ‰ in *Cibicidoides*) were obtained from benthic foraminifera which supported the assumption of Oligocene ice sheets [*Miller and Fairbanks*, 1983, 1985; *Keigwin and Keller*, 1984; *Shackleton et al.*, 1984; *Miller and Thomas*, 1985]. Using the improved $\delta^{18}O$ records and evidence of benthic-planktonic $\delta^{18}O$ covariance, we suggested that there were at least three major intervals of Oligocene glaciation [*Miller et al.*, 1987].

The high-latitude record of Oligocene-middle Miocene

TABLE 2. Definition of Base of Zone Oi1

Parameter	Description
Type Level	Hole 522, 34-3, 43-45 cm, 133.13 msb; *Oberhansli et al.* [1984] report the maximum $\delta^{18}O$ value in *Stilostomella* spp. at this level which is slightly below the increase reported in *Cibicidoides* spp. (Fig.2;*Poore and Matthews* [1984], *Miller et al.* [1988]) due to the sampling interval.
Age Estimate:	35.8 Ma.
Correlation:	
First Order:	Chronozone C13n [*Tauxe et al.*, 1983;]; Zone NP21 [*Percival*, 1984]; near base of Zone P18 of *Berggren and Miller* [1988] based upon foraminiferal biostratigraphy of *Poore* [1984].
Locations Observed:	
Benthic $\delta^{18}O$	Site 19 [*Keigwin and Corliss*, 1986], Site 77 [*Keigwin and Keller*, 1984], Site 277 [*Kennett and Shackleton*, 1976; *Keigwin*, 1980], Site 292 [*Keigwin*, 1980], Site 363 [*Keigwin and Corliss*, 1986], Site 511 [*Muza et al.*, 1983], Site 540 [*Belanger and Matthews*, 1984], Site 574 [*Miller and Thomas*, 1985], Sites 592 and 593 [*Murphy and Kennett*, 1986], Barbados [*Saunders et al.*, 1984], Site 748 [*Zachos et al.*, 1990]. See summary by *Vergnaud-Grazzini* and *Oberhansli* [1986].
Planktonic $\delta^{18}O$	Sites 253, 363, [*Keigwin and Corliss*, 1986], Site 292 [*Keigwin*, 1980], Site 522 [*Oberhansli et al.*, 1984].
Comment:	Sites with probable disconformities (e.g., Site 529, [*Shackleton et al.*, 1984]; (this study)) or possible diagenetic alteration (e.g., Sites 398 [*Vergnaud-Grazzini et al.*, 1978] and 366 [*Miller et al.*, 1989]) have not been listed.

TABLE 3. Definition of Base of Zone Oi2

Parameter	Description
Type Level	Site 529, Sample 16-2, 92-95 cm, 143.92 msb.
Age Estimate	31.5 to 32.5 Ma.
Correlations:	
First Order	Chronozone C11r or lowermost C11n (Fig. 3); upper Zone P20 of *Berggren and Miller* [1988]; between the last occurrences of *"Turborotalia" ampliapertura* and *Subbotina angiporoides* and the first occurrence of *Globigerina angulisuturalis*.
Second Order	middle of Zone NP23.
Locations Observed	
Benthic $\delta^{18}O$ Records	Site 77 [*Keigwin and Keller*, 1984], Site 574 [*Miller and Thomas*, 1985], Site 593 [*Murphy and Kennett*, 1986], Site 703 (*K. G. Miller and M. E. Katz*, unpublished manuscript, 1990).
Planktonic $\delta^{18}O$ Records	Site 253 [*Oberhansli*, 1986], Site 593 [*Murphy and Kennett*, 1986].
Comment	This event has not been observed in low-latitude planktonic records as a result of poor recovery and diagenesis.

TABLE 4. Definition of Base of Zone Mi1

Parameter	Description
Type Level	Hole 522, Sample 15-2, 33-37 cm, 56.93 msb
Age Estimate	23.5 Ma
Correlation:	
First Order	Chronozone C6Cn, in the short reversed between C6Cn2 and C6Cn3; spanning NP25/NN1 boundary (the LO of *Reticulofenestra bisecta* is between 15-2, 100-102 cm and 15-12, 100-102 cm;
Percival [1984])	
Second Order	base of Zone N4a.
Locations Observed	
Benthic $\delta^{18}O$ Records	Sites 558 and 563 [*Miller and Fairbanks*, 1985; this study], Sites 366 and 667 [*Miller et al.*, 1989], Site 703 (K. G. Miller and M. E. Katz, unpublished manuscript, 1990), Site 704 (J. D. Wright et al., unpublished manuscript, 1990).
Planktonic $\delta^{18}O$ Records	Sites 558 and 563 [*Miller and Fairbanks*, 1985].

TABLE 5. Definition of Base of Zone Mi2

Parameter	Description
Type Level:	Hole 608, Sample 36-2 60-64 cm, 331.40 msb.
Age Estimate	16.1 Ma.
Correlation:	
First Order	base of Chronozone C5Br, Zone N8 (*Praeorbulina sicanus* first occurs between Samples 36-4 34-38 cm and 36-5 100-104 cm; *Miller et al.* [1990*b*]).
Second Order	near Zone NN4/NN5 boundary.
Locations Observed	
Benthic $\delta^{18}O$ Records	Site 563 [*Miller and Fairbanks*, 1983, 1985; this study]; Site 289 [*Woodruff et al.*, 1981]; Site 237 [*Vincent et al.*, 1985]; Sites 588 and 590 [*Kennett*, 1986]; Sites 366 and 667 [*Miller et al.*, 1989]; Site 317 [*Woodruff and Savin*, 1989].
Planktonic $\delta^{18}O$ Records	Site 237 [*Vincent et al.*, 1985].

sediments also was limited until recently [e.g., *Tucholke et al.*, 1976]. Prior to 1986, glaciomarine sediments from the Ross Sea, Site 270, provided the only unequivocal record of pre-Miocene glaciation [*Hayes et al.*, 1975; *Leckie and Webb*, 1986]. *Margolis and Kennett* [1971] reported possible Eocene ice rafted detritus off Antarctic, and *LeMasurier and Rex* [1982] reported deposition of Oligocene hyaloclastites in Marie Byrd Land (Figure 6); however, this evidence was believed to be inconclusive by many [*Kennett*, 1982]. A greater antiquity of glacially derived sediments has been revealed by the drilling of the MSST-1 and the CIROS-1 boreholes in the Ross Sea [*Barrett*, 1986; *Barrett et al.*, 1987, 1988, 1989], drilling of the Southern Ocean by Ocean Drilling Program Legs 113, 114, 119 and 120 [*Barker et al.*, 1988; *Ciesielski et al.*, 1988; *Barron et al.*, 1989; *Schlich et al.*, 1989], and studies of sections on King George Island, South Shetland Islands

TABLE 6. Definition of Base of Zone Mi3

Parameter	Description
Type Level:	Hole 608, Sample 32-5 80-84 cm, 297.70 msb.
Age Estimate	13.6 Ma.
Correlation	
First Order	Chronozone C5ABr.
Second Order	Zone N11; according to the *Berggren et al.* [1985] time scale, this should be late Zone NN6. However, *Takayama and Sato* [1986] reported the last occurrence of *S. heteromorphus* (= base of Zone NN6) at Site 608 in Core 32 (near the base of Zone Mi3).
Locations Observed	
Benthic $\delta^{18}O$ Records	Site 563 [*Miller and Fairbanks*, 1983, 1985; this study]; Site 289 [*Woodruff et al.*, 1981]; Site 237 [*Vincent et al.*, 1985]; Sites 588, 590, and 591 [*Kennett*, 1986]; Site 317 [*Woodruff and Savin*, 1989]; Site 709 [*Woodruff et al.*, 1990].
Planktonic $\delta^{18}O$ Records	Site 237 [*Vincent et al.*, 1985].

TABLE 7. Definition of Base of Zone Mi4

Parameter	Description
Type Level	Hole 608, Sample 31-1 85-90 cm, 282.15 msb.
Age Estimate	12.6 Ma.
Correlation	
First Order	base of Chronozone C5Ar, Zones N12-13.
Second Order	Zones N11-N12; Zone NN7.
Locations Observed	
Benthic $\delta^{18}O$ Records	Site 563 [*Miller and Fairbanks*, 1983, 1985; this study]; Site 289 [*Woodruff et al.*, 1981]; Site 77 [*Savin et al.*, 1985]; Site 237 [*Vincent et al.*, 1985]; Sites 588, 590, and 591 [*Kennett*, 1986]; Sites 366 and 667 [*Miller et al.*, 1989]; Site 317 [*Woodruff and Savin*, 1989]; Site 709 [*Woodruff et al.*, 1990].
Planktonic $\delta^{18}O$ Records	not observed.

TABLE 8. Definition of Base of Zone Mi5

Parameter	Description
Type Level:	Hole 608, Sample 28-3 23-27 cm, 255.73 msb.
Age Estimate	11.3 Ma.
Correlation	
First Order	base Chronozone C5r, near Zone N13/N14 boundary (*Globigerina nepenthes* first occurs in Sample 28-2 17-21 cm; *Miller et al.* [1990*b*])
Second Order	Zone NN7
Locations Observed	
Benthic $\delta^{18}O$ Records	Site 563 [*Miller and Fairbanks*, 1983, 1985; this study]; Site 77 [*Savin et al.*, 1985]; Sites 237 and 238 [*Vincent et al.*, 1985].
Planktonic $\delta^{18}O$ Records	Site 237 [*Vincent et al.*, 1985].

(*Birkenmajer* [1987], among others)(Figure 1). We summarize the Oligocene glaciomarine and till records described by these studies, comparing it with a summary of Oligocene benthic foraminiferal $\delta^{18}O$ records (Figure 6). Although the $\delta^{18}O$ summary is for benthic foraminifera, we note that there were concomitant increases in subtropical planktonic foraminifera associated with the bases of Zones Oi1, Oi2, and Mi1 (see below; Figure 6).

Barron et al. [1989] report undifferentiated upper Eocene-lower Oligocene massive diamictites in Prydz Bay, eastern Antarctica (Figures 1 and 6) and suggest that the glacial record of Antarctica may have begun in the middle or late Eocene. Stratigraphic control is poor for the glacial sediments underlying the upper Eocene-lower Oligocene diamictites, and superposition only requires that they be lowermost Oligocene or older. We have argued that the early Eocene was free of ice [*Miller et al.*, 1987], but we believe that the case for middle to late Eocene ice is still uncertain.

The oldest definite ice rafted detritus (IRD) found near Antarctica is from the lowermost Oligocene, where it is found

TABLE 9. Definition of Base of Zone Mi6

Parameter	Description
Type Level:	Hole 608, Sample 25-3 68-72 cm, 227.38 msb.
Age Estimate	9.6 Ma.
Correlation	
First Order	lower Chronozone C5n, Zone N16
Second Order	Zone NN8
Locations Observed	
Benthic $\delta^{18}O$ Records	Site 563 [*Miller and Fairbanks*, 1983, 1985]; Site 289 [*Woodruff et al.*, 1981]; Sites 237 and 253 [*Vincent et al.*, 1985]; Sites 588, 590, and 591 [*Kennett*, 1986]; Sites 366 and 667 [*Miller et al.*, 1989]; Site 317 [*Woodruff and Savin*, 1989]; Site 709 [*Woodruff et al.*, 1990]; Site 704 [*Wright et al.*, 1990; Site 360 [*Wright et al.*, 1990].
Planktonic $\delta^{18}O$ Records	not observed.

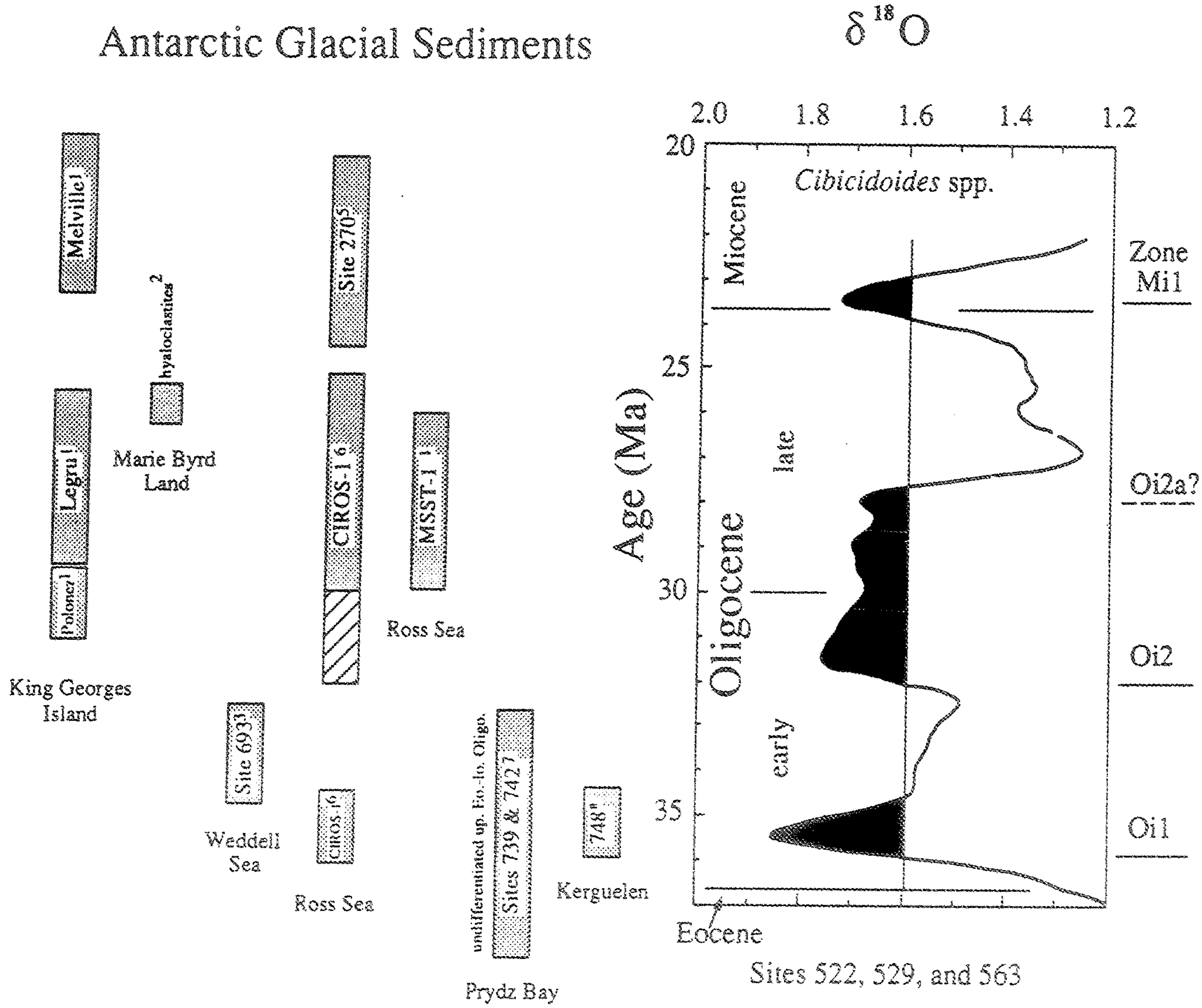

Fig. 6. Oligocene benthic foraminiferal synthesis compared with the record of glaciomarine sediments. Benthic foraminiferal stable isotope data from Sites 522 [*Miller et al.*, 1988], 529 (this study), and 563 [*Miller and Fairbanks*, 1985] were stacked and smoothed with a Gaussian convolution filter in order to remove periods less than 1.0 m.y. Since filtering dampens the amplitude, an arbitrary line was placed through 1.6 ‰ and values higher than this were shaded. Superscripts are 1, *Birkenmajer* [1987]; 2, *LeMasurier and Rex* [1982]; 3, *Barker et al.* [1988]; 4, *Barrett et al.* [1987]; 5, *Leckie and Webb* [1986]; 6, *Barrett et al.* [1988, 1989]; 7, *Barron et al.* [1989]; 8, *Breza et al.* [1988].

not only in the Weddell and Ross Seas (Figures 1 and 6), but also as far north as 58°S (Site 748)[*Breza et al.*, 1988]. The record at Site 748 provides a "smoking gun"; *Zachos et al.* [1990] established that the IRD is associated precisely with the lowermost Oligocene $\delta^{18}O$ increase (the base of Zone Oi1). Lowermost Oligocene glacial sediments have also been reported from the Ross Sea (CIROS-1; Figure 6; *Barrett et al.* [1988, 1989]). It is clear that widespread ice sheet grew during the earliest Oligocene, influencing marine sedimentation far away from the continent itself. Although interpretations of glaciomarine sediments are equivocal (see below), the presence of IRD at a great distance from the Antarctic continent argues for a significant early Oligocene ice sheet.

The "middle" Oligocene has drawn considerable attention since *Vail et al.* [1977] first postulated that the largest sea level fall of the Phanerozoic occurred at circa 29 Ma. There is a 0.8 to 1.0 ‰ $\delta^{18}O$ increase in the "middle" Oligocene (at the base of Zone Oi2; Figure 3), and we estimate the age as circa 32.5-31.5 Ma (Figure 6). Despite our attempts to identify and correlate this event, its age remains the poorest documented of our isotope zones. In fact, it appears from the available data that a second "middle" $\delta^{18}O$ increase occurred at circa 28 Ma (i.e., near the top of Zone P21b in Figure 3; see also *Keigwin and Keller* [1984] and *Miller and Thomas* [1985]). The glacial sediment record is also poorly dated for this "middle" Oligocene interval. At CIROS-1 in the Ross Sea, the lowermost Oligocene glacial sediments are overlain by interglacial sediments which are in turn overlain by "middle" Oligocene glacial sediments between 350 and 160 msb (Figure 6)[*Barrett et al.*, 1988]. *Harwood et al.* [1989] correlate this interval with Chrons C10n to C7n (~30.0-25.5 Ma), although the magnetostratigraphic correlations are sufficiently ambiguous to allow correlation of this interval to Chrons C11n to early C8n (circa 32.0-27.5 Ma). *Harwood et al.* [1989] report a disconformity at CIROS-1 at 350 msb near the base of the "middle" Oligocene diamictites and suggest an age of 34.5 to 30.5 Ma for the hiatus. Again, assuming that the base of the diamictites correlates not with C10n but C11n, the erosion associated with the base of the glacial sediments correlates with the base of Zone Oi2.

Birkenmajer [1987, and references therein] reported two "middle" Oligocene glacial units on King Georges Island, South Shetland Islands (Figure 1), although the precise ages of these units are not known. The tills of the Polonez Cove Formation and Legru Bay Group have been used to define climato-stratigraphic units, the Polonez and Legru Glaciations (Figure 6)[*Birkenmajer*, 1979]. The ages of these were originally believed to be Pliocene based on macroinvertebrates [*Birkenmajer*, 1979, 1982], but later radiometric measurements indicated Oligocene ages. Andesitic lavas at the base of the Legru Bay Group have been dated as 29.5±2.1 Ma while those at the top yielded ages of 25.7±1.3 Ma; this establishes minimum ages for the underlying sediments of the Wesele Cove Formation (type of the Wesele Interglacial), which overlies the Polonez Cove Formation ([summary by *Birkenmajer* [1987]). We speculate that the age of the Polonez Glaciation is equivalent to the $\delta^{18}O$ increase at circa 32.5-31.5 Ma., and that the Legru Glaciation correlates with the poorly known $\delta^{18}O$ increase at circa 28 Ma.

A large $\delta^{18}O$ increase occurred across the Oligocene/Miocene boundary (~23.5 Ma; the base of Zone Mi1; Table 4). *Birkenmajer* [1979, 1987] reported glaciomarine sediments of the Cape Melville Formation on King George Island (his Melville Glaciation; Figure 6). Basalts from the underlying formation have been dated as 23.6±0.7 Ma, while the Cape Melville Formation is intruded by andesitic dikes of 20 Ma [*Birkenmajer*, 1987]. We speculate that the Melville Glaciation is equivalent to the $\delta^{18}O$ increase at circa 23.5 Ma.

An uppermost Oligocene to lower Miocene glaciomarine section has been reported at Site 270 in the Ross Sea [*Hayes et al.*, 1975; *Allis et al.*, 1975; *Barrett*, 1975; *Leckie and Webb*, 1986](Figures 1 and 6). Upper Oligocene to lower Miocene diamictites overlay a glauconitic sand radiometrically dated as 26 Ma ± 2 m.y. [*McDougall*, 1977]. The paleomagnetic polarity patterns suggest that sediments from ~50 to 230 msb are Chronozones C5B (~15.0 Ma) to C6 (20.5 Ma; lower Miocene)[*Allis et al.*, 1975]; ages from 230 msb (20.5 Ma) to 365 msb (26 Ma; the level of the glauconite) are obtained by interpolation. This broad age control prevents firm comparison with the oxygen isotope record, although this unit partially may correlate with Zone Mi1 and two younger zones identified by *Wright and Miller* [1990]: Zones Mi1a (21.7 Ma) and Mi1b (18.2 Ma).

Isotope workers generally agree that there was a buildup of the Antarctic ice sheet in the middle Miocene. We suggest that there were seven intervals of early to middle Miocene ice growth (associated with the bases of Zones Mi1-Mi5, this study; Zones Mi1a and Mi1b; *Wright and Miller* [1990]). However, there is a poor record of lower to middle Miocene glacial sediments [*Tucholke et al.*, 1976] other than that at Site 270 [*Hayes et al.*, 1975]. Given the difficulties in establishing equivalency between glacial sedimentation and ice growth inferred from Miocene $\delta^{18}O$ increases, we regard the Oligocene correlations (Figure 6) as remarkable. Still, we acknowledge that better age resolution for Antarctic glacial sediments is required to establish their correlation to the $\delta^{18}O$ increases in the "middle" and latest Oligocene.

We caution that interpretations of glaciomarine sediments found near the Antarctic continent are debatable, because they may represent calving of an ice sheet or a mountain glacier. For example, the results from the west Antarctic peninsula (King Georges Island; *Birkenmajer* [1987, and references therein]) are areally restricted tills which could be explained by mountain glaciation. Similarly, diamictites in the MSST and CIROS-1 boreholes [*Barrett*, 1986; *Barrett et al.*, 1987, 1988, 1989] could be explained by their proximity to the Trans-Antarctic Mountains which were forming at this time [*Barrett et al.*, 1989]. Still, we note that periods of ice growth inferred from benthic and planktonic $\delta^{18}O$ covariance apparently correlate with intervals of extensive glaciomarine sedimentation (Figure 6). Widespread deposition of massive diamictites in the early Oligocene (Prydz Bay Sites 739 and 742; Figure 6; *Barron et al.* [1989]) argues for extensive continental ice sheets. Similarly, uppermost Oligocene to lower Miocene diamictites blanket the Ross Sea (e.g., Site 270; Figure 6; *Hayes et al.* [1975]; P. J. Barrett, personal communication, 1990), arguing for large ice sheets. The massive diamictites found at MSST and CIROS-1 (Figure 6) can be traced over 200 km into Victoria Land [*Barrett*, 1986; *Barrett et al.*, 1988; P. J. Barrett, personal communication, 1990], suggesting extensive glaciation. Finally, the earliest Oligocene $\delta^{18}O$ increase has been linked directly with a pulse of widespread glaciomarine sediments that not only influenced the Ross Sea, Weddell Sea, and Prydz Bay, but also the central Kerguelen Plateau. Such a widespread distribution suggests that the Oligocene was part of the ice house world.

Oxygen Isotopes and Glacioeustasy

We have appropriate western equatorial planktonic $\delta^{18}O$ data for only three of the increases (Mi2, Mi3, and possibly Mi5) used to define our zones. The lack of suitable low-latitude planktonic $\delta^{18}O$ records severely inhibits our ability to estimate sea level changes using $\delta^{18}O$ records. We note covariance between the benthic and subtropical planktonic $\delta^{18}O$ records for the Oi1, Oi2, and Mi1 events; still, we require western equatorial planktonic records to establish these events as ice volume increases.

The Oi1 event. The $\delta^{18}O$ increase which culminates at the base of Zone Oi1 occurs in benthic foraminifera in every ocean (Table 2); at Sites 522 and 748 this increase is correlated with earliest Oligocene Chron C13n [*Oberhansli et al.*, 1984; *Miller et al.*, 1988; *Zachos et al.*, 1990]. High-latitude planktonic foraminifera (e.g., Sites 592 and 593 [*Murphy and Kennett*, 1986]) recorded the increase, as did middle latitude locations such as Site 522 (fig. 4 by *Miller et al.* [1987]; data after *Oberhansli et al.* [1984] and *Poore and Matthews* [1984]). Although the appropriate low-latitude records are lacking, the link of the IRD at Site 748 with the $\delta^{18}O$ increase establishes that this was a major ice growth event [*Zachos et al.*, 1990]. The maximum benthic foraminiferal $\delta^{18}O$ values attained at Pacific sites (~ 1.8 ‰; Site 77, *Keigwin and Keller*, 1984) are lower than those in the Atlantic (> 2.0 ‰; Site 563, *Miller and Fairbanks* [1985]; Sites 522, 529; Figures 2 and 3) or Southern Oceans (> 2.0 ‰; Site 689; *Kennett and Stott* [1990]). This suggests that Atlantic and Southern Ocean deep waters were 1°C colder or saltier during the earliest Oligocene. The highest Pacific Oligocene $\delta^{18}O$ values were recorded during the Oi2 event; in contrast, Atlantic locations attain approximately equal $\delta^{18}O$ values at the Oi1 and Oi2 maxima. This results in distinctly different Oligocene $\delta^{18}O$ records in the Atlantic versus the Pacific.

The amplitude of the Oi1 event is typically 1.0 ‰ in benthic foraminifera (*Keigwin and Corliss* [1986]; *Vergnaud-Grazzini and Oberhansli* [1986], this study); although planktonic $\delta^{18}O$ data are limited (e.g, Sites 522, 592, 593), the increase is generally also nearly 1.0 ‰ in planktonic foraminifera. The sole exception is at Philippine Sea Site 292 where the planktonic increase is only 0.3 ‰; however, this record is limited by discontinuous coring [*Keigwin*, 1980]. Thus, using the Pleistocene $\delta^{18}O$/sea level calibration (0.11 ‰/10 m; *Fairbanks and Matthews* [1978]), the Oi1 increase may represent as much as 90 m of glacioeustatic lowering (assuming 1.0 ‰ of change in seawater $\delta^{18}O$) or as little as 30 m (assuming only 0.3 ‰ of change in seawater $\delta^{18}O$). Considering the maximum $\delta^{18}O$ of freezing ice, an outside limit for pre-Pleistocene $\delta^{18}O$/sea level calibration was estimated to be 0.055 ‰/10 m [*Miller et al.*, 1987]. If this calibration is used, the $\delta^{18}O$ may represent from 55 to 180 m of glacioeustatic lowering. These high amplitudes provide an extreme upper limit, and actual glacioeustatic falls were probably closer to those estimated using the Pleistocene calibration. Still, this wide range in possible sea level values illustrates the uncertainties in estimating pre-Pleistocene glacioeustatic amplitudes.

The Oi2 event. As noted above, the timing of the $\delta^{18}O$ increase associated with the base of Zone Oi2 is still poorly known. Our best estimate is that it appears to be linked with upper Chronozone C11r or lowermost C11n at Site 529 (Figure 3). We revise the age estimate of this event from circa 31 Ma [*Miller et al.*, 1987] to circa 32.5-31.5 Ma (Table 3). The event has been reported in benthic foraminiferal $\delta^{18}O$ records from five Atlantic and Pacific locations (Table 3). No western equatorial planktonic records are available for this interval, although a similar increase occurred at Indian Ocean Site 253 (fig. 4 by *Miller et al.* [1987]; ~33° S paleolatitude; data after *Oberhansli* [1986]; reconstruction of *Sclater et al.* [1985]). The glacial sediment record (Figure 6) is consistent with this being an ice growth event. We require better correlations and isotope data to confirm that this $\delta^{18}O$ increase resulted from ice growth.

Benthic foraminiferal records indicate that another $\delta^{18}O$ increase occurred in the upper part of Zone Oi2, dividing it into a lower Zone Oi2 and an upper Zone Oi2a. At Site 529, the upper increase occurs near the top of Zone P21b (circa 28 Ma; Figure 3). At Site 522, a single high $\delta^{18}O$ value occurs near the base of Chronozone C8r (circa 28 Ma; Figure 2), correlating with the 28 Ma ("Oi2a") increase at Site 529. The poor representation of this event at Site 522 may be due to dissolution at the critical levels [*Miller et al.*, 1988]; the $\delta^{18}O$ increase at the base of Zone Oi2 is also represented by a single data point at Site 522 (Figure 2). Despite this, records at Sites 522 and 529 both indicate a 28 Ma age for this interval of high $\delta^{18}O$ values. However, since the data (particularly at Site 522) are sparse and relegated to benthic foraminifera, we have not formally subdivided Zone Oi2.

The Mi1 event. A benthic foraminiferal $\delta^{18}O$ increase occurs near the top of the Oligocene at six Atlantic locations, with maximum values (= base of Zone Mi1) attained at the base of the Miocene (Figure 7; Table 4). A similar uppermost Oligocene $\delta^{18}O$ increase occurs in surface-dwelling planktonic foraminifera at subtropical Sites 558 and 563 (Figure 6)[*Miller and Fairbanks*, 1985]. Although western equatorial planktonic records are not available, it seems reasonably certain that the concomitant increase in benthic and subtropical planktonic $\delta^{18}O$ records reflects a major increase in ice volume. This was a relatively rapid (~1.0 m.y.), transient event of large magnitude (~1.0 ‰). Using the Pleistocene $\delta^{18}O$/sea level calibration suggests that 90 m of glacioeustatic lowering occurred between approximately 24.5 Ma and 23.5 Ma.

This event appears to be rapid on the million year scale. Still, many "Milankovitch" bandwidth (20 kyr-400 kyr) glacioeustatic changes may be embedded in this approximately 1-m.y. cycle. This event is not an artifact of signal aliasing of a high-frequency signal [*Pisias and Mix*, 1988], because it is

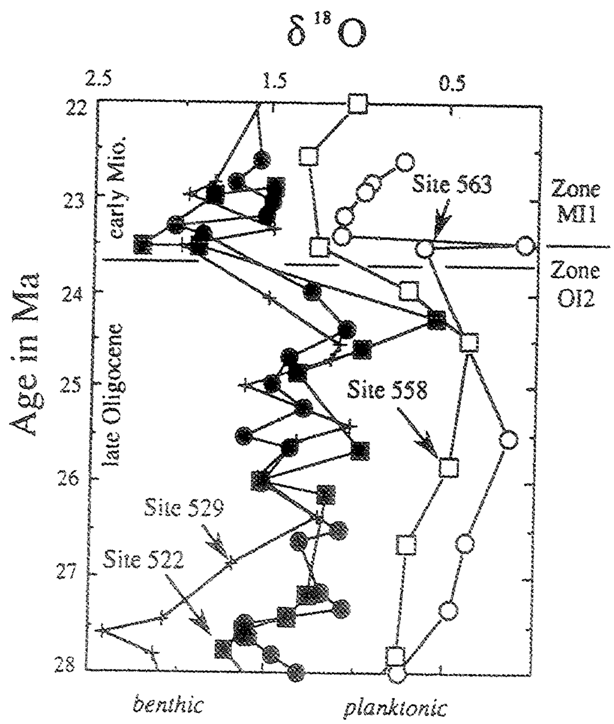

Fig. 7. The Mi1 stable isotope event, circa 24-23 Ma showing benthic foraminiferal (solid) and planktonic (open) foraminiferal $\delta^{18}O$ records. Data from *Miller and Fairbanks* [1985] (Site 563 benthic and planktonic records, represented by solid and open circles, respectively, and Site 558 planktonics, represented by open squares), *Miller et al.* [1988] (Site 522 benthic record, solid squares), and this study (Site 529 benthic record, pluses).

observed at six different locations with different sampling intervals and sedimentation rates. However, the relationship between million year isotope fluctuations and astronomical climate changes is not known, and signal aliasing of a Milankovitch signal may change the amplitude and character of the million year signal [*Pisias and Mix*, 1988]. Nevertheless, our stratigraphic control establishes that these million year $\delta^{18}O$ fluctuations are coeval events among ocean basins.

We revise the chronology of the Mi1 event slightly from *Miller and Fairbanks* [1985]. Based on our studies of Site 522, we realized that the increase occurred not in earliest Chron C6Cn as suggested by the magnetostratigraphy at Sites 563 and 558 [*Miller et al.*, 1985b], but in the short reversed polarity interval between Chrons C6Cn2 and C6Cn3 (23.5 Ma; Table 4)[*Miller et al.*, 1988]. This implies that Chron C6Cn1 or C6Cn2 are concatenated at Sites 558 and 563, consistent with the polarity records of *Miller et al.* [1985*b*]. The $\delta^{18}O$ increase is associated with the first occurrence of *Globorotalia kugleri* at Sites 366, 563, and 667 [*Miller et al.*, 1989], a planktonic foraminifera used to correlate the Oligocene/Miocene boundary [*Berggren et al.*, 1985]. This suggests that the base of Zone Mi1 may be used to approximate (but not precisely locate; Figure 7) the Oligocene/Miocene boundary on the basis of oxygen isotopes [*Miller et al.*, 1989].

The Mi2 event. The $\delta^{18}O$ increase which culminated at the base of Zone Mi2 has not been previously recognized as a global event. We establish that this increase was recorded by benthic foraminifera at numerous locations (Table 7). The increase is synchronous between Sites 563 and 608 within the resolution of magnetostratigraphy (Figure 4), spanning the boundaries between Chronozones C5Cn and C5Br (~16.5-16.0 Ma). This event is smaller than the Oi1, Oi2, and Mi1 events, because the amplitude is only ~0.6-0.8‰. This event is recorded in planktonic $\delta^{18}O$ at western equatorial Indian Ocean Site 237 (Figure 8)[*Vincent et al.*, 1985], where the amplitude and timing are remarkably similar to the global benthic $\delta^{18}O$ record. Such a mimicry of the deepwater $\delta^{18}O$ signal by western equatorial surface waters can most reasonably be ascribed to changes in global seawater composition due to ice growth. Using the Pleistocene calibration, we estimate that sea level was glacioeustatically lowered by ~50-75 m between circa 16.5 and 16.0 Ma.

The Mi3 event. The Mi3 and Mi4 events constitute the classic [*Savin et al.*, 1975; *Shackleton and Kennett*, 1975] middle Miocene $\delta^{18}O$ increase. It has not been generally recognized that this increase occurred as two steps (Figures 4 and 8). The first step forms the basis for Zone Mi3 (Figure 4); the increase apparently begins as low as the base of Chronozone C5AD (circa 14.9 Ma) at both Sites 563 and 608, reaching peak values (= base of Zone Mi3) in Chronozone C5ABr (circa 13.6 Ma; Figure 4, Table 8). As with the previous event, planktonic $\delta^{18}O$ records from western equatorial Indian Ocean Site 237 show a remarkably similar increase (Figure 8)[*Vincent et al.*, 1985]. Both records show increases of between 0.5 ‰ and 0.8 ‰ (Figure 8), suggesting glacioeustatic lowering of ~45-70 m (using the calibration of *Fairbanks and Matthews* [1978]).

Resolution of this increase and the subsequent Mi4 event are limited by sampling coverage. It is not clear from the benthic records if the $\delta^{18}O$ increase began in Chron C5AD. There may actually be three middle Miocene $\delta^{18}O$ increases: one in Chron C5AD (unresolved at present), one in C5AC (Zone Mi3), and one in C5AA-C5Ar (Zone Mi4)(Figure 8). Resolution of the planktonic record at Site 237 is also limited by slow sedimentation rates and drilling disturbance, and the record of *Vincent et al.* [1985] provides the maximum sample coverage possible at this site.

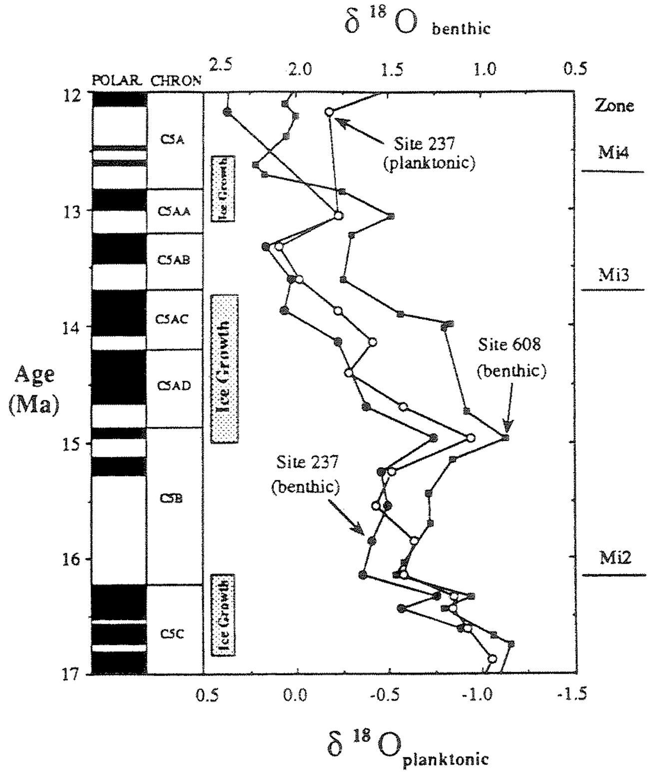

Fig. 8. Late early to middle Miocene oxygen isotope comparison of the Site 608 benthic foraminiferal $\delta^{18}O$ standard (solid circles) with benthic (solid squares) and planktonic (open squares) $\delta^{18}O$ records from Indian Ocean Site 237. Note different scales for benthic and planktonic $\delta^{18}O$ records, and the similarities among the records including increases from circa 16.5-16.0 Ma and circa 14.5-12.5 Ma.

The Mi4 event. This is the second event of the two-step middle Miocene $\delta^{18}O$ increase. The increase occurs across the Chronozone C5AAn/C5An transition at Sites 563 and 608, with peak values observed in Chronozone C5A (circa 12.6 Ma; Figure 4). Benthic records at both Sites 563 and 608 exhibit increases of 0.7‰. No suitable low-latitude planktonic $\delta^{18}O$ data are available that show this increase. The interpretation of the Mi4 event is equivocal, because our strategy requires both benthic and low-latitude planktonic foraminiferal $\delta^{18}O$ records.

The Mi5 and Mi6 events. At both Sites 608 and 563, benthic foraminiferal $\delta^{18}O$ values increase across the C5An/C5r transition and in the lower part of C5n (Figure 4). Maximum values occur near the base of C5r (= base of Zone Mi5; circa 11.3 Ma) and near the middle of C5n (= base of Zone Mi6; circa 9.6 Ma). The amplitudes of both increases are 0.35 to 0.60 ‰ at Sites 608 and 563, respectively. A large (~1.0 ‰) increase in low-latitude planktonic foraminiferal $\delta^{18}O$ values occurs at Site 237 in the interval encompassing Zones Mi5 and Mi6. Based on the biostratigraphy of *Vincent* [1977], it appears that the planktonic increase at Site 237 correlates with the benthic foraminiferal $\delta^{18}O$ increase associated with Zone Mi5 (circa 11.5-11.0 Ma). Biostratigraphic correlations of this interval are particularly difficult, and there may be an unconformity near the level of the planktonic $\delta^{18}O$ increase. Use of nannofossil biostratigraphy suggests that the planktonic $\delta^{18}O$ at Site 237 correlates not with Zone Mi5, but with Zone Mi6 [*Vincent et al.*, 1985]. If the foraminiferal correlations are correct, a glacioeustatic lowering of at least 30-45 m occurred between 11.5 and 11.0 Ma (= base of Zone Mi5). No suitable low-latitude planktonic $\delta^{18}O$ records are available for the Zone Mi6 (circa 9.6 Ma), and it is not clear if this event is due to a temperature decrease, increase in ice volume, or both.

Comparison With Passive Margin Records

Global sea level (eustatic) lowerings cause the development of unconformities on passive continental margins and epicontinental seas [e.g., *Haq et al.*, 1977; *Christie-Blick et al.*, 1990]. *Pitman* [1978] pointed out that the rate of global sea level change controls development of passive margin unconformities rather than the absolute position of sea level. Depositional models [e.g., *Posamentier and Vail*, 1988; *Christie-Blick et al.*, 1990] have relied on the assumption that unconformities develop during the most rapid eustatic falls. This may not be strictly true, because there may be some lead/lag between the time of most rapid eustatic fall and the development of unconformities (*N. Christie-Blick*, unpublished manuscript, 1990). Still, the times of most rapid eustatic lowering should approximately coincide with the lowering of baselevel and the erosion of unconformities [*Christie-Blick et al.*, 1990].

Unconformities (= sequence boundaries) may be detected using seismic stratigraphy [*Vail et al.*, 1977] and physical stratigraphy, the combination of which has been designated sequence stratigraphy [*Haq et al.*, 1987]. Unconformities are also associated with hiatuses. Aubry [1990] presents a detailed discussion of the detection of hiatuses using integrated magnetostratigraphy and biostratigraphy. The coincidence of hiatuses must be documented both within and between basins in order to establish a global cause such as eustatic change.

The challenge to stratigraphers is to document the ages of the unconformities and their associated hiatuses in regions which are most sensitive to eustatic change. This is a difficult task. Neritic sections are the most sensitive to sea level change, but they often lack good biostratigraphic or magnetostratigraphic control (see discussion by *Miller and Kent* [1987]). We have documented that continental slope records often contain unconformities and hiatuses which may be related to eustatic changes [e.g., *Miller et al.*, 1985*a*, *c*; *Poag et al.*, 1985]. However, the relationship between sea level and continental slope erosion is not direct and these slope sections cannot be used as unequivocal evidence for global sea level changes [*Miller et al.*, 1990*a*]. We must focus on delineating hiatuses in continental shelf (neritic) and epicontinental sea sections which are directly affected by eustatic change.

The publicly available magnetobiostratigraphic data base for neritic sections is inadequate for testing the ages of many of the Cenozoic sequence boundaries. Aubry's [1990] study provides a detailed example of the inadequacy of data for the lower/middle Eocene sequence boundary. The ages of most other Cenozoic sequence boundaries are not as well constrained. Ongoing studies of Oligocene-Miocene outcrops and boreholes in the Atlantic and Gulf coastal plains offer the potential for improving age estimates of the sequence boundaries [*Poore and Bybell*, 1988; *Miller et al.*, 1990*a*; and K. G. Miller, D. V. Kent, and P. R. Thompson, unpublished manuscript, 1990]. Still, it is clear that further drilling of these and other margins is required to obtain appropriate material.

We noted that there is a correspondence between the timing of erosional events on passive margins and the oxygen isotope record [*Miller et al.*, 1985*c*, 1987]. One notable exception was the poor correspondence between the middle Miocene oxygen isotope record and the compilation of *Vail et al.* [1977]. The cycle chart of *Haq et al.* [1987] contains many more sequence boundaries than *Vail et al.* [1977], particularly in the middle Miocene. *Haq et al.* [1987] provide detailed age estimates for 11 Oligocene to Miocene type-1 sequence boundaries (Table 10). If there is a causal relationship between ice volume and unconformities, then sequence boundaries should correlate with the inflection points of $\delta^{18}O$ increases in benthic and low-latitude plank-tonic foraminifera. Note that since our oxygen isotope zones are defined on the maximum $\delta^{18}O$ value, there should be a slight lag of the base of the zone from its associated sequence boundary.

We compare the ages of the oxygen isotope increases (inferred glacioeustatic lowerings) with the ages of the *Haq et al.* [1987] sequence boundaries (Table 10) and find good agreement between the two records. Unfortunately, the biostratigraphic data base for most of the sequence boundaries is not available, and we cannot evaluate fully discrepancies between the two records. Still, using the available public data on the ages of these sequence boundaries, we can make the following evaluations.

Oi1 event. The benthic and subtropical planktonic foraminiferal $\delta^{18}O$ increase associated with this zone correlates well with the 36.0 Ma type-1 sequence boundary of *Haq et al.* [1987]. We have independently observed that an erosional event occurs near the base of Magnetochron C13n (circa 35.9 Ma) in Alabama (Bay Minette and St. Stephens Quarry boreholes; K. G. Miller, D. V. Kent, and P. R. Thompson, unpublished manuscript, 1990). Due to concatenation of this and the "middle" Oligocene sea level lowerings, the lower Oligocene is poorly represented in most neritic sections.

Oi2 event. We establish that a 0.8-1.0 ‰ $\delta^{18}O$ increase occurred in benthic and subtropical planktonic foraminifera during Chron C11 (circa 32.5-31.5 Ma), while the major sequence boundary in this interval is estimated as 30.0 Ma. However, the "middle" Oligocene is very difficult to correlate biostratigraphically, and published age estimates of the oxygen isotope and sequence boundary have varied by 1-2 m.y. (e.g., the sequence boundary was reported as 29 Ma by *Vail et al.* [1977] and 30.0 Ma by *Haq et al.* [1987]). The difference in ages between the oxygen isotope event and the sequence boundary may be due to a biostratigraphic correlation problem with the *Haq et al.* [1987] record. In support of this, we note that an unconformity at the base of the Chickasawhay Limestone in Alabama (= the 30.0 Ma sequence boundary of *Haq et al.* [1987]; see *Baum* [1986] for details) can be correlated with all of Chron C12n and part of Chron C11r (~32.9-32.1 Ma) [K. G. Miller, D. V. Kent, and P. R. Thompson, unpublished manuscript, 1990].

The second "middle" Oligocene benthic foraminiferal $\delta^{18}O$ increase ("Oi2a"; circa 28 Ma) correlates well with the 28.4 Ma sequence boundary of *Haq et al.* [1987]. We have no isotope evidence for a third "middle" Oligocene $\delta^{18}O$ increase, although *Haq et al.* [1987] report a type-1 sequence boundary at 26.5 Ma. This remains an unsolved problem.

Mi1 event. The distinct Mi1 $\delta^{18}O$ increase began in both benthic and subtropical planktonic foraminifera near the end of the Oligocene. *Haq et al.* [1987] similarly report a type-1 sequence boundary in late Zone P22 (i.e., near the end of the Oligocene)(Table 10). The differences in estimated ages (23.5 versus 25.5 Ma) between these two events result from differences in time scales, because *Haq et al.* [1987] estimate the Oligocene/Miocene boundary as 25.2 Ma, while *Berggren et al.* [1985] report it as 23.7 Ma.

Haq et al. [1987] report a major type-1 sequence boundary at 21.0 Ma (their "mid-Burdigalian" event). *Wright and Miller* [1990] document a benthic foraminiferal $\delta^{18}O$ increase at Site 747 that correlates directly to Chronozone C6Ar (age estimate 21.7 Ma). There are coeval increases at Sites 563 (283 msb) and 608 (380 msb)(Figure 4). *Wright and Miller* [1990] defined the base of Zone Mi1a on this benthic foraminiferal $\delta^{18}O$ increase at Site 563 (Figure 4). This increase correlates to Haq et al.'s 21.0-Ma sequence boundary. Suitable low-latitude

TABLE 10. Comparison Between the Timing of Sequence Boundaries and Oxygen Isotope Zones.

Sequence Boundaries			Oxygen Isotope Zonal Boundaries				
Age	Magnetochron	Foraminiferal Zone	Isotope Zone	Age	Magnetochron	Foraminiferal Zone	Comments
8.2 Ma*	C4A	N16	Mi7	8.5 Ma	base C4n	(N16)	*Wright and Miller* [1990]
10.5 Ma+	C5n base	N14	Mi 6	9.6 Ma	lower C5n	N16	time scale problem
12.5 Ma+	C5AA	N12	Mi 5	11.3 Ma	base C5r	N13-N14	time scale problem
13.8 Ma+	C5ACn	N11	Mi 4	12.6 Ma	base C5Ar	(N11-N12)	minor correlation problem
15.5 Ma+	C5Br	mid-N8	Mi 3	13.6 Ma	C5ABr	(N11)	major correlation problem
16.5 Ma+	mid-C5Cn	base N8	Mi 2	16.1 Ma	C5Br	N8	excellent correlation
17.5 Ma*	C5Dn	N7	Mi1b	18.1 Ma	C6Dr	N6	*Wright and Miller* [1990]
21.0 Ma+	top C6An1	mid-N5	Mi1a	21.2 Ma	C6Ar	top N4b	*Wright and Miller* [1990]
22.0 Ma*	C6AAn	lower N5	no evidence				
25.5 Ma+	C6Cr	top P22	Mi 1	23.5 Ma	C6Cn	(N4a)	time scale problem
26.5 Ma+	C7An	mid-P22	no evidence				
28.4 Ma+	C9n	upper P21	"Oi 2b"	28.0 Ma	(top C9n)	top P21b	good agreement on timing
30.0 Ma+	C10n	mid-P21	Oi 2	32.2 Ma	C11r	upper P20	major correlation problem
33.0 Ma*	base C12n	P19/20	no evidence				
36.0 Ma+	C13n	P18¥	Oi 1	35.8 Ma	C13n	lower P18	excellent correlation

Sequences taken from *Haq et al.* [1987].
+ Type 1 sequence boundaries.
* Type 2 sequence boundaries.
¥ Using the zonal criteria of Berggren and Miller [1988].
Oxygen Isotopes: Zones, age estimates, correlations to magnetochrons, and correlations to planktonic foraminiferal zones are from Tables 2-9.
Second-order correlations are indicated in parentheses.

planktonic data are not available at present, and we caution that the cause of this increase is not yet known. *Wright and Miller* [1990] also note a benthic foraminiferal $\delta^{18}O$ increase at ~18.2 Ma at Sites 563, 608 (Figure 4), and 747 which they used to define Zone Mi1b. This age estimate is similar to that of a minor (type 2) sequence boundary of *Haq et al.* [1987]. Again, suitable low-latitude planktonic data are lacking for the Mi1b increase, and its interpretation is equivocal.

Mi2 to Mi6 events. In the 6 m.y. of the middle to early late Miocene, we report five oxygen isotope events (= inferred glacioeustatic lowerings), three of which have been observed in western equatorial planktonic records. *Haq et al.* [1987] similarly report five type-1 sequence boundaries in this interval. There are differences in age estimates of the oxygen isotope events versus the sequence boundaries, with differences ranging from 0.4 m.y. to 1.9 m.y. We ascribe these differences to biostratigraphic and time scale problems. The difference between the *Berggren et al.* [1985] and *Haq et al.* [1987] time scales for this interval has been dealt with by *Gradstein et al.* [1988]. We cannot evaluate the biostratigraphic control on the ages of these middle Miocene sequence boundaries because the only biostratigraphic data provided for this interval by Exxon is that in *Greenlee and Moore* [1988]. Their data are clearly insufficient to warrant the precision of the age estimates provided by *Haq et al.* [1987]. Despite the differences in age estimates, it is clear that both passive margins and benthic oxygen isotopes recorded five events during the middle Miocene.

Wright and Miller [1990] report one additional $\delta^{18}O$ increase in the late Miocene, associated with Zone Mi7 (Figure 4; circa 8.5 Ma). This event apparently correlates with the 8.2-Ma type-2 sequence boundary of *Haq et al.* [1987]. Low-latitude planktonic data are lacking for this $\delta^{18}O$ increase, and its interpretation as an ice volume increase is equivocal.

This summary establishes that there are some remarkable similarities between the Cenozoic oxygen isotope record and the sequence stratigraphic record of *Haq et al.* [1987](Table 10). However, further work is needed not only to seek undetected oxygen isotope events but also to confirm and correlate the ages of sequence boundaries on passive margins.

CONCLUSIONS

We have not unlocked the secrets of the ice house. The timing and nature of Oligocene-Miocene $\delta^{18}O$ fluctuations and major sequence boundaries require further documentation. However, comparisons of the $\delta^{18}O$ and passive margin records have improved since *Vail et al.* [1977]. In 1982, Miller questioned the "Vail record" for its identification of a "middle" Oligocene eustatic fall (manuscript written in 1982, published as *Miller et al.* [1985a]). A "middle" Oligocene $\delta^{18}O$ increase was subsequently discovered (= Zone Oi2) [*Keigwin and Keller*, 1984; *Miller and Thomas*, 1985], and *Miller et al.* [1985c] noted a good correspondence between the "middle" Oligocene $\delta^{18}O$ record and passive margin erosion. In 1987, Miller et al. suggested that five Oligocene-Miocene $\delta^{18}O$ increases were correlated to passive margin erosion events. In this contribution we (1) identify and date 12 Oligocene to Miocene oxygen isotope increases; (2) note that six of the 12 $\delta^{18}O$ increases also occur in tropical or subtropical planktonic foraminifera (the other six lack suitable data); using Pleistocene strategies, these increases may best be attributed to growth of ice sheets; (3) correlate these $\delta^{18}O$ increases with sea level changes inferred from continental margin stratigraphy [e.g., *Haq et al.*, 1987] and ice growth events inferred from glaciomarine sediments, establishing a mechanism which links all three records. *Haq et al.* [1987] report 11 Oligocene-early late Miocene type-1 sequence boundaries. Of the 12 $\delta^{18}O$ increases, 10 correspond with type-1 sequence boundaries, and two correspond with a type-2 sequence boundaries. Of the type-1 sequence boundaries, only the 26.5-Ma boundary lacks a corresponding $\delta^{18}O$ increase. We may not have unlocked the ice house, but we are knocking at the door.

Three types of data are needed to evaluate the ice house link between $\delta^{18}O$ and sequence boundaries. First, we need to document sequence boundaries on margins with different subsidence histories and to establish firmly the timing of these events (see Aubry [1990] for examples). Second, the physical record of glacial advances and retreats must be better documented. Third, we need to characterize fully the oxygen isotope record. This task is not complete even for the 10^6-year scale, because we lack the critical planktonic foraminiferal records; in addition, we need to consider the "climate" scale (10^4-10^5 year). It can be argued that sedimentation on most margins responds to sea level changes on the million year scale, because higher-frequency sea level events are not detectable. Still, there are many settings with expanded sections which record sea level variations on the 10^4-10^5 year scale, and these records must be compared with the higher-order $\delta^{18}O$ record.

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