

OLIGOCENE TO MIOCENE CARBON ISOTOPE CYCLES AND ABYSSAL CIRCULATION CHANGES

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Abstract. Three cycles of $\delta^{13}\text{C}$ occurred in Oligocene to Miocene benthic and planktonic foraminifera at western North Atlantic Sites 558 and 563. Intervals of high $\delta^{13}\text{C}$ occurred at about 35-33 Ma (early Oligocene), 25-22 Ma (across the Oligocene/Miocene boundary), and 18-14 Ma (across the early/middle Miocene boundary). Similar carbon isotopic fluctuations have been measured in benthic and planktonic foraminifera from the Atlantic, Pacific, and Indian oceans, suggesting that these cycles represent global changes in the $\delta^{13}\text{C}$ of mean ocean water. The average duration of the carbon cycles is 50 times greater than the residence time of carbon in the oceans. Therefore, the mechanism controlling these cycles must be tied to changes in the input ratio of organic carbon to carbonate from weathering rocks or to changes in the output ratio of organic carbon to carbonate in marine sediments. Following a strategy used to study modern and Pleistocene oceans, benthic foraminiferal $\delta^{13}\text{C}$ differences between the Atlantic and Pacific are used to infer Oligocene through Miocene abyssal circulation changes. The Atlantic was most enriched in ^{13}C relative to the Pacific from about 36-33 Ma (early Oligocene) and 26-10 Ma (late Oligocene to late Miocene). We interpret this as indicating supply of nutrient-depleted bottom water in the North Atlantic, perhaps analogous to modern North Atlantic Deep Water. High benthic foraminiferal $\delta^{13}\text{C}$ values at about 36-35 Ma, 31-28 Ma, 25-24 Ma, and younger than 15 Ma indicate the presence of ice sheets at these times. Covariance between benthic and planktonic foraminiferal $\delta^{18}\text{O}$ records of 0.3-0.5‰/‰ at 36 Ma, 31 Ma, and 25 Ma suggests that three periods of continental glaciation caused eustatic (global sea-level) lowerings of 30-50 m during the Oligocene epoch. The $\delta^{13}\text{C}$ cycles do not correlate with sea-level changes deduced from oxygen isotopic data, nor do they correlate with other proxy indicators for sea level.

Introduction

Stable isotopic studies of the shells of microfossil foraminifera can provide information

on changes in global carbon reservoirs, abyssal circulation, ice volume, and temperature. For example, synchronous planktonic and benthic foraminiferal $\delta^{13}\text{C}$ changes observed in different ocean basins and oceanographic settings provide evidence for changes in the $\delta^{13}\text{C}$ of mean ocean water. Deep-sea benthic foraminiferal $\delta^{13}\text{C}$ records can also provide a monitor of bottom-water $\delta^{13}\text{C}$ conditions, allowing reconstruction of Atlantic and Pacific carbon isotopic differences caused by abyssal circulation. Oligocene (36.6 to 23.7 Megannus (Ma) = 10^6 years B.P.) to Miocene (23.7 to 5.3 Ma) North Atlantic isotopic records with good chronostratigraphic control were not available prior to drilling of Deep Sea Drilling Project (DSDP) Sites 558 and 563. In this study, we measured benthic and planktonic foraminiferal $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ from western North Atlantic Sites 558 and 563 in order to compare the Atlantic data with published Pacific isotopic records. Oxygen isotopic records are used as a proxy indicator of sea level, which we compared with the oceanic carbon cycles and changes in abyssal circulation deduced from $\delta^{13}\text{C}$ studies.

Previous studies of Tertiary carbon isotopes have been hampered by the fact that most existing data are based upon analyses of benthic assemblages or taxa which do not accurately record modern deep-sea $\delta^{13}\text{C}$ distribution. Studies have shown that the benthic foraminifera Cibicidoides accurately reflects deep-water $\delta^{13}\text{C}$ variations [Graham et al., 1981; Belanger et al., 1981]. Fortunately, this taxon is abundant in the Tertiary fossil record recovered by the DSDP. Deep- and bottom-water $\delta^{13}\text{C}$ variations may be caused by (1) changes in global carbon reservoirs [Broecker, 1970, 1982; Shackleton, 1977]; (2) abyssal circulation changes [Bender and Keigwin, 1979; Belanger et al., 1981; Curry and Lohmann, 1982, 1983; Boyle and Keigwin, 1982; Shackleton et al., 1983]; (3) local productivity changes, especially on continental margins [Sarnthein et al., 1982]; and (4) changes in the preformed chemistry of the source region [Mix and Fairbanks, 1984].

Miller and Fairbanks [1983] first delineated three cycles in Oligocene to middle Miocene $\delta^{13}\text{C}$

TABLE 1. Oxygen and Carbon Isotopic Data, DSDP Sites 558 and 563

Core-Section/ Depth, cm	Depth, msb	Age, Ma	$\delta^{18}\text{O}_{\text{PDB}}$	$\delta^{13}\text{C}_{\text{PDB}}$
<u>Cibicidoides spp., Site 558</u>				
1-2/90-94	160.40	8.27	2.12	1.10
2-1/80-85	168.30	8.53	2.26	1.09
3-1/82-87	177.82	8.83	2.05	1.07
4-1/81-86	187.31	9.19	2.12	1.25
5-1/103-109	197.03	9.58	2.24	0.81
6-1/79-84	206.29	9.95	1.84	0.95
7cc	215.80	10.93	2.14a	1.09a
8-1/82-86	225.32	11.02	1.66a	0.69a
9-1/16-22	234.16	11.74	1.63a	0.96a
11-1/103-109	254.03	13.36	1.56a	1.06a
12-1/80-85	263.30	14.11	1.44a	1.35a
13-1/103-109	273.03	14.91	1.11a	1.21a
14-1/67-71	282.17	15.65	1.21a	1.43a
15-1/131-135	292.31	16.48	1.56	2.08
16-1/79-83	301.29	17.53	1.24	1.01
16-4/68-73	305.68	18.52	1.21	0.85
17-1/128-132	311.28	20.60	1.52	0.61
17-4/56-61	315.06	21.54	1.03	0.89
18-1/92-96	320.42	22.88	1.33	1.50
18-4/16-21	324.16	23.66	1.71	1.49
18-4/81-86	324.81	23.77	1.57	0.58
18-5/23-28	325.73	23.94	2.09	1.34
19-1/2-6	329.02	24.64	1.10	1.03
19-1/141-145	330.41	24.78	2.15	1.04
19-2/34-39	330.84	24.86	1.18	0.60
19-3/69-74	332.69	25.19	1.08	0.66
19-4/118-122	334.68	25.55	1.40	0.93
19-5/110-115	336.10	25.80	1.57	0.59
20-1/118-122	339.68	26.44	1.29	0.86
20-2/66-71	340.66	26.62	0.74	0.44
20-3/14-19	341.64	26.79	1.31	0.55
20-4/122-126	344.22	27.26	1.66	0.89
21-1/46-49	348.46	28.02	1.40	0.80
			1.09	0.39
21-1/132-136	349.32	28.17	1.71a	1.23a
21-2/102-108	350.52	28.39	1.85	0.65
21-3/13-16	351.13	28.50	1.79	0.86
22-1/34-38	357.84	29.70	1.77	0.46
22-2/84-88	359.84	30.03	2.05a	1.38a
22-3/21-26	360.71	30.13	1.60	0.61
22-4/3-8	362.03	30.28	1.84	0.62
22-4/78-81	362.78	30.37	1.77	0.69
23-1/52-57	367.52	30.89	1.65	0.49
23-2/51-55	369.01	31.06	2.03	0.52
23-3/65-69	370.65	31.24	1.30b	0.09b
23-4/49-53	371.99	31.39	1.36b	0.34b
23-5/22-26	373.22	31.53	1.65b	0.62b
23-6/45-49	374.95	31.72	1.82b	0.30b
24-2/80-85	378.80	32.13	1.10b	0.69b
24-3/79-84	380.29	32.25	1.29b	0.82b
24-4/75-79	381.75	32.38	1.77b	0.67b
25-1/117-121	387.17	32.83	1.29b	0.77b
25-4/85-89	391.35	33.63	1.59b	0.85b
26-1/93-97	396.43	34.73	1.13b	1.23b

TABLE 1. (continued)

Core-Section/ Depth, cm	Depth, msb	Age, Ma	$\delta^{18}\text{O}_{\text{PDB}}$	$\delta^{13}\text{C}_{\text{PDB}}$
<u>Cibicidoides spp., Site 558 (continued)</u>				
26-4/22-26	400.22	35.51	1.51b	1.13b
27-1/81-85	405.81	36.50	1.04b	1.48b
<u>Catapsydrax spp., Site 558</u>				
16-5/88-92	307.38	18.91	1.80	0.82
16-6/87-91	308.87	20.05	1.76	0.96
17-1/128-132	311.28	20.64	1.34	1.46
17-2/100-106	312.50	20.94	1.47	1.40
17-3/80-85	313.80	21.26	0.95	1.03
17-4/7-12	314.57	21.45	0.93	1.15
17-4/56-61	315.06	21.57	1.19	1.45
17-5/23-28	316.23	21.86	1.13	1.75
18-1/28-33	319.78	22.73	1.36	1.70
18-1/92-96	320.42	22.88	1.56	1.80
18-2/126-130	322.26	23.32	1.19	1.44
18-3/80-86	323.30	23.50	1.23	1.51
18-4/16-21	324.16	23.66	1.79	1.76
18-4/81-86	324.81	23.77	1.51	1.64
18-5/23-28	325.73	23.94	1.32	1.50
19-1/2-6	329.02	24.53	0.97	1.52
19-1/66-70	329.66	24.64	0.98	1.61
19-1/141-145	330.41	24.78	1.49	1.07
19-2/34-39	330.84	24.86	0.81	1.13
19-3/69-74	332.69	25.19	1.24	1.19
19-4/10-14	333.60	25.35	1.23	0.88
19-4/118-122	334.68	25.55	1.66	1.28
19-5/110-115	336.10	25.80	1.45	0.89
20-1/118-122	339.68	26.44	1.52	1.12
20-2/66-71	340.66	26.62	1.52	0.95
20-3/14-19	341.64	26.79	1.69	1.02
20-4/12-16	343.12	27.06	1.48	1.10
20-4/122-126	344.22	27.26	1.85	1.16
20-5/27-32	344.77	27.36	1.64	1.26
21-1/46-49	348.46	28.02	1.66	1.06
21-1/132-136	349.32	28.17	1.80	1.22
21-2/102-108	350.52	28.39	1.73	1.00
21-3/13-16	351.13	28.50	1.63	1.06
22-1/34-38	357.84	29.70	1.57	0.78
22-1/97-101	358.47	29.82	1.55	1.02
22-2/84-88	359.84	30.04	1.33	0.85
22-4/3-8	362.03	30.28	1.15	0.39
22-4/78-81	362.78	30.37	1.27	0.70
23-1/52-57	367.52	30.89	1.53	1.32
23-2/51-55	369.01	31.06	1.55	1.11
23-3/65-69	370.65	31.24	1.63	0.69
23-5/22-26	373.22	31.53	1.60	0.79
23-6/45-49	374.95	31.72	1.80	0.71
24-1/34-38	376.84	31.93	1.49	0.63
24-2/80-85	378.80	32.13	1.89	1.13
24-3/79-80	380.29	32.25	1.57	0.83
24-4/74-79	381.75	32.38	1.70	1.38
24cc	382.40	32.73	1.43	1.12
25-1/117-121	387.17	32.83	1.62	1.19
26-4/22-26	400.22	35.56	1.68	1.52

TABLE 1. (continued)

Core-Section/ Depth, cm	Depth, msb	Age, Ma	$\delta^{18}\text{O}_{\text{PDB}}$	$\delta^{13}\text{C}_{\text{PDB}}$
<u>Globorotalia opima nana, Site 558</u>				
18-1/92-96	320.42	22.88	1.02	1.73
18-4/16-21	324.16	23.66	1.29	1.81
19-1/141-145	330.41	24.78	1.22	0.87
19-3/69-74	332.29	25.12	0.72	1.14
19-4/118-122	334.68	25.55	0.38	0.93
20-2/66-71	340.66	26.62	0.47	0.69
20-4/122-126	344.22	27.26	0.71	0.84
21-1/132-136	349.32	28.17	0.76	0.69
21-2/102-108	350.52	28.39	0.85	0.79
22-1/34-38	357.84	29.70	1.14	0.87
22-4/3-8	362.03	30.28	1.36	0.84
23-1/52-57	367.52	30.89	1.40	0.95
24-4/75-79	381.75	32.38	1.25	1.22
25-1/117-121	387.17	32.83	1.05	1.20
25-4/85-89	391.35	33.63	1.06	0.98
<u>Globigerina anguliseturalis, Site 558</u>				
19-4/118-122	334.68	25.59	0.80	1.41
20-4/122-126	344.22	27.25	0.82	1.47
<u>Concentrated Crystals From Benthic Foraminifera, Site 558</u>				
26-4/22-26	400.22	35.35	1.89c	1.32c
<u>Cibicidoides spp., Site 563</u>				
1-1/119-125	157.69	8.06	2.27	1.29
1-3/18-24	159.68	8.29	2.28	1.03
2-1/33-38	166.33	8.73	2.30	1.05
2-4/34-39	170.84	9.08	2.41	1.04
3-1/35-40	175.85	9.47	2.40	1.23
3-4/36-41	180.36	9.82	2.23	1.25
4-1/37-42	185.37	10.22	2.03	0.84
4-4/30-36	189.80	10.58	1.61	0.94
5-2/130-136	197.30	11.25	2.24	1.44
5-5/130-136	201.80	11.66	2.25	1.36
6-2/130-136	206.80	12.10	2.24	0.87
6-5/121-127	211.21	12.50	1.91	1.07
7-2/131-136	216.31	12.95	2.19	1.15
7-5/110-116	220.60	13.33	2.12	1.10
8-2/114-120	225.64	13.77	2.27	1.12
8-5/114-120	230.14	14.17	2.14	1.19
8-1/21-26	232.71	14.40	2.17	1.70
9-2/114-120	235.14	14.62	1.96	1.72
9-3/104-109	236.54	14.74	0.97	1.01
9-4/109-114	238.09	14.88	1.26	1.00
9-5/114-120	239.64	15.02	1.28	1.32
9-6/113-118	241.13	15.15	1.32	1.21
10-2/90-96	244.40	15.48	1.53	1.95
10-3/105-113	246.05	15.67	1.15	1.72
10-4/95-101	247.45	15.82	1.05	1.82
10-5/93-98	248.93	15.99	1.64	2.07
10-5/110-116	249.10	16.01	1.22	1.43
10-6/90-95	250.40	16.15	1.44	1.77

TABLE 1. (continued)

Core-Section/ Depth, cm	Depth, msb	Age, Ma	$\delta^{18}\text{O}_{\text{PDB}}$	$\delta^{13}\text{C}_{\text{PDB}}$
<i>Cibicidoides</i> spp., Site 563 (continued)				
11-1/110-116	252.60	16.46	1.21	2.01
11-2/100-105	254.00	16.67	1.04	1.51
11-3/81-86	255.31	16.87	1.25	1.73
11-4/112-118	257.12	17.14	1.40	1.33
12-2/110-116	263.60	17.86	1.40	1.56
12-5/124-130	268.24	18.24	1.64	1.17
13-2/124-130	273.24	19.82	1.31	0.71
13-5/124-130	277.74	20.49	1.59	0.91
14-2/129-135	282.79	21.23	1.72	1.01
15-2/117-123	292.17	22.62	1.56	1.51
15-4/110-115	295.10	23.05	1.71	1.41
15-6/100-106	298.00	23.59	1.49	1.30
16-1/114-120	300.14	24.08	1.52	1.40
16-2/117-123	301.67	24.43	2.05	1.59
16-3/110-116	303.10	24.76	1.79	1.17
			1.84	1.30
			2.05	1.50
16-5/117-123	306.17	25.45	1.25	1.06
16-6/93-98	307.43	25.75	1.06	0.80
16-7/43-48	308.43	25.98	1.38	0.82
17-1/81-86	309.31	26.18	1.48	0.78
17-2/19-24	310.19	26.38	1.29	0.82
17-2/117-123	311.17	26.61	1.64a	1.34a
17-3/5-9	311.55	26.69	1.36	0.90
17-3/124-129	312.74	26.96	1.53	0.63
17-4/132-137	314.32	27.33	1.09	0.46
17-5/24-30	314.74	27.42	1.30	0.63
18-1/20-25	318.20	28.21	1.18	0.40
18-2/15-21	319.65	28.55	1.06	0.57
18-3/15-20	321.15	28.89	1.61	0.73
18-3/138-144	322.38	29.17	1.62	0.70
18-4/20-25	322.70	29.24	1.59	0.83
18-4/120-125	323.70	29.47	1.45	0.36
18-5/23-29	324.23	29.59	1.31	0.44
18-5/65-67	324.65	29.69	1.68	0.29
18cc	324.70	29.70	1.51	0.55
19-1/12-18	327.62	30.36	1.58	0.46
19-2/7-12	329.07	30.68	1.37	0.48
19-3/12-18	330.62	31.03	1.35	0.55
20-1/24-28	337.24	32.51	1.59	0.79
20-2/41-47	338.91	32.88	1.40	0.93
20-4/20-26	341.70	33.10	1.59	0.70
20-7/4-10	346.04	33.51	1.55	0.93
21-1/110-116	347.60	33.55	1.75	1.29
21-2/112-116	349.12	33.67	1.15	0.63
21-3/90-95	350.40	33.76	1.47	1.06
			1.46	1.04
21-4/31-37	351.31	33.83	1.71	1.40
			1.58	1.18
21-5, 20-25	352.70	33.91	1.33	1.79
			1.21	1.65
21-6/22-28	354.22	34.05	1.98	1.54
22-1/110-116	356.60	34.23	2.43	1.50
22-2/3-8	357.03	34.27	1.56	1.10
			1.63	1.16

TABLE 1. (continued)

Core-Section/ Depth, cm	Depth, msb	Age, Ma	$\delta^{18}\text{O}_{\text{PDB}}$	$\delta^{13}\text{C}_{\text{PDB}}$
<u>Cibicidoides spp., Site 563 (continued)</u>				
22-2/104-108	358.04	34.34	1.46	1.18
22-3/6-11	358.56	34.38	2.28	1.44
22-3/110-116	359.60	34.46	1.67	1.19
			1.74	1.06
22-3/145-149	359.95	34.49	1.74	0.95
			1.86	1.00
22CC	360.10	34.50	1.63	1.03
<u>Catapsydrax spp., Site 563</u>				
13-2/124-130	273.24	19.82	0.82	1.04
13-5/124-130	277.74	20.49	0.78	1.07
14-2/129-135	282.79	21.24	0.99	1.02
15-2/117-123	292.17	22.63	0.97	1.62
15-5/117-123	296.67	23.31	1.05	1.63
16-2/117-123	301.67	24.45	1.22	1.69
16-5/117-123	306.17	25.47	1.05	1.01
17-2/117-123	311.17	26.62	1.26	1.17
17-5/24-30	314.74	27.43	1.48	1.12
18-2/23-29	319.73	28.57	1.51	0.98
18-5/23-29	324.23	29.60	1.55	0.76
19-1/12-18	327.62	30.36	1.76	0.94
19-3/12-18	330.62	31.03	1.14	1.06
20-2/41-47	338.91	32.88	1.23	1.26
20-4/20-26	341.70	33.10	1.30	1.26
20-7/4-10	346.04	33.51	1.08	1.27
21-1/140-146	347.60	33.55	0.91	1.21
21-4/31-37	351.31	33.83	1.08	1.42
22-1/110-116	356.60	34.23	0.84	1.30
22-3/110-116	359.60	34.46	0.91	1.30
<u>Globorotalia opima nana, Site 563</u>				
15-2/117-123	292.17	22.63	0.70	1.72
15-4/110-115	295.10	23.31	0.98	1.63
16-1/114-120	300.14	24.10	1.08	1.82
16-3/110-116	303.10	24.78	1.10	1.59
16-5/117-123	306.17	25.01	0.07	0.83
16-4/109-115	304.59	25.12	0.64	1.13
17-2/117-123	311.17	26.62	0.15	0.95
17-5/24-30	314.74	27.43	0.38	0.68
18-2/23-29	319.73	28.57	0.46	0.91
18-5/23-29	324.23	29.60	0.74	0.84
19-1/12-18	327.62	30.36	0.37	0.48
19-3/12-18	330.62	31.03	1.08	0.78
20-1/24-28	337.24	32.51	1.12	1.11
20-2/41-47	338.91	33.10	1.44	1.18
21-1/140-146	347.60	33.55	0.97	0.95
21-4/31-37	351.31	33.83	0.97	1.27

TABLE 1. (continued)

Core-Section/ Depth, cm	Depth, msb	Age, Ma	$\delta^{18}\text{O}_{\text{PDB}}$	$\delta^{13}\text{C}_{\text{PDB}}$
<i>Globorotalia opima opima</i> , Site 563				
20-2/41-47	338.91	32.88	1.10	1.08
20-4/20-26	341.70	33.10	1.16	0.75
20-7/4-10	346.04	33.51	0.95	1.03
21-4/31-37	351.31	33.83	0.83	1.50

In samples with replicates, mean values were plotted in Figures 1-4. Age estimates were obtained by interpolating between biostratigraphic datum levels and magnetostratigraphic chron boundaries (Figures 1 and 2). Slight differences between $\delta^{18}\text{O}$ values reported here and those of Miller and Fairbanks [1983] are due to use of H. Craig's (personal communication, 1978, to Blattner and Hulston [1978]) correction to PDB standard.

^a Possible recrystallization (see text), data not plotted.

^b Large crystals noted; visible crystals removed from test by sonifying. Data plotted in Figures 2 and 3.

^c Corresponding $\delta^{18}\text{O}$ values for this level: *Cibicidoides*, 1.5⁰/oo; *Catapsydrax*, 1.7⁰/oo.

records of *Cibicidoides* from the Atlantic and Pacific. (We use the term cycle here to refer to periodic or aperiodic sequences of changes in which the initial conditions are reestablished.) The $\delta^{13}\text{C}$ cycles also have been observed in planktonic foraminifera [Shackleton et al., 1984; Vincent et al., 1984; Vincent and Berger, this volume; this study], strongly suggesting that these $\delta^{13}\text{C}$ cycles resulted from global changes in the $\delta^{13}\text{C}$ of mean ocean water. These cycles occurred with periods of 7-12 m.y. and amplitudes of approximately 1.0⁰/oo, and are similar in amplitude to late Quaternary high-frequency (10⁴-10⁵ year) benthic and planktonic foraminiferal $\delta^{13}\text{C}$ cycles [Shackleton et al., 1983; Mix and Fairbanks, 1984; Fairbanks and Mix, 1984].

Differences in the amplitudes between Atlantic and Pacific benthic foraminiferal $\delta^{13}\text{C}$ cycles can be used to reconstruct abyssal circulation [e.g., Shackleton et al., 1983; Mix and Fairbanks, 1984]. Today, Atlantic deep and bottom waters are 1⁰/oo higher in $\delta^{13}\text{C}$ than are Pacific bottom waters, reflecting production of "young," ^{13}C -enriched North Atlantic Deep Water (NADW) [Kroopnick et al., 1972; Kroopnick 1974, 1980, 1984]. Blanc et al. [1980], Blanc and Duplessy [1982], Miller and Fairbanks [1983] and Miller and Thomas [1984] attempted to isolate global from abyssal circulation $\delta^{13}\text{C}$ changes in the Tertiary by comparing carbon isotopic records from the Atlantic and the Pacific. This approach is similar to that of Curry and Lohmann [1982],

Shackleton et al. [1983], and Mix and Fairbanks [1984] for the late Quaternary, and is used in this study. We demonstrate that the Atlantic also was enriched in ^{13}C relative to the Pacific from 36-33 Ma (early Oligocene) and 26-10 Ma (late Oligocene to late Miocene), suggesting supply of ^{13}C -enriched bottom water in the North Atlantic, perhaps in a mode analogous to modern NADW.

Tertiary paleoclimatic studies generally have assumed that the earth was substantially ice free prior to about 15 Ma (middle Miocene), and therefore $\delta^{18}\text{O}$ history prior to 15 Ma reflected only temperature changes [Shackleton and Kennett, 1975; Savin et al., 1975, 1981; Savin, 1977; Woodruff et al., 1981]. However, recent studies of benthic foraminifera indicate that continental ice sheets must have existed in the Oligocene at approximately 36-35 Ma and 31-28 Ma [Miller and Fairbanks, 1983; Keigwin and Keller, 1984; Miller and Thomas, 1984; Poore and Matthews, 1984; Shackleton et al., 1984]. We present further evidence that supports the existence of Oligocene ice sheets at 31-28 Ma, and new data which suggest glaciation at about 25-24 Ma (below the Oligocene/Miocene boundary). The $\delta^{13}\text{C}$ cycles are compared with the record of glaciation to assess whether the cycles were related to sea-level changes as they were in the Quaternary.

Methods and Samples

Isotopic measurements on benthic and planktonic foraminifera were obtained for the Oligocene to

Miocene sections at DSDP Sites 558 (37°46.24'N, 37°20.61'W; 3754 m present water depth) and 563 (33°38.53'N, 43°46.04'W; 3796 m present water depth) in the western North Atlantic (Figures 1-3; Table 1; see also Miller and Fairbanks [1983]). These two sites were selected for their relatively continuous records and for the excellent age control provided by magnetobiostratigraphy [Khan et al., 1984]. Our Atlantic data are compared with benthic foraminiferal isotopic data from the following/ Pacific sites: eastern equatorial Site 574 (03°59.24'N, 114°08.53'W; 4561 m present depth [Miller and Thomas, 1984]), eastern equatorial Site 77 (00°28.90'N, 133°13.70'W; 4291 m present depth [Savin et al., 1981; Keigwin and Keller, 1984]), and western equatorial Site 289 (00°29.92'S, 158°30.69'E; 2206 m present depth [Savin et al., 1981]).

Both Site 558 and Site 563 are located beneath the subtropical gyre. The surface waters in subtropical gyres are generally low in nutrients and nearly uniform in carbon isotopic composition [Kroopnick, 1984]. The carbon isotopic records of planktonic foraminifera from subtropical gyres are not subject to changes caused by variable upwelling rates of waters rich in nutrients and low in $\delta^{13}\text{C}$. Since the North Atlantic subtropical gyre was a stable feature throughout the Tertiary [Berggren and Hollister, 1974], Sites 558 and 563 are particularly suitable locations for obtaining Tertiary global carbon isotopic records.

All benthic foraminiferal isotopic analyses were performed on mixed species of *Cibicidoides* (including *Planulina wuellerstorfi*, which many authors place in *Cibicidoides wuellerstorfi*). *Cibicidoides* spp. (including *P. wuellerstorfi*) secrete calcite tests that are consistently offset from $\delta^{18}\text{O}$ equilibria by approximately 0.65‰ [Shackleton and Opdyke, 1973; Duplessy et al., 1980; Woodruff et al., 1980; Belanger et al., 1981; Graham et al., 1981]. In addition, studies of Holocene core tops have demonstrated that *Cibicidoides* accurately reflects the distribution of ^{13}C of CO_2 in the modern ocean [Belanger et al., 1981; Graham et al., 1981].

Planktonic foraminiferal isotopic analyses were performed on *Globorotalia opima nana* (Figures 1 and 2), the taxon which was generally most depleted in ^{18}O at Sites 558 and 563. Analyses of *G. opima opima* yielded values similar to those for *G. opima nana*, while analyses of *Globigerina angulicostata* yielded values similar to or more enriched than those for *G. opima nana* (Table 1). Poore and Matthews [1984] noted that compared with other species at Site 366 (equatorial Atlantic), *G. opima nana* exhibited intermediate $\delta^{18}\text{O}$ values; however at higher latitudes (Site 522, 26°S), similar to our locations, this taxon (and *G. opima opima*) appears to have been among the most depleted in ^{18}O [Poore and Matthews, 1984]. Therefore, we assume that it calcified in the surface mixed layer at Sites 558 and 563. Analyses of the planktonic foraminifera *Catapsydrax* spp. (Figure 3; Table 1) showed that

this was consistently the most enriched planktonic taxon at this location; Poore and Matthews [1984] similarly noted that this was the most enriched planktonic taxon at Site 366.

Samples were washed with sodium hexametaphosphate and/or hydrogen peroxide in tap water through a 63- μm sieve and air dried. Benthic foraminifera were ultrasonically cleaned in distilled water for 10-20 s, crushed, and roasted at 370°C in a vacuum to remove organic matter. Similar procedures were followed for planktonic foraminifera, although they were not sonified due to susceptibility to breakage. The CaCO_3 was reacted with H_3PO_4 at 50°C, and the gas was analyzed with a VG Micromass 903E mass spectrometer following standard procedures.

Ages are reported using the Berggren et al. [1984a, b] time scale by linearly interpolating between magnetostratigraphic chron boundaries where possible and using planktonic foraminiferal datum levels where magnetostratigraphic data are limited (Figures 1-4). With such magnetobiostratigraphic control [Khan et al., 1984], we believe that, in general, the Atlantic age models are reliable within 0.5 m.y. or better. An exception to this occurs in the lower Oligocene of Site 563, where the age models are uncertain. Pacific age models are based entirely upon biostratigraphy (Figure 4), and are therefore less accurate. Pacific isotope data are sparse, and the Pacific age model is tentative for the time interval 26-21 Ma.

Paleodepths for Sites 558 and 563 are estimated by "backtracking" [Sclater et al., 1971, 1977; Berger and Winterer, 1974] using the empirical western North Atlantic age versus subsidence curve [Tucholke and Vogt, 1979]. These sites subsided from approximately 2200-2400 m during the earliest Oligocene (36 Ma) to 3200 m during the latest Oligocene (24 Ma) and to 3600 m during the late middle Miocene (11 Ma).

The average benthic foraminiferal sampling intervals at Sites 558 and 563 are between 0.3 and 0.4 m.y. Planktonic foraminiferal sampling intervals are typically 0.7 m.y., although we analyzed at 0.3 m.y. sampling interval for *Catapsydrax* spp. at Site 558 (Figure 3). If high-frequency changes similar to Quaternary ^{18}O isotopic fluctuations (on the order of 10^4 - 10^5 years) occurred in the Oligocene-Miocene, our sampling may have introduced aliasing. However, the isotopic records from both sites, including planktonic and benthic data, show similar patterns (Figures 1-3), despite differences in sampling intervals. Therefore, the sampling interval is probably sufficient to resolve isotopic changes on the order of 10^6 - 10^7 years.

Results

Oxygen Isotope Stratigraphy

Oxygen isotope data provide (1) constraints on ice-volume and temperature changes; (2)

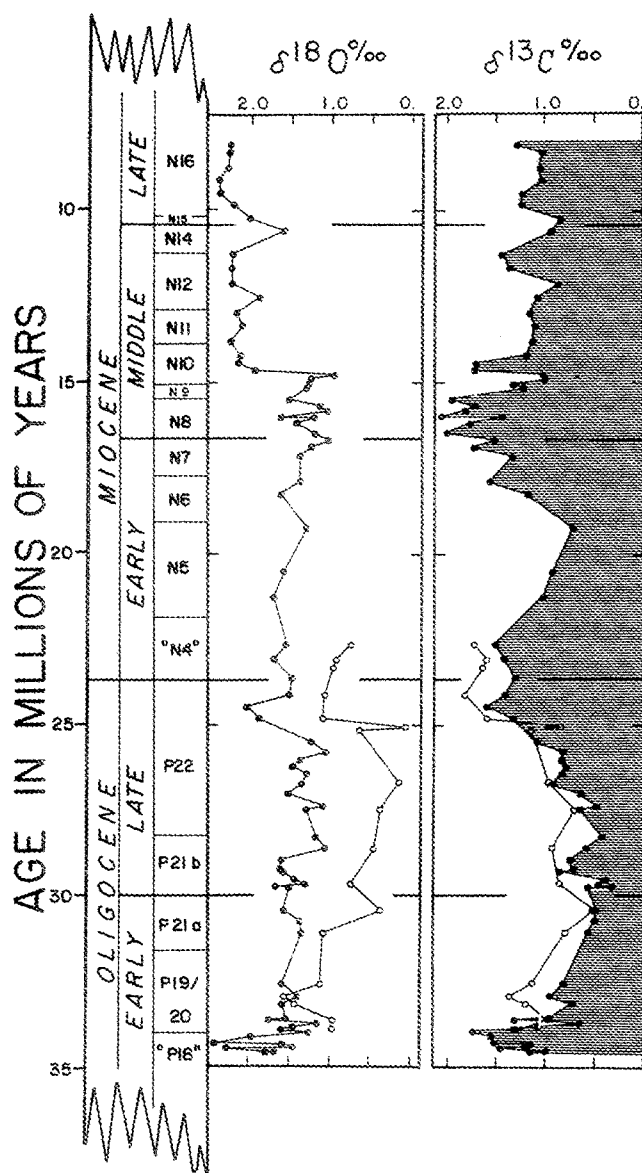


Fig. 1. Site 563 oxygen and carbon isotopic composition for the benthic foraminifera *Cibicidoides* (solid circles) and the planktonic foraminifera *Globorotalia opima nana* (open circles). The age model was obtained by interpolation between the following levels: base chron C4An, 159.5 msb, 8.2 Ma; base chron C5n, 188.0 msb, 10.4 Ma; top chron C5An, 204.5 msb, 11.9 Ma; base chron C5Bn, 242.5 msb, 15.3 Ma; top chron C5Cn, 251.0 msb, 16.2 Ma; top chron C5Dn, 260.0 msb, 17.6 Ma; base chron C5Dn, 267.0 msb, 18.1 Ma; unconformity 270.5 msb from 18.4 to 19.4 Ma; base chron C6n, 277.5 msb, 20.5 Ma; top chron C6Cn, 296.5 msb, 23.3 Ma; last occurrence (LO) of *Chiloguembelina*, 326.0 msb, 30.0 Ma; base chron C12n, 339.0 msb, 32.9 Ma; LO of *Pseudohastigerina*, 353.5 msb, 34.0 Ma.

supplementary stratigraphic control, especially in correlating sites without magnetostratigraphy (e.g., Sites 77, 289, 574); and (3) evidence for the suitability of our samples for carbon isotopic studies, allowing us to screen sections altered by diagenesis.

Site 563. Rapid increases occurred in benthic foraminiferal $\delta^{18}\text{O}$ values (Figure 1) at 14.7 Ma (middle Miocene) and 25–24 Ma (immediately below the Oligocene/Miocene boundary). Peak values at these times were greater than 2.0‰ . The 14.7 Ma event precisely correlates with a well-documented benthic foraminiferal increase noted in the Atlantic, Pacific, and southern oceans [e.g., Savin et al., 1975, 1981; Shackleton and Kennett, 1975; Boersma and Shackleton, 1977], providing control on stratigraphic correlations between basins. The 25–24 Ma increase has been detected elsewhere only at Site 558 [this study]. This apparently results from a previous lack of study of uppermost Oligocene sediments.

An interval of high benthic foraminiferal $\delta^{18}\text{O}$ values (greater than 2.0‰) occurred at other DSDP sites near the early/late Oligocene boundary (between about 31 and 28 Ma (Table 2)); such high values were not observed at Site 563 (Figure 1) (values only attained 1.7‰). However, at Site 563 some of the section representing the time 32–28 Ma probably is missing and/or mechanically disturbed as suggested by (1) poor recovery precluded the identification of magnetochrons C10 and C11 (32.6–29.7 Ma), and it is in these chrons that highest $\delta^{18}\text{O}$ values are expected (see below); (2) chrons C8n and C9n (29.7–26.9 Ma) are concatenated and a hiatus is indicated [Khan et al., 1984]; and (3) drilling disturbance between 335 meters subbottom (msb) and 324 msb (estimated age 31.5–29.5 Ma) may have mixed the samples, lowering the magnitude of peak values.

High benthic foraminiferal $\delta^{18}\text{O}$ values (greater than 2.0‰) occurred near the base of Site 563 (Figure 1). These values may correlate with peak $\delta^{18}\text{O}$ values noted at other DSDP sites at 36–35 Ma following the well-established increase that occurred just above the Eocene/Oligocene boundary [e.g., Savin et al., 1975; Shackleton and Kennett, 1975; Kennett and Shackleton, 1976; Keigwin, 1980; Miller and Curry, 1982; Belanger and Matthews, 1984; Keigwin and Keller, 1984; Miller and Thomas, 1984; Miller et al., 1984a; Oberhänsli and Toumarkine, 1984; Poore and Matthews, 1984; Shackleton et al., 1984; Snyder et al., 1984]. Although the estimated age for these high $\delta^{18}\text{O}$ values at Site 563 is 34.5–34.0 Ma (Figure 1), we caution that the lower Oligocene age model (below 350 msb) at Site 563 is based upon extrapolation of sedimentation rates from biostratigraphic levels. We believe that these peak $\delta^{18}\text{O}$ values at Site 563 correlate with similar high values noted at 36–35 Ma elsewhere; this is within uncertainties in the age control at the base of Site 563.

The planktonic foraminiferal $\delta^{18}\text{O}$ record of *Globorotalia opima nana* also exhibited high values

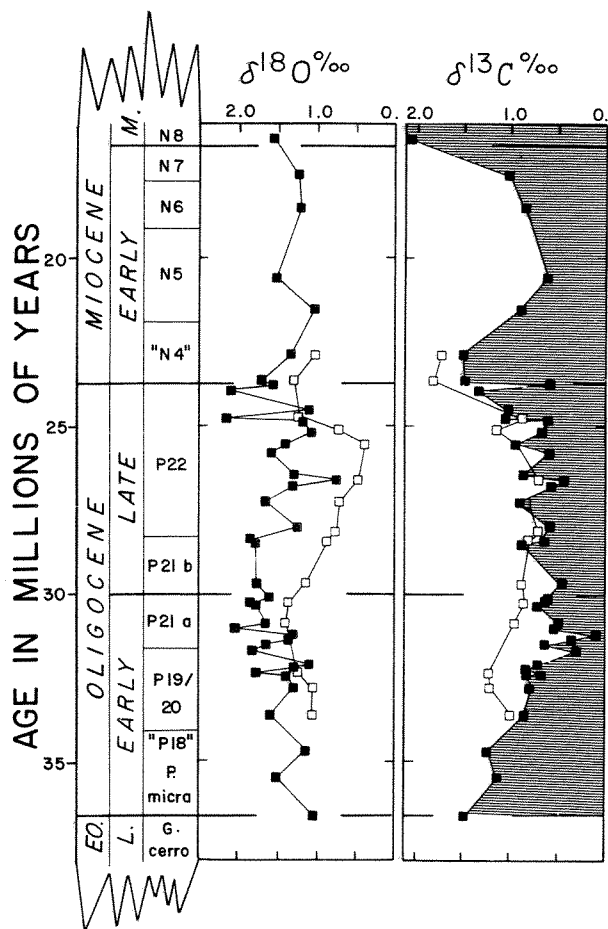


Fig. 2. Site 558 oxygen and carbon isotopic composition for the benthic foraminifera *Cibicidoides* (solid squares) and planktonic foraminifera *Globorotalia opima nana* (open squares). The age model was obtained by interpolation between the following levels: top chron C5n, 180.0 msb; 8.9 Ma; base chron C5n, 218 msb, 10.4 Ma; base chron C5n, 277.5 msb, 15.3 Ma; base chron C5n, 298.5 msb, 17.0 Ma; top chron C5n, 301.5 msb, 17.6 Ma; base chron C5n, 304.0 msb, 18.1 Ma; unconformity at 307.5 msb from 19.0 to 19.7 Ma; base chron C6n, 310.5 msb, 20.5 Ma; top chron C6n, 322.0 msb, 23.3 Ma; middle chron C10, 359.5 msb, 30.0 Ma; base chron C11n, 378.0 msb, 32.1 Ma; base chron C12n, 388.0 msb, 32.9 Ma; top chron C13n, 399.0 msb, 35.3 Ma; base Oligocene, 405.8 msb, 36.5 Ma.

at 25–24 Ma (Figure 1). Covariance of planktonic and benthic foraminiferal $\delta^{18}\text{O}$ at 25–24 Ma can be estimated as a minimum of $0.5^\circ/\text{oo}$. Using a strategy applied to Quaternary studies [Shackleton and Opdyke, 1973], we interpret planktonic-benthic covariance as resulting from ice-volume changes. High $\delta^{18}\text{O}$ values also occurred in *G. o. nana* at about 34–31 Ma when benthic values were low

(Figure 1). Changes in planktonic $\delta^{18}\text{O}$ not reflected in benthic foraminifera (e.g., between 34 and 31 Ma (Figure 1)) are interpreted as sea-surface temperature or regional surface-water $\delta^{18}\text{O}$ changes, which is the same convention used in Pleistocene studies [Shackleton and Opdyke, 1973]. Site 558. Oxygen isotope analyses of the middle Miocene section (292–225 msb, 16.5–11.0 Ma) (Table 1) yielded anomalously low values. The prominent middle Miocene $\delta^{18}\text{O}$ increase and subsequent high values which were noted at Site 563 and many other locations (see above) were not recorded at Site 558. Examination of the section at Site 558 shows that between approximately 290 and 220 msb, sediments are well indurated chalks; samples from this section were difficult to process, and preservation is poor. We suspect that the low $\delta^{18}\text{O}$ values observed (1.1 – $1.7^\circ/\text{oo}$ versus consistently greater than $2.0^\circ/\text{oo}$ for the correlative section at Site 563 and elsewhere) reflect partial recrystallization.

The section below 290 msb at Site 558 is less lithified, consisting of oozes and slightly indurated chalks; this section yields $\delta^{18}\text{O}$ values which are similar to those noted at correlative levels at Site 563. In addition, the oxygen isotopic record below 290 msb (older than 16 Ma) reveals patterns and values which are similar to records at Sites 77 and 574 [Keigwin and Keller, 1984; Miller and Thomas, 1984]. We believe that although diagenesis altered the overlying section (290–220 msb), most of the section representing the interval between 31 and 16 Ma is unaltered, since the close agreement of $\delta^{18}\text{O}$ records occurs among sites with different burial depths and diagenetic histories. The exception to this occurs near the base of Hole 558 (below approximately 370 msb; older than 31.2 Ma (Table 1)); here, large (greater than $150\ \mu\text{m}$) crystalline overgrowths of calcite were observed. By ultrasonically cleaning the samples for longer intervals we were able to separate the crystals from the exterior test walls (analyses of concentrated crystals given in Table 1). However, the data from below 370 msb must be considered less reliable, and comparisons using these data (Figure 4, older than 31 Ma) are tentative.

The benthic foraminiferal $\delta^{18}\text{O}$ record at Site 558 (Figure 2) shows that high values occurred at 25–24 Ma; high values also occurred immediately above and below the early/late Oligocene boundary (Figure 2; values exceeding $2.0^\circ/\text{oo}$ in both intervals). Planktonic foraminiferal (*Globorotalia opima nana*) $\delta^{18}\text{O}$ values also were high at 25–24 Ma and near the early/late Oligocene boundary (Figure 2). The high benthic and planktonic foraminiferal $\delta^{18}\text{O}$ values noted between 25 and 24 Ma at Site 558 correlate with similar peak values at Site 563 (Figure 1). Although planktonic and benthic records may exhibit covariance of greater than $0.5^\circ/\text{oo}$ at 25 Ma (Figure 1; Table 2), the increase in planktonics at about 25 Ma apparently began prior to the benthic foraminiferal $\delta^{18}\text{O}$ increase (Figure 2).

This apparent lead-lag relationship may be an artifact of sampling and/or aliasing of the record.

The record recovered at Site 558 representing the time interval 32–28 Ma is much more complete than that same interval at Site 563. Magnetochrons C9 through C12 (approximately 32–28 Ma) are well represented at Site 558 [Khan et al., 1984], suggesting that the record is complete. High benthic and planktonic foraminiferal $\delta^{18}\text{O}$ values occurred near the early/late Oligocene boundary (Figure 2); oxygen isotopic values increased prior to 31 Ma at Site 558 and reached maximum values from 31 to 30 Ma, with benthic values remaining high until 28 Ma (Figure 2). At Sites 574 and 77 in the Pacific, similar high $\delta^{18}\text{O}$ values in *Cibicidoides* were noted at 29–30 Ma [Keigwin and Keller, 1984; Miller and Thomas, 1984]. This slight apparent time discrepancy can be attributed to uncertainties in Pacific biostratigraphy. Given such uncertainties, we estimate the timing of this $\delta^{18}\text{O}$ event as follows. High values at Site 558 are associated with chron C11n (top) to C9n and planktonic foraminiferal Zones P21a and P21b, constraining peak values to about 31–28 Ma. This is consistent with the observation of high values in Zones P21a/P21b at Sites 77 and 574. The timing of the change from lower to higher values at Site 558 apparently occurred in chron C11 at about 32–31 Ma (Figure 2), although the data older than 31 Ma may have been affected by diagenesis. However, this estimate is supported by data from Site 77 [Keigwin and Keller, 1984], where the change began within Zone P21a, constraining the increase to about 31 ± 0.5 Ma.

At both Site 558 and Site 563, *Catapsydrax* spp. are consistently the most isotopically heavy planktonic foraminifera. The $\delta^{18}\text{O}$ record of this taxon (Figure 3) shows little correspondence to the record of benthic foraminifera or to *Globorotalia opima nana* (Figures 1–3).

Global $\delta^{13}\text{C}$ Cycles

Three cycles of benthic foraminiferal $\delta^{13}\text{C}$ occurred at Site 563 in the Oligocene–Miocene (Figure 1). High values occurred at 36–33 Ma (within the early Oligocene), 26–22 Ma (across the Oligocene/Miocene boundary), and 18–14 Ma (across the early/middle Miocene boundary (Figure 1)). Nearly the same pattern occurred in planktonic foraminifera (in both *Catapsydrax* spp. and *Globorotalia opima nana*) at both subtropical gyre Sites 558 and 563 during the Oligocene to early Miocene (Figures 1–3). At Site 558, the same pattern can be seen in Oligocene to middle Miocene benthic foraminifera (Figure 2; Table 1).

The early to late Oligocene cycle also has been noted in benthic foraminifera from the Pacific [Keigwin and Keller, 1984; Miller and Thomas, 1984], South Atlantic [Poore and Matthews, 1984; Shackleton et al., 1984], Gulf of Mexico [Belanger and Matthews, 1984], and Bay of Biscay [Miller and

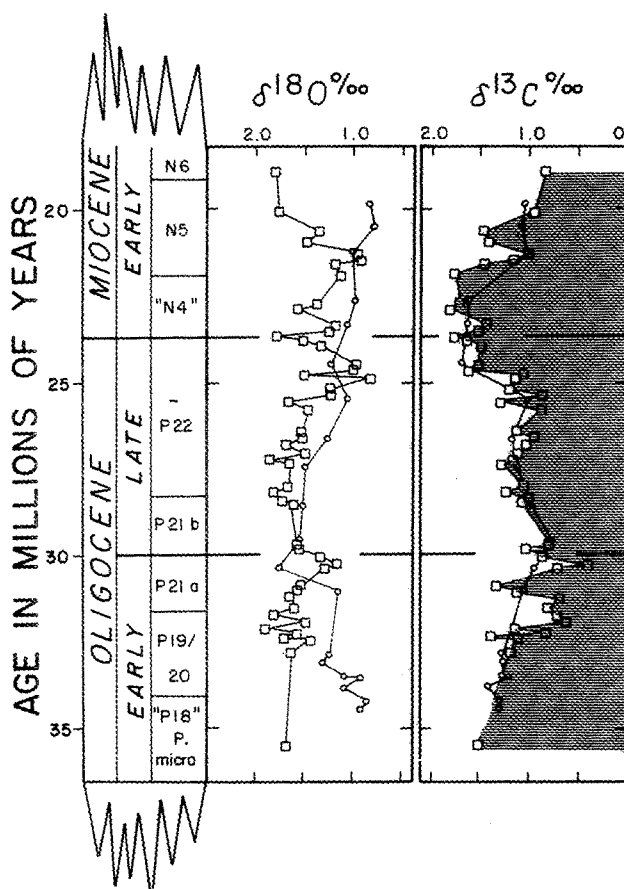


Fig. 3. Site 558 (open squares) and Site 563 (open circles) oxygen and carbon isotopic composition for the planktonic foraminifera *Catapsydrax*. Age models are as in Figures 1 and 2.

Curry, 1982]. Shackleton et al. [1984] also noted a similar change in Oligocene South Atlantic planktonic foraminifera. In the Miocene, the $\delta^{13}\text{C}$ cycles can be observed in both benthic and planktonic foraminifera from the Pacific [Savin et al., 1981; Vincent et al., 1984], Indian [Vincent and Berger, this volume], and South Atlantic [Shackleton et al., 1984] oceans.

An increase in benthic foraminiferal $\delta^{13}\text{C}$ also occurred across the Eocene/Oligocene boundary at Pacific and North Atlantic sites [Miller and Curry, 1982; Keigwin and Keller, 1984; Miller and Thomas, 1984; Miller et al., 1984a]. Keigwin [1980] and Oberhänsli and Toumarkine [1984] noted a $\delta^{13}\text{C}$ increase in both benthic and planktonic foraminifera across the Eocene/Oligocene boundary in the Pacific and South Atlantic oceans, respectively.

It is apparent that benthic and planktonic foraminifera recorded three worldwide Oligocene–Miocene $\delta^{13}\text{C}$ cycles with amplitudes of approximately 0.5 – 1.0 ‰. Choosing $\delta^{13}\text{C}$ maxima as the boundaries (Figure 4), cycle 1 lasted about

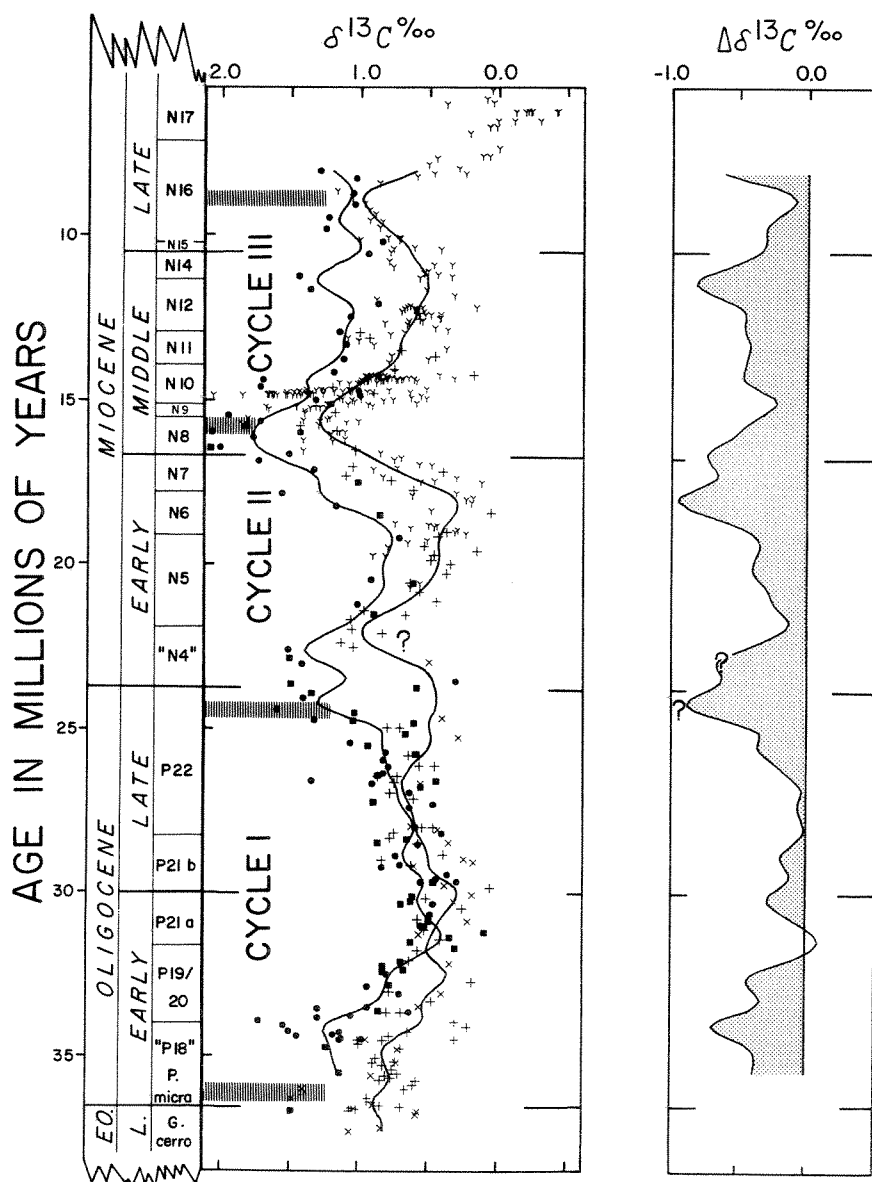


Fig. 4. Comparison of benthic foraminiferal (*Cibicidoides*) $\delta^{13}\text{C}$ from Atlantic Sites 558 (solid squares) and 563 (solid circles) with Pacific Sites 574 (crosses (data from Miller and Thomas [1984])), 77 (pluses (data from Keigwin and Keller [1984] and Savin et al. [1981])), and 289 (letter Y (data from Savin et al. [1981])). Curves were obtained by separately interpolating Atlantic and Pacific data to constant time intervals (0.2 m.y.) and smoothing with an 11-point Gaussian convolution filter, eliminating frequencies less than approximately 1/m.y. Global cycle boundaries were chosen as the most enriched interval constrained by the Pacific benthic record, except the cycle boundary at about 24.5 Ma, which was chosen on the basis of the Atlantic record (Pacific age model is unreliable at this time). The difference curve was obtained by subtracting the smoothed Atlantic $\delta^{13}\text{C}$ record from the smoothed Pacific record. The solid line is drawn through the 0.0‰ difference, and regions less than zero are shaded. Pacific isotope data are sparse, and Pacific age models are considered unreliable in the time interval 26–21 Ma. Age models for Atlantic sites are from Figures 1 and 2. Those for Pacific sites are as follows: Site 574, after Mayer, Theyer et al. [1984]; Site 77, after Miller and Fairbanks [1983] and Miller and Thomas [1984]; Site 289, model obtained by interpolating between last occurrence of *Globorotalia margaritae* (156.5 msb, 5.6 Ma) [Srinivasan and Kennett, 1981] and other biostratigraphic datum levels cited by Miller and Fairbanks [1983].

12 m.y. (from about 36.5 to 24.5 Ma), cycle 2 lasted about 8.5 m.y. (from about 24.5 to 16 Ma), and cycle 3 lasted about 7 m.y. (from about 16 to 9 Ma).

Since the cycles delineated here occurred in benthic and planktonic foraminifera from many locations, they must have resulted from global changes in $\delta^{13}\text{C}$ of mean ocean water. Several mechanisms have been proposed to explain Pleistocene $\delta^{13}\text{C}$ variations with shorter periodicities (10^3 – 10^4 years), including changes in the size of the terrestrial organic carbon reservoir [Shackleton, 1977] and amount of organic carbon deposited on continental margins, presumably in response to sea-level changes [Broecker, 1982]. However, since the residence time of carbon in the oceans is approximately 1.8×10^5 years [Froelich et al., 1982], these mechanisms cannot be invoked to explain the cycles delineated here. For example, a one-time transfer of organic carbon from the terrestrial biosphere to the ocean would result in a rapid $\delta^{13}\text{C}$ change in planktonic and benthic foraminifera, such as that measured in the Pleistocene. However, the average carbon isotopic composition of carbonate sediments would eventually be restored to the average carbon isotopic composition of incoming carbon derived from weathering rocks. The time it takes to replace the ocean carbon inventory is short (1.8×10^7 years) compared with the average period of the carbon isotopic cycles (9×10^6 years) delineated here. Thus, Oligocene-Miocene $\delta^{13}\text{C}$ cycles were not controlled by transfer of organic carbon between the terrestrial biosphere and ocean, and therefore must represent changes in $\delta^{13}\text{C}$ of riverine bicarbonate (input) or changes in the ratio of the amount of carbon buried as calcium carbonate to the amount buried as organic carbon (output).

Ronov [1968] estimated that 18% of sedimentary carbon is stored in the geological record as organic carbon and that the remaining 82% of carbon is stored in the form of carbonate. Sedimentary organic carbon is approximately -25‰ in $\delta^{13}\text{C}$ [Craig, 1953], and carbonate carbon is approximately 0.5‰ [Keith and Weber, 1964], resulting in the following estimate for riverine (riv) CO_2 :

$$\delta^{13}\text{C}_{\text{CO}_2(\text{riv})} = \frac{\text{input}}{\text{output}} = f(\delta^{13}\text{C}_{\text{CaCO}_3}) + 1 - f(\delta^{13}\text{C}_{\text{CHOC}_2}) = 4.1\text{‰}$$

where f is the fraction of carbon stored in the form of carbonate.

The 1‰ changes noted in the Oligocene-Miocene $\delta^{13}\text{C}$ record (Figures 1–4) can be explained by changes in the ratio of CaCO_3 carbon to organic carbon in riverine input. Alternatively, the 1‰ enrichment can be explained by a decrease in the fraction of CaCO_3 carbon deposited in marine sediments from 0.82 to 0.79. These estimates assume that organic carbon $\delta^{13}\text{C}$ remained constant. A decrease in $\delta^{13}\text{C}$ of organic carbon from -25‰ to approximately -30‰ would also explain a 1‰ increase in $\delta^{13}\text{C}$ in CaCO_3 . We know of no

measurements of organic carbon variability which support this possibility. The carbon isotope mass balance is determined here by the ratio of organic carbon to carbonate carbon and implies nothing about the total amount of carbon stored in the oceans.

Discussion

We attempt to determine if there was an underlying oceanographic cause for the Tertiary global carbon cycles by comparing proxy indicators for sea-level and climatic changes with the $\delta^{13}\text{C}$ cycles. Such proxy indicators are provided by Tertiary records of ice volume– $\delta^{18}\text{O}$ history [this study], changes in depositional patterns on continental margins [Vail et al., 1977], sea-floor spreading rates [Pitman, 1978], and climatic changes recorded by terrestrial vegetation [Wolfe, 1978].

Glacio-eustasy and the $\delta^{18}\text{O}$ Record

Based upon warm bottom-water temperatures (5° – 9°C) inferred from analyses of mixed assemblages of benthic foraminifera, initial studies concluded that the earth was substantially ice free prior to the middle Miocene (about 15 Ma) [Shackleton and Kennett, 1975; Savin et al., 1975, 1981; Savin, 1977; Woodruff et al., 1981]. Matthews and Poore [1980] questioned the assumption of an ice-free world prior to 15 Ma, and Oligocene isotopic data presented here and elsewhere support their contention. The $\delta^{18}\text{O}$ data discussed here from Atlantic and Pacific DSDP sites show that high benthic foraminiferal values (greater than 2.0‰ in *Cibicides*) occurred at 36–35 Ma, 31–28 Ma, 25–24 Ma, and younger than 15 Ma (Figure 1; Table 2). These $\delta^{18}\text{O}$ values indicate that in the Oligocene, bottom-water temperatures were as cold as or colder than at present (typically 2°C), assuming an ice-free world [Shackleton and Kennett, 1975; Graham et al., 1981]. Such cold bottom waters require a cold source region and are incompatible with an ice-free world. Therefore, significant continental ice volume must have existed in the Oligocene and from the middle Miocene to present [Miller and Fairbanks, 1983; Keigwin and Keller, 1984; Miller and Thomas, 1984; Poore and Matthews, 1984; Shackleton et al., 1984].

Covariance of planktonic and benthic foraminiferal $\delta^{18}\text{O}$ is used in the Quaternary as an ice-volume indicator [Shackleton and Opdyke, 1973]. Increases of at least 0.3 – 0.5‰ apparently occurred in both planktonic and benthic foraminifera at about 36 Ma, 31 Ma, and 25 Ma (Table 2), suggesting that ice volume increased at these times. Using the Quaternary sea level– $\delta^{18}\text{O}$ calibration [Fairbanks and Matthews, 1978], sea level fell by 30–50 m at 36 Ma, 31 Ma, and 25 Ma. This implies the development of ice sheets at these times equal to one half of the size of modern ice sheets. Savin et al. [1975] suggested that the major middle Miocene (15 Ma) benthic foraminiferal $\delta^{18}\text{O}$ increase (Figure 1) may not

TABLE 2. Magnitude of Peak $\delta^{18}\text{O}$ Values for *Cibicidoides* (except as noted) and Planktonic-Benthic $\delta^{18}\text{O}$ Covariance

DSDP Site	Location	Peak $\delta^{18}\text{O}$ ‰/oo, PDB	Covariance	Reference
<u>Latest Oligocene (24-25 Ma event)</u>				
Site 558	western North Atlantic	2.1	0.5-0.9	this study
Site 563	western North Atlantic	2.1	0.4-0.9	this study
<u>"Middle" Oligocene (28-31 Ma Event)</u>				
Site 77	equatorial Pacific	2.3	?0.4	Keigwin and Keller [1984]
Site 574	equatorial Pacific	2.3	a	Miller & Thomas [1984]
Site 558	western North Atlantic	2.1	?0.4 ^b	this study
<u>Earliest Oligocene (35-36 Ma Event)</u>				
Site 292	Phillipine Sea	2.3 ^c	?0.3	Keigwin [1980]
Site 277	southern ocean	1.6 ^c	1.0	Keigwin [1980]
Site 522	South Atlantic	2.5 ^d	0.5	Oberhänsli and Toumarkine [1984]
Site 119	Bay of Biscay	2.0	a	Miller & Curry [1982]

^a no planktonic foraminiferal data available.

^b based, in part, upon data with some recrystallization.

^c = *Oridorsalis* spp.

^d = *Stilostomella* spp.

have been accompanied by an increase in planktonic foraminiferal $\delta^{18}\text{O}$. However, Vincent and Berger [this volume] and Vincent et al. [1984] noted a synchronous, albeit smaller, increase in planktonic $\delta^{18}\text{O}$ at low latitude Pacific and Indian Ocean sites. Shackleton and Kennett [1975] noted covariance between benthic and planktonic foraminifera at high latitudes at about 15 Ma, and suggested that ice sheets developed at this time. We tentatively accept the premise that the middle Miocene (15 Ma) benthic foraminiferal increase was associated, in part, with an increase in ice volume. Thus, we infer that ice-volume increases caused sea-level falls at about 36, 31, 25, and 15 Ma.

Comparison of our inferred glacio-eustatic history with the carbon isotopic cycles shows no consistent response of carbon isotopes to sea-level changes:

1. The global $\delta^{13}\text{C}$ increases that occurred at the beginning of the Oligocene (about 36 Ma) and near the Oligocene/Miocene boundary (about 25 Ma)

were associated with $\delta^{18}\text{O}$ increases interpreted as sea-level lowerings.

2. The inferred glacio-eustatic lowering of 31-28 Ma is not associated with a change in $\delta^{13}\text{C}$.

3. The global $\delta^{13}\text{C}$ increase beginning at 18 Ma leads a major $\delta^{18}\text{O}$ increase (inferred sea-level fall) by 2-3 million years (see also Vincent and Berger [this volume]). This inconsistent relationship precludes a simple causal relationship between sea-level changes and the $\delta^{13}\text{C}$ cycles.

Other Proxy Indicators of Sea level

Pitman [1978] used variations in the rate of sea-floor spreading to determine a tectono-eustatic record. Comparisons of the carbon isotopic cycles with the record of tectono-eustasy show that the carbon cycles were too rapid to be explained by this mechanism.

Vail et al. [1977] presented a record of changes in coastal onlap observed in seismic

profiles from passive continental margins. They interpreted the onlap record as a record of sea level. Although Pitman [1978] and Watts [1982] questioned a direct causal relationship between onlap and sea-level records, we acknowledge that there may be a relationship between the glacio-eustatic record and the onlap record, albeit an indirect one [see Miller et al., 1984b]. In the Oligocene, major sea-level falls inferred by Vail et al. [1977] and Vail and Hardenbol [1979] occurred at 31 Ma and 25 Ma. Therefore, there is little correlation between the sea level record inferred from seismic stratigraphy and the carbon cycles.

We conclude that the low-frequency (10^6 years) Tertiary $\delta^{13}\text{C}$ cycles were not directly tied to global sea-level changes like the high-frequency (10^4 - 10^5 years) Pleistocene cycles.

Terrestrial Climate

The timing of the global carbon isotope cycles appears to have been similar to changes in inferred terrestrial climate. Based upon changes in terrestrial vegetation, Wolfe [1978] suggested that periods of lowered mean annual temperature occurred in the early Oligocene, across the Oligocene/Miocene boundary, and in the middle Miocene, in good agreement with periods of high global $\delta^{13}\text{C}$ (Figure 4).

We suggest that the apparent correlation of the global $\delta^{13}\text{C}$ cycles with the terrestrial plant record may be invalid due to problems in correlation of the terrestrial record. For example, the period of greatest Paleogene at about 31-32 Ma, as constrained by radiometric dates [Wolfe and Hopkins, 1967]. Wolfe and Hopkins assigned this cool interval to the middle of the Oligocene, which agrees with the time scale of Berggren et al. [1984b]. Wolfe [1978] later reinterpreted this same cooling with the Eocene/Oligocene boundary, suggesting that the terrestrial cooling is synchronous with the global increase in $\delta^{13}\text{C}$ noted here. However, we suggest that the terrestrial cooling occurred near the early/late Oligocene boundary as indicated by the radiometric dates of Wolfe and Hopkins [1967]. In this interpretation, the $\delta^{13}\text{C}$ cycles do not correlate well with the terrestrial plant record.

Abyssal Circulation History:

The Pacific-Atlantic $\delta^{13}\text{C}$ Difference

Comparison of the carbon isotopic record at Atlantic Sites 558 and 563 with Pacific Sites 77, 289, and 574 (Figure 4) shows that the Atlantic sites were generally higher than the Pacific sites, with greatest differences between oceans from about 36 to 33 Ma (early Oligocene) and from 26 to 10 Ma (latest Oligocene-Miocene). We interpret this as reflecting supply of ^{13}C -enriched bottom water to the North Atlantic, perhaps in a manner analogous to NADW. The difference between smoothed Pacific minus smoothed

Atlantic $\delta^{13}\text{C}$ values (Figure 4) illustrates gross changes in production of bottom water. However, the difference curve is highly sensitive to the age models used and to stratigraphic uncertainties. The mean difference for the interval examined was about one half of the present-day difference (0.4 versus 1.0‰); peak differences that occurred at about 12-10 Ma, 18-16 Ma, and 25-22 Ma approached 1‰ (Figure 4). We caution that the variations in the difference curve may be entirely due to slight inaccuracies in Pacific biochronology. Still, the peaks at 12-10 Ma and 18-16 Ma correlate with periods of bottom-water erosion noted in the North Atlantic [Miller and Tucholke, 1983; Mountain and Tucholke, 1984]. We speculate that increased flow of North Atlantic deep water resulted in bottom-water erosion at these times; more Atlantic data and better stratigraphic control in the Pacific are needed to test this conclusion. The peak difference noted at 22-25 Ma (Figure 4) is insignificant due to sparse Pacific isotopic data and uncertainties in the Pacific age models for this interval. The reduced $\delta^{13}\text{C}$ difference noted in the Oligocene (33-26 Ma) can be interpreted as reflecting reduced production of North Atlantic deep water [Miller and Thomas, 1984] or increased preformed nutrient levels in bottom-water source regions [Mix and Fairbanks, 1984]. Unlike in the Pleistocene, there apparently are no correlations among the Atlantic-Pacific $\delta^{13}\text{C}$ difference record, global changes in $\delta^{13}\text{C}$, and eustatic changes during the Oligocene-Miocene.

Conclusions

1. Three cycles of benthic and planktonic foraminiferal $\delta^{13}\text{C}$ occurred in the Oligocene to Miocene with periods of 7-12 m.y.; these cycles resulted from global changes in the $\delta^{13}\text{C}$ of mean ocean water. The $\delta^{13}\text{C}$ cycles were independent of changes in sea level indicated by changes in spreading rates (ocean volume), seismic stratigraphy, and oxygen isotopic stratigraphy. The most likely causes of these cycles were small changes in the input or burial ratio of carbonate carbon to organic carbon.
2. Pacific benthic foraminifera were depleted in ^{13}C relative to the Atlantic from 36-33 Ma (early Oligocene) and 26-12 Ma (late Oligocene to middle Miocene); we interpret this difference between basins as reflecting supply of ^{13}C -enriched bottom waters to the deep North Atlantic at these times. Reduced $\delta^{13}\text{C}$ differences between 33 and 26 Ma (early to late Oligocene) may reflect reduced production of "NADW" or an increase in preformed nutrient content of "NADW" source regions.
3. High benthic foraminiferal oxygen isotopic values noted at about 15 Ma (middle Miocene), 25-24 Ma (near Oligocene/Miocene boundary), 31-28 Ma (near early/late Oligocene boundary), and 36-35 Ma (just above the Eocene/Oligocene boundary) suggest the existence of significant continental

ice at these times; covariance between planktonic and benthic foraminiferal $\delta^{18}\text{O}$ records is used to estimate that development of ice sheets lowered sea level at 36 Ma, 31 Ma, and 25 Ma by 30–50 m. Unlike those in the Pleistocene, Tertiary global $\delta^{13}\text{C}$ changes did not correlate with glacio-eustatic changes.

Acknowledgments. We thank W.B. Curry, L.D. Keigwin, and A.C. Mix for discussions, W.B. Curry and A.C. Mix for reviewing the manuscript, A.C. Mix for data-smoothing programs, A.D. Chave for plotting programs, I. Ivanciu and M. Katz for technical assistance, and G. Kolibas, A. Ivanciu, D. King, and B. Gruder for performing isotopic analyses. Samples were provided by the DSDP. We thank E. Sundquist and W. Broecker for inviting us to participate in the Chapman Conference and the American Geophysical Union for travel support. This work was supported by National Science Foundation grants OCE 83-10086 (K.G.M.), OCE 82-08784 and OCE 82-00717 (R.G.F.), and a grant from the ARCO Foundation. This is Lamont-Doherty Geological Observatory of Columbia University contribution 3742.

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Geophysical Monograph 32

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and Atmospheric CO₂:
Natural Variations
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**E. T. Sundquist and W. S. Broecker
Editors**

American Geophysical Union
Washington, D.C.
1985