19. PALEOCEANOGRAPHIC CHANGES AND REEF GROWTH OFF THE NORTHEASTERN AUSTRALIAN MARGIN: STABLE ISOTOPIC DATA FROM LEG 133, SITES 811 AND 817, AND LEG 21, SITE 209[1]

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ABSTRACT

Stable isotopic data obtained from planktonic and benthic foraminifers were used to study paleoceanographic changes along the northeastern Australian margin from late Miocene (10 Ma) to Holocene time, and to evaluate the influence of these changes on reef growth. The data indicate that variations in surface-water temperatures may have had an important effect on the reef complexes on the Queensland Plateau and possibly off the northeastern Australian margin. Three sites were studied: Leg 21, Site 209 on the eastern edge of the Queensland Plateau, and Leg 133, Site 811 on the western margin, and Site 817 on the lower southern slope of the plateau.

Shallow-water bioclasts recovered from Holes 811A and 817A indicate extensive reef growth on the Queensland Plateau during the middle Miocene (before 12 Ma), signifying surface-water temperatures of 20°C or greater. The amount of reefal detritus produced during the late Miocene (10.0–5.2 Ma) decreased progressively, resulting in a reduction in area of the reef complexes. The isotopic data from planktonic foraminifers in these late Miocene age sediments indicate the presence of relatively cool surface waters (16°–19°C), which may have been a major factor contributing to the demise of the reefs on the Queensland Plateau. Surface waters remained cool until the middle Pleistocene (1.2–0.5 Ma), when the surface-water temperature apparently increased to approximately 25°C, recorded both in the isotopic data and by renewed reef growth. This increase occurred simultaneously (within the error of the age model) with the initiation of the Great Barrier Reef. We propose that cooling of surface waters during the early late Miocene contributed to reef decline on the Queensland Plateau, and that subsequent warming of surface waters during the middle Pleistocene promoted the initiation of reef growth on the northeastern Australian margin. Reef development on the Queensland Plateau never recovered to the middle Miocene extent because of a combination of tectonic (accelerated subsidence of the plateau) and paleoceanographic (the cooler surface waters present from the late Miocene throughout the Pliocene) factors.

Variations in seafloor $\delta^{18}O$ appear to be controlled by regional factors, as indicated by the similarity of data from Sites 811 and 817 to those from Site 590 on Lord Howe Rise.

INTRODUCTION

Two objectives of Leg 133 were to study the temporal and spatial evolution of the Queensland Plateau and associated reef systems, and to determine the effects of changing sea level, climate, and oceanic conditions on carbonate platform growth. Material recovered during Leg 133, and its correlation to seismic profiles, showed extensive reef growth on the Queensland Plateau before 12 Ma (Davies, McKenzie, Palmer-Julson, et al., 1991). Reef growth began to decline during the late middle Miocene, as shown by a decrease in the amount and type of bank-derived components in the slope sediments (Davies, McKenzie, Palmer-Julson, et al., 1991). This decline may have been the result of plate subsidence, although there was also a well-documented, global regression at this time (Haq et al., 1987). Therefore, although sea level was decreasing, the reefs apparently could not keep up with the subsidence. This indicates that other factors, such as paleoceanographic controls, may have been important in contributing to the reef demise.

To investigate possible effects of paleoceanographic changes on the Queensland Plateau margin, core material from three sites was selected: Site 811 to the west, Site 817 to the south, and Site 209 (DSDP Leg 21) to the east (Fig. 1). Oxygen isotopic ratios for both benthic and planktonic foraminifers from these sites were analyzed to evaluate possible variations in water temperature and current flow, which may have affected the growth and decline of reef complexes on the northeastern Australian margin.

[1] McKenzie, J.A., Davies, P.J., Palmer-Julson, A., et al., 1993. *Proc. ODP, Sci. Results,* 133: College Station, TX (Ocean Drilling Program).

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Modern Regional Oceanography

Surface-water circulation in the western Coral Sea is dominated most of the year by the slow southern edge of the westward-flowing South Equatorial Current (SEC) (Fig. 2; Pickard et al., 1977). This current enters the western Coral Sea and diverges; the southern branch becomes the East Australian Current near 20°S (Orme, 1973), and the northern branch flows into the Solomon Sea (Brandon, 1973; Orme, 1973; Pickard et al., 1977; Wyrtki, 1960). As a result of the influx of currents into this area, the water masses build up against the Australian continental margin and sink. Therefore, no significant upwelling occurs in this area (Pickard et al., 1977). In the austral summer, the influence of the Northwest Monsoon causes increased southerly currents near the northeastern Australian margin (Fig. 2). This flow is from surface waters moving across the Torres Strait and also possibly from waters flowing from the Solomon Sea (Wyrtki, 1960). Sea-surface temperatures in the western Coral Sea have an annual range of 22° to 28°C (Orme, 1973). This temperature range and surface-water salinity are favorable for high carbonate ion concentrations, thus promoting calcification and reef growth in areas of the Queensland Plateau and the Great Barrier Reef (Orme, 1973).

Two subsurface water masses affect the sites we studied: Subtropical Lower Water (SLW) and Antarctic Intermediate Water (AIW) (Pickard et al., 1977). SLW occurs from the surface to 250 m water depth and has a temperature of 18°–25°C (Brandon, 1973; Wyrtki, 1962). This water mass has a general southwesterly flow across the Queensland Plateau (Fig. 2). SLW usually exists below the SEC water, but, where the flow of SEC water is restricted (for instance by present-day reef complexes), SLW may reach the surface. AIW occurs at depths of 650 m to 1100 m and has a temperature of 1.7° to 4.2°C (Brandon, 1973; Pickard et al., 1977). AIW flows around and across the Queensland Plateau in areas deeper than 650 m (Fig. 2).

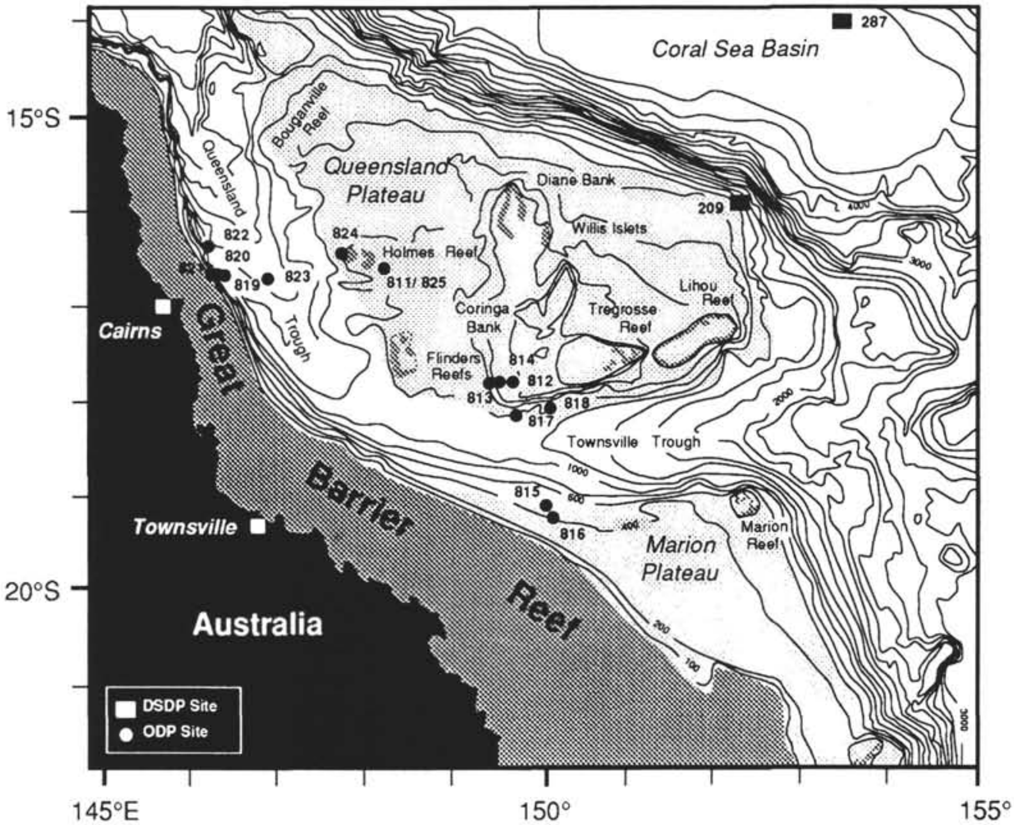

Figure 1. Map of the northeastern Australian margin after Davies et al. (1989). Leg 133 sites are designated with circles, and nearby Sites 209 and 287 with squares. Light gray outlines the generalized shape of the Queensland and Marion plateaus. Dark gray indicates areas of modern reef growth off the northeastern Australian margin.

Present-day wind circulation is dominated by southeast trade winds. In the austral summer, monsoonal northeast winds also may be important (Orme, 1973).

SITE DESCRIPTIONS

Site 209 (15°56.19′S, 152°11.27′E) was drilled during Leg 21 on the eastern edge of the Queensland Plateau in a water depth of 1428 m (Fig. 1). Core recovery was poor because of the varying hardness and sandy nature of the sediments (Burns, Andrews, et al., 1973). Middle Miocene to Holocene sediments are predominantly carbonate, with only small amounts of terrigenous material. Most of this carbonate is composed of foraminifers and calcareous nannofossils. In general, the microfossil preservation in the sediments is good, with most of the solution-susceptible species (e.g., the foraminifer *Globigerinoides sacculifera fistulosus*) remaining well preserved (Kennett, 1973). The foraminiferal faunas indicate a tropical environment from the mid-Pliocene to the Holocene (Kennett, 1973). A hiatus is present at this site from 11–3.3 Ma (between Cores 21-209-5H and 21-209-6H, approximately 45 mbsf).

Site 811 (16°30.98′S, 148°9.44′E) is located on the western margin of the Queensland Plateau, 3.5 nmi east of Holmes Reef in a water depth of 937 m (Fig. 1). Recovery in Hole 811A was high (99.4%), with a nearly continuous stratigraphic penetration of middle Miocene to Holocene sediments. The sediments recovered are predominantly foraminifer nannofossil oozes, with increased bank-derived material from the upper Pliocene to Holocene (Davies, McKenzie, Palmer-Julson, et al., 1991). Preservation of microfossils varies through the section. Microfossils in sediments with increased bank-derived material, such as Core 133-811A-3H, exhibit considerable infilling and overgrowths. This is caused by an increase in carbonate remineralization resulting from the presence of unstable carbonate minerals, such as aragonite and high-magnesium calcite, in the bank-derived material from the lower Pleistocene (Davies, McKenzie, Palmer-Julson, et al., 1991). Samples from predominantly pelagic sediments exhibit good preservation with limited infilling and overgrowths of microfossils.

Site 817 (18°9.50′S, 149°45.49′E) is located on the lower southern slope of the Queensland Plateau, in a water depth of 1016 m (Fig. 1). Recovery in Hole 817A was 84.8%. Drilling recovered a nearly continuous upper Miocene to Pleistocene sequence of highly bioturbated pelagic oozes that become increasingly intercalated with micritic periplatform deposits from the upper Pliocene to the Pleistocene. Most samples contain pyrite and pyritized foraminifers. As with microfossil samples from Hole 811A, preservation in samples from Hole 817A varied depending on the amount of micritic bank-derived material present in the sediments.

Lithologic facies at all three sites indicate a deepening of water depths from upper bathyal (200–600 m) in the middle Miocene to modern middle to lower bathyal depths (1000–2000 m) (Katz and Miller, this volume). This deepening resulted from the subsidence of the rift-bounded Queensland Plateau (Burns, Andrews, et al., 1973; Davies, McKenzie, Palmer-Julson, et al., 1991).

METHODOLOGY

Splits of each sample were oven-dried at 60°C, weighed, and disaggregated in a hot solution of buffered sodium hexametaphosphate. Samples were washed over a 63 μm sieve and dried at 60°C.

The planktonic foraminifer *Globigerinoides ruber* was used to estimate sea surface carbon and oxygen isotopic values. Although *G. ruber* may migrate within and below the mixed layer, it calcifies most of its test near the surface and therefore records near-surface-

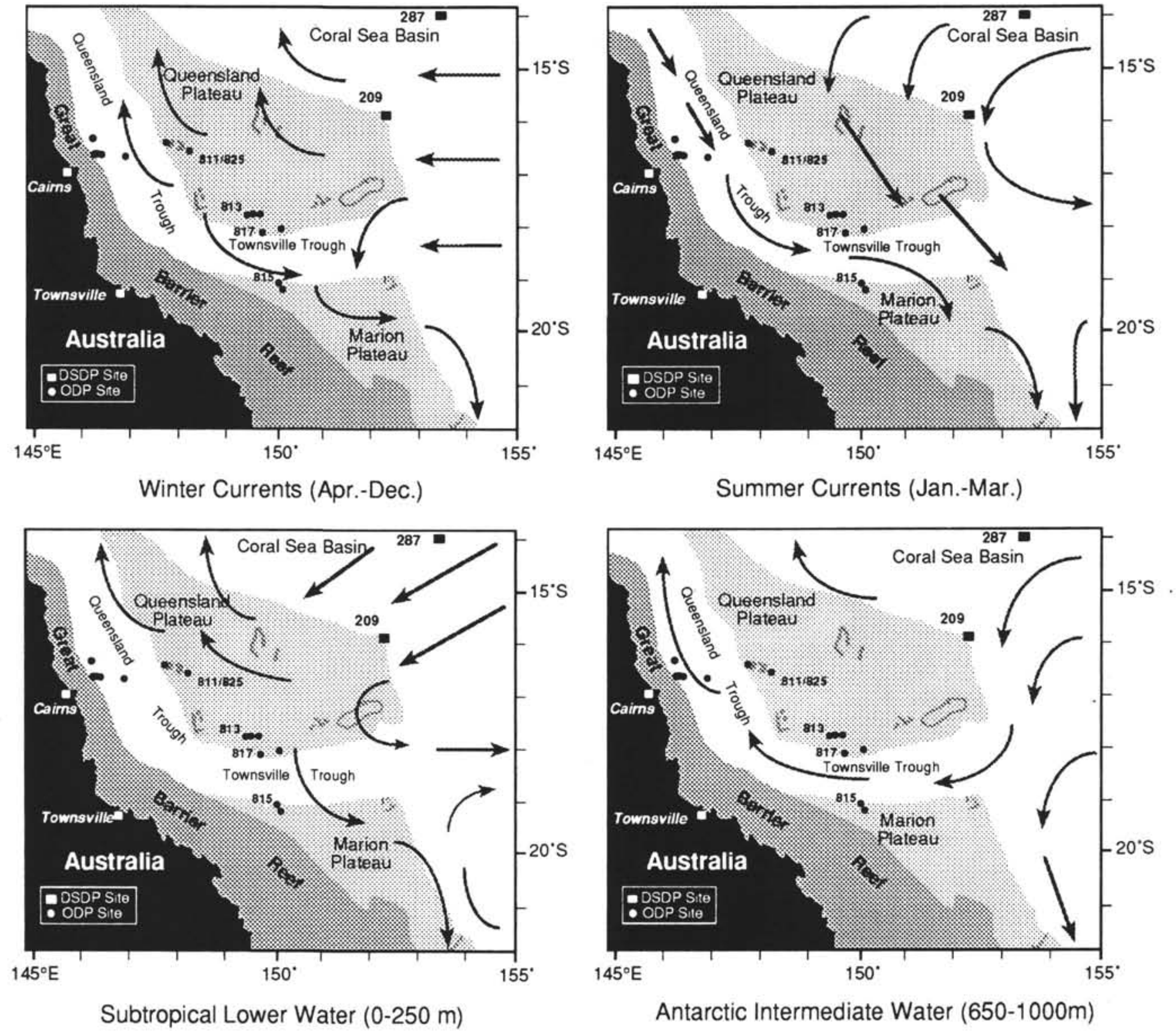

Figure 2. Generalized circulation patterns for the western Coral Sea (after Pickard et al., 1977). The winter currents are the normal surface-water flow paths for most of the year, with the summer surface-water current flow representing the northwest monsoon. For both winter and summer, the dominant current flowing to the west is the South Equatorial Current (SEC).

water isotopic values (Duplessy et al., 1981; Erez and Honjo, 1981; Goodney et al., 1980; Savin et al., 1975). *G. ruber* samples were chosen from the 180–250 μm size fraction to minimize changes in the isotopic signal resulting from age variations of individual planktonic foraminiferal specimens (Duplessy et al., 1981; Killingley et al., 1981). Offsets from isotopic equilibrium values with seawater (vital effects) for *G. ruber* are not well characterized, and estimates range from –0.2 to –0.6‰ (Fairbanks et al., 1982; Duplessy et al., 1981; Deuser, 1987). Because of the inconsistencies in the estimates, no corrections were applied to the data.

Cibicidoides spp., from the >250 size fraction, were used to estimate deep-water isotopic conditions. *Cibicidoides* spp. has been found to precipitate its calcite out of equilibrium with seawater. The average offset for $\delta^{18}O$ is –1.02‰ (Graham et al., 1981). The offset for $\delta^{13}C$ is –1.11‰, which is near the equilibrium offset between dissolved bicarbonate and calcium carbonate at temperatures similar to those found in deep waters (Emrich et al., 1970; Graham et al., 1981), and, accordingly, Cibicidoides is a good indicator of equilibrium ^{13}C values. The different species of Cibicidoides show similar isotopic values within the error of the method, and, therefore, the results from different species are comparable (Woodruff et al., 1980).

Picked samples were sonicated for one minute in methanol to remove any contamination in the tests, such as coccoliths. Hydrogen peroxide was used to remove organic matter. This treatment was continued until the reaction ceased (Grossman and Ku, 1986). The addition of hydrogen peroxide to foraminiferal samples has been found to have no effect on the isotopic values (Ganssen, 1981). Once the organic matter was removed, the samples were washed in distilled water, followed by methanol, and allowed to dry.

For isotopic analysis, approximately 0.5 mg of sample was placed into a glass sample boat and crushed in methanol. The samples were reacted at 82°C in a 100% phosphoric acid common acid bath using a VG Isogas Autocarbonate preparation system. The gas evolved was cryogenically distilled to remove water produced in the reaction. Samples were run on a VG precision isotope ratio mass (PRISM). Accuracy of the laboratory standard, Carrara marble, was ±0.07‰

for oxygen and ±0.03‰ for carbon. Precision for replicate analysis of foraminiferal samples was ±0.10‰ for oxygen and ±0.15‰ for carbon. All values obtained are in standard δ notation, tabulated relative to PDB. The data are presented in Tables 1 through 3.

STRATIGRAPHY

For the Leg 133 sites, a stratigraphy and sedimentation rate model was developed using shipboard biostratigraphic data (Davies, McKenzie, Palmer-Julson, et al., 1991), stable isotopic data (this paper), and strontium isotopic data (McKenzie et al., this volume) (Table 4, Fig. 3). Because of the type of sediments collected from Holes 811A and 817A, magnetic reversal data measured on the cores were not useful for stratigraphic correlation.

Oxygen isotope data from Site 811 were correlated with well-dated isotopic records from Site 586 on the Ontong Java Plateau (Whitman and Berger, 1992) and Site 590 on Lord Howe Rise (Elmstrom and Kennett, 1986). The three records are similar in the interval from 2.0 to 3.5 Ma. Thus, because the isotopic excursions have well-constrained ages in the Site 586 and 590 sediments, they can be used to further confine ages at Site 811 (Table 4). Three prominent isotopic shifts were used: 2.4, 3.3, and 3.6 Ma (Fig. 4).

Within the upper Miocene sediments at Site 811, $\delta^{13}C$ values shift 1‰ to lighter values (Figs. 5 through 8). This shift is the globally well-documented late Miocene carbon shift, which has been estimated to occur between 6.2 and 6.4 Ma (Bender et al., 1981; Vincent et al., 1980). The initial decrease in $\delta^{13}C$ can be placed either at 122 or 125 mbsf because of uncertainties about where this increase began. The point at which the $\delta^{13}C$ ceases to decrease occurs at 88 mbsf. As a result of uncertainties in the initial depth of the $\delta^{13}C$ event, the 6.2 Ma datum placement has an error of 3 m (104–107 mbsf). Using the average sedimentation rate for this interval, this corresponds to an error of 120 k.y. The magnitude of the shift is 1.0 to 1.3‰ in the planktonic foraminifers and 0.9‰ in the benthic foraminifers. The carbon shift provides another stratigraphic marker at Site 811.

For most of the section, the different dating methods at Site 811 correlate well (Table 4). The exception is the interval from 2.4 to 4 Ma (50–100 mbsf), where the strontium and stable isotopic dates are younger than the biostratigraphic ages. The sediments in this interval appear to have been deposited rapidly as a result of increased downslope transport from the Queensland Plateau. This rapid deposition is seen at other Leg 133 sites, and we think it is responsible for the observed age discrepancies. The sediments within this interval are coarse; therefore, winnowing also may have biased the location of nannofossil datums. Evidence suggests that similar age discrepancies occur at Site 817, although the lack of isotopic dates in the younger portion of the section makes identification uncertain. Between 2.4 and 4 Ma, greater weight was given to the isotopic ages (both Sr and O) when developing the sedimentation rate model, as we think they indicate actual sedimentary ages more closely. Hiatuses may occur in Core 133-811A-3H (lower Pleistocene), as the expected sequence of nannofossil biohorizons was not seen in the lower Pleistocene (Davies, McKenzie, Palmer-Julson, et al., 1991). Increased amounts of large size fraction, bank-derived carbonate indicate the possibility of winnowing in this interval.

Sedimentation rates vary at Site 811, mainly owing to changing fluxes of bank-derived material from the Queensland Plateau. During much of the Pliocene, when the influx of bank-derived carbonates was high but variable, the sedimentation rates fluctuated between 1.4 and 12 cm/k.y. Within the dominantly pelagic interval, the sedimentation rates are relatively constant, averaging 1.5 cm/k.y. (Fig. 3).

Drilling in Hole 817A provided an extended Pleistocene and upper Pliocene sequence of pelagic sediments. From 0 to 47 mbsf (0–1 Ma), the sedimentation rate is 5.2 cm/k.y.; from 47 to 75 mbsf (1–2.3 Ma) 2.1 cm/k.y.; and from 75 to 151 mbsf (2.3–3.5 Ma) it is 6.2 cm/k.y. (Fig. 3). Sedimentation rates in the lower Pliocene and upper Miocene sequence are much lower, 1.6 cm/k.y. (151 to 196 mbsf; 3.5–6.2 Ma;

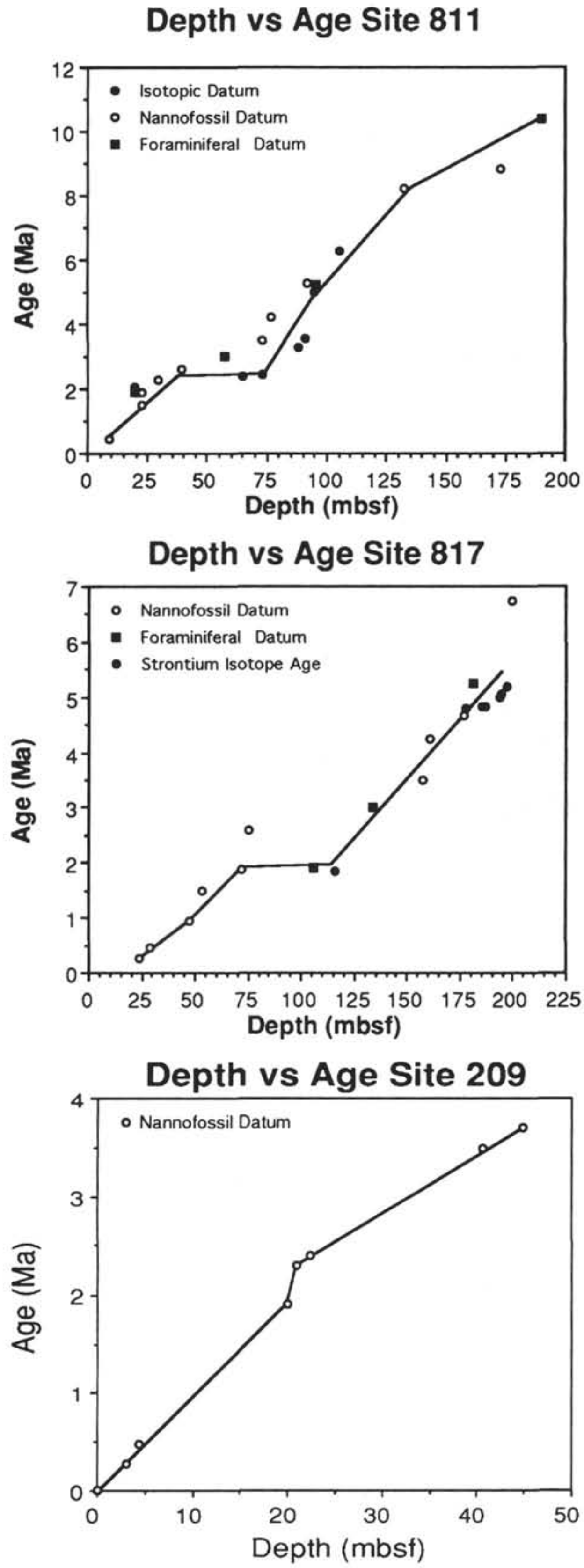

Figure 3. Depth vs. age plots for Sites 811, 817, and 209. Lines on the figure are the visually best fit lines to the data from which the sedimentation rates were calculated. Isotopic dates were weighted most strongly, followed by nannofossil datums. Foraminiferal datums were considered to be the coarsest ages obtained.

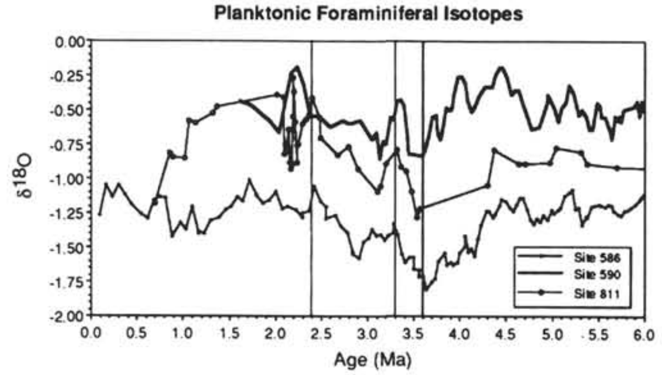

Figure 4. Correlation of isotopic data for Sites 590, 586, and 811, to show how the planktonic oxygen isotopic data were used to refine the stratigraphy of Site 811. All data were smoothed using a three-point running average. Data from Site 586 were obtained on *G. sacculifer*, whereas those of Sites 811 and 590 were from *G. ruber*.

Fig. 3). Carbon isotopic data for both planktonic and benthic foraminifers indicate a hiatus because of the absence of the late Miocene age carbon shift in the isotopic record. Strontium isotope dating of foraminiferal samples from this interval confirms that part of the record is missing, possibly as much as 4 m.y. (McKenzie et al., this volume; Table 4). Below the hiatus, foraminiferal specimens are more overgrown, infilled, and reworked. Despite the expanded nature of the section, it appears that downslope transport and reworking of sediments from the plateau were significant, thus decreasing the accuracy of the dating methods (especially from 2.4 to 4 Ma).

Poor recovery, drilling disturbance, and the absence of paleomagnetic reversal data at Site 209 made it difficult to establish a high-resolution stratigraphic model for calculating ages (Burns, Andrews, et al., 1973). However, for the Pleistocene to Holocene samples, excellent microfossil preservation permitted good placement of nannofossil and foraminiferal datums. The age vs. depth plot for Site 209 indicates that the sedimentation rates were 1.11 cm/k.y. until 2 Ma; 0.4 cm/k.y. until 2.4 Ma; and afterward increase to 1.53 cm/k.y. below 2.4 Ma (Fig. 3). A hiatus exists at Site 209 between Cores 5H and 6H (45 mbsf), estimated to be from 11 to 3.3 Ma (Kennett, 1973). This hiatus makes it difficult to determine stratigraphic ages for the sediments in Core 6H, although nannofossils from this core broadly place the sediment in the upper middle Miocene (approximately 11 Ma).

In Figures 5 through 8, biostratigraphic datums for Holes 811A and 817A and Site 209 are shown along with the isotopic data for both planktonic and benthic foraminifers.

RESULTS

Sample Preservation

Microfossil preservation in Holes 811A and 817A varied inversely with the amount of platform-derived carbonate in the samples. The dissolution and reprecipitation of bank carbonates increases the extent of calcite overgrowths and infilling. Samples from intervals having elevated bank carbonate compositions were not used if even

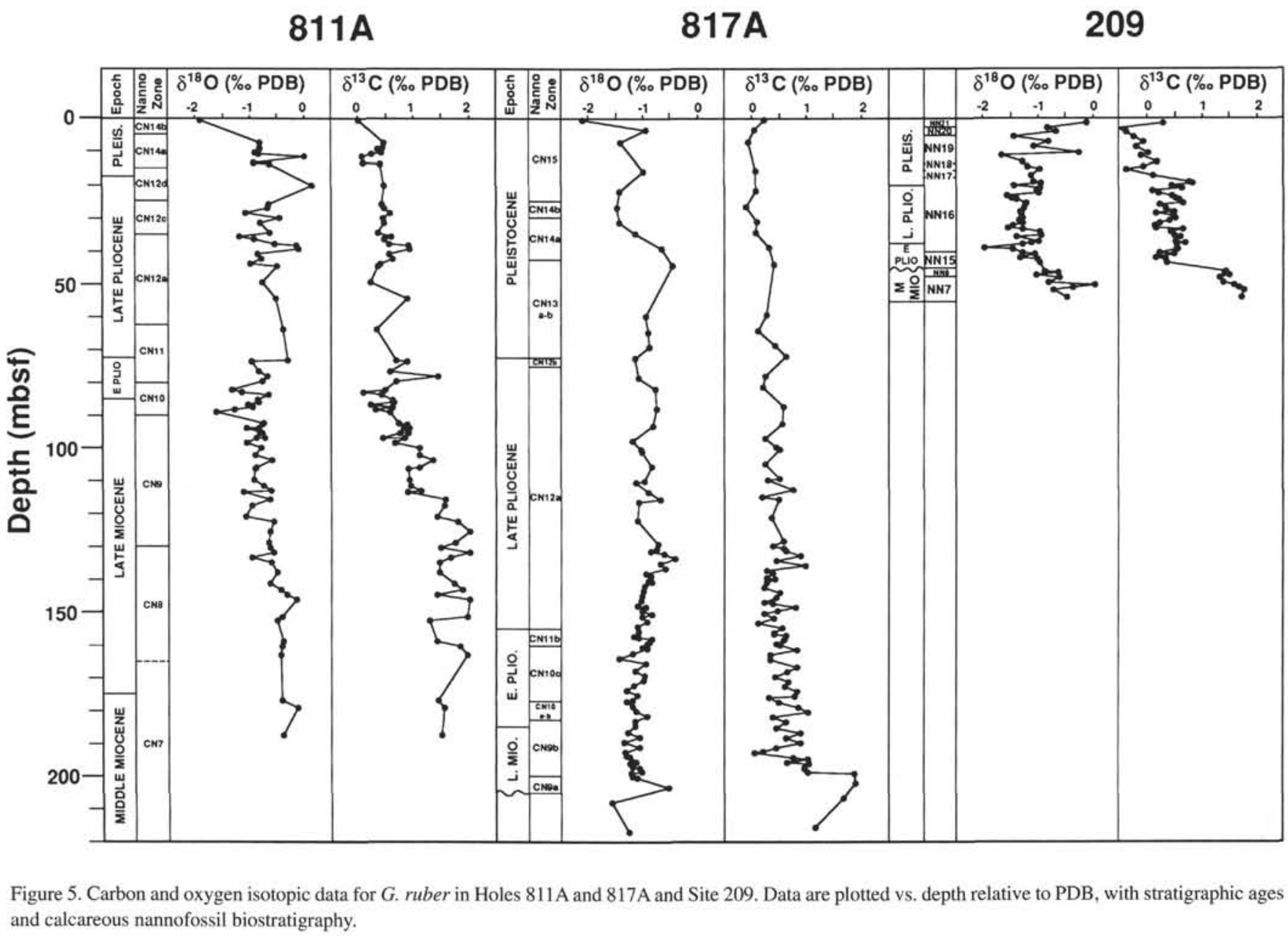

Figure 5. Carbon and oxygen isotopic data for *G. ruber* in Holes 811A and 817A and Site 209. Data are plotted vs. depth relative to PDB, with stratigraphic ages and calcareous nannofossil biostratigraphy.

Table 1. Hole 811A isotopic data.

Core, sample, interval (cm)	Depth (mbsf)	Age (Ma)	∂18O (PDB) *G. ruber*	∂13C (PDB) *G. ruber*	∂18O (PDB) *Cibicidoides sp.*	∂13C (PDB) *Cibicidoides sp.*	∂18O (PDB) *Cibicidoides sp. (eq.)*	Δ∂18O (PDB)	Δ∂13C (PDB)
1H-1, 4 - 6	0.04	0.00	-1.92	-0.02	2.44	0.84	3.55	5.48	0.87
2H-1, 125-127	6.75	0.70	-0.81	0.44	3.10	1.13	4.21	5.01	0.69
2H-2, 131-133	8.31	0.86	-0.81	0.42	2.34	1.55	3.45	4.26	1.13
2H-3, 12 - 14	8.62	0.89	-0.82	0.32	2.68	1.29	3.79	4.61	0.97
2H-3, 125-127	9.80	1.02	-0.90	0.39	2.82	1.68	3.93	4.84	1.30
2H-4, 26-28	10.26	1.06	-0.84	0.21					
2H-4, 96-98	10.96	1.14	0.00	0.03	2.84	1.44	3.95	3.95	1.41
2H-5, 126-128	12.76	1.32	-0.93	0.06	2.82	1.20	3.93	4.86	1.14
2H-6, 20-22	13.20	1.37	-0.64	0.38					
3H-1, 9 - 11	15.09	1.56			2.67	1.14	3.78		
3H-2, 12 - 14	16.62	1.72			2.78	1.19	3.89		
3H-3, 145-147	19.45	2.02	0.14	0.44					
3H-4, 20-22	19.70	2.04			3.01	0.85	4.12		
3H-6, 130-132	23.80	2.08			2.33	1.15	3.44		
3H-7, 40-42	24.40	2.08			2.46	1.27	3.57		
4H-1, 85-87	25.35	2.09	-0.67	0.40	2.28	1.29	3.39	4.06	0.89
4H-2, 25-27	26.25	2.10	-0.68	0.44	2.80	1.07	3.91	4.59	0.62
4H-3, 26-28	27.76	2.11	-1.10	0.56	2.51	1.43	3.62	4.72	0.88
4H-4, 24-26	29.24	2.12	-0.47	0.42	3.15	0.96	4.26	4.73	0.54
4H-5, 43-45	30.93	2.14	-0.81	0.44	3.01	1.01	4.12	4.93	0.57
4H-7, 50-52	34.00	2.16	-0.64	0.32	3.07	0.86	4.18	4.81	0.54
5H-1, 85-87	34.85	2.17	-1.21	0.57	1.97	1.30	3.08	4.29	0.73
5H-2, 25-27	35.75	2.18	-0.94	0.44	2.75	1.26	3.86	4.79	0.82
5H-3, 20-22	37.20	2.19	-0.55	0.53	2.06	1.33	3.17	3.72	0.80
5H-3, 73-75	37.73	2.19	-0.14	0.88	2.38	1.26	3.49	3.63	0.38
5H-4, 29-31	38.79	2.20	-0.10	0.90	2.60	1.17	3.71	3.81	0.27
5H-5, 30-32	40.30	2.21	-0.87	0.52	2.04	1.22	3.15	4.02	0.70
5H-6, 25-27	41.75	2.23	-0.79	0.59					
5H-7, 25-27	43.25	2.24	-0.99	0.38	2.51	0.92	3.62	4.61	0.55
6H-1, 74-76	44.24	2.25	-0.50	0.33	2.10	0.58	3.21	3.71	0.25
6H-3, 23-25	46.68	2.27			2.06	1.28			
6H-4, 110-112	49.05	2.29	-0.77	0.19					
6H-5, 110-112	50.55	2.30			2.06	1.08			
7H-1, 90-92	53.90	2.33	-0.54	0.87	1.55	1.08	2.66	3.19	0.21
8H-1, 69-71	63.19	2.40	-0.39	0.31	2.21	0.83	3.32	3.70	0.52
9H-1, 71-73	72.71	2.48	-0.31	0.66	1.66	0.97	2.77	3.08	0.31
9H-1, 110-112	73.10	2.49	-0.97	0.86					
9H-3, 110-112	76.10	2.68	-0.84	0.55	1.74	0.88	2.85	3.69	0.33
9H-4, 110-112	77.60	2.79	-0.68	1.42	1.75	1.09	2.86	3.55	-0.33
9H-5, 16-18	78.16	2.83			1.83	1.31			
9H-5, 110-112	79.10	2.90	-0.78	0.66	1.90	0.87	3.01	3.79	0.21
10H-1, 42-44	81.92	3.10	-1.34	0.46	1.11	1.09	2.22	3.56	0.64
10H-1, 108-110	82.58	3.14	-1.16	0.05	1.93	1.03	3.04	4.20	0.98
10H-2, 41-43	83.41	3.20	-0.66	0.40					
10H-2, 108-110	84.08	3.25			1.38	0.41	2.49		
10H-3, 41-43	84.91	3.34	-0.86	0.59	2.06	0.88	3.17	4.03	0.29
10H-3, 109-111	85.59	3.42	-0.85	0.62	1.17	1.31	2.28	3.12	0.68
10H-4, 41-43	86.41	3.44	-1.04	0.20	1.67	1.11	2.78	3.81	0.90
10H-4, 108-110	87.08	3.49	-0.95	0.59	1.23	1.36	2.34	3.29	0.77
10H-5, 41-43	87.91	3.55	-1.28	0.28	2.20	0.81	3.31	4.59	0.53
10H-5, 100-102	88.50	4.30	-1.62	0.55	1.79	1.15	2.90	4.52	0.60
11H-1, 80-82	91.80	4.37	-0.76	0.70	1.08	1.52	2.19	2.95	0.83
11H-1, 111-113	92.11	4.45	-0.77	0.86	1.41	1.48	2.52	3.29	0.62
11H-2, 80-82	93.30	4.70	-0.84	0.80	1.55	1.25	2.66	3.49	0.45
11H-2, 111-113	93.61	4.80	-1.06	0.91	1.63	1.36	2.74	3.80	0.44
11H-3, 80-82	94.80	5.04	-0.77	0.89	1.28	1.42	2.39	3.16	0.54
11H-3, 111-113	95.11	5.13	-0.81	0.74	0.83	1.24	1.94	2.75	0.51
11H-4, 80-82	96.30	5.37	-0.73	0.42	1.80	0.97	2.91	3.65	0.55
11H-4, 111-113	96.61	5.47	-0.88	0.82	1.49	0.85	2.60	3.48	0.03
11H-5, 111-113	98.11	5.80	-1.06	0.65	1.77	1.21	2.88	3.94	0.55
11H-6, 111-113	99.61	6.05	-0.80	1.10	1.40	1.08	2.51	3.31	-0.02
12H-1, 110-112	101.60	6.14	-0.90	1.10	1.50	1.18	2.61	3.51	0.08
12H-2, 110-112	103.10	6.20	-0.60	1.33	2.02	1.29	3.13	3.73	-0.05
12H-3, 110-112	104.60	6.26			1.68	1.37			
12H-4, 55-56	105.30	6.28	-0.89	1.09	1.41	1.43	2.52	3.40	0.34
12H-4, 67-69	105.67	6.30	-0.92	0.88					
12H-5, 67-69	107.17	6.36			1.67	1.17			
12H-5, 110-112	107.60	6.38			0.86	1.48			
12H-6, 110-112	109.10	6.45	-0.93	0.90	1.59	1.14	2.70	3.63	0.24
13H-1, 109-111	111.09	6.55	-0.76	0.93	1.75	1.24	2.86	3.62	0.31
13H-2, 109-111	112.59	6.64	-0.61	1.11	1.85	1.25	2.96	3.57	0.14
13H-3, 15-17	113.15	6.67	-1.14	0.86					
13H-4, 80-82	115.30	6.75	-0.64	1.55	1.26	1.13	2.37	3.00	-0.42
13H-5, 109-111	117.09	6.89	-0.98	1.53	0.99	1.42	2.10	3.08	-0.11

Table 1 (continued).

Core, sample, interval (cm)	Depth (mbsf)	Age (Ma)	∂18O (PDB) *G. ruber*	∂13C (PDB) *G. ruber*	∂18O (PDB) *Cibicidoides sp.*	∂13C (PDB) *Cibicidoides sp.*	∂18O (PDB) *Cibicidoides sp. (eq.)*	Δ∂18O (PDB)	Δ∂13C (PDB)
14H-1, 110-112	120.60	7.07	-1.10	1.40	1.41	1.38	2.52	3.62	-0.02
14H-2, 110-112	122.10	7.17	-0.57	1.78	1.32	1.74	2.43	3.00	-0.03
14H-4, 110-112	125.10	7.34	-0.64	2.01	1.84	2.11	2.95	3.59	0.10
14H-6, 138-140	128.38	7.52	-0.66	1.75					
15H-1, 110-112	130.10	7.60	-0.64	1.48	1.42	1.53	2.53	3.17	0.05
15H-2, 110-112	131.60	7.69	-0.57	2.01	1.61	1.88	2.72	3.29	-0.13
15H-3, 110-112	133.10	7.77	-0.97	1.66					
15H-4, 80-82	134.30	7.83	-0.63	1.44					
15H-6, 110-112	137.60	8.03	-0.51	1.44					
16H-2, 80-82	140.80	8.21	-0.65	1.71	1.87	1.90	2.98	3.64	0.20
16H-2, 140-142	141.40	8.22			1.58	1.77			
16H-3, 110-112	142.60	8.30	-0.44	1.87	1.89	2.05	3.00	3.44	0.18
16H-4, 110-112	144.10	8.39	-0.32	1.41	1.66	1.78	2.77	3.10	0.37
16H-5, 110-112	145.60	8.47	-0.14	2.00	2.11	1.99	3.22	3.36	-0.01
16H-6, 110-112	146.10	8.55			1.20	1.55	2.31		
17H-2, 140-141	150.90	8.78	-0.41	1.97	1.31	2.02	2.42	2.83	0.05
17H-3, 110-112	152.10	8.86	-0.50	1.26	1.36	1.94	2.47	2.97	0.69
17H-4, 110-112	153.60	8.95			1.96	1.61	3.07		
17H-5, 110-112	155.10	9.00			1.47	1.59	2.58		
18H-1, 110-112	158.60	9.19	-0.40	1.41	1.87	2.23	2.98	3.38	0.81
18H-2, 110-112	160.10	9.28	-0.41	1.82	1.41	1.79	2.52	2.93	-0.04
18H-4, 79-81	162.79	9.93	-0.44	1.96	1.56	1.72	2.67	3.11	-0.25
19H-4, 93-95	172.43	10.03			-0.33	2.23	0.78		
19H-5, 110-112	174.10	10.17			1.35	1.38	2.46		
19H-7, 22-24	176.72	10.30	-0.42	1.43					
20H-2, 98-100	178.98	10.38	-0.12	1.54					
20H-3, 100-102	180.48	10.46			1.71	1.22	2.82		
20H-4, 97-99	181.97	10.75			1.38	1.74	2.49		
21H-1, 110-112	187.10	10.98	-0.40	1.49					

a moderate amount of diagenesis was observed. For pelagic sediments, preservation was good to very good for both planktonic and benthic foraminifers (Fig. 9). Sr/Ca data obtained from Hole 815A samples showed that diagenetic overgrowths and alterations of microfossils were not significant (Chivas, pers. comm., 1992). However, visually, Site 815 microfossils showed equal to greater infilling and overgrowths. Therefore, we feel that the isotopic data obtained from microfossils at Sites 811 and 817 provide an accurate estimate of paleoceanographic conditions from this area. Site 209 showed the best preservation of the sites studied. Well-preserved, dissolution-susceptible species were found in most samples. Overgrowths and infillings were minimal, apparently because of the absence of metastable bank-derived carbonate.

Oxygen Isotopes

Oxygen isotopic data were collected on *G. ruber* from Holes 811A and 817A and Site 209. Values range from –2 to 0‰ (Figs. 5 and 7).

In the upper Miocene (10–5.2 Ma), in Hole 811A there is a general decrease in $\delta^{18}O$ of –0.2‰ to –0.8‰, with fluctuations of up to 0.5‰ (Figs. 5 and 7). In the lower Pliocene sequence (5.2–3.4 Ma), in both Holes 811A and 817A rapid fluctuations of about 0.25‰ are seen in the oxygen isotopic values. At the lower Pliocene/upper Pliocene boundary (3.4 Ma), the oxygen isotopic values at Hole 811A decrease sharply by 1‰. This change is not seen in the Hole 817A data. In both Holes 811A and 817A, the upper Pliocene sediments are characterized by increasing values (~0.75‰) until 2.7 Ma, where they decrease by about 0.2‰ to the Pliocene/Pleistocene boundary (1.6 Ma) (Figs. 5 and 7).

The most dramatic change in the oxygen isotopic values occurs in the lower Pleistocene, where a 2.0‰ shift to lighter values is seen (Figs. 5 and 7).

Planktonic foraminiferal isotopic data at Site 209 are, until the upper Pliocene boundary (1.6 Ma), generally similar to those from Holes 811A and 817A. The main difference among the three sites is that the Pleistocene shift to lighter values is not seen here (Figs. 5 and 7).

Benthic isotopic data were collected from all three sites (Figs. 6 and 8). The range of values for Holes 811A and 817A is similar (~2‰), whereas the data for Site 209 are offset to higher values by about 1‰. In the upper Miocene, oxygen isotope values from Hole 811A fluctuate around 1.6‰, with variations of 0.5‰. In both Hole 811A and 817A data, the lower Pliocene is characterized by a slight increase in isotopic values of about 0.25‰. This increase also is seen in the Site 209 data, but here the change is greater (0.6‰). This general increase in $\delta^{18}O$ continues at all sites from the upper Pliocene through to the Holocene sequence.

In general, all isotopic data from Site 817 showed a lower amplitude signal than data from Sites 209 and 811, caused by a greater amount of downslope transport and reworking at this site. We postulate, therefore, that the data from Sites 811 and 209 are more indicative of the paleoceanographic trends in the vicinity of the Queensland Plateau.

Calculation of Paleotemperatures

Paleotemperatures were calculated from planktonic foraminifer oxygen isotope sequences using the equation,

$$T = 16.9 - 4.38\ (\delta_c - \delta_w) + 0.10\ (\delta_c - \delta_w)^2 \quad (1)$$

for temperatures estimated to be above 7°C (Shackleton and Kennett, 1975). Benthic foraminifer oxygen isotope data were used to calculate seafloor paleotemperatures using the equation,

$$T = 16.4 - 4.2\ (\delta_c - \delta_w) + 0.13\ (\delta_c - \delta_w)^2 \quad (2)$$

where temperatures are expected to be less than 7°C (Grossman, 1982). For both equations, T is temperature in °C, δ_c is the isotopic composition of biogenic carbonate precipitated in isotopic equilibrium with the surrounding seawater, and δ_w is the isotopic composition of the

Table 2. Hole 817A isotopic data.

Core, sample, interval (cm)	Depth (mbsf)	Age (Ma)	∂18O (PDB) *G. ruber*	∂13C (PDB) *G. ruber*	∂18O (PDB) *Cibicidoides sp.*	∂13C (PDB) *Cibicidoides sp.*	321.00 *Cibicidoides sp. (eq.)*	Δ∂18O (PDB)	Δ∂13C (PDB)
1H-1, 82-84	1.02	0.10	-0.12	0.28	3.26	0.57	4.37	4.49	0.29
1H-2, 76-78	2.46	0.23	-0.85	-0.59	3.37	0.26	4.48	5.33	0.85
1H-3, 70-72	3.90	0.37	-0.69	-0.38	3.96	0.58	5.07	5.76	0.95
1H-4, 60-62	5.30	0.50	-1.45	-0.25	2.24	0.34	3.35	4.80	0.59
1H-5, 70-72	6.90	0.65	-0.83	-0.07	3.14	0.83	4.25	5.08	0.90
1H-6, 70-72	8.40	0.79	-1.08	-0.22	3.27	0.68	4.38	5.46	0.90
2H-1, 110-112	10.10	0.95	-0.27	0.01	2.62	0.10	3.73	4.00	0.08
2H-2, 60-62	11.10	1.05	-1.67	-0.12	2.78	0.84	3.89	5.56	0.96
2H-3, 70-72	12.70	1.20	-1.28	0.18	3.17	0.79	4.28	5.57	0.61
2H-4, 69-70	14.19	1.34	-1.21	-0.07	3.50	0.83	4.61	5.82	0.90
2H-5, 21-23	15.21	1.43	-0.98	-0.38	3.19	0.34	4.30	5.28	0.72
2H-6, 44-46	16.94	1.60	-1.14	0.10	2.89	0.81	4.00	5.14	0.71
3H-1, 66-68	18.68	1.76	-1.10	0.76	2.97	0.86	4.08	5.19	0.10
3H-1, 117-119	19.17	1.81	-0.96	0.83	2.88	0.95	3.99	4.95	0.12
3H-2, 45-46	19.95	1.88	-1.44	0.43	2.92	0.25	4.03	5.47	-0.18
3H-2, 110-112	20.60	2.00	-0.98	0.61	2.90	0.99	4.01	4.99	0.38
3H-3, 40-42	21.40	2.15	-1.05	0.09	3.43	0.42	4.54	5.59	0.33
3H-3, 106-108	22.06	2.27	-1.00	0.19	2.97	0.16	4.08	5.07	-0.03
3H-4, 40-42	22.90	2.42	-1.59	0.44	2.46	0.83	3.57	5.15	0.38
3H-4, 106-108	23.56	2.46	-1.49	0.50	2.41	0.47	3.52	5.01	-0.02
3H-5, 28-30	24.28	2.51	-1.40	0.58	2.48	0.82	3.59	4.99	0.25
3H-5, 107-109	25.07	2.56	-1.23	0.64	2.40	0.91	3.51	4.74	0.27
3H-6, 28-30	25.78	2.60	-1.25	0.21	2.60	0.48	3.71	4.96	0.27
3H-6, 107-109	26.57	2.65	-1.26	0.33	2.93	0.81	4.04	5.30	0.48
4H-1, 91-93	27.91	2.74	-1.34	0.48	2.46	0.50	3.57	4.91	0.02
4H-1, 128-130	28.28	2.76	-1.32	0.15	2.57	0.54	3.68	5.00	0.39
4H-2, 33-35	28.83	2.79	-1.32	0.47	2.61	0.70	3.72	5.04	0.23
4H-2, 121-123	29.71	2.85	-1.28	0.50	2.56	0.82	3.67	4.95	0.32
4H-3, 37-39	30.37	2.89	-1.36	0.39	2.68	0.82	3.79	5.15	0.43
4H-3, 120-122	31.20	2.94	-1.28	0.22	2.64	0.72	3.75	5.03	0.50
4H-4, 37-39	31.87	2.99	-1.48	0.12	2.76	0.80	3.87	5.35	0.69
4H-4, 120-122	32.70	3.04	-1.57	0.14	2.39	0.80	3.50	5.08	0.66
4H-5, 27-29	33.27	3.07	-1.30	0.65	2.48	0.98	3.59	4.89	0.33
4H-5, 102-104	34.02	3.12	-0.97	0.42	2.28	0.79	3.39	4.36	0.37
4H-6, 36-38	34.86	3.18	-0.96	0.48	2.67	0.62	3.78	4.74	0.14
4H-6, 107-109	35.57	3.22	-1.40	0.60	2.03	0.62	3.14	4.54	0.02
5H-1, 91-93	36.91	3.31	-0.99	0.51	2.65	0.69	3.76	4.74	0.18
5H-1, 137-139	37.37	3.33	-1.13	0.68	2.23	0.65	3.34	4.47	-0.03
5H-2, 31-33	37.81	3.36	-1.29	0.53	2.54	1.28	3.65	4.94	0.75
5H-2, 113-115	38.63	3.41	-1.99	0.54	2.26	0.91	3.37	5.36	0.37
5H-3, 27-29	39.27	3.45	-1.46	0.56	2.22	0.98	3.33	4.79	0.42
5H-3, 117-119	40.17	3.51	-1.30	0.22	2.59	1.42	3.70	5.00	1.20
5H-4, 27-29	40.77	3.55	-1.06	0.49	2.35	0.97	3.46	4.52	0.48
5H-4, 117-119	41.67	3.61	-1.34	0.15	2.02	1.30	3.13	4.47	1.15
5H-5, 36-38	42.36	3.65	-1.03	0.33	2.34	1.11	3.45	4.48	0.78
5H-5, 122-124	43.22	3.70	-0.98	0.34	1.96	0.58	3.07	4.05	0.24
6H-1, 66-68	45.96		-0.88	1.43	1.83	1.08	2.94	3.82	-0.35
6H-1, 122-124	46.52		-0.65	1.41	2.46	1.55	3.57	4.22	0.14
6H-2, 37-39	47.17		-1.04	1.50	2.16	1.53	3.27	4.31	0.03
6H-2, 122-124	48.02		-0.62	1.32	1.96	1.33	3.07	3.69	0.01
6H-3, 37-39	48.87				2.01	1.47	3.12		
6H-3, 122-124	49.52		-0.81	1.39	1.72	1.23	2.83	3.64	-0.16
6H-4, 26-28	50.06		0.02	1.58	1.91	1.45	3.02	3.00	-0.13
6H-4, 123-125	51.03		-0.37	1.67	2.21	1.43	3.32	3.69	-0.23
6H-5, 26-28	51.56		-0.73	1.76	1.99	1.75	3.10	3.83	-0.01
6H-5, 123-125	52.53				1.00	1.05	2.11		
6H-6, 32-34	53.12				1.79	1.58	2.90		
6H-6, 122-124	54.02		-0.49	1.72					

Table 2 (continued).

Core, sample, interval (cm)	Depth (mbsf)	Age (Ma)	∂18O (PDB) *G. ruber*	∂13C (PDB) *G. ruber*	∂18O (PDB) *Cibicidoides sp.*	∂13C (PDB) *Cibicidoides sp.*	∂18O (PDB) *Cibicidoides sp. (eq.)*	Δ∂18O (PDB)	Δ∂13C (PDB)
19H-1, 70-72	167.90	4.44	-1.14	0.85	1.92	1.31	3.03	4.16	0.46
19H-2, 70-72	169.40	4.53	-0.95	0.67	1.75	0.93	2.86	3.82	0.26
19H-3, 70-72	170.90	4.63	-0.98	0.43	1.96	1.00	3.07	4.05	0.57
19H-4, 70-72	172.40	4.72	-1.15	0.68	1.80	1.13	2.91	4.07	0.45
19H-5, 70-72	173.90	4.81	-1.28	0.63	1.97	1.00	3.08	4.36	0.36
19H-6, 70-72	175.40	4.91	-1.09	0.85	1.59	1.05	2.70	3.79	0.20
19H-7, 70-72	176.90	5.00	-1.19	0.80	1.96	1.03	3.07	4.26	0.23
20H-1, 70-72	177.40	5.03	-1.29	0.32	1.45	0.42	2.56	3.86	0.11
20H-2, 70-72	178.90	5.13	-1.18	0.51	1.58	0.96	2.69	3.87	0.45
20H-3, 70-72	180.40	5.22	-1.11	0.86	1.74	0.96	2.85	3.96	0.11
20H-4, 70-72	181.90	5.31	-0.91	1.05	1.79	1.10	2.90	3.81	0.05
20H-5, 70-72	183.40	5.41	-1.14	0.39	1.55	0.81	2.66	3.80	0.42
20H-6, 70-72	184.90	5.50	-1.14	0.64	1.61	0.94	2.72	3.85	0.30
21H-1, 70-72	186.90	5.63	-1.27	0.47	1.79	0.62	2.90	4.18	0.16
21H-2, 70-72	188.40	5.72	-1.05	0.91	1.91	1.08	3.02	4.07	0.17
21H-3, 70-72	189.90	5.81	-1.34	0.64	1.48	1.05	2.59	3.92	0.41
21H-4, 70-72	191.40	5.91	-1.05	0.91	1.90	0.85	3.01	4.06	-0.05
21H-5, 70-72	192.90	6.00	-1.31	0.47	1.68	0.89	2.79	4.11	0.43
21H-6, 20-22	193.90	6.06	-1.30	0.21	1.68	1.02	2.79	4.09	0.81
21H-6, 70-72	194.40	6.09	-1.23	0.06	1.71	0.65	2.82	4.05	0.59
21H-7, 70-72	195.60	6.17			1.80	1.21	2.91		
22H-1, 20-22	195.90	6.19	-1.11	0.77					
22H-1, 70-72	196.40	6.22	-1.23	1.05	0.38	0.96	1.49	2.72	-0.09
22H-2, 20-22	197.40	6.28	-1.18	0.66					
22H-2, 70-72	197.90	6.31	-1.04	1.06	0.74	1.43	1.85	2.88	0.38
22H-3, 20-22	198.90	6.38	-1.00	0.97	1.68	1.19	2.79	3.79	0.22
22H-3, 70-72	199.40	6.41	-1.20	0.97	1.35	1.15	2.46	3.66	0.17
22H-4, 20-22	200.40	6.47	-1.18	1.04	1.48	1.11	2.59	3.77	0.07
22H-4, 70-72	200.90	6.50	-1.10	1.89	0.59	2.11	1.70	2.80	0.22
22H-6, 70-72	203.90		-0.50	1.91	1.25	1.91	2.36	2.86	0.01
22H-7, 20-22	204.90				1.70	2.31	2.81		
23H-2, 133-135	208.30		-1.56	1.70	0.99	1.88	2.10	3.67	0.18
24H-2, 130-132	217.50		-1.24	1.19	1.19	2.10	2.30	3.54	0.91

seawater in which the carbonate was precipitated. The estimation of δ_w contributes much of the error in paleotemperature calculations. δ_w varies through time as the result of two main factors: changes in the oceanic reservoir of ^{16}O as a result of storage in continental ice ("ice volume effect"), and localized changes in $\delta^{18}O$ caused by salinity variations related to river runoff, evaporation-precipitation, and surface currents. The latter effect is usually small and, therefore, is often ignored in paleoceanographic reconstructions (Savin, 1977). The ice volume effect, on the other hand, is significant, as the isotopic composition of seawater is thought to differ by as much as 1.0‰ between a pre-glacial, ice-free earth and modern glacial conditions (Craig, 1965; Shackleton and Kennett, 1975). As a result, the isotopic difference caused by changes in ice volume, in many cases, can be more significant than those caused by changes in temperature. Thus, the difficulty is in estimating the amount of ice present during the geologic past, where continental ice was not as extensive as today. By comparing low- and middle-latitude planktonic and benthic isotopic data, it is possible to estimate changes of δ_w (Miller et al., 1987; Shackleton and Opdyke, 1973; Vincent and Berger, 1981). These estimates can be constrained further by sedimentary evidence about the extent of glaciations. To estimate the paleotemperatures as accurately as possible for this study, we used variable values for δ_w from the late Miocene to Holocene in the paleotemperature calculations. These values were obtained from estimates of ice volume found in the literature (e.g., Feary et al., 1991). Despite the attempt to remove the effect of ice-volume variations, remnants still are visible in our calculations of seafloor temperatures, as they indicate temperatures less than zero in the late Pliocene and Pleistocene, which is not possible (Fig. 10). This is the result of an increased frequency of glacial fluctuations after the early Pliocene (400 k.y. prior and 100 k.y. after), which makes it difficult to accurately account for ice volumes. As a result, the accuracy of the calculated seafloor and sea-surface temperatures in this interval is suspect (Fig. 10).

Other difficulties in the accurate estimation of paleotemperatures are (1) the presence of diagenetic alteration of biogenic carbonate; (2) disequilibrium precipitation of carbonate (vital effects); and (3) variations in the depth stratification of planktonic foraminifers in the water column because of seasonality and age of the organisms (Berger and Gardner, 1975; Savin, 1977; Shackleton and Opdyke, 1973; Vincent and Berger, 1981). The importance of these effects varies between planktonic and benthic foraminifers. In planktonic foraminifers, vital effects are present but are relatively small (Berger and Gardner, 1975; Curry and Matthews, 1981; Duplessy et al., 1981; Killingley et al., 1981; Williams et al., 1981), although variations in isotopic composition, caused by seasonality of growth and migration in the water column with age, can be significant (Savin et al., 1985; Savin et al., 1975). This effect is reduced if samples are chosen from a small size range (Duplessy et al., 1981; Killingley et al., 1981; Savin and Douglas, 1973). On the other hand, vital effects in benthic foraminifers are significant and may produce as much as 3.0‰ offset from ambient seawater (Graham et al., 1981; Savin, 1977; Shackleton, 1974; Vincent et al., 1980; Woodruff et al., 1980). However, these offsets are fairly well known for many benthic foraminiferal species, and, therefore, values can be adjusted to equilibrium (Graham et al., 1981; Woodruff et al., 1980). As a result, monospecific samples must be used when analyzing benthic foraminifers.

Table 3. Site 209 isotopic data.

Core, sample, interval (cm)	Depth (mbsf)	Age (Ma)	∂18O (PDB) *G. ruber*	∂13C (PDB) *G. ruber*	∂18O (PDB) *Cibicidoides sp.*	∂13C (PDB) *Cibicidoides sp.*	321.00 *Cibicidoides sp. (eq.)*	Δ∂18O (PDB)	Δ∂13C (PDB)
1H-1, 82-84	1.02	0.10	-0.12	0.28	3.26	0.57	4.37	4.49	0.29
1H-2, 76-78	2.46	0.23	-0.85	-0.59	3.37	0.26	4.48	5.33	0.85
1H-3, 70-72	3.90	0.37	-0.69	-0.38	3.96	0.58	5.07	5.76	0.95
1H-4, 60-62	5.30	0.50	-1.45	-0.25	2.24	0.34	3.35	4.80	0.59
1H-5, 70-72	6.90	0.65	-0.83	-0.07	3.14	0.83	4.25	5.08	0.90
1H-6, 70-72	8.40	0.79	-1.08	-0.22	3.27	0.68	4.38	5.46	0.90
2H-1, 110-112	10.10	0.95	-0.27	0.01	2.62	0.10	3.73	4.00	0.08
2H-2, 60-62	11.10	1.05	-1.67	-0.12	2.78	0.84	3.89	5.56	0.96
2H-3, 70-72	12.70	1.20	-1.28	0.18	3.17	0.79	4.28	5.57	0.61
2H-4, 69-70	14.19	1.34	-1.21	-0.07	3.50	0.83	4.61	5.82	0.90
2H-5, 21-23	15.21	1.43	-0.98	-0.38	3.19	0.34	4.30	5.28	0.72
2H-6, 44-46	16.94	1.60	-1.14	0.10	2.89	0.81	4.00	5.14	0.71
3H-1, 66-68	18.68	1.76	-1.10	0.76	2.97	0.86	4.08	5.19	0.10
3H-1, 117-119	19.17	1.81	-0.96	0.83	2.88	0.95	3.99	4.95	0.12
3H-2, 45-46	19.95	1.88	-1.44	0.43	2.92	0.25	4.03	5.47	-0.18
3H-2, 110-112	20.60	2.00	-0.98	0.61	2.90	0.99	4.01	4.99	0.38
3H-3, 40-42	21.40	2.15	-1.05	0.09	3.43	0.42	4.54	5.59	0.33
3H-3, 106-108	22.06	2.27	-1.00	0.19	2.97	0.16	4.08	5.07	-0.03
3H-4, 40-42	22.90	2.42	-1.59	0.44	2.46	0.83	3.57	5.15	0.38
3H-4, 106-108	23.56	2.46	-1.49	0.50	2.41	0.47	3.52	5.01	-0.02
3H-5, 28-30	24.28	2.51	-1.40	0.58	2.48	0.82	3.59	4.99	0.25
3H-5, 107-109	25.07	2.56	-1.23	0.64	2.40	0.91	3.51	4.74	0.27
3H-6, 28-30	25.78	2.60	-1.25	0.21	2.60	0.48	3.71	4.96	0.27
3H-6, 107-109	26.57	2.65	-1.26	0.33	2.93	0.81	4.04	5.30	0.48
4H-1, 91-93	27.91	2.74	-1.34	0.48	2.46	0.50	3.57	4.91	0.02
4H-1, 128-130	28.28	2.76	-1.32	0.15	2.57	0.54	3.68	5.00	0.39
4H-2, 33-35	28.83	2.79	-1.32	0.47	2.61	0.70	3.72	5.04	0.23
4H-2, 121-123	29.71	2.85	-1.28	0.50	2.56	0.82	3.67	4.95	0.32
4H-3, 37-39	30.37	2.89	-1.36	0.39	2.68	0.82	3.79	5.15	0.43
4H-3, 120-122	31.20	2.94	-1.28	0.22	2.64	0.72	3.75	5.03	0.50
4H-4, 37-39	31.87	2.99	-1.48	0.12	2.76	0.80	3.87	5.35	0.69
4H-4, 120-122	32.70	3.04	-1.57	0.14	2.39	0.80	3.50	5.08	0.66
4H-5, 27-29	33.27	3.07	-1.30	0.65	2.48	0.98	3.59	4.89	0.33
4H-5, 102-104	34.02	3.12	-0.97	0.42	2.28	0.79	3.39	4.36	0.37
4H-6, 36-38	34.86	3.18	-0.96	0.48	2.67	0.62	3.78	4.74	0.14
4H-6, 107-109	35.57	3.22	-1.40	0.60	2.03	0.62	3.14	4.54	0.02
5H-1, 91-93	36.91	3.31	-0.99	0.51	2.65	0.69	3.76	4.74	0.18
5H-1, 137-139	37.37	3.33	-1.13	0.68	2.23	0.65	3.34	4.47	-0.03
5H-2, 31-33	37.81	3.36	-1.29	0.53	2.54	1.28	3.65	4.94	0.75
5H-2, 113-115	38.63	3.41	-1.99	0.54	2.26	0.91	3.37	5.36	0.37
5H-3, 27-29	39.27	3.45	-1.46	0.56	2.22	0.98	3.33	4.79	0.42
5H-3, 117-119	40.17	3.51	-1.30	0.22	2.59	1.42	3.70	5.00	1.20
5H-4, 27-29	40.77	3.55	-1.06	0.49	2.35	0.97	3.46	4.52	0.48
5H-4, 117-119	41.67	3.61	-1.34	0.15	2.02	1.30	3.13	4.47	1.15
5H-5, 36-38	42.36	3.65	-1.03	0.33	2.34	1.11	3.45	4.48	0.78
5H-5, 122-124	43.22	3.70	-0.98	0.34	1.96	0.58	3.07	4.05	0.24
6H-1, 66-68	45.96		-0.88	1.43	1.83	1.08	2.94	3.82	-0.35
6H-1, 122-124	46.52		-0.65	1.41	2.46	1.55	3.57	4.22	0.14
6H-2, 37-39	47.17		-1.04	1.50	2.16	1.53	3.27	4.31	0.03
6H-2, 122-124	48.02		-0.62	1.32	1.96	1.33	3.07	3.69	0.01
6H-3, 37-39	48.87				2.01	1.47	3.12		
6H-3, 122-124	49.52		-0.81	1.39	1.72	1.23	2.83	3.64	-0.16
6H-4, 26-28	50.06		0.02	1.58	1.91	1.45	3.02	3.00	-0.13
6H-4, 123-125	51.03		-0.37	1.67	2.21	1.43	3.32	3.69	-0.23
6H-5, 26-28	51.56		-0.73	1.76	1.99	1.75	3.10	3.83	-0.01
6H-5, 123-125	52.53				1.00	1.05	2.11		
6H-6, 32-34	53.12				1.79	1.58	2.90		
6H-6, 122-124	54.02		-0.49	1.72					

Diagenetic effects are ubiquitous to all calcium carbonate deposited on the seafloor. Usually, these effects take the form of dissolution, which preferentially removes the lighter isotopes in the carbonates, thus biasing results to isotopically heavier values indicating "cooler" temperatures (Killingley, 1983; Savin, 1977; Savin et al., 1973). Diagenetic overgrowths of calcite also are important and were the dominant form of diagenesis observed in the sediments that we studied. Overgrowths form in the sediments at temperatures cooler than those found in surface waters, and similarly bias planktonic isotopic data to "cooler" values. If overgrowths and infillings of foraminiferal tests were noted as moderate or greater, these samples were disregarded. Scanning electron microscope (SEM) photomicrographs of samples

Table 4. Age datums used for the calculation of sedimentation rates.

Site	Depth (mbsf)	Age (Ma)	Datum Type a	Biohorizons b
811	8.9	0.465	N	CN14b-CN15 (NN19-NN20)
	19.5	2.03	Sr (foraminifers)	
	15.0-24.5	1.9	F	N22-N21
	22.9	1.48	N	C. macintyrei
	22.9	1.88	N	CN12-CN13 (NN18-NN19a)
	29.4	2.29	N	CN12d-CN12c (NN17-NN18)
	38.9	2.60	N	CN12b-CN12a
	53.0-62.5	3.0	F	N21-N18/N19
	64.8	2.4	Oxygen Isotopes	
	72.6	3.51	N	CN11b-CN12a (NN15-NN16)
	72.7	2.47	Sr (foraminifers)	
	76.9	4.24	N	CN10c-CN11a
	88.0	3.30	Oxygen Isotopes	
	90.0	3.60	Oxygen Isotopes	
	91.6	5.26	N	CN9b-CN10a (NN11b-NN12)
	94.8	5.02	Sr (foraminifers)	
	91.0-100.5	5.2	F	N18/N19-N16/N17
	104-107	6.2-6.4	Carbon Shift	
	132.6	8.20	N	CN8-CN9a (NN10-NN11a)
	172.6	8.85	N	CN7-CN8a (NN9-NN10)
	180.5	.00-11.7	Sr (foraminifers)	
	186.0-195.5	10.4	F	N14-N15
817	24.0	0.27	N	CN14b-CN15
	29.0	0.465	N	CN14a-CN14b
	47.0	0.93	N	CN13b-CN14a
	53.0	1.48	N	CN-13a-CN13b
	72.0	1.88	N	CN12b/d-CN13a
	75.0	2.6	N	CN12a-CN12b/d
	100.7-110.2	1.9	F	N21-N22/23
	115.7	1.86	Sr (foraminifers)	
	129.2-138.7	3	F	N18/19-N21
	157.0	3.51	N	CN11b-CN12a
	161.0	4.24	N	CN10c-CN11b
	176.7-186.2	5.2-5.3	F	N16/17-N18/19
	177.0	4.6-4.72	N	CN10a/b-CN10c
	177.4	4.8	Sr (foraminifers)	
	181.0	5.26	N	CN9b-CN10a/b
	184.9	4.82	Sr (foraminifers)	
	186.9	4.84	Sr (foraminifers)	
	193.9	4.99	Sr (foraminifers)	
	194.4	5.05	Sr (foraminifers)	
	197.4	5.2	Sr (foraminifers)	
	200.0	6.74	N	CN9a-CN9b
	201.9	.42-12.0	Sr (foraminifers)	
	205.2-214.7	10.2	F	N10/14-N16/17
209	3.0	0.27	N	NN20-NN21 (CN14b-CN15)
	4.5	0.465	N	NN19-NN20 (CN14a-CN14b)
	20.0	1.88	N	NN18-NN19 (CN12d-CN13a)
	21.0	2.29	N	NN17-NN18 (CN12c-CN12d)
	22.5	2.42	N	NN16-NN17 (CN12b-CN12c)
	40.5	3.51	N	NN15-NN16 (CN11b-CN12a)

[a] N = nannofossil datum. F = foraminiferal datum.
[b] First datums listed are those that were given the sediments studied; those in parentheses are the approximate equivalent.

from Holes 811A and 817A were taken to illustrate average and "highest acceptable" amounts of dissolution and infilling (Fig. 9).

The combination of all errors inherent in paleotemperature calculations leads to a possible 3° to 4°C difference between measured and actual temperatures. This difference is usually toward "cooler" temperatures (Berger and Gardner, 1975; Savin, 1977; Savin et al., 1973). We think that through careful data collection we were able to estimate temperatures to within this limit and, therefore, can provide a first-degree estimate of seawater temperature changes off the northeastern Australian margin during the late Neogene.

DISCUSSION

Changes in Surface-water Temperature and Circulation

Paleotemperatures calculated from oxygen isotopic data have allowed us to reconstruct a general paleotemperature curve for the Queensland Plateau from the late Miocene to the Holocene. Correlating these data with sedimentary facies changes enabled us to look at the possible effect that paleoceanographic changes had on the growth and demise of the reef complexes in this area.

Middle Miocene

Samples of middle Miocene age taken on the Queensland Plateau predominantly contain reefal detritus, periplatform deposits, and shallow-water, lagoonal-type sediments (Davies, McKenzie, Palmer-Julson, et al., 1991) (Fig. 11). The sediment facies (and their correlation to seismic data) indicate that, during this time, extensive reef complexes existed on the Queensland Plateau (Davies, McKenzie, Palmer-Julson, et al., 1991). The tropical nature of the bioclasts is evidence that surface-water temperatures must have been at least 20°C and that paleoceanographic conditions must have been advantageous for reef growth. Stable isotopic data could not be collected in this time interval because of the extensive remineralization of the sediments resulting from high amounts of reefal detritus and metastable, bank-derived carbonate. Therefore, the surface-water temperatures could not be estimated more exactly.

The position of the study area at this time (~15 Ma) was between 24°S and 29°S (present-day position is 15°S to 20°S) (Davies, McKenzie, Palmer-Julson, et al., 1991). This places the sites on the tropical/subtropical climate boundary, relative to modern-day climatic conditions. The presence of warm-water faunas in the recovered sediments indicates that the conditions were tropical.

Late Miocene

Late Miocene age sedimentary facies recovered during Leg 133, at sites on and around the Queensland Plateau, show an increase in pelagic sedimentation at the deeper sites and an increase in periplatform sedimentation at the shallower sites. This indicates an alteration of reefal development on the plateau (Davies, McKenzie, Palmer-Julson, et al., 1991). Planktonic foraminifers from the youngest sediments of late Miocene age overlying the middle Miocene reef deposits at Hole 811A record isotopic ratios that indicate relatively cool surface-water temperatures (15° to 18°C; Fig. 10). Similar temperatures are recorded in upper middle Miocene sediments (11.5–10.8 Ma) from Site 209. Temperate-water nannofossil and foraminiferal assemblages and temperate-water carbonates found in the sediments from this time provide further evidence for lower temperatures than in the middle Miocene (Davies, McKenzie, Palmer-Julson, et al., 1991). Calculated temperatures were below those needed for productive reef growth, and this may have been an important cause for the back-stepping of the reef complexes.

Calculated surface-water temperatures increased throughout the late Miocene at Hole 811A. By 5.5 Ma surface-water values were approximately 19°C at Site 811 and 20°C and Site 817.

Paleotemperatures were calculated using oxygen isotopic data from Site 590 (Elmstrom and Kennett, 1986) and Site 586 (Whitman and Berger, 1992) to compare our data set with those from open-ocean sites. Both Site 590 and 586 data show a trend similar to that for Holes 811A and 817A from 6 to 5 Ma. This may indicate either a regional cooling of surface waters in the South Pacific or the effect of ice-volume changes overprinted on the paleotemperature calculations.

Pliocene

During the Pliocene (5.2–1.6 Ma), surface-water temperatures at Holes 811A and 817A continued to be cooler than modern temperatures; yet, compared with those of late Miocene age, temperatures were somewhat warmer, with an average of 18° to 21°C. Paleotemperature data from Sites 590 and 586 have similar patterns of change to those of Hole 811A and Site 209 from 3.8–1.6 Ma (and to a lesser extent Hole 817A). This again indicates either a regional change in the surface-water temperature or the ice volume effect.

The continuation of somewhat cool surface-water temperatures in the Pliocene may explain why only a localized reinitiation of reef growth is seen on the Queensland Plateau at this time (Davies, McKenzie, Palmer-Julson, et al., 1991). The effect of cooler surface waters is

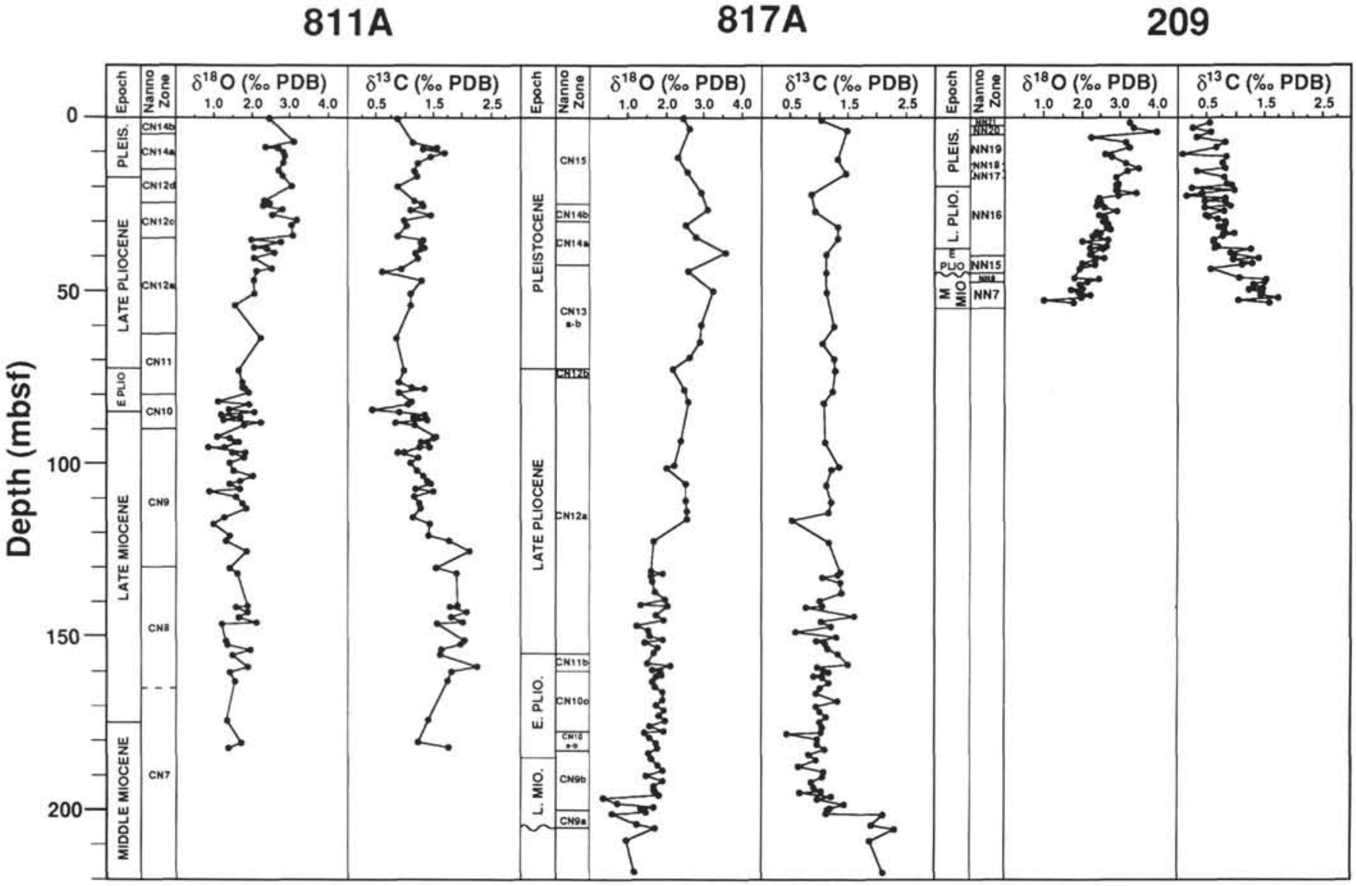

Figure 6. Carbon and oxygen isotopic data from *Cibicidoides* spp. in Holes 811A and 817A and Site 209. Data are plotted vs. depth relative to PDB, with stratigraphic ages and calcareous nannofossil biostratigraphy.

compounded with rapid subsidence during the late Miocene and early Pliocene (Katz and Miller, this volume; Symonds et al., this volume). Therefore, a combination of tectonic and paleoceanographic factors may have prevented the reinitiation of extensive reef development.

Pleistocene

A dramatic increase in the inferred surface-water temperatures (to 25°C) can be seen in the Pleistocene at Holes 811A and 817A (Fig. 10). No data are available from Site 590 for the Pleistocene, and data from Sites 209 and 586 do not show this change.

This apparent temperature increase is not the result of ice-volume changes, as ice volumes were generally increasing and this would make the inferred temperatures cooler. The other possible cause for this change is a decrease in salinity either because of an increase in precipitation or increased monsoonal flow of water from the Torres Strait. This effect, if present, is not likely to be large enough to produce an isotopic shift of the magnitude seen.

The increase in inferred temperature recorded in the Pleistocene occurs concurrently (within the error of the age models) with the initiation of the Great Barrier Reef. This indicates that once surface waters warmed, reefs were able to develop on the Australian continental margin. Reef growth did not become increasingly extensive on the Queensland Plateau because, by the Pleistocene, much of the plateau surface was below the photic zone as a result of continued subsidence.

Mechanisms for Changes in Surface-water Temperature

The causes for the inferred temperature variations recorded in the oxygen isotopic data are difficult to determine with the limited data set available. Further data will be needed to fully assess the mechanisms for temperature change.

The finding that tropical surface-water temperatures have changed substantially over time has important implications for estimates of paleo ice volumes. It has been argued, using Holocene $\delta^{18}O$ records, that the temperature of surface waters in tropical regions changes little, if at all, during glacial-interglacial cycles. Therefore, it has been proposed that these records can be used as indicators of ice volume, as temperature changes are not important (Matthews and Poore, 1980). Interpreting tropical $\delta^{18}O$ records in this way suggests that ice sheets may have existed since the Eocene. Our findings show that the temperature of tropical surface waters is not constant over time, as $\delta^{18}O$ records indicate substantial temperature changes that are supported by faunal variations in the sediments.

Evidence shows that cooler surface waters existed on the Queensland Plateau from the late Miocene to the late Pliocene (10.4 to 2.0 Ma). The question is if this change from warmer mid-Miocene temperatures is local or regional. The fact that Sites 590 and 586 data show changes similar to Hole 811A in the Pliocene indicates either that these changes were regional or caused by ice-volume changes. Yet, the absence of longer time-series records from Sites 590 and 586 makes a conclusion difficult. Further investigation of other drill sites in the area may show conclusively whether these paleoceanographic changes are local or regional.

If these changes were local, variations in sea level, combined with subsidence of the Queensland Plateau, may have caused local temperature fluctuations because of restriction of circulation in the western Coral Sea. During times of high sea level (such as the early middle Miocene), and after the subsidence of the platform, warm waters from the SEC may have been able to easily enter the area across the platform

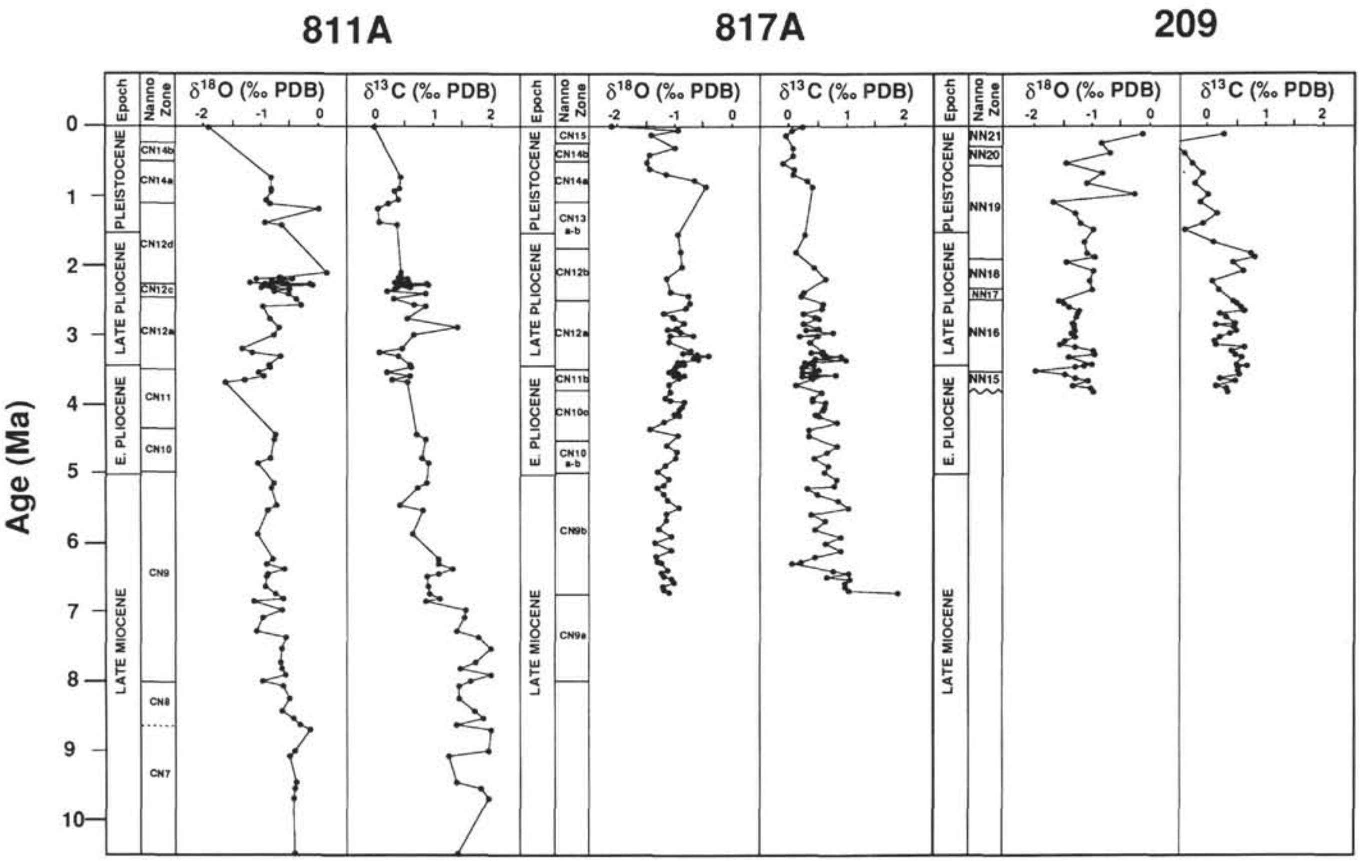

Figure 7. Carbon and oxygen isotopic data for *G. ruber* in Holes 811A and 817A and Site 209. Data are plotted vs. age relative to PDB, with stratigraphic ages and calcareous nannofossil biostratigraphy.

(Fig. 12a). Active reef growth and warmer sea-surface temperatures on the northeastern Australian margin during these times have been observed in the record. At times of low sea level, before the subsidence of the plateau (such as during the late Miocene), circulation over the platform may have been restricted by shallow and exposed areas on the Queensland Plateau (Fig. 12b). This may have allowed the cooler SLW water mass to surface in this area or cooler water to inflow from the south.

If the intrusion of cooler waters occurred on a more regional scale, it may have been because of middle Miocene circulation intensification resulting from increases in the pole-equator temperature gradient in the South Pacific. As a result of this change, the SEC may have narrowed and possibly allowed the intrusion of cooler water from the south into the area.

The sudden oxygen isotopic depletion seen in the Pleistocene at both Sites 811 and 817 is not seen in either Site 209 or 586 data. Therefore, this change is local. It is difficult to determine if this isotopic shift is dominantly the result of a temperature increase or a change in salinity. Three lines of evidence indicate the former: (1) the calculated sea-surface temperatures for the Pliocene are approximately 4°–5°C cooler than modern measured temperatures for the area, and, therefore, temperatures must have increased in the Pleistocene to the modern value; (2) the appearance of the Great Barrier Reef is nearly synchronous with this increase in temperature, indicating that the warmer water allowed the development of productive reef growth on the margin; (3) the effect of salinity variations on the isotopic values, if present, is likely to be small. This apparent temperature change is not the result of northward movement of the Australian continent, because the change was too rapid. But, it is important to note that alteration of circulation also could change $\delta^{18}O$/salinity relationships, and this would influence the interpretation of the paleotemperatures.

The cause for the mid-Pleistocene increase in temperature, as stated above, may be the result of sea-level variations and subsidence of the Queensland Plateau allowing warmer water to flow across the platform.

Changes in Seafloor Temperature

Prior to Leg 133, we believed that the analysis of benthic foraminiferal $\delta^{18}O$ would allow us to study variations in deep-water temperatures resulting from subsidence of the Queensland Plateau. The analysis of benthic foraminiferal paleoenvironmental data has shown that this is not the case, however, as by the late Miocene, the age from which we are first able to obtain reliable isotopic data, the platform had already subsided to within the range of AIW (Katz and Miller, this volume). At Site 811, near 8.5 Ma, there is an apparent facies deepening from upper bathyal (200–600 m) to mid-bathyal (600–1000 m) depths. Yet, within this interval, the benthic isotopic values do not change, indicating that the seafloor was most likely within AIW flow immediately before the apparent depth transition. At Site 817, sediments above the hiatus present near 6.2 Ma were all deposited at mid-bathyal depths (600–1000 m) and, therefore, should be recording changes occurring within AIW. Thus, as changes in oxygen isotopic values are not reflecting a deepening of the seafloor, they are indicating changes in ice volume and temperature.

Of the three sites analyzed in this paper, both Sites 811 and 817 are within the flow of AIW, whereas Site 209 is most likely between the flow of AIW and Antarctic Deep Water (ADW). Site 586 on the Ontong Java Plateau is also a deep-water site most likely within the ADW range. Site 590, however, is within the range of AIW. When the calculated paleotemperatures for all of these sites are examined (Fig. 10), all five records are similar to a first-degree, most likely as

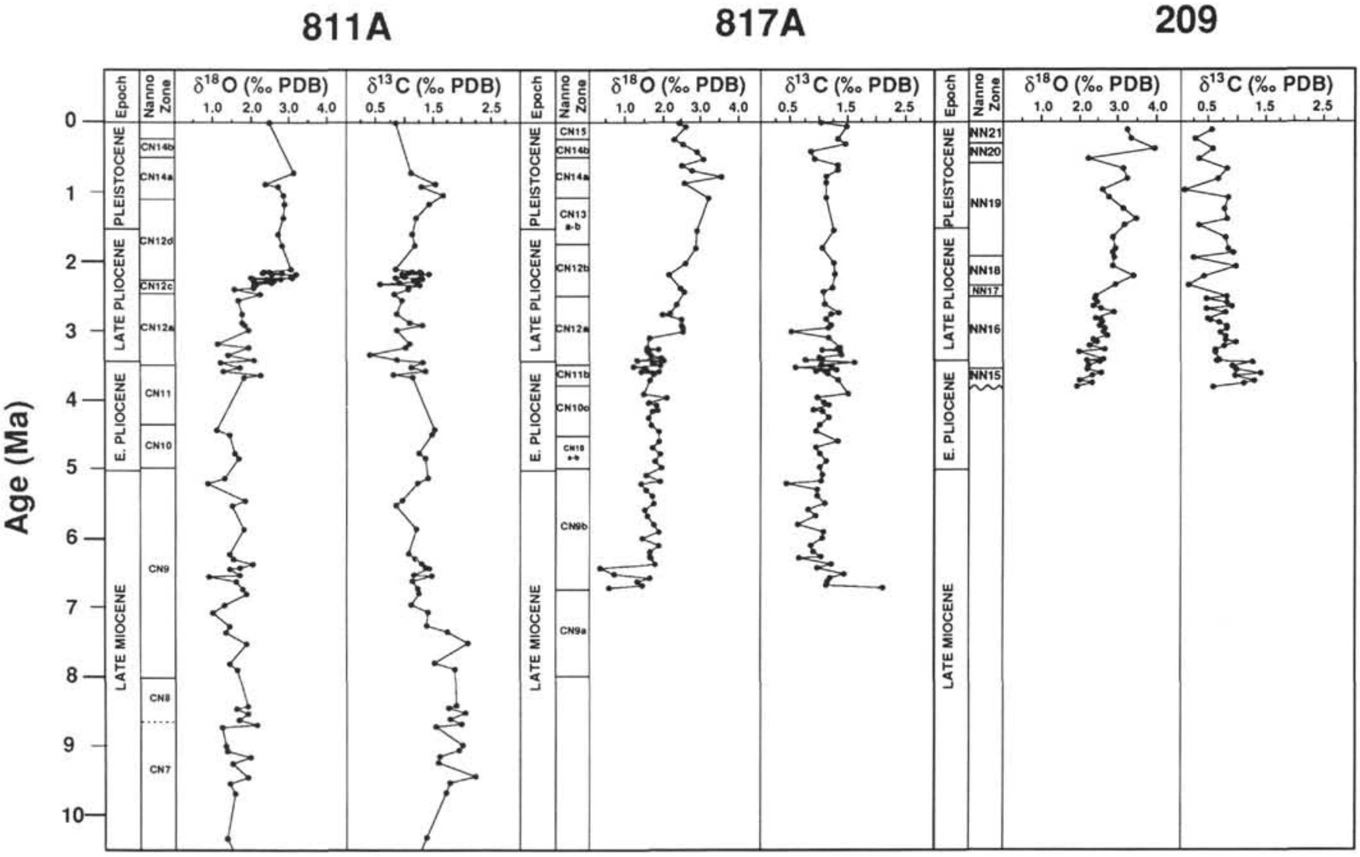

Figure 8. Carbon and oxygen isotopic data from *Cibicidoides* spp. in Holes 811A and 817A and Site 209. Data are plotted vs. age relative to PDB, with stratigraphic ages and calcareous nannofossil biostratigraphy.

a result of the ice-volume effect. Yet, the three sites occurring within AIW show a stronger correlation. Therefore a portion of the $\delta^{18}O$ signal seen in the three records is most likely the result of changes in AIW temperature over time. The fact that the Site 590 data are similar to the Sites 811 and 817 data is an indication that local changes in seafloor temperature were not a dominant component of the benthic foraminiferal signal.

CONCLUSIONS

Sedimentary facies and isotopic data from sites surrounding the Queensland Plateau show the following:

1. Surface waters in the middle Miocene were warm. Sediments recovered during Leg 133 showed that extensive reef growth occurred on the Queensland Plateau and, therefore, temperatures were probably greater than or equal to 20°C, the limit for healthy reef growth.
2. Isotopic data from planktonic foraminifers in the sediments overlying the reefal deposits indicate cooler surface-water temperatures, in the range of 16° to 19°C, from the late Miocene to the late Pliocene. These cooler surface waters, with temperatures below those needed for productive reef growth, may have been a factor contributing to attenuation of reef development on the Queensland Plateau.
3. The middle Pleistocene warming of surface waters to ~25°C, indicated by the planktonic foraminiferal isotopic data, may have promoted the initiation of the Great Barrier Reef.
4. Benthic foraminiferal isotopic data from Sites 811 and 817 appear to covary with those of Site 590. As all of these sites are within the range of AIW, it is an indication that local variations in bottom-water $\delta^{18}O$ were overshadowed by regional changes.

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* Abbreviations for names of organizations and publications in ODP reference lists follow the style given in *Chemical Abstracts Service Source Index* (published by American Chemical Society).

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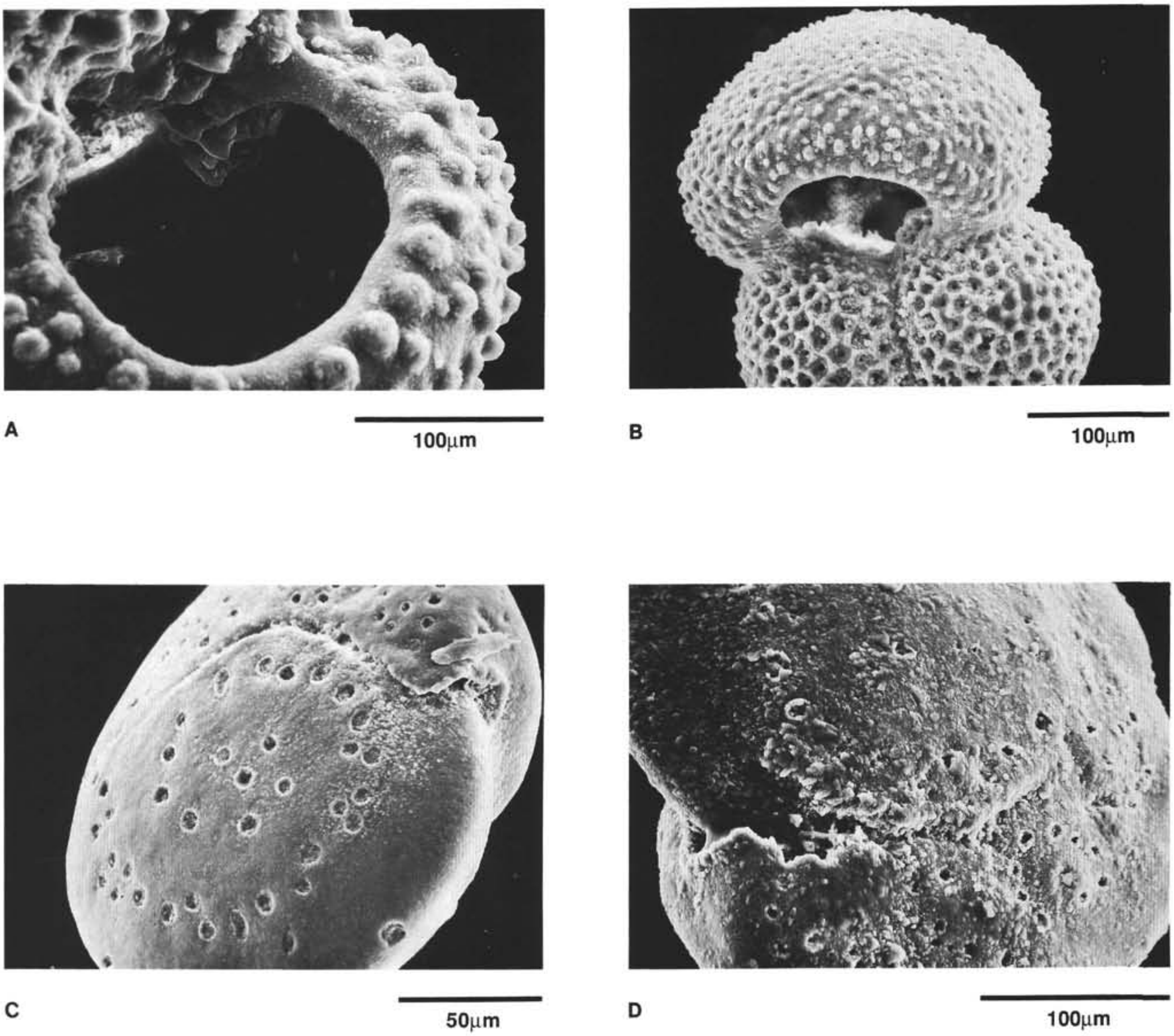

Figure 9. SEM photomicrographs of foraminifers from Leg 133 samples used for isotopic analysis. **A.** An example of a well-preserved *G. ruber* (133-811A-15H-2, 110–112 cm). Minor amounts of calcite overgrowths are present, and pores visible through the aperture of the foraminifer are not infilled. **B.** A *G. ruber* sample from 133-817A-16H-1, 20–22 cm, which is an example of the maximum amount of infilling considered acceptable. A greater amount of overgrowth is seen on the surface of the foraminifers, yet the pores are still relatively open. Inside the aperture, coccoliths may be seen. As this photo was taken before the sample was sonicated, much of this infilled material likely would have been removed before isotopic analysis. **C.** An example of a well-preserved *Cibicidoides* sp. (133-817A-15H-4, 20–22 cm). Although some overgrowths and infilling may be seen in the aperture and in the pores, in general the foraminifers are well preserved. **D.** A *Cibicidoides* sp. sample from 133-811A-16H-3, 110–112 cm, which illustrates the maximum amount of diagenesis considered acceptable for isotopic analysis of a benthic foraminifer sample. More overgrowths around the aperture and the pores are seen, yet the pores are not totally infilled, as they would be with poorly preserved material. As with the *G. ruber* photo (9B), this was taken before sonicating, which undoubtedly would have removed some of the infilling.

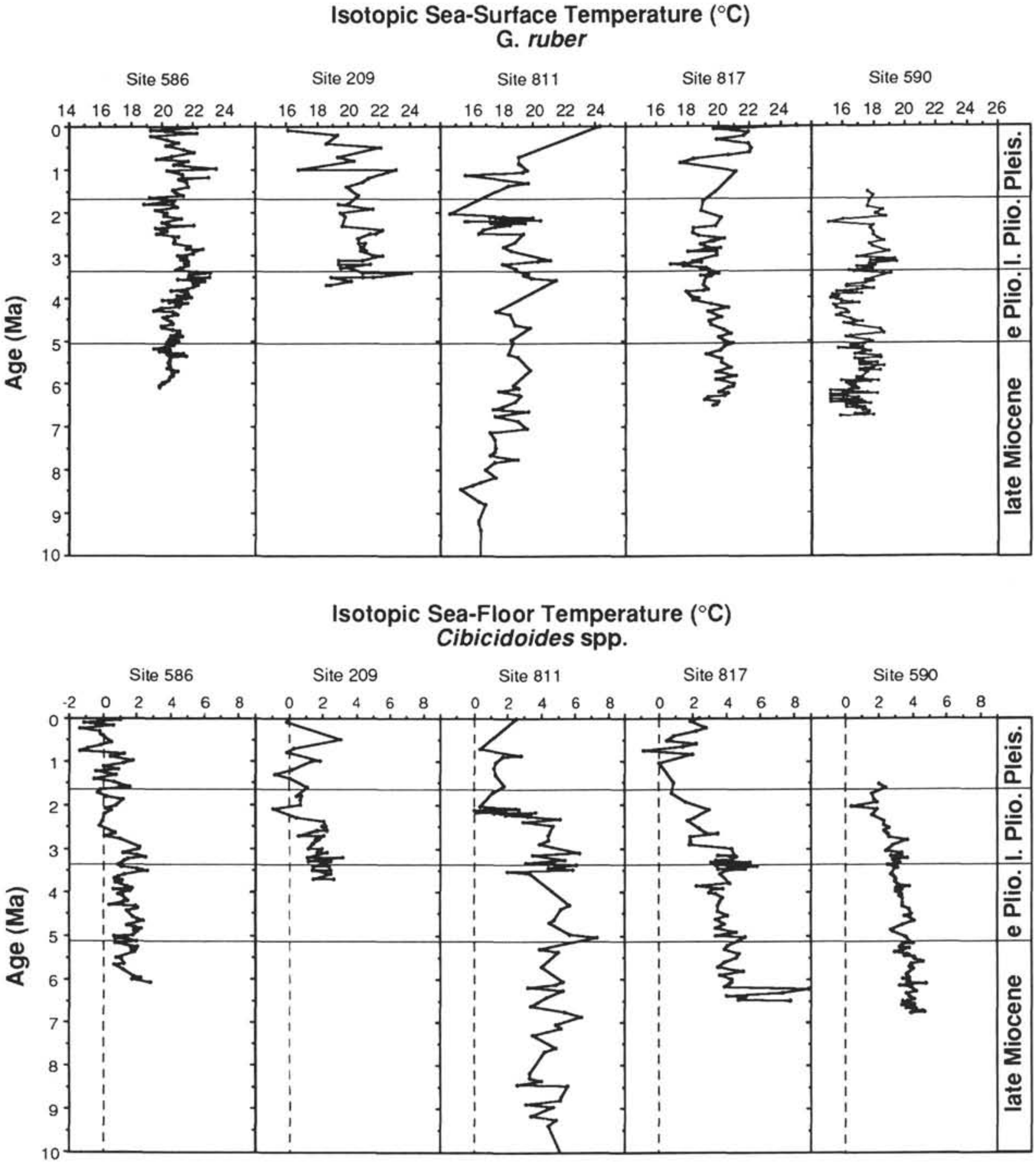

Figure 10. Calculated sea-surface (*G. ruber*) and seafloor (*Cibicidoides* spp.) isotopic paleotemperatures from Holes 811A and 817A and Sites 209, 586, and 590 plotted vs. age. Lines on the figure show age boundaries.

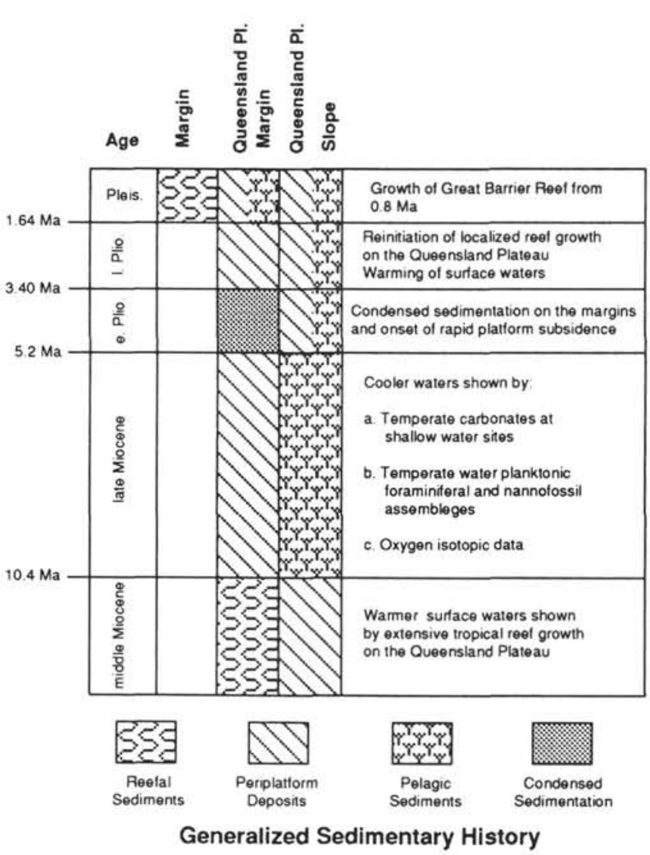

Figure 11. Summary of shipboard stratigraphic facies data for sites along the margin, Queensland Plateau margin, and the Queensland Plateau slope. A simplified representation of the sedimentary changes at each of the locations is shown. This figure is presented to provide a general picture of the facies changes that were observed during Leg 133, with a brief description of the paleoceanographic and sedimentary changes.

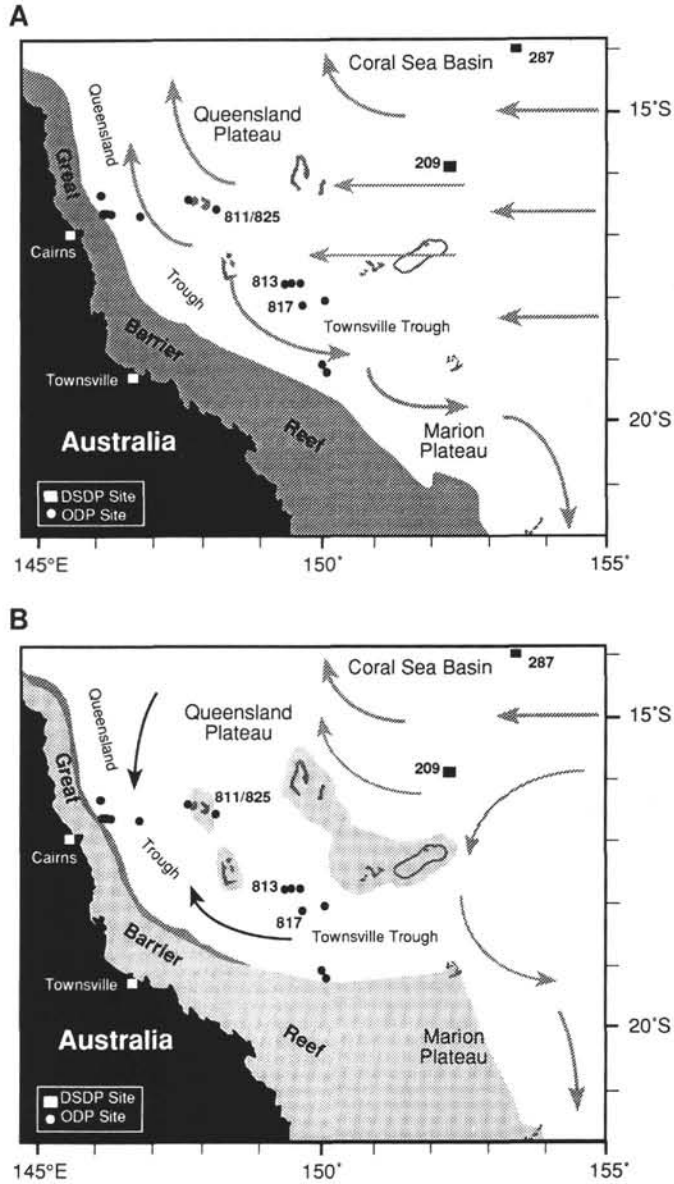

Figure 12. Possible paths of current flow on the northeastern Australian margin during (**A**) higher sea level and (**B**) relatively low sea level periods. Solid arrows indicate possible surfacing of cooler SLW waters with lower sea level (**B**). As with previous figures, the darker shaded areas represent reef growth, whereas the lightest shading follows a general outline of the Queensland and Marion plateaus. The medium gray shading in the low sea level frame (**B**) indicates areas that, if sea level were lowered by 80 m, would have been very shallow or exposed.