Quaternary climate change as revealed by calcium carbonate fluctuations in western Equatorial Atlantic sediments*

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Abstract—Fluctuations of total calcium carbonate content in eight western Equatorial Atlantic cores are used to evaluate Quaternary climate change. Both pelagic Mid-Atlantic Ridge and terrigenous-rich continental-rise cores that span the last 130,000 years show identical carbonate fluctuations which accurately reflect climatic oscillations during the Holocene, Wisconsin Glacial, and Last Interglacial. The carbonate fluctuations reveal a 20,000-year periodicity for warm-cold (interstadial-stadial) cycles with carbonate maxima reflecting interstadials and minima reflecting stadials. The carbonate fluctuations correlate in detail with other time scales for climatic oscillations deduced previously from radiometrically-dated sea-level maxima, solar insolation fluctuations, and North American ice-margin fluctuations.

Pelagic cores spanning the last 500,000 years do not clearly reflect the 20,000-year cold–warm cycles but do show the 100,000-year glacial–interglacial cycles previously revealed by oxygen-isotope variations. The carbonate fluctuations in another core, which contains a continuous record of the last 1,800,000 years, indicates that 15 to 20 of these 100,000-year cycles have occurred during this period. The carbonate fluctuations of two cores suggest that the Atlantic circulated faster during glacial than during interglacials. A core from the Demerara Plateau (2000-m depth) has anomalous carbonate fluctuations plus expanded interglacial and shortened glacial sections that suggest faster circulation of surface water masses during glacial. The sediments of a pelagic core, which were deposited beneath the Antarctic Bottom Water, have undergone severe carbonate dissolution during glacial. These cycles apparently indicate an increase in production and faster circulation of the Antarctic Bottom Water during glacial.

INTRODUCTION

The calcium carbonate content of deep-sea sediments has long been known to fluctuate in response to Quaternary climate change. Schott (1935) recognized that the carbonate content of Late Wisconsin sediments in the Equatorial Atlantic was lower than that of the overlying Holocene sediments. He concluded that a relatively lower rate of carbonate deposition coupled with a relatively higher rate of clay deposition during the Wisconsin accounted for this relationship. Numerous studies have confirmed that in Equatorial Atlantic and Caribbean sediments the carbonate content is relatively higher in interglacial than in glacial sequences and has fluctuated in response to Quaternary climatic oscillations (Correns, 1937; Wiseman, 1954, 1956, 1965; Olausson, 1965, 1967; Broecker, Turekian and Heezen, 1958; Turekian, 1965; Needham, Conolly, Ruddiman, Bowles and Heezen, 1969; Ruddiman, 1971; Hays and Perruzza, 1972; Damuth, 1973; Gardner, 1973; Prell, 1974).

The exact cause of the observed carbonate fluctuations has not yet been completely resolved because as many as three separate factors control the total carbonate content at any given location: (1) productivity of carbonate-secreting organisms, (2) dissolution of calcareous tests during and after deposition, and (3) dilution by non-calcareous material. Wiseman (1956, 1965) attributed the fluctuations of total carbonate to fluctuations in surface productivity. Unfortunately, Wiseman was forced to assume that the accumulation rate of

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non-calcareous material (dilution) was constant, an assumption previously questioned by Kullenberg (1953). Subsequently, Broecker, Turekian and Heezen (1958) demonstrated that, at least for the Equatorial Atlantic, the rate of non-calcareous clay accumulation was the least constant of all three factors. Their study of core A180-74 revealed that the rate of non-calcareous clay deposition decreased by a factor of 3-7 from latest Wisconsin to earliest Holocene. Furthermore, Broecker and others discovered that the rate of carbonate accumulation decreased from latest Wisconsin to early Holocene by a factor of 2-1, a finding exactly the reverse of Wiseman's.

More recent studies of eastern Equatorial Atlantic sediments support the conclusions of Broecker, Turekian and Heezen (1958) and suggest that total carbonate content has been primarily controlled by changes in rates of accumulation of non-calcareous material (largely wind-blown detritus) (Needham, Conolly, Ruddiman, Bowles and Heezen, 1969; Ruddiman, 1971; Hays and Perruzza, 1972). However, Gardner (1973, 1975) has shown that dilution does not always account for the observed carbonate fluctuations in this region but that dissolution, especially during glacial cycles, has had a major effect on total carbonate content. Recent studies of western Equatorial Atlantic sediments by Damuth (1973) and Bé, Damuth, Lott and Free (in press) also suggest that both dilution and dissolution have been important in determining the total carbonate content of the sediments.

Although these studies have not resolved the relative contributions of dilution, dissolution, and productivity to the total carbonate content in Equatorial Atlantic sediments, they have demonstrated that the carbonate fluctuations are time-stratigraphic throughout the region and reflect the timing of glacial-interglacial climatic cycles. How accurately the carbonate fluctuations reflect the nature and timing of the climatic oscillations is a major concern of the present study.

During a detailed examination of western Equatorial Atlantic sediments (Damuth, 1973), eight piston cores from various physiographic provinces were analyzed for total carbonate content because close-interval carbonate determinations had previously been reported for only a few western Equatorial Atlantic cores (Olausson, 1960; Turekian, 1956) with little no discussion of the relationship of the observed fluctuations to climate change. The eight cores were selected because they contain continuous, undisturbed records of sediment accumulation during the past 175,000 to 1,800,000 years and contain sediment types representative of the various physiographic provinces of the region. The detailed nature and timing of carbonate fluctuations in the cores are compared with previously published evidence of Quaternary climatic oscillations including faunal and oxygen-isotope variations, solar insolation curves, radiometrically dated high sea-level stands, and eastern North American continental deposits. Two of the cores also provide evidence for changes of the intensity of circulation in the Atlantic in response to glacial-interglacial climatic fluctuations.

### STRATIGRAPHY OF PISTON CORES

The locations of the cores are shown in Fig. 1 (Table 1). Cores A180-73, V25-59, V25-60, and V25-45 are from seamounts of the Mid-Atlantic Ridge (Fig. 1) and are composed of light brown to light orange-tan, high carbonate (> 50%), pelagic foraminiferal ooze and marl. Cores A180-73 and V25-59 penetrate Quaternary sediments approximately 175,000 years old, whereas cores V25-60 and V25-45 penetrate sediments approximately 400,000 and 1,800,000 years old. Core V25-42

### Table 1. Locations, depths (corrected meters), and lengths of piston cores.

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>Length (cm)</th>
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<tbody>
<tr>
<td>V15-168</td>
<td>00°12'N</td>
<td>39°54'W</td>
<td>4219</td>
<td>1101</td>
</tr>
<tr>
<td>V25-42</td>
<td>12°33'N</td>
<td>50°39'W</td>
<td>4704</td>
<td>1035</td>
</tr>
<tr>
<td>V25-45</td>
<td>11°32'N</td>
<td>42°38'W</td>
<td>3455</td>
<td>1136</td>
</tr>
<tr>
<td>V25-56</td>
<td>03°33'S</td>
<td>35°14'W</td>
<td>3512</td>
<td>735</td>
</tr>
<tr>
<td>V25-59</td>
<td>01°22'N</td>
<td>33°29'W</td>
<td>3824</td>
<td>841</td>
</tr>
<tr>
<td>V25-50</td>
<td>03°17'N</td>
<td>34°50'W</td>
<td>3749</td>
<td>877</td>
</tr>
<tr>
<td>A180-73</td>
<td>00°10'N</td>
<td>23°00'W</td>
<td>3748</td>
<td>490</td>
</tr>
<tr>
<td>V25-76</td>
<td>08°29'N</td>
<td>53°10'W</td>
<td>2107</td>
<td>684</td>
</tr>
</tbody>
</table>
is from a small basement hill that protrudes approximately 300 m above the Demerara Abyssal Plain (Fig. 1). The core consists of interbedded pelagic brown clay (CaCO₃ < 25%) and light brown nannofossil marl (CaCO₃ = 25 to 60%); it penetrates sediments approximately 500,000 years old.

Cores V25–56 and V15–168 are from the continental rise (Fig. 1) and are largely composed of gray hemipelagic clay rich in terrigenous and organic detritus. Core V15–168 contains interbeds of silt and sand that were deposited by turbidity currents. The upper 40 to 50 cm (Holocene sections) of these cores consist of light brown pelagic foraminiferal ooze, which reflects the interruption of terrigenous sediment supply to the lower continental margin throughout the Holocene because of the high sea-level stand (DAMUTH, 1973; McGary and DAMUTH, 1973). Cores V25–56 and V15–168 penetrate sediments approximately 200,000 and 130,000 years old. Core V25–76 is from the upper continental slope just below the outer edge of the Demerara Plateau (Fig. 1). The core is composed of gray hemipelagic clay and penetrates sediments approximately 200,000 years old. More detailed descriptions of the locations and lithologies of these cores can be found in DAMUTH (1973).

Stratigraphic control was established by determining variations in abundance of foraminifera of the Globorotalia menardii complex at 10-cm intervals down each core using the method of ERICSON and WOLLIN (1956, 1968). Ericson and Wollin have been able to construct correlative, reproducible climatic curves for the entire Quaternary from Equatorial Atlantic and Caribbean cores and have designated the warm and cold climatic zones observed by the letters Z to Q in order of increasing age according to the system begun by ERICSON (1961). The Z zone (0 to 11,000 yr B.P.) contains abundant G. menardii complex and corresponds to the Holocene; the Y zone (11,000 to 85,000 yr B.P.) contains no G. menardii complex and approximately corresponds to the Wisconsin Glacial; the X zone (85,000 to 127,000 yr B.P.) contains abundant G. menardii complex and approximately corresponds to the Last Interglacial, etc. Dates for climatic-zone boundaries back to 400,000 yr B.P., which are used throughout the present study, are those proposed by BROECKER and VAN DONK (1970) with the exception of the X/Y boundary,
which is approximately 85,000 yr B.P. in the western Equatorial Atlantic (DAMUTH, 1973), rather than 75,000 yr B.P. as proposed by Broecker and Van Donk. Dates for climatic zone boundaries older than 400,000 yr B.P. are from ERICSON and WOLLIN (1968) and are based on magnetic reversal stratigraphy.

The date of 85,000 yr B.P. used throughout the present study for the X/Y boundary was determined in the following manner: radiocarbon dating placed the Y/Z boundary at 11,000 yr B.P. (BROECKER, EWING and HEEZEN, 1960), whereas the W/X boundary has been radiometrically determined by BROECKER and VAN DONK (1970) to be 127,000 yr B.P. If the total sediment thicknesses of the X and Y zones within core A180–73 (340 cm) and core V25-59 (325 cm) of this study are each divided by 116,000 years (the time difference between the W/X and Y/Z boundaries), then average sedimentation rates of 2.9 and 2.8 cm (10^4 yr)^{-1} are obtained for A180–73 and V25-59. These rates can then be used to calculate the age of the X/Y boundary in each core by dividing the thickness of the Y zone (215 cm for A180–73 and 205 cm for V25-59) by the sedimentation rate for each core (giving 74,000 and 73,000 years), and then adding 11,000 years to account for the Z zone. This procedure yields an age for the X/Y boundary of 85,000 yr B.P. for A180–73 and 84,000 yr B.P. for V25–59. This date is consistent with the 84,000 yr B.P. reported by PRELL (1974) for the age of the X/Y boundary in Caribbean sediments.

The eight cores were sampled for calcium carbonate determinations at 10-cm intervals. Samples were ground to a fine powder and desiccated in an oven (≈ 80°C) for at least 24 h. After cooling, approximately 0.1 g of dried sample was weighed for CaCO₃ analysis. The percentage of CaCO₃ was measured by the rapid gasometric technique described by HÜLSEMANN (1966). The experimental error is less than 3%. Measurements were made on reagent grade CaCO₃ (100%) standard samples prior to the first and after every few core sample determinations to insure proper functioning of the apparatus. No correction was made for salt content of the dry cores. The percentage of coarse fraction (> 62 μm) in each sample was determined prior to faunal analysis. Calcium carbonate, coarse fraction, and G. menardii/weight values for each sample of the cores can be found in DAMUTH (1973; Appendix C).

CARBONATE FLUCTUATIONS AND CLIMATE DURING THE LAST 130,000 YEARS

Four cores reveal the detail of calcium carbonate fluctuations during the last 130,000 years (Fig. 2). The two Mid-Atlantic Ridge cores, V25–59 and A180–73, are composed entirely of pelagic marl and ooze, whereas the two continental rise cores, V25–56 and V15–168, are composed of hemipelagic clay that is rich in terrigenous detritus. V15–168 also contains several turbidites composed of terrigenous sand and organic material. These redeposited beds have been omitted from Fig. 2 so that the carbonate curve is continuous; however, the location and thickness of each bed are noted beside the curve by small arrows and numbers.

Although the cores are from different sedimentary regimes, the four carbonate curves are quite similar in shape and each curve contains a number of peaks and depressions that are correlative from core to core (Fig. 2). The peaks are numbered in succession downwards by P1, P2, P3, etc., to P10, the depressions by D1, D2, etc., to D7. In some instances a numbered peak (or depression) actually consists of a group of two or more smaller peaks (or depressions), e.g. P8. This is especially true for the two continental rise cores, which contain relatively expanded sections because of the high influx of terrigenous sediment. For these expanded sections the time interval between the 10-cm samples is relatively shorter than in the pelagic cores and should reveal relatively greater detail in the carbonate fluctuations. Also, slight changes in rates of influx of terrigenous sediment could cause minor peaks and depressions that would not occur in the pelagic sediments.

The Holocene (Z zone) section of each core contains relatively high carbonate contents and
peak P1 is generally the maximum value observed in each core (54 to 84%). The upper few centimeters of cores A180-73 and V15-168 are apparently missing, thus the maximum values for peak P1 in these two cores may be a few percent higher than actually observed.

A broad, deep carbonate depression (D1) characterizes the upper Y zone (Fig. 2). Carbonate values are as much as 25% lower than for peak P1 in the pelagic cores (V25-59 and A180-73) and as much as 45% lower than peak P1 of the hemipelagic cores (V25-56 and V15-168). Depression D1 actually has two minimum values (D1a and D1b), which are separated by a moderately sized peak (P2). Depression D1 apparently correlates with the Late Wisconsin Glacial Maximum (Stage 2 of Emiliiani, 1955).

Relatively high carbonate contents characterize the middle of the Y zone, but these values are normally lower than those observed in the Z and X zones (Fig. 2). Three correlative peaks (P3, P4, and P5), which have values 5 to 20% higher than the values of the depressions (D2, D3) which separate them, are recognized. The zone of relatively high carbonate formed by peaks P3, P4, and P5 appears to correlate with the Middle Wisconsin (Stage 3 of Emiliiani, 1955).

The lower Y zone is marked by a deep carbonate depression (D4), which apparently correlates with the Early Wisconsin Glacial Maximum (Stage 4 of Emiliiani, 1955). Depression D4 is similar to Depression D2 in that it has two minimum values (D4a and D4b) separated by a small peak (P6). The lowermost Y zone contains a steep, sharp peak (P7) that straddles the X/Y boundary; however, the apex of peak P7 is within the Y zone (Fig. 2).

In addition to a portion of peak P7, the X
zone contains two large carbonate peaks (P8 and P9) that are bounded by deep depressions (D5, D6 and D7) (Fig. 2). Values at the bottoms of the depressions are 20 to 40% lower than at the apices of the peaks. Peak P8 generally contains two or three small peaks at its apex. For the pelagic cores (V25-59 and A180-73), the maximum carbonate values (70 to 85%) of peaks P8 and P9 approach those of peak P1. The X zone apparently correlates with the Last Interglacial (Stage 5 of EMILIANI, 1955).

Carbonate values are relatively low in the W zone, although a moderate peak (P10) is observed in the middle of the zone. The W zone apparently corresponds to the glacial maximum (Stage 6 of EMILIANI, 1955) immediately preceding the Last Interglacial.

The nearly identical shape of the four carbonate curves of Fig. 2 indicates that the total carbonate content of sediments has fluctuated in a uniform manner throughout the western Equatorial Atlantic during at least the last 130,000 years. Furthermore, carbonate fluctuations reported for the eastern Equatorial Atlantic (GARDNER, 1973, 1975) and for the Caribbean Sea (PRELL, 1974) are of similar nature and identical timing to those shown in Fig. 2. Although the average carbonate content is lower for the continental rise sediments (cores V25-56 and V15-168) than for the pelagic Mid-Atlantic Ridge sediments (cores V25-59 and A180-73) because of the high influx of terrigenous sediment to the continental margin, the timing and magnitude of the carbonate fluctuations are not significantly different between these two significantly different sedimentary provinces. Thus, although the proportional contribution of each factor controlling total carbonate (dilution, dissolution, and productivity) must vary between sedimentary provinces (e.g. higher dilution on continental rise than Mid-Atlantic Ridge), the three factors must always vary in such a way that their combined effect yields carbonate fluctuations of similar nature and timing across the entire western Equatorial Atlantic without regard to sedimentary province. At present it is impossible to determine the relative contribution of each factor in any core.

The character and timing of the calcium carbonate fluctuations during the past 130,000 years suggests that the fluctuations may accurately reflect the nature and timing of climatic oscillations during this period. For example, the shape of the carbonate curve (Fig. 2) is similar to that of sea-level curves for the last 130,000 years, which have been deduced by dating tectonically uplifted coral reefs on Barbados (STIEDEN, HARRISON and MATTHEWS, 1973, Fig. 6) and New Guinea (BLOOM, BROECKER, CHAPPELL, MATTHEWS and MESOLELLA, 1974, Fig. 5). Therefore, it is of interest to compare in detail the timing of carbonate fluctuations with previously published evidence for climatic change during the last 130,000 years. In Fig. 3, the carbonate fluctuations are compared with radiometrically-dated high sea-level stands, $^{18}$O fluctuations in Equatorial Atlantic sediments, northern hemisphere insolation changes, and ice-margin fluctuations inferred from the continental stratigraphy of eastern North America.

To correlate realistically the carbonate fluctuations with these other lines of evidence, it was necessary to construct an accurate plot of carbonate content versus time for the past 130,000 years (Z, Y, X zones) using cores V25-59 and A180-73. A separate time scale had to be constructed for each core because of the slightly different sedimentation rates in each core. In addition, although the sedimentation rate within each core is essentially constant below the top 1 m of the core, biostratigraphic relationships and radiocarbon dating (Table 2) have demonstrated that rates for approximately the top 1 m

<table>
<thead>
<tr>
<th>Table 2. Radiocarbon dates.</th>
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<tbody>
<tr>
<td>Core A180–73 (from ERICSON et al., 1961)</td>
</tr>
<tr>
<td>0–8 cm 2960 ± 200 yr B.P.</td>
</tr>
<tr>
<td>30–38 cm 15,300 ± 300 yr B.P.</td>
</tr>
<tr>
<td>80–88 cm 27,500 ± 1500 yr B.P.</td>
</tr>
<tr>
<td>Core V25–59 (from Bé et al., in press)</td>
</tr>
<tr>
<td>15–18 cm 6490 ± 130 yr B.P.</td>
</tr>
<tr>
<td>33–38 cm 11,950 ± 280 yr B.P.</td>
</tr>
<tr>
<td>64–68 cm 17,240 ± 530 yr B.P.</td>
</tr>
<tr>
<td>101–108 cm 31,450 ± 1350 yr B.P.</td>
</tr>
</tbody>
</table>
Fig. 3. Correlation of carbonate fractions versus time for cores A180-73 and V25-59 with radiometrically-dated high sea-level stands from islands, oxygen-isotope variations in deep-sea sediments (core A180-73), mean annual rate of change in winter solar irradiation for 25° to 75°N (KUKLA and KUKLA, 1972), and ice-margin fluctuations (stadial-interstadial periods) in the eastern Great Lakes region of North America. Stadial-interstadial chronology (WARM-COLD PERIODS) and subdivisions of the Wisconsin and Last Interglacial (AGE-SUBAGE) are those proposed by MÖRNER (1972a). Section showing ice-margin fluctuations between Quebec and Ohio is simplified and redrawn from MÖRNER (1972b). Bars for sea-level maxima indicate range of ages reported, not error in single age determinations.
of core are apparently higher than for the rest of the core. This apparent increase in rate occurs because the sediments of the uppermost 1 m of the sea floor have not yet completely dried out or been compacted as much as the underlying sediments. Fortunately, the radiocarbon dates (Table 2) for several intervals within the upper 1 m of each core permit accurate time scales to be constructed for this portion of each core.

A time scale for each core was constructed by assuming a constant sedimentation rate for each interval of sediment between the W/X (127,000 yr B.P.), X/Y (85,000 yr B.P.), and Y/Z (11,000 yr B.P.) boundaries, and the radiocarbon-dated intervals, and then interpolating dates within each of these intervals. A date for each carbonate sample was then ascertainable from these time scales and these dates were plotted (Fig. 3) to give accurate curves of carbonate content versus time for cores V25-59 and A180-73.

Recently, uplifted coral reefs on Barbados and New Guinea have been used to document glacio-eustatic sea-level maxima and, hence, climatic oscillations. Broecker, Thunber, Goddard, Ku, Matthews and Mesolella (1968) and Mesolella, Matthews, Broecker and Thunber (1969) have dated sea-level maxima on Barbados at 85,000; 105,000; and 125,000 yr B.P. These maxima correlate well in time with peaks P7, P8 and P9 of the carbonate curve (Fig. 3). More recently, James, Mountjoy and Omura (1971) reported a 60,000 yr B.P. reef terrace, which suggests a sea-level maximum during the Middle Wisconsin and approximately correlates with peak P5 of the carbonate curve. In addition, Steinen, Harrison and Matthews (1973) have presented evidence from the subsurface of Barbados for a low sea-level stand between 105,000 and 125,000 yr B.P. This sea-level minimum is recorded in the carbonate curve as depression D6 (Fig. 3).

On the Huon Peninsula of New Guinea, Veeh and Chappell (1970) dated a flight of raised coral terraces which indicate sea-level maxima at 28,000 to 30,000; 35,000 to 50,000; 60,000; 74,000; and 118,000 to 140,000 yr B.P. More recently Chappell (1974) published slightly revised dates for the same terraces. These dates indicate sea-level maxima at 6,500; 30,000; 40,000 to 50,000; 60,000; 80,000; 105,000, and 120,000 yr B.P. These revised dates correlate well with carbonate peaks P3, P4, P5, P6, P7, and P8, although P5 appears to be slightly younger than 60,000 yr B.P. (Fig. 3).

Konishi, Schlanger and Omura (1970) radiometrically dated coralline limestone formations on the Ryukyu Islands of the western Pacific which indicate sea-level maxima at 38,000 to 44,000 and 59,000 to 64,000 yr B.P. These sea-level maxima approximately correlate with peaks P4 and P5 of the carbonate curve.

In addition to the sea-level maxima of 82,000; 105,000; and 125,000 yr B.P., the dated reefs from Barbados, New Guinea, and Ryukyu indicate a series of sea-level maxima during the Middle Wisconsin (Y zone) between 25,000 and 65,000 yr B.P. Three separate maxima seem to have occurred during this period at approximately 28,000 to 30,000; 35,000 to 50,000, and 55,000 to 65,000 yr B.P. These dates approximately correspond in time to the occurrence of peaks P3, P4, and P5 of the carbonate curve (Fig. 3). There are no sea-level maxima reported for the Late (~ 16,000 to 25,000 yr B.P.) or Early (~ 65,000 to 75,000 yr B.P.) Wisconsin. Veeh and Chappell (1970) reported evidence of a sea-level minimum between 15,000 to 20,000 yr B.P.

The good correlation between carbonate peaks and radiometrically-dated sea-level maxima suggest that the carbonate fluctuations accurately reflect the timing of glacio-eustatic sea-level oscillations and, hence, climatic oscillations. The carbonate curve thus suggests the following chronology for the last 130,000 years. During the Wisconsin (Y zone) large sea-level minima occurred between 12,000 and 24,000 yr B.P. (depression D1, Late Wisconsin Glacial Maximum) and between 61,000 and 74,000 yr B.P. (depression D4, Early Wisconsin Glacial Maximum). Sea level was high for much of the Middle Wisconsin but there were three separate maxima at approximately 28,000 to 32,000; 39,000 to 42,000, and 49,000 to 53,000 yr B.P. as indicated by peaks P3, P4, and P5. These sea-level
maxima are separated by brief lowerings of sea-
level during 35,000 to 38,000 yr B.P. (depression
D2) and 44,000 to 47,000 yr B.P. (depression D3).
During the Last Interglacial (X zone), prominent
maxima occur at about 80,000 to 84,000 yr B.P.
(P.7); 90,000 to 105,000 yr B.P. (P8) and 115,000
to 125,000 yr B.P. (P9). Large sea-level minima
occurred between these high sea-level stands at
about 85,000 to 88,000 yr B.P. (D5); 106,000 to
110,000 yr B.P. (D6); and 126,000 to 131,000 yr
B.P. (D7). Although the carbonate fluctuations
apparently reflect the general magnitude and
timing of the sea-level fluctuations, there is no
indication that the observed magnitudes of the
carbonate fluctuations are directly proportional
to the magnitudes of the corresponding sea-level
fluctuations.

A further indication that the carbonate fluctuations accurately reflect climatic oscillations is indicated by the excellent correlation of the carbonate curves with fluctuations in solar energy received in the northern hemisphere during the last 130,000 years (Fig. 3). KUKLA and KUKLA (1972) presented an insolation curve showing the rate of change in solar irradiation at the top of the atmosphere (25 to 75°N) during the northern winter. The portion of their curve covering the last 130,000 years is redrawn in Fig. 3 beside the carbonate curves. KUKLA and KUKLA (1972) have shown that intervals of positive insolation correlate with warm climatic intervals as recorded in the deep-sea and continental records, whereas intervals of negative insolation correlate with cold climatic intervals. Figure 3 demonstrates that carbonate peaks correlate precisely with positive insolation peaks, whereas carbonate depressions correlate precisely with negative insolation depressions.

The timing of climatic fluctuations during the last 130,000 years has also been inferred from continental deposits in the eastern Great Lakes–
St. Lawrence region of North America. These deposits record ice-margin fluctuations of the Laurentide Ice Sheet, which expanded and shrank in response to climatic oscillations (GOLDTHWAIT, DREIMANIS, FORSYTH, KARROW and WHITE, 1965; DREIMANIS, 1969; DREIMANIS and KARROW, 1972; MÖRNER, 1972a, b; DREIMANIS and GOLDTHWAIT, 1973). Recently, a chronology for these Wisconsin ice-margin fluctuations and corresponding stadal–interstadial periods based on radiocarbon dating of the deposits has been presented (DREIMANIS and KARROW, 1972; DREIMANIS and GOLDTHWAIT, 1973). During the present study this chronology was compared to the carbonate chronology of Fig. 3. A good correlation was apparent back to only about 30,000 yr B.P., with ice-advances coinciding with carbonate depressions and ice recessions coinciding with carbonate peaks. Prior to 30,000 yr B.P., however, there was no correlation. This is not surprising because the effective limit of radiocarbon dating is about 30,000 yr B.P.

MÖRNER (1972a) emphasized the unreliability of radiocarbon dates older than 30,000 yr B.P. and proposed a chronology and classification for stadial–interstadial periods of the last 130,000 years based on the solar insolation chronology of KUKLA (1969). Figure 3 shows Mörner’s stadial–interstadial periods (WARM–COLD PERIODS) and proposed sub-division of the Wisconsin and Last Interglacial (AGE–SUBAGE) together with a generalized section showing ice-margin fluctuations between Quebec and Ohio (redrawn from MÖRNER, 1972b). MÖRNER (1972a, b) emphasized that this record of climatic change from North America also correlates in detail with the record in Europe and thus reflects global climatic changes.

There is excellent correlation between the calcium carbonate curves, Mörner’s stadial–interstadial periods, and ice-margin fluctuations (Fig. 3). Calcium carbonate peaks correlate precisely with interstadials (warm periods) and ice recessions, whereas carbonate depressions correlate with stadials (cold periods) and ice advances. The prominent calcium carbonate depression (D1) in the upper Y zone correlates with the Late Wisconsin Glacial Maximum. Peak P2 may record the Erie Interstade (~15,500 to 16,500 yr B.P.; MÖRNER, 1972a). The zone of high carbonate in the middle of the Y zone correlates peak for peak with the series of three recognized Middle Wisconsin interstades: Peak P3 records the Plum Point; Peak P4, the Port Talbot II; and Peak P5, the Port Talbot I. Depressions (D2 and
D3) between these peaks correlate with short stadials. The prominent carbonate depression (D4) in the lower Y zone correlates with the Early Wisconsin Glacial Maximum. Peak P7 correlates with the St. Pierre Interstadale whereas Peaks P8 and P9 correlate with interstadials of the Last Interglacial (Eemian II and I). Depressions (D5, D6, D7) between these peaks correlate with stadial periods. This excellent correlation suggests that the calcium carbonate fluctuations may accurately reflect northern-hemisphere ice-margin fluctuations and that Mörner’s chronology for the Wisconsin may be essentially correct.

Variations of the $^{18}O$/$^{16}O$ ratio in the shells of planktonic foraminifera have also been interpreted as indicators of glacial–interglacial seawater-temperature fluctuations (EMILIANI, 1955) and ice-volume fluctuations (BROECKER and VAN DONK, 1970). A plot of $^{18}O$ variations versus time for the last 130,000 years for core A180-73 is shown in Fig. 3. The $^{18}O$ curve for core A180-73 as well as published curves for several other Atlantic and Caribbean cores which span the last 175,000 years (see BROECKER and VAN DONK, 1970, Fig. 1) do not consistently show well-defined peaks and depressions that can be correlated from core to core or that correlate with the calcium carbonate peaks and depressions, insolation chronology, or the continental ice-margin fluctuations of Fig. 3. For example, the $^{18}O$ curve for A180-73 fails to show a peak that correlates with peak P8 of the corresponding carbonate curve (although some other $^{18}O$ curves do show this peak). In addition, the $^{18}O$ curve for A180-73 fails to show the detail of the Middle Wisconsin (Y zone) stadial–interstadial complex as is evidenced in the carbonate curve (peaks P3, P4, P5). Most published $^{18}O$ curves fail to reflect the magnitude of the Early Wisconsin Glacial Maximum as reflected by the carbonate curves (depression D4) and continental deposits. This lack of reproducible detail in $^{18}O$ curves suggests that although $^{18}O$ variations provide a general record of world-wide ice-volume fluctuations, they do not provide a precise record of the timing of stadial–interstadial periods and northern hemisphere ice-margin fluctuations as do the calcium carbonate curves.

The correlations described above and shown in Fig. 3 thus suggest that the calcium carbonate fluctuations observed in western Equatorial Atlantic sediments may accurately reflect the nature and timing of climatic oscillations during the last 130,000 years. The carbonate fluctuations suggest that warm–cold (stadial–interstadial) cycles have occurred with a periodicity of about 20,000 years. A similar periodicity has previously been reported for insolation cycles (KUKLA and KUKLA, 1972) as well as for sea-level maxima (BLOOM, BROECKER, CHAPPELL, MATTHEWS and MesoLLA, 1974). Because the Earth has experienced a warm phase (interstadial) during the last 10,000 years, which apparently reached its zenith about 5000 to 6000 years ago (peak P1 of the carbonate curve), the Earth should undergo a period of increased cooling (stadial) for approximately the next 10,000 years.

**CARBONATE FLUCTUATIONS AND CLIMATE PRIOR TO 130,000 yr B.P.**

Calcium carbonate and coarse fraction (> 62 μm) contents along with the frequency of abundance of foraminifera of the *Globorotalia menardii* complex are shown for cores V25-60 and V25-42 in Fig. 4. These cores are continuous back to the U zone (400,000 to 550,000 yr B.P.). Core V25-60 is a fairly shallow (3749 m) Mid-Atlantic Ridge core composed of foraminiferal marl and ooze, whereas core V25-42 is a deep (4704 m) abyssal core from a small basement knoll on the Demerara Abyssal Plain and is composed of pelagic brown clay and marl (Fig. 1).

The carbonate content of V25-60 generally ranges between 45 and 75% with peak-to-depression variations as great as 40% (Fig. 4). For climatic zones Z to W (0 to 165,000 yr B.P.) not all of the ten carbonate peaks and seven depressions described in the previous section are well defined. The absence or poor definition of some peaks may be attributed to the relatively low sedimentation rate of this core ($\sim 2$ cm ($10^3$ yr)$^{-1}$), which effectively increases the time interval between the 10-cm sample locations and also permits greater homogenization of sediments.
Fig. 4. Calcium carbonate and coarse fraction percentages plus abundances of G. menardii complex versus depth in core (meters) for cores V25-60 and V25-42. Lengths of climatic zones in these cores have been normalized to those of core V12-122, which is shown on the left. Dotted lines correlate climatic-zone boundaries from core to core. Dates for climatic-zone boundaries are from Broecker and van Donk (1970) except for the X/Y boundary (Damuth, 1973 and the present study). Oxygen-isotope curve for V12-122 is replotted from Broecker and van Donk (1970); carbonate and coarse fraction curves are from Prell (1974); and G. menardii curve is from Ericson and Wollin (1968). Times of maximum ice build-up on the continents (glacial maxima) as inferred from the oxygen-isotope variations in core V12-122 by Broecker and van Donk (1970) are labelled G1, G2, etc.; interglacial phases are labelled I1, I2, etc.
by benthic organisms. Winnowing by bottom currents may also have altered the carbonate content of certain intervals.

The carbonate content of core V25-42 shows a much wider variation than in V25-60 and has values that range from less than 1 to 74% (Fig. 4). The location and depth, the widely fluctuating carbonate and coarse fraction contents, and the G. menardii curve all suggest that the sediments of V25-42 have undergone cycles of extreme dissolution by the Antarctic Bottom Water. Dissolution is further emphasized by samples from the brown clay (low carbonate) beds of the core in which only a few specimens of the most resistant foraminiferal species (P. obliquiloculata and G. menardii tumida) are preserved.

The nature and timing of climatic oscillations prior to the Last Interglacial are not yet known in as much detail as are oscillations of the last 130,000 years. Temporal variations of $^{18}O/^{16}O$ ratios in deep-sea sediments during the last 600,000 years have revealed glacial–interglacial cycles with a periodicity of approximately 100,000 years (Emiliani, 1955, 1966; Broecker and Van Donk, 1970; Shackleton and Opdyke, 1973). Thus to evaluate the relationship of carbonate fluctuations in cores V25-60 and V25-42 to climatic oscillations during the last 500,000 years, it is necessary to compare these cores to one of similar sedimentation rate and time span for which the $^{18}O$ variations are known. V12-122, a well-studied Caribbean core, has a time span and sedimentation rate similar to those of V25-42 and V25-60. Oxygen isotope (Broecker and Van Donk, 1970), carbonate, coarse fraction (Prelle, 1974), and G. menardii (Ericson and Wollin, 1968) curves have been previously published for V12-122 (Fig. 4). In Fig. 4 the lengths of the climatic zones (Z, Y, etc.) of cores V25-60 and V25-42 have been proportionally expanded or compressed so that they correspond to the equivalent climatic-zone lengths of core V12-122. Thus isochronism of curve peaks and depressions from core to core will be more readily apparent.

Periods of maximum glacial advance (maximum $^{18}O$ content) during the last 500,000 years, as inferred by Broecker and Van Donk (1970) from the $^{18}O$ curve of V12-122, are labelled for convenience as G1, G2, etc., in Fig. 4. Likewise, periods of maximum deglaciation (terminations or interglacials), indicated by minimum $^{18}O$ contents, are labelled I1, I2, etc. Comparison of the carbonate and coarse fraction curves for cores V12-122 and V25-60 with the $^{18}O$ variations indicates that the carbonate and coarse fraction fluctuations in these two cores do not reflect the 100,000-year glacial–interglacial cycles as dramatically as the $^{18}O$ variations (Fig. 4). Although carbonate peaks and depressions appear to correlate with interglacials (I1, I2, etc.) and glacials (G1, G2, etc.) these curves do not show the definitive saw-tooth shape of the $^{18}O$ curve. Perhaps in addition to the 100,000-year cycles these carbonate curves also show some of the 20,000-year climatic cycles revealed by the higher sedimentation-rate cores (Fig. 3) of the previous section. Whether or not this is the case, it would be difficult, if not impossible, to discern the nature and timing of glacial–interglacial cycles from only the carbonate curves of these two cores. In contrast, the G. menardii curves of both cores, especially that of V12-122, seem to mimic more closely the shape of the $^{18}O$ curve (and hence the glacial–interglacial cycles) than do the carbonate or coarse fraction curves.

In contrast to core V25-60, the carbonate fluctuations, as well as the coarse fraction and G. menardii fluctuations of core V25-42, clearly reflect the 100,000-year glacial–interglacial cycles as revealed by the $^{18}O$ curve (Fig. 4). During interglacial phases (I1, I2, etc.) carbonate contents generally reach values greater than 40%, whereas during glacial phases (G1, G2, etc.) carbonate values are often less than 1%. The coarse fraction curve also clearly reflects the six glacial–interglacial cycles, although not so dramatically as the carbonate curve. In addition, interglacial phases I1 to I5 are reflected by sharp peaks in the G. menardii curve, whereas G. menardii are absent during the intervening glacial phases. In normal pelagic sediments such as those of cores V25-60 and V12-122, the G. menardii complex is absent only in the Y, W, and U zones, but it is abundant throughout the V zone, which
they are climatically controlled, i.e. relatively greater dissolution during glacial phases. The ramifications of these observations will be discussed more fully in the following section.

Calcium carbonate, coarse fraction, and *G. menardii* variations for core V25-45 from the Mid-Atlantic Ridge (Fig. 1) are shown in Fig. 5. Comparison of the *G. menardii* curve for this core with the *G. menardii* curve for a nearby core, V16-205 (ERICSON and WOLLIN, 1968), which also has paleomagnetic stratigraphic control, reveals that V25-45 contains a continuous sedimentary record for the last 1,800,000 years (zones Z to Q), or nearly the entire Quaternary. The carbonate content is uniformly high and ranges from 70 to 85%. The relationship of the carbonate variations to Quaternary climatic oscillations is uncertain because the extremely low sedimentation rate \( \sim 0.5 \text{ cm (10^3 yr)}^{-1} \) obscures the detail of the carbonate and coarse fraction curves and makes comparison with the other curves for the last 400,000 years (cores V25-60 and V25-42) difficult; comparison with curves for the last 175,000 years (cores A180-73 and V25-59) is impossible. The carbonate peaks of the X to U zones of V25-45 seem to correlate in frequency and time with the peaks of low \(^{18}O\) content (interglacials) of core V12-122 (Fig. 4). If each major carbonate peak below the T/U boundary in core V25-45 reflects a glacial–interglacial cycle, then approximately 10 to 15 cycles have occurred between the beginning of the Pleistocene and 550,000 years ago (T/U boundary). Thus 15 to 20 glacial–interglacial cycles have occurred during the entire Quaternary.

VAN DONK (1973) made close-interval (10-cm) \(^{18}O\) determinations down core V16-205 from a site approximately 500 km north of V25-45. V16-205 also contains a sedimentary record of the entire Quaternary (ERICSON and WOLLIN, 1968) and has a sedimentation rate \(0.5 \text{ cm (10^3 yr)}^{-1}\) identical to V25-45. Van Donk reports that the \(^{18}O\) record for V16-205 shows about 20 glacial–interglacial cycles during the Quaternary. Thus the carbonate fluctuations of core V25-45 apparently reflect the frequency and timing of Quaternary glacial–interglacial cycles. Close-interval \(^{18}O\) determinations down this core span three complete glacial–interglacial cycles. In contrast, the *G. menardii* complex is absent during glacial phases (G, G4, G5) of the V zone because severe dissolution during the glacial has destroyed the foraminiferal tests. The shapes of the carbonate, coarse fraction, and *G. menardii* curves of core V25-42 thus seem to be controlled largely by fluctuations in the rate of carbonate dissolution by the Antarctic Bottom Water. The timing of these dissolution cycles suggests that
could more precisely reveal the exact frequency and timing of glacial-interglacial cycles during the Quaternary as well as their correlation with calcium carbonate fluctuations.

**Changes in Oceanic Circulation Rates in Response to Climatic Oscillations**

Alternations between glacial and interglacial climate imply changes in the rates of atmospheric circulation, especially in the tradewind belts. Atmospheric changes should, in turn, cause changes in rates of oceanic circulation and mixing. Whether the ocean mixed faster or slower during glacial times than during interglacials is as yet unresolved. Arrhenius (1952) correlated high carbonate zones in eastern Equatorial Pacific sediments with glacial phases and postulated that increased atmospheric circulation during glacial times caused increased upwelling of nutrient-rich deep waters and thus caused increased productivity. In contrast, Broecker (1971) suggested that variations in calcite and aragonite accumulation in the sediments of the World Ocean indicate that carbonate production was greater during interglacials than glacial, and thus that oceanic circulation was faster during interglacials. Evidence has also been presented for increased production and strengthening in circulation of the Antarctic Bottom Water during glacials (Olausson, 1960; Berger, 1968; Johnson, 1973; Gardner, 1975). Two cores from the present study, V25-76 and V25-42, provide evidence that oceanic circulation in the Atlantic was faster during glacial phases than at present or during the previous interglacials.

Core V25-76 apparently provides direct evidence that the surface circulation and possibly the circulation of the underlying Antarctic Intermediate Water were faster during the Wisconsin Glacial (Y zone) and previous glacial (W zone) than during the Last Interglacial (X zone) and the Holocene (Z zone). V25-76 is from the continental slope (2017 m) just below the outer edge of the Demerara Plateau (Fig. 1) and is composed of gray calcareous clay. The coarse-fraction component (> 62 μm) that is disseminated throughout the clay consists largely of foraminifer tests with minor amounts of pyrite, hydroterroilite, organic detritus, and occasional fragments of pteropod or other mollusc shells. Mineral grains are absent or rare. Foraminifer tests are commonly fragmented and broken, suggesting reworking by currents. Coarse elasic beds (silt-sand) are absent from the core with the exception of a thin (< 1 cm), apparently winnowed bed comprised of pteropod and foraminifer tests and fragments at 75 to 76 cm below the core top.

The calcium carbonate, coarse fraction, and G. menardii variations down core V25-76 are shown in Fig. 6. These curves are compared with carbonate, coarse fraction, and G. menardii curves for core V25-59, which was discussed in a previous section (see Figs. 2 and 3). V25-76 is compared with V25-59 because the latter displays the carbonate, coarse fraction, and G. menardii variations that are normally observed in the pelagic and hemipelagic sediments throughout the western Equatorial Atlantic (including cores A180-73, V25-56, V15-168 of this study, Fig. 2; also see Damuth, 1973). When compared with V25-59, it is apparent that the carbonate, coarse fraction, and G. menardii fluctuations, as well as the proportional lengths of the climatic zones (Z, Y, X, W) of V25-76 are anomalous.

For core V25-76, the X zone (Last Interglacial) and Z zone (Holocene) are abnormally expanded, whereas the Y zone (Wisconsin Glacial) and the W zone are abnormally short (Fig. 6). The relationships are exactly the reverse of those normally observed for cores of this region (especially on the continental slope and rise); in nearly all other cores the glacial sections (Y and W zones) are expanded in relation to the interglacial sections (e.g. V25-59). Sedimentation rates for the Y and W zones of V25-76 are only 0.95 and 1.05 cm (10^3 yr)^{-1}. In contrast, normal pelagic rates for the Y and W zones are 2 to 5 cm (10^3 yr)^{-1} and rates for hemipelagic cores of the continental slope and rise are generally greater than 5 cm (10^3 yr)^{-1} (Fig. 2; see also Damuth, 1973; McGeaney and Damuth, 1973; Damuth and Kumar, 1975). Thus at the site of core V25-76,
Fig. 6. Calcium carbonate and coarse fraction fractions plus abundance of *G. menardii* complex versus depth in core (meters) for cores V25-59 and V25-76. Dates for climatic zone boundaries are from Broecker and Van Donk (1970) except for the X/Y boundary which is from Damuth (1973) and the present study.

Sediment accumulation during glacial phases was, for some reason, anomalously low. In contrast, sedimentation rates during the interglacial phases (Z and X zones) at site V25-76 were higher than normal. The rate for the Z zone is 5.9 cm (10³ yr⁻¹), whereas the X-zone rate is 6.7 cm (10³ yr⁻¹). Normally, interglacial rates throughout the western Equatorial Atlantic are less than 5 cm (10³ yr⁻¹) (Damuth, 1973).

The carbonate, coarse fraction, and *G. menardii* curves of V25-76 do not show the characteristic fluctuations normally observed in western Equatorial Atlantic sediments of the last 130,000 years, e.g. V25-59 (Fig. 6). Carbonate and coarse fraction contents are uniformly low (< 15%) during interglacial phases (Z, X, uppermost V zones), except for the basal portion of the X zone and in the Z zone where sharp peaks occur. In contrast, glacial phases (Y and W zones) are marked by increased carbonate and coarse fraction contents. This is the opposite of the relationship normally observed in pelagic and hemipelagic sediments (e.g. V25-59) of the western Equatorial Atlantic.

The only explanation that appears to account for the anomalous relationships observed in core V25-76 is that the surface waters of the western Equatorial Atlantic circulated faster during glacial than during interglacial phases. Presently, at depths of 0 to 1000 m, strong northwestward-setting currents flow parallel to the South American coast between the Amazon River and Trinidad. At the surface, the Guiana Current attains speeds of 50 to 150 cm s⁻¹ (1 to 3 knots), but speeds decrease with depth to the top of the Antarctic Intermediate Water, which flows northwestward between 700- and 900-m depth at speeds of 6 to 12 cm s⁻¹ (Defant, 1961; Wüst, 1963). These two currents, especially the Guiana Current, transported large quantities of fine terrigenous sediment from the Amazon River to the region of core site V25-76. Deposition of this sediment diluted the normal pelagic sediments being deposited to produce the somewhat expanded Z, X, and uppermost V zones observed in V25-76, as well as the low carbonate and coarse fraction contents. The low carbonate and coarse fraction contents of these interglacial zones cannot be explained simply by carbonate dissolution because (1) the core site is too shallow...
(2107 m) and (2) dissolution alone would produce shortened interglacial sections, rather than the observed expanded sections. The abundant disseminated silt and shell fragments within the core are further evidence that dilution by suspended terrigenous sediment has expanded the interglacial sections.

In contrast to the expanded interglacial sections, the shortened glacial sections (Y and W zones) of core V25-76 indicate that deposition at the core site slowed drastically during the glacial phases. The only explanation for these abnormally low rates seems to be that the currents over the core site increased in speed during the glacial phases to such an extent that they prevented a large fraction of the fine terrigenous sediment, which was in suspension, from being deposited. This effect greatly reduced glacial accumulation rates on the outer Demerara Plateau and produced the anomalously short glacial sections of core V25-76. The stronger currents may also have eroded some sediment after deposition at the core site. The winnowing of a large fraction of the fine terrigenous sediment also caused the observed increases in the carbonate and coarse fraction (largely foraminifera) contents of the Y and W zones of core V25-76.

Alternative explanations for the shortened glacial sections might be (1) that current speeds actually decreased during the glacial periods (relative to the interglacials) because of a general decrease in the rate of oceanic mixing; or (2) that less terrigenous material was injected into the western Equatorial Atlantic during the glacial periods than during the interglacial periods. In either case, less terrigenous sediment would have been transported to the outer edge of the Demerara Plateau (V25-76 coring site) and shortened glacial sections would have resulted. Neither of these explanations is valid, however, because (1) the highest rates of terrigenous sedimentation in the western Equatorial Atlantic occurred when sea level was lowered during the glacial phases, not during interglacials (DAMUTH, 1973; McGEEARY and DAMUTH, 1973; DAMUTH and KUMAR, 1975); and (2) even if removal of terrigenous sediment to the deep-sea provinces had been completely stopped during glacials, the normal pelagic sedimentation rate [~ 3 to 5 cm (10^3 yr)^{-1}] for the Demerara Plateau region should still have been several times greater than the observed rates [~ 1 cm (10^3 yr)^{-1}], and thus glacial sections would be thicker than those actually observed.

Therefore, the only explanation that seems to account for the stratigraphic relationships observed in core V25-76 is that circulation within the surface waters of the Atlantic must have been faster during the Pleistocene glacial phases than during the present or previous interglacial phases. If this interpretation is correct, then core V25-76 provides direct evidence that rates of oceanic mixing in the Atlantic Ocean have fluctuated in response to climatic change.

Evidence that the deep waters, especially the Antarctic Bottom Water, circulated faster during glacial phases than during interglacial phases is provided by core V25-42 (Fig. 4). As described in the previous section, the shapes of the carbonate, coarse fraction, and G. menardii curves of this core reflect periods of intense carbonate dissolution. The timing of these dissolution cycles correlates with glacial–interglacial climatic oscillations. Periods of extreme dissolution (carbonate and coarse fraction depressions) occurred during glacial (G1, G2, etc., Fig. 4), whereas relatively minor dissolution (carbonate, coarse fraction, and G. menardii peaks) occurred during interglacials (II, I2, etc.).

The location and depth (4704 m) of the V25-42 core site suggest that dissolution was accomplished by the northward-flowing Antarctic Bottom Water (AABW). A potential temperature of 1.848°C was recorded at 25 m above the sea floor during coring of V25-42, indicating that the core site lies under the AABW (potential temperature = < 1.9°C). The relatively high carbonate content (> 45%) of the Holocene section of V25-42 (II, Fig. 4) thus suggests that the AABW has been less corrosive during this interglacial period than it was during the Wisconsin Glacial (G1, Fig. 4) when carbonate-poor (< 25%) sediments were accumulating at the core site. Thus the carbonate dissolution cycles observed down core V25-42 must have resulted
from oscillations in the corrosive strength of the AABW. The fact that the AABW was more corrosive during glacial than during interglacial phases indicates that the AABW underwent increased production and hence increased circulation during glacial phases. Thus the dissolution cycles observed in core V25-42 provide evidence that circulation of the deep waters of the Atlantic increased during glacial phases relative to interglacial phases.

Carbonate dissolution cycles that are apparently similar in timing to those of V25-42 have recently been described for eastern Equatorial Atlantic sediments by Gardner (1973, 1975). Gardner attributed the dissolution cycles to increased production and circulation of AABW during glacial phases and possibly to the development during glacial phases of a corrosive North Atlantic Bottom Water that flowed southward along the eastern Atlantic basins.

CONCLUSIONS

Despite the differences in sedimentary environment, identical carbonate fluctuations are observed for the last 130,000 years in both pelagic oozes from the Mid-Atlantic Ridge (cores V25-59 and A180-73) and terrigenous-rich hemipelagic clays from the continental margin (cores V25-56 and V15-168). In each of the four cores, ten correlative carbonate peaks and seven depressions accurately reflect the timing of climatic oscillations during the Last Interglacial, Wisconsin Glacial, and Holocene. Carbonate maxima and minima reveal a 20,000-year periodicity for warm–cold (interstadial–stadial) cycles. The timing of the carbonate fluctuations correlates in detail with other evidence for climatic oscillations including sea-level maxima inferred from radiometrically-dated coral reefs; fluctuations in solar irradiation received in the northern hemisphere; and North American ice-margin fluctuations inferred from continental deposits.

The carbonate fluctuations in cores that span the last 500,000 to 1,800,000 years do not clearly show the 20,000-year climatic cycles (as do the cores that span only the last 130,000 years) because of the low sedimentation rates [< 2 cm (10^3 yr)^{-1}]. However, comparison of the carbonate curves of cores V25-60 and V25-42, which span the last 500,000 years, with published δ^18O variations for this time period reveals that the carbonate fluctuations seem to reflect the 100,000-year glacial–interglacial cycle that has been inferred from oxygen isotope studies. This 100,000-year periodicity is especially apparent in core V25-42 because the glacial-age sediments of the core have undergone severe carbonate dissolution. In core V25-45, which contains a continuous sedimentary record of the last 1,800,000 years, the carbonate fluctuations also appear to reflect the 100,000-year glacial–interglacial cycle. If so, then the carbonate fluctuations indicate that between 15 and 20 cycles have occurred during the Pleistocene.

In addition to providing evidence for the nature and timing of climatic oscillations, the carbonate fluctuations also provide evidence that the Atlantic Ocean circulated faster during glacial phases than during interglacial phases. A shallow (2000-m) continental-slope core, V25-76, from the outer edge of the Demerara Plateau contains anomalous carbonate, coarse fraction, and G. menardii fluctuations plus expanded interglacial and abnormally shortened glacial sections. The only explanation that appears to account for these anomalies is that the surface waters over the core site circulated faster during glacial phases than at present or during previous interglacials. The sediments of core V25-42, a deep (4700-m) abyssal core, were deposited beneath the Antarctic Bottom Water and have undergone several cycles of carbonate dissolution. Periods of severe dissolution occurred during glacial phases, whereas periods of relatively minor dissolution occurred during interglacials. These dissolution cycles have apparently been caused by an increase in production and hence faster circulation of the Antarctic Bottom Water during glacial.

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