Glacial–Holocene stratigraphy, chronology, and paleoceanographic observations on some North Atlantic sediment drifts

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Abstract—Several sediment drifts in the deep North Atlantic have been studied using oxygen isotopes, percent CaCO₃, clay mineralogy, tephra content and Accelerator Mass Spectrometer (AMS) radiocarbon dating. Because enhanced deposition of fine-grained terrigenous sediment increases rates of sedimentation by as much as 100 times the regional mean, sediment drifts contain high-resolution records of climate and oceanographic change. The major results of this study are that: (1) patterns of sedimentation rate are regionally variable. In the Western North Atlantic (Bermuda Rise, Blake–Bahama Outer Ridges), rates of sedimentation have been relatively low during the Holocene, but were as high as 200 cm/1000 years during the latest glacial episode, reflecting a greater supply of sediment due to glaciation and sea level lowering. Deep circulation patterns were probably similar to today. In the northern North Atlantic (Gardar Drift), which is dominated today by Norwegian Sea Overflow Water (NSOW), sedimentation rates were lowest during the glacial, despite the availability of sediment, because NSOW production was stopped. Rates peaked during the early Holocene as NSOW resumed. Glacial and Holocene rates of sedimentation were roughly comparable in the northeast Atlantic (Feni Drift) because sedimentation at that location is influenced less by NSOW than by Northeast Atlantic Deep Water (NEADW), which has a large southern-source component. (2) Stable isotopic events of deglacial age are preserved with unprecedented clarity in the sediment of North Atlantic drifts. Where the Younger Dryas cooling and the Vedde Ash occurred together, at 56°N on Feni Drift, direct AMS dating suggests that the radiocarbon age difference between the surface ocean and atmosphere reservoirs in the northeast Atlantic was the same 10,500 years ago as it is today (about 400 years). A δ¹⁸O maximum of Younger Dryas age indicates cooler or more saline surface waters above the Bermuda Rise at 33°N and as far south as 28°N on the Bahama Outer Ridge. Two δ¹⁸O minima at 12,000 and 13,500 radiocarbon years BP on the Bahama Outer Ridge and the Bermuda Rise probably resulted from lowered surface water salinity. These low salinity events most likely originated as meltwater discharge down the Mississippi and into the Gulf of Mexico, followed by advection into the open North Atlantic via the Gulf Stream.

INTRODUCTION

CONSTRUCTIONAL sedimentary deposits in the deep sea, known as sediment drifts, offer a unique opportunity for high-resolution study of the paleo-ocean once a detailed stratigraphy and chronology are established for representative cores. In this paper we describe the stratigraphy and chronology of cores from several sediment drifts in the North Atlantic Ocean, and use our stratigraphic data to draw some preliminary conclusions about short-duration events of climate and ocean chemistry change during late Quaternary time.

Sediment drifts are found in many locations in the North Atlantic because of the abundant supply of sediment to that basin and the diverse sources of deep water that

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redistribute and focus the sediment (Fig. 1; MCCAVE and TUChOLKE, 1987). In fact, there is at least one sediment drift associated with every deep water mass in the North Atlantic. Most of the sediment in these drifts is in the fine silt and clay fractions, which are distributed by currents that are generally less than 15 cm s\(^{-1}\) (MCCAVE and TUChOLKE, 1986). Larger particles, mostly microfossil tests, are generally deposited \textit{in situ}. Thus, study of the fine fraction (<63 \mu m) of drift sediment has the potential to reveal the provenance of the sediment making up the drift deposit, and hence current direction, and study of the coarse fraction (mostly foraminifera) can tell about surface and deep ocean conditions.

In general, the ability to resolve brief events in the sediment record of ocean and climate history is limited by the intensity of bioturbation and by the rate at which sediment accumulates. Ideally, one would like to study sediment cores from areas with the most rapid sediment accumulation and with the least bioturbation. But where sediment accumulates most rapidly, on continental margins, there is usually an active benthos and sedimentation is marked by downslope transport of coarse material. Furthermore, most nearshore locations probably reflect local oceanographic conditions rather than open ocean conditions. Offshore locations with high pelagic rain rates usually underlie fertile waters that support an active benthos. Given these conditions, sediment drifts have the most to offer for high resolution paleoceanographic studies.

Despite the potential of high temporal resolution, the paleo-ocean record from sediment drifts has received relatively little attention. One of the first studies to exploit the high resolution achievable in cores of drift sediment was that of PISIAS et al. (1973). Two of the three cores this group examined came from the northeast Atlantic (V23-81, V23-82; Fig. 1), and significant spectral peaks were found in sediment composition in the several hundred to several thousand year range. Resolving such short periodicities was possible because sedimentation rates in the upper parts of those cores exceed 10 cm/1000 years, whereas typical rates in the North Atlantic are 2–3 cm/1000 years. By focusing on the continental rise off Chesapeake Bay, BALSAM and HEUSSER (1976) reported details in the records of terrestrial and marine climate which were unattainable in typical North Atlantic sediment cores. In cores from the Gardar Drift, on the eastern flank of the Reykjanes Ridge, RUDDIMAN and BOWLES (1976) found very large changes in sedimentation rates during the earliest Holocene. These were attributed to changes in the supply of sediment from northern sources and changes in deep-sea circulation patterns, which would not have been evident in cores dominated by pelagic sedimentation.

DUPLESSY et al. (1981) also studied events in the marine record of climate, using stable isotopes and radiocarbon in cores from the northeast Atlantic Ocean. The key core in their paper, CH73-139C, came from near V23-81 on the Feni Drift, and has been useful in subsequent chronological investigations using accelerator radiocarbon dating (DUPLESSY et al., 1986; BARD et al., 1987). BROECKER et al. (1988) have continued the high resolution examination of glacial to Holocene change using accelerator radiocarbon techniques, building on the faunal studies of Feni Drift core V23-81 presented by PISIAS et al. (1973) and RUDDIMAN et al. (1977). Most recently, JOHNSON et al. (1988) studied fluctuations in the western boundary undercurrent on the Blake Outer Ridge.

All together, there are at least 18 sediment drifts in the North Atlantic Ocean, and probably more yet to be identified (MCCAVE and TUChOLKE, 1986). In this paper we discuss stratigraphic results of cores from six sediment drifts, with special emphasis on the northeastern Bermuda Rise (Fig. 1) where a large diameter piston core (GPC-5) was
Sediment drifts in the North Atlantic

Fig. 1. Equal area projection of the North Atlantic region showing generalized deep circulation, core locations as solid dots, and names of sediment drifts described in this paper. AABW, Antarctic Bottom Water; NADW, North Atlantic Deep Water; NEADW, Northeast Atlantic Deep Water; NSOW, Norwegian Sea Overflow Water; BBOR, Blake–Bahama Outer Ridge. Feni Drift cores are V23-81, CH73-139C, KNR 51 GGC-11 and PG-13; Gardar Drift core is KN714-15; Bermuda Rise cores are KNR 31 GPC-5 and EN120 GGC-1; continental rise core is V26-176; and BBOR cores are KNR 31 GPC-7 and GPC-9.

available. Previous paleoceanographic and sedimentological studies of this core were reported by Silva et al. (1976), Schnitker (1979), Laine and Hollister (1981) and Keigwin et al. (1984). Use of large diameter cores is most important because on many sediment drifts high rates of sedimentation are achieved through advection of fine-grained terrigenous material which reduces the abundance of the foraminifera needed for faunal, chemical and isotopic analysis. Our results from the Bermuda Rise and other sediment drifts show that with proper attention to stratigraphy and chronology, exceptionally detailed records of sediment and ocean history can be recovered from sediment drifts. Although we recognize the danger in generalizing about the history of a large sediment body on the basis of one or a few cores, for some drifts ours is the first detailed investigation.

MATERIALS AND METHODS

Of the eight cores described here (Table 1) three are Giant Piston Cores (GPC’s) from the Blake–Bahama Outer Ridge complex (GPC-7 and 9) and the northeast Bermuda Rise (GPC-5) (Silva et al., 1976; Laine, 1978; Flood, 1978). On cruise 51 of R.V. Knorr, GPC-13 and a gravity core (GGC-11) were taken on Feni Drift. GPC-13 was rigged as a conventional piston core and was triggered by a gravity core (PG-13). No previous studies of these Knorr 51 cores have been published.
Table 1. Core locations used in this study

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (m)</th>
<th>Lat. (°N)</th>
<th>Long. (°W)</th>
<th>Sedimentary feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>KNR 31 GPC-5</td>
<td>4583</td>
<td>33 41.2</td>
<td>57 36.9</td>
<td>NE Bermuda Rise</td>
</tr>
<tr>
<td>KNR 31 GPC-7</td>
<td>4935</td>
<td>28 17.9</td>
<td>72 17.8</td>
<td>Nose of Blake Outer Ridge</td>
</tr>
<tr>
<td>KNR 31 GPC-9</td>
<td>4758</td>
<td>28 14.7</td>
<td>74 26.4</td>
<td>West flank, Bahama Outer Ridge</td>
</tr>
<tr>
<td>KNR 51 PG-13</td>
<td>2665</td>
<td>54 28.5</td>
<td>15 17.9</td>
<td>Feni Ridge</td>
</tr>
<tr>
<td>KNR 51 GPC-13</td>
<td>2665</td>
<td>54 28.5</td>
<td>15 17.9</td>
<td>Feni Ridge</td>
</tr>
<tr>
<td>KNR 51 GGC-11</td>
<td>2680</td>
<td>58 45.7</td>
<td>25 57.3</td>
<td>Gardar Drift</td>
</tr>
<tr>
<td>KN714-15</td>
<td>2598</td>
<td>58 45.7</td>
<td>25 57.3</td>
<td>Gardar Drift</td>
</tr>
<tr>
<td>V26-176</td>
<td>3942</td>
<td>36 02.8</td>
<td>72 23</td>
<td>N. American continental rise</td>
</tr>
</tbody>
</table>

Following reports that Holocene sediment accumulated on the northeast Bermuda Rise at a rate of 20 cm/1000 years (Silva et al., 1976; Keigwin et al., 1984), the percent carbonate was determined at 4 cm spacing using the technique of Jones and KATCHEWS (1983). Using the same technique, %CaCO₃ was determined every 5 cm at KN714-15, but at remaining sites we used a more rapid and less precise modification of this technique, in which acid digestion was done at atmospheric pressure. Stable isotope analyses were done at the same levels as carbonate analyses, using procedures similar to those reported elsewhere (Keigwin, 1979).

Typical samples from core GPC-5 were 20 g (dry), but samples as large as 50–100 g were sometimes required from GPC-7 and GPC-9 depending on carbonate dilution by silt and clay-sized terigenous particles. Samples were dried, weighed and disaggregated in distilled water on a shaker table for 20 min. They were then briefly sonicated and washed over a 63 μm screen. The coarse fraction was shaken again in distilled water, and this process was repeated three times. Next, the coarse fraction was dried, weighed and stored in a glass vial. A split of a sample was put aside for micropaleontological studies, and specimens were picked from the remainder for stable isotope analysis and radiocarbon analysis.

Where possible, abundance peaks of monospecific benthic and planktonic foraminifera were chosen for accelerator mass spectrometer (AMS) radiocarbon dating using the tandem accelerator at the National Accelerator Facility at the University of Arizona (Jones et al., 1989). For comparison of our marine records with climatic events occurring on land, we subtract 400 years from our time series data to account for the radiocarbon age difference between the atmosphere and the surface ocean (Stuiver et al., 1986). The appropriateness of this correction will be discussed below.

The size fraction <63 μm was further size-sorted by Stokes Law settling to isolate the fraction <2 μm for clay-mineral analysis. The sediment was prepared according to the method outlined in Jones (1983), which is a slight modification of the methods described by Heath and Pisias (1979). The analyses were performed on a Phillips 3500 XRD using a Cu ka X-ray tube operated at 40 kV and 20 ma. All data were collected digitally with an IBM AT computer.

RESULTS AND DISCUSSION

Core stratigraphy

Bermuda Rise. Previous radiocarbon dating of Bermuda Rise sediments yielded anomalous results, which were traced to detrital contamination in the silt and clay size
fraction (Keigwin et al., 1984). The abundance of two important species of planktonic foraminifera reveals several peaks in the upper 250 cm of GPC-5 which we have AMS-dated, but deeper in the core foraminiferal abundance decreases and mixed planktonic species have been AMS-dated (Fig. 2). Dating of monospecific foraminiferal samples from peaks in their abundance (Table 2) avoids the detrital contamination problem and minimizes age shifts introduced by the effects of bioturbation. The oldest reservoir-corrected date on a monospecific sample of planktonic foraminifera (10,750 ± 310 years; Table 2) comes from *N. pachyderma* (dex.), the coolest faunal element found at this middle latitude location. A foraminiferal assemblage with a significantly cooler aspect, and a reservoir-corrected date insignificantly different from the brief return to glacial conditions in northern Europe, clearly identify this as the time of the Younger Dryas cooling (10,600 years; Mangerud et al., 1984). Our age model for the Holocene section

Fig. 2. Number per gram of the planktonic foraminifera *Globigerinoides ruber* (solid circles) and *Neogloboquadrina pachyderma*, dextral (open circles) from core GPC-5, northeast Bermuda Rise. Radiocarbon-dated levels are indicated by the following notation: MP, mixed planktonic species; GR, *G. ruber*; GS, *Globigerinoides sacculifera*; PAR; *N. pachyderma* (dex.). Where possible, peaks in abundance of a species were radiocarbon dated by accelerator mass spectrometry (AMS).
Table 2. Radiocarbon results of this study

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth (cm)</th>
<th>Species</th>
<th>Radiocarbon age*</th>
<th>Accession no.</th>
</tr>
</thead>
<tbody>
<tr>
<td>GPC-5</td>
<td>0-2</td>
<td>G. ruber white</td>
<td>4540 ± 10</td>
<td>AA-2482</td>
</tr>
<tr>
<td>GPC-5</td>
<td>29-31</td>
<td>Mixed plank.</td>
<td>1890 ± 140</td>
<td></td>
</tr>
<tr>
<td>GPC-5</td>
<td>64-66</td>
<td>G. ruber white</td>
<td>2940 ± 65</td>
<td>AA-1806</td>
</tr>
<tr>
<td>GPC-5</td>
<td>102-104</td>
<td>G. sacculifera</td>
<td>1260 ± 55</td>
<td>AA-2483</td>
</tr>
<tr>
<td>GPC-5</td>
<td>100-104</td>
<td>G. sacculifera</td>
<td>4710 ± 140</td>
<td>WHG-211</td>
</tr>
<tr>
<td>GPC-5</td>
<td>154-156</td>
<td>G. ruber</td>
<td>7770 ± 90</td>
<td>AA-1808</td>
</tr>
<tr>
<td>GPC-5</td>
<td>174-176</td>
<td>G. ruber</td>
<td>9210 ± 150</td>
<td>AA-1806</td>
</tr>
<tr>
<td>GPC-5</td>
<td>200-204</td>
<td>N. pachyderma (dex.)</td>
<td>11,150 ± 310</td>
<td>AA-1804</td>
</tr>
<tr>
<td>GPC-5</td>
<td>282-284</td>
<td>Mixed plank.</td>
<td>12,800 ± 130</td>
<td>WHG-176</td>
</tr>
<tr>
<td>GPC-5</td>
<td>388-390</td>
<td>Mixed plank.</td>
<td>14,800 ± 200</td>
<td>WHG-185</td>
</tr>
<tr>
<td>GPC-5</td>
<td>540-542</td>
<td>Mixed plank.</td>
<td>15,500 ± 140</td>
<td>WHG-177</td>
</tr>
<tr>
<td>GPC-5</td>
<td>650-652</td>
<td>Mixed plank.</td>
<td>16,200 ± 390</td>
<td>WHG-178</td>
</tr>
<tr>
<td>GPC-5</td>
<td>730-732</td>
<td>Mixed plank.</td>
<td>17,000 ± 140</td>
<td>WHG-179</td>
</tr>
<tr>
<td>KN714-15</td>
<td>0-2</td>
<td>Mixed plank.</td>
<td>2320 ± 65</td>
<td>AA-1813</td>
</tr>
<tr>
<td>KN714-15</td>
<td>4-6</td>
<td>Mixed plank.</td>
<td>2560 ± 55</td>
<td>AA-1812</td>
</tr>
<tr>
<td>KN714-15</td>
<td>44-46</td>
<td>Mixed plank.</td>
<td>4130 ± 85</td>
<td>AA-1811</td>
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<tr>
<td>KN714-15</td>
<td>98-100</td>
<td>Mixed plank.</td>
<td>6250 ± 110</td>
<td>AA-1810</td>
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<tr>
<td>KN714-15</td>
<td>162-164</td>
<td>Mixed plank.</td>
<td>8070 ± 120</td>
<td>WHG-204</td>
</tr>
<tr>
<td>KN714-15</td>
<td>224-226</td>
<td>Mixed plank.</td>
<td>9280 ± 140</td>
<td>AA-1990</td>
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<tr>
<td>KN714-15</td>
<td>297-300</td>
<td>Mixed plank.</td>
<td>8690 ± 120</td>
<td>WHG-205</td>
</tr>
<tr>
<td>KN714-15</td>
<td>364-366</td>
<td>Mixed plank.</td>
<td>10,100 ± 90</td>
<td>AA-2189</td>
</tr>
<tr>
<td>KN714-15</td>
<td>392-394</td>
<td>N. pachyderma (sin.)</td>
<td>12,100 ± 110</td>
<td>WHG-206</td>
</tr>
<tr>
<td>KN714-15</td>
<td>424-426</td>
<td>N. pachyderma (sin.)</td>
<td>17,900 ± 180</td>
<td>AA-2190</td>
</tr>
<tr>
<td>KN714-15</td>
<td>601-603</td>
<td>Mixed plank.</td>
<td>30,400 ± 640</td>
<td>AA-2481</td>
</tr>
<tr>
<td>GPC-9</td>
<td>1-6</td>
<td>Mixed plank.</td>
<td>2260 ± 100</td>
<td>WHG-192</td>
</tr>
<tr>
<td>GPC-9</td>
<td>77-82</td>
<td>Mixed plank.</td>
<td>10,500 ± 85</td>
<td>WHG-305</td>
</tr>
<tr>
<td>GPC-9</td>
<td>108-110</td>
<td>Mixed plank.</td>
<td>11,100 ± 120</td>
<td>WHG-193</td>
</tr>
<tr>
<td>GPC-9</td>
<td>175-177</td>
<td>Mixed plank.</td>
<td>14,000 ± 140</td>
<td>WHG-194</td>
</tr>
<tr>
<td>GPC-9</td>
<td>296-301</td>
<td>Mixed plank.</td>
<td>15,200 ± 190</td>
<td>WHG-195</td>
</tr>
<tr>
<td>GPC-9</td>
<td>395-399</td>
<td>Mixed plank.</td>
<td>19,700 ± 190</td>
<td>WHG-197</td>
</tr>
<tr>
<td>GPC-9</td>
<td>522-527</td>
<td>Mixed plank.</td>
<td>23,800 ± 310</td>
<td>WHG-212</td>
</tr>
<tr>
<td>V26-176</td>
<td>225</td>
<td>Mixed plank.</td>
<td>12,600 ± 140</td>
<td>WHG-214</td>
</tr>
<tr>
<td>V26-176</td>
<td>455</td>
<td>Mixed plank.</td>
<td>42,800 ± 1300</td>
<td>WHG-213</td>
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<tr>
<td>13 PG</td>
<td>2</td>
<td>Mixed plank.</td>
<td>2400 ± 75</td>
<td>WHG-208</td>
</tr>
<tr>
<td>13 PG</td>
<td>128</td>
<td>N. pachyderma (sin.)</td>
<td>11,100 ± 100</td>
<td>WHG-218</td>
</tr>
<tr>
<td>GPC-13</td>
<td>9</td>
<td>Mixed plank.</td>
<td>28,500 ± 450</td>
<td>WHG-209</td>
</tr>
</tbody>
</table>

* Radiocarbon ages before reservoir correction. Ages mentioned in the text and used in the figures have been reservoir-corrected by subtracting 400 years.
† These samples were probably switched, which we verified with a replicate analysis (WHG-211) at 100-104 cm.
‡ Keigwin et al. (1984).

of the Bermuda Rise is based on interpolation between the Younger Dryas aged event and reservoir-corrected dates at other levels in GPC-5 (Fig. 2). Likewise, for the Late Pleistocene section where increased flux of clay- and silt-sized terrigenous grains dilutes carbonate (discussed below), the age model is based on interpolation between dates on mixed assemblages of planktonic foraminifera to about 17,000 years BP and extrapolation to about 19,000 years BP (Fig. 3).

Detailed calcium carbonate percentage records from the Bermuda Rise added to the preliminary results of Silva et al. (1976) provide a means of detailed correlation between cores. The striking correlation of our results from core GPC-5 with the results of Boyle and Keigwin (1987) from a second Bermuda Rise core (EN120, GGC-1) is best illustrated when both data sets are plotted vs time using the age model described above.
Sediment drifts in the North Atlantic

Fig. 3. Age–depth plot of cores from four sediment drifts in the North Atlantic. All dates are based on AMS radiocarbon analyses of hand-picked mixed planktonic or monospecific planktonic foraminifera, excepting the 29–31 cm interval at GPC-5, which was a beta-decay analysis (Keigwin et al., 1984). Results from Feni Drift cores V23-81 and CH73-139C are from Broecker et al. (1988) and Duplessy et al. (1986), respectively. It is evident that Gardar Drift core KN714-15 has lowest linear rates of sedimentation during glacial times and that Bermuda Rise (GPC-5) and Bahama Outer Ridge (GPC-9) cores have low rates during the Holocene time. Feni Drift cores (V23-81 and CH73-139C), however, have relatively constant linear rates of sedimentation. These patterns reflect the supply of sediment and the availability of currents to advect it.

(Fig. 4). In general, minimum carbonate values are characteristic of the glacial interval, with maximum values in the Holocene, typical of an “Atlantic-type” carbonate stratigraphy (Gardner, 1975). Details in the Bermuda Rise record of the Holocene (20 cm/1000 years; Fig. 3) may reflect climatic influence on sedimentation which is not normally resolved in North Atlantic cores with typical pelagic sedimentation rates of approximately 2 cm/1000 years. From the Bermuda Rise record, the Holocene is seen to be marked by three minima in percent carbonate, which suggest variation in the production, dissolution, and/or dilution of carbonate.

As with %CaCO$_3$, additional sampling for stable isotopes has resolved additional detail in the record of Holocene climate change not resolved by Keigwin et al. (1984). The base of the Holocene is marked by a rapid decrease in $\delta^{18}O$ values of nearly 1.5‰ from the maximum at the Younger Dryas cooling (10,500 years BP) to minimum values by about 9000 years BP, reflecting continued melting of the Laurentide ice sheet and warming of near-surface waters (Fig. 5). This is the step in deglaciation known as Termination 1B (Duplessy et al., 1981). A small maximum in $\delta^{18}O$ (~0.3‰) is evident at about 1500 years BP, where we have sampled and replicated the record intensively. Overall, the late Holocene pattern of slightly increasing $\delta^{18}O$ with decreasing age for G. ruber resembles the pattern displayed by slightly decreasing percent carbonate (compare Figs 4 and 5).
Gardar Drift. As expected, the %CaCO₃ at core KN714-15 increased and δ¹⁸O (on G. bulloides, 180–300 μm) decreased from glacial to interglacial time (Fig. 6). Maximum oxygen isotope ratios on G. bulloides were achieved about 17,000 years ago, and although the δ¹⁸O was unusually large (4.20%), that analysis replicated well. However, the maximum on G. bulloides reported by Duplessy et al. (1981) for their Feni Drift location is close to 3.5%, in agreement with our next largest maximum. Very little detail is evident in the stratigraphic record older than about 10,000 years BP, most likely because of lowered rates of sedimentation (about 5 cm/1000 years). In particular, there is no good evidence for the Younger Dryas cooling. Ruddiman and McIntyre (1981) proposed that surface waters north of 50°N never warmed enough prior to the Younger Dryas readvance of the polar front to show a reduction in polar fauna dominance. Thus, the Younger Dryas cooling is thought to be manifested as a pause in sea-surface warming at higher latitudes.

To verify this assertion, we examined samples below 350 cm at KN714-15 for both evidence of Ash Zone 1 and the Younger Dryas cooling (Fig. 7). Ash Zone 1 is an
Fig. 5. Oxygen isotope results on the surface-dwelling planktonic foraminifera *G. ruber* from the Bermuda Rise compared to the earlier results on *Globorotalia inflata* (Keigwin et al., 1984) using the age model described in the text. Prominent events of $^{18}$O enrichment at about 10,700 and 14,500 years BP and events of $^{18}$O depletion of about 12,000 and 13,500 years BP mark the isotopic termination. The event at about 10,700 years BP corresponds to the Younger Dryas cooling, previously thought to be restricted to higher latitudes in the North Atlantic region. Delta $^{18}$O minima correspond in age exactly with meltwater discharge events known from the Gulf of Mexico, and probably reflect lowered surface ocean salinity in the northern Sargasso Sea. Symbols for AMS dated levels as in Fig. 2.

interval of dispersed, clear, rhyolitic shards, which has been dated on land to 10,600 years (Mangerud et al., 1984). In this core Ash Zone 1 is centered at about 380 cm (Fig. 7) but shards are in low abundance, which explains why this zone was not identified in previous studies (Ruddiman and McIntyre, 1981). By interpolation between radiocarbon dated levels at 365 and 393 cm (Table 2), we calculate the reservoir-corrected age of Ash Zone 1 in KN714-15 to be 10,600 years BP, exactly the same as the terrestrial age. The percent abundance of *N. pachyderma* (sin.) does not change across this interval, in agreement with the model of Ruddiman and McIntyre (1981).

Feni Drift. Of our three KNR 51 cores from the Feni Drift, only core GGC-11 contains both the glacial and interglacial intervals (Fig. 8A). Unfortunately, the only core
Fig. 6. Oxygen isotope and percent calcium carbonate results from core KN714-15, from the Gardar Drift. In general, this core displays the typical Atlantic Ocean pattern of anticorrelation between $\delta^{18}O$ and $\%CaCO_3$. During deglaciation at about 10,000 years BP, however, the two parameters become uncoupled, due to the enormous flux of terrigenous material to this location as Norwegian Sea Overflow Water production resumed. AMS dated levels are marked by the following notation: MP = mixed planktonic species, PAL = N. pachyderma (sin.)

material available was small water content samples at 10 cm spacing, so a more detailed record than that reported here cannot be produced. The location of nearby cores GPC-13 and PG-13 appears more suitable for high resolution study. However, AMS radiocarbon dating of these cores indicates that PG-13 terminates in sediment of about 11,000 years age and that GPC-13 is missing the entire interval younger than 28,000 years (Table 2). From GGC-11 we find that the maximum $\delta^{18}O$ on N. pachyderma (dex.) is 3.75‰ at 82 cm, giving a glacial–interglacial amplitude of 2.75‰ for this species. This amplitude compares favorably with the 2.5‰ range of G. bulloides reported by Duplessy et al. (1981) for their Feni Drift core and indicates a temperature component of 5 or 6°C assuming an ice volume effect of 1.3‰ and insignificant salinity change. Such an isotopic estimate of paleotemperature change is probably a minimum estimate, because the high
Fig. 7. Details of the latest glacial termination at Gardar Drift core KN714-15. The dispersed rhyolitic ash shards (>150 μm) of Ash Zone 1 (Vedde Ash) are present, but in low abundance. There is no evidence, however, for the coeval Younger Dryas cooling in either the record of \textit{N. pachyderma} (sin.) abundance or δ\textsuperscript{18}O. These results support the observation of Ruddiman and McIntyre (1981) that at this location there was insufficient surface water warming prior to 11,000 years BP for the Younger Dryas event to register as an oscillation of the polar front. AMS dated levels as in Fig. 6.

latitude North Atlantic had lower salinity during the glacial maximum (Keigwin and Boyle, in press). Both summertime (Duplessy et al., 1981) and wintertime (Bard et al., 1987) paleotemperature estimates based on transfer functions from a nearby Feni Drift core indicate glacial–interglacial temperature changes of about 8°C.

Core PG-13 has an excellent record of Termination 1B, the Younger Dryas event and Ash Zone 1 (Fig. 8B). Between 10,000 and 9,000 years BP, decreasing δ\textsuperscript{18}O of \textit{N. pachyderma} (dex.) and increasing %CaCO\textsubscript{3} mark Termination 1B (Fig. 8B). Close sampling near the beginning of this trend allowed us to identify the peak of Ash Zone 1, and the peak in abundance of \textit{N. pachyderma} (sin.), which are indistinguishable stratigraphically at this site as well as at nearby V23-81 (Ruddiman and McIntyre, 1981). An AMS date on \textit{N. pachyderma} (sin.) from the Younger Dryas peak in abundance of
this species gives a reservoir corrected age of 10,700 years BP, virtually identical in age to the Younger Dryas event on the Bermuda Rise and the Ash Zone 1 date on the Gardar Drift.

**U.S. east coast continental rise.** The chronology and stratigraphy of core V26-176 from the lower continental rise in the western North Atlantic have been discussed previously by BALSAM and HEUSSER (1976). Their age control came from radiocarbon determination on bulk sediment at 50 and 160 cm, giving reservoir corrected ages of 3430 ± 140 years and 7780 ± 200 years, respectively. Although these radiocarbon analyses were on bulk sediment, any detrital effect on the dates should be relatively minor because the samples were from intervals of maximum carbonate composed primarily of planktonic foraminifera. Additional age control in the BALSAM and HEUSSER (1976) paper came from second order correlation with dated levels in another core and on land, but these are inconsistent with our AMS radiocarbon date of 12,200 years BP at 225 cm (Table 2).

Our oxygen isotope results on the planktonic foraminifera *G. ruber* (150–250 μm) from V26-176 show a typical (but “noisy”) pattern of glacial to interglacial change, with values enriched in 18O below about 200 cm (Fig. 9). Based on δ18O, the glacial maximum

![Diagram](image-url)
Fig. 8. (A) Oxygen isotope results from core GGC-11 from the Feni Drift. These results indicate that the sedimentation rate at this location is only about 5 cm/1000 years, despite the proximity to other locations with much higher rates. (B) Stratigraphic results at Feni Drift core PG-13. At this location the linear rate of sedimentation is 10 cm/1000 years, and Ash Zone 1 and the Younger Dryas cooling are well-represented. AMS dated levels are marked by symbols as in Fig. 6.

is likely to be between 250 and 350 cm, but our next deepest radiocarbon date (42,400 years BP at 455 cm) is not close enough to date accurately the age of the $\delta^{18}$O maximum by interpolation. Nevertheless, our results do not support the chronology of BALSAM and HEUSSER (1976), which assumed that the 18 ky level is below about 400 cm (equivalent to 35,000 years BP interpolating between our AMS dates). The interesting pattern of $\delta^{18}$O variability suggests that some of the same events we find on the Bermuda Rise may be recorded on the continental margin as well, but it has been difficult to get enough large, closely spaced samples for analysis from this conventional-diameter piston core. Unlike other cores used in this study, at V26-176 we found two samples with well-sorted quartz sand (Fig. 9). These suggest that even the best cores on the continental margin are likely to be affected by downslope transport.

Blake–Bahama Outer Ridge (BBOR). FLOOD (1978) previously established a preliminary Late Quaternary stratigraphy of many cores from the BBOR, based on lithology,
faunal ranges and radiocarbon dating of bulk sediment. We have concentrated our efforts on the upper few meters of GPC's 7 and 9, adding detail to Flood's earlier work.

Of many available BBOR cores, we concentrated on GPC's 7 and 9 for the following reasons: (1) preliminary stratigraphy was available, and each core recovered Holocene sediment, (2) each appeared to have had high rates of sedimentation during the latest glacial and subsequent termination, (3) there are no paleoceanographic records from such great depths (4700–4900 m) in the western North Atlantic, and (4) large volumes of sediment were collected by the GPC's.

At the deepest core, GPC-7 (4935 m), from the east flank of the Blake Outer Ridge, % carbonate is too low to generate a detailed stratigraphy, and surface-dwelling planktonic and benthic foraminifera are too rare for $\delta^{18}O$ analysis in the Pleistocene.

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Fig. 9. Oxygen isotope results on the pink (solid circles) and white (open circles) phenotypes of the surface-dwelling planktonic foraminifera G. ruder at core V26-176 from the lower continental rise, U.S. east coast. Although there is some evidence of "events" in the record of deglaciation (i.e. low $\delta^{18}O$ at about 250, 280 and 310 cm) we did not get large enough samples from this conventional piston core to generate closely spaced data. Two layers of fine, well-sorted quartz sand are marked by "XXX." Radiocarbon dated levels on bulk sediment (BALSAM and HEUSSE, 1976) are marked by "B," and AMS dated levels on mixed planktonic foraminifera are indicated by "MP".
section. Although only about 200 m shallower than GPC-7, GPC-9 (4758 m on the Bahama Outer Ridge) has significantly higher carbonate percentages, and a detailed oxygen isotope stratigraphy was obtainable in the upper 270 cm of this core (Fig. 10). Seven AMS radiocarbon dates based on mixed species of planktonic foraminifera provide age control (Table 2). As in our Bermuda Rise core (Fig. 5), $\delta^{18}O$ on white *G. ruber* shows a glacial to interglacial range of about 2%, and significant events are found in the record of deglaciation. Those isotopic events with the best age control include maxima at 14,500, 13,000 and 10,500 years BP, and minima at 13,500–14,000 years BP, and at about 12,000 years BP. Percent CaCO$_3$ gradually increases from a minimum 20,000 years ago, with a rapid increase at about 12,000–13,000 years BP, and a second minimum at 9000–10,000 years BP, prior to the late Holocene maximum.

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**Fig. 10.** Oxygen isotope and % carbonate results from core KNR 51 GPC-9, Bahama Outer Ridge. Distinct $\delta^{18}O$ minima occur at 12,000 years BP and between 13,500 and 14,000 years BP, the same as on the Bermuda Rise (Fig. 5) and in the Gulf of Mexico (Broecker et al., 1988). These events are inferred to reflect a meltwater influence from the Mississippi River. Mixed planktonic AMS dates indicated by "MP."
Regional patterns of sedimentation

Cores of each sediment drift we have examined display the familiar Atlantic Ocean pattern of decreased carbonate percentages during times of oceanic enrichment in $^{18}O$. As has been observed previously, this is due in large part to dilution of carbonate by fine-grained terrigenous sediment during glacial intervals (GARDNER, 1975). Our detailed stratigraphy and chronology, however, reveal that rates of sedimentation were regionally variable (Fig. 3). The end members of this variability are illustrated by patterns of sedimentation on the Bermuda Rise and the Gardar Drift (Fig. 11). When the linear sedimentation rate for dated intervals at these locations is normalized to the average rate for the last 18,000 years, variability within and between these locations is emphasized. Although some of the variability in linear rate of sedimentation can be ascribed to changing physical properties on the Bermuda Rise (SILVA et al., 1976), the very large range in rates of sedimentation on both the Bermuda Rise and the Gardar Drift must reflect variability in the supply of sediment as well as variability in the current controlling its deposition.

On the Gardar Drift sedimentation rates were lower during the late glacial than the Holocene (Figs 3 and 11). Our results from KN714-15 support the observation of RUDDIMAN and BOWLES (1976) that there were enormous changes in sediment deposition during the transition from glacial to interglacial time on the Gardar Drift. At KN714-15 these rates range from a maximum of nearly 90 cm/1000 years about 9 ky ago to a minimum of about 5 cm/1000 years earlier in the deglaciation. In the late Holocene, sedimentation is relatively constant at close to 20 cm/1000 years. The four-fold increase in sedimentation (relative to the average for the last 18,000 years) in the late deglacial and early interglacial on the east flank of the Reykjanes Ridge was due to both the ample supply of sediment during the glacial interval and the reinitiation of Norwegian Sea

![Fig. 11. Linear rates of sedimentation on the Gardar Drift (from core KN714-15) and the Bermuda Rise (from core KNR 31 GPC-5) normalized to the average for the past 18,000 years. Pulses of increased sedimentation occurred on the Gardar Drift in the earliest Holocene in contrast to the Bermuda Rise where increased sedimentation occurred during the glacial–early deglacial interval. Peak deposition on the Bermuda Rise most likely occurred in response to erosion in eastern Canada during glaciation. Although Iceland was probably a source of sediment during glaciation, peak deposition on the Gardar Drift was delayed until the earliest Holocene, when Norwegian Sea Overflow Water production resumed.](image)
Overflow Water (NSOW) following retreat of the polar front (Ruddiman and Bowles, 1977).

At the other extreme, drift deposits from the Bermuda Rise had much higher sedimentation rates in the glacial than the interglacial (Figs 3 and 11). The linear rate of sedimentation was lowest for much of the early Holocene at about 14 cm/1000 years, and peaked near 15,000 years BP at a rate of 200 cm/1000 years. Although not constrained by as many AMS dates, the pattern of sedimentation on the BBOR is similar to that found on the Bermuda Rise, with higher rates during the glacial including a peak of nearly 100 cm/1000 years between 14,000 and 15,000 years BP (Fig. 3).

The general pattern of higher linear rates of sedimentation during glacial times displayed on both the Bermuda Rise and the BBOR must reflect the greater supply of sediment during glacial times in the western North Atlantic basin and the continual availability of deep currents to advect it. Sedimentation on the BBOR is dominated by the western boundary undercurrent (Flood, 1978), and the return flow of the Gulf Stream (Worthington, 1976) is thought to be responsible for focusing sediment on the Bermuda Rise (Laine and Hollister, 1981). Despite glacial–interglacial changes in the chemistry of deep water in the North Atlantic (e.g. Boyle and Keigwin, 1985), each of these current systems probably existed during glacial time. Thus, the supply of sediment to the deep western North Atlantic from both erosion on the continents and from shelves exposed by lowered sea level probably controlled the patterns of linear rates of sedimentation we observe during glacial time (Fig. 11).

By about 10,000 years BP, however, assuming deglaciation was half over, sea level must have risen enough to flood the continental shelves and trap much of the terrigenous sediment. Therefore, Holocene patterns of sedimentation in the deep sea are probably controlled by redistribution of sediment within basins (Milliman, 1976). Suman and Bacon (1989) have defined a “focusing factor” which indicates that the increasing linear rate of sedimentation which began about 8000 years BP was due to increased current activity in the deep western North Atlantic. From this sedimentological evidence on the Bermuda Rise it appears that the deep Atlantic has already begun to return to glacial levels of circulation intensity.

Between the extremes of maximum (Bermuda Rise) and minimum (Gardar Drift) linear sedimentation rates during glacial time, Feni Drift cores have relatively constant rates (Fig. 3). Constant rates, at least during the deglacial interval, may have resulted from compensation of decreased carbonate production by increased terrigenous flux at locations with more pelagic influence on sedimentation (Ruddiman and McIntyre, 1981). Terrigenous material on Feni Drift is derived primarily from the Irish continental margin to the east (Lonsdale and Hollister, 1979; Flood, 1978).

Lonsdale and Hollister (1979) described the interplay between sediment supply and mid-depth circulation in the region of Feni Drift. These authors found evidence for two currents that are likely to play a role in sedimentation on Feni Drift: southerly flowing NSOW, on the western side of Rockall Trough, and northeasterly flowing Northeast Atlantic Deep Water (NEADW) which flows along the Irish continental margin. Lonsdale and Hollister (1979) suggested that this latter current forms a cyclonic loop and transports fine-grained terrigenous particles from the continental margin to Feni Drift. Their hydrographical data show the water of southern origin (AABW) contributes significantly to NEADW, with greatest influence below about 2500 m. The general circulation scheme described by Lonsdale and Hollister (1979) is supported by recent
long-term current mooring data (Dickson and Kidd, 1986). Both Lonsdale and Hollister (1979) and Dickson and Kidd (1986) conclude that NEADW is more important than NSOW in shaping Feni Drift. Thus, for Feni Drift glacial and interglacial rates of linear sedimentation are roughly comparable because water of southern origin was still active in sediment transport during glacial time even as NSOW production was diminished.

**Preliminary paleoceanographic observations**

Our detailed stratigraphic data lead to some preliminary observations on short-duration events in the Late Quaternary history of climate and ocean chemical change in the North Atlantic region. Although these observations will be developed more thoroughly elsewhere, on the basis of the present data set we find evidence for: (1) significant climate variability within the Holocene; (2) occurrence of the Younger Dryas cooling farther to the south than previously reported; and (3) occurrence of events of lowered surface water salinity in the central North Atlantic during deglaciation.

As discussed above, the large glacial to interglacial changes in rates of sedimentation and %CaCO₃ reflect the accumulation of clay- and silt-sized terrigenous material. In general the underlying process (glaciation and continental erosion) results in anti-correlation of %CaCO₃ and δ¹⁸O on the glacial-interglacial timescale, but the same relationship also holds for shorter-duration events. For example, where there is sufficient CaCO₃ and oxygen isotopic evidence for the Younger Dryas cooling as on the Feni Drift and the Bahamas Outer Ridge, the two climate proxies remain closely linked (Figs 8B and 10, respectively). The same also seems to occur on the Bermuda Rise during the Holocene, where we observe small but significant % carbonate variability (Fig. 4). The last 2500 years of the Bermuda Rise record has been sampled very intensely for δ¹⁸O, and it is evident that a small maximum in δ¹⁸O of ~0.3‰ occurred at 1500 years BP (Fig. 5), coincident with a carbonate minimum (Fig. 4). This δ¹⁸O maximum is not likely to reflect seawater compositional change due to continental ice growth because there is no evidence for sea level changes of a few tens of meters in the late Holocene. It most likely represents a small cooling (~1°C) or salinity increase (~0.5‰) at the sea surface.

The results of Suman and Bacon (1989) contribute to understanding brief events in the carbonate stratigraphy of the Bermuda Rise. From a downcore profile of excess ²³⁰Th concentration and the assumption of a constant flux of ²³⁰Th to the seafloor, Suman and Bacon (1989) concluded that the supply of carbonate remained nearly constant throughout the Holocene (Fig. 12A). The variations in %CaCO₃ observed in the Holocene of GPC-5 are due mainly to variations in the primary supply of low carbonate terrigenous sediment. It is reasonable to assume that the flooded continental shelves acted as effective sediment traps during the last several millenia. Thus the terrigenous sediment which was deposited about 1500 and 3500 years BP on the Bermuda Rise must have been redistributed from within the North Atlantic basin.

Constraints on the provenance of this dilutant can be placed by the clay mineralogy of the <2 μm fraction from the Bermuda Rise. These data indicate that during Holocene times of low carbonate percentages the percentage of chlorite increased (Fig. 12B). Chlorite is produced largely as a result of mechanical weathering, and Biscaye (1965) showed that in the North Atlantic maximum input of this clay mineral appears to originate in eastern Canada. Therefore, the proposed dilutant must have relatively high chlorite content, a lower %CaCO₃ than Holocene sediments of the Bermuda Rise and relatively little excess ²³⁰Th. Milliman (personal communication) suggested continental
Sediment drifts in the North Atlantic

Fig. 12. Comparison of %CaCO₃, % clay, carbonate flux and clay flux at Bermuda Rise core GPC-5. (A) Fluxes of clay and carbonate (after SUMAN and BACON, 1989) indicate that carbonate percentage on the Bermuda Rise is dominated by clay flux. (B) Increased chlorite percentages (talc-normalized) during times of increased clay flux (lower % CaCO₃) suggests a northern source for this increased fine-grained sediment dilutant on the Bermuda Rise. Advection of continental slope sediment off Nova Scotia may be responsible for these late Holocene clay flux variations.

Although %CaCO₃ is too low to recognize significant late glacial and deglacial variability on the Bermuda Rise, events that have occurred every 2000 years or so are
evident in our $\delta^{18}$O record. Most widely occurring among these events is the enrichment in $^{18}$O associated with the Younger Dryas cooling, which we record unambiguously on the Bermuda Rise (Fig. 5), and probably as far to the south as the BBOR at 28°N (Fig. 10).

The Younger Dryas cooling is closely associated with the Vedde Ash at many locations in Scandinavia, and both are consistently dated at 10,600 years BP (MANGERUD et al., 1984). It is important to establish the age of the Vedde Ash marine records because it will allow a direct comparison of dates from the marine and terrestrial carbon reservoirs (BARD, 1988). Surface marine waters in the modern North Atlantic are about 400 years older than modern terrestrial carbon, and about the same difference is maintained for most of the Holocene (STUIVER et al., 1986). This difference results from the upwelling and diffusion of older waters from the intermediate ocean to the surface ocean. If this upward flux of older water was different in the glacial ocean, then a reservoir correction of 400 years might not be appropriate.

Since the distribution of *N. pachyderma* (sin.) and ash shard abundance are nearly identical on the Feni Drift (Fig. 8, BROECKER et al., 1988), the ash peak can be directly dated. Our direct dating of this event on Feni Drift (KNR 51-13PG; 10,700 ± 100 years BP, reservoir-corrected; Fig. 8B) agrees very well with the age of the Vedde Ash on land (10,600 years). On the Gardar Drift, however, where there is no distinct peak in *N. pachyderma* (sin.) abundance during Younger Dryas time (Fig. 7), it may be more appropriate to date the weighted mean of the ash distribution (at ~380 cm) than the peak in ash abundance (at ~385 cm). In any case, at the present time we are dating the ash on Gardar Drift by interpolation, giving reservoir-corrected ages of ~10,600 ± 100 years BP (weighted mean) or 10,900 ± 100 years BP (peak in ash abundance). Since the age of the Vedde Ash and the cooling on land agree so well with the marine age in our directly dated sequence when corrected for the modern reservoir effect, it seems likely that, at least at Younger Dryas time, the 400 year adjustment is appropriate.

Although the Younger Dryas is the best-known event of the deglaciation, our Bermuda Rise results show that other events were equally prominent (Fig. 5). In particular, events of low $\delta^{18}$O which are centered on 12,000 and 13,500 years BP demand an explanation. Earlier, before an accurate chronology and detailed sampling were available, KEIGWIN et al. (1984) speculated that the older of these events reflected a glacial meltwater pulse in the northern Sargasso Sea. They reasoned that it was unlikely that the surface of the North Atlantic could have warmed substantially (about 4°C) early in the deglaciation, so lowered salinity was offered as an explanation. New evidence indicates that these events are widespread in the North Atlantic. As on the Bermuda Rise, the BBOR record has events of low $\delta^{18}$O, with similar ages (about 12,000 and 13,500 years BP). Finally, in the Gulf of Mexico events of low $\delta^{18}$O, which have been associated with Laurentide ice sheet meltwater (LEVENTER et al., 1982), are also found to occur close to 12,000 and 13,500 years BP, when reservoir-corrected (BROECKER et al., 1988). Thus, we speculate that events of meltwater discharge from the Mississippi reached as far as the central North Atlantic, although we cannot rule out salinity changes caused by changes in the balance between evaporation and precipitation over the North Atlantic.

**CONCLUSIONS**

The expanded sedimentary sections in deep-sea sediment drifts offer great potential for resolving the finer features of paleo-ocean and paleoclimatic change. Although
Sediment drifts have high rates of sedimentation due to reworking of sediment, our study of several representative drifts in the North Atlantic region indicates that sediment transport and reworking primarily affect silt and clay-sized particles. Thus, sand-sized particles, which in our cores are mostly planktonic foraminifera, accumulate particle-by-particle from overlying surface waters and dilution by finer terrigenous particles prevents loss of their fine-scale paleoceanographic information due to bioturbation. Where the source of the finer, advected component of drift sediment is known, variation of its accumulation has the potential to reveal the direction of deep-sea currents, and variability in the source of the sediment. From our detailed stratigraphic and chronologic studies, we find that:

(1) Sediment drifts in the North Atlantic can be characterized on the basis of variation in their rates of sedimentation. At one extreme, the Gardar Drift has lower rates of sedimentation during glacial than interglacials. This pattern probably results from decreased production of NSOW during glacial time (Ruddiman and Bowles, 1976). Core KN714-15 is a good example of this group, from which our AMS radiocarbon dating indicates that rates of sedimentation increased by a factor of 20 from the latest glacial to the Holocene. Maximum rates of sedimentation approached 100 cm/1000 years between 8 and 10 ky ago. The same trends are found in two other Gardar Drift cores (Ruddiman and Bowles, 1976), so this pattern is likely to be regional. Other northern drift locations (the Feni Drift) show relatively little change in linear sedimentation rate from glacial to interglacial time. The reason for this may be that, although NSOW production was dramatically decreased, water of southern origin (NEADW) continued to advect fine-grained material to Feni Drift.

Lower latitude sediment drifts in the western North Atlantic define the other extreme, with higher rates of sedimentation during glacial time. Increased supply of terrigenous material eroded from the continent during glaciation and sea-level lowering probably resulted in the higher rates. Linear rates of sedimentation have increased during the past 8000 years, when sea level was high and continental shelves were trapping sediment; increased advection by deep-sea currents is postulated (Suman and Bacon, 1989). In the western North Atlantic the basic patterns of deep-sea circulation were probably similar in both glacial and interglacial time, although the mixture of water masses undoubtedly changed with time. Thus, there was probably always a return flow of the Gulf Stream helping to build the Bermuda Rise, and a deep western boundary undercurrent carrying sediment to drifts along the continental margin.

(2) The record of changing %CaCO₃ on the Bermuda Rise is dominated by dilution by clay- and silt-sized terrigenous particles. Carbonate fluxes remained constant during the last 12,000 years (Suman and Bacon, 1989), and episodes of increased terrigenous flux, possibly advected from the continental slope off Nova Scotia, account for lowered carbonate percentages 1500 and 3500 years ago. A link between surface-ocean climate and supply of sediment to the deep ocean is suggested by the coincidence of carbonate dilution and ¹⁸O enrichment in surface-dwelling planktonic foraminifera at 1500 years BP on the Bermuda Rise.

(3) The oxygen isotopic record of deglaciation in the western North Atlantic is extremely complicated. The Younger Dryas cooling of about 10,000–11,000 years BP, which was previously identified in marine sediment cores only in the northeast Atlantic, is evident in sediment drifts as far south as 28°N. Brief events of low δ¹⁸O at about 12,000 and 13,500 years in the record of surface-dwelling planktonic foraminifera suggest
lowered salinity of surface waters overlying the Bermuda Rise and the BBOR. These events may have resulted from advection (via the Gulf Stream) of glacial meltwater which drained into the Gulf of Mexico from the Mississippi River.

(4) Variations in Holocene carbonate content (0–10,000 years BP) and deglacial δ¹⁸O (10–15,000 years BP) occur approximately every 2000 years. Unfortunately, due to the dramatic environmental changes associated with the glacial termination, we are unable to get a long enough series of oscillations in either climatic proxy to determine whether or not they are actually cyclic. Where %CaCO₃ becomes too low to show significant variability, however, δ¹⁸O variability increases in significance. Thus, the temporal variability of our marine-based climate proxies is similar enough to the terrestrial record of glacial advances and retreats described by Denton and Karlen (1973) and the glacial–Holocene oscillations in ice-core δ¹⁸O described by Dansgaard et al. (1984) to support their observation of widespread climatic forcing on non-orbital timescales. It will be important to better understand this millennial scale variability in climate in order to assess possible anthropogenic effects on climate.

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