Late glacial warming prior to Heinrich event 1: The influence of ice rafting and large ice sheets on the timing of initial warming

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ABSTRACT

High-resolution faunal, isotopic, and sedimentologic data from North Atlantic core V29-191 show that sea-surface temperatures increased from 17.5 to 17.3 ka, prior to the Heinrich event 1 (H1) ice-rafting event ca. 17–16 ka. These new data support previous studies that showed that warming predated the extensive climatic warming ca. 15.9 ka. This relationship indicates that H1 occurred after warming had begun. Loss of latent heat during iceberg melting created near-glacial sea-surface temperatures during H1 and kept much of the subpolar North Atlantic cold despite increasing Northern Hemisphere insolation. The cold-ocean conditions influenced proximal terrestrial climates and may also have affected remote regions.

INTRODUCTION

The mechanisms of climate change during the last glaciation are widely debated and the hypotheses hinge on the timing of climate change worldwide. Some data suggest that contemporary climate changes occurred in terrestrial and marine environments within the circum-North Atlantic region (e.g., Bond et al., 1992, 1993; Grimm et al., 1993) as well as globally (e.g., Denton and Hendy, 1994; Lowell et al., 1995; Kennett and Ingram, 1995). In contrast, other studies show that some Antarctic climate changes occurred prior to those in Greenland (Sowers and Bender, 1995; Blunier et al., 1998).

Asynchrony in the timing of climate change may occur if some locations are affected by local climatic conditions, which appears to be the case in some Northern Hemisphere records. Ice-core oxygen isotope records of H_2O from Greenland (e.g., Dansgaard et al., 1993) are often taken to represent Northern Hemisphere climates, primarily because a strong correlation exists between Greenland's climate data and climate data from the subpolar North Atlantic and northern and western Europe (e.g., Ruddiman and McIntyre, 1981; Oeschger et al., 1984; Bond et al., 1993). In contrast, several other Northern Hemisphere studies show that climatic warming or deglaciation had begun 1–2 k.y. or more before the large-scale climate change ca. 15.9 ka (e.g., Bard et al., 1987; Jones and Keigwin, 1988; Lister, 1988; Reille and Lowe, 1993; Giraudi and Frezzotti, 1997). An early warming trend is present but interrupted by a return to cold conditions in the Greenland isotope records (e.g., Dansgaard et al., 1993). This warming is missing or subdued in ice-proximal terrestrial records and in some often-referenced sea-surface temperature records from the northern North Atlantic (e.g., Oeschger et al., 1984; Bond et al., 1992, 1993). Consequently, this early warming is rarely invoked in global correlations, which instead let the large-scale warming ca. 15.9 ka represent the onset of warming in the Northern Hemisphere (Sowers and Bender, 1995; Blunier et al., 1998).

Local climatic influence from ice sheets and regional events, superimposed on the globally correlated long- and short-term climate cycles, complicate the timing issue. During the last glacial period, climate cycles of 6–10 k.y. culminated with cold air temperatures over Greenland, cold sea-surface temperatures in the North Atlantic, and large-scale discharges of icebergs into the North Atlantic (termed Heinrich or H events) (Heinrich, 1988; Broecker et al., 1992; Bond et al., 1992, 1993), distributed primarily in a zone of maximum ice-rafted detritus (Ruddiman, 1977). Thus, the H events appear to be connected with the climaxes of cooling cycles in the Northern Hemisphere (Bond et al., 1992, 1993). However, evidence for warming ca. 17.7 ka or before (e.g., Jones and Keigwin, 1988; Reille and Lowe, 1993; Giraudi and Frezzotti, 1997; Zahn et al., 1997) implies that the H1 and H0 events, peaking ca. 16.5 ka and ca. 12 ka, respectively (Bond et al., 1993; Bond and Lotti, 1995), do not conform to this hypothesis.

To investigate the discrepancy between various records in the circum-North Atlantic region concerning the warming ca. 17.7 ka, particularly its absence in some marine records, we examined a high-sedimentation-rate core from the subpolar North Atlantic, on the edge of the maximum ice-rafted detritus zone, with respect to faunal assemblages, oxygen isotopes, and lithic sedimentation. The goal was to obtain a detailed sea-surface temperature record and a proxy record for ice rafting from the same stratigraphic sequence at high temporal resolution.

MATERIAL AND METHODS

Core V29-191 (54°16′N, 16°47′W) was taken from the Feni drift during a *Vema* cruise organized by the Lamont-Doherty Earth Observatory of Columbia University, New York. The core interval studied had a sedimentation rate of ~10 cm/1000 yr (Lagerklint, 1995), making it ideal for high-resolution studies. Planktonic foraminifera were dated by ^{14}C using accelerator mass spectrometry (AMS). Resulting ages were corrected by 400 yr for the reservoir age of the surface ocean (Table 1) (Stuiver and Braziunas, 1993) and to allow comparison with other marine records with the same reservoir correction (e.g., Bond et al., 1992, 1993; Zahn et al., 1997). All ages mentioned in this text have been converted to calendar years by using the U/Th conversion scale (Bard et al., 1993) and the equation provided in Bard et al. (1997), and are expressed in thousands of calendar years before present (ka). Sampling intervals of 1–2 cm provided temporal resolution of 20–250 yr. The number of lithic grains (>150 μm) per gram of sediment served as a proxy for ice-rafted detritus, consistent with Bond and Lotti (1995). The relative abundance of the planktonic foraminifera *Neogloboquadrina pachyderma* (sinistral), counted from at least 300 foraminifera (>150 μm), provided a proxy for polar surface-water conditions (Bé, 1977). Oxygen isotope values ($\delta^{18}O$) based on one to three measurements of *N. pachyderma* (dextral) at each depth level reflected both sea-surface temperature and the

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isotopic composition of surface water ($\delta^{18}O_{water}$) (Epstein et al., 1953; Shackleton, 1967) in subpolar regions. The laboratory precision (1 σ) was 0.06‰ for the $\delta^{18}O$ values (and 0.04‰ for $\delta^{13}C$) of the NBS-20 standards analyzed with the samples. The $\delta^{18}O$ values are reported with respect to PDB (Peedee belemnite) and were calculated by using the value of −4.14‰ for NBS-20 from Coplen et al. (1983). All data are reported in Appendix 1.[1]

RESULTS

Two intervals with high concentrations of ice-rafted detritus occur in the part of core V29-191 studied here (Fig. 1A). The highest detritus peaks at 173 and 121 cm have ages of 16.6 ka and 12.2 ka, respectively (Table 1), identifying them as the H1 and H0 ice-rafting events observed in other North Atlantic cores (e.g., Bond et al., 1992, 1993). A peak in volcanic shards at 119 cm identified as the Vedde Ash, which has been dated as ca. 11.9 ka (Birks et al., 1996), supports the age for the H0 peak.

The glacial interval in V29-191 was identified by high percentages of *N. pachyderma* (sinistral) between 210 and 183.5 cm (20.3–17.5 ka) (Fig. 1B). An abrupt decrease in *N. pachyderma* (sinistral) percentages recorded an increase in sea-surface temperatures from 183.5 to 179.5 cm (17.5–17.3 ka). During the following H1 interval, *N. pachyderma* (sinistral) percentages increased, marking the return of colder conditions, interspersed with another warming signature. Above the last cold peak (at 169 cm, 16.1 ka), *N. pachyderma* (sinistral) percentages decreased significantly, recording the widespread deglacial warming ca. 15.9 ka. Warmer surface water conditions prevailed until sea-surface temperatures again decreased during the H0 event at 12.2 ka.

The *N. pachyderma* (dextral) $\delta^{18}O$ record shows high but variable values for the glacial interval, indicating generally cold conditions (Fig. 1C). Prior to H1, a 1.2‰ decrease in $\delta^{18}O$ values marks the first warming, coinciding with the warming in the faunal record beginning at 17.5 ka (Fig. 1B). A subsequent increase in the $\delta^{18}O$ values reflects the return to cold sea-surface temperatures during H1 (Fig. 1C). However, within the H1 layer, $\delta^{18}O$ values decreased again, indicating a combination of the second warming and meltwater (with low $\delta^{18}O_{water}$ values) from icebergs. The $\delta^{18}O$ values remained low despite the return to cold conditions recorded by the second peak in *N. pachyderma* (sinistral) percentages (Fig. 1B), illustrating the influence of low $\delta^{18}O$ meltwater on the isotope record. Low planktonic foraminiferal $\delta^{18}O$ values during H1 have previously been recorded in other records (e.g., Bond et al., 1992; Keigwin and Lehman, 1994). The *N. pachyderma* (dextral) $\delta^{18}O$ values remained low during the warm interval. Higher values initially recorded the cold conditions associated with the H0 event but were replaced by low $\delta^{18}O$ values reflecting the meltwater from icebergs.

The decreases in *N. pachyderma* (sinistral) percentage and *N. pachyderma* (dextral) isotope values, independently recorded from 17.5 to 17.3 ka in the same stratigraphic sequence, are evidence for warming prior to H1. A concern with close-interval sampling of ocean cores is the possibility that bioturbation may have affected the chronology. However, three observations argue against significant impact from bioturbation in this core interval. First, monospecific foraminifera from four depth levels across the warming and H1 interval were ^{14}C dated. The dates are in strictly chronological order (14 800 ^{14}C yr at 183.5 cm; 14 650 ^{14}C yr at 179.5 cm; 14 100 ^{14}C yr at 173 cm; 13 700 ^{14}C yr at 169 cm) and one level is from the warm interval. This indicates that the proxy records in this interval were not compromised by bioturbation. Second, a number of other North Atlantic records show a warming predating the time of H1 (e.g., Bard et al., 1987; Maslin et al., 1995; Zahn et al., 1997). This shows that warming has been detected in subpolar North Atlantic cores other than V29-191, which supports our interpretation. Third, $\delta^{18}O$ measurements of the species *N. pachyderma* (sinistral) from the same depth levels do not show similar excursions to low values (Lagerklint, 1995). Instead, *N. pachyderma* (sinistral) $\delta^{18}O$ values decrease gradually, whereas *N. pachyderma* (dextral) values decrease abruptly. Similar excursions to low $\delta^{18}O$ values in both *N. pachyderma* (sinistral) and *N. pachyderma* (dextral) would be expected if younger material with a considerably warmer signature had been incorporated at these levels.

TIMING OF WARMING IN THE CIRCUM-NORTH ATLANTIC AND THE ROLE OF ICE

Our high-resolution multivariate data from the same stratigraphic sequence in North Atlantic core V29-191 show a significant warming from 17.5 to 17.3 ka, preceding the H1 ice-rafting event and beginning at least

[1]GSA Data Repository item 9996, Appendix I, V29-191 faunal, isotopic, and lithic data, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, editing@geosociety.org, or at www.geosociety.org/pubs.drpint.htm.

TABLE 1. AMS ^{14}C DATES FOR V29-191

Depth (cm)	Corrected ^{14}C age (yr)	Error (yr)	Calendar years (yr)	Species	NOSAMS* accession number
111	9 750	±55	11 189	Mixed planktonics	OS-3536
121	10 550	±50	12 206	*N. pachyderma* (sin)	OS-3537
125	11 050	±45	12 838	*N. pachyderma* (sin) and (dex) *G. bulloides*	OS-3538
151	12 150	±50	14 218	*N. pachyderma* (sin)	OS-3539
155	9 800	±45	11 253	Mixed planktonics	OS-3540
163	12 450	±45	14 592	*G. bulloides*	OS-3541
169	13 700	±55	16 138	*N. pachyderma* (sin)	OS-3542
173	14 100	±55	16 629	*N. pachyderma* (sin)	OS-3544
179.5	14 650	±55	17 301	*N. pachyderma* (sin)	OS-3543
183.5	14 800	±90	17 484	*N. pachyderma* (sin)	OS-4143
209.5	17 150	±95	20 311	*N. pachyderma* (sin)	OS-4142

Note: Dates have been corrected with subtraction of 400 yr for the reservoir effect and then converted to calendar years (Bard et al., 1993, 1997).

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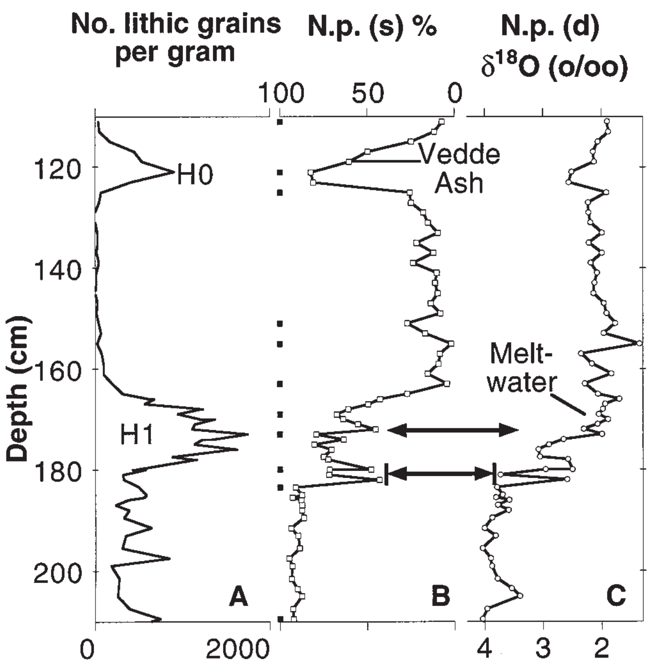

Figure 1. Records from North Atlantic core V29-191. A: Number of lithic grains per gram of sediment. B: Relative abundance of *N. pachyderma* (sinistral) in percent (squares). C: Oxygen isotopes of *N. pachyderma* (dextral) (circles). Arrows denote two brief warming events prior to and within H1 interval. Filled squares mark ^{14}C-dated levels listed in Table 1; Vedde Ash layer is marked. Lithic record illustrates that H0 and H1 ice-rafting events both occurred after warming had begun, as recorded in percentage of *N. pachyderma* (sinistral) and isotope records.

1.6 k.y. before the widespread warming ca. 15.9 ka (Fig. 2A). Numerous other studies support our interpretation. In the eastern North Atlantic between ~55° and 60°N, faunal records show that sea-surface temperatures began to warm prior to the H1 time frame (BOFS 11K, 17K, and 14K in Maslin et al., 1995; Oppo and Lehman, 1995). Off the coast of Portugal, the $\delta^{18}O$ values in core SO75-26KL decreased prior to the ice-rafted detritus maximum (Zahn et al., 1997), indicating that warming began before H1 (Fig. 2B), and faunal and $\delta^{18}O$ records from SU81-18 show warm surface conditions ca. 17.7 ka (Bard et al., 1987). Low $\delta^{18}O$ values from the Norwegian Sea, Fram Strait, and the Arctic Ocean imply increased fresh-water discharge from surrounding ice sheets prior to or between 17.7 and 17.1 ka (e.g., Jones and Keigwin, 1988; Lehman et al., 1991; Stein et al., 1994). This evidence indicates that H1 occurred after climatic warming had begun, not at the end of a cooling cycle.

Terrestrial records in the circum-North Atlantic region also recorded this early warming. Pollen, oxygen isotope, and beetle records from Spain, southern France, and the Western Alps feature a warming ca. 17.7 or 17.1 ka (Reille and Lowe, 1993; synthesis in Walker, 1995). Deglaciation had started in parts of Switzerland by ca. 17.7 ka (Lister, 1988) and glaciers had retreated in the Swiss Alps by ca. 16.5 ka (Schlüchter, 1988). The Laurentide ice sheet retreated from a maximum position between 21.3 or 24.8 and 16.5 ka in the Gulf of Maine (Belknap and Shipp, 1991). The Scandinavian ice sheet began to retreat from its maximum position at 21.3 ka (Lundqvist and Saarnisto, 1995), interstadial conditions prevailed on the Irish coast at 19.8 ka and between 18.5 and 17.4 ka (McCabe and Clark, 1998), and glaciers retreated stepwise in Italy between 20.1 and 18.9 ka and between ca. 17.7 and 16.5 ka (Giraudi and Frezzotti, 1997). The recession of the ice margins resulted in a 10–15 m sea-level rise recorded at Barbados between 21 and 17.5 ka (Fairbanks, 1989).

In contrast to the numerous warming and/or deglaciation signatures at or before 17.7 ka, North Atlantic cores from the zone of maximum ice-rafted detritus (e.g., Ruddiman and McIntyre, 1981; Bond et al., 1992, 1993; BOFS 5K and 8K in Maslin et al., 1995), for example the often-referenced V23-81 (Fig. 2C) (Bond et al., 1993; Bond and Lotti, 1995), and ice-sheet proximal records from northern and western Europe (e.g., Oeschger et al., 1984) do not show a clear warming until between 15.9 and 15.3 ka, post-dating H1. The Greenland isotope records indicate a warming trend beginning ca. 22 ka, but this reverted to cold conditions during the H1 time frame (Fig. 2D) (e.g., Dansgaard et al., 1993).

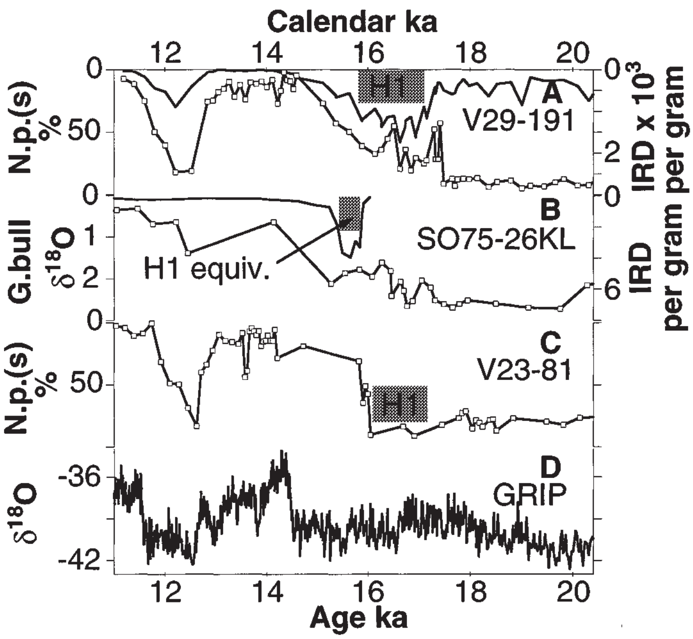

Figure 2. A: ***N. pachyderma*** **(sinistral) percentage and ice-rafted detritus (IRD) in V29-191. B:** ***G. bulloides*** **$\delta^{18}O$ record and IRD from SO75-26KL off Portugal (digitized from Zahn et al., 1997). C:** ***N. pachyderma*** **(sinistral) percentage in V23-81 (courtesy G. Bond). D: Greenland Ice Core Project (GRIP) ice core $\delta^{18}O$ record from Greenland (e.g., Dansgaard et al., 1993; data archived at World Data Center-A for Paleoclimatology, Boulder, Colorado, USA). Shaded areas show depths of H1 in marine cores and illustrate that cold ice-rafting event occurred after warming had been registered in V29-191, SO75-26KL, and GRIP core.**

The inconsistencies in timing of the warming in the North Atlantic region likely arise from a combination of unevenly distributed climatic impact from the ice sheets and a cold ice-laden ocean, and temporal resolution insufficient to detect the warming ca. 17.7 ka. We propose that loss of latent heat from the ocean during the melting of H1 event icebergs cooled the surface water in the zone of maximum ice-rafted detritus. This chilling effect created polar conditions in much of the subpolar North Atlantic and locally interrupted the warming that was under way in distant terrestrial and marine locations (similar to McCabe and Clark's [1998] suggestion), most likely in response to the increased insolation in the Northern Hemisphere. Because the largest effects of the iceberg-induced cooling were in the maximum ice-rafted detritus zone, cores in this zone fail to show a clear warming until ca. 15.9 ka, after H1 had ended (e.g., Bond and Lotti, 1995; BOFS 5K and 8K in Maslin et al., 1995). Because the climatic warming was disrupted by the iceberg-induced surface cooling, it appeared only as a short-lived event in this zone. Therefore, records with low temporal resolution in the maximum ice-rafted detritus zone do not show the warming signature, but instead show continuously cold conditions until ca. 15.9 ka. The downstream effect brought back near-glacial conditions in adjacent regions, and caused glacier readvances and cooling in Europe and on Greenland. The proximity of the large ice sheet and cold ocean prevented the warming from being recorded in ice-proximal regions in northern and western Europe. In contrast, north and south of the maximum ice-rafted detritus zone, warming began prior to H1 and the records show modest or no cooling during the H1 time frame (Oppo and Lehman, 1995; BOFS 11K and 17K in Maslin et al., 1995; Zahn et al., 1997). Likewise, in southern parts of Europe, areas distal to the North Atlantic and the ice sheet, the early warming was recorded and the cool H1 period caused merely a halt of the warming.

The warming-cooling-warming sequence that began at 17.5 ka in the North Atlantic (in V29-191) also occurs in data from widely remote areas. The North Pacific polar front retreated shortly after 17.6 ka with a brief halt or readvance at 15.7 ka (Thunell and Mortyn, 1995). Variations in an alkenone record from the Indian Ocean suggest two warming intervals between ca. 19.5 and ca. 17.5 ka prior to a cooling during the H1 period (Bard et al., 1997). In the Southern Hemisphere, a warming in the Chilean Andes beginning at 17.3 ka featured a stepwise pattern (Moreno et al., 1999). Our data cannot clarify whether the H1 event was entirely produced by local surface cooling from latent heat loss or if the local cooling was superimposed on a millennial-scale temperature cooling, as suggested by Alley (1998). Nor can it identify the cause of the early warming, whether it is increased insolation or influx of tropical water that resulted in ice-sheet melting and iceberg discharge, as proposed by McIntyre and Molfino (1996). In each case, the large physical impact of icebergs cannot be disregarded. The mechanism connecting the North Atlantic climate changes and remote regions may lie in the large-scale effects of the surface-water conditions and the ice rafting, which not only altered local sea-surface temperatures and atmospheric circulation, but also surface-water salinities, potentially affecting the thermohaline circulation (e.g., Bond et al., 1993; Keigwin and Lehman, 1994; Zahn et al., 1997).

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REFERENCES CITED

Alley, R. B., 1998, Icing the North Atlantic: Nature, v. 392, p. 335–337.

Bard, E., Arnold, M., Maurice, P., Duprat, J., Moyes, J., and Duplessy, J.-C., 1987, Retreat velocity of the North Atlantic polar front during the last deglaciation determined by ^{14}C accelerator mass spectrometry: Nature, v. 328, p. 791–794.

Bard, E., Arnold, M., Fairbanks, R. G., and Hamelin, B., 1993, ^{230}Th-^{234}U and ^{14}C ages obtained by mass spectrometry on corals: Radiocarbon, v. 35, p. 191–199.

Bard, E., Rostek, F., and Sonzogni, C., 1997, Interhemispheric synchrony of the last deglaciation inferred from alkenone palaeothermometry: Nature, v. 385, p. 707–710.

Bé, A. W. H., 1977, An ecological, zoogeographic and taxonomic review of recent planktonic foraminifera, *in* Ramsay, A. T. S., ed., Oceanic micropalaeontology, Volume 1: London, New York, Academic Press, p. 1–88.

Belknap, D. F., and Shipp, R. C., 1991, Seismic stratigraphy of glacial marine units, Maine inner shelf, *in* Anderson, J. B., and Ashley, G. M., eds., Glacial marine sedimentation: Paleoclimatic significance: Geological Society of America Special Paper 261, p. 137–157.

Birks, H. H., Gulliksen, S., Haflidason, H., Mangerud, J., and Possnert, G., 1996, New radiocarbon dates for the Vedde Ash and the Saksunarvatn Ash from western Norway: Quaternary Research, v. 48, p. 119–127.

Blunier, T., Chappellaz, J., Schwander, J., Dällenbach, A., Stauffer, B., Stocker, T. F., Raynaud, D., Jouzel, J., Clausen, H. B., Hammer, C. U., and Johnsen, S. J., 1998, Asynchrony of Antarctic and Greenland climate change during the last glacial period: Nature, v. 394, p. 739–743.

Bond, G. C., and Lotti, R., 1995, Iceberg discharges into the North Atlantic on millennial time scales during the last glaciation: Science, v. 267, p. 1005–1010.

Bond, G., Heinrich, H., Broecker, W., Labeyrie, L., McManus, J., Andrews, J., Huon, S., Jantschik, R., Clasen, S., Simet, C., Tedesco, K., Klas, M., Bonani, G., and Ivy, S., 1992, Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial period: Nature, v. 360, p. 245–249.

Bond, G., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J., and Bonani, G., 1993, Correlations between climate records from North Atlantic sediments and Greenland ice: Nature, v. 365, p. 143–147.

Broecker, W., Bond, G., Klas, M., Clark, E., and McManus, J., 1992, Origin of the northern Atlantic's Heinrich events: Climate Dynamics, v. 6, p. 265–273.

Coplen, T. B., Kendall, C., and Hopple, J., 1983, Comparison of stable isotope reference samples: Nature, v. 302, p. 236–238.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gundestrup, N. S., Hammer, C. U., Hvidberg, C. S., Steffensen, J. P., Sveinbjörnsdóttir, A. E., Jouzel, J., and Bond, G. C., 1993, Evidence for general instability of past climate from a 250 kyr ice-core record: Nature, v. 264, p. 218–220.

Denton, G. H., and Hendy, C. H., 1994, Younger Dryas age advance of Franz Josef glacier in the Southern Alps of New Zealand: Science, v. 264, p. 1434–1437.

Epstein, S., Buchsbaum, R., Lowenstam, H. A., and Urey, H. C., 1953, Revised carbonate-water isotopic temperature scale: Geological Society of America Bulletin, v. 64, p. 1315–1326.

Fairbanks, R. G., 1989, A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation: Nature, v. 342, p. 637–642.

Giraudi, C., and Frezzotti, M., 1997, Late Pleistocene glacial events in the central Apennines, Italy: Quaternary Research, v. 48, p. 280–290.

Grimm, E. C., Jacobson, G. L., Jr., Watts, W. A., Hansen, B. C. S., and Maasch, K. A., 1993, A 50,000-year record of climate oscillations from Florida and its temporal correlation with the Heinrich events: Science, v. 261, p. 198–200.

Heinrich, H., 1988, Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years: Quaternary Research, v. 29, p. 142–152.

Jones, G. A., and Keigwin, L. D., 1988, Evidence from Fram Strait (78°N) for early deglaciation: Nature, v. 336, p. 56–59.

Keigwin, L. D., and Lehman, S. J., 1994, Deep circulation change linked to HEINRICH event 1 and Younger Dryas in a middepth North Atlantic core: Paleoceanography, v. 9, p. 185–194.

Kennett, J. P., and Ingram, B. L., 1995, A 20,000-year record of ocean circulation and climate change from the Santa Barbara Basin: Nature, v. 377, p. 510–514.

Lagerklint, I. M., 1995, Abrupt late Quaternary climate changes and seasonality variations in the eastern North Atlantic [M.S. thesis]: Orono, University of Maine, 104 p.

Lehman, S. J., Jones, G. A., Keigwin, L. D., Andersen, E. S., Butenko, G., and Østmo, S.-R., 1991, Initiation of Fennoscandian ice-sheet retreat during the last deglaciation: Nature, v. 349, p. 513–516.

Lister, G. S., 1988, A 15,000-year isotopic record from Lake Zürich of deglaciation and climatic change in Switzerland: Quaternary Research, v. 29, p. 129–141.

Lowell, T. V., Heusser, C. J., Andersen, B. G., Moreno, P. I., Hauser, A., Heusser, L. E., Schlüchter, C., Marchant, D. R., and Denton, G. H., 1995, Interhemispheric correlation of late Pleistocene glacial events: Science, v. 269, p. 1541–1549.

Lundqvist, J., and Saarnisto, M., 1995, Summary of Project IGCP-253: Quaternary International, v. 28, p. 9–17.

Maslin, M. A., Shackleton, N. J., and Pflaumann, U., 1995, Surface water temperature, salinity, and density changes in the northeast Atlantic during the last 45,000 years: Heinrich events, deep water formation, and climatic rebounds: Paleoceanography, v. 10, p. 527–544.

McCabe, A. M., and Clark, P. U., 1998, Ice-sheet variability around the North Atlantic Ocean during the last deglaciation: Nature, v. 392, p. 373–377.

McIntyre, A., and Molfino, B., 1996, Forcing of Atlantic equatorial and subpolar millennial cycles by precession: Science, v. 274, p. 1867–1870.

Moreno, P. I., Lowell, T. V., Denton, G. H., Jacobson, G. L., Jr., and Andersen, B. G., 1999, Abrupt changes in vegetation and climate during the last glacial maximum and last termination in the Chilean Lake District: A case study from Canal de la Puntilla (41°S): Geografiska Annaler, v. 80A, p. 285–311.

Oeschger, H., Beer, J., Siegenthaler, U., Stauffer, B., Dansgaard, W., and Langway, C. C., 1984, Late glacial climate history from ice cores, *in* Hansen, J. E., and Takahaski, T., eds., Climate processes and climate sensitivity: American Geophysical Union Geophysical Monograph 29, p. 299–306.

Oppo, D. W., and Lehman, S. J., 1995, Suborbital timescale variability of North Atlantic Deep Water during the past 200,000 years: Paleoceanography, v. 10, p. 901–910.

Reille, M., and Lowe, J. J., 1993, A re-evaluation of the vegetation history of the eastern Pyrenees (France) from the end of the last glacial to the present: Quaternary Science Reviews, v. 12, p. 47–77.

Ruddiman, W. F., 1977, Late Quaternary deposition of ice-rafted sand in the subpolar North Atlantic (lat 40° to 65°): Geological Society of America Bulletin, v. 88, p. 1813–1827.

Ruddiman, W. F., and McIntyre, A., 1981, The North Atlantic Ocean during the last deglaciation: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 35, p. 145–214.

Schlüchter, C., 1988, The deglaciation of the Swiss-Alps: A paleoclimatic event with chronological problems: Bulletin de l'Association Française pour l'Étude du Quaternaire, v. 2/3, p. 141–145.

Shackleton, N., 1967, Oxygen isotope analyses and Pleistocene temperatures reassessed: Nature, v. 215, p. 15–17.

Sowers, T., and Bender, M., 1995, Climate records covering the last deglaciation: Science, v. 269, p. 210–214.

Stein, R., Nam, S.-I., Schubert, C., Vogt, C., Fütterer, D., and Heinemeier, J., 1994, The last deglaciation event in the eastern central Arctic Ocean: Science, v. 264, p. 692–696.

Stuiver, M., and Braziunas, T. F., 1993, Modeling atmospheric ^{14}C influences and ^{14}C ages of marine samples to 10,000 BC: Radiocarbon, v. 35, p. 137–189.

Thunell, R. C., and Mortyn, P. G., 1995, Glacial climate instability in the Northeast Pacific Ocean: Nature, v. 376, p. 504–506.

Walker, M. J. C., 1995, Climatic changes in Europe during the last glacial/interglacial transition: Quaternary International, v. 28, p. 63–76.

Zahn, R., Schönfeld, J., Kudrass, H.-R., Park, M.-H., Erlenkeuser, H., and Grootes, P., 1997, Thermohaline instability in the North Atlantic during meltwater events: Stable isotope and ice-rafted detritus records from core SO75-26KL, Portuguese margin: Paleoceanography, v. 12, p. 696–710.

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