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Glacial/Interglacial response rate of subpolar North Atlantic waters to climatic change: the record in oceanic sediments

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Estimated rates of response of northeast Atlantic surface waters to large-scale palaeoclimatic changes have been reconstructed in two forms: (1) as changes through time of estimated temperature at selected points in space, and (2) as movements through space of the polar front during particular intervals in time. Three periods of rapid change between the glacial and interglacial climatic extremes were examined: a short cold episode during isotopic stage 7 (the next-to-last interglaciation); the deglacial warming into the last interglaciation at the isotopic stage 6/5 boundary; and the most recent deglacial warming into the present interglaciation at the isotopic stage 2/1 boundary.

Changes in sea-surface temperature of 7–11 °C (estimated from transfer function analysis) are characteristically registered in these cores in a few thousand years. The corresponding temporal rates of cooling or warming recorded usually average 1–5 °C/1000 years (a) for the complete climatic shift. During local passage of the polar front, these rates are even higher. Regional advance and retreat rates of the polar front along a NW/SE axis from Cape Farewell, Greenland, to Spain characteristically range from 200 to 1600 m/a during these intervals. These estimates represent the rates of change presently recorded in the sediments. The actual (faster) rates of palaeo-oceanographic change in the overlying North Atlantic surface waters will only be determined once the smoothing effects of vertical mixing can be removed.

Introduction

The world ocean is important both as (1) a permanent geologic monitor of late Quaternary climatic change and (2) an integral part of the land–ice–water–air system in which the changes occur. One critical factor in the kinematics of that system is the response rate of the surface ocean to climatic change. Modern annual oceanic responses are but small oscillations around the interglacial configuration and do not bear directly on larger scale variations and longer terms rates of response. In this paper we will primarily focus on the rate of change of the surface ocean between the extreme glacial and interglacial configurations in both time and space.

The record left in the sediments by subpolar North Atlantic surface waters shows a strong response to past oscillations of mean global climate. This response has been interpreted as temporal changes in the palaeobiota and in estimated sea-surface palaeotemperature at selected points in space (Bramlette & Bradley 1941; McIntyre, Ruddiman & Jantzen 1972; Sancetta, Imbrie & Kipp 1973). It also reflects spatial changes in watermass and oceanic frontal positions during selected time intervals (McIntyre *et al.* 1972; Ruddiman & McIntyre 1973, 1976). These studies show that palaeotemperature changes and watermass migrations in the subpolar North Atlantic are at least as large as in any other area of the world ocean.

These strong changes, combined with the high quality of the sedimentary record, make the subpolar North Atlantic an ideal location to define the maximum oceanic response rates recorded in deep-sea sediments. With this as our objective, we have taken from the record of the last 200 000 years three intervals registering prominent and rapid palaeoclimatic shifts between the extreme glacial and interglacial oceanic configurations. By examining faunal evidence of these changes in several cores, we have attempted to overcome local biases at individual core sites. Future studies will be necessary to remove the blurring effects imposed by the mixing activities of burrowing organisms in order to derive the actual rates of oceanic response.

Selection of climatic levels for study

In the subpolar North Atlantic, changes in percentage of the single foraminiferal species with polar watermass preferences – *G. pachyderma*, left-coiling – are an excellent first-order indicator of oceanic sensitivity to climatic change. Changes in abundance of this species as a percentage of the total foraminiferal fauna are shown in figure 1 for a portion of one core spanning the last 600 000 years (Ruddiman & McIntyre 1976). High percentages indicate low temperatures and salinities (glacial climates), and low percentages signify higher temperatures and salinities (interglacial climates).

The palaeoclimatic changes that we will examine are located by arrows in figure 1. They fall within faunal stages numbered to equate with the isotopic stages of Emiliani (1955, 1966) and Shackleton & Opdyke (1973). The correspondence of isotopic and faunal stages throughout the last 600 000 years has been demonstrated by Ruddiman & McIntyre (1976). The intervals that we have selected are: (1) an intensely cold climatic episode within interglacial stage 7; (2) the glacial-to-interglacial change across the stage 6/5 boundary (isotopic termination II of Broecker & van Donk (1970), followed by a short (stage 5d) cooling about 17 000 years later; and (3) the glacial-to-interglacial shift across the stage 2/1 boundary (termination I), including the very brief Younger Dryas cooling within the lowest portion of isotopic stage 1. Because the Younger Dryas cold episode lasted less than 1000 years, the initial sampling resolution of about 10 cm (2500 years) in figure 1 missed it entirely.

All three intervals have broad climatic significance. The stage 7 cold episode during the next-to-last interglaciation represents one potential analogue for future climate. It is the fastest complete interglacial-to-glacial-to-interglacial oceanic event that we have verified in this suite of northeast Atlantic cores. The stage 6/5 boundary was selected because the fauna consistently showed the fastest complete deglacial termination of oceanic glacial conditions of any of the major Brunhes climatic oscillations (Ruddiman & McIntyre 1976). The stage 5d cooling that followed represents a second analogue for future climate. The stage 2/1 boundary warming was chosen because it is the most recent deglacial termination, is accessible to ^{14}C dating, and is thus directly comparable to continental data during the same period. In addition, the subsequent Younger Dryas cold episode at 10200 a B.P. is an important indication that a very brief but intense climatic change can still be discriminated in cores with high deposition rates.

The cores in which we have examined one or more of these changes are located in figure 2 and table 1. These cores have relatively high deposition rates (2.4–15 cm/1000 a), virtually undissolved carbonate fractions, and excellent long-term stratigraphy.

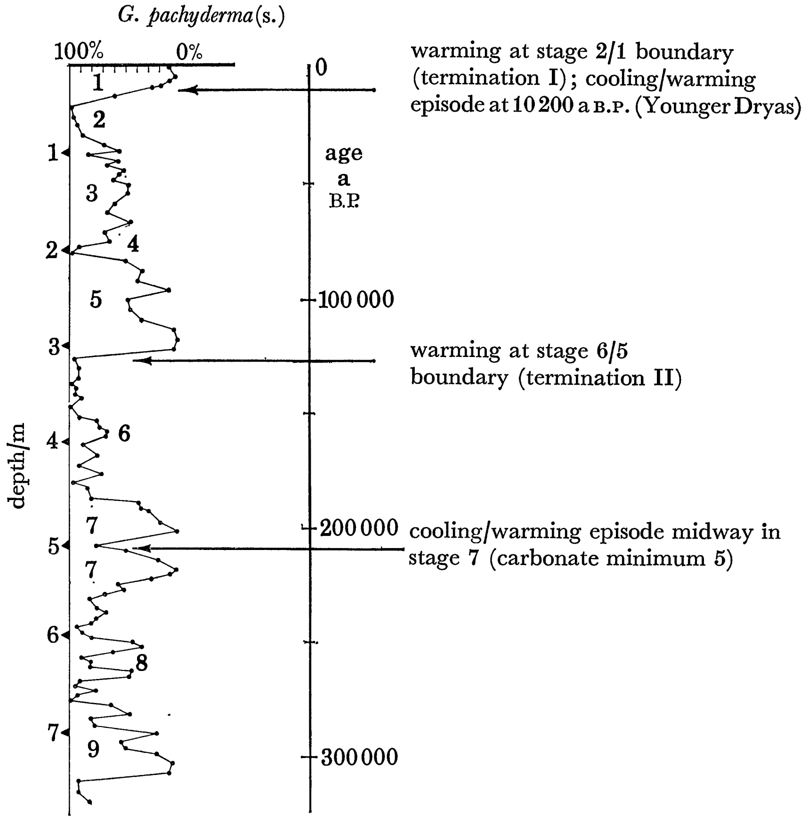

FIGURE 1. Placement of three studied intervals in long-term record of climatic change. Polar fauna as a percent of total planktonic foraminifera larger than 149 μm after Ruddiman & McIntyre (1976). Large numbers next to curve are designations of faunal/isotopic stages. Small numbers to left are metres of depth in core K708-7.

TABLE 1. LOCATIONS AND DEPTHS OF NORTHEAST ATLANTIC CORES

core	north latitude	west longitude	depth/m
K708-1	50° 00′	23° 45′	4053
K708-4	49° 59′	25° 01′	3346
K708-6	51° 34′	29° 34′	2469
K708-7	53° 56′	24° 05′	3502
K708-8	52° 45′	22° 33′	4009
RC9-225	54° 59′	15° 24′	2334
RC9-228	52° 33′	18° 45′	3981
V23-81	54° 15′	16° 50′	2393
V23-82	52° 35′	21° 56′	3974
V23-83	49° 52′	24° 15′	3871

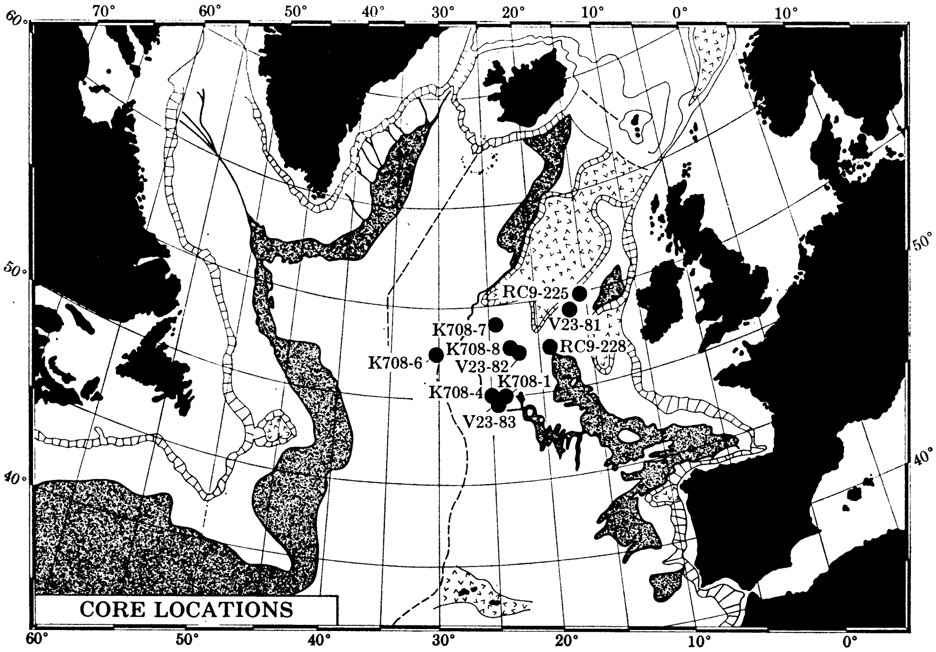

FIGURE 2. Locations of cores studied. Stippled areas are turbidite provinces. Regions with V pattern are shallow plateaux. Continental slopes are shown by short hachured lines. Axes of Mid-Atlantic and Iceland–Faeroe Ridges are dashed.

TABLE 2*a*. TYPE OF STRATIGRAPHIC CONTROL LEVELS FOR STAGE 2/1 WARMING AND YOUNGER DRYAS COLD EPISODE

nature of stratigraphic level	age of stratigraphic level/a	method	references
core top	0	assumed	—
ash zone 1 peak	9300	^{14}C on surrounding carbonate in several cores	1, 2
Younger Dryas cold peak	10220	^{14}C on 61–77 cm in core V23-82	1, 3
faunal termination 1	~13500	interpolation between ^{14}C dates above and below faunal change in several cores	1, 3, 4

1. Ruddiman & McIntrye (1973).
2. Ruddiman & Glover (1975).
3. Sancetta *et al.* (1973).
4. McIntyre *et al.* (1972).

TABLE 2*b*. DEPTHS OF STAGE 2/1 CONTROL LEVELS IN CORES STUDIED

	depth in core/cm				
control level (age)	K708-1	V23-82	RC9-228	V23-81	RC9-225
core top (0)	0	0	0	0	0
peak of ash zone 1 (9300)	62	63	—	—	70
Younger Dryas cold peak (10220)	(70†)	70 (^{14}C date)	(92†)	(157†)	(75†)
faunal termination 1 (~13500)	(85†)	88	120	213	93
other ^{14}C dates	—	—	—	40 (4010 a)	—

(†) These levels not used as time control to plot figures 7–8.

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Stratigraphy and temporal plotting assumptions

For the climatic change at the stage 2/1 boundary, we used actual ^{14}C dates for primary control,† as well as faunal or mineralogic levels correlative with ^{14}C-dated horizons in other cores (table 2). We then made simple linear interpolations between all listed control levels to transform each sample depth into an estimate of actual age.

The major uncertainty in the plotted levels lies in the age of the faunal warming near the stage 2/1 boundary. Ruddiman & McIntyre (1973) calculated an extrapolated age of 13500 a B.P. from overlying sedimentation rates in several cores, but did not date the transition directly using ^{14}C. We recently obtained a ^{14}C age of 14825 ± 365 a B.P. for the section of core V23-81 from 200–185 cm, which would place the faunal boundary (212 cm) at 16 300 a B.P. We regard this date as unrealistically old, probably because of contamination by older glacial-age carbonate, and have used the 13500 a B.P. date.

Table 3*a*. Type of stratigraphic control levels for calculating and testing long-term sedimentation rates in northeast Atlantic cores

nature of stratigraphic level	age of stratigraphic level	method of age determination	references
stage 6/5 boundary (faunal equivalent of isotopic termination II)	126000 a B.P. ($\pm$ 6000 a)	interpolation from $^{230}Th/^{234}U$ and $^{231}Pa/^{235}U$ control	1, 2
stage 2/1 boundary (faunal equivalent of isotopic termination I)		see table 2*a*	
ash zone 1 peak		see table 2*a*	
Barbados 1 high sea level; isotopic stage 5*a* (faunal equivalent)	82000 a B.P. ($\pm$ 4000 a)	$^{230}Th/^{234}U$ $^{231}Pa/^{235}U$	1

1. Broecker *et al.* (1968).
2. Broecker & Ku (1969).

Table 3*b*. Depths of long-term control levels (cm) and rates of sedimentation (cm/1000 a)

	cores							
	K708-1	K708-4	K708-6	K708-7	K708-8	RC9-225	V23-82	V23-83
stage 6/5 boundary (127000 a B.P.)	858	358	445	307	550	920	805	675
approx. stage 2/1 boundary (9300 B.P.)	62	34	28	28	45	70	63	70[a]
mean rate of sedimentation (stage 6/5–stage 2/1)	6.77	2.75	3.54	2.37	4.21	7.22	6.30	5.33
Barbados 1 high sea level faunal equivalent (82000 a B.P.)	460	235	290	215	365	600	470	400

[a] Faunal termination 1 (13500 a B.P.) used instead of ash zone 1.

† Stuiver & Borns (1975) have shown that no age correction is necessary for ^{14}C ages from deep-sea sediments due to equal and exactly offsetting errors caused by (1) biological fractionation and (2) mixed layer composition.

Table 4. Comparison of interglacial and glacial sedimentation rates in cm/1000 a for northeast Atlantic cores

core	rate during last glaciation	rate during last interglaciation	ratio interglacial/glacial
K708-1	5.47	9.29	1.70
K708-4	2.75	2.76	1.00
K708-6	3.60	3.44	0.95
K708-7	2.57	2.04	0.79
K708-8	4.40	4.11	0.93
RC9-225	7.29	7.11	0.98
V23-82	5.60	7.44	1.33
V23-83	4.78	6.11	1.20
	$\bar{x}$ = 4.56	$\bar{x}$ = 5.29	$\bar{x}$ = 1.16

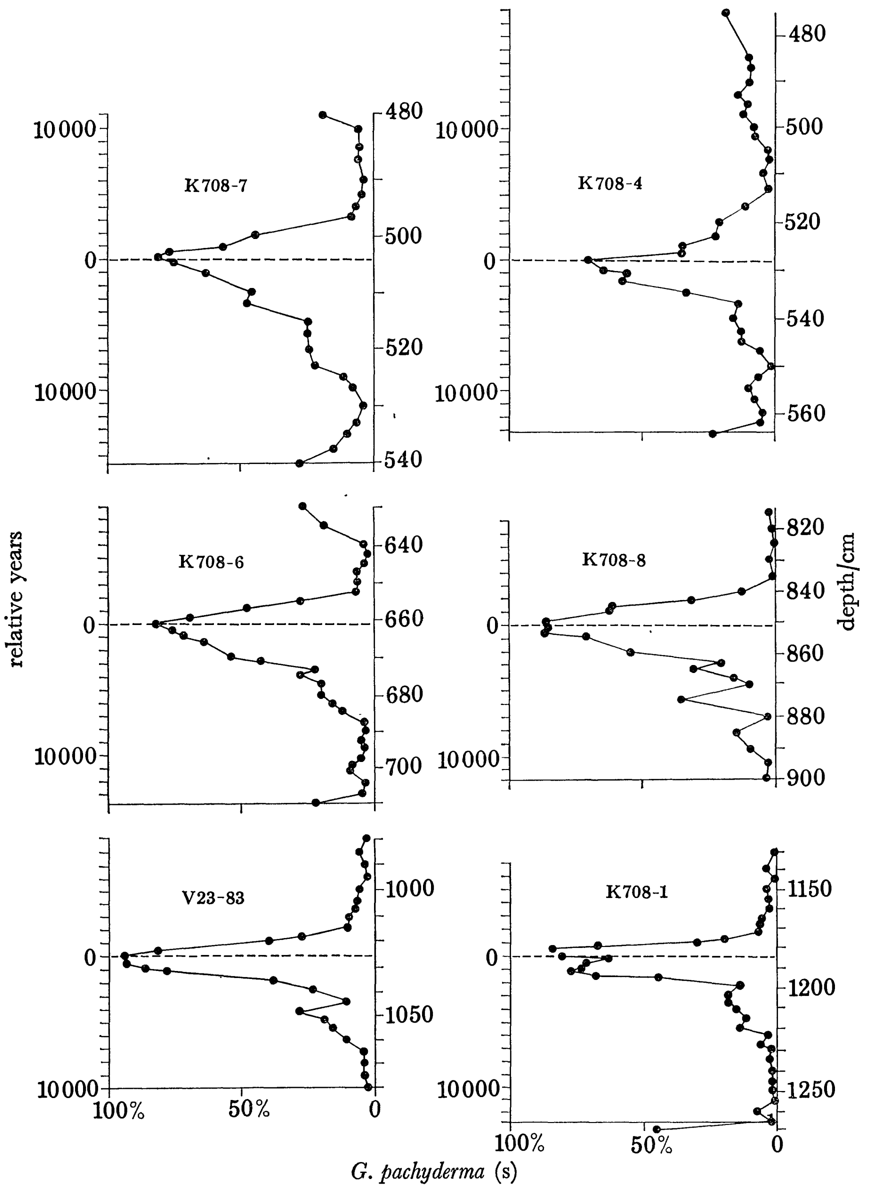

Figure 3. Down-core percentages of *G. pachyderma* (s.) (●) as a percent of total planktonic foraminifera larger than 149 μm during cold episode in middle of stage 7. Time plotted in relative years increasing in both directions away from midpoint of cold episode set at 0 a. Time plotting assumptions in table 3 and text. Core depth in centimetres shown on right.

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For the two earlier episodes (in stage 7 and at the stage 6/5 boundary), we made no attempt to specify absolute ages, although the cold episode in stage 7 occurred at roughly 195000 a B.P. (Ruddiman & McIntyre 1976) and the stage 6/5 boundary at about 127000 a B.P. (Broecker & Ku 1969; Ku, Bischoff & Boersma 1972). Instead, we chose to express the duration of these changes in a relative time framework. To do this, we determined in each core the long-term rate of sedimentation during the late Quaternary and assumed that this rate prevailed during the two intervals examined.

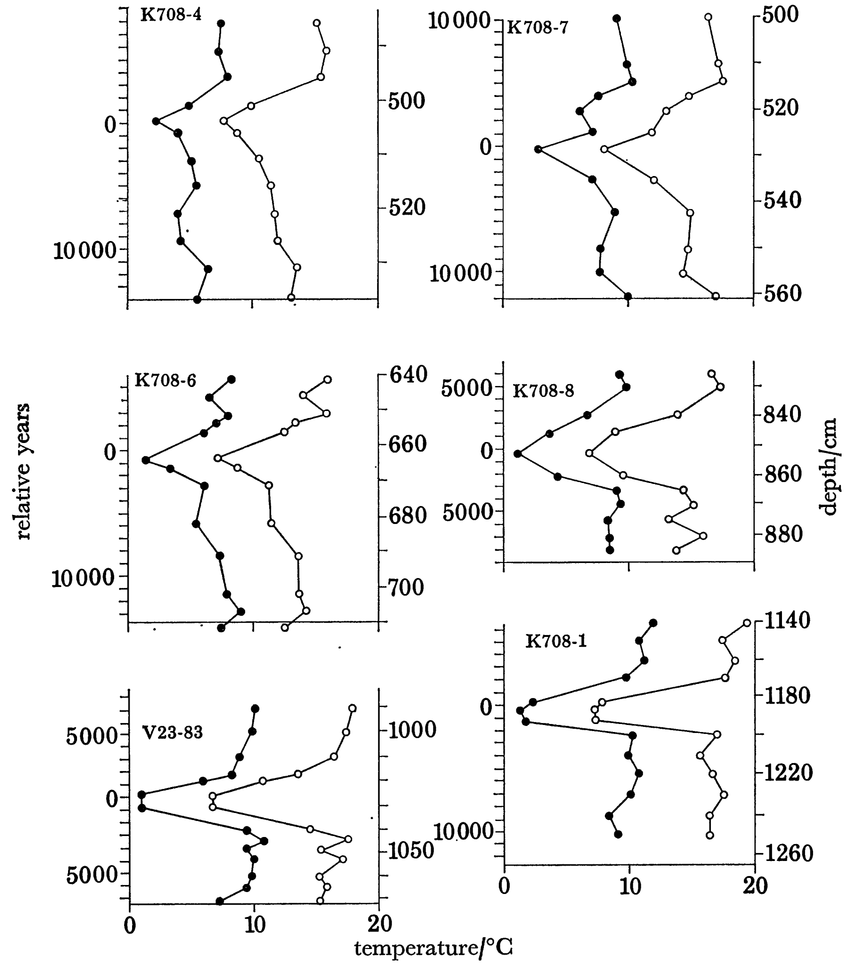

FIGURE 4. Estimates of summer (○) and winter (●) palaeotemperature (°C) during stage 7 cold episode. Time and depth plotted as in figure 3. Estimates based on transfer function equation F13B-4CE.

We selected the climatic cycle between stage boundaries 6/5 and 2/1 as the best long interval to provide a mean sedimentation rate (table 3*a*). The isotopic stage 6/5 boundary has been dated in other oceanic cores at 126000 ± 6000 a B.P. (Broecker & van Donk 1970) and correlated with North Atlantic faunal, floral and lithologic changes (McIntyre *et al.* 1972; Sancetta *et al.* 1973; Crowley 1975). The faunal stage 6/5 boundary is at most about 1000 years older than its isotopic counterpart in this region of the subpolar North Atlantic.† For the

† In these cores, the faunal stage 2/1 boundary precedes the isotopic stage 2/1 boundary dated at low latitudes by about 2000 years. Since the stage 6/5 deglaciation appears to have been faster, we estimate no more than a 1000-year lead time.

younger control level we used in most cases ash zone 1 at 9300 a B.P. (Ruddiman & Glover 1972). In one core (V23-83) where the ash zone was not examined, we used the faunal stage 2/1 boundary estimated at ~13500 a B.P. by Ruddiman & McIntyre (1973). Control levels and average sedimentation rates for this long interval (stage 6/5 to 2/1) are listed in table 3*b*.

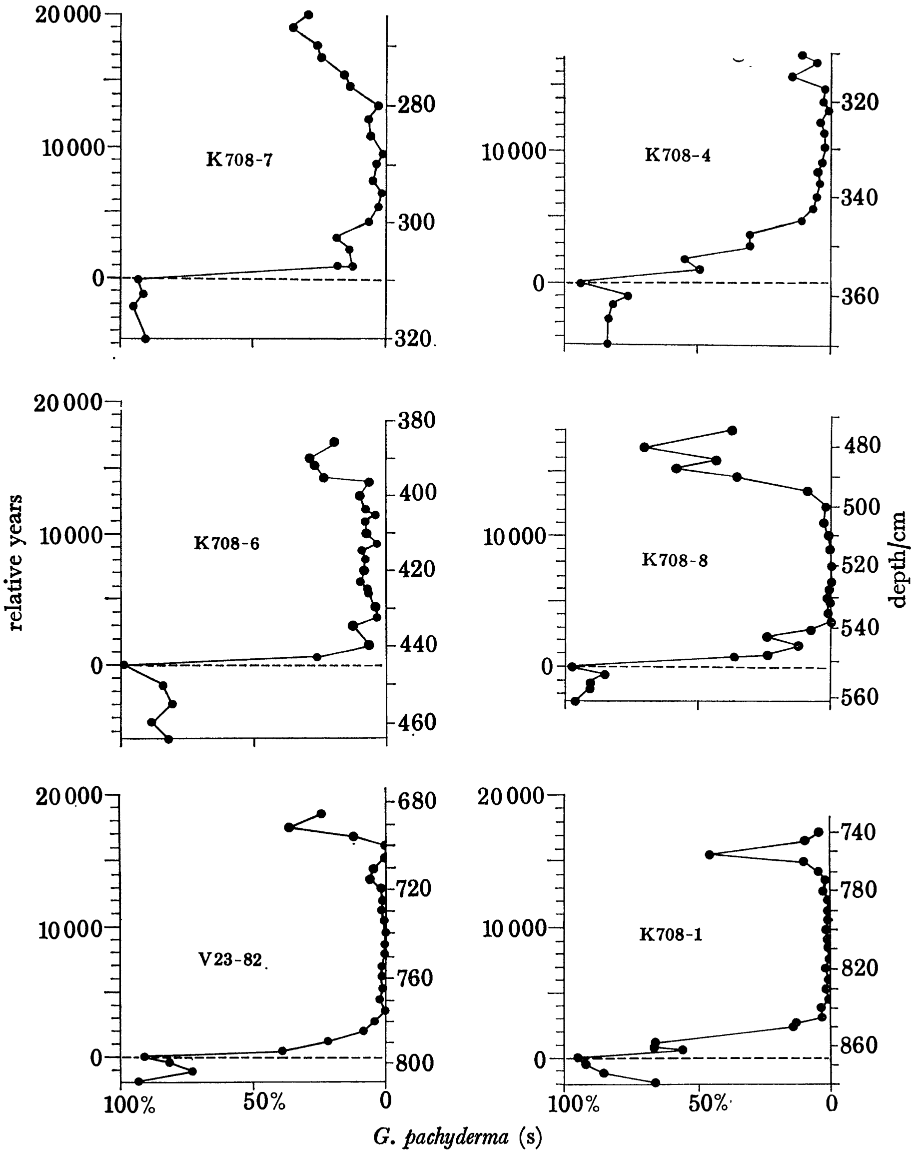

FIGURE 5. Down-core percentages of *G. pachyderma* (s.) (●) as a percent of total planktonic foraminifera larger than 149 μm during climatic warming at stage 6/5 boundary. Time plotted in relative years increasing upwards from uppermost full-glacial sample set at 0 a. Time plotting assumptions in table 3 and text. Core depth in centimetres shown on right.

To test the assumption of constant sedimentation rates, we used an additional control level located just below the stage 5/4 boundary and roughly midway within the interval used to calculate the mean sedimentation rates (table 3*b*, figure 1). We previously correlated this peak with the Barbados 1 coral reefs dated by U/He methods and indicative of a high stand of sea level at 82000 a B.P. (McIntyre *et al.* 1972; Mesolella, Matthews, Broecker & Thurber

1969). This correlation has been supported by other marine stratigraphic work (Sancetta *et al.* 1973; Crowley 1975).

The 82000 year level is equivalent to substage 5*a* of Shackleton (1969) and lies just below the stage 5/4 boundary. Beneath the stage 5a level lies an earlier interglacial interval (127000–82000 a B.P.) of highly calcareous coccolith-foraminiferal ooze mainly derived from overlying surface waters. Above it is a glacial interval (75000–10000 a B.P.) dominated by moderately

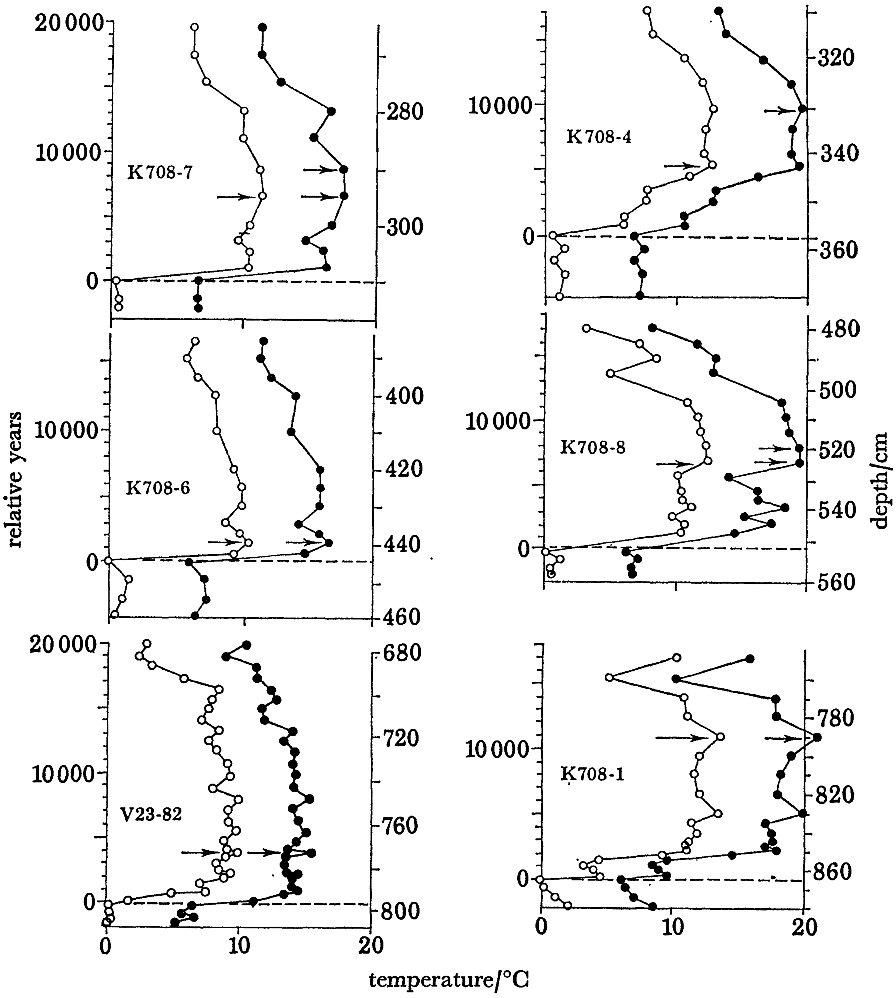

FIGURE 6. Estimates of summer (●) and winter (○) palaeotemperature (°C) during stage 6/5 warming. Time and depth plotted as in figure 5. Estimates based on transfer function equation F13B-4CE.

calcareous sandy lutites ice-rafted from continental sources. These two contrasting lithologic units provide a meaningful test of constancy in depositional rates during varying climate. As shown in table 4, the sedimentation rates are usually constant to within about 30 % and often to within 10 % (4 out of 8 cores). The interglacial mean for all cores is 16 % larger than the glacial average, indicating that sedimentation rates stay reasonably constant during changing climates on at least a long-term scale.

A finding of nearly constant sedimentation rates seems at odds with the gross changes in lithologic composition and sediment source during these two climatic intervals. Nevertheless, Ruddiman & McIntyre (1976) have shown that the constant rates of sedimentation indicate a coincidental but almost exactly balanced input of the two major components. Increases in the absolute input of one component balance and offset decreases in the absolute input of the other.

Because the sedimentation rates in these cores are reasonably constant over long intervals of time (10^5 a), we have assumed that they are also constant over the shorter intervals (10^3–10^4 a) encompassing the two early climatic episodes. We then plotted these episodes in units of relative time. We set the midpoint of the cooling episode in stage 7 at 0 a (not B.P.) and plotted relative years in both directions away from the midpoint to separate out visually the durations of the initial cooling and later warming phases (figures 3–4). For the stage 6/5 boundary, we selected the uppermost full-glacial sample as 0 a and plotted relative years upwards away from this level (figures 5–6).

The cores were sampled at time intervals of about 1000 years for the initial counts of polar fauna and at a spacing of about 200 years for the counts on which palaeotemperature estimates were made. With this spacing, no high-amplitude climatic episode recorded in the sediments should escape detection.

Procedures

Taxonomy

The taxonomy follows that of Kipp (1976), with three exceptions discussed by Ruddiman & Glover (1975). First, no attempt is made to distinguish intergrade forms between *Globigerina pachyderma* (right-coiling, or dextral) and *Neogloboquadrina dutertrei*. All intergrades are assigned to *G. pachyderma* (right-coiling), except for those with an umbilical tooth. Second, *Globorotalia menardii* and *G. tumida* are counted together. Third, forms of *Globigerinoides sacculifer* with and without sac-like last chambers are combined. Since the last two species account for no more than 0.5 % of any count, these taxonomic decisions have no vital effect on the palaeoecological interpretations. All counts from the three climatic episodes are listed in appendixes 1–3.†

Counting procedure

Counts of the planktonic foraminifera in the > 149 μm fraction were made in two ways: the first, a faster and less rigorous technique on samples at close-spaced intervals to search for short-term changes; the second, a slower but comprehensive technique run on samples indicated by the first technique as adequately representing the palaeoecologic trends.

The first technique requires counting only the single polar-water foraminifer *Globigerina pachyderma* (left-coiling, or sinistral) as a percentage of all foraminifera present in the > 149 μm fraction. Using portions of a sample aliquot strewn across a tray marked with perpendicular lines, counts were made along a diagonal series of boxes leading from the tray margin to its centre. The number counted ranged from 150 to 610 and averaged 275.

The technique used to count full aliquots of > 149 μm planktonic foraminifera for palaeotemperature analysis is that of Imbrie & Kipp (1971) and Sancetta *et al.* (1973). The number of individuals counted in all samples averaged 414 and ranged from 191 to 629.

† Appendixes 1–6 appear on the accompanying microfiche.

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Transfer function analysis

The basic methods of transfer function analysis were first derived by Imbrie & Kipp (1971) and later modified slightly by Imbrie, van Donk & Kipp (1973). To convert the foraminiferal counts listed in appendixes 1–3 to estimates of palaeotemperature, we used transfer function equation F13B-4CE (Ruddiman & Glover 1975).† Equation F13B-4CE is based on the same 191 core tops as the F13 set of Kipp (1976) but incorporates the three taxonomic changes. Relevant statistics in Ruddiman & Glover (1975) show lower standard errors and higher correlation coefficients than equation F13 on estimates of seasonal palaeotemperature, but higher standard errors and lower correlation coefficients on palaeosalinity estimates. None of the differences were large. The standard errors of estimate for equation F13B-4CE are ± 1.157 for winter palaeotemperature (T_w) and $+1.357$ for summer palaeotemperature (T_s). This represents roughly 10 % of the observed range of temperature variation (appendixes 4–6). Communalities (defined in Imbrie *et al.* 1971) for most samples were very high (appendixes 4–6), indicating that excellent analogues for most palaeoassemblages can be found in the model generated from the set of 191 core tops and used to calibrate the surface-sediment fauna against the modern oceanographic environment.

Berger & Gardner (1975) have discussed problems in separating past variations in temperature from those in salinity. If the strong correlation in the modern ocean between temperature and salinity were broken or shifted in some way, any palaeoestimating technique using the modern ocean for calibration would produce errors in palaeoestimates. In such a case, the ecological variable that is causally dominant would be best estimated, with larger errors for secondary variables. In this paper, we chose to make estimates only of palaeotemperature, the variable normally considered dominant (Imbrie & Kipp 1971).

The temperature estimates in this paper have an additional problem; there is an artificial lower limit due to the monospecific composition of the fauna in polar waters. Because there is no variation possible beyond the essentially 100 % of *G. pachyderma* (s.) found just behind the modern Polar Front, no temperature estimates lower than -0.1 °C in winter and 5.6 °C in summer can be made for older levels. All estimates at or near these limits should be understood to represent 'less than or equal to' values. Additional work along the lines initiated by Kennett (1968, 1970) and by Vilks (1975) on morphologic variations within the species *G. pachyderma* (s.) might extend the range of temperature estimates to somewhat colder values.

Observed rates of change: temporal

Down-core changes in estimated palaeotemperature during the specified intervals of time (figures 4, 6 and 8) are the basis for establishing the recorded rates of change in oceanic temperature at points in space. Procedurally, we reconstructed these rates of change in two ways. First, we measured the full temperature change and time duration between the coldest (full-glacial) and warmest (full-interglacial) sample in a chosen interval. This yields the mean rate of change across the entire climatic transition (table 5).

Unfortunately, interglacial temperature estimates tend to vary within a rather narrow range around a mean value for long intervals of time. The warmest sample usually does not stand out above the other temperature estimates by an amount exceeding the standard error of

† Equation F13 was used for estimates in cores V23–81 and V23–82.

Table 5. Average amplitudes, durations and rates of oceanic coolings and warmings recorded for three climatic episodes

	estimated summer temp./°C†	estimated winter temp./°C†	average duration of change between peaks/a†	estimated rate of cooling °C/1000 a†‡	estimated rate of warming °C/1000 a†‡
stage 7 episode					
initial warm peak	16.0	9.4			
cold episode peak	7.3	1.7	9200 (4400)	0.9 (1.4)	—
later warm peak	17.3	9.8	5800 (2900)	—	1.6 (2.3)
stage 6/5 boundary					
stage 6 cold peak	6.2	0.2			
stage 5 warm peak (5e)	18.4	11.7	7000 (1700)	—	1.7 (5.2)
stage 2/1 boundary: entire transition					
stage 2 cold peak	6.6	0.9			
stage 1 warm peak	16.4	10.8	10100 (2800)	—	1.0 (2.6)
stage 2/1 boundary: separate parts					
initial stage 2 cold peak	6.6	0.9			
last stage 2 warm peak	14.3	9.0	2700 (1500)	—	2.9 (3.9)
Younger Dryas cold peak	7.4	1.8	700 (400)	11.0 (13.2)	—
stage 1 warm peak	15.2	10.1	7700 (500)	—	1.1 (12.1)

† Values for the two oldest climatic episodes based on six-core averages calculated from data in figures 4 and 6. Values for stage 2/1 boundary based on five-core average, except for Younger Dryas cold peak, which is based only on cores V23-81 and RC9-225.

‡ Values on parentheses apply to the 75% of temperature change occurring toward or away from the coldest sample. Based on the average temperature change of the winter and summer temperature estimates.

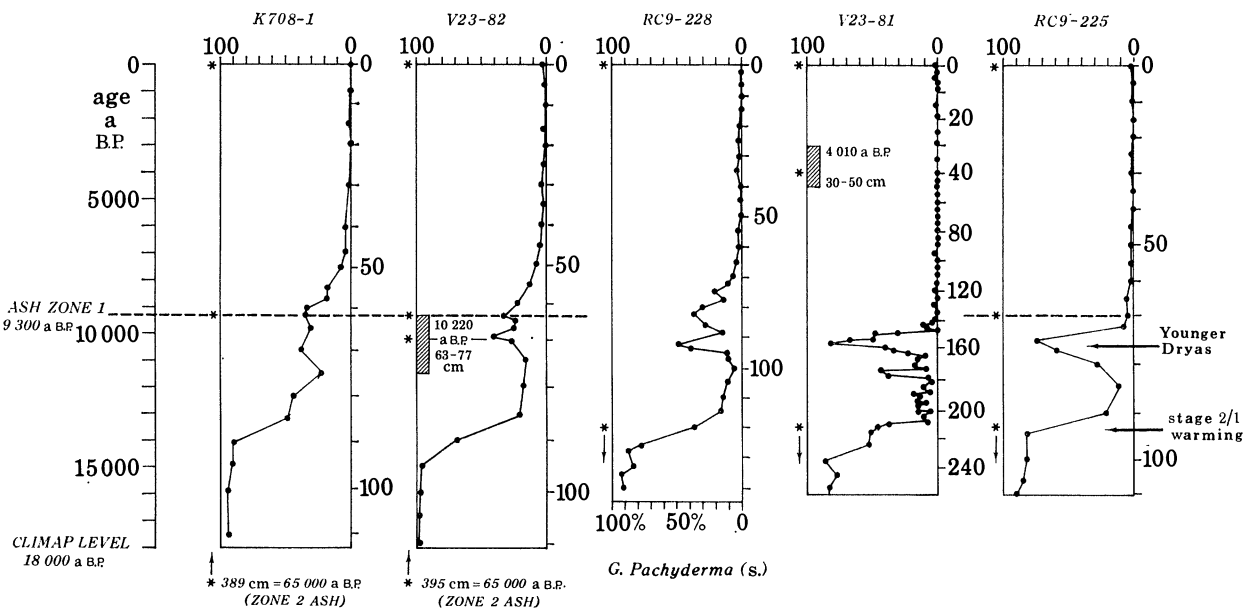

Figure 7. Down-core percentages of *G. pachyderma* (s.) as a percent of total planktonic foraminifera larger than 149 µm during climatic warming at stage 2/1 boundary, including Younger Dryas cold episode. Time plotted in absolute ^{14}C a B.P. by interpolation between levels listed in table 2.

estimate (±1.2 °C). In these cases, the choice of a 'warmest' sample thus seems somewhat arbitrary, and its interpolated age from core to core can vary considerably (for example, note the discrepancies among the arrows in figure 8 locating the warmest Holocene temperature estimates). In addition, the largest part of the temperature changes in most cores is recorded in a relatively short period of time, usually occurring near the cold peak.

Consequently, we have also measured the amount of time required for 75 % of each warming or cooling to take place (table 5, values in parentheses). We then calculated the substantially faster rates of change across these shorter intervals.

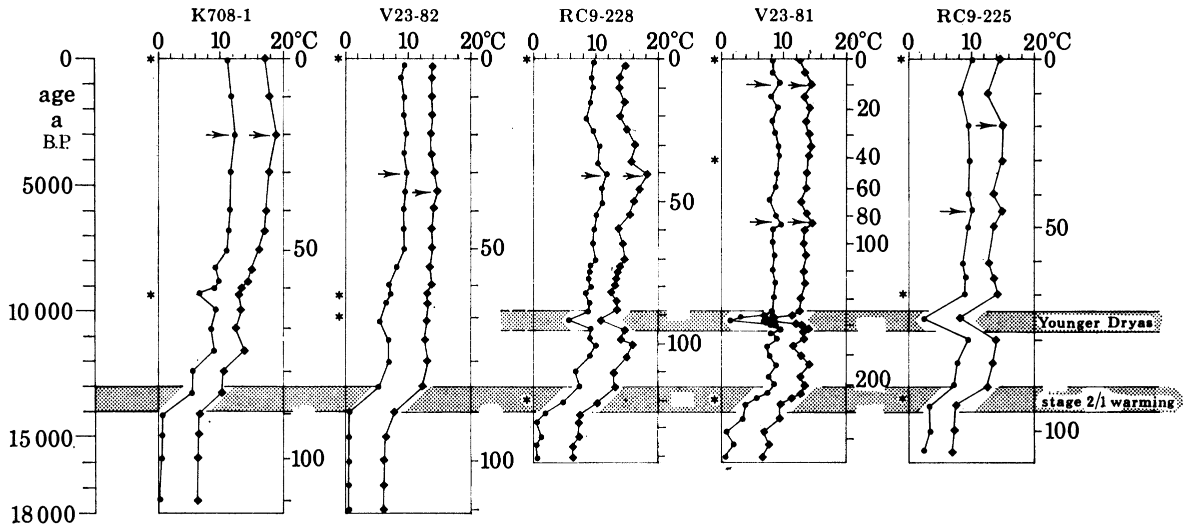

FIGURE 8. Estimates of summer (●) and winter (◆) palaeotemperature (°C) during stage 2/1 warming and Younger Dryas cold episode. Time and depth plotted as in figure 7. Estimates based on transfer function equation F13B-4CE. Arrows show warmest stage 1 estimates in all five cores.

Because post-depositional mixing has presumably lowered the amplitudes and lengthened the durations of the original oceanic variations, we stress that all these rates are merely our best compilation of the smoothed record left in the sediments. They are probably not the actual rates of oceanic response.

Stage 7 episode

The stage 7 cooling episode is slightly asymmetrical, with a slower initial cooling and a faster subsequent warming (figures 3–4). The estimated summer temperature (averaged in six cores) started at 16.0 °C early in stage 7, fell to 7.3 °C in the middle of the cold episode, and warmed to 17.3 °C late in stage 7 (figure 4). At the same time, estimated winter temperature cooled from 9.4 to 1.7 °C and then warmed to 9.8 °C. These changes reflect an initial gradual decrease, followed by a more rapid increase in polar fauna (figure 3).

The initial temperature shift from the warmest sample in the bottom half of stage 7 to the peak cold sample is recorded in an average of 9200 years, with the later change from the cold peak to the warmest overlying sample in 5800 years (figure 4, table 5). We also calculate that 75 % of the initial cooling is registered in an average of 4400 years, and an equal portion of the later warming in an average of 290C years (table 5).

For the entire initial cooling in the stage 7 episode, the recorded rates of temperature change averaged just under 1 °C/1000 a (table 5). Along the steeper slopes near the cold peak, the rates rise to slightly higher values. Similarly, for the later peak-to-peak warming in the upper half of the stage 7 episode, the recorded rates of change average 1.6 °C/1000 a (table 5) but reach peak values above 2 °C/1000 a.

The duration of the stage 7 cold episode appears briefest in cores K708-1 and V23-83, longest in core K708-7, and intermediate in the three other cores. We infer that these apparent differences in part actually reflect relative glacial/interglacial changes in sedimentation rate. The long-term changes listed in table 4 correctly predict that glacial episodes will be somewhat under-represented (shorter in appearance) in cores K708-1 and V23-83, over-represented in K708-7, and reflected fairly accurately for the other three cores.

Increased percentages of ice-rafted sand and lutite in the stage 7 cold episode of these cores suggest that there was significant glacier growth at this time over eastern North America and Scandinavia (Ruddiman & McIntyre 1976). Isotopic data from benthic foraminifera show a prominent glacial pulse at or just below the stage 7 midpoint (see, for example, Shackleton & Opdyke 1973). Cores with high sedimentation rates show that during this episode isotopic δ values in benthic foraminifera grew from levels typical of interglaciations to values as much as 80 % towards the glacial extreme and then fell again to interglacial levels (N. J. Shackleton, personal communication).

These northeast Atlantic cores thus registered virtually a complete interglacial-glacial-interglacial temperature oscillation in no more than 15000 years. During this period of time (or less), continental glaciers grew to 80 % of their maximum (full-glacial) size and wasted back again.

Stage 6/5 *boundary warming* (*termination II*) *and stage* 5d *cold episode*

A faster and even stronger interglacial-to-glacial warming occurred at the stage 6/5 boundary (figures 5–6). From estimated winter temperatures of 0.2 °C and summer temperatures of 6.2 °C in the coldest levels late in stage 6, the warming brought estimated sea-surface temperatures to 11.7 °C and 18.4 °C, respectively. This peak-to-peak change is recorded in an average of 7000 years; the first 75 % of the change is registered in just 1700 years (table 5). The corresponding rates of warming average 1.7 °C/1000 a for the entire peak-to-peak shift but reach 5.2 °C/1000 a in the more rapid change in the initial portions of the warming (figures 5–6).

The stage 6/5 boundary warming carried global climate from full-glacial conditions to an unusually extreme interglacial state. Isotopic data from Shackleton & Opdyke (1973) indicate that the uppermost portion of stage 6 may not be quite as intensely glacial as earlier stage 6 levels or as the stage 2 isotopic maximum. Still, a fully glacial oceanic climate late in stage 6 must be inferred from the totally dominant polar faunas at these levels (figure 7). Isotopic data indicate that during stage 5e the mass of continental ice was smaller than during any other Brunhes interglaciation (Shackleton 1969). This is supported by a stage 5e fauna and flora warmer than those of any interglaciation in the last 600000 years (Ruddiman & McIntyre 1976).

In summary, we find that a full glacial-to-interglacial oceanic warming in the northeast Atlantic, and apparently correlative with a total wasting of continental ice, was registered in no more than 6500 years, with most of the change recorded in less than 2000 years.

The stage 5d cold pulse that ended the extreme interglacial warmth of stage 5e reached its peak intensity an average of 17000 years after the beginning of termination II. Using slightly different stratigraphic assumptions from those in this paper, Sancetta *et al.* (1973) calculated an age of 110000 a B.P. for the peak cooling in core V23-82. The age of stage 5d in the 6 cores in this study ranges from 109000 to 111000 a B.P., with a standard deviation of just ±1100 a around the average of 110000 a B.P.

Stage 2/1 boundary warming (termination I and Younger Dryas)

Considered in its entirety, the deglacial warming into the Holocene was less abrupt than that at stage 6/5. The warmest temperatures were not reached until late in stage 1, about 9200 years after the beginning of deglacial warming at 13500 a B.P. (figures 7–8). The lengthened interval required to reach the peak interglaciation may be related to a reversal of deglacial warming centred at about 10200 a B.P. This short cold episode was detected and dated in core V23-82 and correlated to the Younger Dryas (Ruddiman & McIntyre 1973; Sancetta *et al.* 1973). It occurred at a time when the deglacial warming of the northeast Atlantic was not complete, and it briefly returned that ocean to an almost glacial state.

Viewed as a single climatic warming, the entire stage 2/1 shift registered a rise in mean summer temperatures from 6.6 to 16.4 °C and in winter temperatures from 0.9 to 10.8 °C in 9200 years (table 5). This represents a mean rate of 1.0 °C/1000 a. Roughly 75% of that change occurred in the 2800-year span between 13700 and 11900, at a mean rate of 2.6 °C/1000 a.

Viewed in terms of its separate parts, the long interval of deglaciation at the stage 2/1 boundary actually consists of three separate and rapidly recorded climatic changes: (1) an initial warming at about 13500 a B.P.; (2) a cooling beginning after 11500 a B.P. and culminating at about 10200 a B.P.; and (3) a later warming occurring for the most part before 9300 a B.P. but not reaching peak interglacial warmth until about 5000 a B.P.†

This general chronology is largely supported by climatic evidence in immediately adjacent continental regions. Coope (1975) used insect assemblages recorded in peat layers on Great Britain to estimate climatic changes at land sites only 10–15° of longitude (500–750 km) east of oceanic cores V23-81 and RC9-225. He inferred a strong and abrupt warming at about 13000 a B.P. (bracketed between a glacial insect assemblage dated at 14460 a B.P. at the Glanllynnau site and a warm coleopteran fauna dated at 12940 a B.P. from the site at Robert Hill). Mangerud & Gulliksen (1975) quote work in preparation on coastal molluscs asserting that warm North Atlantic waters reached western Norway before 12600 a B.P. These estimates substantiate a deglacial warming at about 13500 a B.P. that affected the easternmost North Atlantic and westernmost Europe and Scandinavia. If this is correct, it challenges the conventional palaeoclimatic interpretations of European (or at least British) pollen data. Specifically, it can be argued that the British arboreal pollen response to an actual atmospheric climatic warming that occurred about 13500 a B.P. lagged as much as 1500 years or more behind the faster response of the insects (Coope 1975), the aquatic pollen (Iversen 1954), the coastal molluscs (Mangerud & Gulliksen 1975) and the open-ocean planktonic foraminifera (Ruddiman & McIntyre 1973).

Coope (1975) inferred a more gradual cooling into the Younter Dryas peak, culminating shortly before 10000 a B.P., followed by a rapid warming out of the Younger Dryas. This equates more closely with the conventional European interpretation of the Younger Dryas cooling, as does the oceanic record.

In the oceanic record, estimated winter ocean temperatures (calculated from the average

† In other contributions to the symposium, the initial warming is referred to as the 'Windermere Interstadial' (Britain) and the Woodgrange Interstadial (Ireland). The later warming into stage 1 is the 'Flandrian' (Britain) and the Littletonian (Ireland). Stage 2 glacial conditions are 'Late Devensian'.

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of all five cores) rose during the initial (13500 a B.P.) warming from 0.9 to 9.0 °C, as summer sea-surface temperature increased from 6.6 to 14.3 °C.† This part of the deglacial warming was recorded in an average of 2700 years, for a mean rate of warming of 2.9 °C/1000 a (figure 8, table 5). The first 75 % of the warming, however, is registered in 1500 years, for a rate during that period of almost 4 °C/1000 a.

Analysis of the short Younger Dryas cooling is more complex: the episode has different peak intensities in the five cores sampled in detail (figures 7–8). This mostly reflects actual differences in the overlying surface waters during the Younger Dryas. The climatic changes discussed so far have minimized regional differences because they all document abrupt shifts from one climatic extreme to another. As a result, all cores have shown virtually the same climatic response. Because the Younger Dryas cold episode was more moderate in intensity, it affected the three northern cores far more severely than the two southern cores. This will be explained from a geographic viewpoint in the next section.

Mixing may have altered in a differential manner the record of the Younger Dryas cooling in these five cores. If so, the cores with highest sedimentation rates should be least mixed and register the smallest loss of resolution of short events. Some such effect on the cores is suggested in figures 7 and 8, which show strongest Younger Dryas peaks in the three cores to the north with highest sedimentation rates; however, core RC9-225 to the north has a rate of deposition comparable to the two southern cores and yet registers a much more intense Younger Dryas peak. We conclude that the north-to-south temperature gradient is for the most part real, and that the Younger Dryas cooling was far more intense in cores north of 53° N.

An additional complication is climate-related variations in sedimentation rate. Even though we demonstrated constancy of sedimentation rate for most of the cores in this study, the core with the highest depositional rate (V23-81) clearly breaks the pattern. The Würm sedimentation rate from 75000 to 13500 a B.P. is 8.8 cm/1000 a. From 13500 a B.P. to the Younger Dryas at 10200 a B.P., the rate rises to 21 cm/1000 a; from the Younger Dryas to the ^{14}C date at 4010 a B.P., it remains high at 18.9 cm/1000 a; and from the ^{14}C date to the present it falls to 10 cm/1000 a. Because the sedimentation rates vary by a factor of two or more due to climate, we think that the duration of the Younger Dryas cooling on a time plot could be somewhat compressed relative to the remainder of the Holocene record in this core. This is analogous to the differing apparent durations of the stage 7 cooling discussed earlier.

Weighing all these complexities, we have used only the two far northern cores (V23-81, RC9-225) to compute mean rates of change. During the warm-to-cold shift into the Younger Dryas, estimated winter temperatures cooled from 8.2 to 1.8 °C and summer temperatures from 14.3 to 7.4 °C (figure 8; table 5). This change is recorded in 700 years, giving a mean rate of cooling of 11 °C/1000 a. Recorded rates of cooling near the cold peak reached values of over 13 °C/1000 a. During the subsequent warming away from the Younger Dryas, estimated winter temperatures rose from 1.8 to 10.1 °C and summer temperatures from 7.4 to 15.2 °C (figure 8; table 5). The total change peak-to-peak is registered in 7700 years, giving a mean rate of warming of about 1.1 °C/1000 a. Most of the change (~6 °C) is recorded within 500 years, giving rates of change of over 12 °C/1000 a.

† Roche, McIntyre & Imbrie (1975) reported palaeotemperature estimates on coccolith assemblages in the upper sections of cores RC9–228 and V23–81. Recent work by one of us (A.M.) shows that these estimates were several degrees Celsius too high, in part because of the lack of a cold end-member species. Work is underway to try to incorporate such an end-member into a new transfer function equation.

In summary, peak-to-peak rates of warming and cooling for North Atlantic surface waters ranging from 1 to 5 °C are registered in these cores for three intervals of relatively rapid climatic change. During shorter intervals, the recorded rates ranged as high as 13 °C/1000 a.

Inferred rates of polar front migration

To this point, we have documented down-core changes in estimated sea-surface temperature at selected points in space. From a geographic perspective, these changes also reflect large-scale horizontal migrations of water masses and frontal systems. In the North Atlantic, these migrations generally occur along the NW/SE axis documented by Ruddiman & McIntyre (1973) for the last deglacial warming (figure 9).

The most prominent feature in this complex march of isotherms back and forth across the North Atlantic is the polar front (convergence). Its position varies from the full-glacial position along 41° N to 44° N documented by CLIMAP (McIntyre, Moore *et al.* 1976) to the modern location as a coastal current restricted to the continental shelf of Greenland (figure 9). The intermediate positions shown in figure 9 have been defined by reconstructions at the levels of two ash layers deposited at 9300 a B.P. and roughly 65000 a B.P. (Ruddiman & Glover 1972, 1975, in preparation).

These geological reconstructions suggest that although the intensity (thermal gradient) of the polar front has varied with time, it has retained the strongest gradient in the subpolar North Atlantic. Consequently, passage of this front across a given core site will constitute the dominant thermal event within any general climatic cooling or warming. As a result, these areal migrations will leave in the temporal record of any one core short intervals of rapid temperature change imposed upon general periods of more gradual cooling or warming.

We suggest that the most rapid rates of temperature change calculated for the three climatic intervals in this paper occurred during periods of passage of the polar front. Such rates may be locally very rapid while the response in adjacent waters is slower. To develop an integrated geographical perspective, we have thus examined the rates of horizontal migration of the polar front in the subpolar North Atlantic.

As in the previous procedures for studying temporal rates of change, we have broken the spatial rates of change into two categories: (1) the average long-term rates of change from one climatic extreme to the other, and (2) the faster short-term rates prevailing during more rapid portions of the climatic transitions.

Three parameters must be established to measure recorded rates of polar front migration during the three climatic episodes we are examining: (1) a reference axis along which to measure the migrations; (2) polar frontal positions along the reference axis during the periods of interest, and (3) the timing (absolute or relative) of changes between these positions.

Choice of a reference axis

The polar front moves in part as a line hinged in the western North Atlantic and sweeping out an increasingly broad arc towards the northeastern North Atlantic (figure 9). In the western North Atlantic southeast of Newfoundland, movement of the front has been contained between the continental margin and the relatively stable western boundary current of the subtropical gyre. As a result, rates of migration in the western Atlantic have been relatively slow. To define maximum rates of migration, we chose as a reference axis a great circle line from Cape Farewell, Greenland, to the northwestern tip of the Iberian Peninsula, a distance of about 2800 km (figure 9). Frontal movements along this axis in the northeast Atlantic were particularly large and the rates of migration very rapid.

Geographic positions of the polar front

Geographical positions at the climatic extremes are relatively easy to establish for the three climatic changes we are studying†. Polar faunal percentages of 90 % or higher are found during the cold episode in stage 7 and in the glacial levels at the beginning of the stage 6/5 and 2/1 warmings. These polar-dominated faunas indicate polar front locations well southeast of the cores in this study, and probably close to the glacial position at 18000 a B.P. defined by CLIMAP (McIntyre, Moore *et al.* 1976). These inferred extreme glacial positions near the coast of the Iberian Peninsula are shown in figure 9.

The location of the Younger Dryas polar front is bracketed north of cores K708-4 and V23-82, slightly south of RC9-225 and V-23-81, and very near the location of core RC9-228. Other data show that it also lies slightly north of cores K708-7 and K708-8. This evidence seems to demand the serpentine twist to the Younger Dryas polar front shown in figure 10.‡ This is a signficant departure from the usual pattern of advance and retreat shown in figure 9.

The extreme northwestern retreat positions of the polar front during warm interglaciations are also reasonably easy to infer. We have already documented the polar front positions at 11000 and 6000 a B.P. (Ruddiman & McIntyre 1973). Work in progress with the CLIMAP project has shown that the polar front in isotopic stage 5e must have been restricted to a band on the East Greenland Continental Shelf even thinner than that during stage 1 (figure 9). The stage 7 interglaciation was not so intense as that in stage 5e; we infer that the polar front positions both below and above the cold episode were close to the modern position (figure 9).

We do not have direct evidence of where the polar front was located during most of the shorter intervals of more rapid temperature change registered in these northeast Atlantic cores. We do, however, know one of these positions positively – the deglacial retreat position at 9300 a B.P., shortly after the Younger Dryas readvance. This position is based on the zone 1 ash reconstruction of Ruddiman & Glover (1975). The 9300-year datum in the northeast Atlantic cores occurs at roughly the level of 75 % warming from the glacial stage 2 maximum to the interglacial stage 1 maximum. The polar front retreat at this point was roughly 90 % complete (figure 9).

Based on the polar front position at this important datum, we have assumed that during any other advance or retreat in which these northeast Atlantic cores show temperatures roughly 75 % toward the interglacial extreme, the polar front location was also (at least) 75 % toward the interglacial extreme near Greenland. To calculate the short-term migration distances, we thus multiplied the total migration distances (peak-to-peak between the climatic extremes) by 75 %.

† Faunal data from 15 additional cores were used to supplement the inferences on polar front locations made here.

‡ This position predicts interesting consequences for the intensity of Younger Dryas cooling on the British Isles, pointing to a much larger downwind cooling in the maritime climate over Scotland than in the southern British Isles. It also indicates that warm North Atlantic Drift flow into the Norwegian Sea ceased during the Younger Dryas.

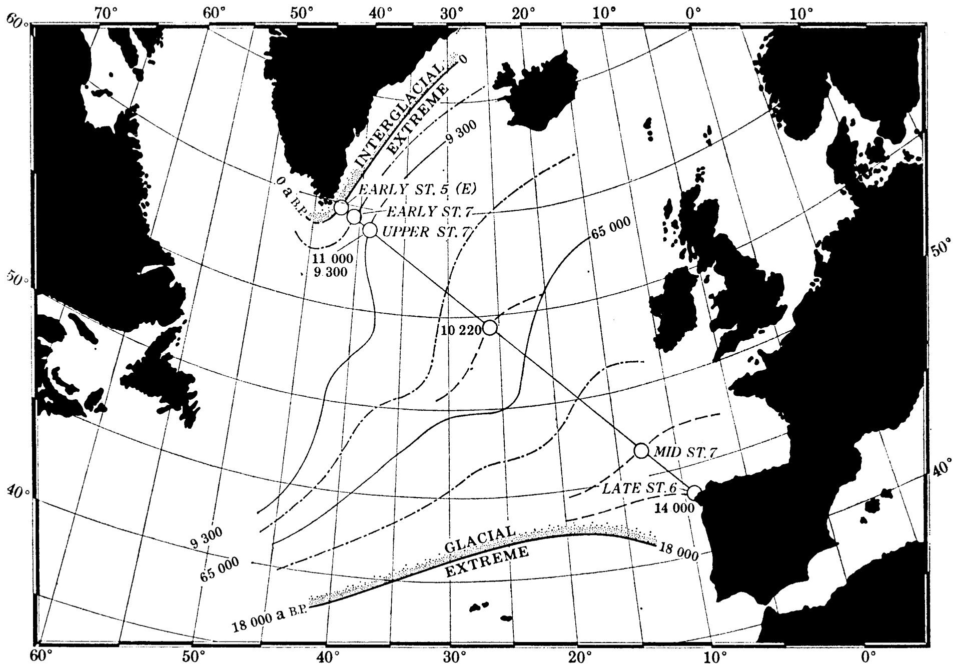

FIGURE 9. Past migrations of the North Atlantic polar front. Straight line is reference axis along which migrations are measured. Curving lines give general advance and retreat pattern of the polar front, including specific positions during the climatic intervals under study. Circles show intercepts of front with reference axis. Positions are keyed as follows. For the cold episode in stage 7 (*ca.* 195000 a B.P.), the positions 10000 a before, at, and 5000 a after the cold peak are shown as early st. 7, mid st. 7 and upper st. 7. For the stage 6/5 boundary warming (*ca.* 127000 a B.P.), the last cold level just prior to the change is shown as late st. 6 and the warmest level at the end of the change as early st. 5 (E). Positions during the stage 2/1 deglacial warming and Younger Dryas cold episode are specified in absolute ^{14}C a B.P.

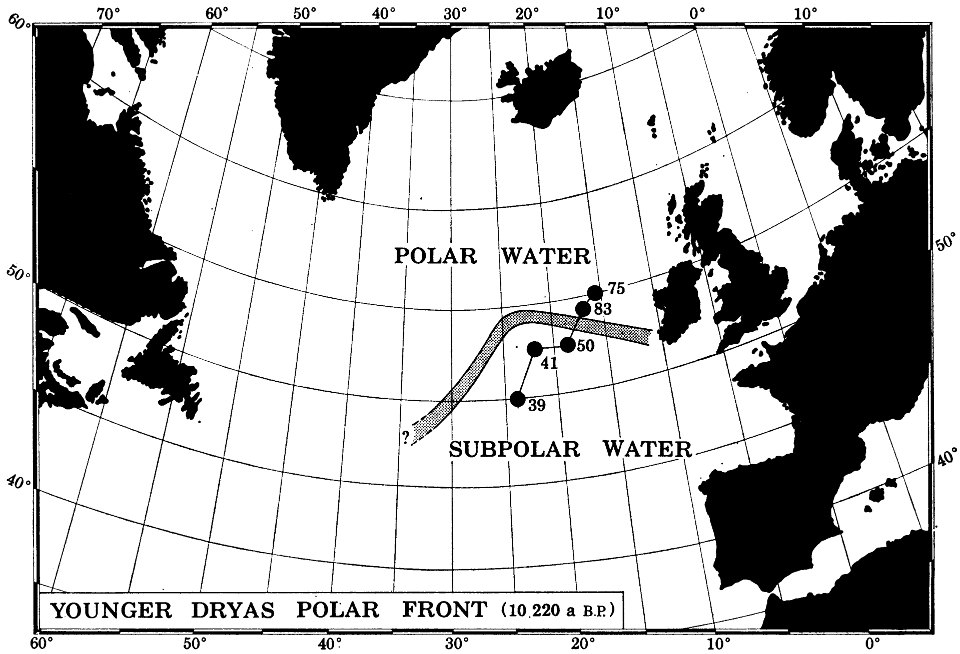

FIGURE 10. Location of the Younger Dryas Polar Front (stippled pattern). Percentages of polar fauna (*G. pachyderma*, s.) as total of planktonic foraminifera larger than 149 μm at Younger Dryas cold peak are shown for five cores also plotted in figure 7.

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In actuality, the front was probably located even farther to the northwest at these times; thus the short-term rates calculated here for rapid migrations early in the climatic warmings and late in the coolings are probably minimum estimates.

Timing of migrations

The time of occupation of these positions in either an absolute or relative time sense is more difficult to establish. For the time of occupation of extreme polar front positions during the stage 2/1 warming and the Younger Dryas cold episode, we have used the ages published by Ruddiman & McIntyre (1973). Even these ages are to some extent linear interpolations and extrapolations between available ^{14}C control levels.

Table 6. Rates of migration of the polar front during three climatic episodes

	migration distance/km	time/a	rate of migration/(m/a)
mid-stage 7			
cooling	2000 (1500)	9200 (4400)	220 (340)
warming	2100 (1575)	5800 (2900)	360 (540)
stage 6/5 boundary			
warming	2700 (2025)	7000 (1700)	390 (1190)
stage 2/1 boundary: entire transition			
warming	2300 (1725)	10100 (2800)	230 (620)
stage 2/1 boundary: separate portions			
initial deglacial warming	2000 (1500)	2700 (1500)	740 (1000)
Younger Dryas cooling	800 (600)	700 (400)	1140 (1500)
later stage 1 warming	1100 (825)	7700 (500)	140 (1650)

No absolute dating technique can measure events 100000 to 200000 years old with the accuracy necessary to detect advances and retreats no more than a few thousand years in duration. For the stage 7 and stage 6/5 climatic changes, we have thus made the simplifying assumption that the down-core records already examined in these northeast Atlantic cores can provide us with the necessary information on the relative timing of watermass migrations elsewhere in the subpolar Atlantic. We have assumed that each southeasternmost penetration of the polar front is time-correlative with the highest percentages of polar fauna and the coldest temperature estimates in the cores (table 6). Similarly, we have assumed that the levels of warmest temperature estimates in the northeast Atlantic cores equate in time with the northwesternmost retreat locations of the polar front. These data enable us to calculate net rates of change across each entire climatic shift (table 6).

This approach to calculating rate estimates leaves uncertainties caused by the lack of strongly definitive warm peaks in many cores (as discussed previously). Their placement in time can be uncertain to within as much as a few thousand years. To circumvent this problem, we also chose the average levels of 75 % interglacial warmth in the temperature curves of each climatic transition. These easily picked time levels were assumed to equate to the periods when the polar front was located 75 % towards the interglacial extreme position. From these, we calculated the short-term rates of migration recorded in table 6 in parentheses. These faster rates tended to prevail early in the warmings and late in the coolings.

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Rates of migration

First, we have calculated peak-to-peak rates for each climatic transition. During the cooling episode in stage 7, the polar front advanced a total distance of 2000 km southeastward in 9200 years and then retreated 2100 km northwestward in 5800 years.† The corresponding mean rates of migration for the peak-to-peak changes are 220 m/a (= km/1000 a) for the entire initial cooling and 360 m/a for the full later warming (table 6). During the more rapid stage 6/5 warming, the polar front retreated 2700 km in 7000 years, for a mean retreat rate of 390 m/a. There was a total retreat of 2300 km in 10000 years during the entire stage 2/1 warming; this gives a mean retreat rate of 230 m/a. In short, recorded advance and retreat rates of the polar front from one climatic extreme to the other averaged from 200 to 400 m/a and were sustained over several millennia.

The very brief Younger Dryas cooling produced a total southeastward polar front advance of 800 km in 700 years, followed by a northwestward retreat of 1100 km in 7700 years. The advance rate averaged 1140 m/a; the retreat occurred at a mean rate of 140 m/a.

The recorded rates of migration during selected portions of each of these climatic changes were considerably faster. During the last 4400 years of the cooling episode in stage 7, the front advanced 1500 km at a mean rate of 340 m/a. During the initial warming out of this cold peak, there was a retreat of 1575 km in 2900 years, a rate of migration of 540 m/a. The extremely rapid warming early in the stage 6/5 transition produced a polar front retreat of 2025 km in 1700 years, an average rate of 1190 m/a. The entire stage 2/1 warming began 14000 a B.P. and was 75 % completed by 11900 a B.P. During this period, the polar front retreated 1725 km, for a mean rate of 620 m/a for the fastest part of the warming.

As for the separate advances and retreats, the deglacial retreat early in the stage 2/1 warming carried the polar front northwestward some 1500 km in 1500 years; this gives a mean retreat rate of 1000 m/a. The very brief Younger Dryas cooling produced a southeastward polar front advance of 600 km in 400 years, followed by a northwestward retreat of 825 km in 500 years. The advance rate averaged 1500 m/a; the early retreat rate averaged 1650 m/a.

In summary, polar front advances and retreats in the more rapid portions of these large climatic changes are recorded at average rates of 300–1600 m/a over periods of a millennium or more.

Significance of vertical mixing in blurring the actual rates of change

In the North Atlantic, vertical mixing is the greatest single impediment to the exact reconstruction of oceanic palaeoclimates. Infauna burrow into the uppermost 5–30 cm of sediment, which in the Atlantic usually contain a record of climatic change spanning anywhere from 3000 to 15000 years. The vertical exchange of sediment that results from this burrowing has the effect of blurring and smoothing the climatic record. The amplitudes of climatic oscillations are reduced, with the greatest losses of original amplitudes occurring for the shortest (highest frequency) oscillations. Prominent climatic transitions that were originally very abrupt may be smoothed into considerably more gradual forms. All of this reduces the oceanic rates of change recorded in the sediments.

The problem of quantitatively modelling vertical mixing in order to remove its effects is complex, and work on this is presently underway. The purpose of this paper is to document as

† Mean times are taken from calculations in last section and represent an average value for all cores examined.

carefully as possible the rates of change preserved in the sediments. When precise quantification of mixing is available, these oceanic data will then be 'unmixed' in a quantitative way. Until then, the actual (more rapid) rates of change must largely remain a matter of conjecture, as must detailed comparisons to those portions of the continental record lying within the range of ^{14}C dating. In particular, we see the need for work to improve the oceanic and continental records in the northeast Atlantic and on Great Britain. This is one of the few areas of the world where a really detailed oceanic record of high quality is available near a continent whose climate is dominantly maritime.

Summary and conclusions

1. We examined several northeast Atlantic cores for evidence of rates of change during three prominent palaeoclimatic shifts during the last 200 000 years.

2. During a cooling episode midway through interglacial isotopic stage 7, the northeast Atlantic cores registered virtually a complete interglacial-to-glacial-to-interglacial oscillation in no more than 15 000 years.

3. At the isotopic stage 6/5 deglaciation (termination II), a full glacial-to-interglacial oceanic warming was recorded in no more than 6500 years; 75 % of the shift was registered in less than 2000 years.

4. The slower overall deglaciation at the isotopic stage 2/1 boundary was related to a short reversal of deglacial warming centred at 10 200 a B.P.; during this Younger Dryas cooling, the North Atlantic very briefly returned to a nearly full-glacial state.

5. The cores register average rates of change in estimated surface water palaeotemperature of 1–5 °C/1000 a for the entire peak-to-peak transition; the rates range from 2 to 13 °C/1000 a during shorter intervals of faster change early in the warmings and late in the coolings.

6. The recorded rates of polar front movement along an axis running from Cape Farewell, Greenland, to the Northern Iberian Peninsula average 200–400 m/a for the several thousand years of change between the climatic extremes; the rates range as high as 400–1650 m/a during shorter intervals of more rapid change.

7. Further work to remove the mixing effects of burrowing organisms will be necessary to derive the actual (faster) rates of oceanic surface water change.

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PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

Discussion

G. Manley (3 *Whitwell Way, Coton, Cambridge*). This establishment by Professor McIntyre and his colleagues of the characteristics of the North Atlantic in glacial times is indeed gratifying as it confirms the general scheme that I presented in the *Geographical Journal* (vol. 117, p. 62) in 1951 as a likely explanation on 'best-fit' principles, of what was at that time known about the climate. The further extension of the cold water as far south as the latitude of northern Spain that he has demonstrated is helpful with regard to mountain glaciation and the vegetation of western France. We now need to elucidate the factors – meteorological or oceanographical, or both – that led to the renewed forward surge of warm mid-Atlantic surface water past Ireland into the Norwegian Sea. Such surges provide an economical way of explaining many minor climate episodes. The squeezing of the subpolar and transitional water masses that he has shown brings to mind the present east–west banding of the surface water masses in the North Pacific about 45°–50° N.

A. Dreimanis (*Geology Department, The University of Western Ontario, London, Canada*). McIntyre reported the beginning of the late-glacial warming of northern Atlantic Ocean about 13500 a ago, while the terrestrial record from Britain, as indicated by the beetle studies (Shotton, Coope) suggests a delay of the beginning of this warming episode to about 13000 a B.P. Farther inland, in the Baltic region, the late glacial warming began again at least several centuries prior to 13000 a B.P., as indicated by some interstadial sites. The most detailed palaeobotanical investigations have been published in the early sixties from the Raunis Interstadial, northern Latvia, and subsequently three ^{14}C dates have been obtained on it: 13390 ± 500 (Mo-196), 13250 ± 160 (TA-177), and 13320 ± 250 (Ri-39) a B.P. Of a similar age, more than 13000 ^{14}C a B.P., is also the Meiendorf Interval in northwestern Germany, and the Vintapper Interstadial of southern Sweden.

In the Great Lakes Region of North America, the Cary-Port Huron or Mackinaw Interstadial culminated 13300 a B.P., and several of its correlatives, or at least rapid glacial retreats began prior to 13000 a B.P., in other areas of North America.

If a more or less synchronous climatic warming began prior to 13000 a B.P. in the Atlantic ocean and widely scattered areas over the continental Europe and North America, a question arises of why the beginning of the Late-Glacial Interstadial was delayed in Britain for several centuries.

H. H. Lamb (*Climatic Research Unit, University of East Anglia, Norwich NR4 7TJ*). What chances are there of obtaining further data points for the ocean surface temperatures in ice-age times in the Atlantic nearer the European seaboard? This area must be of great interest in connection with the climates prevailing in the British Isles and all western and northern Europe.